I. GEOLOGY OF THE WINCHESTER-HEMET AREA, RIVERSIDE COUNTY, CALIFORNIA

II. GEOCHEMICAL INVESTIGATIONS OF AN ARKOSIC QUARTZITE OF THE WINCHESTER-HEMET AREA, CALIFORNIA

Thesis by

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ABSTRACT

I.

An area of about 100 square miles located 20 miles southeast of Riverside, California, has been geologically mapped. The oldest rocks of the area are phyllites and quartzites correlated with the Bedford Canyon formation (Triassic). Conformably overlying these rocks is a 13,000 foot thick section of quartzite, schist and rare amphibolite, here named the French Valley formation. The metasedimentary rocks were formed from a series of shales, shale-clast conglomerates, poorly sorted feldspathic, calcareous and arkosic sandstones and rare basalt flows or tuffs. Relict sedimentary structures and textures are locally well preserved.

The sedimentary rocks were intruded with a series of basic and ultrabasic magmas and were tightly folded prior to late Cretaceous time. Late in the period of folding they were metamorphosed to the hornblende-hornfels facies (200°-600°C, 3-5 kilobars), producing andalusite, cordierite, sillimanite and garnet in the schists and hornblende-plagioclase (-diopside, -garnet) assemblages in the amphibolites. Zones of progressive metamorphism are mapped trending N-S and cutting the NW-SE structural trend of the folded rocks. In the late Cretaceous these rocks were intruded by basic to intermediate plutonic igneous rocks of the southern California batholith, with no appreciable contact metamorphism.

II.

A series of seven samples of arkosic quartzite from a single homogeneous member of the French Valley formation has been collected along a profile of increasing metamorphic rank. The component minerals and aliquots of the total rocks have been analysed by X-ray fluorescence, emission spectrographic and wet chemical procedures. For most elements studied, concentrations in the total rock, biotite and ilmenite appear to be constant through the series. These elements either did not vary beyond the experimental uncertainty or they varied without apparent trend over a small range slightly exceeding the uncertainty. With increasing degree of metamorphism the following changes were suggested but fell within or only slightly exceeded the assigned limits of error: a) the total rock gained CaO, Mn, Sr, Co, La, Y, and Ti(?) and lost K_2O, Na_2O, and V; b) the biotite gained Mn, Ti, Ba, and possibly Co, Cr, V, and La and lost CaO; c) ilmenite gained Mn and Ni and lost Cr, Ba, Zr(?), Y, and Pb. Other changes observed with increasing metamorphic rank were: a) coarsening and homogenization of the rock texture; b) decrease in the ratio Fe^{+++}/Fe^{++} in both rock and biotite; c) decrease in 2V and basal spacing of the biotite; and d) variations in albite content of plagioclase and potash feldspar that suggest loss of Na_2O from rock.

For most elements fractionations between biotite and rock appear to vary directly with the concentration in biotite. Fractionations between ilmenite and biotite are shown to be related to ionic size and charge.

The composition of the rock is such that partial fusion would yield a liquid with a composition close to that of the total rock. This may be the cause of the small variations in major element composition with metamorphic grade.

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PART I. GEOLOGY OF THE WINCHESTER-HEMET AREA, RIVERSIDE COUNTY, CALIFORNIA

Plate 1. Frontispiece

Domenigoni Valley looking south from Searls Ridge

I. INTRODUCTION

A. GEOGRAPHIC FEATURES

1. Location and Size

The Winchester-Hemet area is located in Riverside County, California about 20 miles southeast of the city of Riverside. The boundaries of the area are defined on the north and west by the north and west edges of the Winchester quadrangle (U. S. Geological Survey topographic map, 1:24,000, 7.5', 1953). On the south and east the area overlaps into parts of the Bachelor Mountain, Sage and Hemet quadrangles (see fig. 1). The total enclosed area is 92 square miles of which about one third is underlain by alluvium.

2. Topography

The area of study can be divided in terms of topographic features into two parts. The southeast quarter of the area, in the vicinity of Rawson Ranch, is a series of rounded hills and ridges, interspersed with small flat-bottomed valleys. The hills rise to elevations of 3000 feet, dominated by Black Mountain (3030 feet) near the southeast corner of the area. North and west of this high country the area consists of isolated hills and ridges surrounded by an alluvial plain. The plain is at an elevation of 1500 ± 100 feet, with a gentle slope to the west. The ridges and hills are up to 2200 feet in elevation. Adjoining the area on the northwest are the Lakeview Mountains, surmounted by a broad undulating surface at an elevation of about 2100 feet.

3. Culture

The area includes the towns of Winchester, Homeland and parts of the town of Hemet. State highway 74 passes along the north side of

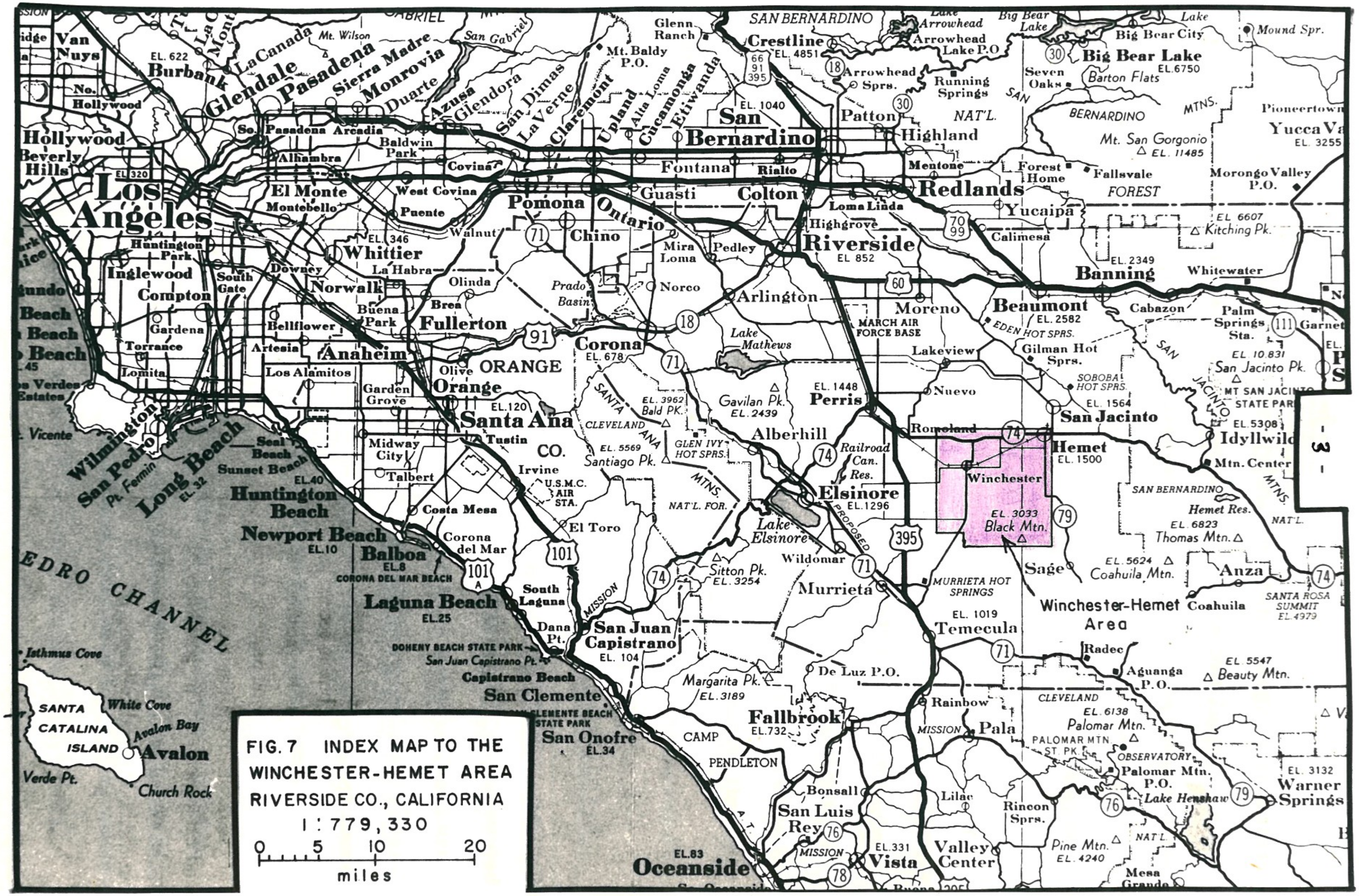

FIG. 7 INDEX MAP TO THE WINCHESTER-HEMET AREA RIVERSIDE CO., CALIFORNIA

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the area. A spur of the Santa Fe railroad passes through Winchester and terminates in Hemet. The alluvial plain of the San Jacinto Valley is the site of extensive agricultural development, both in livestock and produce. Fruit is one of the most important crops in the vicinity of Hemet. The hills, because of their low water supplies are not generally suitable for grazing.

4. Climate

The climate of the area is typical of most of southern California west of the mountain ranges. The average rainfall is about 10 inches. Maximum summer temperatures are about 110° and minimum winter temperatures are about 20°. Rainfall is almost entirely restricted to the months of October to April, except for very rare summer thunderstorms.

5. Vegetation

The alluvial plain is largely cultivated. The ridges and hills are generally only sparsely covered with chapparral and grass. Locally, however, very dense growths of brush up to 10 feet high have prevented the author's access to several geologically critical areas. The more level parts of the highland around Rawson Ranch are commonly grass-covered and have a thin soil cover. Scattered clumps of oak are found on this highland, especially along the Rawson Ranch fault zone.

6. Exposures

About one third of the area is underlain by alluvium overlying various types of bedrock. Of the remaining two-thirds, a considerable fraction consists of areas of low-relief and very poor exposure, underlain principally by igneous rocks and low-rank metamorphic rocks.

Much of the latter portion of the area is actually cultivated and exposures are restricted to small knobs scattered through ploughed fields. Exposures are fair to excellent on Double Butte, Searls Ridge and in the general vicinity of Rawson Ranch (see inset on lower right of fig. 1). In the last named area, comprising the bulk of exposures of metamorphic rocks in the Winchester-Hemet area, the quality of exposures varies radically from place to place. In the areas of relatively low relief, such as near and in Crown Valley, the outcrops are deeply weathered and commonly covered with a layer of soil several feet thick. Here again one must rely on scattered isolated knobs. Elsewhere, on the flanks of the higher and more precipitous topographic features exposures are commonly continuous over wide areas but deeply weathered, so that, for the purposes of sampling of rock types it is necessary to seek exposures on bold cliff faces or in stream bottoms.

7. Accessibility

The area is traversed by many roads and no part of the area is more than 1 mile from a usable road. The major paved roads, State Route 74, Winchester road, Newport road (partly unpaved) and Florida Ave. remain for the most part on the alluvial plain. Farm roads of variable quality are found in the vicinity of Rawson Ranch, to the south in Auld Valley, and along the base of Bachelor mountain.

B. PURPOSE OF THIS INVESTIGATION

The writer has dealt with the geology of the Winchester-Hemet area from two interrelated aspects: first as a location for study of the stratigraphic, structural and metamorphic history of the Peninsular range prior to introduction of the southern California batholith (Part I);

and secondly as a location for study of the geochemistry of progressive regional metamorphism of a homogeneous metasedimentary unit (Part II). The detailed purposes of the geochemical research undertaken are explained in the introduction to Part II.

From earlier work by Larsen (1948) and others it is known that the igneous rocks of the southern California batholith intruded a series of metasedimentary rocks of Mesozoic age. Some detailed work (see below) on these metasediments has been done, especially in the Santa Ana mountains, but there and in other areas of exposures of metamorphic rocks intruded by the batholith it has been found difficult or impossible to work out the detailed picture of sedimentation, metamorphism and pre-intrusive structure. The difficulties have been largely due to poor exposures and dense brush. Distinctive lithologic marker units of appreciable lateral extent are rare and the exposures are generally isolated septa or inclusions in the surrounding igneous rock complex.

The Winchester-Hemet area was outlined by Larsen as containing one of the largest continuously exposed blocks of metasedimentary rock in the batholithic province. Some question was left by Larsen as to the age relations of the metasediments exposed here to those described elsewhere in this province. It nevertheless appeared to the writer that this would be a critical locality in which to study the pre-intrusive history of the province.

Among the specific problems which have been considered are the following:

1.) What was the nature of the original sediments from which the present metamorphic rocks were derived and what light does their composition shed on the paleogeography of the Peninsular ranges?

2.) What were the P-T conditions of metamorphism?

3.) What was the deformational history of these sediments and how does it relate to their metamorphism?

4.) To what extent did the intrusion of the southern California batholith influence the distribution of metamorphic zones and disturb the structure of these sediments?

In addition, problems of more local import will be discussed in the light of new evidence collected in this study. These include: 1.) the nature of the contact between the Bedford Canyon formation and Larsen's "Paleozoic schists and quartzites" and the relative ages of these two rock units; 2.) the possible significance of Webb's (1939) fossil-locality at the Hemet magnesite mine; 3.) the geology of the Hemet magnesite mine; and 4.) a reconsideration of the relative ages of some of the igneous rock types described by Larsen.

C. PREVIOUS WORK

The first fairly detailed studies of the metamorphic rocks of this area were those of Fraser (1931) who studied those parts of the area included in the Hemet quadrangle and Dudley (1935) who studied the crystalline rocks of the Perris block as far east as Double Butte. In both cases the petrology of these rocks was dealt with in very general terms.

Webb (1939) reported the occurrence in this area of a fossil of Paleozoic age, and suggested this age for the "older" metasedimentary rocks of this area.

The metamorphic rocks of this area were later studied by W. J. Miller in his compilation of crystalline rock types of southern California (1946). Miller concluded that the Diamond Valley series (included in the French Valley formation of this report) was probably distinct from and older than the Santa Ana metamorphic series, now known as the Bedford Canyon formation.

Finally, in 1948, Larsen published a description of parts of the batholithic province of southern California including most of the Winchester-Hemet area. He assigned the general name "Paleozoic schists and quartzites" to the apparently older metasedimentary rocks of the eastern part of this area, accepting Webb's age determination as valid for the whole series.

Related work on similar metamorphic rocks has been carried on in localities west and south of here. See, for example, Moore (1930), Engel (1943, 1949) and Gray (1954) on the Santa Ana Mountain metamorphic rocks and Hudson (1922), Merriam (1946) and Everhart (1951) on the metamorphic rocks of the Julian-Cuyamaca area.

D. METHODS

The area was mapped on a scale of 1:24,000 using as a base the Winchester quadrangle and parts of the Bachelor Mountain, Sage and Hemet quadrangles, all U. S. G. S. topographic sheets, 7.5', 1953. Aerial photographs were used only for location purposes. Geologic features could not be detected on them in general, although outcrops are extensive. About 100 field days were spent in preparing the geologic map. The Hemet magnesite mine was mapped on a scale of 400 feet = 1 inch using as a base an enlarged aerial photograph of the mine area. The information was then transferred to a correspondingly enlarged and slightly modified portion of the topography of the Winchester quadrangle.

About 250 thin sections and grain mounts were prepared from rocks of this area. In addition minerals of many rocks were studied by oil-immersion techniques. Index of refraction measurements were made with Na_D light, using immersion oils calibrated, if possible, with an Abbe refractometer. Most determinations are valid to ± 0.002. Plagioclase was generally determined by use of the Michel-Levy extinction angle method, determining λ^+. In a few cases, where twinning was rare, indices of refraction on cleavage flakes were determined and the composition found from the curves drawn by Tsuboi (Winchell and Winchell, 1951, p. 280). Optical data on pyroxene, hornblende, anthophyllite, epidote, andalusite, and chlorite were related to composition using the data of Winchell and Winchell (1951). Garnet compositions were determined (approximately) on the basis of a_o, index of refraction and density measurements, using the values of these parameters for the end members of the garnet series as given by Levin (1950). Density determinations were made in water in a 5 cc pycnometer, using traces of alcohol as a wetting agent, but all the values appear to be too low.

Where necessary, x-ray powder patterns were obtained to identify minerals for which optical data was equivocal. For this purpose a Debye-Scherrer-type camera with a radius of 57.2 mm. was used. Interplanar spacings were measured using a calibrated plastic strip and "d" values were checked against ASTM file cards. For determination of a_o of garnet, higher order reflections were used and calculated back to the unit cell dimensions. The reflections were indexed by comparison with the indexed pattern for grossularite given by Yoder (1950).

A study of sillimanite was made using infra-red absorption data to attempt to identify possible mullite solid-solutions. The sillimanite was separated from its host rock using heavy liquids and a Franz isodynamic separator. It was ground in an agate mortar to -250 mesh size. The powder was mulled with nujol and spread between two NaCl disks. The mull was analysed on a Perkin-Elmer Model 12-A infra-red absorption spectrophotometer, with an NaCl prism, in the wavelength region 8-16 micra. The absorption patterns were compared with those obtained by Roy and Francis (1953) and with patterns from known samples of mullite and sillimanite. The same unknown samples were analysed by x-ray diffraction, using the data of G. C. Kennedy of the Institute of Geophysics, University of California at Los Angeles (pers. comm., 1959) to determine the degree of solid-solution of mullite in the sillimanite.

Universal stage measurements were made to determine quartz C-axis orientations in two thin sections. A Leitz 4-axis universal stage was used for this purpose.

Two magnetometer traverses were made across Domenigoni Valley, with the assistance of Mr. T. N. Irvine. For this purpose a Sharpe Dl-M Reconnaisance Magnetometer was used. The vertical component of the magnetic field was determined.

E. EXPLANATION OF THE GEOLOGIC MAP

1. Stratigraphic Keys

In the southeast and south-central part of the area near the type locality of the French Valley formation it is possible to trace individual lithologic members of this formation for great distances. In this area the formation has been divided into three members and 18 subunits. A typical map key is of the form "fvB3" where "B" refers to the member and "3" to the subunit.

Lithologies exposed further to the east cannot always be correlated with these numbered subunits because metamorphism has obliterated distinguishing textural and compositional features, outcrops are less abundant and, to an undefined extent, facies changes have occurred along strike. In such areas the general map key will be "fvAq" where "A" again refers to a member and "q" (or "s", "am", etc.) refers to a lithology described in the key on the map sheet. In this way the question of possible time-equivalence of subunits of differing primary sedimentary lithology has been left open where evidence is uncertain or lacking.

Additional keys of the usual type are described on the map sheet.

2. Progressive Metamorphism

An overlay to the geologic map has been prepared (fig. 2) showing the distribution of critical metamorphic mineral assemblages through the area, as well as other evidence pertaining to the metamorphism. With one exception the author has refrained from incorporating this data into the geologic map itself. The one exception is the mapping of migmatitic gneiss as a discordant zone cutting across many of the pelitic schist units. Even here an attempt has been made to correlate stratigraphy through the zone of migmatization by following quartzite markers.

3. Contacts

Five types of contacts are distinguished on the geologic map. The solid and dashed lines are as usually defined. The dotted line contact has been used to indicate three types of contact relations:

a.) contacts buried by alluvium or other units

b.) contacts under cultivated land

c.) contacts in areas made inaccessible by heavy brush.

Near Black Mountain and south and west of Goodhart Canyon the geologic contacts are based in part on inferences from observations made on aerial photographs. Such contacts are labelled by a "P" along the contact.

4. Coloring of the Map

The coloring of the metasedimentary map units is not based on a stratigraphic code but is rather based on the inferred lithologic composition of the parent sedimentary rock. Where metamorphism has obliterated primary sedimentary features the colors are used as a more generalized indication of primary lithology and should be understood to be highly inferential in such cases.

E. DEFINITION OF SOME TERMS USED IN THIS THESIS

In general the definition of geologic terms used in this thesis is in accord with the definitions given in "Glossary of Geology and Related Sciences" published by the American Geological Institute, 1957. The meaning of some specific words as used in this thesis may give rise to doubt and are therefore defined below:

"a" tectonic axis: direction of tectonic displacement of matter; more generally, a line tangential to a fold and perpendicular to the fold-axis.

"b" tectonic axis: a line parallel to the axis of a fold; the axis of local rotational movement in folding.

fibrolite: sillimanite occurring as clusters of very thin fibers.

grain size:

- in metamorphic rocks:
  - very fine: less than 0.01 mm. average grain size
  - fine: 0.01-0.10 mm.
  - medium: 0.10-0.50 mm.
  - coarse: 0.50-5.0 mm.
  - very coarse: greater than 5.0 mm.
- in igneous rocks:
  - fine: less than 1 mm. average grain size
  - medium: 1-5 mm.
  - coarse: greater than 5 mm.

pyralspite: a garnet whose composition is on the pyrope-almandine-spessartite side of the miscibility gap of the field of naturally occurring garnets.

ugrandite: a garnet whose composition is on the grossularite-andaradite-uvarovite side of the miscibility gap in the field of naturally occurring garnets.

wacke: a poorly sorted clastic sedimentary rock consisting of subequal proportions of angular to subangular sand-sized grains and silt to clay size grains.

F. GEOGRAPHIC NAMES AND COORDINATES

In addition to the geographic names given on the various topographic maps used as base maps for this work, it has been convenient to refer to the following places by names not shown on these maps:

The ridge north of Diamond and Domenigoni Valleys is called "Searls Ridge".

The ranch located 0.3 miles southwest of Crown Valley is called "Rawson Ranch".

These names are shown on fig. 1.

For convenience in locating points of interest on the geologic map, and for economy of nomenclature, locations are indicated according to a grid system combined with the usual range-township-section notation. The townships have been abbreviated by letters A through D, as follows:

"A": T6S, R2W
"B": T6S, R1W
"C": T5S, R1W
"D": T5S, R2W

The sections are numbered as usual. Within each section one may imagine a 4x4 grid system set up, with the resultant subdivisions numbered as in Cartesian coordinates, starting from the southwest corner of each section. Thus a typical location designation will be "A16(3, 4)": this is equivalent to "the northwest quarter of the northeast quarter of section 16, T6S, R2W". This system of nomenclature is illustrated in fig. 8.

FIG. 8 Coördinate system used in this thesis.

Each section of the county land grid is divided into subunits as shown below:

(1,4)	(2,4)	(3,4)	(4,4)
(1,3)	(2,3)	(3,3)	(4,3)
(1,2)	(2,2)	(3,2)	(4,2)
(1,1)	(2,1)	(3,1)	(4,1)

II. STRATIGRAPHY OF THE LAYERED METAMORPHIC ROCKS

A. INTRODUCTION

Because the stratigraphy of this area has been little understood this section is principally an attempt to reconstruct the primary stratigraphic features, neglecting for the time being the metamorphic phenomena that are superimposed. A separate section will deal with the metamorphism.

All the metamorphic rocks of this area show some variation in rank of metamorphism as they are traced through this area. In this reconstruction principal use has been made of the lowest rank exposures, where primary sedimentary features are best preserved. Even these rocks have undergone extensive recrystallization, as will be seen.

Both primary sedimentary facies and metamorphic rank appear to vary along strike, and it is not always possible to distinguish the effects of these two types of variation. Consequently some of the deductions about regional stratigraphy are tentative. In particular one might cite two such effects of metamorphism: squeezing of the metasedimentary units during folding resulting in apparent thinning (and perhaps thickening) of these rock units; and weak metasomatism in the higher rank rocks obscuring original chemical compositions.

The stratigraphic section is divided into two major units, the older Bedford Canyon formation and the younger French Valley formation. The latter is in turn divided into three members called A, B and C, each of which is further subdivided into subunits for mapping purposes (see fig. 1).

B. BEDFORD CANYON FORMATION

1. Occurrence

Outcrops of the Bedford Canyon formation are scattered through the southwest corner of the Winchester quadrangle and across the northern edge of the Bachelor Mountain quadrangle. The outcrops are characteristically rounded hills of very low relief. Locally, where there has been contact metamorphism by igneous intrusions the Bedford Canyon formation rocks form higher hills. This phenomenon is best observed just to the west of the Winchester-Hemet area in the outcrops north of Menifee Valley. The lower hills are generally covered with sparse vegetation. In the Winchester quadrangle and to the west of here they are commonly covered with cultivated fields so that contacts between outcrop and alluvium are very obscure.

2. Name

The Bedford Canyon formation was named by Larsen (1948, p. 18) for its occurrence in Bedford Canyon on the north slope of the Santa Ana Mountains, approximately 30 miles from the nearest exposures in this area. The same formation as exposed in nearby localities in the Santa Ana block and across the Elsinore trough to the east, has been described and named by at least five other authors. Their descriptions of the formations they have named are essentially identical. They are probably dealing with overlapping portions of a very thick monotonous lithologic section. Briefly the history of the nomenclature may be tabulated as follows:

1893	J. P. Smith	Santa Ana slates
1930	B. N. Moore	Silverado formation
1933	E. M. Irving	Elsinore series
1935	P. Dudley	Elsinore series
1948	E. S. Larsen	Bedford Canyon formation
1950	R. Engel	Santa Ana formation

The last name is pre-empted, as pointed out by Larsen (1948, p. 19) and therefore Larsen's nomenclature will be used in this discussion since it is the next most recent.

3. Age and Correlation

The only paleontologic evidence indicative of the age of this formation comes from fossils in a few small limestone lenses which are interbedded with typical Bedford Canyon type rocks in the Santa Ana Mountains. The age of the fossils appears to be Triassic. Mendenhall assigned an age of upper Triassic on the basis of a fauna including the following genera:

Rhynconella
Spiriferina
Terebratula

Smith (1914, p. 145) called the limestones upper Triassic on the basis of a new species of the pelecypod genus Daonella, D. sanctae anae with affinities to D. böckhi of the Alpine Norian. Popenoe (1955) gives the most recent view on the subject as follows:

> "The limestone lenses from the upper part of the [Bedford Canyon] section have yielded a sparse fauna including the brachiopod genus Halorella and probably the ammonite genus Juvavites (S. W. Muller, pers. comm.). Muller considers these to indicate a Norian upper Triassic age for the containing beds."

Our ignorance of the regional structure prevents us from relating the fossiliferous horizons of the Santa Ana Mountains to that part of the section exposed in the area under study. For the purposes of this paper the age of the whole formation will be assumed to be upper Triassic. Larsen suggests that part of the formation may be Jurassic though he gives no evidence to support this except the great thickness of section above the fossiliferous horizon. In view of the eugeosynclinal character of much of the section it is likely that the sediments were deposited quite rapidly in a rapidly sinking basin and there are numerous examples of basins filled with tens of thousands of feet of such sediments within a single epoch.

The rocks indicated on the map as "Bedford Canyon formation" have been correlated with the formation described by Larsen purely on the basis of lithologic similarity. The correlation to this area was first made by Larsen. There are numerous possible objections to this correlation. First of all, one cannot trace the rocks of this formation in continuous outcrop from the type locality to the Winchester-Hemet area. Between these two sites are interposed the Elsinore fault of undetermined lateral displacement, several intrusive bodies and extensive areas of alluvial cover. A second objection is that, as will be shown below, there are appreciable differences in the lithology of the Bedford Canyon formation as exposed in the Santa Ana Mountains and its exposures in this area. A third objection might well be that Larsen's original "definition" does not truly define the formation in the usually accepted sense. Larsen gives no thickness for the formation, describes neither the upper nor lower contacts of the formation and does not give a columnar section for the type locality.

In consideration of the first objection stated above, one may appeal at best to the need for obviating unnecessary confusion in the nomenclature of a relatively unknown area. Admittedly the numerous structural, intrusive and stratigraphic barriers interposed between the various exposures of the "Bedford Canyon formation" introduce a measure of doubt as to its unity. But, on inspecting widely dispersed localities of this formation the writer has found no clear basis on which to suspect that what he has seen does not represent what was once a single continuous stratigraphic unit of approximately uniform or gradational lithology. It appears that, prior to the intrusion of the southern California batholith the region of the Peninsular Ranges was underlain by a sequence of slates and quartzites which extended over the entire area of present exposure of the Bedford Canyon formation. The fact that this sequence has subsequently been isolated into structural units with abrupt discontinuities between them does not justify assigning a separate stratigraphic name to the rocks of each unit. It is hoped that in the future a detailed picture of the stratigraphy of this area may be arrived at. In the meantime the similarities in lithology which are self-evident in the diverse localities should be sufficient justification to group all these exposures into a single formation.

In regard to the second objection, namely the appreciable differences in lithology from the type locality, it seems that the lithologies of the two areas are in fact gradational into one another and the differences are simply the result of gradational changes in environment of deposition and source material. Again the interruptions in exposure between the two localities prevent one from establishing this gradation with certainty.

Further study of the formation, especially comparing its exposures on opposite sides of the Elsinore trough, should be undertaken to test this hypothesis.

The third objection, perhaps the most serious in the strict sense of stratigraphic nomenclature, is probably an unavoidable one. In accepting Larsen's nomenclature over that of any of his predecessors it should be understood that this is a purely arbitrary decision. For the present purposes any one of the names proposed in the list given above would be acceptable (excepting R. Engel's, of course). The character of the rocks under consideration, the poor quality of their exposures, the structural complexity of the Santa Ana Mountains and the stratigraphic facies change evident in this area and probably present in the Santa Ana Mountains form an almost insuperable obstacle to establishing a "type section" in the accepted sense. The formation is almost certainly of very great thickness, and is certainly extremely monotonous in lithology. Larsen's type section should be considered "typical" only in that it does in fact typify the bulk of the formation, in a general sense. In all probability neither the top nor the bottom of the formation as thus defined are exposed in the Santa Ana Mountains.

Rocks of similar lithology have been observed scattered throughout the Peninsular Ranges and have been correlated by Larsen with this formation, on the basis of lithologic similarity. Inspection of some of these localities by the author has confirmed their general similarity to rocks of the Santa Ana Mountain exposures. As mentioned above, rocks of this lithology occur over much of the erosional surface across the

Elsinore trough from the Santa Ana Mountains. Irving (1935) has described a section typical of this formation from southeast of Corona, which he calls the Elsinore series. Dudley, in his description of the Perris block, has described the same assemblage of slates grading into schists and locally, migmatite gneiss. He maps this formation as extending to Double Butte and the hills on the south and west flanks of the Lakeview Mountains, where the present author distinguishes a later series of metasedimentary rocks. Dudley reports the occurrence of metavolcanic rocks and limestone as minor lithologic types in this same series. Both he and Irving note the occurrence of metachert and cherty slates in the section, not associated with volcanic rocks. Larsen has mapped these same rocks in the vicinity of Railroad Canyon (east of Elsinore) as Bedford Canyon formation. Both Dudley and Larsen have noted distinctive coarser-grained schists and gneisses occurring as septa in the igneous complex of this area. These rocks generally have a bulk composition close to that of the typical phyllite-quartzite assemblage of the Bedford Canyon formation.

As can be seen from the above discussion the correlation of various exposures of this formation has been very vague and there is great uncertainty as to the stratigraphic position of individual localities. Lacking well defined markers or fossil bearing horizons the above authors have been forced to consider the Bedford Canyon formation (or Elsinore series, etc.) an undifferentiated lithologically monotonous sequence of unknown but great thickness. Dudley refers to the presence of isoclinal folds east of Elsinore while other authors, including Gray (oral comm.) and Moore (1930) who have studied the formation as exposed west of the Elsinore trough have found no repetition of the section due to folding. Bedding attitudes in the Santa Ana Mountains, although not all in concordance, suggest a general homoclinal north-east dip. If these sedimentary rocks formed a continuously homoclinally dipping series from the Santa Ana Mountains to Winchester in pre-batholithic time there is an immense thickness of this same lithology. Assuming a homoclinal dip of 50°, somewhat less than the average observed along a typical traverse of this section, one arrives at a thickness of 98,000 feet or slightly more than 18 miles. It is likely that this exceptional thickness may be accounted for in part by prying apart of the isolated units of metasedimentary rock by the interposed batholithic masses. In part this apparent thickness may be due to isoclinal folding as yet undeciphered by any students of this area. The writer has found it possible to trace individual marker zones within the Bedford Canyon formation a mile or more along strike, without any reversal in trend, in the Winchester-Hemet area, bearing out the hypothesis of a homoclinal structure. However, it is evident that the overlying French Valley formation has been tightly folded in this same locality and it is not improbable that similar tight folds may lie unrecognized to the west of here, in the Bedford Canyon formation.

F. S. Hudson (1922) has correlated the Julian schist series of San Diego County with the Triassic rocks of the Santa Ana Mountains. It is more likely that the Julian series is correlative with the overlying French Valley formation described below.

4. Lithologic Description

a. General Description

In the Winchester-Hemet area the Bedford Canyon formation is represented by a section of thinly laminated metasedimentary rocks ranging in texture and composition from phyllite to fine-grained quartzite. Near the top of the section coarse-grained, locally conglomeratic quartzites are found. In general the rock types of this section may be described as representing a continuum in lithology between the two end members, phyllite and quartzite, although individual adjacent laminae in the section may differ markedly in their compositional position along this continuum. The diverse laminae may be distinguished by their varying proportions of mica and quartz. Feldspar is relatively rare as a relict clastic mineral. Meta-cherts occur as distinctive units interlaminated with and in part gradational into the phyllite-quartzite continuum.

In this area the formation has not been subdivided although a few quartzite beds have proved useful as markers in mapping.

b. Lowest Rank Occurrences

In areas of low rank metamorphism exposures of this formation are small and deeply weathered. They consist principally of discontinuous "ribs" of quartzite along the crests of rounded hills. Deeply weathered, paper-thin phyllite laminae are exposed between these ribs. In the few places where streams have cut broad rock exposures on the sides of their channels the various rock types show little variability in subsceptibility to erosion. These channel walls are the best exposures for studying the section, aside from road cuts (Plate 2, appendix).

The outcrops are always colored a homogeneous brownish gray to black and from a distance show little sign of lithologic variation. When viewed closely, however, especially in road cuts, the rock exhibits a finely spaced foliation and compositional banding. With the onset of weathering this layering is accentuated due to the preferential weathering of mica-rich layers. Intra-stratal small-scale deformation such as folding, faulting and brecciation is readily identified in this foliated rock. The meta-cherts and quartzites form continuous bands that stand out in low relief and are grooved parallel to bedding where very thin laminae of micaceous rock have weathered out.

The range of mineralogic composition of the phyllite-quartzite sequence can be tabulated as follows:

Per Cent by Volume		
Max	Min	
99	40	Quartz: very fine to medium-grained anhedral equant grains.
50	0	Biotite: z = greenish brown to reddish brown. x = pale yellow to tan. Very fine to medium-grained subhedral flakes containing inclusions of graphite and ilmenite (?).
60	10	Muscovite: very fine to medium-grained subhedral flakes.
5	0	K-Feldspar: untwinned, non-perthitic
4	tr	Graphite
2	tr	Limonite
tr	0	Magnetite: subhedral to euhedral fine grains
tr	0	Tourmaline: a = brownish green; c = pale yellow. Fine euhedral grains in cross-orientation to schistosity.

The ratio of muscovite to biotite is highly variable; there is some suggestion that it is proportional to the graphite content. Plagioclase was not observed in any of the thin sections studied, though very fine grains of untwinned oligoclase may be present in the micaceous quartzites, indistinguishable from quartz.

Nearly pure quartzites with more than 90 per cent quartz form characteristic beds that make up less than 1 per cent of the Bedford Canyon formation. They are often greenish gray on a fresh surface due to finely disseminated epidote. Calcite and sphene are minor constituents. A relict granular texture is visible on weathered surfaces suggesting that they were derived from fine grained sandstones.

In the lowest rank rocks primary sedimentary compositional layering is preserved in detail. The foliation parallels relict bedding in almost every outcrop observed. In the phyllites, mica occurs in very fine flakes or tablets with moderate to excellent planar orientation, enclosing fine blebs of quartz in a poikiloblastic texture. Graphite and limonite are concentrated in planar aggregates of very fine grains (Plate 3a, appendix).

In both the phyllites and the arenaceous rocks quartz and feldspar are in anhedral round grains varying widely in grain size. This size distribution is presumably inherited from the sorting characteristics of the parent sediment. The purer quartzites are more equigranular. However, even in the lowest rank phyllites and quartzites there is considerable recrystallization of the quartz to produce dendritic infiltrating branches emanating from individual grains, and cause consolidation of the very-fine-grained matrix into coarser irregular interstitial grains.

The average grain size of rocks in the sequence phyllite to quartzite increases with increasing proportion of quartz. This is also probably a textural feature of the parent sediments. However throughout the Bedford Canyon formation as here exposed there occur pure, very-fine-grained quartzites containing fine laminae of phyllite. No relict clastic features are visible in hand specimen, and in thin section the quartz grains appear highly irregular in outline and intimately intergrown. These layers are thought to be meta-chert. Rare porphyroblasts of pale yellow-brown garnet (grossularite) and epidote occur in them enclosing fine quartz grains. The meta-chert occurs in layers varying in thickness from one foot down to a fraction of an inch.

Sorting in the parent sediments was apparently best in the shaly members and decreased with increasing grain size. The phyllites at present contain up to 60 per cent very fine-grained quartz in addition to the fine-to medium-grained quartz of clastic origin. It is either intermixed with the mica or in discrete very fine laminae. This quartz is probably in part recrystallized from primary amorphous silica such as that which made up the cherts.

c. Higher Rank Occurrences

The Bedford Canyon formation has been traced southeast along the southern edge of the area into zones of higher regional metamorphic rank where characteristic mineralogic and textural changes are observed. In addition in sections 21 and 22 of T6S, R2W where this formation is in contact with the Domenigoni Valley granodiorite altered porphyroblasts occur as relicts of contact metamorphism.

The former occurrences are principally along the northern edge of the Bachelor Mountain quadrangle in the valley north of Bachelor Mountain. In this area the phyllites are converted to schist and gneiss. The principal megascopic changes are increase in grain size to a medium grained texture and appearance of sub-concordant veinlets of leucocratic rock. As a result of these changes the rock loses its fissility and tends to crop out in rounded bouldery exposures. Towards the eastern limit of exposures the leucocratic material apparently increases in abundance, and takes on the appearance of pegmatitic veins. Much of this material may be derived as a "sweat product" from the metasedimentary rocks themselves but it is noted that this effect increases with proximity to the Woodson Mountain granodiorite and Lakeview Mountain tonalite intrusive masses, suggesting that some of this pegmatitic material may be of exotic derivation.

Under the microscope the textural changes described above can be seen to correlate with the appearance first of andalusite as fine porphyroblasts and subsequently of sillimanite in fibrolite bundles replacing biotite, muscovite and andalusite. These mineralogic changes are discussed in greater detail in the section on metamorphism (Chapter VI).

The contact metamorphic effects, as distinct from the regional metamorphism, are vaguely localized at the contacts of the low-rank schists with the later intrusives. In road cuts near the igneous contact in Sec. 21, R2W, T6S the phyllite is converted to a biotite-muscovite schist that contains planar elliptical aggregates of sericite about 2 mm. by 1 mm. These are apparently pseudomorphous after poikiloblasts of an aluminum silicate, probably andalusite. Less commonly there occur knots of biotite and quartz suggesting pre-existent porphyroblasts of cordierite. These schists are fine-to medium-grained and some show lineations due to crinkling of the foliation laminae and alignment of elongate clumps of biotite and muscovite.

In one outcrop adjacent to Washington avenue in section 22, R2W, T6S the schist has been converted to a migmatite in an area of concordant injection of granodiorite dikes. The migmatite is medium-grained, weakly foliated and shows some ptygmatic folding. The dike material itself, Domenigoni Valley granodiorite, is unfoliated and medium-grained.

5. Primary Lithology

The Bedford Canyon formation was originally deposited as a series of thinly laminated fine-grained clastic and chemical (the latter possibly biogenic) sediments. The sequence of rock types from phyllites to impure quartzites represents a parent sequence of shales, sandy shales and shaly sandstones. Quartz is at present the most abundant mineral in the recrystallized rocks. It represents two primary sources: very fine-grained clastic quartz, still texturally distinguishable; and clay- to silt-sized amorphous silica, probably of biogenic origin, now completely recrystallized. Clay minerals present in the primary shale component of the sediments have been converted to mica. Carbonaceous material of biogenic origin was disseminated through the shales and has been converted to graphite. There is no pyrite present in the phyllites at present, though sulfur may have been expelled during metamorphism. Feldspar was present in subordinate quantities in the sandy shales and shaly sandstones. The present alkali-rich composition of the minor feldspar component of the quartzites is probably representative of its primary composition, at least in areas where the rocks are below the andalusite grade of metamorphism. Above this rank potash feldspar is formed by decomposition of mica. In summary the original sediments can be inferred to have been very fine-grained, poorly sorted, consisting of very fine- to medium-grained clastic quartz and minor feldspar in a matrix of clay, limonite, silica and carbonaceous debris.

The meta-chert layers and the cherty component of the phyllites may possibly be derived from stratiform accumulations of biogenic skeletal amorphous silica such as diatom frustules, sponge spicules, etc. There is no evidence of volcanic deposits associated with these layers. No other possible chemical sources are suggested by the associated sediments. The general sedimentary assemblage suggests a marine origin. Calc-silicate minerals (grossularite, epidote, etc.) associated with the thicker meta-chert units may represent biogenic calcite (foraminiferal tests, for example) precipitated simultaneously with the silica.

Primary sedimentary structures have been well preserved in the lower rank metamorphic rocks of this formation. Primary bedding lamination as fine as 0.2 mm. thickness per bed can be distinguished, frequently marked by extreme variations in the ratio of clay to quartz (silica) between individual laminae. The sediments were poorly sorted in many horizons. Graded bedding is not observed. Very little definite evidence of current action is present except for truncation of shale laminae by coarser wacke at shallow angles of contact. Such features are localized to a few inches along any given outcrop and channelling by this mechanism is less than 3 centimeters deep. No evidence of penecontemporaneous deformation of the sediments was observed. Small-scale deformation during tilting of the section has been assumed to account for the crumpling and shearing within individual strata.

However there occur within this formation, near its upper contact, lenses of conglomeratic arkose containing chips and finer flakes of shale as the only coarse clasts. These lenses have been observed to occur in the screen of metasedimentary rock near the southwest corner of the Winchester quadrangle. They are interlayered with fine-grained pure quartzite. As this type of sedimentary structure occurs more extensively in the overlying French Valley formation, its significance will be discussed in the description of that formation.

6. Thickness

The base of the Bedford Canyon formation is not exposed in the Winchester-Hemet area and indeed is not defined in its type locality. An aggregate thickness of about 3500 feet is exposed in the limited area of outcrop in the southwest corner of the Winchester quadrangle and the northern edge of the Bachelor Mountain quadrangle, assuming no structural duplication, but the section has been observed to extend conformably to the south and west beyond the limits of the present study, dipping always to the northeast.

7. Relation to Overlying and Underlying Beds

No direct evidence as to the orientation of "tops" of beds is observed in outcrops of the Bedford Canyon formation in this area. However the formation is observed to underly in conformable fashion the

French Valley formation for which such evidence is available. On this basis it is possible to deduce that the French Valley formation is younger.

Since the base of the formation is nowhere exposed no direct statement can be made about its basal contact. Conglomerate units within the formation contain only clasts of shale or wacke such as might be derived by intraformational erosion and redeposition. They give no information as to the nature of the bedrock exposed in the vicinity during deposition. Outside of the Winchester area there is some evidence on this matter. Irving (1935) reports clasts of "granite" to occur in conglomerates of this formation near Corona. Rankin (1928) has observed pebbles of vein quartz, "diorite" and "porphyritic rocks" (possibly intraformationally derived volcanic rocks) in exposures in the Santa Ana Mountains. The predominant coarse clasts in these conglomerates are of chert, quartzite and fine-grained volcanic rocks. A more thorough study of the provenance of the rock fragments in these units would be helpful in determining the nature of the basement underlying the Bedford Canyon formation, and the nature of its depositional environment.

As defined in this report the top of the Bedford Canyon formation is conformably overlain by the French Valley formation everywhere that it is exposed, with possible local angular unconformity.

8. Sedimentary Facies Change

As exposed in this area and between here and the type locality the Bedford Canyon formation is lithologically monotonous on a large scale but highly variegated in detail. Individual laminar units of quartzite or wacke may extend along strike for 50-200 feet and rarely pure quartzites can be traced along strike for distances up to a mile. Thus in one sense facies change within the formation is extreme while in a broader view the formation is of uniform widespread lithology.

As noted earlier there are appreciable differences between the Bedford Canyon as described at its type locality by Larsen and as exposed in the Winchester-Hemet area. In effect these differences represent facies changes over long distances, and it is presumed, though not well established, that the lithology of this formation varies by imperceptible degrees from the one locality to the other. Intermediate exposures such as those at Railroad Canyon near Elsinore should exhibit lithologic characteristics intermediate between the extremes of the Santa Ana Mountains and the Winchester-Hemet area. Whether this is so remains to be tested by future students of Peninsular Range geology.

Detailed study of rocks of the type locality and vicinity reveals interesting contrasts in lithology with rocks of the Winchester-Hemet area. The author has briefly studied the Bedford Canyon formation as exposed along the crest of the Santa Ana Mountains and near Mesa de Burro southeast of the Elsinore Mountains. Also, through the courtesy of Cliffton Gray of the California Division of Mines and Charles Woodward of UCLA he has been permitted to study thin sections of some of these rocks. Additional descriptions of sedimentary types encountered in the central part of the Santa Ana block are contained in the works of B. N. Moore and W. D. Rankin. Comparisons of the Santa Ana-type sediments, as they will be called, with the Winchester type are complicated by the higher degree of metamorphism in the Winchester-Hemet area. The Santa Ana sediments are only mildly recrystallized. Interstratified basalts contain relict pyroxene only partly replaced by chlorite. There are local aureoles around intrusives in which higher rank hornfelses are encountered, but as a whole the mineralogy of the rocks indicates a rank of lower greenschist facies (Turner 1948, pp. 93-98). The inferred primary sedimentary types of the two areas are tabulated for comparison below:

	Santa Ana Mountains	Winchester-Hemet area
1. Arenites		
Mineralogy:	quartz ferromagnesian minerals (partly converted to chlorite) chlorite sericite plagioclase limonite rock fragments (aphanitic volcanics, chert, etc.)	quartz (sericite) clay chlorite (now converted to biotite) graphite (carbonaceous matter) traces of alkali feldspar
Texture	very coarse- to very fine-grained, poorly sorted angular grains of sand some polymict conglomerate	medium- to very fine-grained, poor to moderately well sorted subangular (?) grains of sand rare intraformational conglomerate
Structures	massive to thinly laminated (20 mm.) graded bedding and other current structures	very thinly laminated (0.2-5.0 mm.) seldom massive no graded bedding
2. Cherts	impure, shale stringers interlaminated	impure, shale stringers interlaminated
3. Limestone	thin lenticular laminae	absent
4. Volcanics	interbedded with clastics, but less than 20 per cent of section basalt and andesite flows, andesitic tuffs, (?)	absent
5. Pelites	thinly laminated shale quartz, clay, limonite, graphite, chlorite	same, but very thinly laminated

Contrasting these two sections as a whole it can be seen that the Santa Ana section is more typically eugeosynclinal (as described, for example by Kay (1951, p. 86), or Krumbein and Sloss (1953, p. 367)): greywackes and dark shales predominate with lesser amounts of limestone, chert and volcanics. The Winchester type is a transitional type closely related to the latter, i.e. miogeosynclinal. As will be seen below, the overlying French Valley formation, while containing elements of the Winchester type of sedimentary assemblage, represents a still more characteristically miogeosynclinal or unstable shelf type of assemblage (Krumbein and Sloss, p. 362).

These are of course very broad generalizations based on relatively little published data and supplemented by a minimal amount of field investigation by the author. The problem merits further study along both stratigraphic and structural lines.

C. THE FRENCH VALLEY FORMATION

1. Introduction

The Bedford Canyon formation as exposed in the Winchester-Hemet area is conformably overlain by a series of about 20,000 feet of quartzites, schist, amphibolite and chert, here designated as the French Valley formation.

Due to the marked increase in metamorphism observed in rocks progressively eastward through the area it has not been possible to carry individual units within this formation through the more complex structure of the eastern part of the area with any degree of certainty. Therefore, although the stratigraphic description to follow, based on low-rank outcrops, is fairly detailed and unequivocal, the map units within the formation have been highly generalized in some places. This generalization permits a more realistic portrayal of the reliability of mapping in areas where similar rock types cannot be definitely correlated across faults or under covered or brushed-over areas.

The formation is subdivided into three members, A, B and C, of generally uniform internal lithology. They will be discussed separately except in respect to the problems of age and correlation.

2. Name

The French Valley formation is named for its exposures and type section located along the east edge of French Valley (a name no longer in use by natives of the area, though represented on the most recent maps). The rocks of this formation were first studied by D. M. Fraser (1931) who examined exposures in the Hemet quadrangle in the vicinity of Mica Butte. He called the rocks "undifferentiated metamorphics". W. J. Miller, studying the whole series as exposed along Searls Ridge named it the "Diamond Valley series". Larsen (1948, p. 16) refers to the same rocks including the type sections of the present French Valley formation as undifferentiated "Paleozoic schist and quartzite". It will be seen below that these rocks are probably not Paleozoic.

Although Miller has given a name to this formation, he has made no attempt to describe the stratigraphy of these rocks and since all the exposures in Diamond Valley are of higher metamorphic rank and not well suited to specific stratigraphic description it has been thought proper to abandon his name and choose a new name from a geographic locality where a columnar section could be worked out. At the same time it should be understood that the French Valley formation as here defined is wholly included within Miller's Diamond Valley series.

3. Stratigraphic Description

a. Member A

i. General description

Member A of the French Valley formation is a sequence of interbedded pure- to arkosic-quartzites and schists with pronounced lateral variation in lithology throughout the section. The member is generally characterized (with respect to its primary sediments) by poor sorting and gradational transitions in composition and texture both along and across the inferred time lines.

The base of the formation is defined at the first thick feldspathic quartzite overlying the Bedford Canyon formation phyllites. In general Member A of the French Valley formation is distinguished from the Bedford Canyon formation by the occurrence of abundant thick meta-arenaceous units, commonly of feldspathic composition, interlayered with meta-pelitic rocks (resembling those of the Bedford Canyon formation).

ii. Occurrence and topographic expression

Member A appears as an arcuate strip about one mile wide along the east side of French Valley and extending south into the valley north of Bachelor Mountain, toward the head of Tucalota Creek.

In the area of the type section of this member, sections 14, 15, 22 and 23 of T6S and R2W outcrops of this formation form an ascending series of rounded ridges. These rise westward in a moderately steep bluff to an elevation of 2100 feet. The bluff is underlain by a conglomerate near the top of the member, and is prominently visible from French Valley from where it and its foreland ridges appear dark gray and sparsely vegetated. The break in topography from the low, rounded hills of French Valley to the area of higher rounded ridges roughly marks the boundary between this member and the Bedford Canyon formation.

Within this member there are a few units which form high cliffs. A few lenticular quartzites form low transverse cliffs across the stream-cut ridges. Generally, however, the boundaries between lithologic subunits of this formation (as mapped and discussed below) have very little topographic expression. The uppermost unit in this member, a schist, underlies a gentle linear depression near the top of the mountain block south of Domenigoni Valley, roughly marking the base of Member B.

iii. Descriptive petrology (lowest rank occurrences)

Quartzite

The meta-arenites of this sequence range in composition from pure quartzite to conglomeratic, arkosic quartzite. The pure quartzite is restricted to thin laminae, laterally discontinuous and more abundant near the base of the section. In outcrop these pure quartzites commonly form low ridges and weather to a pale pinkish brown. They are yellow-gray on a fresh surface. These pure quartzites are interlaminated with arkose and schist so that the whole range in composition from pure quartzite to meta-wacke can be found in layers a few inches apart. The pure quartzites contain trace amounts of magnetite and graphite. Map unit fvA3 consists of a pink-weathering massive rock resembling the typical pure quartzite in outcrop but containing an assemblage of diopside, hornblende, oligoclase and sphene in addition to about 70 per cent quartz. It appears greenish gray and flinty on a fresh surface (Plate 4a, appendix).

The purer quartzites are fine-grained and equigranular. The quartz has a granoblastic texture with no evidence of orientation of the C crystallographic axes when observed in thin section. Individual beds are up to 50 feet thick and massive throughout.

As the amount of feldspar in the quartzite increases there is a concurrent decrease in degree of sorting of the relict clastic grains and there appears in increasing amounts a third component, shale fragments, now converted to schist. These arkosic quartzites form mottled gray-weathering outcrops which weather preferentially along the planes of weakness induced by the shale chips. On a fresh surface the rock has a dull grayish cast and a salt-and-pepper appearance (Plate 4c, appendix).

Mica is mostly oriented parallel to relict bedding in these impure quartzites. Most of this mica is enclosed in the shale chips which themselves are oriented in this plane. The orientation of the mica flakes with respect to the shale chips may be in part inherited from a primary orientation of incipient mica flakes in the parent shale layers. This orientation has been subsequently further developed during the metamorphism of the quartzites. The shale chips themselves appear in general to have been deposited lying flat on the exposed depositional surface. Later deformation has streaked out these clasts into undulating micaceous streaks less than 0.1 mm. thick. Mica in the sandy matrix is generally less well oriented.

The typical arkose of this member contains about 15 per cent untwinned potash feldspar, 10 per cent oligoclase, 70 per cent quartz and variable amounts of biotite and muscovite. The latter two minerals occur, together with very fine-grained graphite and quartz in the shale chips. Several per cent of the mica also occurs as discrete crystals in the arkosic matrix. Accessory minerals in the arkose are zircon, magnetite, ilmenite, graphite and sphene. Several percent of granule-sized polycrystalline grains are observed in some specimens, consisting of intergrown quartz and feldspar, or of several quartz units in different optical orientations. These are probably relict rock particles.

The texture and composition of these quartzites are comparable to what are called "lithic wackes" by some authors (Turner, Williams and Gilbert, 1953, p. 301). They have the poorly sorted characteristics of wackes but lack the argillaceous matrix of greywackes or subgreywackes. In this case the rock in question is a metamorphic rock but the low rank exposures display the textural characteristics of the parent sedimentary rock in an excellent state of preservation.

Transitional between the true arkoses with shale chips and the orthoquartzites are appreciable quantities of feldspathic quartzite with up to 15 per cent total feldspar and little or no mica. These rocks are poorly sorted and massive to faintly foliated. Relict coarse feldspar clasts weather out on the outcrop surfaces. Unit fvA2 consists in part of such quartzites, with admixed calcareous material.

The texture of the shale-chip conglomeratic arkoses is a very uniform characteristic of this lithology. The matrix of the rock is very poorly sorted arkose consisting of medium to very coarse angular grains of quartz and feldspar in an interstitial submatrix of very fine-grained quartz. Mica occurs as rudely oriented fine tablets throughout the matrix. Embedded in this sand-sized material are clasts of shale ranging in size from granule to pebble and in shape from spheres to rectangular or irregular polyhedra. Within the same hand-specimen the chips themselves are widely diverse in composition, ranging from light gray quartzose phyllite to black, glistening graphitic slate. There is a continuous transition from this lithology to conglomerate of similar texture and composition.

While, in these lowest rank arkoses, the biotite and muscovite grains are very fine, the majority of the quartz and feldspar grains are medium- to very coarse-grained. They are subangular to subrounded and only slightly recrystallized from primary clastic shapes.

Meta-wackes such as were observed in the Bedford Canyon formation are observed in rare instances in Member A. Mineralogically these rocks closely resemble the conglomeratic meta-arkose; only the texture has changed. A specimen from one of the lowest rank exposures of this lithology at the base of subunit fvA3, contains 40 per cent sericite, 5 per cent biotite, 25 per cent quartz, 20 per cent potash feldspar and 10 per cent oligoclase. The rock consists of angular medium-grained clasts of quartz and feldspar embedded in a very fine-grained matrix of sericite and biotite. Individual angular equant clasts of quartz intergrown with feldspar of both types occur as a minor constituent, apparently fragments of a medium-grained granitic rock. Strata of this lithology occur at the base of fvA3 and 600 feet above the base of fvA6, as well as in minor lenses in the arkose units (Plate 4d, appendix).

Conglomerate

The shale-fragment arkoses can be found to grade locally into conglomerates, that is, rocks in which the pebble-sized clasts make up more than 50 per cent of the rock. Rocks of this texture make up less than 3 per cent of the section of Member A. They form small lenses and stringers in subunits fvA2, fvA6 and fvA7. A very persistent lens of this sort occurs near the top of subunit fvA7 extending about 3000 feet along strike south from Domenigoni Valley. The most abundantly conglomeratic unit is fvA6 which contains lenses and stringers of pebble to boulder conglomerate interlaminated with conglomeratic quartzite and lenses of graphitic schist.

In outcrop the conglomerates weather like the schist which they resemble compositionally, and are very weakly resistant to erosion. Except on steep slopes or where the beds cross ridges, the outcrops are poorly exposed or completely covered by soil and slumped material. The rare outcrops are dark dull gray and the characteristic conglomeratic texture is completely obscured by muddy stains of the decaying schist-clasts which cover the outcrop surface.

The microscopic texture of the clasts does not indicate extensive shearing (displacement of matrix, breaks in pebble outline, etc.) and it appears that the shape of the clasts at present is representative of their primary clastic shape. The flatness, $\frac{A+B}{2C}$, is greater than 10 for most of the clasts. It is observed that as the abundance of pebble-sized clasts decreases in a given stratum the flatness increases. Thus, in the slightly conglomeratic arkoses the typical "pebble" is actually only a two-dimensional streak of shaly material, usually only recognizable where the rock has parted along its relict bedding. In the true metaconglomerates discussed here, a fractured surface perpendicular to bedding exposes roughly elliptical, highly angular pebble- to cobble-sized clasts of schist with flatness ratios estimated as low as 3. Rare stringers contain subrounded to rounded clasts of high spericity with flatness ratios less than 3. A conglomerate lens at the base of subunit fvA6 in section A14(2, 3),[1] for example, contains slabby clasts of phyllite up to one foot across, in addition to well-rounded cobbles of the same material. There is in general a negative correlation between clast size and flatness.

As mentioned above the conglomerates are essentially intraformational. The coarser clasts consist in all cases of schist or phyllite indistinguishable megascopically and microscopically from pelitic rocks occurring as continuous laminae above and below them in the section (Plate 4b, appendix). A thin section of a typical conglomerate has the following estimated composition:

Pebbles:	25% biotite	Matrix:	45% quartz
	30% quartz		3% oligoclase
	40% muscovite		2% K-feldspar
	5% andalusite		25% muscovite
	2% graphite		20% biotite

However as in the case of the conglomeratic arkoses there is considerable variation in the composition of the pebbles, in terms of the ratio of quartz to mica. The pebbles range in color from light gray to black and some of the lighter-colored pebbles show a cherty fracture surface. Again as in the case of the arkoses the sand-sized clasts themselves are

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[1] See Chapter I for key to coordinate system.

embedded in a sub-matrix composed in this case of very fine-grained biotite and muscovite and minor quartz. The sand-sized clasts of quartz and feldspar are coarse and angular. Many have been fractured in place without rehealing.

The schist and phyllite clasts are only weakly fissile. More fissile types such as are encountered as small pebbles in the meta-arkoses are not as a rule found in the true conglomerates. Evidently these have been comminuted by current action into fine mud and enter into the sub-matrix of the conglomerates.

The conglomerates do not in general immediately overlie their source rocks but are rather interbedded with pebbly arkoses. An exception to this rule is the series of conglomerate lenses in the biotite schist at the very top of Member A (subunit fvA7). In this unit it is observed that the conglomerates, often containing less than 20 per cent quartzitic matrix, are made up of locally derived schist (shale) fragments, platy and angular in form. These conglomerates can be traced northwest into conglomeratic arkose. Nowhere in the conglomerates are there found clasts of obviously exotic origin, that is, not identical to some lithologic type occurring elsewhere in the section of Member A and the Bedford Canyon formation.

In places the conglomerate units clearly channel and truncate the underlying arkose and schist. Such a channel occurs at the base of a lens in A14(2, 3). Here the coarser clasts are roughly oriented in an arc parallel to the walls of the channel.

Well defined grading in the size and abundance of schist clasts was observed in only one locality, in subunit fvA3 in A23(1, 4). This unit grades upward from 50 per cent large shale-pebbles into arkose with 2 per cent granule sized chips. Elsewhere in Member A conglomerates are found to grade above and/or below into less pebbly finer-grained arkose.

Schist and Phyllite

In the area of low rank exposure at least 15 per cent of the total thickness of Member A is composed of biotite-muscovite-quartz-graphite schist or phyllite. However rocks in this lithology are generally poorly exposed in this area, forming gentle slopes covered with float of schist flakes and pebbles overlying a characteristic gray powdery soil. Only where interlayed with quartzite do the schist units form continuous outcrops. In outcrop the schist weathers to a dull brownish gray streaked with orange. On fresh surfaces it is a dull gray showing a silvery sheen on parting-surfaces.

The schists and phyllites vary from rocks containing 80 per cent or more mica and graphite to rocks of subgreywacke composition (described above). The principal type accounting for over half the rocks designated as schist contains about 20 per cent biotite, 10 per cent muscovite, 0 to 5 per cent graphite and a remainder of quartz. Feldspar is lacking in the lower rank rocks although with the appearance of andalusite minor untwinned potash-feldspar appears. The accessory minerals are unidentified ores and anhedral tourmaline.

In texture the argillaceous rocks range from phyllite to medium-grained schist. The lowest rank occurrences, also the lowest in the section (fvA1), are identical to phyllites of the Bedford Canyon formation. Orientation of mica in these phyllites is weakly defined parallel to relict bedding. It is more consistent in the biotite flakes than in the muscovite. The rocks are all very fine-grained and equigranular. Spotting is ubiquitous; ellipsoidal patches up to 3 millimeters in maximum dimension occur in which biotite is absent and muscovite is coarse and poikiloblastic. The spots may be surrounded by rims of more biotite-rich material. These spots may represent relict poikiloblasts of andalusite.

Graphite ranges in abundance from trace amounts to 5 volume per cent. It occurs as a fine dust, and locally as streaks and patches up to a millimeter across. A distinctive unit with phyllitic texture occurs about 80 feet above the base of fvA5 interlayered with arkose. In thin section it appears to be a subgreywacke with medium-grained angular clasts of quartz and potash-feldspar in a matrix of very fine-grained quartz and graphite. The foliation of this rock is entirely defined by the streaks of graphitic material. Graphite makes up about 5 per cent of the rock by volume. Such lithology has also been recognized in certain schist laminae in Member C. It may be even more abundant in Member A but in outcrop it is difficult to distinguish from the true argillaceous rocks (mica schists).

The schists occur predominantly as laminae or lenses a few inches to several tens of feet thick. The one prominent exception is unit fvA7, 540 feet thick. This unit contains much meta-conglomerate. The lenses and laminae are often intimately interlayered with stringers of quartzite. These stringers are usually deformed into complex folds not relatable to the regional structure. In many outcrops one may find thin sandy layers that have been squeezed into small flexures or extruded into pockets displacing adjacent schist. Likewise shaly material is found to have been extruded from between quartzitic layers or intruded in dike-like fashion across the layers, often isolating boundin-like angular clasts of quartzite. The schistosity of the shaly material cuts across these intricate deformational features at steep angles to the apparent flow direction of the shale. These features are seemingly pre-metamorphic.

Chert

Though much less abundant than in the Bedford Canyon formation, scattered lenses of meta-chert, recognizable by its flinty fracture-surface and very fine-grain size, occur interbedded with arkose in subunit fvA4. In addition, it is probable that much of the quartz of the phyllitic units is derived from amorphous silica, as suggested by its extremely fine-grain size. The chert units are lenticular bodies of less than 100 feet length along strike and a maximum thickness of 30 feet.

iv. Higher rank occurrences

The andalusite isograd cuts through the French Valley formation near its type locality east of French Valley. Continuing further south the andalusite rank-boundary transgresses the entire member as does the sillimanite boundary, but rocks of appropriate composition to determine the andalusite isograd are rare in the critical areas immediately north of Bachelor Mountain.

Within Member A the first signs of progressive metamorphism are observed in the very poorly sorted argillaceous rocks (meta-sub-greywackes). They can be found to grade along strike into medium-grained schists of identical mineral composition or containing small porphyroblasts of andalusite. The major significant change is in the texture: growth and reorientation of the mica into fine flakes roughly oriented parallel to the relict bedding. The megascopic effect of this recrystallization is the appearance of a foliation in these weakly fissile or massive argillaceous rocks. The rare pure shales of the lower part of Member A show a similar recrystallization, less pronounced since they are moderately schistose in even the lowest rank exposures.

In the arkoses and conglomerates advancing recrystallization is accompanied by increasing deformation of the pebbles and granules of shale (now muscovite-biotite schist). These are progressively stretched out and flattened, imparting a weak schistosity to the more conglomeratic arkoses. Even in the highest rank rocks, however, the diffuse outlines of the original clasts can be recognized. Andalusite does not develop in the shale clasts of the arkoses.

With further increase in metamorphic rank in the pelitic schists the andalusite porphyroblasts continue to grow poikiloblastically including fine rounded quartz grains and graphite flecks. The coarsest andalusite porphyroblasts, 0.2 to 0.5 centimeters across contain euhedral clear central patches, elongate parallel to [001]. Muscovite persists in all the andalusite-bearing schists in moderately well oriented flakes or irregular poikiloblasts. The andalusite porphyroblasts distort the foliation defined by the muscovite and biotite flakes although at the grain boundaries of the andalusite mica flakes abut sharply against the andalusite crystal faces.

Fine anhedral poikiloblasts of cordierite occur in the andalusite schists of fvA7 and fvA8. They are filled with inclusions of quartz and graphite. They make up less than 2 per cent of the rock in which they are enclosed.

The quartzite units in this member undergo little change in mineralogy with increasing rank since their assemblage of minerals is so simple. An exception is subunit fvA2, a limy quartzite, containing at its lowest rank exposures (below the andalusite isograd) an assemblage of hornblende ($z \wedge C = 17^{\circ}$, y = grass green, x = pale yellow, z = bluish green), diopside, oligoclase, sphene, calcite and quartz. This calc-silicate assemblage is, beside the aluminum silicate phase transition, the only critical mineral association by which to determine the rank of metamorphism. The assemblage diopside-hornblende-plagioclase has been observed by Eskola to coexist with biotite-muscovite-plagioclase rocks of the Orijarvi region of Finland, in the cordierite anthophyllite subfacies of the amphibolite facies as defined by Turner (1948, p. 77).

In the southeast corner of section A23, rocks correlative with Member A though differing in detailed stratigraphy can be traced into the sillimanite zone. In thin section the andalusite appears to be replaced by fine sillimanite needles. There is no further increase in grain size corresponding to this increase in metamorphic rank. Continuing along strike in these beds into section 24 the schists develop a marked lineation produced by orientation of mica flakes and crenulation of schistosity. Sillimanite becomes an abundant phase and relict bedding features are obliterated by recrystallization of the schists to medium grain size. Both because of these textural changes and because of presumed sedimentary facies change it is impossible to correlate exactly the stratigraphic subunits into the sillimanite-zone and almandine-zone exposures of the southern corner of the Winchester quadrangle and the northeast corner of the Bachelor Mountain quadrangle.

The meta-arkose and feldspathic quartzites of Member A persist with no change in mineralogy and slight recrystallization as one proceeds into the sillimanite zone. But the calcareous quartzites of fvA2 develop a new phase-assemblage: quartz-andesine-grossularite-sphene-diopside (-clinozoisite). Quartzite with this mineralogy occurs in A25(2, 4). It has a very marked lineation due to: 1) elongation of quartz grains and trains of quartz granules; and 2) lenses and streaks of garnet and intergrown very fine sphene crystals. This rock is interlaminated with biotite-muscovite-sillimanite schist, sharing the sense of lineation as defined by elongation of mica and quartz and crenulation of the mica layers.

The mineralogy described above is stable throughout the amphibolite facies as defined by Turner (1948, p. 76) and is not critical for any specific subfacies in his opinion. It may merely represent a bulk composition with a higher $Ca/(Mg+Fe)$ ratio than the quartzites seen in the andalusite zone. The higher CaO content of the plagioclase (An_{42} as against An_{20}) may represent higher temperatures of metamorphism (see the hypothetical phase diagram of Ramberg, 1952a, p. 51).

The schists of the presumed stratigraphic extension of Member A into the southeast portion of the area become progressively more coarse-grained and eventually grade into gneiss which extends around the nose of the Domenigoni Valley syncline. Andalusite and muscovite disappear; sillimanite occurs in fiber-bundles which attain medium-grain size (0.5-2 mm.) and show characteristic basal fractures across the width of the fiber-bundles. The sillimanite has a pleochroism identical to that of biotite which it appears to replace (z = pale brown, x or y = brown). This may be due either to fine inclusions of biotite or to structural Fe_2O_3 in the sillimanite lattice, replacing Al_2O_3. The mineralogy of the higher rank units is virtually constant with increasing grade beyond the sillimanite isograd: biotite in medium-grained tablets anhedral normal to [001]; quartz both in equant grains enmeshed in mica and as fine veinlets, tourmaline (ϵ = pale yellow; ω = yellowish green), in fine euhedrons; and fine-grained poikiloblastic muscovite, anhedral and unoriented (retrograde?). In addition variable amounts of untwinned potash feldspar with very fine lammelar perthitic inclusions may be present, increasing in abundance in samples taken successively further east. A banded gneiss from B19(4, 1) was found to contain 20 per cent coarse poikiloblastic potash feldspar, 2 per cent plagioclase (An_{45}) and traces of albite-quartz myrmekite. Garnet fails to appear in the schists of Member A until well around the nose of the Domenigoni Valley syncline where correlation is dubious, although coexistent amphibolites (considered to be metamorphosed intrusive sills) contain a pink almanditic garnet in areas west of here.

The highest rank phase assemblage observed in rocks of Member A is found at the southeast corner of the Winchester quadrangle on the north slope of Black Mountain. Here calc-silicate rocks occur which are roughly correlative with the upper part of Member A (though not represented by lower rank equivalents at the type section). They contain a very coarse-grained assemblage of quartz, diopside, grossularite, wollastonite, plagioclase (An_{40}) and calcite. Rocks of comparable composition in the lower rank assemblages were observed to contain no wollastonite but calcite in equilibrium with excess quartz. This is evidence that temperatures of metamorphism here have exceeded the wollastonite reaction point, assuming that pressures were comparable in the two localities. Similar wollastonite-bearing calc-silicate rocks are found scattered through the eastern limb of the Domenigoni Valley syncline and the adjacent tightly folded metasediments, and probably are correlatives of this part of the stratigraphic section.

The farthest eastern exposures of Member A, adjacent to the Lakeview Mountain tonalite are migmatitic biotite-quartz-plagioclase-potash-feldspar-muscovite gneisses with variable amounts of sillimanite (0-5 per cent) and garnet (0-3 per cent).

In section A24 the southeast-trending section of arkose and schist of Member A is abruptly cut off along a northeast-trending line by a homogeneous mass of distinctive breccia-like sillimanite-bearing migmatite. The metasedimentary component of this rock does not appear to have been derived from the adjacent rocks of the French Valley formation, but rather resembles the Bedford Canyon formation. This lithologic type (migmatite breccia) is discussed separately in Chapter IV. The boundary between Member A and this lithologic unit is locally transitional, especially where schist lenses of Member A approach the contact. Sillimanite in very fine unoriented needles occurs in most of the rocks showing this migmatitic texture, indicating an abrupt rise in metamorphic grade, since the adjacent schists of Member A contain at this locality only muscovite (no andalusite).

v. Primary lithology

The metasedimentary rocks of Member A were derived from a series of interbedded shales and sandstones and minor conglomerates. The various subunits of this member represent various mixtures of these rock types. These various lithologies apparently were formed from sediments made up of varying proportions of three principal end members: 1) arkosic sand consisting of 70 per cent quartz, 20 per cent potash feldspar and 10 per cent plagioclase (sodic oligoclase to andesine); 2) carbonaceous mud consisting of an estimated 60 per cent clay minerals, 10 per cent carbonaceous matter and 30 per cent clay- to silt-sized silica and quartz; and 3) granule- to cobble-sized clasts of indurated shale, siltstone and very fine-grained sandy shale to shaly sandstone. The rare pure or slightly feldspathic quartzites may represent recycled material winnowed out from the primary arkosic sand; relict sedimentary textures show them to have been better sorted and finer grained than the arkoses. Calcite and/or dolomite was a component of some of the sandstones. The fact that it was lacking from the shaly facies indicates that it was added in the arenaceous component of the sediments rather than as a foraminiferal component as was suggested for the calcareous cherts of the Bedford Canyon formation. Perhaps it represents sand-sized material eroded from a littoral molluscan fauna.

The shale-clast constituent may be of two possible origins: 1) contemporaneously deposited shales, siltstones, etc., slightly indurated and then reworked by the currents carrying the arenaceous component of the sediments; or 2) fragments of earlier sedimentary formations such as the Bedford Canyon formation, eroded from strand lines not presently exposed in the outcrops of the French Valley formation. Such strand lines might be represented by angular unconformities between the Bedford Canyon formation and the French Valley formation. The abundance of shale clast debris appears to be uniform throughout great thicknesses of arkosic quartzite (as will be seen especially well exemplified in Member B). The shale clasts are uniformly angular and poorly sorted as are the relict clasts of quartz and feldspar. Although direct channeling of shale layers by conglomerate or conglomeratic sandstones is not observed, these latter units are invariably interlaminated with and in some instances even grade laterally into what were shales, siltstones and sandy shales of composition identical to that of the clasts that they contain. None of these facts are conclusive proof that the clasts were not of exotic origin. Indeed, the rocks which formed the shoreland of the basin in which these sediments were deposited were probably in fact of just the lithology represented by these clasts. That is, the Bedford Canyon formation may well have been exposed as a sedimentary rock at the time the French Valley formation was being deposited. However it is evident that it is not necessary to invoke such exotic sources and therefore we shall neglect them in this discussion, invoking Occam's Razor.

If we neglect the possible contribution of exotic clasts of shale to the French Valley sediments, the observed sedimentary types can be explained as mixtures of the two remaining end-members (sand and mud) produced in two ways. Principally, shaly laminae must have been eroded by turbulent water movement, fracturing the layers and liberating clasts of shale to be deposited with the arkosic sand. This process, working on source material of varying competence, produced deposits ranging from slightly pebbly arkose to boulder conglomerate. Individual units such as fvA6 show complete transitions through these lithologies. The second process apparently involved mixing of unconsolidated mud with arkosic sand in varying proportions. This was a more infrequent phenomenon and resulted in the rarer wacke-type shaly sandstones and sandy shales such as those in fvA2 and fvA6 (and smaller unmapped lenses). The difference between these two types of mixed sediments may have been due largely to differences in degree of consolidation of the shaly material before erosion by sand-laden currents.

Current features bearing out this hypothesis of submarine erosion are observed in the relict sedimentary structures of the rocks. Graded beds of conglomeratic quartzite are found showing concentration of shale clasts at the base grading upwards into purer arkose. Cross-bedding and gentle truncation of underlying beds are ubiquitous in the quartzites (Plate 4e, appendix). Graded bedding of the more common variety, showing gradation from shaly sand to shale (schist) is observed in many of the meta-wacke layers. These features also help bear out the proposed age relation between the French Valley formation and the Bedford Canyon formation since they indicate the section to be right-side up. However, one critical type of erosional feature which is rarely observed is channeling of shale (schist) layers by conglomerate lenses. That is, local derivation of the shale clast material is not conclusively demonstrated. At the center of section A23 is one possible such occurrence; at the top of subunit fvA3 a poorly exposed lamina of shale appears to be channeled by a conglomeratic quartzite containing chips of the same lithology.

In the thicker shale units, especially fvA7, very fine-grained moderately well sorted slightly shaly arkose forms innumerable lamina a few millimeters thick on the average intercalated with the shale. This is a lithology which was seen below in the Bedford Canyon formation and occurs again above here in fvB2 and various parts of Member C. It is marked by evidence of penecontemporaneous deformation of the shale-sand interlaminations, as described earlier. Presumably this represents quieter water deposition of successive layers of shale and fine sand such as might have been winnowed out of the purer quartzites. These fine sand layers are consequently much better sorted than coarser sands of equivalent composition elsewhere in the section.

The rare cherts in Member A occur as small lenses interbedded with conglomeratic arkose (fvA5). They may be erosional remnants of more extensive layers. They are presumed to be biogenic in origin for the same reasons as the Bedford Canyon cherts. Biogenic silica may have been a constituent of the shales as well although the present degree of recrystallization does not permit distinguishing this component from clastic silt-sized quartz.

vi. Sedimentary facies change

It can be seen from the geologic map (fig. 1) that within the roughly defined boundaries of subunits of Member A there is great change in lithology along strike. For example, subunit fvA2, a calcareous and feldspathic quartzite where it crosses the type section, has become an arkosic quartzite with up to 10 per cent shale clasts at the south edge of the Winchester quadrangle and persists with this lithology to the edge of the mapped area. Similarly fvA6 grades continuously from slightly pebbly arkose at Domenigoni Valley to a densely conglomeratic arkose

(25-50 per cent shale clasts) with lenses of conglomerate where it is cut off by the migmatite breccia, 2 miles to the southeast. Locally even more striking variation can be inferred from the pinch-out of distinctive marker units of wacke and conglomerate. The most persistent units laterally are, as in the Bedford Canyon formation, the pure to feldspathic quartzites which can be followed for distances of a mile or more along strike. It seems therefore that the stratigraphic description given in the columnar section (fig. 3) must be considered at best a general description of the type of sediments to be found in this Member. It cannot be expected that the subunits used in this area can be correlated to areas of exposure more than about 5 miles away.

b. Member B

i. General description

Member B consists of over 5000 feet of massive poorly-sorted meta-arkose with interlayered schists and amphibolite lenses. As in Member A shale chips are a ubiquitous component of the meta-arkoses, usually less than 10 per cent by volume. Extreme facies change of the whole Member and thinning of the quartzite units is inferred in the eastern part of the area where metamorphism has obliterated much critical stratigraphic evidence.

ii. Occurrence and topographic expression

Outcrops of Member B form a broad central strip extending southeast from Domenigoni Valley. This Member also crops out at two localities on Searls Ridge, at the western tip and south of the magnesite mine. In the eastern part of the area the distribution of this unit is largely inferred from scattered outcrops of meta-arkose.

The quartzites of this area form relatively resistant outcrops and therefore underlie higher elevations than the schists of Member A. Two prominent peaks, one in A14(4, 3) and another in A13(2, 4) are underlain by this formation. Within the Member there is little contrast in lithology and consequently little topographic expression of the folded structure of the bedrock. The quartzites form broad rounded ridges and hills, locally densely brushed-over.

iii. Descriptive petrology: low-rank exposures

Quartzites

The quartzites of Member B are lithologically similar in most respects to those of Member A. They form rounded light gray bouldery outcrops resembling exfoliating granite. This resemblance is more than accidental. Except for the addition of variable amounts of shale clasts, the quartzites do in fact resemble a weakly foliated fine-grained granite in both composition and texture.

The two major quartzite units are called fvBl and fvB4. Subunit fvBl is a biotite-bearing arkose in its lower half grading abruptly (?) upward into a hornblende-bearing arkose. Distinction between these two types cannot always be made in the field since the dark minerals are so fine-grained. An estimated modal analysis of typical specimens of each rock is given below:

	biotite quartzite	hornblende quartzite
quartz	50 vol. per cent	65 vol. per cent
potash-feldspar	25	30
plagioclase	17 (An_{10})	1 (An_{20})
hornblende	0	4
biotite	8	tr
zircon	tr	tr
muscovite	1	0
opaques	tr	tr

The rocks are similar in texture: very poorly sorted coarse- to medium-grained quartz and feldspar in a matrix of fine-grained equigranular quartz, feldspar and mica or hornblende. The coarsest grains are angular equant quartz, weathering out on the surface of outcrops. In the biotite arkoses the mica is concentrated together with much of the opaque material in swirling strips corresponding to the relict clasts of shale, Potash-feldspar occurs as dendritic fingers extending away from a central anhedral equant grain, enclosing quartz and plagioclase in equant fine grains. This tendency for the potash-feldspar to recrystallize and redistribute itself in interstices becomes more apparent in the higher-rank rocks. The potash-feldspar contains fine lammelar perthite blebs.

Among the coarser clasts are distinctive polycrystalline aggregates of quartz, intergrown feldspar and quartz and two-feldspar intergrowths. The quartz grains show smooth, curving internal grain boundaries such as are commonly encountered in high rank granoblastic quartzites. They may represent rock clasts from such a source area. The two-feldspar and quartz-feldspar grains have more sinuous internal grain-boundaries suggesting igneous rock textures (fine-grained granitic rocks). Pebble-sized clasts of such rock types are lacking. All these rock clasts are angular to subangular where least recrystallized.

Muscovite is absent or present only in trace quantities in the quartzites although it is a common mineral (2-10 per cent) in the intercalated schists. It is strange therefore that even the shale chips within the quartzites contain only traces of this mineral.

Interbedded with the meta-arkoses are thin, persistent ridge-forming quartzites. They contain traces of calc-silicate minerals and are generally less feldspathic than the arkoses. One such quartzite near the top of fvB4 (in A2(1, 2)) contains the assemblage epidote-grossularite-scapolite-potash-feldspar-calcite in addition to 70 per cent quartz. In outcrop this rock appears to be gradational into meta-arkose.

Accesory minerals in the typical meta-arkose of Member B are: zircon, subhedral, zoned with colorless unoriented inclusions; magnetite in euhedral fine octahedrons; sphene, euhedral colorless to yellowish brown; and graphite as inclusions in biotite or as anhedral very fine grains clustered around biotite-rich streaks. No aluminum silicate minerals occur in the arkoses.

Relict current structures such as those seen in Member A are found here, too. However bedding features of any type are indiscernible over a great part of the section. In places the only hint of bedding attitudes is given by the attitude of the shale clasts. In many of these cases there is some question as to whether later shearing has re-oriented the clasts. In fact a well defined lineation, related to the direction of major fold axes, is defined in places in the quartzites by elongation of the plates of mica schist or by streaks of mica representing smeared out granule-sized clasts. In one locality (A11(2, 2)) extreme shearing has crushed and streaked out all clasts including the coarse quartz grains, resulting in a flaser-gneiss texture. Because of this confusion between sedimentary and tectonic structures there is a possibility that some folding may have been overlooked within the major quartzite units. The general structural pattern of the quartzite units is however one of broad open folding.

Schists

Lenses of schist occur intercalated with the quartzite. Only the thicker, more persistent such lenses have been mapped but innumerable others down to a few inches in thickness and a few tens of feet of strike-length are scattered through the section. For the most part they are identical to the schist units described above in the discussion of Member A.

The three major schist units are fvB2-3, fvB4s_1 and fvB4s_2. The first appears to thin to the southeast while the other two thicken markedly to the southeast. The lowest-rank exposures of fvB2 and fvB4s_1 are andalusite-muscovite-biotite-quartz schists with up to 15 per cent graphite. Andalusite is in coarse poikilitic euhedral porphyroblasts (Plate 5a, appendix). Muscovite occurs as flattened oval patches suggesting pseudomorphism after an aluminum silicate. Thin very fine-grained well sorted arkoses are interlaminated. Near the top of fvB2 a distinctive lithology of quartzite clasts in a shale matrix occurs, a sort of inverse intraformational conglomerate. This immediately underlies an amphibolite. The clasts are somewhat smeared out, suggesting that they were not completely indurated at the time of deposition.

Where the schist lenses are interlayered with hornblende arkose the schists themselves are amphibolitic. One such schist lens near the top of fvB1 contains about 10 per cent hornblende (x = yellow-green, z = grass green, y = yellow), 60 per cent quartz, 15 per cent graphite and 10 per cent potash-feldspar. This rock has a pronounced schistosity, evidently defined by planar aggretates of graphite. The hornblende is in unoriented fine needles dispersed through the rock. The quartz is very fine-grained and equigranular, suggesting a meta-chert. Subunit fvB3 in A11(3, 1) is a schist containing the assemblage biotite-andalusite-hornblende-muscovite-quartz. Hornblende disappears from this unit as it is traced southeast. Thin amphibolitic quartzites are interlaminated. The hornblende in these schists may be recrystallized from the components of detrital mafic minerals derived from the intercalated flow seen in A13(2, 2).

Micaceous quartzites, higher-rank equivalents of the subgreywacke type observed in Member A occur here transitional in texture and composition between the arkoses and schists. They are distinguished from the schists chiefly by their content of coarse quartz and potash-feldspar, in vaguely recognizable clastic grains. Very fine euhedral garnet (spessartite?) occurs in some of these rocks, visible only in thin section.

Conglomerate

As in Member A, a continuous transition from pure arkose to densely-packed shale-clast conglomerate may be observed in many places throughout the quartzite sequence. The schist sequence fvB2-fvB3 in A11(3, 1) shows an exceptionally well preserved case of local fracturing and redeposition of shaly material as clasts in a sandy matrix, in a lens pinching out to the southeast. The angular platy characteristics of the pebble-sized and coarser shale clasts is faithfully preserved in the lower-rank rocks. In general, conglomerate (with more than 50 per cent pebble-sized clasts) makes up about 2 per cent of Member B by thickness, scattered throughout the section.

Amphibolite

The lowest occurrence of definitely meta-volcanic rock in the French Valley formation section is a body of amphibolite intercalated between fvB2 and fvB3 in A13(1, 2). A mode of this rock is given below: (volume per cent)

plagioclase	47.6	
hornblende	49.8	(yellow-grass green-colorless)
ore minerals	2.3	(probably magnetite)
biotite	0.4	(brown-pale yellow)

The plagioclase averages An_{49} in composition. It is in relict phenocrysts showing strong zoning with Ca increasing outwards (reversed). A fine-grained groundmass feldspar also occurs, of indeterminate composition, though more sodic than the phenocrysts. The hornblende is also in two textural habits: coarse clumps of medium-grained anhedral crystals interlocking without common orientation; and fine disseminated stubby euhedral prisms. These two textural types in both plagioclase and hornblende indicate that this rock was derived from a porphyritic igneous rock, probably a basalt flow (see Plate 15, appendix).

This amphibolite forms a lens of 100 feet maximum thickness and 4000 feet length along strike. Much smaller lenses occur above and below it in fvB2 and fvB3. In addition, very fine-grained resistant amphibolitic quartzites occur in fvB3.

Outcrops of the amphibolite weather black with spots of reddish plagioclase altered to clay. The rock is uniformly fine-grained and massive, showing no relict flow-banding.

Subunit fvB4

The uppermost arkose unit of Member B has been sampled for the purposes of detailed chemical and mineralogical studies of the phenomenon of progressive metamorphism. This unit will therefore be discussed in somewhat greater detail in the second part of this paper.

iv. Higher rank occurrences

The andalusite isograd crosses Member B near its type locality and the northern limit of exposures so that most of the critical assemblages are of andalusite grade or higher at their lowest rank exposures.

The arkose units persist virtually unchanged in mineralogy and only slightly modified texturally into the highest-rank environment. The major effect of metamorphism is the redistribution of biotite from its original well-defined concentration as relict clasts of shale. This process has gone to completion in the highest rank rocks, partly accelerated by plastic flowage of the whole rock (shown by ptygmatic folding of fine quartz veinlets). The sub-matrix of fine sand present in the low-rank arkoses is recrystallized to a uniform medium-grain size in rocks of the almandine zone. The only significant mineralogic change is the appearance of tourmaline, up to 2 per cent by volume plus traces of pink almanditic garnet in the highest rank arkoses.

The schist units undergo the same changes in mineralogy and texture as those described for Member A. Again, near the contact with the migmatite breccia of section A24 there is a continuous gradation from rocks of this type into unbrecciated andalusite schists of subunit $fvB4s_1$ in section A13(4, 1). Since pelitic rocks replace the arenites of Member B as one proceeds southeast along strike the typical rock types of Member B in the sillimanite zone and higher are schists and gneisses.

Along the Rawson fault zone there appears to be a zone parallel to the isograds in which the mineralogy of the schists is reverted from an andalusite-bearing assemblage to a muscovite-bearing one. The muscovite forms coarse poikiloblasts that texturally resemble andalusite and may be replacing that mineral. In these same rocks sillimanite begins to appear replacing biotite in fine- to medium-grained knots of fibrolite. A typical rock of this texture from B19(4, 1) has the following estimated composition: (volume per cent)

biotite	10	
sillimanite	8	
muscovite	2	
quartz	30	
potash-feldspar	20	
plagioclase	30	(An_{10})
accessories: apatite, ilmenite, zircon, tourmaline	tr	

Potash-feldspar first appears as a major phase in the schists of Member B coincidentally with the first appearance of sillimanite.

In the highest rank exposures near the crest of the Domenigoni Valley syncline (section B21) schists of Member B have become biotite-sillimanite gneisses with variable amounts of pink garnet and muscovite. These rocks have a migmatitic texture although their bulk composition relative to the low-rank schists does not indicate addition of granitic material. Rather they appear to have sweated out veinlets of pegmatitic material which grade in texture into the main body of the rock. Potash-feldspar is more abundant than plagioclase in these rocks.

The amphibolites of fvB3 pinch out completely before reaching the sillimanite isograd, and no higher-rank equivalents are seen further south in the section although intrusive amphibolites cut the section in the vicinity of Crown Valley.

c. Member C

i. General description

Of the three members of the French Valley formation distinguished here, Member C most resembles in overall lithology the Bedford Canyon formation. It consists primarily of interlayered biotite schists and thin arkosic to pure quartzites plus minor lenticular amphibolites and associated hornblende schist. It is readily distinguished in the field from the arkoses of Member B, even in the high-rank occurrences. It differs from Member A in the absence of conglomerate.

ii. Occurrence and topographic expression

The type section of Member C is in the core of the Domenigoni Valley syncline. This core is well exposed north and south of Domenigoni Valley. Essentially all of Searls Ridge is underlain by this unit as is the eastern half of Double Butte.

Because this member consists of interlayered less resistant schists and more resistant quartzites, the topography of the area it underlies is in part controlled by structure. South of Domenigoni Valley the structure of the major syncline can be roughly discerned from the form of the arcuate parallel ridges underlain by this Member. On Searls Ridge the effect of stratigraphy on topography is less marked due to the relative homogeneity of the rock types. In general the banded gneisses of subunit fvC7 form the highest ridges athwart the main trend of Searls Ridge and some folded structures in this unit can be seen expressed in the curving form of the down-slope ridges. The eastern eminence of Double Butte consists of interlayered banded gneiss and homogeneous graphitic schist; the relative resistance of the former rock is expressed in the southeast trend of the off-slope ridges.

iii. Descriptive petrology

Quartzite

Except for lenticular thin quartzites interlayered with the schists, quartzite is restricted to two subunits of this Member, fvC3 and fvC5. The arkoses of fvC3 resemble in all respects those of Member B. The quartzite of fvC5 is generally only slightly feldspathic. It contains less than 10 per cent biotite. Locally as for example on the east flank of the Domenigoni Valley syncline a calc-silicate assemblage of diopside-grossularite-plagioclase-quartz is present in this subunit. But this unit is also found to grade through arkose to shale clast conglomerate at a few places, for example in B7(3, 3) very near an outcrop of calc-silicate quartzite. The majority of outcrops of this quartzite are quite pure, contain relict well-sorted clastic quartz and lack the characteristic coarse angular clasts of the arkoses.

Thin layers of pure to feldspathic or calcareous quartzite are found interlayered at various places with the schists of Member C. Estimated modes of two such quartzites are given below: (volume per cent)

WH GQ		
quartz	65	
plagioclase	20	(An_{30})
grossularite	8	
diopside	3	
calcite, sphene zircon, magnetite	tr	

WH 176		
quartz	70	
plagioclase	25	(An_{75})
diopside	7	
calcite	1	
hornblende	tr	(grass-green, pale yellow)

Schist

The lowest rank occurrences of most of the schists of Member C are in the andalusite zone (section A12 and Searls ridge, west end). The least recrystallized rocks are finely banded quartz-biotite-graphite-muscovite schists interlayered with very fine-grained biotite quartzites in layers less than one millimeter thick. Near the base of the section of Member C features may be seen suggestive of penecontemporaneous deformation such as those described in fvA7: fracturing of the sandy layers and injection of shaly material. The thin quartzites are commonly deformed into complex ptygmatic-like folds over distances of a foot or less.

Hornblende schist resembling true amphibolite in its dark, rusty-weathering outcrops forms a 100 foot thick layer in fvCl. It contains in addition to hornblende and plagioclase, variable amounts of biotite and quartz and appears to grade laterally into biotite schist with traces of hornblende.

Within units fvC6 and fvC7h it can be observed, as in the Bedford Canyon formation, that the graphite content varies inversely with the biotite content. The extreme end members of this variation are muscovite-quartz-potash-feldspar schists. With increasing quartz content such schists vary from lustrous gray, well-foliated knotted schists (as for example on the crest of Searls Ridge in D35(4, 2)) to massive, blue-gray unfoliated rocks with pitted weathering surfaces. (For example in D36(2, 2), see Plate 5b, appendix.) Andalusite is present in the higher rank equivalents of these rocks. Also present as porphyroblasts in the more quartzose graphitic rocks is an unidentifiable mineral with the following properties: isotropic, n = 1.455, brownish orange translucent in patches, colorless and clear elsewhere, no cleavage; the crystals show a striation in one direction; they are full of opaque inclusions. X-ray study of specimens of this mineral showed only lines of quartz and potash-feldspar, both recognizable as inclusions in the porphyroblasts. This mineral occurs together with andalusite.

A typical graphite schist has the following modal composition (estimated volume per cent):

	Whβ
muscovite	30
quartz	55
graphite	15
hematite, limonite, tourmaline	tr

Such schists are not readily distinguished from the biotite schists interlayered with them in the field and it is not definitely known what proportion of the total thickness of Member C they constitute. Possibly over half of subunits fvC6 and fvC7 are composed of graphitic schists with little or no biotite.

Conglomerate

This rock type is virtually absent from Member C. Subunit fvC5 contains local lenses of very poorly sorted arkose densely crowded with pebble-sized shale clasts (Plate 5c, appendix). Although the schist units appear to grade locally into arkosic quartzite the gradation appears to represent mixture of unconsolidated mud with sand (wackes) rather than broken up shaly layers disseminated as clasts.

Amphibolite

There is some uncertainty as to how much of the amphibolite seen interlaminated with rocks of Member C represents flows or tuffs contemporaneous with the sediments of the French Valley formation. Without much doubt the amphibolites and amphibolitic quartzites in fvC1 and fvC2 (shown in the columnar section) were laid down with the shales of this Member. Hornblende schists suggesting complete gradation between basaltic tuff and shale occur here too. The amphibolites themselves are fine-grained and equigranular and show no relict porphyritic texture. A mode is given below: (volume per cent)

	WH 55
plagioclase (An_{54})	29.3
hornblende	65.9
ores	0.6
sphene	4.2
	(1607 points)

The amphibolites weather to a rusty brown soil and are not resistant relative to the schists, so that exposures are poor and contacts are based largely on float.

Some of the amphibolite layers can be traced for distances of up to a mile and the general zone of amphibolitic rocks including amphibolitic quartzites and hornblende schists can be recognized all along the base of Member C. The quartzites in question appear to grade from almost pure quartz rocks into typical amphibolites. For example the following two modes were made on rocks obtained from outcrops about 100 feet apart along a continuous zone of variably amphibolitic rock:

	WH 156	WH 157 (estimated)
plagioclase	42.3	10
hornblende	55.7	1
magnetite	0.6	tr
sphene	1.4	tr
calcite	-	7
diopside	-	2
quartz	-	70
clinozoisite	-	3
garnet	-	1

It can be seen that WH 157 resembles more closely a calcareous quartzite than one contaminated with amphibolitic material. However as one traces out this stratum the amount of hornblende can be seen to rise continuously. It is possible that this amphibolite and many of the others in this zone of Member C may represent metamorphosed shaly dolomite with varying amounts of quartz admixed. However one would expect that at least locally the carbonate content of the primary sediments would have been sufficient to form a marble while in fact calcite is rare or absent from the amphibolites themselves and marbles are completely lacking from the stratigraphic section. This association may possibly represent meta-cherts formed by alteration of basaltic tuffs or flows.

Besides these amphibolites of probable French Valley age there are others found interlayered with Member C along Rawson Canyon (section B8) which are thought to be metamorphosed intrusive sills. They resemble the extrusive amphibolites very closely in texture and composition and occur structurally in about the same part of the section. Distinction is based principally on their form and spatial distribution, and lack of associated metasediments of mixed volcanic-clastic origin. It can be seen from the geologic map that these supposed intrusive rocks pinch and swell and terminate abruptly against faults with no apparent continuation on the opposite side. Similar amphibolites with comparable spatial distribution and form are found intruding rocks of all parts of the section in the eastern half of the area. Admittedly the distinction cannot be drawn exactly here, however, due to poor outcrops and lack of definitely correlatable stratigraphic markers. Compositionally the two types of amphibolite are almost indistinguishable (see modal analyses in Chapter III). Other arguments as to their intrusive origin are given in Chapter III.

iv. Higher rank occurrences

Most of the exposures of Member C are of andalusite grade or higher. The andalusite isograd is recognized first by the appearance of relict ellipsoidal patches of muscovite enclosing quartz, weathering out as pock-marks on the schist outcrops. In thin section these patches are found to enclose relict scraps of andalusite. A little farther into the andalusite zone andalusite appears as euhedral skeletal porphyroblasts, commonly with a cross-shaped outline. In outcrop the andalusite stands out as pinkish gray crystals usually rimmed with reddish sericitic alteration, contrasting in color with the dark gray biotite-schists. Swirling streaks of inclusions fill the crystals: fine biotite flakes; very fine round grains of quartz; and fine opaque dust (graphite) sometimes weakly concentrated in a chiastolitic pattern. Commonly parts of a single crystal will be densely crowded with inclusions while irregularly shaped patches are completely clear. The pink color is not visible in thin section.

Muscovite occurs in the andalusite schists in two forms: as coarse- to medium-grained subhedral poikiloblasts oriented parallel to the schistosity; and as fine unoriented flakes rimming the andalusite crystals, presumably as a retrograde alteration feature. Further east toward the sillimanite isograd the zone described in Member B recurs where muscovite appears to replace andalusite pseudomorphously.

Together with the first appearance of andalusite, potash-feldspar appears in the schists as coarse poikiloblastic anhedral grains. However this phase remains a minor constituent until the appearance of sillimanite. Cordierite is present in some of the andalusite schists as coarse, round, highly poikiloblastic networks between quartz and mica grains.

Sillimanite first appears in the schists of Member C replacing biotite in bundles of fibers oriented parallel to the schistosity. These bundles then serve as nuclei for further growth and clots up to 1 centimeter across and 1-2 mm. thick grow about these centers.

Garnet, a pink almanditic variety, is locally present, especially in the banded gneisses where it occurs as fine anhedral grains intergrown with biotite.

Texturally for the most part the increased metamorphic grade is associated with only slight coarsening of the schist, especially noticeable in the more quartzose varieties. However one phenomenon that seems to be related to the increase in metamorphic grade is the appearance of a compositional banding of the schists due to concentration of mica into distinct layers (Plate 5d, appendix). All gradations occur from a homogeneous poorly foliated schist to a gneiss composed of alternating bands of feldspathic quartzite (1-5 inches thick) and slightly quartzose biotite-andalusite (or sillimanite) schist (0.2-2 inches thick).

The contact between the banded and homogeneous types is never sharp. Andalusite, sillimanite and garnet, where they occur are restricted to the micaceous layers (Plate 6, appendix). On the west end of Searls Ridge such a banded gneiss contains within its schist layers about 1 per cent of chlorite, pleochroic in shades of brownish yellow. The bulk composition of the banded gneiss is, as well as can be estimated, equivalent to that of the typical homogeneous biotite schist. However the homogeneous schists interlayered with banded gneiss in some localities, especially Double Butte, contain more graphite and less biotite on the average than the typical banded gneiss.

The spatial distribution of this rock type seems to be somehow structurally controlled. It is at present however not clear where to predict the occurrence of banded gneiss except that it seems to be more often than not associated with local tight drag folding in high-rank rocks. There are certainly no occurrences of this type below the andalusite grade of metamorphism. Layers which show banding in high rank localities can be seen to lose this texture as they are traced into low-rank environments. Similar banding has been observed in schists in subunits of the other two Members of the French Valley formation. The origin of this rock will be discussed in the section on metamorphism.[1]

On Double Butte the trace of the isograds appears to parallel the trend of foliation so that individual units cannot be traced into higher metamorphic zones. Banded gneiss occurs here even in the relatively low rank rocks of the south-central part (D28(3, 4)). Andalusite forms

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[1] Only the occurrences of banded schist in rocks stratigraphically equivalent to Member fvC7 are shown on fig. 1. Other areas are shown on the overlay map, fig. 2.

medium-grained skeletal porphyroblasts within the mica bands and to a lesser extent in the homogeneous schist units. The homogeneous schists (fvC7h) are almost unfoliated. Muscovite is more abundant than biotite in these rocks and graphite is present in amounts up to 15 volume per cent. The more graphitic rocks have a well developed foliation due to orientation of the graphite flakes.

There occurs in outcrop on the very southeastern tip of Double Butte a weakly foliated homogeneous biotite-sillimanite-quartz-potash-feldspar schist containing up to 3 per cent cordierite in 4 mm. rounded porphyroblasts. The cordierite grains contain streams of inclusions indicating rolling of the grains during growth. Sillimanite is included in the cordierite.

The rare calcareous quartzites of Member C do not change appreciably with increasing metamorphism. As can be seen in the estimated modes on page227 the An content of plagioclase in these rocks increases with increasing grade of metamorphism. Wollastonite occurs together with diopside, garnet and quartz in a quartzite enclosed in gneiss on the isolated hill southeast of the Lakeview Mountains (section D14(4, 2)), in what may be a stratigraphic equivalent of Member C.

v. Primary sedimentary lithology

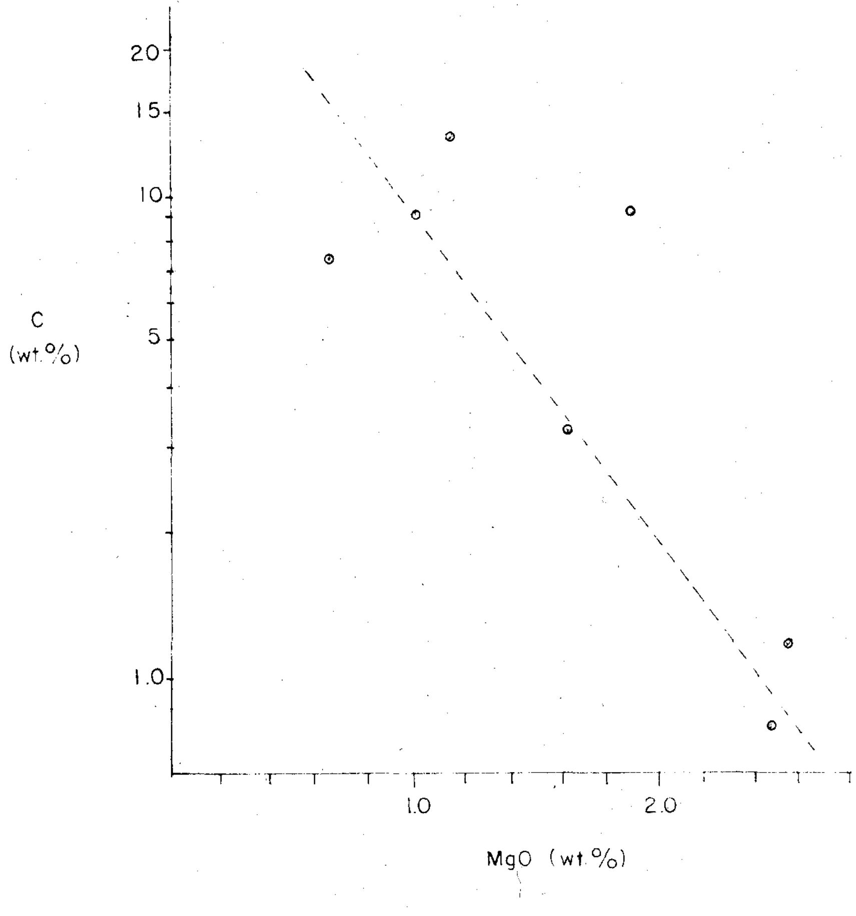
The schists and quartzites of Member C are derived from shales and arkosic to pure sandstones such as were described in reference to Member A. The amphibolites, as discussed above, were probably derived from basaltic flows and tuffs although calcareous shales and calcareous shaly sandstones cannot be ruled out as possible primary rock types. The rare calc-silicate quartzites found interlaminated with the schists of Member C were probably derived from such calcareous shaly sandstones, and their present lithology is in some cases indistinguishable from rocks which have been inferred to be metavolcanic in derivation.

The absence of conglomeratic rocks from this section in contrast with the underlying lithologic units of the French Valley formation suggests a return to quiet water sedimentation like that during Bedford Canyon time. Even the rare conglomerates in fvC5 are usually associated with relatively well sorted sandstones (quartzites) and must represent very local zones of conglomerate deposition.

The graphite schists found high in Member C are derived from carbonaceous muds similar to those making up the graphitic Bedford Canyon metasediments. The inverse correlation of graphite to biotite is more strongly in evidence here than in the Bedford Canyon formation; much of the graphite schist is completely lacking biotite although it contains up to 30 per cent muscovite. Hematite and magnetite are present. The mud from which this rock was derived was apparently very poor in MgO compared to the parent shales of most of the French Valley formation.

Chemical analyses of black shales as compared with typical non-carbonaceous shales are shown graphically in fig. 9. It can be seen that there is a suggestion of an inverse correlation between carbon content (excluding carbonate) and MgO content. In order to explain this trend one must know first of all in what form magnesium is entering the shales. Studies of modern marine sediments (Grim and Johns, 1954) indicate that much of the MgO may be in the form of chlorite, either

FIG.9 Carbon and magnesium content of shales.



data from Pettijohn (1956, p.362, p.344).

authigenic or detrital. An alternate possibility is that the Mg enters the sediments as Mg^{++} ions in replaceable sites in montmorillonites and other clay minerals. The carbonaceous muds which formed these graphitic schists were presumably deposited in an anaerobic environment, with high concentrations of H_2S and NH_3 (Rittenberg et al., 1955) in the water overlying the sediments. It is possible that either: a) extreme concentrations of H_2S, NH_3 or other substances in the oxygen deficient waters overlying the sediments may have inhibited the growth of authigenic chlorite; or b) possibly high concentrations of NH_4^+ may have displaced Mg^{++} from exchangeable sites on the clay minerals of the sediments.

vi. Sedimentary facies change

The base of Member C can be recognized as a break in sedimentation type throughout the Winchester quadrangle. Above this horizon certain markers may be traced for several miles. The amphibolitic zone near the base of the Member is present from the west end of Searls Ridge to section B6(1, 1) on the east flank of the Domenigoni Valley syncline, a distance along strike of about 6 miles. Subunit fvC6 can be recognized as a discrete lithologic unit over the same distance although it varies considerably in thickness. In general sedimentary facies are quite persistent in Member C comparing them to the highly lenticular form of individual rock units in both Members A and B.

In the exposures of Double Butte the map shows there to be a successive thinning of each lithologic unit from southeast to northwest. The units mapped here are mostly banded gneiss and homogeneous graphitic schist. The apparent thinning may be due to pinching of the section during intrusion of the adjacent sub-concordant Domenigoni Valley Granodiorite. There is certainly evidence in French Valley of deformation produced by the force of the intrusion. The thinning may however in fact represent facies pinch-out in the parent sediments. If we completely accept the hypothesis that the banded gneisses are compositionally equivalent to the homogeneous schist, differing only in metamorphism-induced structure, then the possibility of stratigraphic pinch-out seems less likely. In fact the truth probably lies somewhere between these two extremes: the banding of the gneiss probably represents some sort of pre-existent sedimentary banding, absent from the presently homogeneous schists, which has been accentuated by a process of metamorphic differentiation, and the layers of banded gneiss in that case would correspond to sedimentary units (with somewhat gradational boundaries).

4. Age

Direct evidence as to the age of the French Valley formation is lacking in this area. The age of the "Paleozoic schists and quartzites" described by Larsen, essentially identical to the French Valley formation, is given by Larsen to be upper Paleozoic, probably Mississippian. This is based on a fossil locality described by R. W. Webb (1939). A pair of well preserved coral calyxes were discovered, to quote Webb,

> "on a talus pile at the base of a large marble bed in the upper quarry of the Winchester Mine."

The mine referred to is discussed later in this report, where it is referred to as the Hemet Magnesite deposit. The corals were identified by G. H. Girty and G. A. Cooper of the U.S. Geological Survey as being cyathophillid corals of Mississippian age. The actual discovery of the fossil was made by a layman mineralogist, not by Webb.

The validity of these fossils as an indication of the age of the formation is questionable for the following reasons. First of all, the "marble bed" which Webb assumes to be the source of these fossils is in fact a coarsely crystalline calcite marble containing abundant calc-silicate minerals (diopside, forsterite, etc.). It is evidently a recrystallized siliceous dolomite, and it shows what appears to be flow banding parallel to its contacts. This unit is discussed in some detail in Chapter III where it is concluded that it is probably a carbonatite intruded into the associated ultrabasic meta-igneous rocks. Secondly, this is the only occurrence of marble (other than two small inclusions in the San Marcos gabbro) noted in the Winchester-Hemet area and in the French Valley formation as far as it has been studied by this author. For these two reasons it seems questionable that a well preserved coral could have been recovered from calcareous rocks of the Hemet magnesite deposit or that, if it were, it would have any bearing on the age of the enclosing schists and gneisses. Unfortunately the fossils in question have been lost, so that it is not possible to say whether they bear a textural resemblance to the marble of this locality.

The French Valley formation conformably overlies the Bedford Canyon formation in French Valley. The Bedford Canyon formation is thought to be upper Triassic in the Santa Ana Mountains and is probably somewhat younger in this area, as suggested by the general eastward dip of this formation. The French Valley formation is in turn intruded by igneous rocks of Cretaceous age (Larsen et al., 1958). Thus the age of the French Valley formation is between upper Triassic and (middle upper) Cretaceous.

A possible fossil locality has been found by the writer though not thoroughly investigated. It may yield further information on the age of this formation. On a hill south of Florida Avenue in the western part of section D14 there is found a series of biotite migmatite gneisses with rare intercalated calcareous quartzites. The quartzite, composed of diopside, hornblende, garnet and quartz, contains wollastonite masses, apparently pseudomorphously replacing pelecypod or brachiopod shells. The shells are of indeterminable biologic affinity but further investigations may yield pseudomorphic aggregates with forms well enough preserved to be informative. Unfortunately the rock is very hard and unfolifated and consequently fossils would not be readily separable from the matrix.

5. Correlation

Because of the obvious high degree of facies change along strike within the French Valley formation it would be hopeless to attempt to correlate the formation with other exposures except on the basis of general lithologic affinities.

It has been seen that the Bedford Canyon formation can be correlated in isolated exposures along the southeast trend of the Santa Ana Mountains. The author has attempted to trace a parallel belt of sediments of French Valley type with little success. Using the Santa Ana sheet of the geologic map of California (1955) as a guide the writer has visited areas of "pre-Cretaceous metamorphic rock" as far south as Palomar Mountain, and including the Julian schist exposed around the town of Julian in San Diego County.

Only one locality south of the map area yielded exposures resembling typical French Valley lithology. This was an exposure of shale-clast conglomerate in a stream channel on the north slope of Palomar Mountain, 2 miles south of Aguanga. Elsewhere in this vicinity the typical metamorphic rock is a biotite-quartz-feldspar gneiss or schist which could be as well of Bedford Canyon affinity.

Merriam (1946) describes a section of metamorphic rocks in the Ramona quadrangle as being composed of: 1) bands and lenses of quartzite less than 300 feet thick but up to 3000 feet locally, consisting of 80 per cent quartz and a remainder of potash feldspar, andesine and muscovite plus local calcareous facies with calcite, grossularite, diopside and wollastonite; 2) biotite-muscovite-quartz-feldspar schist grading into quartzite; locally sillimanite-garnet schist, sillimanite-cordierite schist, rare lenses of muscovite-sillimanite quartz schist (lacking biotite); and 3) amphibolites, making up less than 10 per cent of the section. These rocks can be correlated to the south with the Julian schist (Donnelly, 1934; Everhart, 1951). This unit, which underlies an area of more than 75 square miles in San Diego county is described by Everhart to consist of: 1) quartz-biotite schist; 2) pure quartzites; and 3) amphibolite. Schist makes up about 90 per cent of the section.

The section as described by Merriam is very similar in lithology to the French Valley formation. The quartzites of this series are described by Merriam as being derived from cross-bedded arkoses. The quartz-mica schists are not distinctive as they could equally well be Bedford Canyon equivalents. The association of calcareous quartzites and quartzose amphibolites is also probably not distinctive since low-rank equivalents of these types are known in the Bedford Canyon formation (tuffaceous and calcareous sandstones).

Northwest of the Winchester-Hemet area the contact between Bedford Canyon and French Valley metasedimentary rocks is obscured by extensive alluvial cover and plutonic igneous intrusive bodies. The nearest possible equivalents of the French Valley metasedimentary rocks are the interbedded schists and quartzites of the Jurupa Mountains. These again contain calcareous, as well as arkosic quartzites. Marble (limestone) occurs in this section distinguishing it from most of the metasedimentary rocks of the Perris block, of which this area is a structural continuation. These rocks may be rather related to the Paleozoic schists of the San Bernardino Mountains (Woodford and Harriss, 1928) and the thick marbles of Crestmore and Colton.

D. DEPOSITIONAL HISTORY OF THE WINCHESTER-HEMET AREA

1. Bedford Canyon time

The Bedford Canyon formation was deposited in a sea of moderate to great depth, probably in a basin enclosed from the open ocean. Positive indication of this is given by the presence of shale to silt sized, clastic sediments, locally rich in carbonaceous matter, such as are attributed to deep quiet-water marine environments (Krumbein and Sloss, 1953, p. 178). The absence of current features or ripple marks and the rarity of coarse clastic sediments are counterindications of near shore deposition. There presumably was a shelf of deposition between the Bedford Canyon basin and shore, trapping coarser terrigenous sediments. Very little volcanic material was added to this basin in this area although to the west (in the Santa Ana Mountains) this same basin of deposition was partly filled with volcanics and volcanic-derived sediments. There is no assurance that this period of volcanism was contemporaneous with the deposition of terrigenous clastics in the Winchester-Hemet area.

No fossil material was found in the Bedford Canyon formation in this area. It is thought that the chert units found in abundance in this formation may represent deposits of the tests of siliceous micro-organisms inhabiting the Bedford Canyon sea. Evidently the fertility of this sea fluctuated with time for the chert tends to occur in fairly well-defined thin layers. Possibly the siliceous material was diluted with clastics at other times, and is therefore unrecognizable.

2. French Valley time

The end of Bedford Canyon time was marked by the advance of shelf conditions referred to above into the Winchester-Hemet area. Whether the shoreline actually crossed into this area at any time is not clear. The coarse poorly sorted arkosic sands, shaly sands and shale-clast conglomerates represent various degrees of mixing of quiet water sedimentation with wave sorting and fluviatile sedimentation. With the term fluviatile must also be understood submarine current transport, marked by graded bedding and cross-bedded sediments extending laterally over great distances. Well defined stream channels are not common. The conglomeratic rocks of this formation apparently represent a recurrent type of sedimentation characteristic of this period. A possible sequence of events forming these units has been outlined above. They presumably represent periodic re-entry of coarse-clastic sedimentation into a basin or lagoon temporarily isolated from sources of coarse clastics and partly filled with silt and mud. The stream velocities of the re-entering clastic-laden currents were sufficient to break up the shaly sediments that had been previously deposited and to strip them down to the original surface of deposition. It is rare that one observes such shaly layers caught in the process of being stripped away, although shale-clast conglomerates are common in much of the section.

These conditions could be found on and near a delta such as that of the Mississippi where local stream deposition would temporarily build up natural levees blocking access of the sand-laden streams to parts of the delta. Under flood conditions these levees would be breached and stripping of the shale could occur.

The small amounts of carbonate material, restricted to arenaceous layers, presumably represent reworked bioclastic debris, stripped by wave action from a nearby shore, and diluted with arkosic sand.

Some volcanic material is found in this section, interlayered with shales marking a period of recurrent quiet-water deposition. The volcanics may have been deposited as crystal tuffs (noting the relict porphyritic texture). Alternately they may have been submarine flows.

The great thickness of the French Valley section indicates that down-buckling of the crust in this area continued through French Valley time. The rate of sinking must have been lower than in Bedford Canyon time, since the area was evidently intermittently built up above the wave base to permit erosion and redeposition of shale units.

3. Source of the feldspathic sediments

The feldspathic sands were derived from a granitic or metamorphic terrane, probably east of this area. Pre-Cambrian igneous and metamorphic rocks are exposed in the San Gabriel Mountains north of here (L. T. Silver, pers. comm., 1959). Feldspathic gneiss of probable pre-Carboniferous age occurs in the San Bernardino Mountains north of the Winchester-Hemet area (Gillou, 1953). Undetermined lateral movement on the great strike-slip faults east of here may have brought other possible sources into juxtaposition in Triassic time. No Triassic marine sediments are known to the east or north of this area for great distances. It is not impossible that land may have extended to the edge of the French Valley basin at this time as the character of the sediments certainly indicates.

The abundance of feldspar in these sediments, although not unequivocal, suggests rapid stripping of material from a granitic or metamorphic highland. However considerable stream transport may have taken place without much altering of the ratio of feldspar to quartz. There are no obvious pre-Triassic sources of such feldspar-rich sands exposed in the immediate vicinity at present. All the nearby granitic rocks have been attributed to a single period of Cretaceous plutonic activity, excepting the pre-Cambrian exposures referred to above.

The high ratio of potash-feldspar to quartz may be wholly or largely the effect of selective weathering of the plagioclase fraction during erosion and transport and may not be representative of the ratio in the parent rock. Similar arkoses in the Tertiary rocks of southern California contain a plagioclase-rich feldspar fraction that reflects the intermediate to basic composition of the igneous rocks from which they were derived. The arkoses of Member B apparently have a mineralogical composition very close to that of a adamellite although no rocks of this composition are known in the pre-Triassic crystalline rocks of southern California.

A possible source of the feldspathic material in these sediments may have been rhyolitic or dacitic tuff. Against this possibility two facts may be marshalled. First it is noted that the feldspar of even the lowest rank meta-arkoses appears to be a very pure potassium feldspar, as inferred from the paucity of perthitic exsolution material present in it. This is in contrast with the typical potash-feldspar of extrusive acidic rocks, a sanidine of very mixed composition. It is conceivable, but unlikely, that all or most of the albite molecule once dissolved in this feldspar may have exsolved as a discrete phase, now observed as anhedral oligoclase grains in the rock. The second fact casting doubt on the volcanic origin of this feldspar is the lack of an argillaceous matrix (which would be represented in the present metamorphosed sedimentary rocks as a micaceous matrix). That such a matrix should be present were the rocks derived from tuffs is strongly suggested by the common development of bentonitic clays in "fossil" ash beds. It is almost invariably true that acidic ash falls are principally composed of glass shards with a smaller amount of crystalline material admixed. This glassy material rapidly alters to a mixture of clay minerals, with concurrent leaching of alkalies and lime. It is possible however that sorting of the altered bentonitic material by stream action (such as is attested to by cross- and graded-bedding in these metasedimentary rocks) might have "reassembled" the altered tuffaceous layers respectively into sandy and argillaceous layers.

Zircon concentrates from a series of samples of arkose from this area (Part II of this study) have been studied. They are very inhomogeneous with respect to shape, color, type of inclusions, etc. This is a suggestion that the sediments were scavenged by stream action over a broad area rather than derived from a single igneous or metamorphic source. Evidence in Part II of this study shows that the ratio Pb/K_2O is quite constant in the meta-arkoses. This is attributed to an initially homogeneous potash-feldspar component in the parent sediment, suggesting derivation from a single source. Both these characteristics may however have been modified by later metamorphism.

Finally, it is noted that coarse angular clasts of feldspar and quartz, often containing grains in several orientations and both quartz and feldspar in the same grain, are found in these sediments as a minor constituent. This feature suggests a rather close source, since extended stream transport would probably have resulted in rounding of these grains. They are not spurious locally derived material; the average composition of these coarser grains is about typical of the arkoses as a whole.

In conclusion it appears that the sand-sized fraction of the French Valley formation was derived from a nearby granitic (or gneissic?) highland, possibly of great relief. The concentration of potash-feldspar relative to plagioclase, atypical for any rocks of this part of the crust, may be either representative of an earlier period of granitic plutonic activity in this area or may be due to climatic conditions which permitted very rapid selective weathering of plagioclase. The latter seems unlikely. Improvement of our knowledge of the possible pre-tertiary displacements on the San Jacinto, Elsinore and San Andreas fault zones may provide us with a clue as to possible bed-rock sources of these sediments, by localizing the portion of the crust that may have been adjacent to this area in Triassic times. At present this question is certainly unanswered.

III. INTRUSIVE METAIGNEOUS AND IGNEOUS ROCKS

A. INTRODUCTION

Three groups of intrusive rocks of igneous origin are found in this area. They are, in sequence of decreasing age: a) a series of basic sills which have been metamorphosed to amphibolites; b) small intrusive masses of ultrabasic rock (and minor associated carbonatite) largely metamorphosed to hydrous magnesian silicates; and c) large plutonic masses of igneous rock ranging in composition from basic to granitic which are locally metamorphosed. Of these three types the second two have been previously studied by workers in this area. The ultrabasic rocks of the Hemet magnesite deposit (also known as the Winchester magnesite mine) have been discussed, principally in relation to the associated magnesite deposits, by F. L. Hess (1908, pp. 38-39) and H. S. Gale (1912, pp. 483-520). Pegmatites associated with these rocks were described by Murdoch (1933). The batholithic igneous rocks have been studied extensively by Larsen (1948).

The first group, amphibolite sills, are dispersed through a large part of the area and rocks of similar composition and texture have been observed in exposures of pre-batholithic metamorphic rock south and east of this area. Ultrabasic rocks are relatively rare in southern California, as far as is known, and, except for a small serpentine body near Santiago Peak in the Santa Ana Mountains, the ultrabasic intrusives described in this area are the only ones known in this part of the Peninsular Ranges. The large plutonic igneous bodies found here are a small part of the batholith of southern California which extends in its entirety from the latitude of Riverside to the southern tip of Baja California.

B. AMPHIBOLITE

1. Occurrence and Topographic Expression

Intrusive bodies of amphibolite are found intruded into the French Valley formation in scattered localities east of a line trending north-south through the Hemet magnesite mine (C31(2, 3)). The one exception is a small area of poorly exposed outcrops in section A26(4, 2). There does not seem to be any tendency toward concentration of these intrusives in any area within this general region, though they may increase in abundance toward the east. The individual bodies range in size from a plump lens 1300 feet long and 600 feet wide in Diamond Valley down to sheets too small to be mapped, less than ten feet wide and a few tens of feet in length. The large lens mentioned above underlies a prominent hill 100 feet high surrounded by alluvium. Other smaller intrusives within the metasedimentary section commonly underlie the crests of low ridges standing above the surrounding deeply weathered gneisses. However even in such exposures the actual outcrop area is small. The soil on the crests of these ridges is densely crowded with float pebbles and cobbles of amphibolite. The width of such amphibolite bodies as mapped may be somewhat exaggerated.

2. Descriptive Petrology

The intrusive amphibolites are found to intrude rocks of sillimanite grade but no lower. The lowest rank exposures of amphibolite are homogeneous to banded, fine-grained, massive and weakly jointed. They are covered by a thin reddish soil and the outcrops are usually stained dark reddish-brown. In thin section the amphibolite is seen to be equigranular and lacking any preferred mineral-orientation. A modal analysis of a typical homogeneous amphibolite from B8(2, 2) is given below: (volume per cent)

plagioclase	36.4	
hornblende	47.1	
clinopyroxene	15.2	
sphene	1.39	
ores	0.1	1303 points

The hornblende is in stubby euhedral prisms. In the lower rank (sillimanite grade) amphibolites it is pleochroic with X = pale brownish yellow, Y = olive green to greenish brown and Z = greenish brown. In a specimen from a high-rank locality in section C28 the pleochroism was X = pale yellow, Y = dark olive-green, and Z = bluish green.

The pyroxene in the lowest rank rocks is disseminated in fine anhedral to subhedral grains. It is an augitic diopside with $2V = 60^\circ$ and $Z \wedge C = 40^\circ$. It has a faint greenish pleochroism.

Plagioclase in the amphibolites is only rarely twinned (making accurate determination difficult in thin section). Even where no relict igneous texture is apparent the grains are zoned, especially the coarser ones that occur in thin veinlets cutting the rock. The composition varies from An_{43} (in the lowest rank specimens) to An_{77} in the highest rank ("Pyroxene-granulite") occurrences in C33.

Sphene is present in most of the specimens looked at, disseminated in fine subhedral grains. Magnetite is a rare constituent in most of the amphibolites as is apatite.

The amphibolites are commonly banded to some degree, even in the lowest rank exposures, and as one progresses further east the degree of segregation of the minerals increases. Amphibolite layers intruded into migmatite gneisses of Diamond Valley (section C33(2, 2)) have the appearance of a finely laminated sediment with alternating layers possessing varying ratios of hornblende to plagioclase.

A modal analysis of such a banded amphibolite is given below: (volume per cent)

	dark band	light band
plagioclase	20.6	40.7
hornblende	66.5	2.0
pyroxene	9.4	51.7
magnetite	0.2	0.2
sphene	3.2	5.4
calcite	-	0.1

The major difference between the bands can be seen to be in the relative amounts of pyroxene and hornblende. This is a common characteristic of all the banded amphibolites. There is little tendency for preferred mineral orientation parallel to the banding, except for local elongation of elliptical patches of pyroxene. Within the lighter bands the pyroxene grain size increases toward the center of the bands. Whether this banding is due to some original compositional layering or to metamorphic differentiation is not clear. The individual layers persist for the entire length of an outcrop in general, from 10 to 50 feet strike length, except in the more finely banded rock where they extend for only a few feet. Veinlets of pure plagioclase transect the rock (Plate 7, appendix).

In the higher rank exposures (garnet zone) a pale pink garnet occurs in the amphibolite, preferentially segregated into the lighter bands. The amphibolites of section C33 in Diamond Valley north of the Searls Ranch, are partly enclosed in tonalite. They are very thickly banded and coarse-grained and individual bands are essentially garnet-pyroxene granulites. This rock-type has a granoblastic texture and is generally medium- to coarse-grained. Individual bands are highly contorted and compressed. Garnet of a dark reddish hue is concentrated in irregular pods rimmed with layers of epidote-plagioclase granulite. A mode of a typical medium-grained granulite from one of the lighter bands is given below: (volume per cent)

plagioclase	35.7
garnet	28.9
pyroxene	31.8
sphene	2.1
epidote	0.8
calcite	0.2
apatite	0.3

The pyroxene in this rock is pale grass-green in thin section, but not pleochroic. The extinction angle is $37^{\circ}(+)$; $2V(+) = 70^{\circ}$; $n_z = 1.728$, $n_y = 1.715$, $n_x = 1.703$ (calc); $n_z - n_x = 0.025$ estimated in thin section. From Winchell's four-component graph (Winchell and Winchell, 1957, p. 415) the composition is estimated roughly as $Di_{45}Hd_{35}Ja_{20}$. The garnet has a refractive index of 1.80 and is presumably in the pyralspite series (see Chapter VI).

Rock of this composition makes up about half of the volume of the metamorphosed basic igneous rock outcrops in the highest-rank exposures. The remainder is an amphibolite containing 60 per cent hornblende (Z = dark green, Y = light olive-green, X = yellow) and 40 per cent plagioclase (An_{70}) plus traces of pyroxene, sphene and rutile. Evidently the bands of pyroxene plagioclase rock which segregated (?) under lower-rank conditions were the locus of the growth of garnet at the highest grade of metamorphism. Epidote, which is locally abundant in the granulite bands (one thin section contained about 15 per cent) appears to be an alteration product of both the plagioclase and the garnet. The country rock of these amphibolite-granulite sills is biotite-sillimanite-oligoclase-(almandine) migmatite gneiss.

3. Structural Relations of the Amphibolites

For the most part the amphibolites of the east half of the Winchester-Hemet area are concordant bodies whose length along strike is many times their maximum width. The exceptions to this are the problematical "fault-slices" just west of Rawson grade, discussed in the stratigraphic description of Member C of the French Valley formation. Intrusive contacts between amphibolite and gneiss are always buried by soil and float. In cases such as these "fault slices" there is a concommitant uncertainty about the true nature of the contacts, especially where the sill appears to end abruptly rather than pinching out along strike. The two adjacent bodies in B8(3, 1) for example have been traced around their northern terminations by following float pebbles. There is no evidence that they are faulted off since stratigraphic units can be traced past these terminations in the enclosing gneiss. However the sills end abruptly along a common line trending N75E. The thicker of the two sills is over 400 feet thick where it terminates. A much thinner sill appears to continue on to the north as an extension of the eastern part of this sill. There is no apparent displacement of the gneiss around this large intrusive body and intrusion of the primary basaltic magma must have been accomplished by stoping of the gneiss (or its parent rock) rather than prying apart along the schistosity.

In every place where the enclosing metasediments have been folded the amphibolites are also folded. Their weak foliation and banding is everywhere parallel to that of the country-rock. From this fact and their mineral assemblages it can be inferred that they were intruded before metamorphism and before most of the folding. However, in outcrops exposed in and south of Crown Valley (B17(4, 2) and B16(1, 1)) the amphibolites exhibit some discordance to relict bedding in the enclosing gneisses and at one point one appears to bifurcate. This may be an illusion produced by local, poorly exposed, complex folding.

4. Origin of the Amphibolites

It has already been shown (Chapter II) that there do exist in the French Valley formation stratigraphically well-defined volcanic units of unquestionable extrusive origin. That there also exist exposed in this area metamorphosed basic igneous rocks of intrusive origin requires further proof and it is unfortunately true that for the most part critical evidence is lacking. To an undeterminable extent there may be some confusion in identifying such units in the field, especially since they are restricted to parts of the area characterized by poor exposures and poor stratigraphic control.

The question of whether these bodies are actually metamorphosed intrusive sills can be best settled by considering the evidence stated above plus one other fact, namely their areal distribution. As was first pointed out, the amphibolites are not restricted to one part of the columnar section. Rather they are distributed through a broad belt of tightly folded metasedimentary rocks representing most of the French Valley section. They may increase somewhat in abundance to the east. Such amphibolites or lower rank equivalents are unknown anywhere in the low-rank exposures of the Bedford Canyon formation of this area and it can be shown that similar amphibolites in the lower-rank French Valley formation are probably interlayered flows or tuffs (or meta-dolomitic shales) rather than intrusives. Thus it can be seen that their present distribution cross-cuts earlier sedimentation patterns and is more strongly related to later superimposed patterns of folding.

The facts above have led the writer to conclude that the amphibolites were intruded before metamorphism and that they made room at least in part by stoping though perhaps controlled by local stratigraphic zones of weakness. Their concordance argues perhaps more effectively for forcible intrusion by prying apart of adjacent strata, but the abruptness of the terminations of several especially thick sills would then require separate explanation.

Admittedly, however, the same pattern of areal distribution could be accounted for by a local site of extrusive igneous activity active throughout a long period of French Valley sedimentation. However, one would expect, in this case, to find some examples of admixed volcanic-clastic sedimentary material represented in the adjacent metamorphic rocks, as is observed associated with the clearly extrusive units of the French Valley formation. Such is not the case; the enclosing schists and gneisses are typical of the pelitic-arenaceous assemblage noted elsewhere in the area, with no exceptional enrichment in MgO or CaO, as deduced from their mineral assemblages.

The overall picture that appears from these observations is as follows: a) after the French Valley and Bedford Canyon sections had been folded to roughly their present configuration, but previous to intrusion of the plutonic igneous rocks, basalt sills were intruded conformably into the sediments from a magmatic source underlying the eastern part of the area; b) the basalt and enclosing sediments were further folded and finally regionally metamorphosed to amphibolite and gneiss respectively.

C. META-ULTRABASIC AND ASSOCIATED ROCKS

In four scattered localities through the Winchester-Hemet area are found meta-igneous rocks derived from a suite of ultrabasic rocks. Two of these are small bodies of tremolite-anthophyllite schist, one in section B20(4, 2), and the other in B16(1, 4). The other occurrences are of metaperidotite showing relatively well-preserved igneous textures, one on the east edge of Double Butte (section D22(1, 3)) and the other, a large complex sill, in section C31 on the crest of Searls Ridge, at the site of the Hemet magnesite mine.

1. Tremolite-anthophyllite Schist

This is a lime-green medium-grained rock with a pronounced schistosity and lineation due to orientation of the prisms of amphibole. Outcrops of this rock are fresh and bare and stand out above the surrounding gneisses. In thin section the schist is seen to be composed of about 25 per cent glassy, green tremolite ($Z \wedge C = 14^{\circ}$, $n_z = 1.630$) and 65 per cent pale brownish anthophyllite ($n_z = 1.638$) intergrown in a network of well-oriented subhedral prisms. The tremolite is in general somewhat coarser-grained and can be distinguished by this characteristic in thin section. The relative amounts of the two amphiboles are however very uncertain. The rock also contains small amounts of talc, brownish chlorite and chromite (?). It is probably derived from a two-pyroxene pyroxenite.

The schistosity and lineation of this rock conforms to that of the surrounding gneiss and the rank of metamorphism exhibited by the mineral assemblage tremolite-anthophyllite is probably in equilibrium with the biotite-sillimanite-almandine assemblage of the gneisses (Fyfe, Turner and Verhoogen, 1958, p. 230). Therefore it appears that this rock was metamorphosed concurrently with the gneisses. It is not known when the parent rock was intruded but it may be significant that one of these bodies (that in Crown Valley) lies along strike about one hundred feet north of an intrusive amphibolite. The parent ultrabasic rock in this case may have been a crystal accumulation from the basaltic magma ancestral to the amphibolite.

2. Ultrabasic and Associated Rocks of the Hemet Magnesite Mine

a. Distribution and Occurrence

The Hemet magnesite mine is a quarry cut into a highly altered ultrabasic intrusive body on the crest of Searls Ridge. The ore-body is one of two metaperidotite intrusives intruded into the axial region of a syncline in the French Valley schists. The other body, much less altered and lacking magnesite, forms a bulbous mass near the axis of the syncline with a tail-like sill extending about one-half mile northwest along the west flank of the fold.

In addition to the metaperidotites there are associated intrusives of pyroxenite, amphibolite (meta-gabbro) and carbonatite and minor dikes of tremolite-rock, pegmatite and vein quartz. A geologic map of the mine area is shown in fig. 6 (in pocket).

b. Descriptive Petrology

i. Peridotite and dunite

The major intrusive body in this group is a complex of peridotite and dunite, both partly altered to hydrous and carbonate minerals. The two rock types appear to grade into one another in the field and have not been mapped separately. Peridotite makes up about 85 per cent of the complex. The rock forms reddish-brown angular massive outcrops speckled with coarse glistening flakes of talc. There are actually two adjacent intrusives of this composition separated by a thin septum of quartzose gneiss and a dike of amphibolite. The northernmost body, the smaller of the two, has been far more drastically altered by hydrothermal activity and is the ore body of the magnesite mine.

Though the primary texture of this rock has been partly obliterated by growth of hydrous phases it is evident that the original rock was a coarse-grained equigranular igneous rock with no linear or planar fabric. The primary pyroxene (now replaced by amphibole) may have been in poikilitic intergrowth around the olivine. There is no evidence of crushing or squeezing during or after the primary crystallization. The rock consists of the following minerals, listed in decreasing order of abundance:

1) Olivine: Fo_{90}
2) Talc: $n_z = 1.570$, $2V = 15^\circ(-)$; greenish white in hand specimen
3) Actinolite: $Z \wedge C = 18^\circ \pm 2^\circ$; $n_z = 1.648$; $Tr_{75}Ft_{25}$(Winchell and Winchell, 1951, p. 433)
4) Chlorite: $n_z = 1.590$ brownish gray in coarse grains; $2V = 15^\circ(+)$ Prochlorite (?)
5) Limonite
6) Chrysotile
7) Magnetite

The olivine has been replaced along fractures by chlorite and/or serpentine and minor amounts of magnetite and limonite (Plate 8a, appendix). The tremolite and talc occur as radiating clumps of coarse crystals interstitial to the olivine. The tremolite is inferred to have formed from a clinopyroxene with an equal or higher Ca:Mg ratio. The other minerals are alteration products of the olivine and possibly of traces of orthopyroxene which may have been present. Even in rock which is almost completely converted to hydrous minerals serpentine is a minor phase, less than 5 per cent by volume. However local patches within the intrusive body are completely converted to serpentinite. These were presumably dunite patches in the original intrusive. Anthophyllite occurs in the more highly altered peridotite with $n_z = 1.640$, $n_y = 1.625$, giving a composition of $Mg_{.75}Fe_{.25}$-anthophyllite assuming no aluminum to be present (Rabbitt, 1948).

The parent igneous rock is inferred to have consisted of 80 to 100 per cent olivine of Fo_{90} composition and a remainder of clinopyroxene (augitic diopside) and orthopyroxene. About half of the olivine is left in the rock now; the pyroxene is completely replaced by amphibole, etc.

Replacement by carbonate minerals is extensive in the smaller intrusive (the magnesite ore-body) though minor amounts of carbonate are present disseminated through the larger intrusive. The principal carbonate phase is presumably magnesite (not distinguished optically from calcite in this study). A composite sample of ore from the quarry was analysed by Gale (1912), and the analysis is given below:

SiO_2	6.17	
Al_2O_3 + Fe_2O_3	0.80	
CaO	tr	
MgO	43.80	
CO_2	45.02	
undet.	4.14	(H_2O?)

This is representative only of the vein material in the ore-body, but it can be seen that this material could be produced by addition of H_2O and CO_2 and removal of SiO_2 from a dunite or orthopyroxene peridotite. In thin section the carbonate ore is seen to contain fragments of olivine (Fo_{95}) and enstatite (which is not completely replaced in this unit) embedded in a very fine-grained matrix of silica, limonite and magnesite. Talc is disseminated in fine- to coarse-grains, as in the other peridotite body. Veinlets of siliceous magnesite cut this rock, ranging in thickness from less than a millimeter to 4 centimeters. Some of the finer veinlets are lined with chrysotile being replaced by magnesite. In outcrop on the quarry wall the thicker white magnesite veins stand out in relief against the dull ocher friable altered peridotite. The veins have undulating planar walls and intersect an angles close to 90°. There is no tendency towards orientation of the veins relative to one another or to the intrusive contacts, suggesting that they were formed by internal expansion due to growth of new crystals, at low confining pressures.

One dike of possibly highly altered peridotite was found on the north flank of the mine ridge, consisting largely of dolomite. This dike shows a zonal pattern of alteration like that observed by Phillips and Hess in serpentines in New England (Phillips and Hess, 1936). The outer margins of the dike adjacent to the country rock (biotite-sillimanite gneiss) are composed of matted crystals of green and colorless chlorite in a layer 1/8 inch thick. Within this zone is a zone of radially clustered anthophyllite needles (n_z = 1.640). The core of the dike is fine-grained dolomite with a dispersed unidentifiable fibrous mineral, n_z = 1.460 to 1.450, parallel extinction, yellowish white, H = 5. The chlorites referred to have the following properties:

green: $n_x = 1.580$ $2V(-) = 10^\circ$
colorless: $n_x = 1.560$ $2V(-) = 15^\circ$ (antigorite)

ii. Pyroxenite

The pyroxenite is invariably coarse-grained, containing single crystals of pyroxene up to 8 inches long and 2-3 inches wide. The crystals are arranged in a random interlocking pattern or rarely in radiating clusters.

The principal constituent mineral is orthopyroxene, making up over 90 per cent by volume of the rock. It is tan to olive brown in hand specimen; n_z = 1.670; density = 3.1 ± 0.1 (determined on a Jolly balance). Uncertainty in the density arises from the presence of hydrous alteration minerals enclosed in the crystals. The estimated composition is En_6Fs_4 (Winchell and Winchell, 1951, p. 406). The optical properties do not appear to vary with locality or grain size.

Anthophyllite and talc are intergrown with the pyroxene. The relative amount of anthophyllite in this rock cannot be determined very well in thin section because of the similarity of its optical properties to those of pyroxene. X-ray diffraction patterns of the bulk rock show only very weak reflections from the strongest anthophyllite lines, relative to those of orthopyroxene.

iii. Calc-silicate marble (carbonatite)

A strip of marble containing calc-silicate minerals extends continuously (except for alluvial cover) along the upper (northeastern) contact of the peridotite sill on the west flank of the syncline. Near the axis of the syncline it terminates in a bulbous mass. A single small outcrop of marble has been found on the northeast flank of the main fold.

This rock is generally massive or weakly foliated with silicate minerals homogeneously distributed throughout (Plate 8d, appendix). Locally there are pods or laminae of silicate-rich rock. The more pure marble weathers to a brownish gray and the silicate accumulations stand out as dark brown streaks and clots. The following minerals are present, in order of decreasing abundance:

calcite: rare lammelar twinning
olivine: $2V(+) = 85^{\circ}$
diopside: $Z \wedge C = 35^{\circ}$ $2V(+) = 60^{\circ}$
tremolite: $2V(+) = 80^{\circ}$; $n_z = 1.644$ colorless
chlorite: $2V(+) = 10^{\circ}$; colorless; $n_x = 1.575$
polysynthetic twinning; + penninite
magnetite
pyrrhotite
pentlandite intergrown in fine equant grains

The latter three minerals have been observed by mineragraphic study of the marble in polished sections.

The calcite is in medium-grained polyhedral crystals. Olivine and diopside form subhedral crystals disseminated through the calcite matrix. Tremolite is in inclusions in the diopside. The chlorite occurs as medium-grained clumps of fine crystals clustered around magnetite, suggesting pseudomorphous replacement of a silicate mineral (pyroxene?) with a higher Fe:Mg ratio than the chlorite.

The $MgCO_3$ content of the calcite from this unit has been determined using the (112) interplanar spacing according to the method of Goldsmith and Graf (1958). Two samples of calcite thus analysed gave $MgCO_3$ contents of 6.6 per cent and 7.5 per cent respectively. No dolomite lines were observed in the powder patterns. Using the calcite-dolomite solvus determined by Goldsmith and Graf, the minimum temperature at which these calcites might have crystallized in equilibrium with dolomite would have been 570° C and 620° C respectively. Dolomite is not present in this rock but the system is evidently saturated with respect to MgO since no quartz or other non-magnesian silicates are present. These temperatures are in general agreement with the suggested stability field of the assemblage calcite-diopside-forsterite (-quartz) given by Bowen (1925).

The marble unit is in contact with the peridotite-dunite complex over most of its length of outcrop. The contact is knife-sharp. In contrast the contact with the schist or gneiss is vague and marked by a zone of quartz-tremolite rock due to reaction between the marble and the gneiss. In one locality two small pods of peridotite are enclosed by the marble.

Where banding is developed in the marble it is seen to conform to the contacts bounding the marble body from all other units. In particular, within the boot-shaped area in the axial region of the syncline banding conforms closely to the marble-peridotite contact. Considering the irregular shape of this contact the conformity of the banding suggests that it was formed during plastic deformation of the marble such as might be caused if it had been intruded into its present location. Such flow banding is not uncommon in marbles of the amphibolite facies, however. It is significant to note here that the adjacent peridotite has no such flow-banding features nor any other textural features to suggest post-crystallization deformation. These facts plus the fact that this is the only occurrence of limestone or marble of any sort in the Winchester Hemet area or its immediate environs, suggests that this marble is actually an intrusive rock, that is, a carbonatite, and that it was intruded after the crystallization of the peridotite. Further evidence on behalf of this hypothesis is offered below.

iv. Amphibolite

Near the intersection of the axis of the fold with the ultrabasic complex two roughly elliptical bodies of dense black amphibolite occur. In addition small concordant dikes occur in the adjacent country rock resembling in all respects the typical intrusive amphibolite described earlier in this chapter. The larger bodies are composed of rock whose composition is given below: (volume per cent, estimated)

plagioclase	35
hornblende	60
augite	5
magnetite	1
pyrrhotite	tr

The rock is equigranular, granoblastic and fine-grained. A striking textural feature of the rock is its brecciated appearance on weathered surfaces, indiscernible in the fresh rock. The breccia consists of 1 inch angular clasts of normal amphibolite in a plagioclase-rich matrix. This texture grades into a mosaic texture with only fine veinlets of plagioclase separating the clasts.

Patches of hornblendite of similar appearance occur within the marble. They contain no plagioclase and about 3 per cent dark brownish-red garnet in fine porphyroblasts surrounded by a concentric halo of colorless amphibole (tremolite, $2V(-) = 85^\circ$, $Z \wedge C = 18^\circ$).

The larger mass of amphibolite terminates on the south in a dike of the same composition that strikes north-south and follows the trace of a small fault cutting the peridotite. This is taken as evidence that the amphibolite is younger than the peridotite. However the amphibolite itself is somewhat sheared along this fault developing a weak foliation. Evidently movement along this fault continued after intrusion of the igneous parent of this amphibolite.

A distinctive type of amphibolite intrudes the marble in one small poorly exposed pod on the northwest limb of the intrusive complex. It is a porphyritic rock consisting of 75 per cent plagioclase (An_{70}) and 25 per cent brownish-gray actinolite, faintly pleochroic in shades of pink, and traces of calcite, chlorite and sphene. The amphibole is in euhedral grains several centimeters long and 1 cm. wide, embedded in a matrix of 1 mm. plagioclase grains.

v. Contact effects

A halo of porphyroblastic cordierite-biotite-quartz gneiss surrounds the peridotite (Plates 8b and c, appendix). Cordierite is in porphyroblasts up to 3 centimeters across. Sillimanite is virtually absent from this rock although it makes up about 10 per cent of the adjacent unaltered gneiss.

A typical specimen of this rock contains the following estimated mineral proportions: (volume per cent)

quartz	15	
K-feldspar	10	
cordierite	40	
plagioclase	10	(An_{45})
biotite	20	(reddish-brown; pale orange)
sillimanite, tourmaline, apatite	tr	

Relict sillimanite is enclosed in very coarse subhedral poikiloblasts of cordierite. These are surrounded by medium-grained potash-feldspar crystals, also poikiloblastic. Quartz occurs as round equant inclusions in all phases. Biotite is in fine euhedral laths. The texture is granoblastic with a faint foliation defined by opaque inclusions in the cordierite grains (visible in thin section) and by oriented biotite flakes. Biotite tends to be segregated to the perimeters of the large cordierite porphyroblasts.

The zone of porphyroblastic gneiss can be traced continuously around the contact between the peridotite-dunite complex and the biotite-sillimanite gneiss. The zone is from 10 to 30 feet thick.

It is thought that this zone represents the outward diffusion of MgO from the intrusive and its reaction with sillimanite and quartz of country rock as follows:

$$\underbrace{2MgO + 2Al_2O_3 + 3SiO_2}_{\text{sill. + quartz}} = \underbrace{Mg_2Si_5Al_4O_{18}}_{\text{cord.}} \qquad (1)$$

The work of Yoder (1952) on the system Al_2O_3-MgO-SiO_2-H_2O indicates cordierite to be stable over a very wide range of temperature and composition in equilibrium with magnesium silicates. The porphyroblastic zone is not present adjacent to the marble suggesting that the marble unit was present at the time of the diffusion outward of magnesium and that it was impervious to the flow of these ions. The magnesium silicate minerals and traces of nickel found in the marble may be in part derived from the diffusion process.

Immediately adjacent to the metaperidotite on its southeast flank is a thin zone separating the ultrabasic from the cordierite gneiss. In this zone, less than a foot wide, the gneiss has been converted into rock of the following estimated composition (volume per cent):

biotite	60
sillimanite	30
tourmaline	5
quartz	3
potash feldspar	2

The sillimanite forms very coarse (3-5 cm.) poikiloblasts in which are embedded randomly oriented coarse tablets of light orange-brown biotite and medium-grained euhedral prisms of green tourmaline. Quartz and feldspar are restricted to the interstices between coarse sillimanite grains. The biotite has a refractive index (n_z) of 1.636, slightly less than that typical of the biotite in the gneissic country rock (about 1.645).

The mineralogy of this zone can be related to that of the cordierite rock by a reaction such as the following:

$$\underset{\text{K-feldspar}}{2KAlSi_3O_8} + \underset{\text{cordierite}}{3Mg_2Al_4Si_5O_{18}} + H_2O = \underset{\text{phlogopite}}{2KMg_3Si_3AlO_{10}(OH)_2} + \underset{\text{sillimanite}}{6Al_2SiO_5} + \underset{\text{quartz}}{6SiO_2} \qquad (2)$$

However it is to be noted that the system represented by the mineral assemblages shown above was open to diffusion of MgO and SiO_2 as discussed above. Therefore any simple closed reaction such as the above may have been modified during contact metamorphism by diffusion of one or more of the components. In particular, viewing the scarcity of quartz in this contact zone it appears that SiO_2 has diffused into the ultrabasic, driving this reaction to the right, while the silica has participated in one of the reactions discussed above converting the primary igneous minerals of the peridotite to a metamorphic assemblage.

c. Origin of the Ultrabasic and Associated Rocks

There were apparently three or four separate intrusions of ultrabasic rock into the area of the magnesite mine. The first two, probably simultaneous, were injections of mixed peridotite and dunite sills along the flanks of the syncline, swelling out somewhat near the axis. The primary peridotite was composed of olivine, clinopyroxene and enstatite; the smaller peridotite body may have consisted of olivine and orthopyroxene only.

Later a series of dikes of pyroxenite was intruded, in part along the contact between the peridotite and the gneiss. It appears from the structure of the intrusive that the country rock yielded plastically to all these intrusions and the contacts are mostly conformable with foliation of the gneiss.

There is no zonal or layered pattern to indicate differentiation in place to produce the three ultrabasic rock types. Although they have been subsequently metamorphosed each preserves enough of its primary texture to show it to be a homogeneous igneous-looking rock. It seems reasonable to conclude that each crystallized in place from a magma of ultrabasic composition. Perhaps all three types crystallized from the same magma at different temperatures though it seems as if some structural indication (zoning or layering) would be preserved to indicate this relationship.

The origin of the carbonatite, if such it is, is more problematical. Two hypotheses have been entertained by the author:

1) Intrusion by plastic flow of a siliceous dolomite stripped from unexposed sedimentary rocks at depth;

2) prior intrusion of an ultrabasic rock of appropriate composition (augite-peridotite), subsequent total replacement by hydrothermal solutions forming a silica-dolomite rock and finally, contact metamorphism during intrusion of a later ultrabasic sequence.

A third and seemingly equally plausible hypothesis, namely that the marble is a sedimentary unit in the French Valley formation, must be considered unlikely. This is the only occurrence of such marble over the whole area studied. It would be extremely fortuitous for this sole exposure also to be the sole locus of large-scale ultrabasic intrusion.

In support of the first hypothesis, it is known that limestones and some siliceous dolomites occur in the stratigraphic pre-batholithic section of southern California (Woodford and Harriss, 1928; McKevett, 1950). In addition the pronounced layering in the marble, although in part attributable to primary sedimentary layering, appears to have been plastically deformed in a fashion suggesting flow banding. It is thought that, by this mode of origin, the ultrabasic magma as it was intruded, melted a limestone or dolomite unit at depth. This carbonate magma would in turn be capable of being intruded as a sill adjoining the peridotite sill. Experimental work on carbonatite magmas by P. J. Wyllie (Tuttle and Wyllie,

1958) indicates that such a magma could exist at temperatures of 740° C at a partial pressure of water of 1 kb. This indicates that a carbonate melt could be produced adjacent to an ultrabasic magma. The texture of the marble, and in particular the small degree of lamellar twinning and gliding in the calcite grains suggests that the marble did not flow appreciably after final recrystallization.

Support of the second hypothesis lies in the similar occurrence of silica carbonate rocks in association with serpentines through the Coast Ranges and in the Santa Ana Mountains. Such an occurrence has been studied by the writer where serpentine intrudes the Bedford Canyon formation near Santiago Peak in the Santa Ana Mountains. This rock type has been described by A. C. Lawson (1893), E. H. Bailey (1942, p. 151) and A. Knopf (1906) from various California localities. In a typical occurrence it consists of very fine-grained calcite veined with amorphous silica. Traces of serpentine, limonite and talc are scattered throughout. Such a rock type is thought by the above authors to have been formed by leaching of a peridotite with CO_2-rich solutions, selectively removing MgO and leaving a residue of calcite, silica, and hydrated iron oxide. However no instances of high-temperature mineral assemblages associated with the carbonate are reported in the literature of this type of deposit. In fact, this phenomenon seems to be more commonly one of a low temperature, low pressure environment, associated with hydrothermal low-grade alteration of the adjacent igneous and metamorphic rocks. If this is a generally valid observation then it would imply something like the following sequence of events: 1) intrusion of an augite-peridotite at depth; 2) elevation of the crust and hydrothermal alteration of the peridotite to silica-carbonate rock; 3) reburial of the silica-carbonate rock (perhaps followed by intrusion of a second ultrabasic body) and thermal metamorphism of the silica-carbonate rock to a calc-silicate high-temperature assemblage. Such a sequence of uplift and reburial is not otherwise recorded in the geological record of this area and does not seem very plausible.

The amphibolites of this locality were probably intruded as a gabbroic magma before metamorphism of the ultrabasic rocks but after intrusion of the carbonatite. This is inferred from the fact that they now contain a mineral assemblage of equivalent grade to that of the ultrabasic rocks and carbonatite.

At the time of intrusion of the ultrabasic magma the adjoining sediments had already undergone folding and some metamorphism. The lack of stress-induced fabrics in any of the ultrabasics (in contrast with the pronounced foliation and lineation of the gneisses) indicates that both primary crystallization from a magma and later metamorphism took place under essentially isotropic stress conditions. A minor amount of shearing could have taken place during primary crystallization without being recognizable at present due to later metamorphic textural changes. Possibly the metamorphism of the calc-magnesian silicates of the igneous rocks took place immediately on cooling from magmatic temperatures. Water may have been in part derived from the adjacent gneisses. As described above there is a zone of magnesium metasomatism around the peridotite indicating that there was diffusive transfer through the walls of the magma chamber. No extensive dehydration reactions can be shown to have taken place in the border zone though the production of cordierite may be in part the result of a reaction such as the following:

biotite + sillimanite = cordierite + K-feldspar + vapor (3)

The biotite content of the porphyroblastic border zone is not appreciably less than that of the unaltered gneiss.

d. Metamorphism of the Ultrabasic Rocks

The mineral assemblages found in the peridotite and pyroxenes may be considered in the light of the system $MgO-SiO_2-H_2O$ as discussed by Bowen and Tuttle (1949) and the system $MgO-SiO_2-Al_2O_3-H_2O$ discussed by Yoder (1952). In the present discussion we must neglect the role of CaO (which is evidently present in major amounts) since systems containing this component have not been studied extensively.

The following mineral assemblages are observed:

a) in the peridotite:

1) T + F + E + S + magnesite + Q
2) F + Ch + Tr
3) S + Tr + Q

b) in the pyroxenite

E + T + A

c) in the dunite

F + T + S

where F = forsterite, S = serpentine, Ch = chlorite, T = talc, E = enstatite, Tr = tremolite-actinolite, A = anthophyllite and Q = quartz (including amorphous silica).

In the first peridotite mineralogy listed if we examine the possible assemblages that this composition would permit (neglecting magnesite and quartz) we see that at a randomly given temperature for example, S + F + T or F + E + T could coexist but not all four. This is of course required by the phase rule, where c = 3. If one considers Al_2O_3 as a component capable of entering serpentine, however, the four phase assemblage may have a divariant stability field. The other two peridotite assemblages may similarly be at equilibrium though it is not known what relation the stability field of tremolite-actinolite would have to the other minerals. The pyroxenite assemblage as far as is known could be stable at the same P-T conditions as the three peridotite assemblages. However in order to form anthophyllite from a pure enstatite pyroxenite SiO_2 must be added:

$$SiO_2 + 7MgSiO_3 + H_2O = Mg_7Si_8O_{22}(OH)_2 \qquad (4)$$

That is, pure enstatite could not be altered isochemically (except for gain of water) to anthophyllite. The stability field of anthophyllite is also unknown.

The dunite assemblage could likewise coexist stably with the others listed. However in order to produce this assemblage from a pure dunite it is necessary to add SiO_2:

$$2Mg_2SiO_4 + 3H_2O + 4SiO_2 = \underset{\text{serp}}{Mg_3(OH)_4Si_2O_5} + \underset{\text{talc}}{Mg_3(OH)_2Si_4O_{10}} \qquad (5)$$

neglecting alumina in the serpentine. A possible source of silica is suggested by the textural evidence of complete replacement of diopside by tremolite:

$$5CaMgSi_2O_6 + H_2O = Ca_2Mg_5(OH)_2Si_8O_{22} + 2SiO_2 + CaO \qquad (6)$$

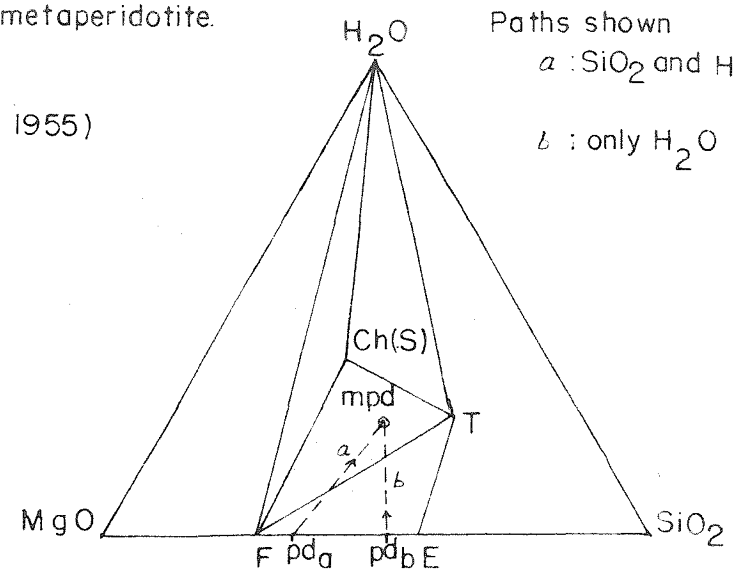
The CaO would be carried off in solution and the silica would be available to react with magnesium silicates. Alternately in the presence of CO_2, calcite could be a product of this reaction. In any case the textural evidence here is striking. Diopside, if it was present at all, is now completely absent. Tremolite forms a matrix throughout the peridotite, surrounding grains of forsteritic olivine which are themselves being replaced by chlorite and talc. Some or all of this tremolite may have been formed by the reaction

$$\underset{2D}{2CaMgSi_2O_6} + \underset{3E}{3MgSiO_3} + \underset{Q}{SiO_2} + \underset{V}{H_2O} = \underset{Tr}{Ca_2Mg_5(OH)_2Si_8O_{22}} \qquad (7)$$

The extent of this reaction would appear to depend on the amount of enstatite present initially. Possibly traces of enstatite may be present in the rock now, but not more than 1-2 per cent by volume. If this reaction rather than (6) is the source of the tremolite then it is evident that silica would not be made available for reaction with forsterite and in fact would be consumed.

A possible path for the metamorphism of the peridotite system is shown in figure 10 (after Yoder, 1955, p. 517). Assuming the system to be open to both water and SiO_2, the path of the system would be as shown by "a". This is equivalent to assuming some source of SiO_2 within the rock such as the reaction of diopside to form tremolite as

Fig. 10 Phase-diagram for the system $MgO-SiO_2-H_2O$ for the conditions of metamorphism of the metaperidotite.

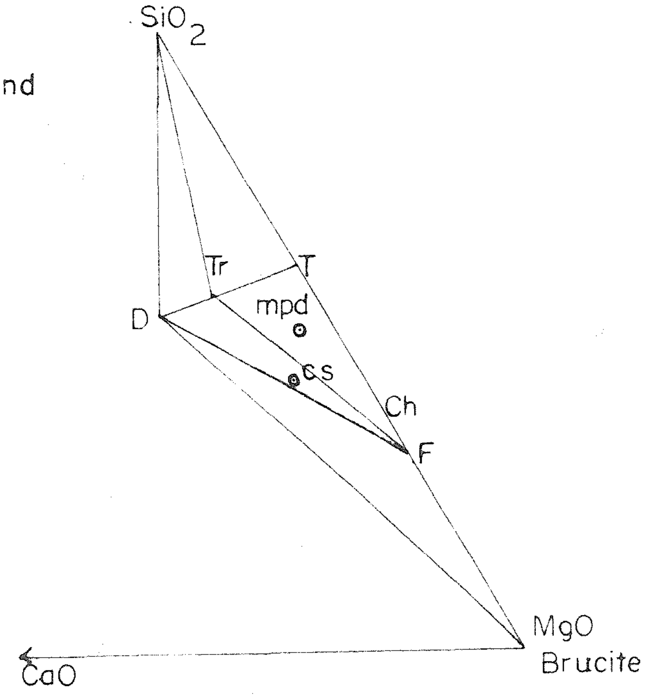


(after Yoder, 1955)

Paths shown
a : SiO_2 and H_2O added
b : only H_2O added.

pd_a, pd_b : initial peridotite ; mpd : final metaperidotite

Fig. 11 Phase diagram of the system $CaO-MgO-SiO_2$ at 500°-600° C and P_{h_2O} = 1-2 kb.



(after Fyfe, Turner and Verhoogen, 1958)

mpd : metaperidotite ; cs: calc-silicate marble (+calcite+Ch).

shown above. Alternately one may assume a different initial composition and arrive at the same final phase assemblage by simply adding water (as shown by path "b"). However it is evident that to accomplish this one must begin with a rock with considerably higher ratio of E:F. The final composition of the system, shown by point "mpe" has been estimated on the basis of inspection of thin sections of the metaperidotite and represents a composition of F:T:Ch = 1:2:2. It is seen that, with progressive addition of water (and SiO_2) the following assemblages will occur in sequence:

F + E (+ D)
F + E + T (+ D + Tr)
F + T + E + Ch (+ D + Tr)
F + T + Ch + (+ D + Tr) (observed assemblage)

"Chlorite" and "serpentine" cannot be sharply distinguished here. The alumina concentration, the final temperature of recrystallization and the pressure (P_{tot} and P_{H_2O}) may be critical factors in determining whether either or both phases may be present. Yoder's work (1952) suggests that for aluminous systems above a temperature of about 500° C serpentine is unstable and will be replaced by an aluminous chlorite.

From the above series of steps it can be seen that at a given T and P_{H_2O} the limiting factor in this series of reactions is the amount of silica available for reaction. That is, the ratio of Ch:F + E is roughly fixed by the P_{H_2O} (at constant temperature) while the ratio of T:F is largely dependent on the amount of silica available for the reaction F + Q + V = T + S.

Another possible source of silica is the country rock. It has been shown that the igneous contact was permeable to MgO and probably to H_2O as well. It is reasonable to presume that SiO_2 may also have migrated through this boundary. The metamorphism of the pyroxenite, in which there is no evidence of diopside having been present, would require addition of silica. Some outside source must therefore be inferred. Study of the contact halo described above indicates that SiO_2 has indeed been removed from the country rock.

Paired reactions similar to the ones described above have apparently gone on in the silicate phase of the marble producing the assemblage F + D + Ch + Tr. Talc may be excluded as a possible phase because of the high CaO concentration and the possibility of the reaction

$$4SiO_2 + 5Mg_3Si_4O_{10}(OH)_2 + 6CaCO_3 = 3Ca_2Mg_5Si_8O_{22}(OH)_2 + 2H_2O + 6CO_2 \quad (8)$$

The relatively small degree of alteration of the pyroxenite and in particular the complete absence of forsterite is puzzling in light of the previous discussion. Evidently we cannot consider the pyroxenite to have recrystallized under the same P_{H_2O}, T conditions as the peridotite since the reaction

$$E + V = F + T \qquad (9)$$

has not taken place to an appreciable extent. Yoder (1954, p. 517) has drawn a hypothetical plot of the sequence of divariant curves around the invariant point in the P-T diagram of the system $MgO-H_2O-SiO_2$ using the Morey-Schreinemakers principle. From this schematic plot it can be seen that one would expect the assemblage E + T + V to be stable at lower pressures and temperatures than the assemblage E + F + T. It is conceivable that retrograde metamorphism has selectively obliterated the higher-temperature assemblage (E + F + T) in the pyroxenite, leaving that in the peridotite. The relatively small degree of hydration of the pyroxenite (less than 10 per cent hydrous minerals) bears out this argument. As mentioned earlier any reaction to produce talc and anthophyllite from pure enstatite must involve addition of silica, perhaps by diffusion from the country rock.

Magnetite, found disseminated in all the meta-ultrabasic rocks, is a product of the conversion of pyroxene and olivine to hydrous silicates with a higher Mg:Fe ratio than that of the parent minerals. This can be seen exemplified in the replacement of forsterite by serpentine, where a fine film of magnetite is observed to run down the middle of the serpentinized cracks in the olivine.

The fact that serpentine is present in some of these assemblages indicates that the temperature of metamorphism was probably below 500° C (Bowen's curve II). This once more must neglect the effect of Al_2O_3. However the phase diagrams obtained by Yoder for a system containing alumina tend to confirm this observation since he finds the upper stability limit of aluminous serpentine to be near 500°C at 15,000 psi P_{H_2O}.

The assemblage magnesite-silica-serpentine found in the ore of the magnesite mine can be conclusively shown on the basis of textures to be later than all the others. Magnesite and silica form veins which cut every other mineral. Pseudomorphous masses of magnesite appear to replace olivine, enstatite, talc and serpentine in the matrix of the ore body. Serpentine, locally intergrown with the magnesite, is the only magnesian silicate phase that appears to be in equilibrium with it. Possibly this represents the reaction of Bowen's curve I, modified by the presence of CO_2:

$$\begin{aligned} &F + V = S + B \\ &B + CO_2 = \text{Magnesite} + V(H_2O) \end{aligned} \qquad (10)$$

However the presence of opaline silica intergrown with the magnesite indicates that reactions such as the following must also have occurred:

$$\underset{\text{serp}}{Mg_3Si_2O_5(OH)_4} + 3CO_2 = \underset{\text{magnesite}}{3MgCO_3} + \underset{\text{silica}}{2SiO_2} + 2H_2O \qquad (11)$$

This reaction as well as the previous one involve large volume changes. For carbonation of serpentine the volume change is 24 cc per mole or approximately 23 per cent increase in the solid phase. There is abundant textural evidence for the magnesite veins having forced apart the enclosing friable matrix. There is no clear indication on the basis of the structure of the ore-body that the magnesite-bearing peridotite has expanded as a whole.

The source of CO_2 for the formation of magnesite is not clear. It would seem that CO_2 should be present as liberated by the decarbonation reactions in the calc-silicate marble. However magnesite is absent from that unit itself, where it could have replaced forsterite. It is inferred from its texture and mineralogy that the magnesite alteration was produced at lower temperatures and later than the formation of the typical mineral assemblages of the calc-silicate marble and the non-magnesitic metaperidotite ("pd" as shown in fig. 6). In fact magnesite formation could have gone on at pressures and temperatures approaching those of surface conditions or mild hydrothermal alteration. Carbon dioxide may have been supplied by further decarbonation of dolomitic marble going on at depth in buried units similar to the one exposed here. The low pressures and temperatures of formation are also suggested by the texture and structure of the magnesite ore. The minerals filling the veins are very fine-grained (in contrast with the coarse-grained rocks which they replace). Opaline silica is intergrown with the magnesite. The irregular stockwork form of the veins suggests formation at low confining pressure and low shear stress, where the shape of the vein was principally controlled by the expansion of the vein material itself. The surfaces of the vein fillings have a colloform appearance, as if they were formed in open spaces and yet they evidently had to displace altered igneous rock during their growth. This suggests that confining pressures were very small while they were growing.

Considering the effect of CaO as a component we may plot the final assemblages on a diagram of the system CaO-MgO-SiO_2 (fig. 11) taken from Fyfe, Turner and Verhoogen (1958, p. 210). The partial pressure of CO_2 is assumed sufficient to convert all brucite to magnesite. The latter authors suggest that this diagram is valid for temperatures of 500-600° C and pressures of 1-2 kb. It is evident that the phase assemblages of the marble (F + D + Tr) would be stable at the same conditions as those observed in the metaperidotite (F + T + Ch + Tr). The two assemblages differ only in the amount of CaO present in the system. The assemblage observed in the pyroxenite is evidently metastable at these same conditions (as noted above).

D. INTRUSIVE ROCKS OF THE SOUTHERN CALIFORNIA BATHOLITH

1. General Character of the Batholith

The major intrusive igneous bodies of the Winchester-Hemet area are composed of plutonic rocks ranging in composition from pyroxenite to leucogranodiorite. Free quartz appears in all rocks more acidic than gabbro. Tonalite is the most abundant type in area of exposure, followed by granodiorite, gabbro and pyroxenite in that order. These plutonic masses are a part of the southern California batholith, a series of granitoid plutons which is exposed from the latitude of Riverside to the southern tip of Baja California in a strip paralleling the coast. The general character of this batholith and the petrology of its igneous rocks has been discussed by E. S. Larsen (1948).

The plutonic igneous rocks of the Winchester-Hemet area have been described by Larsen. The present discussion will be in part a review of his description with additional details added where especially pertinent to the problem of metamorphism in this area.

Larsen distinguished five plutonic igneous rock types in this area: San Marcos gabbro, Bonsall tonalite, Lakeview Mountain tonalite, Woodson Mountain granodiorite and Domenigoni Valley granodiorite. The type localities of two of these rocks are in the area under study: the Lakeview Mountain tonalite and the Domenigoni Valley granodiorite. The other rock types are named for localities generally far removed from this area. The San Marcos gabbro type locality, for example, is 25 miles away. In correlating these igneous rock types over great distances with discontinuous exposures inbetween it is not always implied that there was a single synchronous period of emplacement of the given rock type over the whole area from a single magmatic source. Rather the rocks represent igneous types with common textural and compositional features and with a common sequence of emplacement with respect to one another.

Evidence has been found in this area that in some details the sequence of emplacement proposed by Larsen may not be valid. In particular it appears that the San Marcos gabbro may be younger than the tonalites of this area, and perhaps younger than some of the granodiorite as well. In some other details the present discussion will diverge in interpretation from that of Larsen, though for the most part evidence obtained in this present study has confirmed Larsen's observations in this area.

2. Descriptive Petrology

a. Lakeview Mountain Tonalite

The Lakeview Mountain tonalite is a light gray, medium-grained rock with from 0 to 25 per cent dark inclusions. It has a weak planar structure due to orientation of hornblende and biotite crystals, increasing in intensity near the intrusive contacts. Although it is principally a true quartz diorite in composition, it grades locally into granodiorite. Six modal analyses from scattered localities in the Winchester-Hemet area are given in table 1. A graph of various mineral abundances against quartz content is shown in fig. 12.

Lakeview Mountain tonalite is exposed in two major plutons, one in the Lakeview Mountains and the other south of Hemet. These two may be connected beneath the alluvium of the San Jacinto Valley. A single mound-shaped exposure south of Florida Ave. in section C18 is made up of

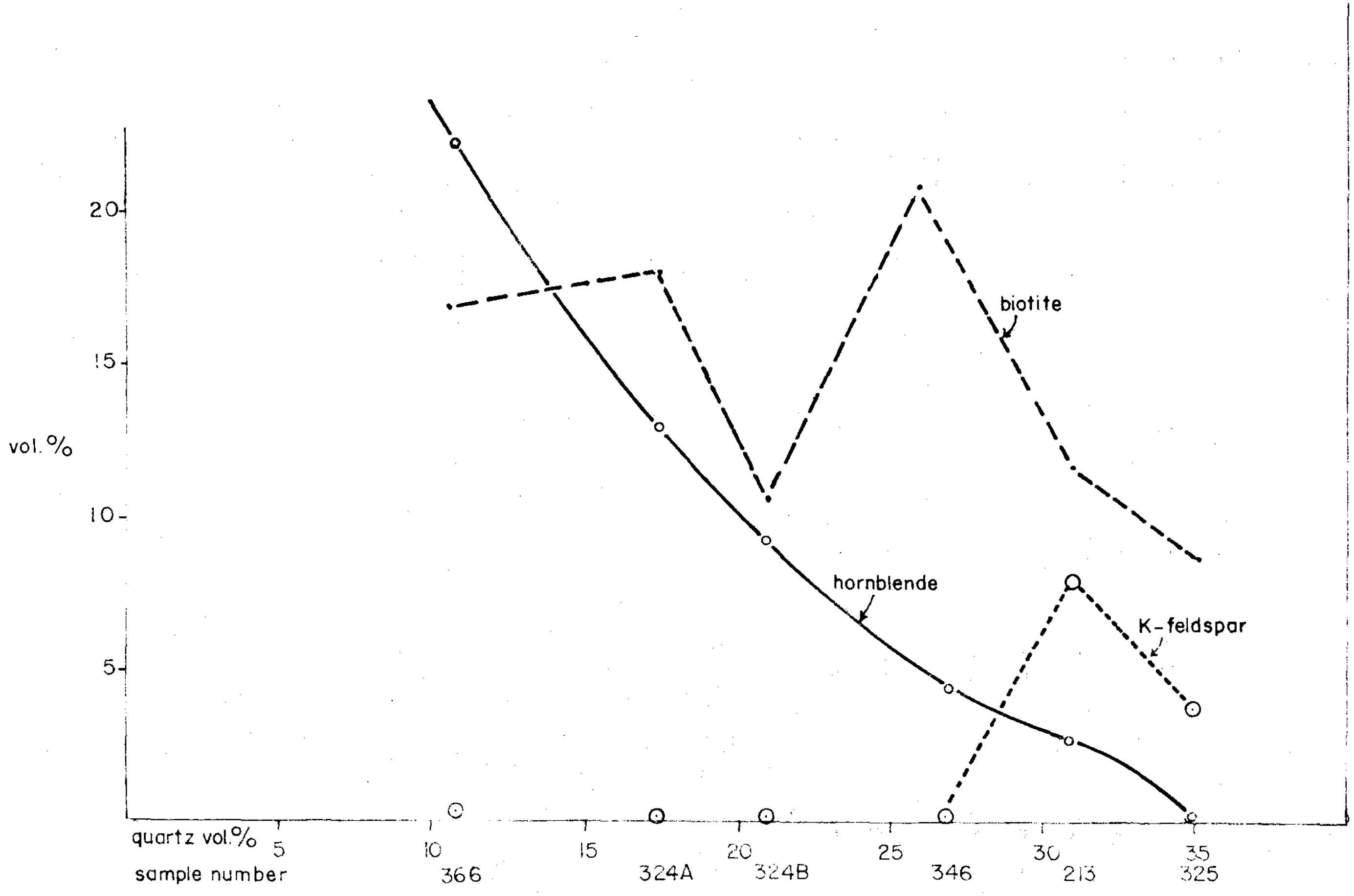
TABLE 1. Modal Analysis of the Lakeview Mountain Tonalite, Winchester-Hemet Area, California

	336	324A	324B	325	213	346	Average
Plag.	51.57	52.33	60.13	52.43	46.94	49.09	52.08
Quartz	11.61	18.37	21.83	35.30	31.72	27.95	24.46
Biotite	15.74	16.33	9.38	7.53	10.24	18.65	12.98
Hornblende	19.74	11.30	7.82	0.17	2.34	3.63	7.50
Magnetite	1.07	0.99	0.60	0.00	0.10	0.11	0.48
Apatite	0.07	0.37	0.12	0.28	0.10	0.34	0.21
K-Feldspar	0.20	0.00	0.00	3.90	8.23	0.11	2.07
Chlorite	0.00	0.21	0.12	0.06	0.00	0.06	0.08
Myrmekite	0.00	0.00	0.00	0.28	0.05	0.00	0.06
Epidote	0.00	0.10	0.00	0.06	0.10	0.06	0.05

Sample location:

	East side of Black Mountain, B28(3, 2), Sage quadrangle
324A	Pixley Canyon road, B15(4, 2), Hemet quadrangle
324B	Pixley Canyon road, B15(4, 1), Hemet quadrangle
325	Gibbel road, C27(3, 1)
213	Ramona Bowl, C23(3, 2) east of map area in Hemet quadrangle
346	Southeast of Crown Valley, B21(3, 4), Hemet quadrangle

FIG.12 Mineral variation diagram, Lakeview Mountain tonalite.



this rock. In addition an elongate plug of tonalite intrudes the metasedimentary rocks 0.6 miles west of the main intrusive mass in the Hemet quadrangle (section B21).

Outcrops of this rock form huge white boulders of disintegration characteristic of most of the granitic rocks of the southern California batholith. The tonalite underlies broad mound-like hills rising to elevations of over 2500 feet east of the Winchester-Hemet area.

The Lakeview Mountain tonalite consists of medium-grained subhedral crystals of slightly zoned plagioclase, euhedral medium- to coarse-grained hornblende and biotite and anhedral interstitial quartz. There may be present lesser amounts of fine-grained interstitial microcline and myrmekite (albite and quartz), zircon, sphene, apatite, magnetite, chlorite and epidote. One section contained traces of colorless pyroxene enclosed in hornblende (WH 324A).

Intrusive contacts with the gneisses of the French Valley formation are knife-sharp and sub-concordant. No exposed contacts between tonalite and other igneous rocks were seen. Where Florida Ave. passes south of the Lakeview Mountains there is found a well-foliated zone in the Lakeview Mountain tonalite. This zone is exceptionally rich in inclusions of both mafic igneous rocks and quartzose gneiss, sheared out into schlieren. This rock is also cut by numerous concordant aplite and pegmatite dikes. This type of rock forms a border zone around the major body of weakly foliated tonalite. Larsen considered this to be a distinct intrusive of Bonsall tonalite, a rock type which underlies much of the Perris block. Bonsall tonalite is distinguished from Lakeview Mountain tonalite solely on the basis of its greater content of dark inclusions (Larsen, p. 59).

Since there appears to be a continuous (though rapid) gradational decrease in abundance of inclusions from this zone into the central tonalite mass there seems to be no basis for distinguishing this rock as a separate intrusive. Such examples of schlieren-rich border zones around plutonic intrusives are common. Since this is the only locality referred to by Larsen as establishing the relative ages of these two rock types it seems possible that elsewhere the true "Bonsall" type may be an inclusion-rich part of the Lakeview Mountain tonalite magma body. This inclusion-rich zone is shown on fig. 1, where it is referred to as " 'Bonsall' tonalite, border phase of Lakeview Mountain tonalite".

This same border zone is in contact with San Marcos gabbro just south of the Lakeview Mountains, in D15(4, 2). The actual contact is buried by alluvium and soil. It is noted however that the well-defined flow banding in the tonalite, defined by smeared-out inclusions, strikes perpendicular to the inferred trend of the tonalite-gabbro contact. This has been inferred by the writer to indicate that the tonalite is older. This fact is also indicated by intrusive relations in the southeast corner of the Winchester-Hemet area where Woodson Mountain granodiorite intrudes the Lakeview Mountain tonalite and is in turn intruded by gabbro. This is contrary to the age sequence deduced by Larsen who considers the San Marcos gabbro to be the oldest rock type in the southern California batholith.

b. Woodson Mountain Granodiorite

The Woodson Mountain granodiorite occurs as a single sub-concordant pluton in the southern part of the Winchester-Hemet area. A thick dike-like extension of this body extends up the valley south of

Rawson Ranch. Outcrops of granodiorite possibly correlatable with the Woodson Mountain granodiorite also occur in the southwestern corner of the map area.

Outcrops of this rock weather to a reddish-gray gruss. Rare boulders of disintegration rest on broad rounded outcrops. Inclusions of darker rock are rare. In the strip south of Rawson Ranch the granodiorite is extremely sheared. An intense vertically-plunging lineation occurs in the rock here, formed by the streaking-out of biotite clusters and all other minerals of the rock. This sheared rock has no foliation however and outcrops eroded perpendicular to the lineation give no hint of the profound shearing (Plate 9a and b). This gneissic phase of the granodiorite is more leucocratic than the bulk of the Woodson Mountain granodiorite. It may contain less than 10 per cent biotite in places. Traces of garnet are present here as trains of fine pink grains. Elsewhere in this area the granodiorite is a medium-grained equigranular weakly foliated rock.

Modal analyses of four specimens of Woodson Mountain granodiorite are given in table 2. The biotite and hornblende are subhedral and commonly intergrown. Plagioclase is weakly zoned and anhedral. Quartz and untwinned perthitic potash-feldspar are interstitial to the other phases. Muscovite and sericite occur as alterations of the plagioclase.

Contacts between the granodiorite and other rock types are seldom exposed. Southeast of the map area along a stream that crosses the contact between tonalite and granodiorite the latter rock is sheared, the degree of shearing increasing with proximity to the contact. The

TABLE 2. Modal Analyses of the Woodson Mountain Granodiorite, Winchester-Hemet Area, California (Volume Per Cent)

	368	155	152	353	Average (W-H)	Average (SCB)
Plagioclase	37.17	27.22	38.39	50.58	38.4	42
Quartz	42.95	41.30	27.00	25.25	34.2	32
K-Feldspar	10.32	25.33	27.67	5.53	17.2	20
Biotite	8.39	4.82	6.62	12.63	18.3	5
Hornblende	0.23	0.09	0.00	5.14	1.4	1
Muscovite	0.00	1.13	0.00	0.00	0.3	
Zircon	tr	0.00	0.08	0.47	0.1	
Garnet	0.00	0.00	0.34	0.00	0.1	
Myrmekite	0.93	0.00	0.00	0.00	0.2	

Location of Samples:

368	South side of Black Mountain, south of map area, B33(2, 1)
155	South of Rawson Ranch, B30(3, 4)
152	South of Rawson Ranch, B19(1, 1)
353	Along southern Rawson Canyon, B30(3, 2)

Average (SCB) is taken from Larsen (1948), Table II, column 8, p. 81, "The average Woodson Mountain granodiorite", converted to volume per cent.

tonalite adjacent to the contact is unsheared. This is interpreted as indicating that the granodiorite was intruded later than the tonalite as a partially crystallized mush and sheared near its contacts by viscous drag. Textural evidence in thin sections of sheared granodiorite indicates that at least some of the potash-feldspar and quartz crystallized after most of the crushing of the coarser grains had taken place.

The contact between gabbro and granodiorite is not exposed though its location can be well-defined. The sheared zone in the granodiorite in section B19(1, 1) is thought to have been produced by drag against the gneisses and schists and the gabbro was intruded later. The age-relation of the granodiorite and gabbro is discussed below.

The general trend of contacts between gneisses and schists and the Woodson Mountain granodiorite is concordant with the foliation in the metasedimentary rocks. Where clearly exposed, as in the high plateau southwest of Black Mountain, the contact is exactly conformable and knife sharp (Plate 9c). There is no obvious relation between degree of migmatization of the gneisses and distribution of Woodson Mountain granodiorite (such as will be discussed in relation to the Lakeview Mountain tonalite in the next chapter). Partially digested inclusions of schist, especially common near the contacts, are the only inclusions noted in the granodiorite.

Smaller bodies of granodioritic composition found scattered through the southeastern part of the area resemble the Woodson Mountain granodiorite. In some cases they may be later than the main mass of granodiorite and may even represent (as in the case of the granodiorite intruding the migmatite-breccia) secretionary or anatectic granitic rock derived locally from the metasedimentary rocks.

In a "decomposed granite" quarry in the southwest corner of the area (sections A19, 30) (indicated as a "gravel pit" on fig. 1) there is exposed a granodiorite texturally similar to the Woodson Mountain granodiorite. It is deeply weathered and poorly exposed. Associated with it are dikes of a distinctive porphyritic granodiorite, apparently cutting the San Marcos gabbro. This dike rock contains the following approximate mineral proportions: (volume per cent)

hornblende	2	
biotite	3	
quartz	30	
plagioclase	40	(An_{20})
K-feldspar	25	

The rock consists of coarse phenocrysts of plagioclase and quartz in a seriate groundmass of fine- to medium-grained quartz, feldspar and biotite. Some of the plagioclase phenocrysts have equant cores of microcline (rapakivi texture); and most of them are weakly zoned. Adjacent to the contact with the granodiorite the gabbro is permeated with quartzo-feldspathic material.

c. San Marcos Gabbro

The San Marcos gabbro occurs in four major plutons: south of the Lakeview Mountains in D14, D23 and D26; an incomplete ring dike south of Rawson Ranch; underlying Bachelor Mountain; and in the southwestern corner of the area along Leon Road. A small plug of pyroxenite in section A23 and various dikes of hornblendite south of Rawson Ranch are thought to be related to the San Marcos gabbro.

Outcrops of the San Marcos gabbro are rounded and smooth. Hills underlain by this rock lack the characteristic weathered boulders of the other plutonic igneous rock types. They are colored light gray although the fresh rock is very dark gray, almost black. A dark reddish-brown soil forms on the gabbro in areas of low relief, permitting one to trace its outcrops below soil cover.

The typical San Marcos gabbro is a hornblende gabbro. In hand specimen it has a mottled uneven texture due to the poikilitic habit of the hornblende. A modal analysis of this rock is given below (from Larsen, 1948, p. 51, column 2)

El-303: southeast slope of Bachelor Mountain

plagioclase	47
hornblende	34
hypersthene	6
augite	5
Fe-ore	8
quartz	tr

In thin section greenish-brown hornblende is seen to form a poikilitic matrix around subhedral plagioclase. A later fibrous blue-green hornblende is less commonly found attacking the brown hornblende. Hypersthene is present as relict patches with poikilitic habit. The plagioclase, though not strongly zoned, is extremely calcic, up to An_{93}. Spinel in vermicular intergrowth with hornblende, stout apatite prisms, anhedral masses of ore minerals and rare pyrite or pyrrhotite are accessory minerals. The plagioclase grains in the typical gabbro are medium-grained. The poikilitic ferromagnesian minerals may be up to 3 cm. across.

Although most of the gabbro masses of this area are composed of such a hornblende, 2-pyroxene gabbro, zones of irregular shape within the main mass have markedly different texture and composition.

Only the body south of the Lakeview Mountains has been studied in sufficient detail to base comment on these features. The varietal types of gabbro are:

Leucogabbro: A zone of leucocratic rock cuts through the main gabbro mass forming a lens about 1000 feet long with gradational contacts against the principal mela-gabbro. This rock consists of 6.2 per cent pale green hornblende, 92.7 per cent plagioclase (An_{92}), 1 per cent augite. It lacks any of the usual accessories. The hornblende is in medium-grained polycrystalline aggregates surrounding cores of colorless pyroxene. The rock is medium-grained and weathers to white bouldery outcrops.

Porphyritic gabbro: Coarse-grained gabbro containing very coarse euhedral phenocrysts of plagioclase occurs as a lens on the south end of the ridge in D23(3, 4). It is compositionally a norite: 20 per cent hypersthene, 5 per cent augite, 15 per cent hornblende (brown to greenish brown), 60 per cent plagioclase (An_{86}) and traces of colorless amphibole (anthophyllite) and vermicular spinel, both rimming hypersthene. Essentially all the plagioclase is in coarse polycrystalline phenocrysts. Hornblende forms a poikilitic matrix around medium-grained pyroxene crystals.

Gabbroic pegmatite: Not shown on the map are lenses of pegmatite, exposed near the crest of the ridge in D23. These contain coarse grained gray plagioclase and hornblende crystals up to 6 inches long and 1-2 inches wide. The lenses parallel the weak banding in the adjacent gabbro.

In addition to these relatively well defined rock types distinguishable within the gabbro, there are more diffuse variations in the relative proportions of dark minerals to feldspar, defining bands whose general trend is often roughly concordant to the igneous contacts. The banding is in part marked by variation in grain size. In the pluton south of the Lakeview Mountains two thick, tabular bodies of fine-grained gabbro have been mapped. The gabbro intruded south of Rawson Ranch is also banded by variations in dark mineral abundance, a feature which appears on aerial photographs as lighter streaks conformable with the intrusive contacts and dipping into the center of the ring dike.

Contacts between the gabbro and adjacent gneiss are generally subconcordant, but locally markedly discordant. Contacts between gabbro and other igneous rocks occur in D14 and D15 south of Florida Avenue and south of Rawson Ranch in A24 and B19. The former locality has been discussed above in relation to the Lakeview Mountain tonalite. South of Rawson Ranch the general form of the gabbro intrusive is a crescent-shaped plug whose arms extend south along the contact between granodiorite and gneiss (Plate 9d, appendix). A smaller plug of gabbro sits within the granodiorite body. Within the gabbro and along the southern margin of the gabbro adjacent to the granodiorite bodies of tonalite are found transitional into quartz-gabbro. These bodies are elongate parallel to the weak banding or foliation of the gabbro, which is in turn parallel to the foliation of the gneisses. The gabbro is not sheared appreciably except in discrete lenses near the contact with the gneiss. However the quartzite of the country rock is strongly sheared just north of the gabbro and it possesses a strong vertically plunging lineation, just like that of the granodiorite (although it is mildly foliated as well).

These facts can be interpreted as indicating the following sequence of events:

1) the granodiorite was intruded into the foliated gneisses and quartzites as a partly consolidated crystal mush; vertical stresses accompanying the intrusion sheared out both the granodiorite itself and the enclosing quartzites and gneisses;

2) after solidification of most of the granodiorite a semicircular dike of gabbro was intruded along the sheared contact between granodiorite and gneiss;

3) the gabbro assimilated acidic material along its contact with the granodiorite forming a concordant zone of unsheared tonalite, and also the gabbro displaced large fragments of granodiorite which it partly assimilated forming internal tonalite bodies.

The last conclusion is highly dubious since the interior tonalite bodies themselves contain abundant schlieren of dark rock and resemble Bonsall tonalite. This would rather tend to indicate a later intrusion of fluid magma into the partly solidified gabbro, incorporating inclusions of gabbro.

The age relations described above are in conflict with those proposed by Larsen and those observed in the quarry in the southwest corner of this area. It is therefore concluded that the age relations of these two rocks, the Woodson Mountain granodiorite and the San Marcos gabbro, are not definitely known here. It is not impossible that the intrusive body south of Rawson Ranch is not Woodson Mountain granodiorite but rather a hitherto undescribed earlier granodiorite. Its locally intensely deformed texture and the presence of significant amounts of muscovite and garnet in the rock distinguish it from the typical Woodson Mountain granodiorite as described by Larsen. This body may be related to the Stonewall granodiorite of the Julian district, 50 miles southeast of here in San Diego County (Everhart, 1951, p. 61) which is cut by the Cuyamaca gabbro, probably related to the San Marcos gabbro.

d. Pyroxenite and Hornblendite

A small plug of pyroxenite and several smaller dikes of hornblende-rock intrude gneiss and gabbro near the southeast corner of the Winchester quadrangle. They are presumably related to the gabbro.

The pyroxenite in A23 is a black coarse- to medium-grained rock with a silky luster due to scattered flakes of chlorite. It is poorly exposed and forms a dark red soil like that on the gabbro. It is composed of interlocking anhedral grains of hypersthene ($2V = 90^{o}$, pleochroism: z = greenish, x = pink) extensively replaced by bluish green to brown hornblende and colorless to pale green chlorite. A schiller texture is preserved in the hornblende grains. Some diopside may have been present in the primary rock since the hornblende is caliciferous.

The hornblendite is identical in texture to the pyroxenite but pyroxene is absent. Locally, as in the dikes of hornblendite cutting the gabbro in A24(4, 2) the rock is schistose and composed of 40 per cent prismatic blue to green actinolite, 55 per cent pale green chlorite and 5 per cent magnetite in patches enveloping the actinolite. There appear to be all gradations within a single dike from unsheared hornblendite to chlorite-actinolite schist. The metamorphism of the primary igneous parent (pyroxenite) into this schist has evidently taken place at lower P-T conditions than that of the enclosing gneiss since chlorite is absent from amphibolites of the country rock. This rock contains a mineral assemblage characteristic of the greenschist facies.

The larger pyroxenite body is discordantly emplaced into the schists with no evidence of deformation of the foliation near its boundaries. Its age relative to the other igneous rocks is unknown but the hornblendite dikes, presumably co-magmatic with it, cut the San Marcos gabbro. A single hornblendite dike is intruded along the contact between Woodson Mountain granodiorite and gneiss in section B19(2, 1). The trend of all of the hornblendite dikes is parallel to regional foliation.

e. Domenigoni Valley granodiorite

The eastern half of the Winchester-Hemet area is underlain by a single pluton of the Domenigoni Valley granodiorite (Larsen, 1948, p. 69). This rock is light gray in outcrop and contains abundant dark inclusions. It is intruded as a single discordant elliptical pluton about 7 miles long and 5 miles wide, with its long axis trending slightly east of north.

Within this rock there is marked variation in texture and some variation in modal composition. Three estimated modes are given in table 4.

Texturally the rock varies from medium-grained equigranular to fine-grained aplitic with medium- to coarse-grained phenocrysts of plagioclase. About 95 per cent of the exposed area of the pluton is underlain by the equigranular phase. Plagioclase is always zoned, varying from calcic oligoclase to albite. About half of the potash-feldspar is microcline microperthite in equant grains with marginal fingers extending away between other crystals. Biotite occurs in two distinct

TABLE 3. Estimated Modal Analyses of the Domenigoni Valley Granodiorite, Winchester-Hemet Area, California

	D19(4, 1)	D8(1, 1)	KDV
Quartz	45	20	37
K-feldspar	15	25	5
Plagioclase	30	40	45
Biotite	8	12	8
Muscovite	2	tr	tr
Chlorite	1	-	tr
Sphene	tr	tr	tr
Magnetite	tr	-	-
Apatite	-	tr	-

Location of samples:

D19(4, 1)	South side of Double Butte, D19(4, 1)
D8(1, 1)	Hill 1571 on section line, D8(1, 1)
KDV	Along Leon road, A5(2, 1)

habits: as equant medium-grained tablets which are dark, almost opaque olive brown in thin section; and as compact medium-grained clusters of lighter colored fine grains, apparently pseudomorphous after hornblende. The latter texture is characteristic of the porphyries. Hornblende, dark grass-green to blue-green, is in subhedral fine grains partially enveloped by biotite. Some of the hornblende appears to rim and replace the tablets of biotite.

On the south edge of the pluton the granodiorite-metasediment contact changes from marked discordance to concordance as one traces it clockwise around the pluton. The "corner" at which the discordance begins apparently represents the point of emergence of a hinge-line along which Bedford Canyon schists were bent as the Domenigoni Valley granodiorite was intruded along the foliation (see cross-section H-H', fig. 3). Locally in the schists near the corner a strong crinkle lineation is developed parallel to the "B" tectonic axis. On the north side of Searls Ridge at its west end the granodiorite cuts sharply across the foliation of the metasediments but elsewhere the discordance is one of less than 30 degrees. In the center of Double Butte the contact is again completely concordant in both dip and strike.

There is no conclusive evidence of the age of the Domenigoni Valley granodiorite relative to the other major igneous rocks. Dark inclusions, now converted to biotite-hornblende-plagioclase rock, are found throughout the intrusive and may be reworked San Marcos gabbro. However, two small dike-like bodies of hornblende gabbro (or amphibolite?) appear to intrude the Domenigoni Valley granodiorite a half mile east of the Winchester road in section A10. These may be relict inclusions. The texture of the gabbro is definitely modified by later metamorphism from an original igneous texture. The plagioclase is An_{35} contrasting with An_{85-95} of the San Marcos gabbro. Pyroxene is replaced pseudomorphously by grains of pale bluish-green hornblende densely crowded in their cores by oriented opaque inclusions. Matted fine-grained biotite and euhedral acicular actinolite form a matrix around the coarser amphibole crystals. The plagioclase crystals have been replaced along fractures by more sodic plagioclase with the same optical orientation. These features suggest reaction of a primary pyroxene-hornblende gabbro with an acidic rock. These features could be equally well explained by: reaction of the host granodiorite with a gabbro intruded into it while the granodiorite was partially crystallized and rich in volatiles; or alternatively, by intrusion of the granodiorite into a pre-existent gabbro mass and stripping of inclusions of gabbro. Other inclusions of mafic rock are rare in this particular locality. The gabbro body is aligned parallel to adjacent dikes of unquestionably post-granodiorite quartz latite. These two facts tend to support the first hypothesis. The gabbro bodies are sheared along their margins and altered to a chlorite-hornblende schist. This is comparable to the shearing of the adjacent quartz-latite dikes.

The general discordance of the Domenigoni Valley granodiorite with both regional foliation and trend of the metamorphic isograds contrasts markedly with the concordance of both the Lakeview Mountain tonalite and the Woodson Mountain granodiorite. This may indicate that the Domenigoni Valley granodiorite is younger. It generally appears to be a finer-grained rock than either of the above rocks and this suggests in turn that it was intruded at a shallower depth (after general crustal uplift) while the country rock was cooler. The discordance itself may be due to the relative rigidity of the schists at a lower P-T regimen in contrast with the plastic mode of yielding during intrusion of the tonalite. The possibility of plastic extrusion of the metasedimentary rock of Double Butte by the force of an intruding magma was discussed in Chapter II.

f. Quartz-latite Dikes

Fine grained dikes of slightly more alkaline composition than the Domenigoni Valley granodiorite intrude the latter rock and the adjacent schists. The dikes are 10 to 30 feet wide and single dikes can be traced for over a mile. They strike uniformly in a N30W direction which is essentially the trend of foliation in the nearby schists. They dip steeply northeast. Where they intrude schists they are in fact sills which persist along single stratigraphic horizons for their entire length. They weather to elongate ridges of dark gray rock spotted with white phenocrysts. These ridges stand up above the more deeply weathered Domenigoni Valley granodiorite and their locations have been mapped from aerial photographs except where they intrude schist.

The texture of this dike rock indicates that it is a metamorphic rock, a meta-latite. It is highly sheared and gneissic in all its exposures. The foliation parallels the dike walls. Medium-grained euhedral phenocrysts of oligoclase make up from 0 to 15 per cent of the rock, embedded in a very fine-grained equigranular groundmass of 35 per cent untwinned potash-feldspar, 40 per cent quartz, 20-25 per cent plagioclase and 5 per cent dark olive brown biotite. Part of the biotite and lesser amounts of muscovite may be clustered together into streaks about 1 mm. long and 0.05 mm. thick. All the fine mica flakes as well as these streaks are very well oriented into a plane parallel with the country-rock contact. The plagioclase phenocrysts are oriented at odd angles to this plane and distort the foliation in their vicinity. Medium-grained ellipsoidal quartz phenocrysts occur in some of the less strongly sheared rocks. Epidote occurs as fine grains replacing plagioclase and intimately intergrown with biotite. In the latter example the epidote may be a reaction product from the breakdown of hornblende:

$$\text{Hb} + \text{K-feldspar} \rightleftharpoons \text{Biotite} + \text{Zoisite} \qquad (12)$$

Hornblende is present as the principal dark mineral in some of the sills intruding the French Valley formation.

In addition to possessing a strong foliation the quartz latite dikes are lineated by streaks of biotite and plagioclase. This lineation consistently plunges 60°, S 60-70 E, in the plane of foliation. The foliation of the dikes dips 60 to 70 degrees northeast.

The chemical and mineralogical resemblance to and the spatial association with the Domenigoni Valley granodiorite suggests that this dike rock is derived from the same magma. Partial chemical analyses of the two rocks are given for comparison below:

	Domenigoni Valley granodiorite (Larsen, 1948, p. 70)	Quartz latite (Larsen, 1948, p. 105)
K_2O	1.86	2.90
Na_2O	3.46	3.50
SiO_2	69.78	71.74

The fact that the quartz latite differs from the granodiorite in being richer in SiO_2, K_2O and Na_2O is compatible with the suggestion that the latite magma is a later differentiate from the granodiorite magma.

The cataclastic texture of these dikes will be discussed in Chapter V (structure).

g. Pegmatites

Two types of pegmatite can be distinguished in the Winchester-Hemet quadrangle: concordant simple granitic pegmatites grading into aplite; and discordant complex pegmatites with associated graphic granite.

The simple, concordant pegmatites occur throughout the area increasing somewhat in abundance to the east. Locally pegmatites of this type are actually discordant, as in D36(3, 1) but here and in other such localities they are of the same texture and composition as the nearby concordant dikes.

The typical pegmatite of this area is composed of about equal amounts of quartz, perthitic potash-feldspar and sodic oligoclase. In addition it may contain from 2 to 10 per cent tourmaline (schorlite), 0 to 5 per cent pale pink garnet (spessartitic), 0 to 2 per cent muscovite, 0 to 10 per cent biotite and traces of epidote and magnetite. Muscovite is apparently a late mineral, forming in planar segregations cutting other phases. The texture of the rock varies from very coarse pegmatitic to very fine-grained aplitic. Within a single outcrop a few feet across all variations from very coarse to very fine-grain size may occur. The various textural types are commonly interlayered in bands parallel to the walls of the dike. They represent variations in the rate of crystallization of the dike material. "Line rock", alaskitic aplite marked by regularly spaced layers of fine-grained garnet, is present in about a third of the dikes. Schorl is commonly clumped in groups of three or four euhedral crystals in the midst of patches of coarse-grained rock. It is rare in the aplites.

The aplite dikes are always sheared to some extent. They are composed of garnetiferous alaskite granite. An estimated mode is given below: (volume per cent)

K-feldspar	20	
plagioclase	30	(An_{15})
quartz	40	
biotite	3	
muscovite	5	
garnet	3	
sillimanite, spinel	tr	

The garnets are a dark reddish brown. They may attain a size of 2 cm. across. They are sometimes sheared out into trains of granule-sized grains. The aplitic rock has a pronounced foliation defined by orientation of mica and squashing out of clusters of quartz. In some instances later milling (following crystallization from a magma) may have appreciably lowered the mean grain size of the rock. Aplites that appear to be profoundly affected by cataclassis are commonly interlayered with schist and gneiss that show no such effects.

Inclusions of schist from the country rock are often incorporated into the pegmatites. They are impregnated by pegmatite minerals and biotite is partially replaced by tourmaline.

Concordant layers of pegmatitic-looking rock are found in the highest-rank metamorphic rocks, grading along strike into quartz-feldspar rich bands of banded gneiss. These layers are of alaskitic composition but are probably formed by anatexis of the host gneiss at high temperatures and P_{H_2O}. They commonly swell out into cylindrical pods at the axes of minor folds. These pegmatites lack garnet or muscovite. They are discussed further in Chapter IV.

The discordant, complex pegmatites are found in two localities. One is in the northern pit of the Hemet magnesite mine. The other is in the "Bonsall" tonalite border zone north of Florida Ave. in D9(4, 1).

The mineralogy of these pegmatites is for the most part the same as the "simple" types. However they contain in addition trace amounts of rare minerals.

In the pegmatite of the magnesite mine, a single pod less than 30 feet long and 10 feet wide, there have been found andalusite (see Murdoch, 1933), topaz, apatite, beryl (?) and columbite, all in amounts less than 1 per cent of the rock by volume. Andalusite is concentrated near the walls of the dike. Its presence here is presumably due to alumina assimilated from the adjacent silimanite gneiss. Other smaller dikes in the vicinity are more typical of the concordant simple variety, but contain traces of andalusite and sillimanite in addition to the usual granitic minerals.

The pegmatites of the southern end of the Lakeview Mountains have been investigated by Mr. Douglas Morton of the University of California at Los Angeles, who has kindly supplied the writer with information about their mineralogy. These pegmatites are bulbous lenses whose maximum exposed dimension is about 200 feet, swelling to a maximum thickness of 30 feet. They dip north about 40 degrees and transect the weak vertical foliation of the tonalite in which they occur. They are zoned somewhat in the fashion of the pegmatites of Pala, San Diego county, California (Jahns, 1954). A border zone rims each pod, consisting of clevelandite albite and perthitic potash feldspar, both in graphic intergrowth with quartz. Biotite occurs in this zone as thin tablets elongate parallel to [010] and with [001] tangential to the dike wall. Quartz increases in abundance toward the core of each pod. The core is composed of massive coarse-grained albite-quartz graphic granite with minor schorlite, muscovite and garnet. The muscovite is pale green, in radiating clusters of coarse tablets, locally concentrated along fractures. Some of the pegmatite bodies have an inner core of pure massive bull quartz. In many of the pegmatites this core has been entirely quarried out.

Traces of the following minerals are found:

allanite
monazite
yttrotantalite
columbite
pyrochlore
microlite
cyrtolite
pucherite
chalcopyrite
chalcocite
sphalerite
bismuthinite and other bismuth minerals

Sulphides also occur in pods replacing the adjacent tonalite. One such pod is 10 inches across and 2 inches thick. The wall rock of the pegmatites is hydrothermally altered. The feldspar is partly converted to clay minerals and the mafic minerals are stained with limonite and chloritized.

3. Age of the Batholithic Rocks

The southern California batholith as a whole was apparently intruded in a very short time interval. Age determinations of various rocks of the batholith from diverse localities in southern California have been made by the Larsen lead-alpha method (Larsen, et al, 1958).

They yield a consistent age of 105 m. y. ± 5 m. y. This age places the time of intrusion in the upper middle Cretaceous on the Holmes time scale. This is in agreement with paleontologic data obtained in Baja California where plutonic rocks intrude and are overlain by fossiliferous sedimentary rocks (Silver et al, 1956). No such paleontologic evidence has been obtained in southern California.

IV. HYBRID ROCKS

A. INTRODUCTION

Three types of rock of hybrid metasedimentary and igneous origin are found in the Winchester-Hemet area. They are the migmatite zone of the eastern edges of the area, the migmatite-breccia body of section A24 and the carbonatite of the magnesite mine. The last has been discussed in Chapter III, and in this chapter we will briefly consider the other two.

B. THE MIGMATITE BRECCIA

Rock of this type occurs in a zone passing through the middle of section A24. The major lithology of the migmatite breccia is a dark gray, weakly schistose rock, weathering to massive, featureless outcrops. On a fresh surface the migmatite breccia is mottled and streaked with light gray rock cutting a black, fine-grained schist. The lighter veinlets have borders against the schist that are variously knife sharp to smoothly gradational. Dark patches of schist range in size from 1/2 inch to 4 inches in maximum dimension. They are subequant or slightly elongate parallel to a "b" lineation defined by the axes of associated ptygmatic folds.

The enclosed schist fragments are invariably composed of fine- to very fine-grained rock. The veinlets increase in grain size as they grow in width until the widest are actually medium-grained dikes of granodiorite or tonalite. The widest dikes are gradational, through a zone of increasing inclusion abundance, into the migmatite breccia proper. They have been mapped as Woodson Mountain granodiorite since they most resemble this rock type, but they may actually be local concentrations of material crystallized from a dilute magma of unknown origin, which was migmatizing the schists. These dikes have been sheared and the typical dike rock is texturally an augen-gneiss. An estimated mode of a specimen from a dike in A24(1, 3) is given below.

WH117

plagioclase	55
muscovite	2
biotite	5
quartz	35
K-feldspar	2
garnet	1
tourmaline, zircon, magnetite	tr

This rock is atypically potash-poor in comparison with most of the dikes and is actually a tonalite. It contains elliptical eyes of plagioclase 1-3 mm. across embedded in a medium-grained equigranular matrix.

The migmatite itself is a fine-grained equigranular schistose rock containing the following estimated mineral proportions:

biotite	25
K-feldspar	2
sillimanite	8
quartz	65
plagioclase, magnetite, apatite, zircon	tr

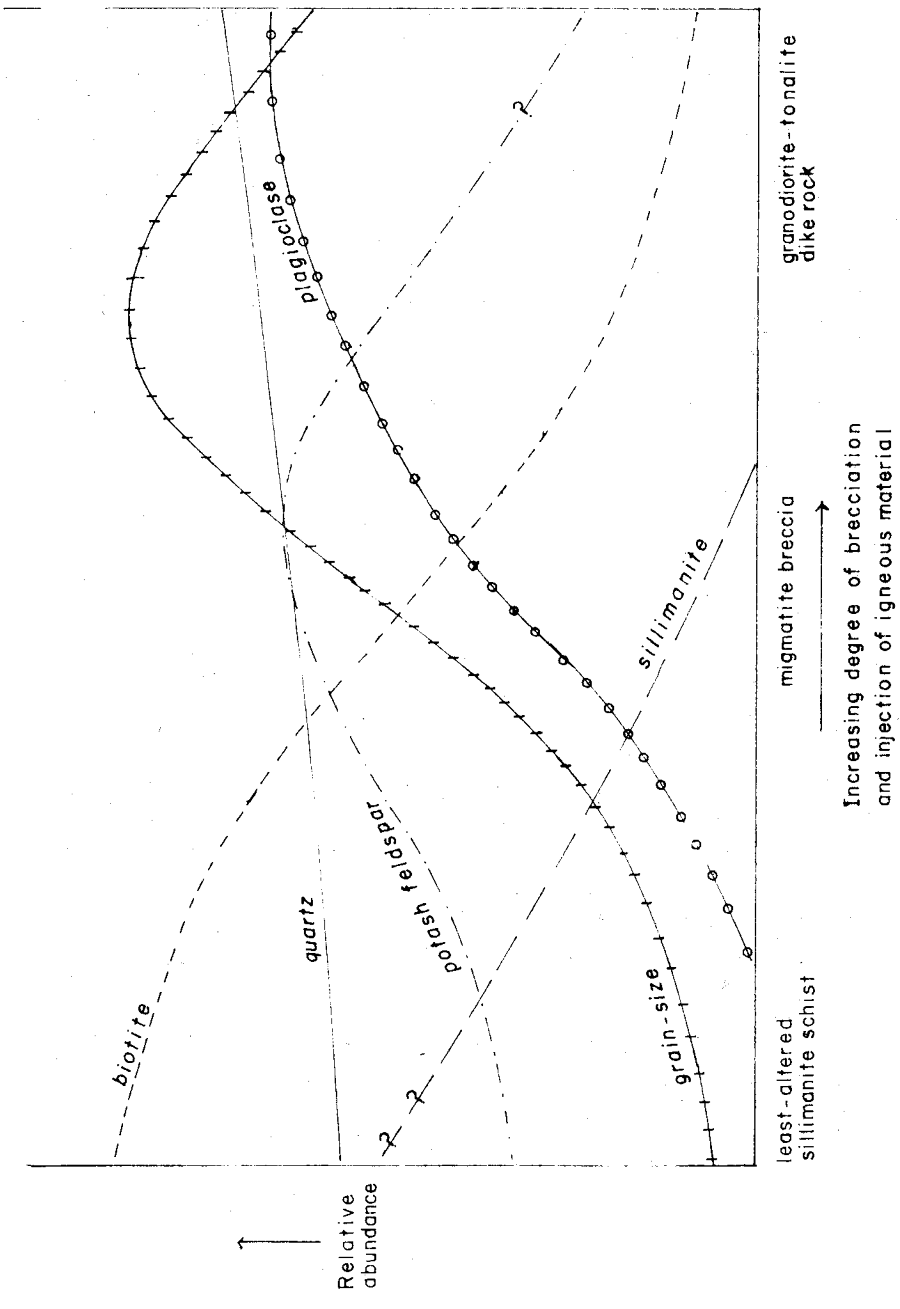
The schist portions of the rock are typically fine-grained biotite sillimanite schist. No relict compositional banding is seen in the schist. The veinitic portions are texturally almost indistinguishable from the schist portions in thin section and differ principally in a slightly higher content of potash-feldspar and lower content of mica and sillimanite. Where a foliation is present in the rock these veinlets cut across the foliation but have themselves been sheared parallel to this plane, resulting in ptygmatic folds. Mica enclosed in the veinlets possesses this same planar orientation.

Much of the migmatite possesses no well-defined schistosity but rather consists of blocks of fine-grained biotite-sillimanite-quartz granulite embedded in a swirling, folded and refolded network of veins of plagioclase, quartz, potash-feldspar and biotite. This rock type is extremely inhomogeneous, even over an area of a few square centimeters. In general the trend of mineral abundance variation from specimen to specimen is from a biotite-quartz-sillimanite-potash-feldspar assemblage to a quartz-plagioclase-potash-feldspar-biotite assemblage. This can be seen illustrated schematically in fig. 13.

The nature of this intermixing of igneous and metasedimentary material is best seen in a photograph of a typical specimen (Plate 11b, appendix). Note especially the lack of orientation of the clasts and the lack of schistose layering.

Contacts of this unit with the adjacent metasediments are gradational where exposed. The most critical contact, however, is almost completely buried. This is the contact separating the migmatite breccia from the quartzites of the French Valley formation, Member A. In places this contact appears to be a fault trending northeast, presumably with a dominant dip-slip component. In section A24(3, 4) the contact is found exposed in a draw on the edge of an open, soil-covered field. Here it is simply an abrupt break in lithology over a distance of a few inches, cutting sharply across relict bedding in the quartzites northwest of the contact. The migmatite southeast of the contact is highly sheared and possesses a steeply plunging "b" lineation. The trend of foliation changes abruptly across the contact (cf. fig. 1).

FIG.13 Schematic illustration of progressive development of the migmatite breccia.



An amphibolitic quartzite occurs wholly included within the migmatite breccia zone, terminating against granodiorite on the southeast and against un-migmatized schist (faulted?) on the northwest. This rock is not migmatized itself and is in sharp contact with the migmatite all along its strike.

The amphibolitic quartzite varies from a quartzite containing only traces of hornblende and plagioclase to a garnetiferous amphibolite with 10 per cent or less quartz. A mode of a specimen from this unit is given below: (volume per cent)

WH106	plagioclase	53.8	(An_{65})
	quartz	12.4	
	hornblende	15.8	
	pyroxene	1.4	diopsidic augite
	garnet	11.1	
	epidote	4.7	
	sphene	1.0	(1027 points)

The garnet in this rock is pale pink in hand specimen, euhedral and coarse-grained. Its refractive index is 1.80. Hornblende, epidote and pyroxene are in fine-grained euhedra. The quartz is in anhedral equant blebs disseminated through the plagioclase. The ratio of plagioclase to total femic minerals (hb + px + gnt) is rather high compared with typical amphibolites of the French Valley formation, which average about 60 per cent femic minerals by volume. Uncertainty in the composition of some of the component minerals prevents us from estimating the chemical composition. The mineralogic composition of this unit is compatible with its derivation from a shaly dolomitic sandstone. No relict sedimentary structures are apparent. The local concentrations of amphibolitic material in this unit perhaps represent relict sedimentary concentrations of carbonate. An alternate hypothesis is that this may be derived from a tuffaceous quartzite in which the tuffaceous component was andesitic in composition. The high CaO content inferred to be present in this rock from the plagioclase composition (An_{65}) and the epidote content would rather support the hypothesis of a marly parent rock.

Origin of the migmatite breccia

The evidence in respect to the origin of this unit is summarized below:

1) The sedimentary component of the breccia is a biotite sillimanite schist resembling some thin units found in the adjacent French Valley formation but more typical of the thick section of phyllite in the underlying Bedford Canyon formation.

2) The injected material is of a granodioritic to tonalitic composition but deficient in mafic components (alaskitic).

3) The contacts with adjacent rock units are abrupt with the exception of the above mentioned diffusive contacts with schist units of the French Valley formation.

4) Except for the amphibolitic quartzite layer the migmatite zone is structurally chaotic; attitudes in the schistose parts of this zone are highly variable on a scale of tens of feet.

5) The texture of the rock shows features of both fracture and plastic flow, apparently occurring in that sequence (since no ptygmatically folded veins are themselves fractured).

6) The inferred metamorphic rank of this unit is discordant with that of the adjacent schists. The latter are andalusite-biotite

schists at their highest rank and more commonly biotite-muscovite schists. The pelitic component of the migmatite-breccia contains a biotite-sillimanite-potash-feldspar assemblage. Elsewhere in the Winchester-Hemet area biotite muscovite assemblages are separated from biotite-sillimanite assemblages by a zone of andalusite schists.

7.) The whole zone of migmatite is spatially closely related to a plexus of intrusive plutonic rock (gabbro and granodiorite).

A sequence of events consistent with the above evidence is as follows:

i) the folded and metamorphosed metasediments of the Bedford Canyon formation, underlying the French Valley formation at this locality, were intruded by igneous rock (tonalite ?) of the southern California batholith;

ii) late in the cooling period of the tonalite intrusive, as P_{H_2O} rose due to crystallization of anhydrous minerals from a hydrous magma, the vapor pressure locally exceeded the confining pressure and boiling took place, fracturing the weaker schists but leaving the more rigid quartzites intact;

iii) this fractured material, impregnated with alaskitic material precipitated from the late water-rich magma, was forcibly intruded into its present site as a plastic mass; partial fusion (anatexis) of the pelitic rocks themselves permitted the breccia fragments to be squeezed and smeared out by viscous drag as the body of breccia moved into place.

The metamorphic rank discordance may be attributed to local heating at depth by the tonalite body. The infiltrating zones of migmatite that extend into schists in the neighboring French Valley formation represent residual alaskitic magma carried up with the breccia intrusive, escaping out into the country rock. One objection to this theory that is immediately apparent is the undisturbed structure of the amphibolitic quartzite, which continues to strike parallel to the regional foliation and is not appreciably fractured, in spite of the chaotic structure of the rocks surrounding it. It is possible that this was a competent unit present higher in the section, a screen immersed in the semi-fluid migmatic breccia. The last stages of injection of the migmatite may have been simply vertical uplift of the whole elliptical mass, bringing the amphibolitic quartzite along with the migmatite to a position where its unmetamorphosed equivalent cannot be seen. This uplift may have been contemporaneous with the late-stage vertical shearing of the Woodson Mountain granodiorite described in the last chapter.

C. MIGMATITE GNEISS

Outcrops of pelitic metasedimentary rocks occurring progressively eastward in the Winchester-Hemet area display a progressive change in texture concurrent with the progressive changes in mineralogy marking rise in metamorphic rank. The originally fine-grained phyllites and interlayered thin quartzites become first more coarsely crystalline and then banded with alternating layers of mica schist and siliceous granulite. The culmination of this progressive change is exemplified by coarse-grained biotite-sillimanite-potash-feldspar-oligoclase-quartz gneisses cut by sub-concordant veins of pegmatitic-textured material of an alaskitic composition. This rock is best described by the term migmatite. A migmatite, as strictly defined, is a schist or gneiss which has been pervasively injected by a granitic magma (Sederholm, 1907). It is texturally defined to be a mixed rock displaying metamorphic textures cut by or concordantly injected in a lit-par-lit fashion by rock with an igneous texture. This in itself may or may not be indication of injection of an externally derived magma. Alternatively the igneous-looking layers may be material taken out of the bulk of the rock by partial fusion or solution in the presence of water at high temperature and pressure.

The rocks of this textural type are distributed in a north-south trending belt adjacent to the tonalite intrusive. This belt swings west near the south edge of the area to conform roughly with the distribution of tonalite and granodiorite in that area. The contact between migmatite and gneiss or schist is drawn somewhat arbitrarily, and roughly bounds an area underlain by rock with over 1/3 veinitic igneous-looking material.

The mineralogic composition of the migmatite is dependent to some extent on the mineralogy of the parent metamorphic rock. In comparison with the primary finely layered schists, however, the typical migmatite gneiss is a very homogeneous rock in general. Samples of rock the size of a thin section show little variation from place to place. Modal analyses are given below in reference to the origin of the migmatite.

The texture of the least-altered part of the migmatite is essentially the same as that of the high-rank gneisses. The bulk of the rock is made up of equant medium-grained quartz grains. Potash-feldspar is in anhedral equant grains containing round fine-grained quartz blebs.

Plagioclase is in equant subhedral grains commonly showing albite twinning. In some thin sections the plagioclase forms rounded elongate grains with limbs of complex shape which follow grain boundaries between other minerals. Biotite and muscovite occur as tablets which are anhedral in the zone [001]. The muscovite contains inclusions of potash-feldspar and appears to be replacing it in part. Biotite contains inclusions of graphite and sillimanite, the latter as a fine fuzzy border around discrete grains of biotite. Sillimanite also occurs as dense contorted medium-grained knots of fibrolite. Myrmekite forms characteristic replacement blebs along the boundaries of quartz and plagioclase. There is apparently no gradation between the albite of the myrmekite and the calcic to sodic oligoclase of the host rock, the two having distinct textural patterns.

As a whole the texture of the rock is granoblastic to gneissose, medium-grained and equigranular. Adjacent to the granitic veins there are commonly zones of biotite-rich rock, but in general the biotite is uniformly disseminated through the rock or vaguely concentrated in bands. Muscovite and biotite have a common poorly defined planar orientation.

The granitic veins themselves range in thickness from 1 mm. to 5 cm. The first such veins to appear are coarse-grained equigranular segregations within the quartz-feldspar bands of the banded gneiss, grading along strike in texture and composition into typical feldspathic quartzite. What is referred to here as increasing degree of migmatization is a combination of textural changes within these lighter bands:

increase in inhomogeneity in the distribution of feldspar; increase in grain size in the feldspathic (granitic) clots; change in grain boundary type in the feldspathic clots from polygonal to irregular sutured contacts; swelling of the feldspathic clots producing distortion of the adjacent micaceous bands; and finally complete disruption of the dark bands by cross-cutting feldspathic accumulations, and envelopment of biotitic scraps by the granitic material. The highest rank migmatites are ptygmatically deformed into folds spaced a few inches apart and granitic material is concentrated at the axes of these folds.

A critical test of whether the vein material is truly exotic granitic material (migmatites in the sense of Sederholm) or whether it is rather partially mobilized gneiss would be to compare the composition of the presumed low-rank unmigmatized equivalents of this rock with that of the migmatite itself. There are two obstacles to making such a comparison: 1) the parent schists are so inhomogeneous laterally and vertically in section that it is impossible to make a sufficiently accurate estimate of their over-all composition to test the hypotheses; 2) in addition to textural changes there are mineralogical changes from unaltered schist to migmatite which cannot be separated from mineralogical changes due to changes in bulk composition; for example, biotite breaks down in the high-rank gneisses into sillimanite, potash-feldspar and ore and it would be difficult to distinguish small amounts of potash-feldspar added in this way from potash-feldspar introduced by an exotic source of alkali-rich solutions. The latter obstacle could be overcome by comparing chemical rather than mineralogical analyses. The obstacle of primary inhomogeneity is more difficult to overcome. In the second part of this thesis a study is made of a relatively homogeneous unit subjected to progressive metamorphism, but this unit has not been carried into the zone of migmatization due to structural complications. Therefore a less detailed attempt has been made here to distinguish between the two hypotheses as to the nature of migmatization in this area.

For this purpose a series of modal analyses of typical biotite-sillimanite gneisses and migmatites has been made. Two groups of analyses are presented: one of a group of specimens of least-altered gneisses from areas of highest-rank mineralogy but lacking abundant vein material; and another of migmatites. The modal analyses of the migmatites were made on specimens selected so as to eliminate as well as possible obvious granitic veins from the counted field. It should be noted that these rocks are by no means considered to represent a single original stratigraphic unit. They are chosen to represent rocks of pelitic derivation whose initial composition is inferred to have been as similar, sample for sample, as possible. The limitations on such a selection process are obvious. In addition to the modal analyses, two analyses of vein material from the migmatite zone are compared. These were made by mounting crushed, sized material in an appropriate immersion medium ($n = 1.540$) and counting grains. These abundance values are not considered to be reliable to better than ± 3 per cent. The analyses are given in table 4.

Considering the modal analyses, there is no noticeable systematic difference in the abundance of the two major minerals, quartz and biotite between the gneiss and migmatite. What differences that occur may be accounted for by the fortuitous choice of specimens for analysis, since

TABLE 4. Modal Analysis of Biotite-sillimanite Gneisses and Migmatite Gneisses and Grain Counts of Crushed Vein Material

	Biotite-sillimanite Gneiss			Migmatite			Veins	
	262	285	196	200	361	360	348	361
Quartz	54.4	72.0	43.5	52.5	63.1	81.4	43	61
Biotite	19.9	16.8	22.7	29.7	22.2	5.0	20	15
K-feldspar	8.2	6.3	10.43	1.3	0.0	0.8	15	6
Plagioclase	3.3	0.3	4.6	4.5	2.8	9.2	11	14
Sillimanite	5.8	3.1	3.4	9.6	9.1	0.5	4	2
Myrmekite	0.1	0.0	1.0	1.4	0.2	0.0	0.4	2
Muscovite	5.2	1.1	9.3	0.4	2.6	1.5	5	0
Opacite	2.7	0.4	0.7	0.5	0.0	0.6	0.5	3
Tourmaline	0.3	0.1	0.0	0.0	0.0	0.0	--	--
Montmorillonite	0.1	--	--	0.1	--	--	1	--
Garnet	--	--	0.9	0.0	--	0.9	--	--

obvious variations in the proportions of these two minerals can be seen in hand specimen or outcrop, representing relict banding.

The abundance of plagioclase is also variable in an unsystematic fashion. There is a suggestion on the basis of these analyses and estimated analyses of other specimens that the abundance of plagioclase increases with degree of migmatization. The modal abundances of plagioclase are always somewhat in question due to the possibility of confusion of untwinned oligoclase with quartz. The plagioclase of the migmatites tends to have an interstitial relation to the other phases, suggesting that it has been in part introduced along grain boundaries. It does not obviously replace the adjacent phases, however.

Potash feldspar, untwinned and non-perthitic, occurs in major amounts in the high-rank schists and gneisses. In the migmatites this mineral is considerably less abundant, while the vein material contains amounts (relative to the other components) comparable with those in the schists. Similarly muscovite is depleted in the migmatites relative to the schists, while it is relatively abundant in the veins.

The analyses of the vein material are too crude to permit any quantitative deductions except to say that the vein material is enriched, relative to the host gneiss, in plagioclase, potash-feldspar and muscovite. The apparently high abundance of biotite in the veinlets is largely an artificial effect due to the contamination of the samples with biotite-rich material concentrated at the walls of the veinlets. In fact it is evident in hand specimen that the veinlets contain less than 5 per cent biotite. The biotite of the veins is yellowish green in contrast with the reddish brown biotite of the gneiss.

Taken together, these facts suggest that the gneisses were subjected to a process of differentiation, which separated at least SiO_2, K_2O and Al_2O_3, redepositing them in the veinlets together with plagioclase as an assemblage of quartz-monzonitic composition. That the Na_2O (as albite) of the veins was likewise derived from the enclosing gneiss is doubtful. The bulk composition of the vein material and gneiss taken together would certainly be enriched in soda relative to the unveined gneiss.

Close to the contact with the tonalite the character of the migmatite changes. The distinction between veinitic material and host rock vanishes and the whole rock becomes a granoblastic-textured gneiss with poorly defined planar accumulations of biotite. A mode of a specimen obtained a few feet away from the contact with the tonalite is given below: (volume per cent)

WH354	plagioclase	26.4	(An_{30})
	quartz	62.4	
	biotite	10.4	
	K-feldspar	0.3	
	graphite	0.5	587 points

The veinlets cutting this rock are essentially composed of a leucotonalite. They contain only traces of potash-feldspar. Evidently the bulk composition of gneiss plus veinitic material closely approaches a tonalite at the very margin of the pluton. However the contact with the tonalite is distinct and knife sharp. The tonalite itself, although gneissic, is not banded and contains a more calcic plagioclase, a brownish green biotite (in contrast with the reddish brown biotite of the gneiss) and a yellow brown hornblende.

Origin of the migmatite zone

Returning to the question of the origin of the migmatites, it is evident from the above description that as far as these observations indicate, both hypotheses of origin are partially valid. That is, the migmatites show evidence of both secretionary (veinitic) and injection (arteritic) processes.

The first process, that of veinitic segregation or differentiation of material out of the bulk of the gneiss and into subconcordant veins, was widespread and affected a broad zone of gneiss roughly concordant with the gneiss-tonalite contact. This process consisted of partial melting of the lowest-melting components of the gneiss (potash-feldspar, quartz, oligoclase and muscovite) and auto-injection of this magma along planes of weakness in the rock simultaneous with folding and plastic flow of the rock. The magma probably did not flow over distances of more than a few feet.

This period of veinitic migmatization may have preceded intrusion of the tonalite magma by a considerable length of time. However, since it represents a period of heating and consequent melting and since the zone of migmatization is parallel to the contact with the tonalite it is equally plausible that the process of differentiation of the veinitic migmatites occurred concurrent with intrusion of tonalite and was supplied with heat from the tonalite magma. Apparently some Na_2O was introduced at this time, perhaps from the tonalite magma.

The process of arteritic migmatization took place in a much more restricted zone, a belt not more than 100 yards wide adjacent to the tonalite contact. A modal analysis of a typical specimen from this zone is given above (WH 354). This rock type was apparently formed by a proccess of pervasive injection of material of leuco-tonalitic composition, rich in Na_2O, SiO_2 and Al_2O_3. This material impregnated the gneiss and formed vaguely defined veins containing the equivalent of a leucotonalite. The material which was thus injected was derived from the tonalite itself and may have been the last crystallizing water-rich fraction of the magma, expelled from the magma chamber by increase in pressure as the magma cooled (Kennedy, 1955b).

Further detailed chemical and modal analyses could be made to test whether:

a) the bulk composition of the veinitic migmatite is constant, taking host rock and vein material together;

b) the bulk composition of the arteritic migmatite could be accounted for by mixing various proportions of a single homogeneous veinitic migmatite with a single leucotonalite magma.

Comparisons with migmatites of other areas

Comparing and contrasting the migmatites of the Winchester-Hemet area with those described in other localities throughout the world, we note two principal distinctions. First, in contrast with the migmatites of Scotland (Cheng, 1944) or Norway (Sederholm, 1923) there is no evidence of large-scale widespread diffusion or percolation of potash into the rock. Rather, potash appears to be taken out of the rocks and redeposited in the veins. The orthoclase porphyroblasts characteristic of high-rank migmatites of these areas are lacking here.

Secondly, the vein material in these classic examples of migmatites is observed to grade directly into massive gneissose igneous rock (granite) of about the same composition. Here, on the other hand, there is a sharp distinction between the typical vein material and the major intrusive igneous rock (tonalite). It may be argued that the zone near the contact with the tonalite is such a transition between igneous rock and migmatite, but here too we have drawn a sharp contact between tonalite and leucotonalitic migmatite.

These distinctions are due largely to the nature of the associated igneous rock. In contrast with the classical Fennoscandian and Canadian granite-migmatite association the intrusive rocks here are intermediate in composition (tonalite and granodiorite). They are certainly low in potash since potash-feldspar is but a minor constituent of most of the tonalite (although rocks of granodioritic composition make up an indeterminate and perhaps significant fraction of the Lakeview Mountain "tonalite"). This fact suggests that metasomatism associated with this tonalite intrusion might be principally Na_2O metasomatism in contrast with the K_2O metasomatism of the typical granitic aureole. The limited spatial extent of arteritic migmatites in this area may be due to the comparative "dryness" of this intrusive (as compared with typical granites). However the more extensive belt of veinitic-appearing migmatites described above would itself require a high water-vapor pressure to permit the partial melting phenomenon envisioned. No completely satisfactory explanation can be provided here to explain the striking differences noted above between the migmatites of the Winchester-Hemet area and those of the "classic" localities.

V. STRUCTURE

A. INTRODUCTION: THE STRUCTURAL FRAMEWORK OF THE WINCHESTER-HEMET AREA

The Winchester-Hemet area lies near the center of the Perris block, a physiographic and structural block bounded on the northeast by the San Jacinto fault and on the southwest by the Elsinore fault. The general features of this block have been discussed by Dudley (1935, 1936). This structural unit has not undergone appreciable deformation since the Cretaceous, except for uniform uplift of the whole block. On its northeast and southwest sides it is bounded by northwest trending mountains. These mountains were apparently formed by subsidiary dip-slip movement on predominantly strike-slip faults. These faults have presumably arisen in the later part of the Tertiary. The San Jacinto block in particular, less than 5 miles northwest of the area, has undergone uplift relative to the Perris block since mid-Tertiary times (Fraser, 1931, p. 537) and movement continues at present on the San Jacinto fault, predominantly in a strike-slip sense.

Within this block there is a general structural grain defined by the northwest trend of folding in the metamorphic rocks. Superimposed on this linear trend is a random distribution of igneous plutons, few of which exhibit a strong linear form, but all of which aggregate to define a broad northwest-trending zone of intrusive rock, part of the southern California batholith. Elongate remnants of the metamorphic rocks persist as screens and roof-pendants within the igneous masses, also defining a northwest grain.

The Winchester-Hemet area, underlain by one of the largest of these metamorphic remnants, exhibits, perhaps in the clearest fashion to be found in the Perris block, the detailed style of deformation that preceded the intrusion of the batholithic rocks. In part the development of the structures discussed here may be correlative with and caused by intrusion, but to a large extent the intrusives themselves have obliterated evidence in regard to this question.

B. FOLDING

From the evidence obtained in this study it appears that the Winchester-Hemet area at one time lay on the southwest margin of a system of steeply-plunging isoclinal folds which became progressively more closely appressed to the northeast. The core of this system of folds is now obliterated by successive intrusions of plutonic igneous rock.

Approaching the area from the southwest one encounters, in the vicinity of French Valley, progressively steepening dips in the metasediments until in section B7 the dips reverse around the nose of the first of the series of folds, the Domenigoni Valley syncline. This structure is well exposed only in section B7. Northwest of here, on Searls Ridge the folding is expressed in banded and homogeneous gneiss and schist in which the overall pattern of synclinal folding has been obscured by a close spaced series of tight subsidiary folds. The homogeneous and undistinctive nature of the schist units in this locality (D36(2, 1), (2, 2))prevents one from following single well-defined marker-units through the series of folds. On fig. 1 dashed lines have been drawn here conformable to the apparent changes in attitude of the foliation, representing the behaviour of a hypothetical marker unit in this locality. These lines are based in part on the outcrop pattern of locally distinctive lithologic zones. The location of the extension of the main synclinal axis across Searls Ridge is very uncertain. Southeast of section B7 the main syncline is obscured by faulting and subsequent drag folding along the Rawson Ranch fault zone. The extension of the Domenigoni Valley syncline across this fault zone can be seen in the structural nose defined by Member B of the French Valley formation as exposed in section B20.

Continuing to the northwest the sediments wrap around an anticline and syncline exposed on Searls Ridge in sections A36 and B31. The plunge of each successive fold in such a northeast traverse increases until the less well-defined folds of the Hemet quadrangle appear to be plunging almost vertically. The dip on the flanks of each successive fold in such a traverse also increases and northeast of the Rawson fault zone and the syncline of the magnesite mine all folding is essentially isoclinal with vertical dips on the flanks. These steeply plunging isoclinal folds are flanked by parasitic drag folds with similar geometry. The folding shown in the Hemet quadrangle is highly generalized due to the difficulty in correlating lithologic units through the areas of migmatization and poor exposure.

Folding due to drag along the Rawson Ranch fault zone is so great compared with the relatively small fault displacement along this zone that it deserves special mention here. This structural feature would be better described as a zone of cross-folding cutting the primary trend of northwest folding, and breached in its last stages by a series of strike-slip faults. The area of Crown Valley in particular exhibits structural features suggesting that the displacement of the axis of the Domenigoni Valley syncline was largely flexural. For example, note the arm-like extension of schist branching south from the lower contact of Member C of the French Valley formation in B17(1, 4). This is apparently a sliver of schist squeezed out of the nose of the syncline by shear stress acting in a right-lateral sense. Poor exposures, especially in the cultivated part of Crown Valley, make the detailed structure of this critical area uncertain.

The folded structures in Members A and B of the French Valley formation north of the pluton of Woodson Mountain granodiorite are excellent examples of folding induced by the force of an intruding magma. Although the detailed structure of these features is not known their general trend oblique to the principal regional trend of folding is well displayed. Apparently the granodiorite entering the metasedimentary rocks from the southeast wedged its way into them, tearing off a strip which was subsequently compressed into a series of folds with E-W trending axial planes. The tonalite intrusive adjacent to the granodiorite on the east no doubt also contributed to this fold pattern.

The isoclinal folding of the eastern part of the area and to a lesser extent the more open folding to the west have caused a structural thinning of the sedimentary units to an unknown extent. Individual units such as the meta-arkose of Member B can also be seen to swell to abnormal thickness at the noses of major folds. Assuming that the arkose of Member B was initially of uniform thickness throughout this area there would have been a structural thinning of this unit to about 1/10 its initial thickness in the eastern part of the area. This is undoubtedly an exaggerated estimate, as part of the thinning is probably due to facies change, but certainly some extrusion of the sedimentary units has occurred. It would seem that the schist units would be subject to such extrusion far more readily than the quartzite units, although there is no way of testing this presumption conclusively.

Small folds down to a few inches in amplitude occur in parallel orientation to the major fold-systems, along their flanks. These should be called drag folds, since they are parasitic to the main folds. It is a matter of conjecture whether they are truly due to slippage between strata along the flanks of the main folds or are simply compressive features of quite the same origin as the major folds. Larger examples of these parasitic folds have similar folds of even smaller amplitude on their flanks. On Searls Ridge in section D36, near the inferred location of the axis of the Domenigoni Valley syncline, the banded gneiss has been contorted into a maze of such small-amplitude folds, so that the major fold-axis cannot be distinguished. A strong axial-plane foliation occurs in the schist layers of these banded gneiss, transecting the bedding (?) plane foliation.

Shear folds due to pure slippage along foliation planes without flexure occur in the schists on the northeast flank of the Domenigoni Valley syncline in D36(4, 3). The amplitude of these folds is usually only a few inches, and they die out quickly along their axial plane. The planes of slippage stand at steep angles to the relict bedding. These shear folds are made visible in this locality by virtue of the close-spaced compositional banding in the schists. In somewhat higher-rank rocks further east along Searls Ridge folding of this sort grades into ptygmatic folding in which the planes of slippage conform more closely to original compositional layering. Only transecting quartz veins in this rock display the folding, as a result. Finally, in the highest grade migmatites, this style of folding grades into fluidal ptygmatic folding of all layered structures in the rock.

C. FAULTING

Two zones of faulting can be distinguished in the Winchester-Hemet area: a zone of E-W trending faults east of French Valley and the Rawson Ranch fault zone extending south from Diamond Valley through the center of the main block of metamorphic rocks.

The faults east of French Valley form a zone of parallel close-spaced breaks. It is not known whether the displacements on these faults were predominantly dip-slip or strike-slip. Displacements are on the order of a few hundred feet. The largest displacement measured is on the southernmost of this series of faults, in section A25(1, 4), a heave of 1270 feet. No drag is found in the schists adjacent to any of these faults.

The common orientation of these faults suggests a common origin. However their sense of displacement is not always the same. They may have been produced by a stress system set up by the forcible intrusion of the nearby Domenigoni Valley granodiorite, although none of the faults actually extends to the igneous contact.

The Rawson Ranch fault zone is a group of faults with predominantly strike slip and lesser dip-slip components of movement. Where it cuts the nose of the Domenigoni Valley syncline the southeast side of the fault-zone has apparently been dropped down, resulting in a widening of the exposure of Member C on the southeast side of the fault. South of Rawson Ranch displacement appears to be largely strike slip, as shown by right-lateral displacements of the vertically dipping beds of subunit fvB4s'. This is in agreement with the sense of displacement due to drag on either side of the fault zone. The amount of displacement across the fault zone at any given point appears to vary markedly along its length, suggesting that some plastic deformation of units normal to the fault trace has taken place either concurrent with faulting or subsequently. Individual faults within the zone have markedly different amounts of displacement associated with them, as well as differing degrees of dip-slip component relative to strike slip. As a whole the major displacement of the zone may be characterized as a zone of right-lateral strike-slip movement. The actual amount of displacement due to faulting is small compared with displacement due to drag (or cross-folding) along the same zone. Up to 2000 feet of heave may occur by faulting.

The age of faulting in this zone is indicated by the fact that the fault zone does not appear to cut the Woodson Mountain granodiorite south of Rawson Ranch. Faulting is apparently later than the period of folding resulting in the Domenigoni Valley syncline and associated folds. At least one of the faults of this zone may have undergone recent movement, however. The easternmost branch, extending from just south of Diamond Valley to the southern edge of the Winchester quadrangle (and perhaps further) is marked by a series of topographic breaks, springs and tree-lines. Such features are lacking along the other branches of the fault zone. Just across the stream from Rawson Ranch this branch is marked by a break in slope, an old fault scarp about 15 feet high. In this vicinity the stream has apparently been rejuvenated by local uplift so that it is cutting down into older alluvium which fills this part of Rawson Canyon. On the topographic map the break can be seen as a straight section of the 2020 foot contour along the east side of Rawson Canyon south of Rawson Ranch (section B18(4, 1) and B19(4, 4)). Tree lines along this fault permit it to be traced on aerial photographs but in mapping metamorphic structures across it south of Rawson Ranch there appear to be less than 50 feet of strike slip displacement along it and in some places none at all. The extension of this fault north of Crown Valley is based on scanty evidence since critical stratigraphic markers are absent along Rawson grade (section B8). Displacement along the fault in this area is largely strike-slip with possibly an appreciable dip-slip component. Where it crosses the axis of the Domenigoni Valley syncline it has an offset in a right-lateral sense of approximately 1300 feet. Evidently displacements along this fault branch die out rapidly to the south. North of Crown Valley the displacement is less. No ground-water blockage or topographic breaks are associated with this fault along Rawson Grade.

The Rawson Ranch fault zone appears to splay out at its northern end. It does not extend across Diamond Valley, although a fault of about 200 feet displacement trending NE cuts marker units in one of the "inselbergs" in the middle of the valley (C32(2, 1)) and is apparently an extension of one of the Rawson Ranch faults.

The linear trend of Diamond-Domenigoni Valleys, cutting as it does across the regional pattern of foliation and folding, suggests that a fault in bedrock may have controlled the original pattern of erosion, as is so commonly true of linear valleys in southern California. In an effort to test this hypothesis a dip-needle survey of the valley was made. Two traverses were run, one across the narrowest part of the valley in A1(4, 3) and the other through A2(4, 2), (4, 3) and (4, 4). No anomalies were observed. It should be added that lithologic correlation across the valley would limit the possible magnitude of strike slip displacement along such a buried fault to less than 2000 feet (left lateral). This is based on estimates of the maximum possible amount of mismatch of units if extended along strike to meet in the center of the valley. More probably the trend of the valley is derived from some antecedent drainage pattern long since obliterated by wider-scale erosion of the old Perris surface (Dudley, 1936, p. 375).

D. STRUCTURAL PETROLOGY OF THE METAMORPHIC ROCKS

1. Foliation

The foliation of the schistose rocks of both the Bedford Canyon formation and French Valley formation is almost invariably coincident with relict sedimentary layering. Exceptions to this rule are restricted to the crests and troughs of folds where, especially in banded gneiss, a second foliation develops parallel to the axial plane. In thin section such folded rocks have two clearly defined planes of preferred mica orientation (Plate 12a, appendix) and in outcrop their intersection is marked by a step-like "b" lineation.

With increasing degree of metamorphism the foliation becomes progressively less well-defined. This is reflected in the microscopic textures of the schists and gneisses by an increasing randomness in the overall orientation of mica, as seen in thin sections cut perpendicular to the foliation.

In the lowest rank phyllites foliation is defined by a uniform planar orientation of mica and disk-shaped fine-grained aggregates of graphite. In addition the foliation is marked by a parting along planes across which mineralogic composition changes abruptly. Both muscovite and biotite are oriented in this plane but in some specimens it is observed that somewhat coarser muscovite porphyroblasts have grown in random orientation.

In the andalusite schists, as andalusite grows in volume the well-defined schistosity of the rock is progressively distorted. This distortion involves twisting and bending aside of streams of fine-grained mica. Within these streams the mica still possesses a well-defined mutual orientation. Such textures are probably due to orientation of growth directions under stress and persist well into the sillimanite zone in spite of the complete recrystallization of the component minerals (Plate 5a, appendix).

In the retrograde muscovite zone described in Chapter II, parallel to and west of the Rawson Ranch fault zone, coarse unoriented books of muscovite develop in the schists. The lack of orientation and the subequant form of the muscovite crystals distinguishes them from the well-oriented muscovite which persists intergrown with biotite in the matrix of the rock. These coarse poikiloblasts evidently were not subject to the anisotropic stress conditions which produced the orientation of the muscovite in the matrix of the same rock. This fact supports the thesis that this mineralogic zone is a zone of retrograde metamorphism.

In the sillimanite schists the sillimanite itself is hardly oriented in its lowest rank occurrences, forming wispy disseminated clouds of fibrolite instead. As it grows in volume at the expense of andalusite and mica it acquires a spatial orientation. The prism axes line up parallel to the "b" tectonic axis and the thick matted clots of fibrolite become more discoidal with their plane in the plane of foliation.

2. Lattice and Dimensional Orientation of Minerals

Throughout this sequence of changes in texture the tectosilicates themselves do not appear to take up a preferred orientation relative to the foliation. Universal stage measurements of the distribution of c-axis orientations of quartz were made on two specimens, a calcareous quartzite from subunit 2 of French Valley formation, Member A and a hornblende meta-arkose from subunit 1 of Member B. No suggestion of preferred orientation was observed on the basis of contoured plots of 75 c-axis in each section. High-rank quartzites in which a strong lineation due to orientation of mica was observed were studied by use of a gypsum plate, as discussed below under "lineation". In any case the external form of the tectosilicate grains is not strongly affected by the shear stress producing the mica orientation. Andalusite tends to be roughly oriented with its longest prism axis in the plane of foliation. Tourmaline is completely random in its orientation, and may in fact be a much later mineral, formed during the period of retrograde metamorphism producing the muscovite.

3. Lineation

Two general types of lineation are observed in the metamorphic rocks of the Winchester-Hemet area: 1) structurally defined lineations such as crinkles, plane intersections, fold axes; and 2) crystalloblastic lineations. Both types of lineation can be related in orientation to the major fold axes. By far the most common orientation of both types of lineation is parallel to the "b" tectonic axis of folding. However local fold patterns not strongly related to the overall structural pattern seem to have a predominant influence on the orientation of these linear elements in some areas. A stereographic projection (equal area) of 35 lineations observed in a 1.5 mile wide strip parallel to the Domenigoni Valley syncline shows a strong clustering of lineations parallel to the presumed axis of the syncline (fig. 14).

Southwest of this strip, lineations of all types cluster around an eastward maximum with a plunge of 50 to 70 degrees. A lineation trend in this direction is seen on Double Butte, on the west end of Searls Ridge and in the low-rank French Valley schists east of French Valley. An equal-area projection of 59 lineations measured in this area (fig. 15) shows a vague concentration of points centering around a plunge of 65° and a strike of N 80° E. In making this plot it is evident that lineations observed in beds striking off the trend of regional foliation (NW) have been rotated with the bed. No attempt has been made to correct for this rotation in plotting the lineations. Apparently the linear element we observe here was developed before the last period of folding of the metasedimentary rocks. No major structures are evident in this locality with which to relate this linear E-W trend. It is suggestive however to note that a similar trend occurs in the southwest corner of the area (section B29) where cross-folds have been produced with steep westward-plunging fold axes. These cross-folds are apparently related to the local compressive stresses applied in a N-S direction by the granodiorite and

FIG 14 Equal-area projection of 35 lineations (axes of minor folds and crumples) measured in a 1.5-mile wide strip along the axis of the Domenigoni Valley syncline. Plotted on lower hemisphere.

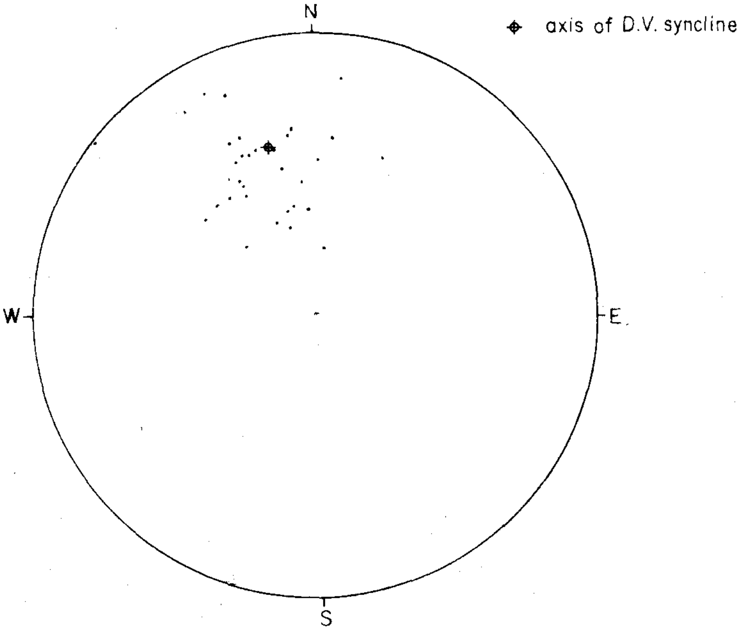
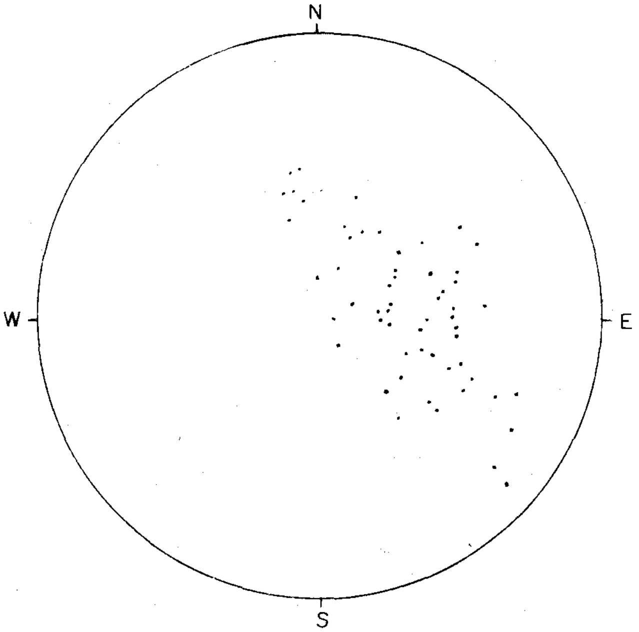


FIG. 15 Equal-area projection of 59 lineations (axes of minor folds and crumples) between French Valley and Double Butte, west of Domenigoni Valley syncline.



tonalite intrusives. On a larger scale the compression indicated by the lineation trend in the western part of the area may represent a N-S stress system set up during the intrusion of the Lakeview Mountain tonalite on the north or the other plutonic bodies on the south.

The lineations, considered in detail are as follows:

1. Structural lineations:
    a) Fold axes: drag folds on the flanks of major structures as well as minor folds not apparently related to any larger structure, observed in the French Valley zone.
    b) Intersection of cross-foliation (axial-plane foliation) with bedding plane foliation, especially in graphitic schists and banded gneisses.
    c) Crinkling of micaceous layers, often gradational into b) above, especially in finely laminated schists. Rarely parallel to "a" tectonic axis.
    d) Boudinage structures due in part to extensive development of axial plane foliation in banded gneiss, isolating elongate masses of quartzite between two sets of micaceous layers intersecting in a "b" lineation.
    e) Streaking in cataclasites: marked by streaked-out clusters of grain fragments (feldspar, quartz, garnet, etc.) milled down from coarser grains (Plate 12b, appendix).
2. Crystalloblastic lineations:
    a) Sillimanite fibers, elongate parallel to "b" in crumpled and tightly folded rocks.
    b) Biotite flakes, elongate in their a crystallographic direction,

oriented parallel to "b" tectonic axis. Measurements of this orientation effect are discussed below.

c) Amphibole prism orientation: most strongly developed in tremolite-anthophyllite schists, very weak in amphibolites.

d) Quartz c-axis orientation: most perfectly developed in sheared quartzites of A24(4, 4) and B19(1, 4). Not present in schists in general except where 2 b) is highly developed.

Various combinations of crystalloblastic and structural lineations are found depending on the mineralogy and texture of the rocks.

Boudinage is common only in the banded gneisses of Member C of the French Valley formation. It is especially well-developed in the tightly folded rocks near the axis of the Domenigoni Valley syncline and on the south side of Double Butte, again in tight folds in banded gneiss. On weathered outcrops the rods of quartzite isolated in these boudins stand out in low relief against the enclosing schist layers. Incipient boudinage of this type is marked by thin fractures parallel to the axial plane, cutting the quartzite bands. As it develops further, flattened rods of quartzite are formed separated from one another by dike-like intrusions of micaceous material with a well-defined axial plane foliation. At the very crest of the folds even the mica in the schist bands (still conforming to relict bedding) is parallel to the axial plane of the fold. Except for the persistence of the quartzite bands bedding plane foliation would be completely obliterated. For this reason there is serious question as to how much unobservable folding of this type has occurred in the homogeneous graphitic and biotite schists interlayered with the banded gneiss. In the absence of such rigid marker units such folding would be indiscernible and essentially equivalent to shear folding.

Crystallographic orientation of mica is present even in the lowest rank phyllites. The earliest formed orientation is, as is typical in mica schists, an orientation of the normals to the basal cleavage of the mica parallel to the normal to foliation and relict bedding. Such an orientation is thought to be largely stress induced although it may be in part derived from inherited flattening of clay-mineral flakes into the bedding planes of the parent shales. An attempt was made to test whether in addition to this orientation there is a linear orientation of any crystallographic direction of the mica in the plane of foliation. This was tested by cutting a thin section parallel to the plane of foliation and measuring the azimuthal orientation of the optic plane of the biotite, the crystallographic plane 010. This is done by viewing the biotite grains in conoscopic illumination on a petrographic microscope and noting the position of the rotating stage at maximum separation of the isogyres. Two specimens were studied in this way. The first specimen, WH 338, is of a homogeneous biotite-sillimanite-andalusite-quartz gneiss from a gneiss lens in subunit 2 of Member A of the French Valley formation, collected at peak 2331 in section A25(4, 2). The specimen was selected for analysis because of a pronounced crystalloblastic lineation developed in both sillimanite and biotite, visible in the hand specimen. The plot of 010 planes confirms this identification of a lineation in the biotite, showing a strong preferred orientation of these planes parallel to the "b" tectonic axis. In this same section the orientation of quartz was estimated by observing the overall color change on rotation of the slide in crossed nicols and with the gypsum plate inserted. There appears to be a girdle of quartz axes steeply inclined to the "b" direction. It is impossible to say, without recourse to universal stage methods what is the apical angle of the cone projecting into this girdle. The second specimen, WH 121, is of banded gneiss from the southeast side of Double Butte. A crinkle-lineation is present in the specimen parallel to local fold axes in the banded gneiss. A plot of 010 planes (fig. 16) shows a very weak concentration of optic planes parallel to "b" almost indistinguishable from a random distribution.

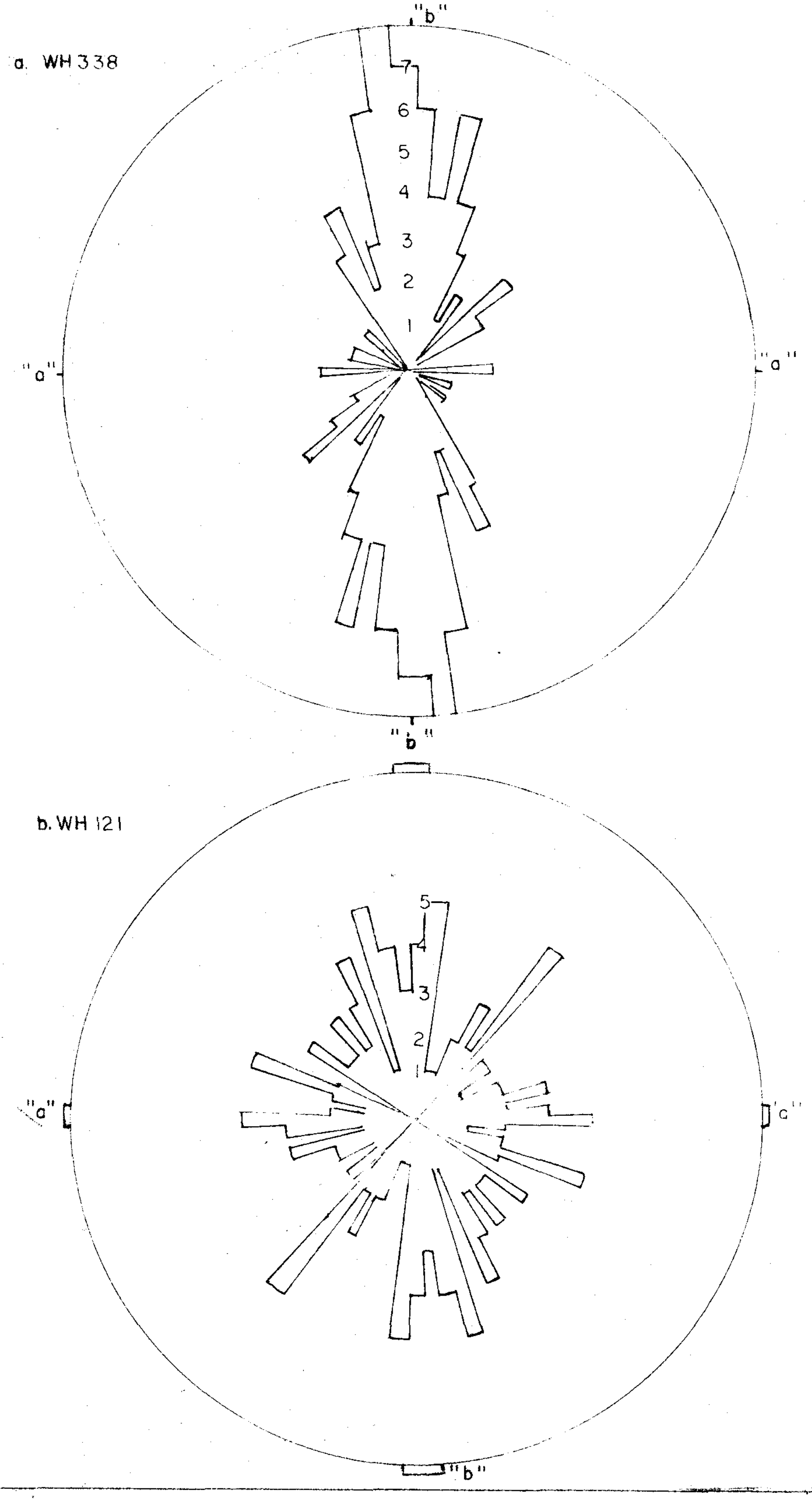
This evidence, albeit very incomplete, indicates a tendency for biotite in these rocks to orient its space lattice with the [100] axis parallel to the "b" tectonic axis, and with the 001 plane in the plane of foliation. We are assuming here that the biotite is monoclinic, the most common crystallographic system of naturally occurring biotites (Eugster and Yoder, 1954). This is in agreement with the large 2V's observed in these crystals. The orientation effects may either be the result of the presence of a preferred glide direction lying in the 001 plane, or a tendency to recrystallize in this orientation in order to minimize the stress-induced chemical potential of its components.

Fairbairn (1949, p. 19), summarizing the literature on investigations of this problem indicates that for muscovite the orientation under stress is just the opposite to that observed here for biotite, with the [100] axis in the foliation plane but perpendicular to "b".

4. Cataclastic Rocks

In addition to the above described petrofabrics originating by deformation and recrystallization under stress, there are scattered occurrences of cataclasites in which recrystallization has been moderate

FIG 16 Orientation of 010 planes of biotite in schist.(Five degree sectors).


a. WH 338

b. WH 121

or lacking following the period of milling-down. This textural type is most commonly found in the quartzites and igneous rocks of this area. For example, in subunit fvA5 in A24(1, 3) southwest of the migmatite-breccia unit, the conglomeratic quartzite of this stratigraphic member has been milled down to a flaser gneiss in which only medium-grained "eyes" of quartz can still be recognized megascopically. This rock forms a lens parallel to relict bedding. The rock is massive and breaks across its weak foliation. A pronounced lineation is observed defined by streaking out of schist clasts to many times their original length.

Cataclastically deformed quartzites similar to this one are found as lenses in various parts of the French Valley section. They are especially common in the arkoses of Member B. They appear to be limited to rocks of andalusite grade or lower, perhaps because higher rank examples have been recrystallized. This would put the time of cataclasis as before the culmination of the regional metamorphic episode. The cataclastic rocks form tabular masses oriented parallel to the foliation. The lineation defined by the streaking out of clasts in the rock plunges generally northward in a few places where it can be observed (for example, in A11(2, 2)). This feature appears to indicate displacement parallel to the axes of the major folded structures during deformation. All mylonitic deformation in the metasedimentary rocks is localized in planes trending northwest but this may be an indication of the relative ease of dislocation along foliation planes rather than an indication of the orientation of the stress system imposed during cataclasis.

Among the igneous rocks cataclastic textures are wide-spread. The concordant aplite dikes are almost invariably flaser gneisses, that is, cataclastic rocks consisting of layers of mylonitic material enclosing lensoid masses of unmilled rock. The strongly sheared Woodson Mountain granodiorite body south of Rawson Ranch has been discussed earlier. It shows internal effects of pronounced shearing after recrystallization. The adjacent quartzites exhibit a steeply plunging lineation marked by milling down of feldspar porphyroblasts and quartz grains (Plate 12b, appendix). These rocks have evidently been sheared by the drag of the rising granodiorite pluton against its walls. This texture disappears a few hundred feet away from the igneous contact. Schists interlayered with these quartzites do not display such cataclastic effects. The Lakeview Mountain tonalite shows no such cataclastic features even near intrusive contacts, nor does the San Marcos gabbro. The Domenigoni Valley granodiorite, while not sheared itself, is cut by a host of highly sheared quartz-latite dikes. These dikes exhibit augen-gneiss and flaser-gneiss textures. There may have been relatively little actual reduction in grain size due to milling in these rocks since even relatively unsheared rocks of this composition are very fine-grained and only slightly porphyritic.

The cataclasis of the latite dikes deserves further mention, in part because it is such a consistent and pervasive feature of this rock type and secondly because, since it affects the youngest rock of the area, it is evidently a result of a relatively recent structural event.

In the first place it is of some interest to know whether the phenomenon exemplified here is actually cataclastic in nature or rather a protoclastic feature developed late in the consolidation of the dikes. In this regard it is significant to note that in all the dikes exhibiting this shearing effect there is some development of a secondary, non-igneous assemblage of minerals, typically chlorite, sericite and epidote. In some of the more mafic-rich dikes (probably dacitic in composition) the rock has been converted to a fine-grained chlorite schist. Although it is not impossible that the shearing episode may have taken place during a period of late deuteric alteration of the original dike rock this association of a low-rank mineral assemblage with the superimposed milled texture suggest post-crystallization metamorphism.

The striking uniformity of trend of the dikes and the parallelism of trend with regional schistosity of the country rock suggests that the pathways along which the latite magma travelled were created by a preferential fracturing of the igneous host rock (principally Domenigoni Valley granodiorite) along a plane of crustal weakness. It is reasonable to presume therefore that the later movements which sheared the dikes also took place along these planes, as is evidenced by the fact that the foliation in the dikes parallels the dike walls. The lineation defined by the streaking out of biotite and feldspar grains in the milled down matrix plunges very nearly due east. The dikes dip in general northeast. Because the dikes dip very steeply the principal displacement associated with the mylonitization process must have been in a vertical sense. Presumably the major principal axis of stress was also oriented in an E-W direction.

The writer has searched for similar expression of late cataclastic deformation related to such a stress system in the nearby metasedimentary rocks. The few examples noted above are all of relatively moderate intensity of cataclasis compared with the latite dikes. Apparently the micaceous rocks could more readily yield by slippage along schistosity planes and in this way release the stress imposed on them. No obvious retrograde metamorphic effects are associated with these mylonitic zones in the metasedimentary rocks.

E. RELATION OF INTRUSIVE BODIES TO THEIR WALL ROCKS

The concordance of the contacts of the Lakeview Mountain tonalite with the foliation of the metasedimentary gneisses in the eastern part of the area suggests that the forcible intrusion of the tonalite may have had a role in compressing the gneisses into the presently observed series of tight isoclinal folds. The tonalite was evidently injected into the core of a zone of progressively metamorphosed and highly deformed rocks, presumably while they were still in a physical state permitting plastic deformation. In this area it is impossible to separate the pre-intrusive deformation from deformation induced by the expanding magma chamber.

Deformation induced by other intrusive bodies has been discussed above. To summarize these effects:

1) Intrusion of the Woodson Monntain granodiorite and Lakeview Mountain tonalite south of Black Mountain apparently deformed the sediments into a group of E-W trending folds.

2) The Domenigoni Valley granodiorite has pried apart the strata of the Bedford Canyon formation in French Valley, causing a sharp change in foliation attitudes from NW strikes to EW

strikes.

3) At Double Butte, the combined effects of intrusion of tonalite on the north and later granodiorite intrusion on the south have squeezed the metasediments, causing them to pinch out to the northwest. The east-plunging "b" lineation in the metasediments may be in part due to this compression.

Other local effects at the borders of the Woodson Mountain granodiorite and the ultrabasic rocks of the magnesite mine indicate that plastic deformation of the enclosing schists was by and large the principal mode of deformation during intrusion.

Evidence of stoping is rare. Except for the "Bonsall" tonalite zone of the Lakeview Mountain tonalite this rock contains few inclusions. The "Bonsall" zone is restricted to exposures south of the Lakeview Mountains ; contacts between the Lakeview Mountain tonalite and metasedimentary rocks on the east end of Diamond Valley do not exhibit such a zone. The Woodson Mountain granodiorite contains no more than 5 per cent by volume of inclusions. These are highly migmatized fragments of gneiss. They are streaked out and do not increase in abundance near the igneous contacts. The Domenigoni Valley granodiorite contains abundant inclusions but again there is no apparent relation between inclusion abundance and distance from intrusive contacts. The inclusions in this rock, as in the "Bonsall" tonalite appear to be derived from basic igneous rock rather than from pelitic or arenaceous metasedimentary rock. In general quartz is less abundant in the inclusions than in the enclosing rock. These inclusions were evidently stripped at considerable depth, perhaps from gabbroic rocks, since they are not for the most part near any apparent source rocks at present.

Another indication of the extent to which stoping provided room for the intruding plutonic igneous rocks is the amount of stratigraphic section which has been removed adjacent to the igneous rock contacts. At the southeast end of the Domenigoni Valley granodiorite pluton it is apparent that virtually the whole of Member A of the French Valley formation has been removed by the intrusive rock. This is the most striking evidence of stoping in the area. In contrast the large Woodson Mountain granodiorite intrusive and the smaller San Marcos gabbro body on its north do not appear to be related to any large removal of section. On its southwest flank this intrusive mass obliterates a small part of Member A of the French Valley formation. Further to the east and south stratigraphic control is not sufficiently precise to permit commenting on possible loss of section due to stoping. The same is true of the Lakeview Mountain tonalite intrusives on both the east and northwest sides of the Winchester-Hemet area. These intrusives are for the most part observed on one side of the metasedimentary rock sequence only and it is not possible to say how much if any of the stratigraphic section once adjacent to the present igneous contact (on the east or north) is now completely removed.

The San Marcos gabbro intrusive of Bachelor Mountain, partly shown on the south edge of fig. 1 (A27, 34, 35), has evidently cut away most of the Bedford Canyon formation section which projects along strike into it. There is no indication of deflection of strike or dip in these beds as they approach the pluton. However immediately adjacent to this intrusive structural attitudes are largely obscured by a soil zone extending around the foot of Bachelor Mountain.

The evidence presented above suggests that stoping of the adjacent metasedimentary rocks was locally important in providing room for the invading igneous plutons. Little evidence of this process is provided by inclusions in the igneous rocks themselves. The abundant inclusions in the "Bonsall" tonalite, although suggesting large-scale stoping, are evidently in part or wholly derived from a basic igneous rock and, except for the local contact with San Marcos gabbro south of Florida Ave., the possible source of these basic inclusions is not exposed. The San Marcos gabbro, where it is in contact with this "Bonsall" zone (as in D14, 15) appears to be younger.

Buddington (1959), in his review of the various modes of "granite" emplacement, refers the southern California batholith to the "transitional epizone-mesozone", stating that (p. 694):

> "The huge batholith of southern California displays predominant discordant contacts and certain other features which ally it with the plutons of the epizone whereas the near-absence of chill zones and the occurrence of internal border foliation and local concordance of structure of country rock to contacts tie it to batholiths of the mesozone."

As described by Buddington (p. 676), the criteria for determining the depth zone of a batholith are based on the nature of the surrounding country rocks and the relation of the pluton to those rocks rather than on the nature of the rocks themselves. Using Buddington's own criteria it would seem on the basis of the present observations (and those to follow in Chapter VI) that the portion of the batholith exposed here is representative of the transitional mesozone to catazone, indicating depths of 7-9 miles and country-rock temperatures of 400-600° C. However it should be clearly noted that this applies only to the larger batholithic bodies, the Lakeview Mountain tonalite and the Woodson Mountain granodiorite in particular. These have been intruded into the higher rank part of the metamorphic block exposed here. The smaller plutons of Domenigoni Valley granodiorite and San Marcos gabbro, though only a few miles distant or actually in contact with the former intrusives, seem to represent transitional epizone-mesozone intrusions as described by Buddington.

The possibility of such differences in depth zone occurring within a single batholith has been noted by Buddington (p. 676). Elsewhere within the southern California batholith examples of epizone intrusion can be found (e. g., the Temescal Wash quartz latite porphyry) and by and large the western half of the batholith appears, on the basis of Larsen's work, to be better fitted to the epi-mesozonal category than the eastern half.

F. AGE OF THE DEFORMATIONS

In the light of the above descriptions it is possible to interrelate to some extent the time of various structural events. The earliest features to develop, perhaps while the sediments were only weakly metamorphosed, were lineations parallel to an E-W trend of folding. The folding itself was not expressed in these rocks except on a small scale and as crinkles parallel to axes of these small folds. Local cataclasis may have taken place at this time, too, localized in the quartzites. This feature was largely obliterated by later metamorphism.

Following the period of N-S compression a period of compression in a NE-SW direction occurred, simultaneous with further increase in metamorphic rank of the schists. The metasediments were deformed into NW-SE trending folds and structural and crystalloblastic lineations were developed, principally parallel to the axes of folding. Intrusion of the major igneous plutons began in this period and continued through it. This would place the age of major folding as being in the Cretaceous. Late in this epoch but while the rocks were still in a P-T environment permitting them to flow plastically, a local right lateral shear-stress system acting along N-S trending planes produced a cross-folding pattern and eventually a series of faults along Rawson Canyon. This event probably followed intrusion of the Lakeview Mountain tonalite and certainly preceded intrusion of the Woodson Mountain granodiorite. Accompanying these intrusions a second period of cross-folding occurred along a set of E-W trending axes, probably due to stresses applied by the two successive intrusions.

The intrusive epoch closed with the intrusion of the Domenigoni Valley granodiorite, further distorting the schists in the vicinity of French Valley, followed by the intrusion of the quartz-latite dikes. Some time after intrusion and crystallization of these dikes they were sheared and milled down, possibly due to an E-W oriented stress system.

The last structural event to occur was minor dip-slip movement on one of the branches of the Rawson Ranch fault zone, resulting in vertical uplift of the east side of the zone. This last event may be a minor effect of stresses applied to the Perris block during vertical movements on the San Jacinto fault.

VI. METAMORPHISM

A. INTRODUCTION

The major metamorphic rock types have already been discussed under "Stratigraphy" (Chapter II). The present chapter will reconsider these types for the sake of the information that their mineralogy and texture yields to an analysis of the metamorphic history of this area. In its final sections we will utilize this data in conjunction with other chemical data and information from hydrothermal synthesis studies to attempt a quantitative evaluation of the physical variables of metamorphism.

An overall survey of the distribution of various types of metamorphic rock through the Perris block and the Peninsular Ranges to the south indicates that in general as one progresses from southwest to northeast the degree of chemical, mineralogical and textural reconstitution increases. Northeast and east of the San Jacinto fault zone the typical metamorphic lithology is highly reconstituted pelitic gneiss and schist, but because of the uncertainty as to the amount of lateral displacement on this fault one would have to restrict such generalizations about regional metamorphic trends to the southwest side of this fault.

The Winchester-Hemet area displays a continuous series of metamorphic lithologies from phyllites to coarse-grained gneisses. In addition the structure of the metasedimentary rocks is such that one may trace a single stratum through all degrees of reconstitution. These two facts make this area a well suited locality for study of the metamorphic history of the pre-batholithic sediments.

B. THE METAMORPHIC ZONES AND THEIR ISOGRADIC BOUNDARIES

1. General Description

The Winchester-Hemet area has been divided into four metamorphic zones based upon the characteristic mineral assemblages found in pelitic schists of each zone. These zones are shown on the overlay to the geologic map, fig. 2. Their boundaries have been determined on the basis of field observations of mineralogy supplemented by some petrographic data. The localities of such petrographic data are indicated on the overlay map. The isograds are drawn to conform to adjacent isograds where no direct evidence is available due to lack of pelitic rocks.

Various chemical and physical parameters are discussed in this chapter with respect to their variation with metamorphic grade. Where these parameters are shown graphically (figs. 23-25) the abscissa of each point is meant to be the inferred relative rank of metamorphism of the sample in which the parameter was measured. This rank is estimated as follows:

The points from which the samples were obtained are projected on a map of the isograds (fig. 2). A set of curves is constructed parallel to the isograds and passing through each of the points. A single (arbitrary) E-W line is constructed crossing all the curves. The curves (and thereby the points from which they were constructed) can then be ordered with respect to the sequence (from west to east) of their intercepts with this straight line. The relative position of these intercepts is used as the abscissa on the plots of the various parameters. The E-W line used in obtaining the ranking shown in figs. 23-25 was a line parallel to the sequence of samples of unit fvB4 (described in Part II). In most cases the sample points themselves fell on or close to the line and estimates of rank are probably quite reliable.

2. Metamorphism of the Pelitic Schists

a. The Muscovite Zone

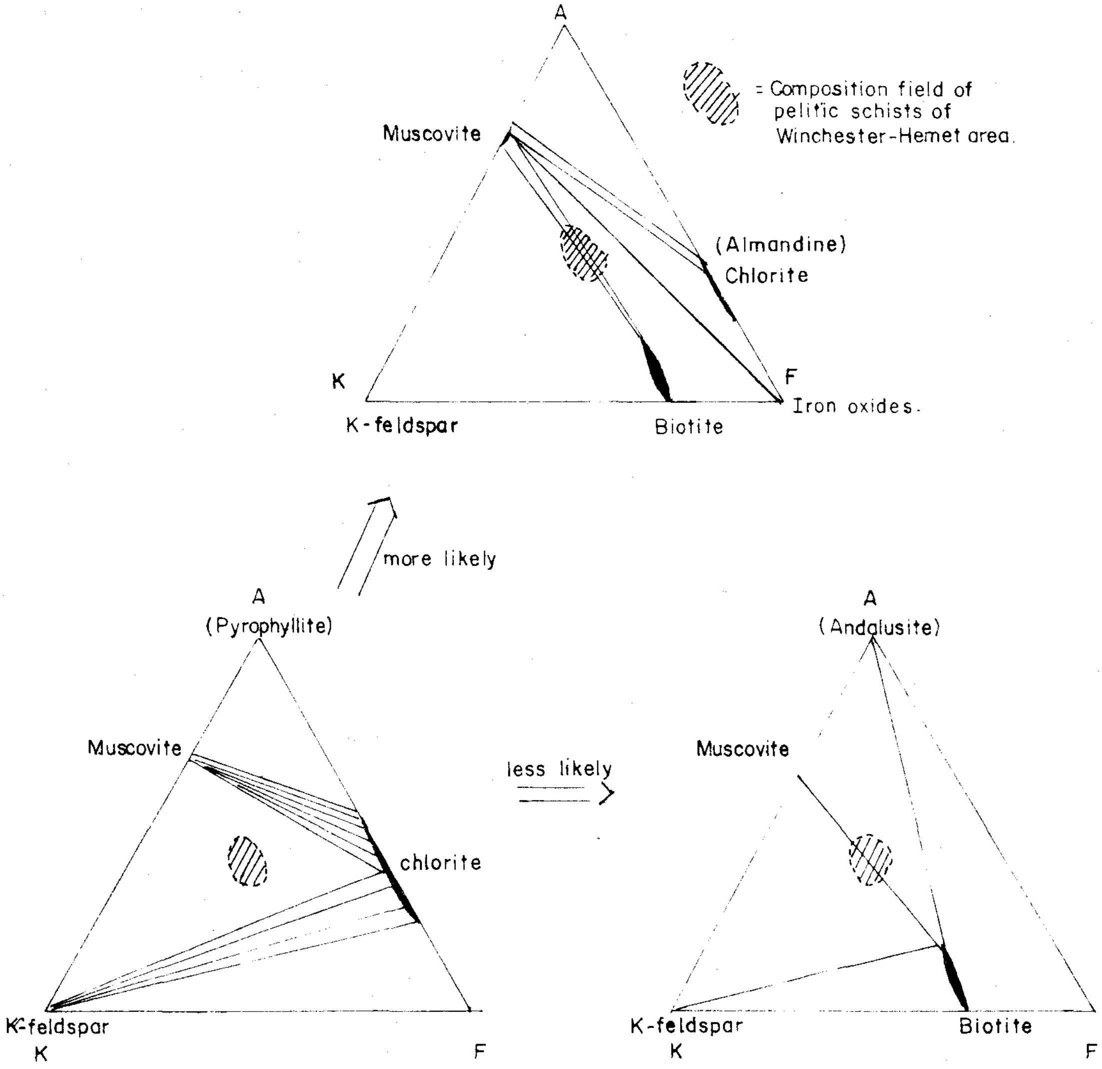
The lowest rank pelitic schists of the Winchester-Hemet area, including most of the Bedford Canyon formation exposures and the lower part of the French Valley formation, contain the mineral assemblage biotite-muscovite-quartz-tourmaline-potash-feldspar-oligoclase. Feldspar of both types is generally present in only accessory amounts. Variable amounts of graphite, magnetite and other unidentifiable opaque accessories are also present in most specimens.

Although lower rank equivalents of this lithology are not exposed in this area it is possible to find schists in the Bedford Canyon formation exposed in the Santa Ana Mountains, about 30 miles to the west, of about the same composition. These have the mineralogy muscovite-chlorite-quartz-albite-epidote. Potash-feldspar varies in abundance and was completely absent from several thin sections of rocks from the Santa Ana Mountains studied by the writer.

Although incomplete reactions producing the assemblage exposed in French Valley cannot be found occurring in any rocks from this area it is possible, on the basis of the present mineralogy to infer what reactions must have taken place. The present assemblage is plotted on an AKF diagram in fig. 17a with possible lower-rank assemblages shown on an adjacent diagram (18b). It is seen that to produce muscovite and biotite from a chlorite schist such as is found in the Santa Ana Mountains it would be necessary to add K_2O, probably in the form of

FIG.17 AKF diagrams of pelitic schists of the muscovite zone showing possible modes of origin.

a. Muscovite zone assemblages found in W-H area



b. Greenschist facies equivalent, found in Santa Ana Mtns.

c. Alternate muscovite zone assemblages, not observed in W-H area.

potash-feldspar. This suggests that potash-feldspar may have been more abundant on the whole in the Bedford Canyon sediments of this area than in their type locality. The chemical equilibrium is such that all of the feldspar (excepting accessory amounts observed in the schists at present) was consumed by the reaction forming biotite:

$$\text{Chlorite} + \text{K-feldspar} = \text{Biotite} + \text{Muscovite} \qquad (1)$$

The other possible reaction forming biotite, as seen in fig. 17 is

$$\text{Chlorite} + \text{Muscovite} = \text{Biotite} + \text{Al-silicate} \qquad (2)$$

This reaction would leave an aluminum silicate phase, not observed in the schists of French Valley. However spots of muscovite-rich material scattered through these rocks may represent pseudomorphs after andalusite. This in turn raises the question of where potash came from to convert the andalusite to muscovite. It is not impossible that some K_2O metasomatism occurred in these sediments during metamorphism. The geochemical studies on the meta-arkose of Member B of the French Valley formation, described in Part II of this thesis, do not suggest this possibility, however.

Chlorite bearing schists crop out in only one locality in the Winchester-Hemet area, on the west end of Searls Ridge. They are found near the andalusite isograd, but are interlayered with typical muscovite-biotite schists. In thin section it is seen that potash-feldspar is absent. These apparently represent rocks in which chlorite is in excess over the amount required to react with all the potash-feldspar originally present in the rock.

b. The Andalusite Zone

The first appearance of andalusite in the pelitic schists marks the lower-rank boundary of the andalusite zone. Most discussions of the paragenesis of andalusite (Ramberg 1952a, p. 48; Fyfe, Turner and Verhoogen, 1958, p. 164; Yoder and Eugster, 1955, p. 270, for examples) attribute the appearance of this phase to the following reaction:

$$\underset{\text{muscovite}}{KAl_3Si_3O_{10}(OH)_2} + SiO_2 = \underset{\text{K-feldspar}}{KAlSi_3O_8} + \underset{\text{andalusite}}{Al_2SiO_5} + H_2O \qquad (3)$$

However this reaction would involve production of one mole of potash-feldspar for every mole of andalusite. In fact very little potash-feldspar is found in the andalusite schists and what little is observed could be accounted for as primary clastic material residual from rection 1 above. Typical field occurrences described in the literature (Hietanen, 1956; Barth, 1936; Eskola, 1914) consist of biotite-quartz-muscovite-andalusite schists lacking potash-feldspar or any other potash-bearing silicate. Three alternative hypotheses may be entertained as to the origin of andalusite in this area:

1) K_2O is carried off as a mobile component.
2) The andalusite schists differ in initial bulk composition from the typical muscovite-biotite schists of the muscovite zone, having excess Al_2O_3 in the form of clay minerals or pyrophyllite.
3) Iron oxide minerals have reacted with muscovite to form a ferrimuscovite or biotite and andalusite.

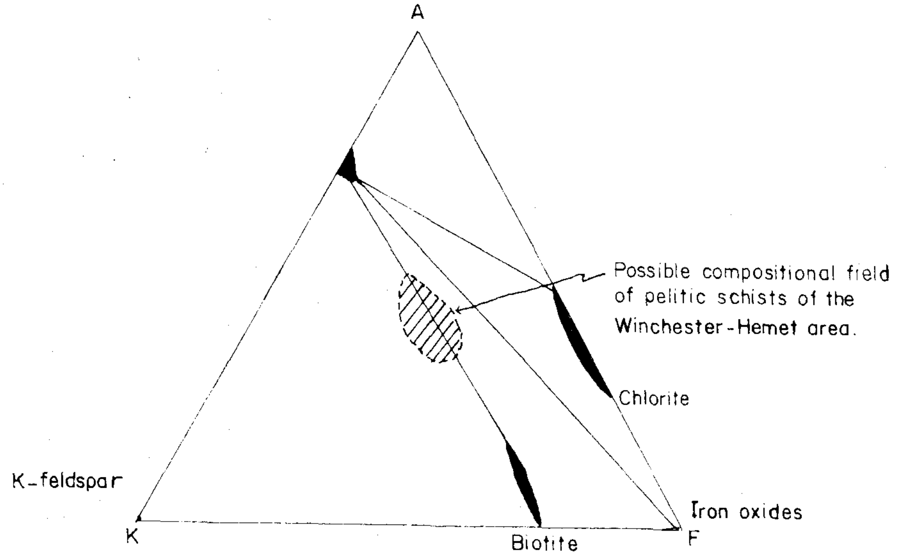
The first hypothesis, although not subject to decisive test without the benefit of chemical analyses, is unlikely since it is evident that some potash-feldspar is stable in the rocks and it is unreasonable that this material should not also have diffused away. Hypothesis 2) above is countered by the observation that the andalusite isograd transgresses lithologic (stratigraphic) boundaries and is roughly parallel to the zones of higher rank assemblages, indicating that it is a true isograd rather than a line of changing bulk composition. No excess aluminous phases are observed in the lower-rank equivalents although pyrophyllite would be almost impossible to distinguish from muscovite.

For the third process a reaction such as the following must take place:

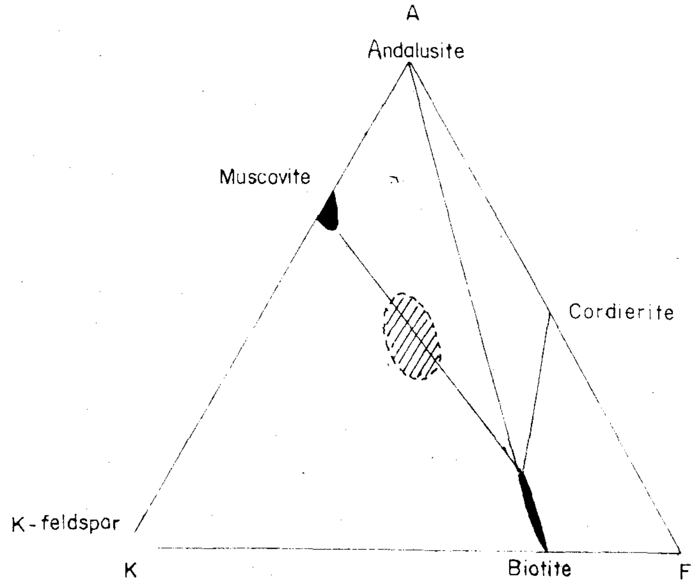
$$\mathrm{SiO_2 + 3FeO + KAl_2Si_3AlO_{10}(OH)_2 = Al_2SiO_5 +} \underset{\text{(annite)}}{\mathrm{KFe_3Si_3O_{10}(OH)_2}} \qquad (4)$$

The phase relations are shown on the AKF diagram in fig. 18. The compositional field shown for the Winchester-Hemet area is based on the observed mineral assemblages. Since biotite is already present in the muscovite-zone rocks this would be equivalent to adding additional Fe to the biotite. There is no evidence that the composition of biotite from the schists changes appreciably on entering the andalusite zone. The index of refraction and color of biotites from either side of the andalusite isograd are essentially identical. The FeO entering into this reaction may be derived from limonite, hematite or magnetite. These may be present as indistinguishable opaque minerals in the lower grade schists. The question is then raised as to what oxidation state of the iron would be stable at the temperatures and pressure of metamorphism encountered here. Evidence from the geochemical study described in Part II of this thesis indicates that with progressive metamorphism Fe in the sediments

FIG.18 Paragenesis of andalusite in the Winchester-Hemet area.



A. Muscovite zone



B. Andalusite zone

tends to become reduced. The reaction forming andalusite may be controlled in part by the availability of FeO as produced by the reaction

$$Fe_2O_3 + H_2 = 2\,FeO + H_2O \tag{5a}$$

or

$$Fe_2O_3 + \underset{\text{graphite}}{C} = 2\,Fe_2O_3 + C \tag{5b}$$

An alternative reaction that may form andalusite by a similar mechanism is:

$$SiO_2 + Fe_2O_3 + KAl_2Si_3AlO_{10}(OH)_2 = \underset{\text{Ferrimuscovite}}{KFe_2Si_3AlO_{10}(OH)_2} + Al_2SiO_5 \tag{6}$$

This would require that the iron content of the muscovite would rise with the appearance of andalusite. Again this is not borne out by any optical data on the muscovites observed on either side of the andalusite isograd. The effect on optical properties of limited solid solution of the ferrimuscovite molecule in muscovite is not well documented, however.

The possibility of aluminum being liberated by substitution of Fe^{3+} in tetrahedral coordination with oxygen is even less likely than the above hypotheses, due to the discrepancy in ionic radius between Fe^{3+} and Al^{3+} (0.65 and 0.51 Å respectively).

Cordierite appears in only minor amounts in the schists of the andalusite zone. Those schists in which cordierite occurs also contain muscovite and biotite so that if we plot their composition on an AKF diagram it appears that this four-phase assemblage is not in equilibrium. Presumably then we cannot assume that MgO and FeO are acting as a single component. The Fe/Mg ratios of coexisting biotite and cordierite must not be identical.

Using the projection of the AKFM tetrahedron proposed by Thompson (1957) this difficulty is reconciled (fig. 19). Potash-feldspar is absent from rocks containing cordierite. The estimated field of composition of pelitic schists of the Winchester-Hemet area is shown by the cross-hatched area on this diagram. This area is estimated on the basis of the mineral assemblages which are actually noted in these schists and is not based on any chemical analysis. It is chosen so as to overlap into all fields for which assemblages are represented. The assemblage almandine-potash-feldspar-biotite although apparently a stable one for certain compositions of rock, does not occur in this area until a higher grade of metamorphism has been reached, probably due to a widening of the compositional field of the garnet.

Cordierite apparently forms by the reaction:

Biotite + Andalusite = Cordierite + Muscovite (7)

Cordierite and andalusite are not found to co-exist in the schists of this area. The alternate possible reaction (see Fyfe, Turner and Verhoogen, 1958, p. 208):

Biotite + Muscovite = Cordierite + K-feldspar (8)

does not appear to take place since neither the assemblages Co-Bi-Kf nor Co-Mus-Kf are found in the cordierite bearing rocks. Reaction 7 should not be taken to represent a discrete reaction which takes place at a specific temperature and P_{H_2O}. As can be seen from figs. 19 and 20 the relative amounts of cordierite and biotite (+ muscovite + andalusite) present are in part dependent on the Mg:Fe ratio of the rock and the reaction shown above will take place at varying T, P_{H_2O} conditions as a

FIG.19 AKFM Diagram of pelitic schists of the andalusite and sillimanite zones.

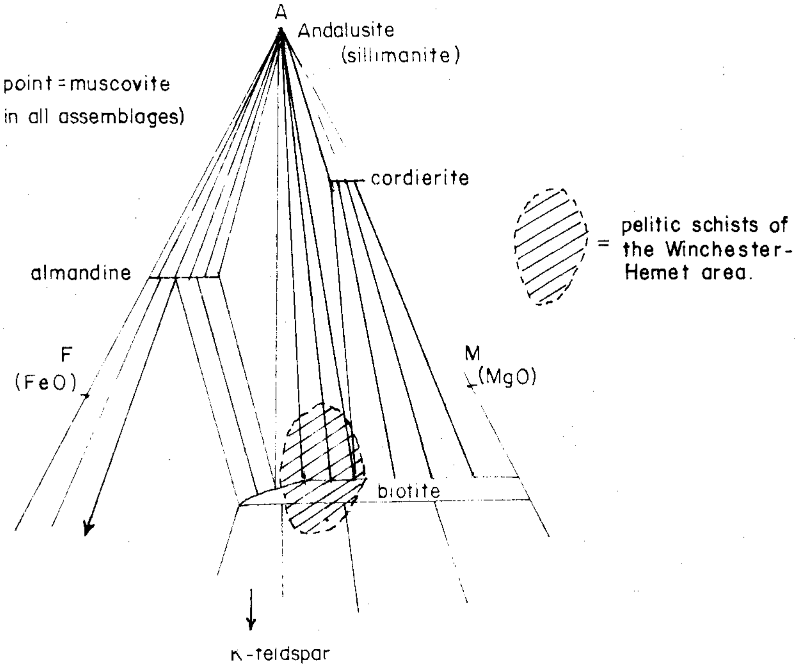
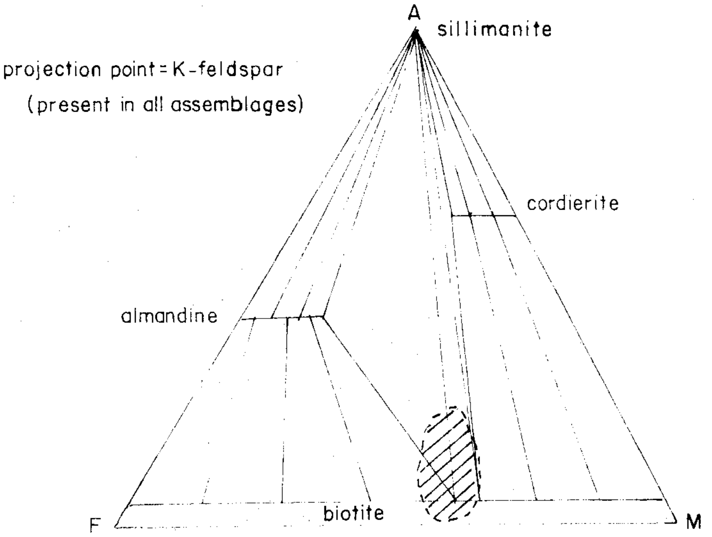


FIG. AKFM Diagram of almandine-garnet zone pelitic schists and gneisses.



function of the chemical composition of the system involved. The fact that cordierite is an extremely rare phase in the schists of the Winchester-Hemet area suggests that the T-P_{H_2O} conditions of metamorphism were such as to permit a relatively small compositional field in which cordierite might coexist with biotite (the right sides of figs. 19 and 20). An alternate conclusion would be that the parent sediments of the rocks exposed here were deficient in MgO relative to FeO.

c. The Sillimanite Zone

Sillimanite first appears as a phase almost coincident with the first appearance of andalusite. However it is present only as a minor phase up to the point where andalusite disappears. This is considered to be a disequilibrium phenomenon and is discussed as such below. The sillimanite isograd in this area is drawn as the locus of points of disappearance of andalusite in each pelitic stratum. All the schists and micaceous gneisses east of this line contain only sillimanite or muscovite-sillimanite or rarely, muscovite alone as an excess-alumina phase.

Sillimanite first appears as fibers replacing biotite and andalusite. Biotite continues to serve as a nucleus for the growth of sillimanite in even the highest rank rocks (Plate 13, appendix). However other discrete clots of fibrolite are also present in rocks with more than 5 per cent sillimanite. Rarely there are euhedral stubby prisms of sillimanite up to 0.1 mm. wide and 1 mm. long. Sillimanite does not appear to replace andalusite pseudomorphously at the sillimanite isograd, but rather forms elsewhere in the rock, especially around biotite, simultaneous with the disappearance of andalusite. The detailed textural relations between sillimanite and andalusite at the sillimanite isograd are obscured by the apparently retrogressive replacement of andalusite by muscovite in a narrow zone paralleling the sillimanite zone and between it and the andalusite zone. The retrogressive nature of this replacement is suggested in part by the fact that this muscovite, occurring as coarse unoriented tablets pseudomorphous after andalusite, is not itself replaced by sillimanite although oriented flakes of muscovite in the same rock are.

The reaction forming sillimanite from biotite is not known on the basis of experimental work but it presumably involves the conversion of a molecule of siderophyllite to feldspar, sillimanite and iron oxide, by reaction with quartz, as follows:

$$2SiO_2 + \underset{\text{"siderophyllite"}}{K_2Fe_5AlSi_5Al_3O_{20}(OH)_4} = 2KAlSi_3O_8 + Al_2SiO_5 + 2H_2O + 5FeO \quad (9)$$

The "siderophyllite" is of course only a component of the biotite. As a result of this reaction the remaining biotite must become more magnesian.

Coincident with the appearance of sillimanite the abundance of potash-feldspar in the schists increases. In the andalusite zone the estimated modal abundance of potash-feldspar is about 2 per cent (average). In the sillimanite zone the schists contain on the average about 15 per cent sillimanite and about 10 per cent potash-feldspar. The modal abundance of the sillimanite is especially difficult to evaluate because of the fibrous habit of this mineral.

The iron oxide liberated by reaction 9 may be partially oxidized to form magnetite. This reaction is at best schematic of the type of reaction forming sillimanite. Other possible reactions would involve the liberation, instead of FeO, of TiO_2 to form rutile or of MgO which would leave the system in solution. The possible reaction products (beside sillimanite and potash-feldspar) are seldom observed in thin section although traces of opaque minerals (magnetite?) are commonly found intergrown with sillimanite where it fringes biotite. Sillimanite associated with biotite is always faintly pleochroic, the intensity of absorption increasing with the rank of metamorphism. This suggests that Fe^{3+} produced in the breakdown of biotite is in part taken up into the sillimanite lattice, replacing Al^{3+}. This is possible since the aluminum is present in an octahedral coordination with oxygen, a position sufficiently large to accept Fe^{3+} as well.

Evidently the formation of sillimanite from biotite goes on at a lower grade of metamorphism than the breakdown of muscovite to form sillimanite and feldspar. Petrographic studies indicate that the latter reaction is just beginning to occur in rocks in which biotite is already densely rimmed with sillimanite. The reaction forming sillimanite from muscovite is

$$SiO_2 + KAl_2Si_3AlO_{10}(OH)_2 = KAlSi_3O_8 + Al_2SiO_5 + H_2O \qquad (10)$$

Perhaps sillimanite formed by reaction 8 may crystallize around the nuclei already formed around biotite, since muscovite is seldom rimmed by sillimanite as is biotite. There is some danger in confusing retrograde muscovite from that which was in equilibrium with sillimanite and biotite. The arguments as to sillimanite paragenesis given here refer to muscovite flakes that are clearly incorporated in the schistose fabric of the rock. In general the muscovite content of the sillimanite schists is about 2 per cent as compared with an average of about 10 per cent for the andalusite schists of comparable bulk composition.

d. Almandine Zone

Pelitic schists containing garnet are rare in the Winchester-Hemet area. It is difficult to define on the basis of their spotty occurrence a zone in which garnet is a stable phase. Evidently most of the schists fall in a compositional field such that the assemblage biotite-sillimanite-quartz-potash-feldspar-(muscovite) is stable up to the upper limit of metamorphism represented in this area. Therefore the almandine isograd has been drawn parallel to the sillimanite isograd and passing through the inferred lowest rank exposures of garnetiferous banded gneiss, those on Searls Ridge in D36(4,1). It is found that this isograd also roughly delineates the boundary of a zone in which garnet appears in rocks of non-pelitic composition (arkose, amphibolite and amphibolitic quartzite). West of this boundary, in the andalusite zone, grossularite is found in calcareous quartzites.

In the western (lower-rank) part of the garnet zone the appearance of garnet is restricted to the vicinity of locally dense swarms of pegmatites. Further to the east garnet is found to occur as a minor phase in schists not close to pegmatites but its distribution is nevertheless spotty. Although no chemical analyses of these garnets are available, some information can be gleaned from their physical properties. Three garnets which have been analysed are compared below to illustrate the changes observed:

	WH194, from aplite on Searls Ridge	WH 94, banded gneiss on Searls Ridge	WH 360, migmatite gneiss, Crown V.
n_{Na_D}	1.810	1.807	1.808
a_o (Å)	11.60 ± 0.01	11.561 ± 0.002	11.540 ± 0.002
color	violet-pink	pink	pink

These analyses have been plotted on a diagram showing the positions of the end-members of the garnet series (fig. 21). It is noted that all three fall very close to spessartite, and that the two samples from gneiss are somewhat closer to the almandite end than the aplite garnet, with perhaps the higher-rank (migmatite) garnet being the more almanditic of the two. Since all three garnets come from rocks with low CaO content it seems justifiable to neglect the ugrandite component in discussing these data.

On the basis of such incomplete evidence with regard to their composition only the most general inferences about the significance of the appearance of garnet can be made. Garnet is probably produced by the following reaction:

biotite + sillimanite = garnet + K-feldspar + vapor (11)

The estimated modal abundance of sillimanite in garnetiferous gneisses and schists is about 2 per cent as compared with the average of 15 per cent in the sillimanite-zone gneisses. No correlation between the abundance of potash-feldspar and that of garnet is observed, presumably because the feldspar is already present as a product of reaction 9.

On the basis of the physical determinations described above it appears that the garnet of the schists is related in composition to that of

(following page 209, facing fig. 21)

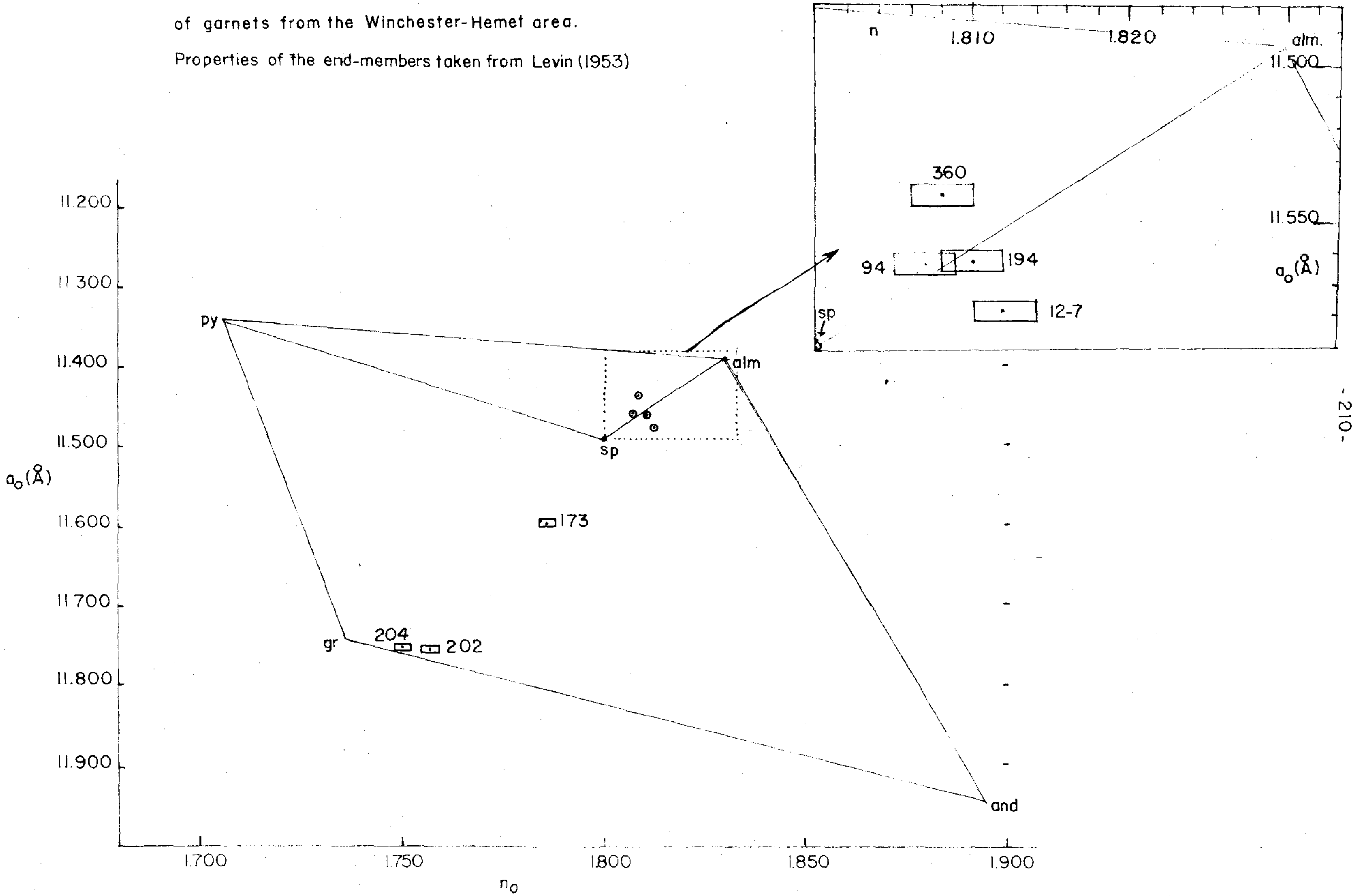
Fig. 21. Unit cell dimensions and indices of refraction of garnets.

Location of specimens:

94	Pelitic gneiss, Searls Ridge C31(1, 4)
194	Aplite, Searls Ridge, C31(4, 1)
360	Migmatite gneiss, east end of Crown Valley, B18(4, 2)
12-7	Meta-arkose (sample 4-7), west end of Crown Valley, B18(4, 2)
173	Amphibolite, Crown Valley, B17(3, 3)
202	Pyroxene granulite, east end of Searls Ridge, C28(1, 4)
204	Calcareous quartzite, Rawson Canyon, B5(2, 1)

FIG.21 Indices of refraction and unit cell dimensions of garnets from the Winchester-Hemet area.

Properties of the end-members taken from Levin (1953)



the pegmatites. Garnet of similar composition is present in aplites intruding non-garnetiferous lower-rank schists west of the garnet isograd. The pegmatites may contribute MnO to the garnet, or may merely act as sources of volatiles catalysing reaction 11. That is, the assemblage garnet-potash-feldspar-biotite-quartz may actually be stable throughout the garnet zone but the assemblage biotite-sillimanite-potash-feldspar-quartz has been preserved metastably due to an activation energy barrier to the completion of reaction 11.

The limited evidence provided by the physical measurements given above is insufficient to suggest any trends in the composition of the garnet with increasing grade of metamorphism.

Sillimanite persists through the garnet zone, perhaps less abundantly in the highest rank migmatites. Muscovite is absent in most of this zone. Locally it may be present as a retrograde alteration of sillimanite. Cordierite is observed in scattered localities, as fine porphyroblastic anhedral grains. The possible assemblages are shown in fig. 20 on an AKFM diagram. In this case the projection is made through the potash-feldspar corner ("K") of the AKFM tetrahedron since all phase assemblages contain excess potash-feldspar (muscovite is not stable in any of these assemblages).

e. Minerals of the pelitic schists

1) Biotite

In order to compare with changes observed in the biotites studied in Part II of this thesis, optical studies were made on biotites from various rocks sampled through the area. Little variation in refractive index was observed. Color variations are small and highly erratic. In general biotite from the muscovite zone is greenish brown while that from the highest garnet zone is reddish brown. Intermediate rocks may show all gradations between these two extremes. This probably reflects an increase in the TiO_2(+ Fe_2O_3 ?) content with increasing metamorphism. The indices of the biotites from various coexistent rocks appear to become more nearly alike with advancing metamorphism, though there is no pronounced increasing or decreasing trend (fig. 22). Using the plot of optical properties vs. composition of Winchell and Winchell (1951, p. 374) the Fe/Mg ratio of the biotites would average about 0.5.

2) Andalusite and sillimanite

Evidently neither the andalusite nor the sillimanite found in these rocks are pure aluminum silicates. Both minerals are colored, the color of the andalusite varying from a pale violet to deep pink, that of the sillimanite being various shades of pale brown. The optical properties of the andalusite match those given by Winchell and Winchell (1951, p. 521) for a slightly ferriferous member of the series Al_2SiO_5 - $(Fe, Mn)_2SiO_5$. A typical andalusite from the middle of the andalusite zone has $n_z = 1.632$, $n_x = 1.642$.

The refractive indices of sillimanite from various localities in the sillimanite zone are considerably different from those given by Winchell or Tröger. Values for n_z are given below:

	n_z	n_x
WH 82	1.672	
WH 164s	1.670	1.648
WH 153A	1.659	
WH 283	1.659	1.640
WH 338	1.657	

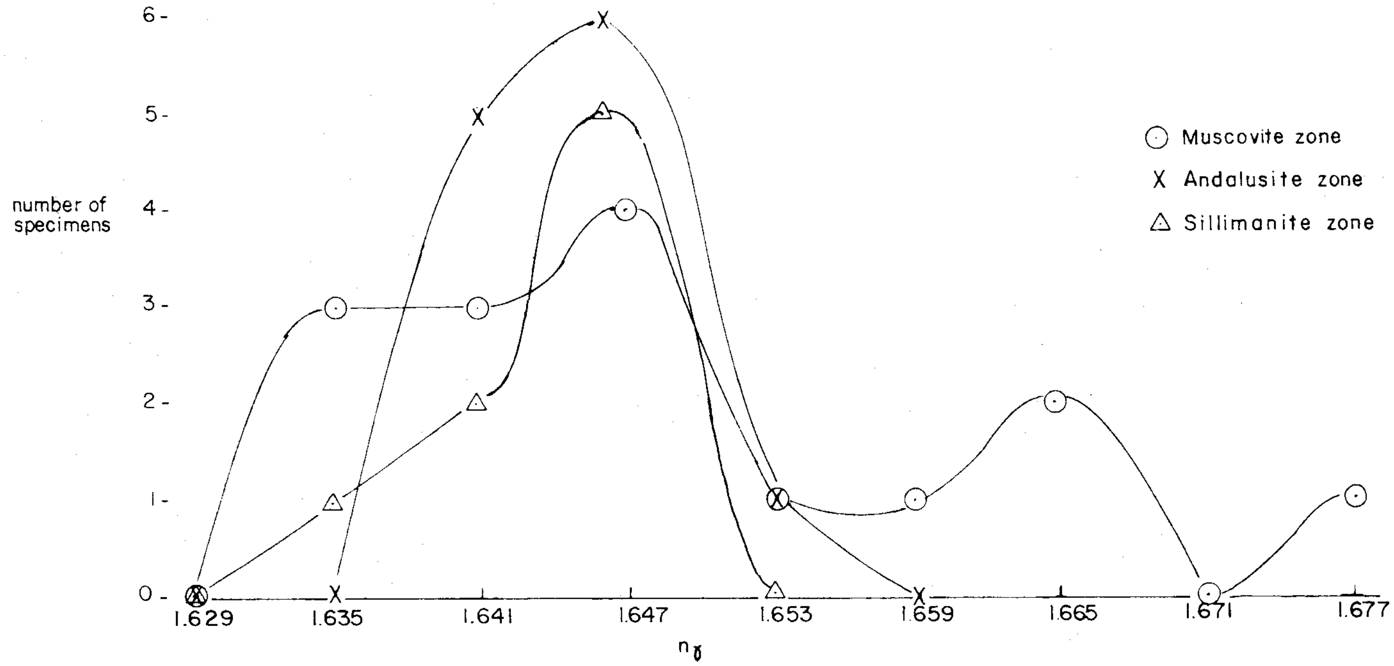


FIG.22 Refractive indices of biotites from pelitic and arenaceous rocks of the Winchester-Hemet area.

These are listed in order of inferred increasing rank of metamorphic grade. It appears that the refractive index n_z decreases with increasing temperature of formation. It was thought that this change might represent the occurrence in these samples of a solid-solution series between sillimanite and mullite as proposed by Kennedy (1955). X-ray patterns of two sillimanites were prepared in the range $2\theta(\text{CuK}\alpha) = 60-62^{\circ}$ as suggested by G. C. Kennedy (personal communication). These patterns were compared with similar patterns for sillimanite from Yorktown, N. Y., and mullite synthesized by the Vitrefax Corporation. The Yorktown sample has $n_z = 1.677$, agreeing perfectly with the values given by Winchell for sillimanite. The X-ray patterns of all three sillimanites in the limited range studied were indistinguishable, and markedly different from that of mullite.

As a further test of the possibility of mullite being present in the samples, a separate of the lowest-index sillimanite listed above (WH 338) was analysed by infra-red absorption analysis. The absorption pattern obtained agreed in all respects with that obtained on the Yorktown sillimanite and with the pattern shown by Roy and Francis (1953). It appears that mullite solid-solution is not the correct mechanism to explain the deviations in optical data. Winchell notes in his description of sillimanite that fibrous varieties have been observed to have lower indices than coarser crystals. It is of course possible that, as in the andalusite described above, some Fe^{3+} is entering into solid-solution in the sillimanite. But it would seem that, as in the case of andalusite, such a mixed crystal would have higher refractive indices than pure Al_2SiO_5. There does not seem to be any correlation in these samples between color and refractive index, the darkest sample being WH 153A. The color may be in part due to finely intergrown flakes of biotite.

3) Feldspar

Plagioclase of the pelitic schists is most commonly untwinned oligoclase in fine anhedral grains. It never constitutes more than 10 per cent of the schists. The composition of this mineral does not appear to vary appreciably from the muscovite zone to the highest garnet zone. The most sodic feldspar observed is An_{10} in the arkoses of the muscovite zone. An_{25} is found in the migmatites of the garnet zone, but in these rocks the composition may vary down to An_{20}.

Immediately adjacent to the Lakeview Mountain tonalite the plagioclase of the gneisses becomes somewhat more calcic, ranging from An_{25} to An_{40}. This may be due to the introduction of material from the tonalite. The plagioclase of the gneisses is never zoned. Sericitic alteration occurs in the middle and high-rank plagioclases.

Potash-feldspar in both the schists and arkosic quartzites is uniform in microscopic appearance throughout the Winchester-Hemet area. It is clear and untwinned and contains traces of very fine lamellar perthitic intergrowths parallel to (010). Microcline twinning is present in only a fraction of one per cent of the grains observed, disseminated about equally through all metamorphic grades. Study of the structural properties of this mineral discussed in Part II of this paper leads to the conclusion that in even the lowest rank rocks the potash-feldspar must be of monoclinic symmetry (sanidinized) in accordance with the optical observation that microcline twinning is absent. It is also observed in that study that the soda content of the potash-feldspar is almost constant with increasing degree of metamorphism from the upper muscovite zone to the lower garnet zone, as determined by unit-cell parameter studies. The soda content is about 5 mol per cent albite. Within the lithologic unit studied (fvB4) it is noted that the albite content of the plagioclase, as determined optically, varies sympathetically with the albite content of the potash-feldspar (fig. 37).

4) Cordierite

Cordierite occurs in coarse, round, untwinned, porphyroblasts containing abundant swirling streams of opaque inclusions, as well as dense clouds of very fine rounded quartz grains. The mean index of refraction is about 1.54; 2V(-) is large.

5) Chlorite

A faintly pleochroic brownish gray chlorite is found as nematoblasts in biotite schist on the west end of Searls Ridge in unit fvC7h. The index of refraction n_y is 1.614 ± 0.004; $2V(+) =$ about 5°; $n_z - n_x = 0.003$. These optical properties correspond to a chlorite with about 40 per cent Fe in the (Mg, Fe) position (Winchell and Winchell, 1951, p. 385). An X-ray pattern of this mineral shows a strong reflection at 7.14 Å corresponding to the 002 reflection. This gives a d(001) of 14.28 Å. This basal spacing corresponds to a composition of 1.0 Al atoms per four tetrahedral sites according to the data of Shirozu (1958).

6) Tourmaline

Tourmaline is present in amounts up to 2 per cent in the schists and quartzites. It is most commonly euhedral and in discrete stubby prisms. Its maximum absorption color varies from reddish brown to pale apple green (ϵ). The colors tend to become lighter with increasing metamorphic grade of the host rock. Dark brown to olive green colors prevail in the muscovite and andalusite zone; olive to apple green colors are common in the sillimanite and garnet zones. Tourmaline is almost absent from the migmatites, and most abundant in schists near the sillimanite isograd. Where it is most abundant it occurs as anhedral grains intergrown with biotite.

7) Zircon

Very fine euhedra of zircon are disseminated through all the schists and quartzites. The coarsest zircon grains are 0.05 mm. long and 0.02 mm. thick. They are uniformly disseminated through the schists although their association with pleochroic haloes in biotite makes them appear more abundant in that mineral. There is no evidence of obvious overgrowth over sedimentary cores; some well rounded grains are present in the lower rank quartzites.

8) Graphite

Fine disseminated flecks of graphite are present in variable amounts in all the schists. With increasing rank of metamorphism, especially in the migmatites, the grain size increases to about 0.1 mm. and the rectangular outline of subhedral plates oriented parallel to foliation becomes the common texture visible in thin section. A striking feature of the coarser-grained graphite is its tendency to grow attached to the (001) face of biotite and muscovite. This may be an example of so-called epitaxis, crystal growth controlled by localization on a crystallographic plane of similar configuration in both minerals.

9) Ores

Both magnetite and ilmenite are found in the schists. They can be recognized by the outline of their euhedral forms in some thin sections. Ilmenite is rare and occurs principally as rod-like inclusions in biotite. Magnetite is probably a common accessory in most of the biotite schists, concentrated in the more biotitic layers. It cannot generally be distinguished from graphite. Inspection of the thin sections in reflected light suggests that both graphite and ore minerals are present in most cases. Some of the latter may be ilmenite. In some sections streaks of fine octahedra can be found paralleling the biotite-rich layers. Hematite is not present except as a rare alteration product, as far as can be determined visually. Presumably the reaction

$$6\,Fe_2O_3 + C = 4(FeO \cdot Fe_2O_3) + CO_2 \qquad (12)$$

can proceed to the right at low temperatures and convert all the hematite or limonite of the primary sedimentary rock into magnetite. A calculation of the ΔF for this reaction at 1 bar and 300° C indicates that the right side is stable and remains so for increasing temperatures. The reaction

$$C + CO_2 = 2CO \qquad (13)$$

which proceeds to the right with increasing temperature would determine the $P(CO_2)/P(CO)$ for the graphitic sediments during metamorphism. Muan (1958, fig. 2) shows that at 1 bar pressure graphite could not be stable in equilibrium with hematite in a system in which the $P(CO_2)/P(CO)$ is controlled by reaction 11. However at increased pressures reaction 13 is driven to the left and the $P(CO_2)$ vs. T curve may intersect the hematite field at low temperatures, that is, hematite may be stable in equilibrium with graphite at moderate temperatures and high pressures.

Since hematite is rare or absent in all but the lowest-rank phyllites of this area, such conditions were apparently not attained.

No sulfide minerals are found in the schists at present. However it may be deduced from the inferred origin of the schists as black shales that appreciable amounts of ferrous sulfide may have been present in the parent sedimentary rock. The sulfides may have had a role in reducing the ferric iron present in the parent sediments; the sulfide ion $S^{=}$ would be oxidized to S^{o} and driven off as a gas during the early stages of metamorphism.

10) Other accessory minerals

Apatite is present as fine euhedral grains in some of the schists. Sphene is found as anhedral blebs intergrown with biotite and ilmenite. Rutile needles are enclosed in quartz and biotite in the highest-rank garnet-sillimanite gneisses.

f. Textural changes with progressive metamorphism

As the schists are progressively metamorphosed into sillimanite gneisses a characteristic inter-related sequence of textural changes occurs which eventually results, in the highest-rank gneisses, in almost total obliteration of schistosity. The sequence is as follows, illustrated by a series of photographs showing each textural type. For the purpose of this discussion we shall consider the micas and graphite as a single mineralogic type, designated by "mica".

1) In the lowest rank schists the micas are relatively coarse-grained in comparison with the tectosilicates. The tectosilicates form a fine matrix which flows between laminar agregates of mica. Mica-mica grain contacts are common (Plate

14a, appendix).

2) In the transition from the muscovite zone to the andalusite zone porphyroblasts of andalusite grow and distort the planar array of mica flakes. The overall grain size of the rock increases. Mica-mica contacts persist. The resultant texture is marked by distorted, swirling or arcuate layers of mica interleaved with discontinuous layers of finer-grained quartzose material (Plate 14b, appendix, showing distortion of a mica-rich layer in the vicinity of an andalusite crystal).

3) As grain size increases to produce fine- to medium-grained schists (andalusite and higher zones), the growth in grain size of tetectosilicates catches up with that of the micas. That is, the very fine matrix of quartz and feldspar recrystallizes into coarser individual grains. Mica-mica grain contacts become less common, partly due to the growth of tectosilicates forcing them apart, and presumably also due to the increased diffusion rates at higher temperatures progressively homogenizing the schist. The net effect is that mica grains are now most commonly interstitial to tectosilicates but arranged so that they are growing along a portion of a tectosilicate-tectosilicate boundary that is parallel to foliation (Plate 14c, appendix).

4) With further recrystallization, entering into the zone of veinitic migmatites, the planar (stress-induced) orientation is modified by plastic flow, evidenced by close-spaced ptygmatic folding of relict compositional bands. Growth of the tectosili-

cates has continued in this environment, but now the tendency of the tectosilicates to attain equant forms has overcome the tendency of the mica grains to orient themselves between the tectosilicate grains and parallel to schistosity. The orientation of the micas becomes more dispersed, first by cylindrical displacements along an axis lying in the foliation and finally by random displacements in the most highly recrystallized and plastically distorted gneiss. The final texture is granoblastic (Plate 14d, appendix).

3. Metamorphism of the Amphibolites

The amphibolites of both intrusive and extrusive igneous origin have undergone progressive metamorphism concurrently with the schists. Mineralogic zones roughly parallel to the zonation of the pelitic schists can be followed out in these rocks although their distribution is too sparse to permit constructing a separate set of isograds.

a. Muscovite Zone

The only amphibolite exposures in the muscovite zone are of the porphyritic meta-basalt flow in Member B of the French Valley formation. This flow extends from a zone of muscovite schists to a zone of andalusite schists with no great change in its mineralogy or texture. In its low-rank exposures west of the andalusite isograd it is composed of plagioclase, hornblende and minor amounts of magnetite, biotite, and a colorless (tremolitic) amphibole.

b. Andalusite Zone

Amphibolites of the andalusite zone are composed of hornblende, andesine and sphene. The traces of biotite present in the muscovite zone exposures are lacking. Sphene may be produced by the reaction of ilmenite with plagioclase though it is equally possible that there is a compositional difference between the rocks containing sphene and the lower grade amphibolites lacking this mineral. Amphibolites of this zone do not show relict igneous tectures.

c. Sillimanite Zone

In exposures of amphibolite in the sillimanite zone a colorless pyroxene appears as a major constituent. The abundance of pyroxene apparently increases with progressive metamorphism. However, due to the banded texture of the higher-rank amphibolites it is difficult to estimate the volume percentage of this phase for the entire metabasalt unit. Sphene diminishes in abundance in comparison with the andalusite zone, perhaps because TiO_2 is now entering into solid solution in the hornblende.

d. Garnet Zone

Beginning at the very edge of the garnet isograd for pelitic schists garnet appears as a constituent of the amphibolites. A single exception to this is garnet found in fragments of amphibolite occurring as float in A26(4, 2), presumably in the andalusite zone. As in the pelitic schists, garnet is only spottily distributed through these rocks. The highest rank exposures of amphibolite in section C29 are banded rocks containing layers of slightly pyroxenic amphibolite intercalated with coarsely crystalline garnet-pyroxene-plagioclase granulite.

Pyroxene continues to be an abundant phase in the garnet zone amphibolites, changing in composition somewhat (as estimated from optical properties) to an augitic and perhaps jadeitic diopside in the highest grade rocks.

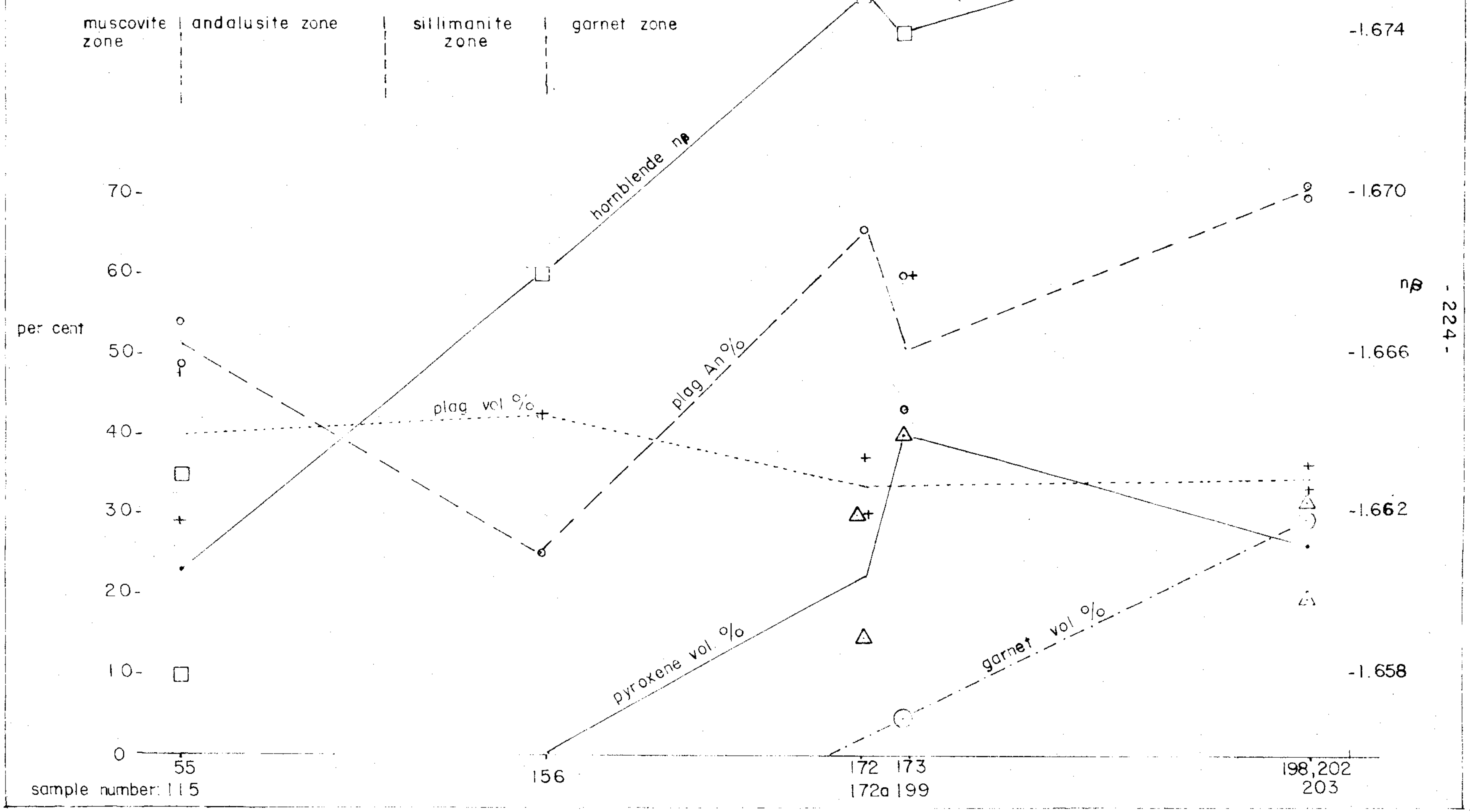
e. Over-all Trends of Progressive Metamorphism of the Amphibolite

The general trend of metamorphism is shown diagrammatically in fig. 23. The various specimens are plotted in order of increasing grade of metamorphism as estimated on the basis of their position with respect to the pelitic isograds. Specimens 172, 172A, 173 and 199 are all essentially of the same metamorphic grade, in the lower part of the garnet zone. The bulk composition of these various specimens is only approximately constant and variations of the various plotted quantities are to some degree due to changes in the chemistry of the systems in which they are measured.

The general trends observed are: 1) increase in the anorthite content of the plagioclase; 2) increase in abundance of pyroxene and garnet; and 3) increase in the β index of hornblende. The hornblende also changes somewhat in color, as indicated below. A similar progressive increase in the β index of hornblende is described by Wiseman (1934) in the progressively metamorphosed epidiorites of the Lewisian in Scotland. The change is correlated with an increase in the FeO/MgO ratio and also perhaps with an increased TiO_2 content.

The increase in the FeO/MgO ratio in the hornblende would be due to the expansion of the stability field and consequent increase in abundance of pyroxene, in which MgO is enriched relative to hornblende. The rise in the CaO/Na_2O ratio in feldspar is concurrent with a gradual decrease in the total amount of feldspar. Presumably while Na_2O is entering hornblende and pyroxene in increasing amounts with progressive

FIG.23 Progressive metamorphism of amphibolites in the Winchester-Hemet area.



metamorphism, CaO is also being used up, first to make pyroxene and finally to make garnet. Traces of rutile are present in the highest rank amphibolites, while sphene is absent, suggesting that CaO available for forming sphene is now deficient although the pyroxene and hornblende cannot dissolve more TiO_2.

The color changes in the hornblende can be generally described as follows:

Color of Max. Abs. Pale green, mottled with brownish-green → Brown and brownish-green → Bluish-green and olive

——increasing metamorphic rank →

The transition from pale green to brownish hornblende may be due to entry of TiO_2 into the lattice as suggested for biotites by Hall (1941a) and Engel (1953, p. 1070). The transition at higher ranks of metamorphism to bluish-green suggests that Na_2O is entering into solution, by analogy with the blue colors of soda-hornblendes. Hornblende in a soda-deficient biotite-hornblende schist from a locality nearby to the banded granulites of C29 is brownish rather than bluish-green.

The composition of the garnet in the amphibolites has been estimated using the index of refraction and unit-cell dimensions. Two samples have been studied, one near the garnet isograd in Crown Valley (WH 173) and the other from a granulite band in C28 (WH 202). They are plotted on fig. 21. Contrary to what might be expected (see for example, Miyashiro, 1953) the garnet from the lower-rank example appears to be of intermediate composition, a mixture of the ugrandite and pyralspite series, while the granulite garnet is a relatively pure member of the ugrandite series, close to the grossularite end. The latter garnet is extensively altered to epidote, preventing precise density determination.

Changes in the composition of pyroxene of the amphibolites have not been studied in detail. The properties of the lower-rank pyroxenes, as observed in thin section match those of pure diopside. A decrease in the extinction angle ($Z \wedge c$) from about 55° to about 39° occurs as one passes up-grade through the garnet zone. This is concurrent with a lowering of birefringence and the appearance of a faint greenish tinge. Pyroxene from the garnet-pyroxene granulite of C29 was indicated in Chapter III to have an approximate composition of $Ja_{20}Di_{45}Hd_{35}$.

4. Metamorphism of the Calc-Silicate Rocks

Calcareous quartzites occur as stratigraphic units in both the Bedford Canyon and French Valley formations. Such units can be traced from the muscovite zone into the garnet zone. Little variation in mineralogy is noted over this range. Estimated modal analyses of representative calcareous quartzites are given in table 5, grouped according to metamorphic grade as estimated from the adjacent pelitic schists. Fig. 24 shows the variation in the composition of the plagioclase and abundance of epidote in these samples. It is seen that the anorthite molecule increases in concentration toward the high-grade end of the metamorphic sequence. Epidote decreases in abundance towards the high-grade end. Epidote found in the garnet zone calcareous quartzites forms spindly anhedral intergrowths around plagioclase and quartz and appears to be of retrograde origin. Wollastonite appears in quartzites of the easternmost garnet zone.

TABLE 5. Estimated Modal Analyses of Calcareous Quartzites

	Gnt.	Calc.	Diop.	Hnbld.	Epid.	Plag.	An%	K-feld.	Wollas.	Scap.
Muscovite Zone										
3	tr	--	--	--	tr	--		--	--	--
19	tr	tr	10	5	--	tr	5(?)	10	--	--
290A	--	tr	--	--	15	tr	5(?)	--	--	--
310	--	tr	--	--	10	--		15	--	1
311	--	--	3	15	--	5	20	30	--	--
Andalusite Zone										
GQ	8	tr	3	--	--	20	30	--	--	--
Sillimanite Zone										
	NO EXAMPLES									
Garnet Zone										
157	1	7	2	tr	3	10	55	--	--	--
106	10	--	tr	10	1	65	65	--	--	--
258	40	--	2	--	tr	2	42	20	--	
176	--	1	7	tr	--	25	65	--	--	
239	30	--	15-25	--	--	20	65	--	15	--
287	25	--	10	--	--	--	--	--	15	

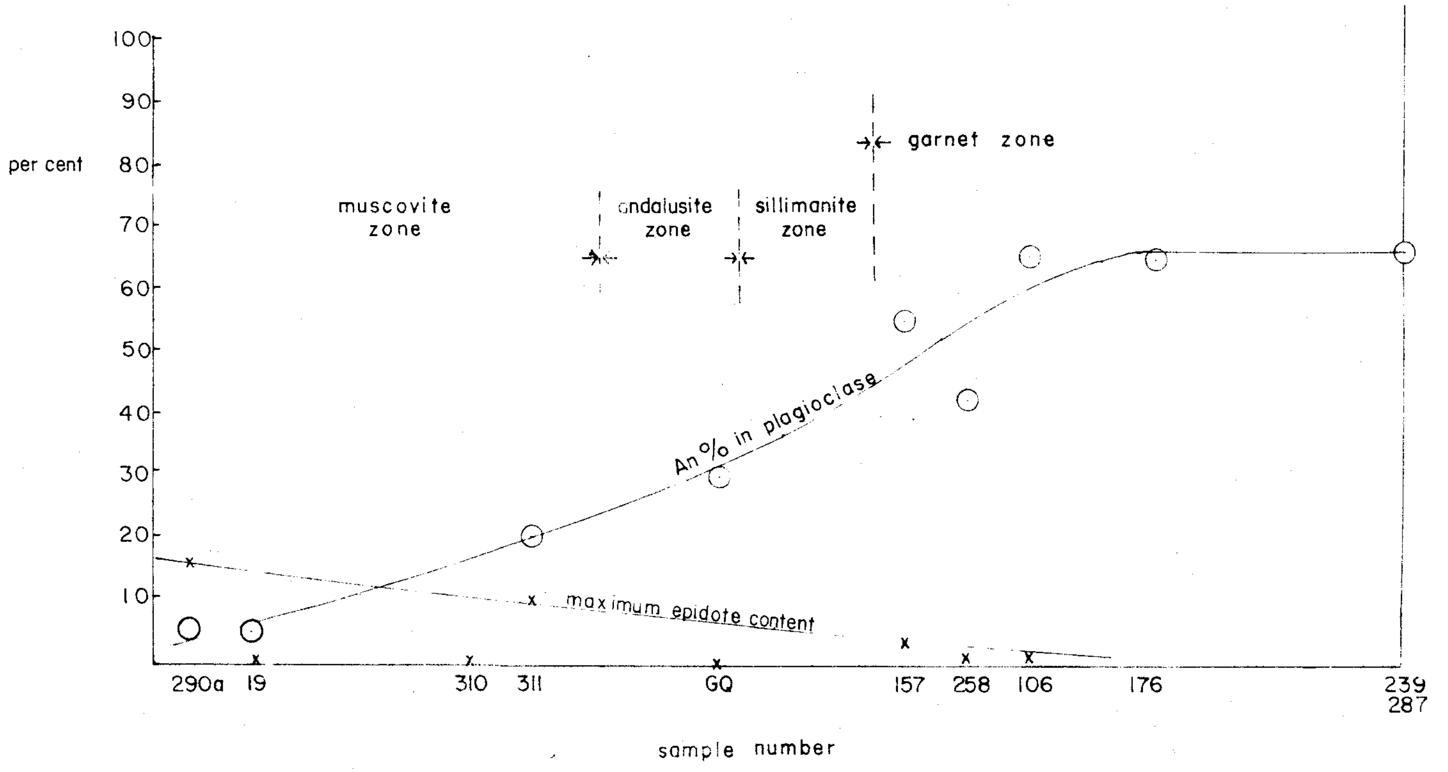
TABLE 5 (Continued). Estimated Modal Analyses of Calcareous Quartzites

Location of samples:

3	calcareous chert from Bedford Canyon formation
19	fvA2, A23(1, 3)
290A	Bedford Canyon formation, quartzite marker, A12(4, 3)
310	fvB4, A3(4, 3)
311	fvB4
GQ	fvC7h "cq" layer, A12(4, 3)
157	fvCs, B6(3, 3)
106	amq, A24(2, 3) (possibly a tuffaceous quartzite)
258	fvA2(?), A25(2, 4)
176	fvC 1-4 q, B7(4, 3)
239	"cq" unit in fvgn (undifferentiated), D14(4, 3)
287	"cq" unit in fvAgn (undifferentiated), B29(3, 4)

Analyses are given in volume percentage. Remainder is quartz.

FIG.24 Plagioclase and epidote in calcareous quartzites.



A garnet from the garnet zone (WH 204, section B5(2, 1), east of Rawson Canyon) has been determined by x-ray and optical methods. It is also plotted on fig. 21. It has the following properties: $n = 1.750 \pm 0.002$; $g = 3.53 \pm 0.03$; $a_o = 11.852 \pm 0.004$. Plots of g vs. n agree with the n vs. a_o plot in showing this garnet to fall close to the andradite-grossularite join with a composition of $Gr_{92}And_8$.

5. Metamorphism of the Ultrabasic Rocks of the Hemet Magnesite Deposit

The metamorphism of these ultrabasic and calc-silicate assemblages has been described extensively in Chapter III. It only remains to relate the period of this metamorphism to that of the enclosing metasedimentary rocks.

The ultrabasic intrusive lies in the garnet zone of metamorphism as defined in the pelitic schist. As has been noted, amphibolites of this zone may contain the assemblage hornblende-diopside-calcic andesine. In the meta-peridotite diopside has been completely replaced by tremolite, suggesting that at comparable conditions of metamorphism diopside should have been hydrated in the amphibolite too. However the presence of calcium-rich plagioclase coexistent with the diopside of the amphibolite may tend to counteract the tendency of the diopside to hydrate since in doing so the ratio of Ca:Mg must decrease from 1:1 to 2:5. Schematically:

$$\begin{array}{lcccccccc} & \text{CaO (in plagioclase)} & + & hb_a & = & \text{diopside} & + & hb_b & (14) \\ \text{Ca:Mg} = & & & (2{:}5) & & (1{:}1) & & (< 2{:}5\ (?)) & \end{array}$$

No such coupled reaction can occur in the ultrabasic rocks since the only lime-bearing phase present to begin with was presumably diopside.

Therefore the diopside of the ultrabasic rock could freely hydrate to form tremolite in the presence of sufficient SiO_2 (as described in Chapter III).

The other critical mineralogical guide to the rank of metamorphism of the ultrabasic rocks is the presence of a contact aureole of cordierite rock. Cordierite is apparently a stable phase in the biotite-sillimanite gneisses of appropriate composition in the garnet zone (in rocks with a high Mg:Fe ratio).

The metamorphic mineral assemblages found here in the ultrabasic rocks (cf. Chapter III, p. 114) have been observed in other geological localities in equilibrium with pelitic schists of the mineralogy of those found here. This fact is discussed by Turner (in Fyfe, Turner and Verhoogen, 1958, p. 210). All assemblages described in Chapter III (with the exception of those of the magnesite ore body) are found stably at this grade of metamorphism, the hornblende-hornfels facies as defined by Turner.

Although the mineralogy of the ultrabasic assemblages are such as could have been in equilibrium with the enclosing gneisses the textures of the meta-ultrabasic rocks do not exhibit the effects of stress concurrent with recrystallization such as are readily seen in the enclosing pelitic gneisses. Rather the textures are typically granoblastic; the individual phases are in subequant, commonly euhedral grains intergrown in random orientation. Even in the marble, which exhibits strong flow banding, the last stage of recrystallization evidently occurred after the conditions of shear-stress which caused the banding, for elongate and platy crystals are found in random orientation through the calcite matrix. Similarly the cordierite gneiss contact aureole is composed of granoblastic rock in which even the biotite lacks pronounced orientation, except in the relict bands representing prior concentrations of well-oriented flakes.

From the textural and mineralogical evidence it appears that the metamorphism of the ultrabasic and associated rocks occurred after the kinematic phase of metamorphism of the country rock but before temperature, confining pressure and P_{H_2O} had fallen below those values characteristic of the garnet zone as defined in the pelitic rocks. The temperatures recorded by the magnesium content of calcite from the calc-silicate marble are roughly in accord with (or slightly higher than) temperatures observed using the muscovite-paragonite geothermometer on pelitic assemblages of the same zone (using the extrapolated geothermal profile of fig. 25). If this conclusion as to the time of metamorphism is correct then the parent ultrabasic igneous rock must have been intruded some time after the kinematic phase of regional metamorphism but before the end of the epoch of elevated temperatures and pressures. Such a distinction between two parts of the progressive metamorphic cycle of this area cannot be recognized in any other metamorphic phenomena observed in this study.

Formation of the magnesite-serpentine-opaline-silica-(talc) assemblage of the magnesite ore body was almost certainly a much later phenomenon representing lower temperatures and confining pressures.

6. Geometric Relations of the Isograds

The isograds shown on fig. 2 are seen to trend in a north to northwest direction. They abut sharply against the major gabbro pluton of Bachelor Mountain and increase in grade away from the Domenigoni

Valley granodiorite pluton. They are however subconcordant to the contact between Lakeview Mountain tonalite and metasedimentary rocks. In all places where tonalite is seen in contact with pelitic metasedimentary rocks the metasedimentary rock is recrystallized to a coarse-grained migmatitic biotite-sillimanite (-garnet) gneiss. The contacts are generally concordant, but locally discordant up to 90°.

What appears to be an isolated elongate mass of andalusite-grade rock has been mapped to the west of the main andalusite zone. This isolated zone can be traced across strike in several schist units intercalated with quartzites which, because of their simpler mineralogy, do not exhibit the critical mineral assemblages of the andalusite zone. The significance of this isolated mass of higher-rank rock is not clear. It is presumably related to some local unevenness in the thermal (or P_{H_2O}) gradient at the time of metamorphism.

The isograds as mapped are of course only the traces of the intersections of surfaces in three dimensions with the earth's surface. There is no direct evidence of the dip of the complete surfaces but it is presumed that they dip steeply in concordance with the steep dip of the tonalite-metasedimentary rock contacts. They transect the folded metasediments at all angles up to 90° and are not obviously displaced by either folding or faulting of these rocks. The major fault zone of Rawson Canyon trends parallel to the isograds, however, and presumably also dips steeply so that only very great displacements along this break would be reflected in discontinuities in metamorphic grade across it. Near the mouth of Rawson Canyon into Diamond Valley such a discontinuity may be present, throwing rocks of the garnet zone on the east against slightly higher rank rocks of the same zone on the west.

The isograds are best determined in the exposures of the main metasedimentary block south of Domenigoni and Diamond Valleys and on Searls Ridge. On Double Butte the isograds trend parallel to the relict bedding but the progression from muscovite phyllitic schist to sillimanite migmatites is recognizable in units of comparable bulk composition progressively higher in the stratigraphic section. The isograds are evidently more closely spaced in this area than to the south, since the entire sequence from phyllitic schist to sillimanite gneiss can be walked out over a distance of one mile.

In general in this area metamorphic grade rises as one progresses higher in the stratigraphic section. Evidently one cannot use the concept of increasing superincumbent load of sediments to explain the sequence of isograds. The principal control over the geometric distribution of the isograds was apparently a source of heat lying to the northeast or east. In fact, since the isograds apparently post-date much of the folding recorded in the metasediments (though not all, as shown in Chapter V), it is likely that they are now in approximately the same orientation with respect to the earth's surface as they were at the time of metamorphism. This cannot be stated with complete certainty since it is impossible to estimate the amount of tilting that the entire crustal block seen here may have undergone prior to or following metamorphism. However the fact that dips in the metasedimentary rocks increase to a maximum of just 90 degrees in the most intensely folded rocks suggests that they stand at present in approximately the orientation in which the last period of folding left them. That is, it seems intuitively unlikely that by coincidence a later period of tilting should have brought these highly folded rocks into a uniform vertically-dipping orientation over a wide area. Near vertically-dipping metamorphic rocks seem typical of the highest grade exposures of many metamorphic terranes throughout the world.

If the present orientation of the metamorphic rocks is indeed representative of their orientation at the end of the period of metamorphism then the sequence of metamorphic assemblages exposed is in effect an isobaric section of the complete metamorphic zonal distribution, neglecting minor variations in elevation due to the present erosional topography. Only one determination of pressure from some critical metamorphic reaction should suffice to determine the pressure of metamorphism for the whole area.

C. PHYSICAL CONDITIONS OF METAMORPHISM

By reference to experimental work on systems of comparable composition it is in theory possible to estimate the temperatures, total pressures, P_{H_2O}, P_{CO_2} of metamorphism in this area. In the following discussion it will be assumed in all cases that $P_{tot} = P_{H_2O}$. This assumption is not independently verified. The calc-silicate quartzites on the other hand were probably not metamorphosed at $P_{CO_2} = P_{tot}$. Locally carbonate is abundant enough to have raised the P_{CO_2} to a considerable fraction of the total pressure but in general the sediments were apparently totally decarbonated well below the lowest rank of metamorphism exposed here.

1. Temperatures of Metamorphism

Of the commonly applied "geothermometers" suggested for use in determining temperatures of recrystallization only two possible methods are readily available here. They are: 1) the Na content of muscovite in equilibrium with albite; and 2) the Mg content of calcite in equilibrium with dolomite. The wollastonite-quartz-calcite equilibrium may similarly be used to determine the pressure of the isobaric section here exposed, as discussed below.

The sodium content of a 2-layer monoclinic muscovite may be determined from the spacing of the 001 planes of the lattice, assuming that all other chemical variables are held constant. That this is a poor assumption is made evident by attempts to reproduce chemical analyses made by wet chemical methods with measurements made by this x-ray technique. Results of this attempt are shown in table 6. Granting that such errors may be minimal in the present investigation, there is still great uncertainty as to the significance of the Na content of muscovites taken from rocks not containing paragonite, since there is no experimental evidence for such a system. In all cases referred to here the muscovite was coexistent with oligoclase or calcic albite.

Temperatures obtained by this technique will probably be a minimum estimate in all cases. The effect of pressure on this temperature estimate is not known. Since the substitution of the smaller Na^+ ion for K^+ in the mica structure lowers the molar volume of the structure it is likely that increased pressure will favor the substitution. Consequently the true temperature of crystallization would be less than that inferred

TABLE 6. Muscovite Analyses

A. Comparison with chemical analyses: $\frac{\text{Na(mols)}}{\text{Na(mols)} + \text{K(mols)}}$

Sample	X-ray	Chemical
PEM-2	0.041	0.071
DV	0.149	0.149

Chemical analyses by Dorothy Settle, using flame photometer.

B. Muscovite analyses from the Winchester-Hemet area

	Δ	Na/Na+K	T degrees C
WH 11	0.00	0.000	--
WH 68	0.07	0.056	340 ± 20
WH 2	0.04	0.032	260 ± 35
WH 63	0.03	0.024	220 ± 40
WH 307	0.05	0.040	290 ± 35
WH 147	0.054	0.043	300 ± 35
WH 12-7s	0.13	0.103	460 ± 15

$\Delta = [\,2\theta(006) - 26.60]$ degrees (CuKα)

$2\theta = 26.60^\circ$ for pure muscovite

$2\theta = 27.86^\circ$ for pure paragonite

$$\frac{\text{Na}}{\text{Na} + \text{K}} = \frac{\Delta}{27.86 - 26.60} = \frac{\Delta}{1.26}$$

Location of specimens:

WH 11	Schist just above base of French Valley formation, fvA1 in A22(3, 3)
WH 68	From fvC5 in A2(1, 4)
WH 2	From Bedford Canyon formation in A22(2, 2)
WH 63	From fvC1 in B12(3, 2)
WH 307	From fvC5 in B18(1, 4)
WH 147	Schist lens in fvB4, adjacent to specimen 4-6 (Part II)
WH 12-7s	From fvB4, adjacent to specimen 4-7 (Part II)

Specimen numbers "4-" referred to above are those in the sample sequence described in Part II of this thesis.

from the solvus given by Eugster and Yoder (1954) if pressures were greater than 2 kilobars (the pressure at which the solvus was determined) at the time of crystallization. An additional effect which cannot be readily taken into account and which is probably large is the degree of substitution of F and Cl for OH^- in the muscovite structure.

X-ray parameters were determined by counting over the peaks of the 004 and 003 reflections with a Norelco x-ray goniometer, determining 2θ to $\pm$ 0.01°. The values for these higher-order reflections were recalculated to the 001 spacing and then converted to mole-percent paragonite using the assumption of linear variation in the basal spacing with composition. This assumption has been experimentally verified by Eugster and Yoder (1954). Temperatures are taken directly from the solvus for paragonite-muscovite, as determined by Eugster and Yoder. The samples analysed span a range of metamorphic grade from the muscovite zone to the lowermost garnet zone. Muscovite from higher-rank rocks appears to be partly or wholly retrograde and has not been analysed by this procedure. The temperatures suggested by these analyses range from 200° for the muscovite zone to 450° for the lowermost garnet zone. They are plotted against estimated position in the metamorphic grade sequence on fig. 25 (described below).

Calcite from the magnesian marble at the magnesite mine on Searls Ridge has been studied by x-ray diffraction methods as discussed in Chapter III. The temperatures obtained from two samples from the locality were 570° and 620° C. There is serious question whether these values, even if correct, represent temperatures of regional metamorphism in the vicinity of the carbonatite body. Structural and textural

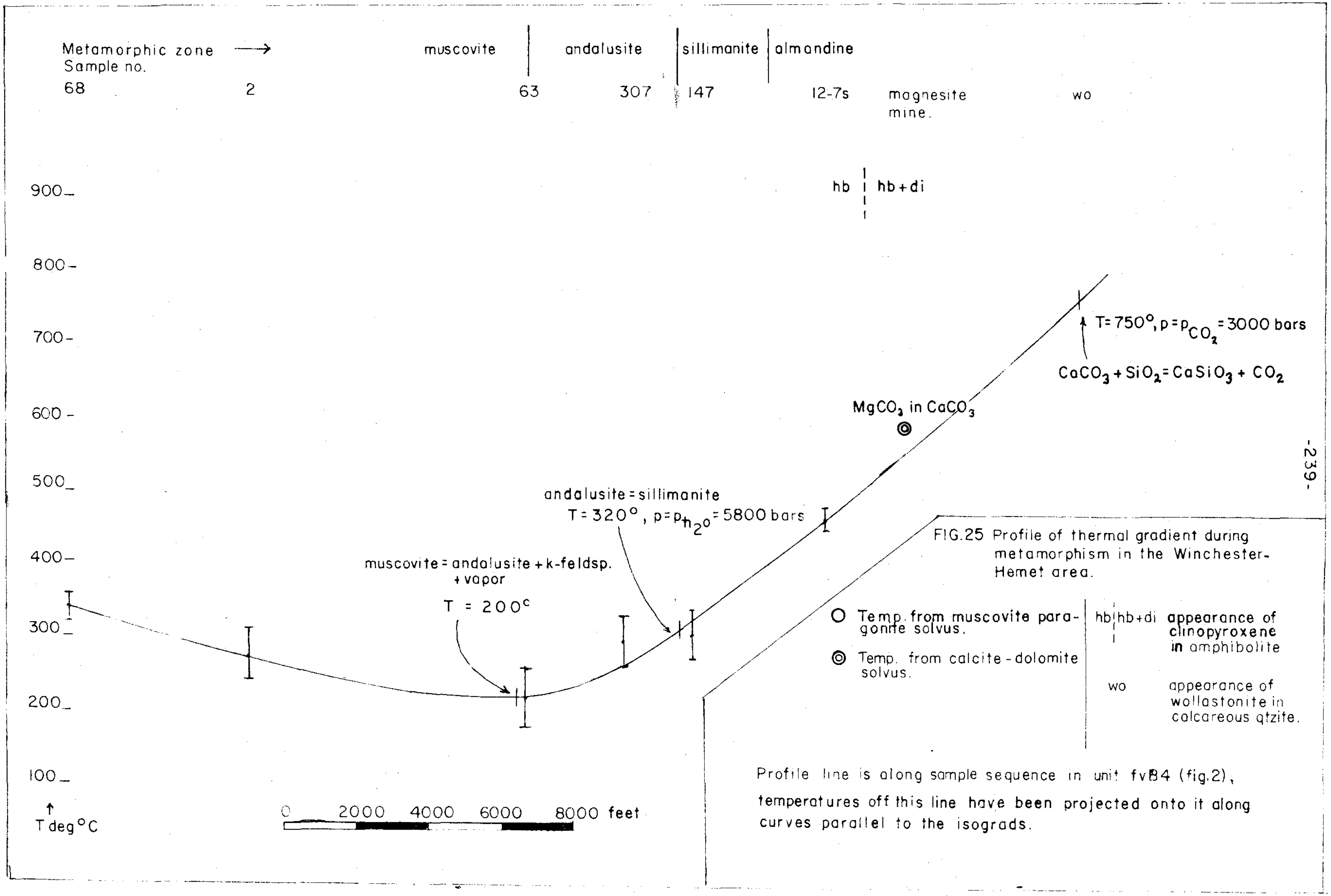


FIG.25 Profile of thermal gradient during metamorphism in the Winchester-Hemet area.

Profile line is along sample sequence in unit fvB4 (fig.2), temperatures off this line have been projected onto it along curves parallel to the isograds.

evidence discussed earlier indicates that the ultrabasic rocks of this locality may have been intruded and metamorphosed after the metamorphism of the country rock. If so, then higher temperatures of the adjacent carbonatite may represent temperatures of contact metamorphism by the intrusive peridotite magma. They may even represent temperatures of crystallization of a primary carbonatite magma. The mine lies about one mile into the garnet zone and a temperature of about 600° would not be an unreasonable extrapolation of the gradient observed up to the garnet zone boundary, as determined by the muscovite geothermometer.

Additional evidence as to the temperature of metamorphism might be obtained from the Na content of potash-feldspar in equilibrium with albite, or from the TiO_2 content of magnetite in equilibrium with ilmenite. These two methods as applied to this area are discussed in Part II of this thesis.

2. The Pressure-Temperature Field of Metamorphism as Inferred From the Data of Hydrothermal Synthesis

The available mineralogical data on this area may be summarized as follows:

1. At the andalusite isograd andalusite forms, probably at the expense of muscovite (equation 4, this chapter).

2. Within the andalusite zone and into the sillimanite zone, cordierite may form at the expense of biotite:

$$\text{biotite} + \begin{matrix}\text{andalusite}\\ \text{sillimanite}\end{matrix} = \text{cordierite} + \text{K-feldspar} + H_2O \qquad (15)$$

3. At the sillimanite isograd biotite breaks down to form sillimanite + Fe-ores + potash-feldspar + more magnesian biotite.

4. At the garnet isograd:

(MnO)? + biotite + sillimanite = garnet + K-feldspar + H_2O (11)

Manganese may be derived in part from adjacent pegmatites.

5. Above the garnet isograd clinopyroxene appears in amphibolites

hornblende = diopside + ferroan hornblende (16)

6. Above the garnet isograd calcite and quartz react to form wollastonite.

7. At or above the garnet isograd, garnet forms in the amphibolites.

8. Kyanite is absent.

Utilizing the temperature data obtained from the muscovite-paragonite solvus we may obtain a tentative profile of the thermal gradient in the Winchester-Hemet area during metamorphism (fig. 25). All but one of the measured samples lies along a line roughly normal to the isograds. The single sample off this line, WH 2, has been projected onto it along an arc parallel to the adjacent andalusite isograd. From this profile estimates of the temperatures of each progressive reaction can be made as follows:

Step 1)	$200^\circ \pm 30^\circ$
Step 3)	320°
Step 4)	380°
Step 5)	400°
Step 6)	750°
Step 7)	500°

The temperatures above 450° were obtained from a linear extrapolation of the lower part of the profile. The $MgCO_3-CaCO_3$ geothermometer temperatures are plotted on this profile too, also projected along a curve parallel to the isograds. Their projection falls about 50° higher than the profile.

Using this temperature gradient and considering the $P_{CO_2}-T$ phase diagram for the reaction calcite + quartz = wollastonite (Danielsson, 1950) we may deduce a minimum pressure of metamorphism for this area of 3000 bars. This assumes that $P_{CO_2} = P_{tot}$ whereas in fact P_{CO_2} was considerably less than total pressure for the various reasons described by Danielsson and by Thompson (1955, p. 96). In this case total pressure as inferred from the phase diagram would be too low. However this is the best estimate of the pressure of metamorphism available to us, using the temperature gradient we have obtained. The absence of kyanite puts an upper limit on the pressure of 15,000 bars, from the P-T curve of the transformation sillimanite = kyanite as determined by Clark, Robertson and Birch (1957).

Experimental work relative to the reactions listed above is meager and in general deals with systems of more restricted chemical composition than the typical schist or amphibolite.

In regard to the occurrence of cordierite, investigations by both Yoder (1952) and Roy and Roy (1957) on the system $MgO-Al_2O_3-SiO_2-H_2O$ indicate that at pressures of 10,000 to 15,000 psi cordierite is stable at temperatures above 550° in its compositional field. The presence of iron will probably lower the equilibrium temperature (see Ramberg (1952), fig. 62). Winkler (1957) investigating the thermal metamorphism of lime-free clay at 2000 bars P_{H_2O} reports that the reaction biotite + musc.+ quartz = cord + sill + K-feldspar + V (17)

takes place at 650°C. In these same experiments Winkler finds partial melting to begin at 720°C. Cordierite in the Winchester-Hemet area appears to form at temperatures as low as 300°C. (in the andalusite zone). The composition of the clay mineral assemblage used by Winkler ("Ton B") is probably similar in composition to the clay-component of the pelitic schists of the French Valley formation. Using an estimated modal analysis of a typical andalusite schist from the French Valley formation a partial chemical analysis has been calculated and is compared below with the analysis of "Ton B" given by Winkler (p. 58):

WH 57 (fvC6, A12(4,3))				"Ton B"
mode (vol. %)		est. comp.		
quartz	30	SiO_2	57.8	60.7
biotite	35	Al_2O_3	23.6	23.8
muscovite	15	Fe_2O_{3t}	7.6	5.6
andalusite	15	MgO	3.6	2.2
cordierite	5	Na_2O	0.1	0.6
		K_2O	4.8	5.2
		TiO_2	1.2	0.8
		total	98.7	98.9

It is seen that "Ton B" has a considerably higher Mg:Fe ratio than the estimated composition of the pelitic schist. The difference is probably within the error of the estimated analysis of the schist. In other respects the two compositions are quite similar, and it would appear that they would behave similarly under hydrothermal metamorphism.

The stability of garnets has been studied in systems of the composition of a garnet by various workers but no work on equilibrium relations between garnet and hydrous minerals has as yet been described in the literature. Yoder has found that almandine is probably stable at atmospheric pressure and temperatures of about 800°. The theoretically obtained slope of the P-T curve for the breakdown of pyrope to cordierite + spinel + forsterite is approximately the same as the experimentally determined slope of the reaction almandine = fe-cordierite + spinel + fayalite. The pyrope P-T curve is about 25 kilobars higher than that for almandine. Evidently addition of Mg to the garnet structure decreases its low-pressure field. Spessartite has been reported by Pettijohn (1949) to be an authigenic mineral in sediments so that its breakdown curve certainly has a zero-pressure intercept at moderate temperatures, in equilibrium with hydrous minerals. Field evidence on the variation of composition of pyralspite garnets (Miyashiro, 1953; Ramberg, 1952, p. 63) suggests that MnO definitely widens the temperature stability range of garnets in natural systems. The reaction for the production of garnet given in step 4 above presumably has the same type of P-T curve as that of the wollastonite reaction.

In regard to the appearance of diopside in amphibolites, the work of Tilley and Yoder on the hydrothermal metamorphism of basalt glasses may shed some light, when complete, on the sub-solidus field of diopside. Engel (pers. comm.) reports that at temperatures of 550° as inferred from various geothermometers, clinopyroxene appears in amphibolites of the Grenville series in New York. In the Winchester-Hemet area diopside first appears at inferred temperatures of about 400°. Pressures of metamorphism in the Grenville province are inferred to have been over 3500 bars so that the reaction forming diopside from hornblende would be expected to occur at a somewhat higher temperature, in the Grenville amphibolites.

The reactions in the calcareous quartzites have not as yet been observed in the laboratory. Epidote has not been synthesized. Grossularite may be formed by the reaction of anorthite with calcite:

$$CaAl_2Si_2O_8 + CaCO_3 + SiO_2 = Ca_3Al_2Si_3O_{12} + 2CO_2 \tag{18}$$

and therefore would have a P-T curve similar to that for wollastonite. Anorthite is probably stable down to 400^o at 1 kilobar P_{H_2O} with respect to the formation of epidote. More albitic feldspars are known to grow authigenically in sediments.

Finally, the appearance of various polymorphs of Al_2SiO_5 and hydrous equivalents of this composition may be considered in the light of experimental work. First it should be remembered that the aluminum silicate minerals found here are probably not pure phases (see discussion earlier in this chapter). Assuming however, for want of analytical data, that these minerals are pure Al_2SiO_5, the significance of the appearance of these minerals with respect to phase diagrams for Al_2SiO_5 is still in doubt. The reactions forming andalusite and sillimanite are apparently by and large dehydration reactions acting on muscovite and biotite. At the sillimanite isograd it appears that most of the andalusite has been replaced by muscovite (though this is probably a retrograde phenomenon). Progressive development of sillimanite is by decomposition of muscovite rather than by polymorphic inversion of andalusite. The characteristic habit of sillimanite is as ellipsoidal bundles of fibrolite enclosed in biotitic streaks and bands. Pseudomorphous sillimanite after andalusite is not observed. This appears to be a generally valid observation in metamorphosed pelitic schists, as discussed by Tozer (1955). The only experimentally determined dehydration reaction forming aluminum silicate is the reaction pyrophyllite = sillimanite + vapor (or mullite + quartz + vapor). This reaction has been studied by Kennedy (pers. comm., 1959). The breakdown curve for pyrophyllite may be analogized to similar but unknown curves for muscovite and biotite although the degree to which hydrous phases obliterate the anhydrous aluminum silicate field in systems of complex composition is clearly dependent on a multitude of factors which cannot be evaluated here.

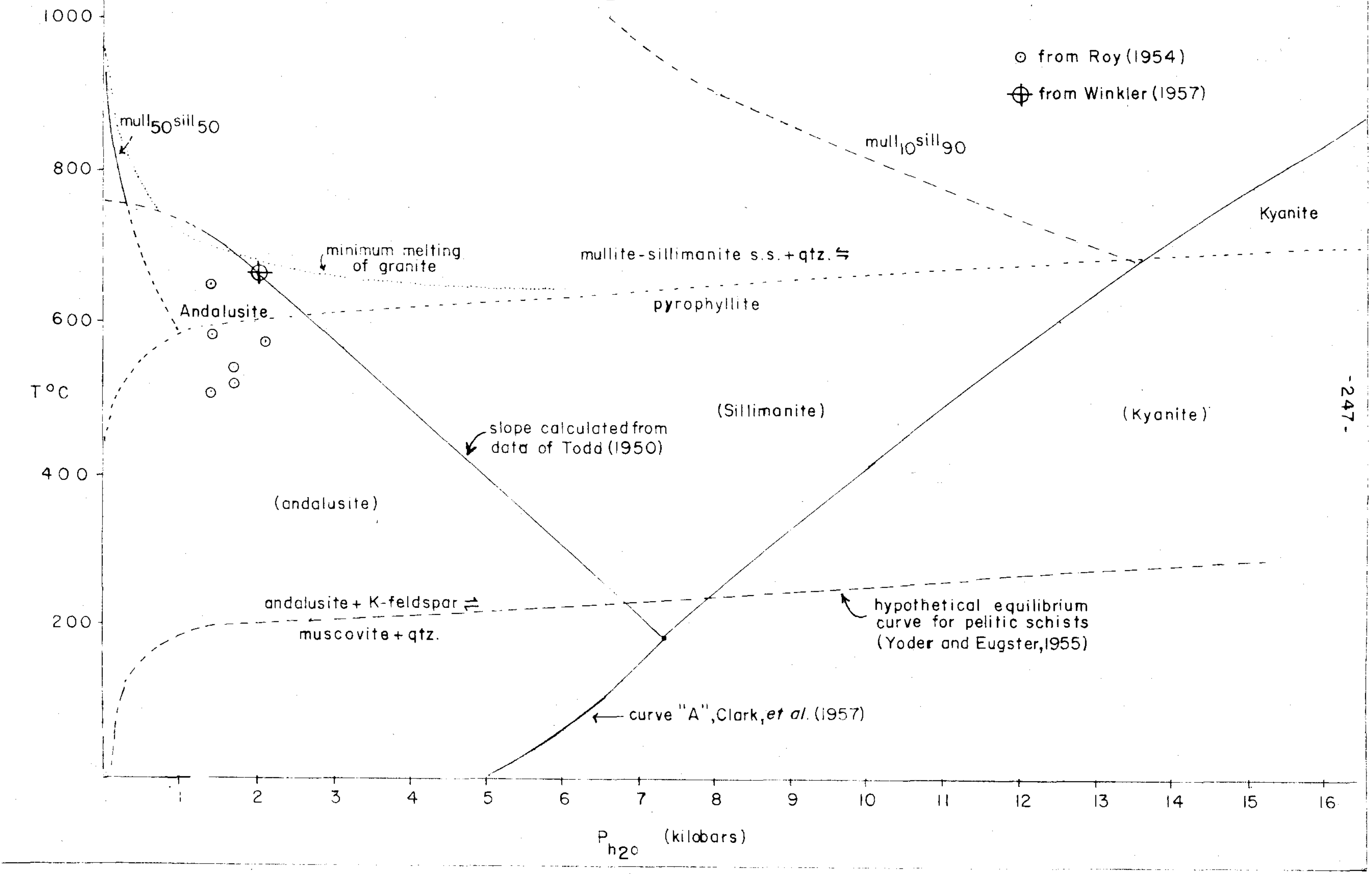
A compilation of the experimental evidence on the Al_2SiO_5 P-T diagram is given in fig. 26. P_{H_2O} is equal to total pressure in all experiments. The points at which andalusite has been synthesized by Roy (1954) and Winkler (1957) are shown. They fall within the pyrophyllite field of Kennedy's data, a fact which may be attributable to the difference in bulk composition of the charges used in the various experiments. The curve andalusite $\rightleftharpoons$ sillimanite and the triple point andalusite-sillimanite-kyanite have been obtained by plotting a straight line with the slope equal to the calculated dP/dT for the reaction and. $\rightleftharpoons$ sill. using the thermodynamic data of Todd (1950):

$$dP/dT = \frac{\Delta S}{\Delta V} = \frac{22.97 - 22.28}{49.0 - 51.6} = -\,0.27\,\frac{\text{cal}}{\text{cm}^3\text{deg. C}} \qquad (19)$$

$$= -\,11.1\,\frac{\text{bars}}{\text{deg. C}}$$

The line has been drawn to pass through the highest temperature at which andalusite has been synthesized and the triple point is obtained at the intersection of this line with the "best" curve for the sill. kyan. reaction as

FIG.26 Hypothetical P_{h_2o}-T diagram for Al_2SiO_5 and muscovite.



given by Clark, Robertson and Birch (1957). The fields of pyrophyllite, mullite and mullite-sillimanite solid solutions are as described by Kennedy (1955 and pers. comm., 1959). There is apparent disagreement between the data of Roy and Winkler and that of Kennedy. Since andalusite was not observed at all by Kennedy it is not possible to compare their results in detail. The field of andalusite may be much smaller than indicated and may only exist in nature in impure systems where the field of pyrophyllite is replaced by a similar but smaller field of muscovite or other hydrous Al-silicates.

Again referring to the inferred temperature gradient for the Winchester-Hemet area we see that andalusite appears to form at a temperature of 200° and sillimanite at a temperature of 320°. The latter temperature may be considered as the temperature of the inversion andalusite = sillimanite. The pressure corresponding to this temperature, as taken from fig. 25 is 5800 bars, 2800 bars higher than that from the wollastonite curve. This should be considered an upper limit to the pressure since the temperature estimate (from the muscovite geothermometer) is a lower limit. From the data of Kennedy it would appear that at temperatures above about 600° sillimanite-mullite solid solutions would be stable at 3000 bars pressure. However, in the presence of excess silica the reaction

$$(2Al_2O_3 \cdot SiO_2) + SiO_2 = 2(Al_2SiO_5) \tag{20}$$

will be displaced to the right. Experimental data from systems with limited silica content are not strictly applicable here. However the P-T curve for the solid solution $sill_{90}$-$mull_{10}$ is at such a high pressure at 650° that it seems likely that some degree of solid solution between these two minerals should be noted. As described earlier in this chapter this likelihood is not confirmed by analyses of sillimanites from this area.

The curve of minimum melting points of a granite (Bowen and Tuttle, 1953) is shown on fig. 26. It is evident that the granitic components of the gneisses could easily be partially melted at temperatures and pressures encountered in the highest rank zones of this area, even using minimum temperatures. In fact it is observed that this zone is an area of veinitic migmatization and the form of the melting curve may be taken as confirmation from experimental evidence of the textural evidence for local derivation of the migmatizing material. As would be expected from the position of this curve, the zone of migmatization is encountered above the sillimanite isograd but below the point of wollastonite formation.

It is also seen from this diagram that for temperatures above 200° the upper limit of pressures of metamorphism is given by the absence of kyanite to be 7 kilobars. Schuiling (1958) has suggested that the kyanite-sillimanite boundary may actually curve towards lower pressures at lower temperatures rather than conforming to the thermodynamically calculated extrapolation of Clark, Robertson and Birch. His conclusions are based on somewhat dubious "geo-experimental" P-T computations (see Clark, Robertson and Birch (1958) for a discussion of his results).

From the provisional stability relations of the aluminum silicates as shown in fig. 26 it can be seen that above pressures of about 7 kilobars andalusite is not stable at all and at pressures below that value the temperature range of its stability field is small compared to that of sillimanite or muscovite. This is nicely borne out in the distribution of andalusite-bearing rocks in the Winchester-Hemet area. The width of the andalusite zone (fig. 2) is small compared to the exposed width of the muscovite zone or the combined sillimanite and garnet zones. On the basis of the geothermal profile deduced above (fig. 25) andalusite is stable over a temperature interval about 100 degrees wide.

D. METAMORPHIC DIFFERENTIATION

It has been found possible in many places in this area to trace apparently homogeneous or finely banded schist into composite gneiss consisting of regularly spaced varve-like layers of feldspathic quartzite and thinner bands of biotite-sillimanite- (or andalusite-) muscovite schist. The transition between banded and homogeneous rock is invariably gradational. The incipient stages of this transition are marked by the appearance of slightly darker bands containing somewhat more biotite than the bulk of the rock, standing out as darker lines on the weathered outcrops. As the process advances the thickness of the bands remains constant as the contrast in color becomes progressively greater. It appears that this banding is most readily developed in rocks whose primary sedimentary structure is one of finely interlaminated quartzose and phyllitic layers, though several localities have been found, especially on Searls Ridge in section D36, where completely homogeneous granulitic biotite-quartz-feldspar-andalusite rock grades into intensely banded quartzite-schist composite gneiss.

Similarly one finds in the amphibolites that homogeneous granoblastic plagioclase-hornblende-diopside amphibolite can be traced along strike into banded rock of about the same bulk composition. In this case the darker bands are composed of hornblende and plagioclase, the lighter bands of plagioclase with lesser amounts of diopside. The segregation is commonly so well developed that the rock has the textural and compositional makeup of interlayered anorthosite and hornblendite. Banding in the amphibolites is usually coexistent with banding in the enclosing pelitic metasedimentary rocks.

It has not been possible to determine in detail whether the bulk composition of rocks of this kind is identical with that of the homogeneous units lying along strike from them. Rough visual estimates do not suggest major differences in composition. However the banding itself prevents making an analysis of mineral volume abundances of the whole rock since the bands are spaced farther apart than a single thin section. Chemical analyses of bulk samples of the rocks would shed light on this problem. An example of the compositional differentiation between the bands is shown in the following estimated modal analyses:

	WH 82 Banded gneiss			WH 76b
	schist	quartzite	4 quartzite + 1 schist	
biotite	70	5	19	12
muscovite	10	0	2	0
sillimanite	20	tr	4	13
K-feldspar	1	5-40	17	15
quartz	1	50-85	58	60
plagioclase	0	1	1	0

Estimating from outcrop appearance that the ratio of light to dark bands is 4:1 the average composition would be as given in the third column. For comparison the estimated mode of a homogeneous schist collected about 1/4 mile west of WH 82 is also given (WH 76b). The two rocks are probably not exact stratigraphic equivalents but are less than 500 feet apart in the section.

Similar banding has been attributed by other authors (Eskola (1932), Turner (1941)) to a process of metamorphic differentiation. That is, under conditions of metamorphism, especially under shear stress, conditions for minimizing the free energy of the system are such that certain minerals, commonly of like physical properties, aggregate together and form layers in otherwise homogeneous rock. This is comparable to the tendency to "concretion" of phases in granoblastic rocks (as described by Eskola) by nucleation and growth from material derived from the surrounding homogeneous rock (e.g., the growth of giant garnet porphyroblasts). Under hydrostatic stress and isothermal conditions of metamorphism it seems as if, starting with a homogeneous rock, one would expect to form a more coarsely crystalline, but still essentially homogeneous rock in which there had been concretion of each phase to a set of stable nuclei disseminated through the rock. If the rock were initially finely banded, it is similarly conceivable that some recrystallization by simple concretion of material to the bands could take place, especially if the bands were composed initially of coarser-grained mica than the disseminated flakes in the quartzose layers. But, by simple concretion an initially homogeneous rock could not be transformed into a banded rock unless some additional constraint were applied to force potential nuclei for concretion to be localized in equally spaced layers through the rock.

It is here proposed that the shear stress prevailing during metamorphism supplies this additional constraint.

The homogeneous schists from which these banded rocks were formed were themselves derived from finely banded phyllitic schists which have been homogenized with progressive metamorphism, due to the action of diffusion of the mineral components through the rock. This has already been described in the section on stratigraphy in this report. When subjected to shear stress the relict planar features of sedimentation act as planes of weakness along which displacements may occur. For a given rock adding planes of weakness along which displacements may occur will reduce the amount of elastic energy stored in the rock while under stress and therefore would be thermodynamically favored. Such planar features could therefore evolve spontaneously. Bands of micaceous rock in which slip of mica flakes across one another could occur would be favored over a homogeneous rock in which mica was isolated between equant grains of quartz. From the textural evidence of the progressive growth of banded features it seems that at first equally spaced zones occur in which mica is more uniformly oriented than in the bulk of the rock. These primary bands are perhaps comparable to the equally spaced slip-bands in metals which form under compression, along planes of maximum shear stress. From this point onward growth of the bands may be entirely due to recrystallization of mica onto nuclei which possess a thermodynamically favored orientation with respect to the stress field. A detailed discussion of the thermodynamics of stressed solids in relation to this problem is not possible at this time.

The field evidence as to the distribution of these banded rocks suggests that they are localized to zones of tight folding, especially to parasitic folds on the flanks of the major folds of this area. Where the banded gneiss is tightly folded the banding continues with no apparent change in thickness around the noses of these folds. If the above arguments as to the role of shear stress in producing banding are at all valid then this banding must have evolved before the period of tight folding, for the planes of maximum shear stress in these rocks during folding would cut sharply across the present banded structures. In fact, the mica within these bands is well oriented parallel to the axial planes of these folds, suggesting that the later folding has reoriented the mica from its initial equilibrium orientation parallel to the banding. Crinkling and sillimanite orientation within these bands parallel the fold axes.

However, if the role of stress in producing the banding is admitted it would seem that some banding transverse to relict bedding should be formed parallel to the axial planes of the folds. In fact, as has been noted (Chapter IV) there is a marked tendency for boudin-like structures to form near the crests of these folds, isolating cylindrical masses of quartzite elongate parallel to the fold axes. The septa between these boudins are planar strips of schist parallel to the axial plane and are in effect bands developed by later shearing along these planes, by the same mechanism as the main bands. Because most of the mica was already accumulated into layers it was not possible for the mica to redistribute itself completely in accordance with the new stress distribution imposed during folding.

It must be noted that most of the above features could be equally well explained by local sedimentary facies changes in the primary sediment. The varve-like character of the banded gneiss may indeed be derived from varved sedimentary deposits of shale and sandstone. No evidence has been found that strictly eliminates this possibility. The spatial distribution of the banded facies, cutting across stratigraphic units as it does, is at best a strong suggestion that this fabric is a superimposed feature of metamorphism. Even the frequent local gradations from banded into homogeneous schists may represent lateral variations from quiet water cyclical sedimentation to more turbulent depositional environments. However the recurrence of such banded features in metamorphic terranes throughout the world and the rarity of comparable sedimentary features in unmetamorphosed pelitic sediments of similar composition would argue for a definitely metamorphic origin.

E. DISEQUILIBRIUM

Mineral assemblages suggesting disequilibrium are found in most pelitic schist samples of andalusite rank or higher. The most striking example of this phenomenon is the common coexistence of sillimanite and andalusite in the schists of the andalusite zone. Whereas coexisting ferromagnesian phases may form permissible equilibrium assemblages because of their complex compositions, it seems reasonable that at a given temperature and pressure only a very small zone should contain rocks containing more than one aluminosilicate. Within a narrow zone it would be possible to find in equilibrium two minerals of the same composition (polymorphs). The zone would represent a "frozen-in" zone of pressure and temperature lying on the univariant transformation curve of the mineral.

It is possible that this apparent disequilibrium phenomenon does in fact represent the preservation in a lower rank assemblage (andalusite-muscovite-etc.) of sillimanite representing an earlier higher rank assemblage. However the habit of the sillimanite in these rocks is quite different from its typical equilibrium form in the higher rank schists, for it forms disseminated wispy fibers, not concentrated on any given site, but equally abundant in both andalusite and mica, as inclusions. More importantly, these sillimanite-andalusite rocks do not show any textural evidence of having gone through the high rank metamorphism of the typical sillimanite schists but rather they contain relatively well-preserved relict sedimentary structures such as are found in the muscovite zone. Within the andalusite zone sillimanite never makes up more than 5 per cent of the volume of the rock (as estimated in thin section) and most commonly is much less than 1 per cent, even in rocks containing 30 per cent andalusite. The random, unoriented wisps of fibrolite contrast markedly in texture to the well-oriented layers of mica of these rocks. Fibers of sillimanite grow across micaceous layers without distortion, while andalusite porphyroblasts enclosed in these layers show swirling inclusion trains indicating that they have been rotated during growth. This textural evidence indicates that sillimanite in these rocks was a late, possibly the last phase to form.

Two possible modes of origin may be suggested. The first is that this sillimanite was formed much later than the major minerals of the rock, and essentially as a contact metamorphic mineral, during the intrusion of the southern California batholith. Unfortunately the spatial distribution of this phase in the andalusite zone is not known, so that it is impossible to relate the distribution of this phase directly to the distribution of igneous rocks in this area. This mechanism would imply that the period of contact metamorphism was short relative to the period of regional metamorphism to which most of the metamorphic phenomena described above have been attributed, since relatively little effect of this process is recorded in the rocks.

Alternately, referring again to fig. 26, it is seen that sillimanite could be produced in rocks of andalusite-grade if the pressure on these rocks was momentarily raised due to some brief orogenic spasm during metamorphism. Such a pulse of higher pressure may have occurred late in the regional metamorphic episode of these rocks, perhaps related to the intrusion of the batholithic rocks and the consequent squeezing of the country rock by the invading magma.

The persistence of muscovite as a phase in pelitic schists of all metamorphic grades is perhaps also symptomatic of failure to reach equilibrium. Although a possible stable phase in the muscovite and andalusite zones, muscovite should be consumed in the sillimanite zone by a reaction such as (3) above. The modal abundance of muscovite does drop markedly from the andalusite zone to the garnet zone, respectively from about 20 per cent to about 1 per cent or less. The greatest drop occurs east of the sillimanite isograd. As noted earlier (Chapter II) there is a narrow zone, parallel to the andalusite zone and between it and the sillimanite zone in which andalusite is entirely replaced by a "shimmer aggregate" of sericite, clots of sericite flakes pseudomorphic after andalusite. East of this zone muscovite occurs principally as disseminated flakes in the gneisses, not generally related in texture to sillimanite. The zone of replacement of andalusite by muscovite and the higher rank occurrences of this mineral are apparently retrogressive metamorphic effects. The muscovite of these rocks is unoriented with respect to the foliation or in some cases two generations of muscovite are present, one consisting of oriented flakes of muscovite, commonly intergrown with biotite, and the other of randomly oriented clots or coarse tablets.

Under the heading disequilibrium phenomena also falls the preservation of lower rank relicts in higher grade rocks and in particular the preservation of relict sedimentary compositional features in the schists and quartzites. Relict textural and structural features as discussed above (Chapter II) are well preserved up to the sillimanite zone. In the lowest rank (muscovite zone) rocks the presumed parent phases of the micas, namely clay and chlorite, are completely obliterated and it is impossible to tell to what extent these phases may represent recrystallized clastic mica. Feldspars, both alkali and plagioclase, lack any distinctive features of a possible high temperature epoch in their parent source rock. The composition of the plagioclase (oligoclase) is compatible with the low grade of metamorphism and the potash-feldspar is clear, untwinned, almost devoid of perthitic exsolution lammellae and very irregular in external form, indicating that it has already recrystallized to a considerable extent in the lowest rank rocks exposed. Studies of the polymorphism of this potash-feldspar discussed in Part II of this text have revealed that in one stratigraphic unit of the French Valley formation (fvB4) the potash-feldspar of even the lowest rank exposures (upper muscovite zone) exhibits monoclinic symmetry, suggesting a high temperature of recrystallization on the basis of hydrothermal synthesis studies of the alkali feldspars (Goldsmith and Laves, 1954). Although this may be an inherited characteristic from the parent potash-feldspar the homogeneity of this phase in all thin sections studied and the consistency of this polymorphic type in progressively higher rank samples suggests more strongly that this is an equilibrium phase in this rock.

The only minerals preserving relict sedimentary texture and composition are the heavy accessories zircon, rutile and, possibly, sphene. However if the arenaceous component of the sedimentary rocks which were ancestral to the quartzites were derived from a terrane of granitic or low grade metamorphic lithology then the primary mineralogy of this component would be essentially an equilibrium assemblage for the whole range of metamorphism exposed here.

F. THE INFERRED HISTORY OF METAMORPHISM

The distribution of the isograds and the geothermometric evidence in the Winchester-Hemet area indicate that the course of progressive metamorphism was dominated by a thermal gradient imposed on the sediments by a heat source lying to the northeast or east. Textures of the metamorphic rocks indicate that for the most part the thermal gradient was in effect simultaneously with the application of shear stress sufficient to form schistose textures in the pelitic rocks. The present record of thermal history presumably represents the highest temperature to occur at each point in the area (Fyfe, Turner and Verhoogen, p. 97). Although there is no clear evidence to this effect at present, one may assume that areas of high rank petrology passed through stages of low and intermediate rank mineralogy and texture before arriving at their present state.

If the heat source for metamorphism is taken to have been the major tonalite intrusive underlying the Lakeview Mountains and the hills south of Hemet then this would place the time of metamorphism in the Cretaceous simultaneous with the period of intrusion of the southern California batholith. However it is noted that similar large tonalite intrusives occur throughout the batholith bordered with rocks of all metamorphic ranks ranging from phyllites such as are found in the lowest rank exposures of this area to migmatite gneisses such as are typical along the tonalite contact here. For example at the type locality of the Bonsall tonalite this rock intrudes fine-grained phyllites and low to medium rank metavolcanics.

Let us on the other hand consider the distribution of high rank metamorphic rocks throughout the whole batholithic province. Regardless of the intrusive history of any given area a consistent trend from phyllites through medium-grained schists to highly recrystallized migmatites prevails from the coastal ranges inland to the San Jacinto-Santa Rosa-Borrego province. If this distribution of metamorphic rock types is compared with Larsen's map of the distribution of igneous rock types it is evident that there is little correlation between degree of metamorphism and intrusive lithology or proximity to intrusive masses. This fact was pointed out by Larsen (1948, p. 19). Larsen also notes that "in general the rocks tend to be more metamorphosed and coarser in grain toward the east" (p. 19). In the Winchester-Hemet area this progressive trend can be walked out within a single block of metamorphic rocks. Usually the interposition of igneous plutons prevents tracing this progression out and we are left with only septa of phyllite or gneiss. The Julian metamorphic series for example is a section of metasedimentary rocks completely isolated by igneous rocks from any nearby less metamorphosed equivalents.

In such a general survey of metamorphic trends it must be remembered that the eastern metamorphic blocks are separated from the Peninsular Ranges by a series of strike-slip faults of unknown displacement, possibly bringing metamorphic assemblages of greatly different origin into juxtaposition. However, traverses made across the San Jacinto fault zone in the vicinity of Anza show no great discontinuity in either bulk composition or metamorphic grade. The development of a complete picture of the pre-batholithic history of the province is contingent on being able to correlate to some degree the metasedimentary rocks on either side of this fault system, but evidence presented here and elsewhere in the literature is as yet insufficient to do more than suggest extensions.

The most general conclusion that may be deduced from the above evidence is that, preceding intrusion of most or all of the batholith, there was a period of regional metamorphism through most of the Peninsular Ranges. Isograds produced during this period trended approximately northwest-southeast. The core of this metamorphic complex (the highest rank assemblages) perhaps underlay the present San Jacinto-Santa Rosa Mountain chain. Regionally metamorphosed sediments of Paleozoic (?) age in the Colorado Desert to the east may have been metamorphosed during the same epoch though transcurrent faults separating the various structural blocks east of the Peninsular Range make any such extrapolation suspect.

Evidence based on geothermometry in the Winchester-Hemet area indicates pressures of metamorphism of 3-6 kilobars, equivalent to burial of 12-24 kilometers in the crust. The geothermal gradient implied by assuming maximum temperatures of 800°C and a maximum pressure of 6 kilobars is 33°C/kilometer. This is close to the "normal" value of 30°C/kilometer given by Birch (1955) but in fact, assuming lower pressures, the gradient may be as high as 55°C/kilometer. The vertical gradient over the colder parts of the metamorphic complex (temperatures about 200°C) was then much less than the "normal" gradient, even assuming the lower estimate of the pressure. The lateral gradient, that which is actually represented by the measured geothermal profile (fig. 25), is as high as 140°C/kilometer (in the garnet zone). This gradient is the apparent gradient of maximum temperatures reached at each point and is actually the envelope of successive gradients impressed during the heating of the area. To a first approximation it can be considered to represent the steady-state gradient for a constant temperature heat source of infinite extent adjacent to the metamorphic rocks on the northeast. It is apparent however that the gradient must flatten out rapidly to the northeast if temperatures did not exceed those of a granitic melt at high P_{H_2O}. In fact it is evident from the steepness of the lateral gradient that the local source of heat represented by the tonalite pluton may have partially obliterated the evidence of an earlier, less steep gradient active during pre-batholithic regional metamorphism.

G. CONCLUSIONS: THE CLASSICAL METAMORPHIC GRADES

Having thus far discussed the metamorphism of the Winchester-Hemet area in strictly mineralogical and physico-chemical terms, it is necessary at this point to refer the metamorphism to the nomenclature of classical metamorphic petrology. Principal reference is made to Chapter VII by F. J. Turner in Fyfe, Turner and Verhoogen (1958) and page references below are to that work unless otherwise indicated.

From a mineralogical standpoint the metamorphic rocks of the Winchester-Hemet area fit most closely to the hornblende hornfels facies as defined by Turner (p. 206). This facies is typified by the metamorphic rocks of the Ythan Valley of Aberdeenshire, described by Read (1952), and called by him the "Buchan" type of metamorphism. The hornblende hornfels facies is considered by Turner to represent metamorphism at lower pressures but essentially the same temperatures as those of the typical "Barrovian" metamorphism. In the writer's opinion the name "hornblende hornfels" is rather inappropriate since in neither the "typical" example of the Ythan Valley metamorphic rocks nor in many of the other examples given by Turner of this facies is a hornfels texture typical of the rocks of this grade.

The stages of metamorphism of pelitic rocks described for this facies by Turner are (p. 205):

1) Knotted andalusite-cordierite-mica phyllites
2) Andalusite-cordierite-mica-plagioclase schists
3) Cordierite-sillimanite-mica-feldspar gneiss, commonly with almandine
4) migmatites

The parallelism with the successive zones of the Winchester-Hemet area is obvious. The mineral assemblages observed here in pelitic, quartzo-feldspathic, calcareous and basic assemblages can be matched with assemblages attributed by Turner to this facies. The critical mineralogical distinction between this facies and the almandine amphibolite facies (of the typical "Barrovian" metamorphism) is the presence here of cordierite and andalusite and the absence of staurolite and kyanite in the pelitic schists. The occurrence of almandine garnet is less critical ("Rather rare anomalous constituents of pelitic hornfelses are red garnet and/or staurolite", p. 207). In calcareous rocks the assemblage diopside-grossularite-wollastonite-quartz is stable, wollastonite occurring "at highest temperatures and lowest P_{CO_2} in this facies" (p. 208). Sillimanite is found in the immediate vicinity of granitic contacts (p. 207).

Thus in all major respects the mineralogy (and texture) of the rocks of the Winchester-Hemet area can be seen to fall into the hornblende hornfels facies. With respect to structure and general form the pattern of metamorphism in this area is somewhat in contrast with the typical hornblende-hornfels aureole as described by Turner. First, as described in the previous section, the metamorphism does not appear to be related to any one single intrusive igneous body. Although the sillimanite-garnet gneisses of the Winchester-Hemet area are distributed in close proximity to the Lakeview Mountain tonalite similar rocks occur in an almost continuous belt to the south, far from the contacts of this particular plutonic mass. The scale of metamorphism in general seems to be too large to be referred to as a contact phenomenon and in general bears great similarity to the metamorphic zones described by Read in the Ythan Valley for which he concludes:

> "...the whole plutonic process is essentially a unified one of regional thermal metamorphism culminating in migmatization" (Read, 1952, p. 278).

The local gradient of metamorphic intensity represented by the transition from the phyllites of French Valley to the migmatites of Diamond Valley is steeper than that usually associated with regional metamorphism, such as that of the Scottish Highlands or Dutchess County, New York.

Other factors suggest that the metamorphism here is rather transitional between Turner's hornblende hornfels facies and the typically regional almandine-amphibolite facies (p. 228). The relatively small width of the andalusite zone as noted above (p. 250 of this text) suggests temperatures approaching the upper limit of the andalusite stability field. Kyanite is however not observed here nor anywhere else in the metamorphic rocks of the Peninsular Range. The development of sillimanite-garnet gneisses over a wide belt is more typical of the almandine-amphibolite facies than of the hornblende hornfels facies. In amphibolites of this area garnet appears in exposures above the garnet isograd, producing the assemblage hornblende-diopside-andesine-garnet, typical of the almandine-amphibolite facies.

In general the writer tends to agree with Turner and Read in their attribution of this type of metamorphism to a process of regional metamorphism at a somewhat lower confining pressure (and P_{H_2O}) than the typical amphibolite facies as described by Turner (1948, 1958). The pressures and temperatures attributed to this zone are about the same as those suggested by Fyfe, Turner and Verhoogen (p. 237), namely, $P_{H_2O} = 3\text{-}6$ kilobars, $T = 300^\circ\text{-}700^\circ C$. In its lower temperature and pressure limits the hornblende hornfels facies grades into the albite-epidote hornfels facies, but this zone is not displayed in this area.

CHAPTER VII. ECONOMIC GEOLOGY

A. INTRODUCTION

The Winchester-Hemet area is not exceptionally well endowed with economically exploitable mineral resources. Except for the intermittent activity at the Hemet magnesite deposit no mining has been done during this century in this area. The present discussion is based mostly on field observations of potential economic deposits. No systematic attempt to investigate the history of the various gold, manganese and quartz prospects has been made.

B. GOLD

Potentially auriferous quartz veins are found cutting both metasedimentary and igneous rocks throughout the area. They range in thickness from a few millimeters up to twenty feet. Prospects have been dug in many localities around Crown Valley. Apparently some ore was removed from a shaft about 20 feet deep 1-1/2 miles west of Crown Valley. The ore was presumably in vein quartz. The country rock is quartzite. An arrastre is present near the head of the draw in A18(1, 1). On Double Butte in section D18(4, 2) three shafts have been driven into sheared granodiorite adjacent to a quartz vein. The shafts are about 15 feet deep. The veins at this locality usually pinch out over distances of about 50 feet, and are a maximum of four feet thick.

C. MANGANESE

Manganese concentrations of possible economic grade are associated with banded gneiss in a few localities scattered through the area. Some of the more prominent exposures are: in D35(2, 3) on the low hill just southeast of the aqueduct; on the south edge of A18(1,1) on either side of the draw; and at the southwest corner of A13(2,2) adjacent to the amphibolite. The last named locality is not in banded gneiss but in a finely laminated schist of about the same composition as typical banded gneiss. The locality in A18(1,1) has been studied in some detail as a representative of this type of deposit. The typical occurrence of manganiferous rock can be recognized by a surficial lustrous black stain on the quartzite weathering surfaces. The quartzite is cut by sub-concordant veins of glassy quartz up to a foot thick. The vein-quartz and the neighboring quartzites are fractured in box-work fashion and replaced along fractures by black manganese oxides. Locally the quartzite within this boxwork has been replaced by a pink colored hornfels. A thin section of a specimen of this rock has the following estimated composition.

rhodonite	80
garnet	5
quartz	15
opacite	1

The rhodonite is in anhedral to subhedral grains arranged in a decussate (hornfelsic) texture enclosing the quartz grains. The garnet is in scattered subhedral fine grains. The opaque minerals fill cracks cutting the rock. In outcrop this rock appears to grade along strike into pure quartzite.

The richest concentrations of manganese are lenses or pods of very fine-grained black manganese oxides within the quartz veins. A specimen of this black material was examined by x-ray diffraction and found to contain the minerals pyrolusite, tephroite and cryptomelane (Fleischer and Richmond, 1943). The dark oxide minerals form botryoidal crusts along open fractures in the quartz veins.

The zone of manganese-bearing rock in the locality of this deposit can be traced along strike for about 1000 feet to the north of the draw in A18(1,1). In this distance it does not diverge more than 50 feet from a single stratigraphic position and may in fact be restricted to one stratum a few feet thick. This appears to be true of each of the manganese occurrences investigated. It is possible that these deposits may represent metamorphosed syngenetic manganese-rich layers in the shale sequence from which these schists were derived. The association with quartz veins is not considered significant in itself since similar but barren veins are found nearby. However, mobilization of manganese to form the concentrations found here may have accompanied the deposition of the quartz veins. During thermal metamorphism of these sediments the primary manganese oxides (?) of the sediments reacted with quartz and mica to form rhodonite and spessarite (?) garnet. Later, perhaps during intrusion of the batholithic igneous rocks, the schist was fractured and veined by quartz and some of the manganese was remobilized and deposited as oxide minerals along the quartz veins. The association of these manganese enrichments with banded gneiss suggests that the process of metamorphic differentiation producing the banding may also have concentrated manganese of the rock into the micaceous bands.

D. QUARTZ AND FELDSPAR

Vein quartz and simple quartz-feldspar-mica pegmatites occur on the south end of the Lakeview Mountains, on the east flank of Double Butte, on the south side of Searls Ridge in D36(1,1), in Crown Valley and in the tonalite south of Hemet. Prospects have been quarried into most of these deposits. The quartz veins are typically composed of white massive bull quartz.

Clevelandite feldspar is present in crystals up to 2 feet across in pegmatites of the Lakeview Mountains but the crystals are intimately intergrown with quartz so that separation of the two minerals would require considerable crushing. Similar coarse feldspar is found in the dike swarm on Searls Ridge. Massive quartz has been quarried from the cores of the pegmatites in the Lakeview Mountains.

The single thick quartz vein in Crown Valley has been quarried to some extent. It should be exploitable as a deposit of pure quartz. It is up to 20 feet thick and 1000 feet long.

E. MAGNESITE

The geologic features of the Hemet magnesite mine have been described in Chapter III. The magnesite ore body is made up of the small metaperidotite intrusive on the northern side of the intrusive complex. The entire intrusive body is veined with magnesite and the limits of the quarry are essentially the contacts of the metaperidotite with its country rock. The individual magnesite veinlets are up to 3 inches thick. They form a stockwork pattern through the earthy altered igneous rock. This matrix is soft enough to be separated by hand from the ore. The abundance of magnesite appears to be uniform throughout the metaperidotite though the individual veins may become slightly thicker with depth into the ore body. The vein material is composed principally of microcrystalline magnesite with minor amounts of opaline silica uniformly dispersed throughout. Magnesite also makes up about

15 per cent of the metaperidotite matrix. The vein material makes up about 10 per cent of the metaperidotite body.

Beginning in 1907 ore was removed from this body by quarrying from the surface. Later a glory-hole was dug, tapped by drifts from the quarry floor. The ore was hand sorted at the mine and shipped to Los Angeles for grinding until 1917 when a mill and calcining furnace was constructed at the mine. Production continued intermittently until 1919. The mill has since been torn down.

The main open cut is 240 feet deep and 250 by 70 feet in projected area. There are two smaller pits south of the main quarry. The tenor of magnesite in the ore presently exposed is probably as high as in that ore which has been quarried out and the deposit is potentially further exploitable. However further quarrying would probably require stripping of waste from the rim of the quarry, since the quarry walls are almost vertical. Very little waste-stripping was required to develop this deposit in the past.

F. NICKEL

Disseminated nickeliferous sulphides have been observed by the writer in the calc-silicate marble unit at the above magnesite mine. A mineral separate of the magnetic constituents of this marble was made and analysed by x-ray fluorescence by Walter Nichiporuk and the author. The analysis obtained was:

FeO Fe_2O_3	87.41	(by difference)
NiO	9.40	
Cr_2O_3	0.67	
CO_2O_3	0.38	

It is estimated that this magnetic fraction makes up about 2 per cent of the total rock. Mineralogically it is composed of magnetite, pyrrhotite and pentlandite in the approximate ratio 2:1:1 (as observed in polished section). The distribution of these minerals through the marble is evidently very erratic and local concentrations are found, marked by dark streaks on the outcrops, in which the sulphide content may be as high as 5 per cent of the rock. It is doubtful that this would be an economically exploitable source of nickel, since most of the sulphides are disseminated so homogeneously through the marble.

G. GRAVEL AND DECOMPOSED GRANITE

The surficial deposits of the San Jacinto Valley north of Searls Ridge include local accumulations of gravel. South of Winchester there is little coarse alluvial debris except in fans at the bases of various bed-rock ridges. However, even in the San Jacinto Valley these gravel beds are poorly exposed due to the smoothness of the alluvial plain and the rarity of deep stream channels. Gravelly layers have been exposed at depths of 10 feet or less in cuts made for the purpose of laying the San Diego aqueduct east of the Lakeview Mountains.

Although there is a surficial zone of decomposition on all the igneous rocks of the area, in most localities erosion has stripped away the granitic gruss at a rate comparable to its rate of formation. At a few places, notably along Gibbel road on the east edge of the area and at the southwest corner of the area along Leon road, friable gruss up to 20 feet thick is exposed. The Leon road locality has been quarried intermittently in the last few years. Other areas of gruss may be expected where tonalite or Woodson Mountain granodiorite underlies areas of very low relief. Domenigoni Valley granodiorite is not subject to such deep decomposition.

H. GROUND-WATER

The ground-water resources of the San Jacinto basin and vicinity have been described in Bulletin 15, "Santa Ana River Investigation" of the State of California Water Resources Board. This report includes an extensive description of water yield conditions throughout most of the area studied here. An older U.S.G.S. Water Supply paper by G. A. Waring (W429, 1919) describes the water resources of the San Jacinto basin in more detail.

The principal source of ground water in this area is from alluvial fill in the various inter-ridge basins. Aquifers are commonly restricted to discontinuous stringers of lenses of clean sand or gravel. These are isolated between relatively impervious silty layers. None of the alluvium is highly indurated. Recharge of the San Jacinto basin is by run-off from the San Jacinto Mountains.

Ground water production from bedrock is of limited extent in this area, and cannot be relied on in periods of low local rainfall. The best production from bedrock is obtained in areas of deeply decomposed granitic rocks or in Bedford Canyon phyllites which have been weathered along their foliation to an appreciable depth, especially in areas of low relief.

CHAPTER VIII. OUTLINE OF THE GEOLOGIC HISTORY OF THE WINCHESTER-HEMET AREA

A. PRE-TRIASSIC

Granitic intrusives of unknown age. Evidence of pre-Triassic plutonic rocks in vicinity.

B. TRIASSIC (?)

1. Bedford Canyon Time

Deep-sea deposition of fine clastic sediments and biogenic silica. Thin arenaceous layers representing elevation to wave base and sorting out of fine material. No evidence of current action and positive evidence of finely laminated sediments indicates quiet-(deep?)-water sedimentation.

2. French Valley Time

a. Member A

Near-shore marine deposition of well- to very-poorly-sorted arenaceous and pelitic sediments. Possibly turbidity current deposition resulting in scouring and truncation of unconsolidated layers and formation of graded bedding. Arkoses appear interlayered with shale and containing shale-clasts. Feldspathic material, moderately to poorly sorted supplied by stream transport from distant granitic or metamorphic highlands, mixed with locally formed shales both as sandy-mud mixtures and as poorly sorted shale-clast conglomerates and conglomeratic sands. Minor contribution of biogenic (or reworked biogenic) calcite in sandy layers.

b. Member B

1) Continued supply of feldspathic sand by terrestial streams.

May represent deltaic deposition: local brackish-water lagoons forming shale layers on top of arkosic sand, then flooding by streams carrying suspended sand, breaking up shale and stripping shale clasts off to be mixed with sand.

2) Volcanic episode: a single flow or ash-fall of basalt or basaltic tuff, during a period of quiet-water sedimentation. Other ash falls may be represented in hornblende-arkoses.

3) After volcanic episode deltaic deposition resumed. In southeastern part of area most of Member A time and Member B time may have been represented by quiet water shale deposition.

c. Member C

Submergence below wave base. Deposition of shale, rare flows or ash falls of basaltic composition, the latter more likely in view of mixing of shale and volcanic constituents. Followed by mud deposition with interspersed periods of sand or sandy mud either as turbidity currents or as product of renewed deltaic deposition. (Evidence of sedimentational history interrupted by structural break.)

C. JURASSIC (?)

Folding of sediments of the Bedford Canyon and French Valley formations simultaneous with metamorphism:

1) Tilting of entire section in a NE-SW direction, possibly by formation of a northeast-dipping homocline.

2) Crustal stress in a N-S direction (direction of major principal stress) producing some local folding and small-scale textures (lineations).

(1 and 2 may be in reversed order.)

3) Stress in a NE-SW direction producing steeply plunging folds isoclinal in eastern part of area.

4) Intrusion of sills of basic igneous rock.

5) Beginning simultaneous with 1) above, rise in crustal temperature, a thermal gradient being developed with maximum downward slope to the west and isotherms trending generally N-S. Metamorphism of sediments:

a. Shales converted to schists
b. Sandstones converted to quartzites
c. Volcanics and basic sills converted to amphibolites
d. Partial melting of lowest-melting constituents of sediments in hotter parts of area, forming veinitic migmatites.

Deformation continued throughout metamorphism, including perhaps some movement along Rawson Ranch fault zone.

D. EARLY CRETACEOUS OR LATE JURASSIC (?)

Intrusion of ultrabasic magma into gneisses at present site of Hemet Magnesite deposit. Intrusion of carbonatite magma and basic rocks, the latter subsequently metamorphosed to amphibolites. Magnesium-metasomatism of adjoining biotite sillimanite gneiss, forming cordierite gneiss.

Metamorphism of ultrabasic rocks; re-equilibration with falling temperature (retrograde metamorphism) continuing in ultrabasic rocks after mineral assemblages of metasedimentary rocks have been "frozen in".

E. MIDDLE TO LATE CRETACEOUS

Intrusion of southern California batholith, culminating period of metamorphism of sedimentary rocks:

1) Lakeview Mountain tonalite
2) Woodson Mountain granodiorite
3) San Marcos gabbro
4) Domenigoni Valley granodiorite
5) Pegmatites and aplites
6) Quartz latite porphyry

sequence uncertain

sequence very uncertain

Depth of rocks which are at present exposed was at this time 2-5 kilometers.

F. TERTIARY

Movement on San Jacinto and Elsinore faults, raising Perris block with simultaneous erosion to a peneplain. Minor faulting may have occurred in Winchester-Hemet area. Mylonitic shearing of quartz-latite dikes, with concurrent retrograde metamorphism.

Further uplift of Perris block during Pliocene and Pleistocene (?) causing rejuvenation of stream cutting. Deep V-shaped valleys formed, draining first to north and then to west as San Jacinto block rose relative to Perris block, cutting off northward drainage. Further blockage of San Jacinto River drainage causing filling of valleys and formation of broad alluvial plain about level with the original Perris surface, leaving "inselbergs" of metamorphic and igneous rock scattered over block.

(The latter summary of Tertiary history is taken in part from Dudley (1936). He would however have valley cutting preceding formation of peneplain, valleys being buried during surface cutting.)

G. QUATERNARY

Continued alluviation of valleys.

PART II. GEOCHEMICAL INVESTIGATIONS OF AN ARKOSIC QUARTZITE OF THE WINCHESTER-HEMET AREA, CALIFORNIA

CHAPTER IX. INTRODUCTION

A. THE PROBLEM

1. Isochemical Metamorphism

In studies of progressive metamorphism it has been commonly assumed that the process of progressive recrystallization goes on isochemically. That is, that the system represented by the rock unit undergoing metamorphism neither gains nor loses chemical constituents. This assumption has been tested for various selected areas of regional metamorphism throughout the world and to a fair degree of approximation found to be correct. It has been shown recently however by the investigations of Shaw (1954a, b) and Engel (1958) that some minor changes in chemical composition can indeed be detected under such circumstances. Studies by these authors have indicated that pelitic and arenaceous rocks can undergo small changes in their minor and major element composition relatable to the effects of progressive metamorphism. The principal changes that are observed are with respect to the mobile components, H_2O, CO_2, etc.

2. Internal chemical equilibrium

Simultaneous with the mineralogic changes which take place during metamorphism, crystallization of new phases and recrystallization of phases with wide stability ranges, there are known to occur chemical re-equilibrations between the various phases which make up the rock. In general these redistributions of components between the phases of a rock should be simply related to the total rock composition, temperature, pressure and activities of all mobile components. Some work has been done establishing tie lines between coexisting phases in rocks of complex composition (e.g., Ramberg (1952), Devore (1955)). A complete picture of the manner in which both major and minor elements undergo changes in their distributions between coexistent phases in a single rock unit during metamorphism is as yet unavailable.

B. THE PURPOSE OF THE PRESENT STUDY

In this work it is proposed to choose a single monotonous and homogeneous lithologic unit and study the geochemical and mineralogic changes in it with progressive metamorphism. The lithologic unit chosen is one from the stratigraphic section of the French Valley formation as described above, subunit fvB4. It is an arkosic quartzite. This unit can be traced from the lowest grade metamorphic zone to the highest grade zone observed in this area. The reasons for its choice are given more specifically below.

It is proposed to show in detail, by means of analysis of successive samples along a traverse within this unit, what changes, if any, in bulk composition occur within the unit with progressive metamorphism. Also, by separating the mineral constituents from this rock and analysing them, it is hoped that one may demonstrate progressive chemical re-equilibrations of these phases with respect to their major and minor chemical components. Time limitations on this research have permitted studying only two phases in detail, biotite and ilmenite, but suggestions as to trends in composition of the other phases are made also, based on analyses of the total rock.

C. CHOICE OF THE UNIT FOR STUDY

The criteria used in deciding the choice of a lithologic unit for study were as follows, listed in order of importance:

1) The unit should be structurally situated so that it transgresses all the metamorphic zone boundaries.
2) The unit should be lithologically monotonous parallel and perpendicular to strike, the latter requirement being imposed to insure that small errors in location relative to marker strata will have negligible effect on the bulk-chemistry of the samples.
3) The overall texture of the rock should be homogeneous enough that sampling effects will not introduce artificial variation in the total rock composition of the sample.
4) The texture and mineralogy of the rock should be such that all important phases could be readily separated by routine mineral processing techniques.
5) The unit should be accessible over the length of the metamorphic gradient so that samples may be conveniently obtained.

Of the characteristic metasedimentary rock types of this area, schist, quartzite and calcareous quartzite, examples of all three were considered for the purposes of this investigation.

Of all the schist units considered objections were raised by the following facts:

a) The individual strata are not traceable over distances of more than two miles along strike, except for some units for which there is, in turn, uncertainty as to their exact stratigraphic continuation due to poorly defined structural relations (e.g., subunit fvA6).

b) There is considerable variation in composition perpendicular to bedding in most units.

c) The typical high rank assemblages (above the sillimanite isograd) contain fibrous sillimanite dispersed through all the phases of the rock and especially through the micas, so that extreme precautions would be necessary in order to obtain pure mineral separates.

The calcareous quartzites were considered unsuitable for the following reasons:

a) No single unit could be traced over a sufficient distance to satisfy requirement 1 above.

b) There is great compositional variation along strike, principally representing varying initial content of $CaCO_3$.

c) The high rank calcareous quartzites have been metamorphically differentiated during metamorphism into layers of varying composition and by concretion of individual minerals; representative bulk samples would have to be exceptionally large.

It will be seen from the description below and the general stratigraphic descriptions of Chapter II that the meta-arkose of Member B satisfies the criteria listed above quite well. The upper-most meta-arkose unit, subunit 4 of this meta-arkose section, has been chosen for study. The study has been made on seven composite samples collected at uniformly spaced intervals along this horizon as it passes from the muscovite zone to the garnet zone. In regard to the criteria of choice listed above, the following possible objections may be made about this unit:

1) Much of its lower-rank and middle-rank exposure is buried below alluvium in Domenigoni Valley, so that a part of the sequence of metamorphic grades will not be represented in any sampling program.
2) The rock is somewhat inhomogeneous, in that it contains discrete clasts of shaly material embedded in a sandy matrix. However the scale of this inhomogeneity is small enough that moderately large samples will suffice to average out the local differences in the ratio of clasts to matrix.
3) No mineralogic changes critical of the progressive metamorphic sequence in this area can be observed to occur in this unit (except for the appearance of traces of garnet) and the appropriate rank of metamorphism must therefore be determined by observation of nearby pelitic schists.

D. SAMPLING PROCEDURE

Samples were collected in all cases from a zone in this unit 200-300 feet below the base of Member C of the same formation. By referring to this marked stratum it was possible to locate the equivalent stratigraphic position of the sample location, except for samples 4-6 and 4-7 discussed below.

Seven composite specimens were obtained, averaging about 12 pounds apiece. Each sample represents a grab sample selected from the freshest exposed rock over a volume of outcrop of about 25 cubic feet. A single-jack was used to strip weathered rock from the surface and to break the fresh specimens for sampling.

The sample locations were spaced at roughly equal intervals along a four-mile traverse. The metamorphic zone assignments of each specimen are as follows:

4-1, 4-2 and 4-3 from the muscovite zone
4-4 and 4-5 from the andalusite zone
4-6 from the sillimanite zone
4-7 from the garnet zone

The sample locations are indicated on the overlay to the geologic map, fig. 2. Specimens of schist from the base of Member C near each sample location were also collected to compare with the metamorphic lithology of the typical pelitic schists of each zone.

There is some uncertainty as to the stratigraphic position of specimens 4-6 and 4-7 because a wedge of schist (fvB4s$_2$) appears at this point in Member B. It was decided on the basis of structural relations that the critical horizon continued below rather than above this wedge. No samples were collected east of sample 4-7 because of uncertainty as to the continuation of the critical horizon through the complex structure west of Crown Valley. However, quartzites of essentially identical mineralogy and texture can be found east of here and approaching to within one mile of the tonalite-metasedimentary rock contact. Probably no great errors in the assumption of primary lithologic uniformity would be encountered in sampling eastward across strike through the major arkose unit.

E. PRESENTATION OF CHEMICAL AND PHYSICAL DATA

The samples of meta-arkose are numbered in order of inferred increasing rank of metamorphism. This has been done, as indicated in section D above, by noting their position relative to the isograds in the adjacent schists. In the following sections graphical representations of various chemical and physical parameters are shown. In each of these the abscissa of each point is meant to be the approximate relative position of the sample in which the parameter was measured, along the line of samples. It is presumed that this relative position corresponds to the relative value of the maximum temperature the sample was subjected to during metamorphism. No attempt has been made to correlate these positions with an absolute temperature scale.

By use of this scheme of graphical representation it is possible to show the relative magnitude of change of each parameter as a function of metamorphic grade. The positions indicated are not strictly spaced in proportion to the spacing of the samples on the sample sequence (cf. fig. 2), and the curves drawn between points should be taken to represent only the gross trend of change.

In figs. 27 and 30 the gap in spacing of sample points between samples 4-1, 4-2 and the remaining samples represents the gap in the exposures of unit fvB4 due to burial by alluvium in Domenigoni Valley.

F. DISEQUILIBRIUM

In any such study as this it is necessary to assume that the physical conditions inferred to have given rise to the observed distributions were imposed simultaneously on each sample, that the prior conditions of temperature, pressure, etc., have negligible effect on the final assemblage and that, most significantly, the various equilibria as observed in a given rock today were all quenched in at the same temperature, pressure, etc. That is, in summary, the metamorphic epoch was a unique, well-defined event. It appears impossible to validate this assumption for most natural occurrences, though instances are found in which obvious violations of these assumptions are known to have occurred. With respect to the area under study at present the problem of such violations has been discussed in Chapter VI under "Disequilibrium". It will be assumed in the subsequent discussion of the geochemistry and mineralogy of subunit fvB4 that such disequilibrium effects are negligible. In support of this assumption it can only be noted that no mineralogic evidence of disequilibrium has been found in the specimens of this unit here investigated. To an unknown extent the unsystematic behavior of some of the variables investigated may be attributed to failure of the samples to satisfy the three assumptions noted above.

CHAPTER X. PETROLOGY OF THE ARKOSE

A. INTRODUCTION

The general features of the meta-arkoses of Member B of the French Valley formation are described in Chapter II. Subunit fvB4 is a slightly conglomeratic meta-arkose typical of much of this member. It is composed of quartz, untwinned potash-feldspar, oligoclase, biotite, ilmenite, graphite, tourmaline, zircon and apatite. Traces of garnet are present in the highest-rank occurrences (samples 4-6 and 4-7). This subunit is exposed on the west end of Searls Ridge, and in a belt extending southeast from the south edge of Domenigoni Valley to Crown Valley. Here it arcs around the nose of the Domenigoni Valley syncline and can be traced north along Rawson Canyon. Typical exposures of this unit are shown in Plate 17 a, b (appendix).

B. TEXTURE

The texture of the lowest rank specimens (4-2 and 4-1) is as follows: subangular relict clasts of quartz and feldspar and angular clots of phyllite or schist are embedded in a matrix of finer-grained quartz, feldspar and biotite. The schist clots are composed of biotite flakes embedded in a matrix of very fine-grained quartz and even finer-grained graphite dust. Graphite is also disseminated as inclusions in the biotite, a fact which has led to difficulty in analysis.

The grain size of the average quartz and feldspar grain increases only slightly from the lowest- to the highest-rank specimen. Estimates of grain size of the various components are given below (in mm.).

Sample No.'s	4-1, 4-2	4-3, 4-4	4-5, 4-6	4-7
Coarse clasts:				
quartz	0.30-0.50	0.50-1.10	0.32-1.10	0.30-1.60
feldspar	0.50-0.60	0.60-0.80	0.70-0.80	0.60-1.00
biotite	0.08-0.20	0.06-0.25	0.08-0.25	0.10-0.40
Matrix:				
quartz and feldspar	0.03-0.25	0.06-0.10	0.05-0.10	0.06-0.20
Phyllite or schist clots:				
quartz	0.03-0.05	0.03-0.05	0.03-0.08	0.08-0.24
graphite	0.010	0.010-0.015	0.015-0.025	0.025-0.05
biotite	0.12	0.14	0.17	0.20-0.48

These analyses are based on ocular micrometer measurements of the mean diameter of grains crossed by a random traverse through a thin section of each sample. Assuming the phyllite and schist clots to represent shale clasts, it is difficult to estimate in any such fashion the relative proportions of "matrix" to "clasts" in each specimen. The measurements shown above suggest that although the maximum grain size of the coarser clasts does not increase greatly, the minimum observed size of grains that are distinctly larger than the "matrix" grains increases. It appears, on the basis of visual estimates that the amount of "matrix" material decreases with progressive metamorphism. The biotite flakes of the schist clots grow in size in parallel fashion to those of the matrix, but are generally somewhat coarser. The general effect that is noted with increasing metamorphism is a homogenization of grain size of the various tectosilicates and increase in grain size of the biotite.

The lowest-rank meta-arkoses of the French Valley formation (in

Member A east of French Valley) exhibit essentially the same texture as the lowest rank specimens of subunit fvB4. The well preserved clastic outline of the shale chips, the graded bedding features and the delicate interlamination of phyllite and quartzite observed in these lower rank rocks (and in fvB4) suggest that the texture of the matrix is also a well preserved relict of the primary sedimentary rock. The alternative hypothesis that the matrix of even the lowest rank specimens is recrystallized from a silty or even tuffaceous glassy (rhyolitic) matrix seems less likely. One would expect that some direct evidence of such an origin would be preserved at least in the lowest rank examples.

Similar quartzites of even higher rank than sample 4-7 are essentially equigranular with a median grain size of about 1.0 mm., except for graphite which never exceeds 0.2 mm. With increasing metamorphism the tectosilicate (quartz and feldspar) grains become more rounded and the number of extremely irregular grains in a thin section decreases. The differences in texture between the lowest and highest rank specimens are quite obvious but the changes along the sequence are very gradual (Plate 18a-d, appendix).

C. THE MINERALS OF THE ARKOSE

a) Quartz

Quartz in this rock is of two types: 1) single grains of a single optical orientation; and 2) complex rounded polycrystalline grains composed of quartz in various optical orientations. Neither quartz type shows strain shadows or lammellae. Very fine rod-like inclusions (rutile ?) are visible in some of the coarser grains.

b) Plagioclase

All the samples of arkose contain a plagioclase of about oligoclase composition. It is anhedral and unzoned. It rarely shows albite twinning. The plagioclase is commonly dusty with traces of clay minerals (in the thin sections of slightly weathered rock). Very rarely wispy patches of antiperthite are found within the grains, about equally commonly at all grades of metamorphism.

c) Potash-feldspar

All the potash-feldspar is untwinned and homogeneous in optical orientation. Fine lammellae of exsolved sodic plagioclase are present in the higher rank (4-4 to 4-7) specimens, oriented parallel to (010). This material makes up less than 0.1 per cent of the feldspar by volume. The feldspar occurs in equant anhedral grains from which spiny limbs extend out between adjacent grains of quartz and plagioclase.

d) Biotite

Biotite is present in subhedral flakes with well developed (001) faces in even the finest grains. It varies in color with increasing metamorphic grade from slightly greenish brown to reddish brown. Inclusions are common in the biotite of some samples. These inclusions are mostly opaque and in most cases probably graphite, although ilmenite may be present. Zircon is very rare as an inclusion although this is a common texture in the adjacent schists. Other included minerals are quartz and rutile.

e) Graphite

In the lower rank arkoses (4-1 to 4-3) graphite is in very fine anhedral grains mostly restricted to the shale chips. With progressive metamorphism it grows in size probably by recrystallization of microcrystalline aggregates into single crystals.

f) Ilmenite

Ilmenite occurs in clumps of very fine subhedral grains. The clumps become more diffuse with increasing metamorphism and the individual grains become more euhedral.

g) Garnet

Traces of pinkish garnet in anhedral equant crystals are present in 4-6 and 4-7. The garnet is commonly adjacent to denser biotite clumps. It is free of inclusions and unaltered.

h) Zircon

Traces of zircon are present in all samples studied. This mineral is anhedral to euhedral, in prisms parallel to the c-axis. Commonly it has the form of a slightly rounded euhedral prism. It invariably contains fine disseminated (unzoned) opaque metallic inclusions, possibly graphite. The zircon is non-magnetic at a setting of 1.7 amperes on a Franz isodynamic separator, indicating that the inclusions are probably non-magnetic. The zircon does not appear to change in texture or color during metamorphism. The population of grains from a single sample is extremely inhomogeneous in form.

i) Tourmaline

This mineral is present in specimens 4-1, 4-6 and 4-7 as fine stubby euhedral prisms associated with the biotite clumps. Its pleochroism is ϵ = olive green, ω = greenish yellow. It is not present as a clastic mineral in the lowest rank quartzites and is apparently a product of metamorphism.

j) Magnetite

Very minute traces of magnetite, observable in the most magnetic mineral separates, are present in most of the samples. Less than 1 mg. was separated from each of the 7-10 pound specimens (compared with 30-50 mg. ilmenite). Very little of the opaque material observed in thin section could be magnetite.

k) Limonite, hematite, etc.

Oxide minerals of iron are present as intergranular films in the lower rank specimens. They may be strictly weathering products as they are not observed in thin sections of fresh rock.

l) Apatite

Fine subhedral prisms of colorless apatite are present in some thin sections of most specimens. This mineral is apparently a relict clastic mineral.

m) Sericite

This mineral is present in a few sections as an alteration of plagioclase. It is very fine-grained and uniformly disseminated through the feldspar grains.

n) "Orange Mineral"

An unidentified orange mineral is present in specimens 4-1, 4-2, 4-4 and 4-5. It is isotropic or weakly anisotropic, translucent and anhedral. Its refractive index is variable from 1.70 to 1.80. Individual grains are composed of aggregates of finer grains in varying optical orientation. Its hardness of the Moh scale is less than 2. An x-ray powder pattern of this mineral matches some reflections for Huttonite but many lines cannot be accounted for. A qualitative emission spectrographic analysis of a hand-picked specimen shows the following elements to be present, in order of decreasing abundance:

Fe, Al, Si, Ti, Mg, Ca, La, Mn, Pb, Y and Yb.

A radiometric analysis of a 20 mg. hand-picked sample was made by B. J. Bolinger. The observed counting rate of 17 cts/hr. is equivalent (for the experimental conditions used) to a uranium concentration of 112 parts per million. This mineral may be a partially metamict allanite, now replaced by clay minerals or hydrous aluminum oxide and iron oxide (Silver and Grunenfelder, 1957).

o) Myrmekite

Equant blebs of intimately intergrown albite and quartz are found along the margins of plagioclase grains in specimens 4-5 to 4-7. This type of association has not been distinguished as myrmekite in modal analyses because the contrast between these blebs and the host plagioclase is so poorly defined in many cases. About 0.1 vol. per cent is present in all thin sections.

D. MODAL ANALYSES

1. Procedure

Chips were taken from both the weathered crust and the fresh core of all specimens. Four thin sections were at first made of each specimen, two weathered and two fresh. The thin sections of the fresh specimens were etched with cold HF fumes and stained with saturated sodium cobaltinitrite solution. It was found however that it was possible to distinguish potash-feldspar in all specimens without benefit of staining, relying on that mineral's low relief. However it was not always possible to distinguish fresh oligoclase from quartz, due to the lack of albite twinning in the feldspar. It was found that in the thin sections made from weathered specimens these two minerals could be in some instances distinguished on the basis of the dusty clay alteration of the feldspar.

Late in the course of research an attempt was made to remedy this deficiency in the modal analysis procedure. A technique of staining was developed whereby both plagioclase and potash-feldspar are given distinctive colors in thin section, permitting the former to be distinguished from quartz. The procedure is as follows:

1. The uncovered thin section is etched with cold HF fumes for about two minutes.
2. The section is immediately immersed in a saturated aqueous solution of sodium cobaltinitrite for 45 seconds.
3. The section is gently washed in a large beaker of cold water.
4. The section is immersed in a 3 per cent solution of eriochrome black to which has been added a few drops of methanol for five minutes.
5. The section is gently washed in a large beaker of cold water, dried for 24 hours and covered.

By this procedure the plagioclase is stained faintly purple while the potash-feldspar is stained a brownish yellow. Some care must be taken to prevent confusion between fine grains of potash-feldspar and biotite. An additional thin section of each of the specimens 4-1 through 4-7 was prepared according to this technique. The section of sample 4-3 was accidentally over-etched during preparation, rendering it unsuitable for analysis.

Modal analyses were made on all the above sections with a Chayes point counter. From 1100 to 1500 points were counted over an area of about

6 square centimeters on each section. No special precautions were taken to orient the path of the traverse relative to the rock fabric since the rocks were texturally almost isotropic, except for a weak planarity introduced by the shale chips. The modal analyses of the unstained and the singly stained (cobaltinitrite) sections, averaged for each rock, are shown in Table 7a. In Table 7b are given the revised modes determined on the doubly stained thin sections (cobaltinitrite and eriochrome black).

2. Analytical Data

It is seen from Tables 7a and 7b that the average composition of this unit is essentially that of an alkaline adamellite. Table 7a differs from the revised modes of Table 7b in having a higher content of quartz and lower plagioclase, as would be expected from the preceding section. Using the feldspar relative abundance of Table 7b the ratio plagioclase:potash-feldspar is about 3:2, typical of many adamellites, while the quartz content, about 40 per cent, is also typical.

Using the chemical analyses of the total rock given in Chapter XII, norms have been calculated for specimens 4-1 and 4-7. These are given in table 8. Comparing these results with Tables 7a and 7b we note that the normative values of quartz and plagioclase agree very well with the revised analyses but not at all with the original analyses (Table 7a). This observation tends to confirm the suspicion that much unaltered, unetched plagioclase was overlooked in the first set of modal analyses. In part the differences between norm and mode are due to the solid solution of albite in the potash-feldspar (note differences in composition between normative plagioclase and that observed (Table 21)). This difference is compensated for in part by the potash of the biotite which

TABLE 7. Modal Analyses of the Meta-Arkose (weight per cent)

a. Averages of unstained and singly stained thin sections:

	4-1	4-2	4-3	4-4	4-5	4-6	4-7	ave. 4-1-7
quartz	43.1	45.7	58.0	54.1	47.6	44.8	54.2	49.6
K-feldspar	23.6	24.4	19.2	20.0	22.8	22.7	18.6	21.6
plagioclase	22.8	15.5	8.4	12.9	11.5	23.0	16.2	15.8
biotite	8.8	12.0	12.7	12.2	16.7	7.8	10.5	11.5
opaques*	1.7	2.4	1.7	0.7	1.3	1.7	0.5	1.4
apatite	0.1	-	0.1	0.1	0.1	-	0.1	0.1

b. Doubly stained thin sections

	4-1	4-2	4-3	4-4	4-5	4-6	4-7	ave. 4-1-7
quartz	37.6	40.6	-	36.1	36.1	37.1	40.0	37.9
K-feldspar	24.4	30.4	-	18.1	18.4	25.4	13.3	21.7
plagioclase	30.4	17.1	-	32.0	32.0	26.7	35.5	29.0
biotite	6.2	9.2	-	12.3	12.6	9.0	9.9	9.9
opaques*	1.4	2.7	-	1.6	1.0	1.8	1.3	1.6

*opaques = ilmenite + graphite, approximately 5:1.

TABLE 8. Norms of Chemically Analysed Rocks

	4-1		4-7	
Q	38.8		40.6	
Or	22.2		19.5	
Ab	31.4	An_{10}	29.3	An_{14}
An	3.6		5.0	
Cor	1.1		5.1	
Hyp	1.1	Fs_{30}	2.3	Fs_{40}
Hem	0.5		-	
Ilm	0.5		0.6	
Mt	-		0.7	

is given as Or in the norms.

The similarity of the average composition (see last column, Table 7a, b) to that of an igneous rock suggests that the parent sediment of this rock was very immature. This question is further discussed below with reference to the chemical analyses.

The variation in modal analyses along the sequence of samples does not have any apparent systematic trend. Qualitatively the variation from one end of the traverse to the other does not appear to be greater than the average variation normal to strike at any given sample locality, as estimated in hand specimens. There is a suggestion in the modal analyses that the abundance of potash-feldspar may decrease slightly toward the higher rank end of the sequence. This is confirmed by the total rock chemical analyses given below. There is a complementary variation in the abundance of feldspar and quartz, due to the fact that these are the major components of the rock.

In general it appears, on the basis of these analyses, that progressive metamorphism has little effect on the mineral composition of this assemblage. It is evident that whatever are the chemical changes due to metamorphism, they are not markedly reflected in these data. The origin of the mineral assemblage as a whole is discussed later in the light of the evidence from chemical analyses.

The existence of discrete clots of phyllitic composition embedded in a matrix of arkosic composition must also be taken into account in the interpretation of the modal analyses. The volume abundance of these clasts is estimated to be about 5 per cent, varying from 0 to 20 per cent in individual thin sections. The chips are quite different in mineralogy

TABLE 9. Accessory Minerals of Arkose Estimated from Modal Analyses and Mineral Separates

	4-1	4-2	4-3	4-4	4-5	4-6	4-7
zircon	x	x	x	x	x	x	x
tourmaline	x	-	-	-	-	x	0.1
rutile	-	-	x	x	-	x	-
orange mineral	x	0.1	-	x	0.2	-	-
garnet	-	-	-	-	-	x	x
apatite	0.1	x	0.1	0.1	0.1	x	0.1
sericite	x	x	0.3	-	0.2	0.2	x

- : less than 0.01 per cent

x : present but less than 0.1 per cent

from the host rock. They are composed of biotite (25-40 vol. per cent), opaque dust (1-10 per cent) and quartz. Feldspar is absent. Because of the latter fact the chips tend to make the composition of the total rock (chips plus matrix) depart from an arkosic composition as they increase in abundance, but, because their abundance is low, the effect is small.

A more serious problem is raised by the fact that, due to their mineralogic composition, the schist clots contain a considerable fraction of the biotite of the rock. Assuming the clots to make up 5 per cent of the rock by volume and to contain 40 per cent biotite, they would contribute 2 per cent biotite to the rock or about 15 per cent of the total amount observed in the modal analyses. Because of the somewhat inhomogeneous distribution of these clots they may tend to add some dispersion to the modal biotite concentrations. It is felt that by averaging over three or four thin sections this source of error is made negligible.

In the following study we will be greatly concerned with the chemistry and mineralogy of the biotite. In the light of the above observations it is important to know if the textural separation of biotite into these two classes, matrix and clot, corresponds to any compositional distinction. In regard to this possibility the following observations have been made. In thin section there is no apparent difference in color, pleochroism, 2V or birefringence between the two biotite groups. By selecting chips of schist from hand specimens and comparing the biotite therein with that of the matrix it has been found that the refractive index n_z is approximately the same in each sample, though erratic variations are observed within single samples of each type. The same is true of 2V measurements. However optical properties are an incomplete estimate of possible compositional differences. Further suggestion of internal homogeneity of biotite in the rock specimens is the fact that the magnetic subsceptibility of the biotite from a given rock is very uniform. This is observed via its behavior in the Franz Isodynamic Separator.

Granting that there may be significant differences in the chemistry (and mineralogy) of matrix and clot biotite it should be noted that by using a single seive size fraction of biotite for analysis there is some tendency to select clot-biotite preferentially over matrix-biotite, since the latter is somewhat finer grained. This effect is most pronounced in the lower rank specimens (4-1 through 4-4). In the higher rank specimens the grain size of the two groups becomes more uniform and in fact the boundaries between clot and matrix become very diffuse. The relative abundance of clots in each specimen (4-1 through 4-7) is about the same and therefore the effect of selective sampling should be uniform except insofar as the grain size of the two groups is homogenized and the ratio of clot- to matrix-biotite in the biotite separates becomes closer to the ratio of their abundance in the rock. All but one of biotite samples separated for analysis were from the +200, -100 size fraction. The diameter of the opening in 200 mesh seive material is about 0.075 mm. We shall assume that the biotite flakes are in general disk shaped and therefore the length of the flakes in thin section is on an average the critical dimension in determining whether a grain can pass through the seive. In specimens 4-1 through 4-4 an estimated 20 per cent of the biotite of the matrix is less than 0.075 mm. across (the mean size varies from section to section from about 0.08 to 0.25 mm.). This fraction will not be present in the size-sample used. In the same series of samples almost none of the schist-clot biotite is less than 0.075 mm. across and therefore ideally all of the biotite from this group except that larger than 0.15 mm. (less than 5 per cent) will be included in the sample.

CHAPTER XI. ANALYTICAL PROCEDURES FOR THE TOTAL ROCK AND ITS CONSTITUENT MINERALS

A. PREPARATION OF THE SAMPLES FOR ANALYSIS

The manner of collection of the samples has been described in Chapter IX. The seven samples of member fvB4 were each prepared for mineral separation and total rock analysis according to the following procedure:

1) The sample was culled free of deeply weathered fragments.
2) The remaining fragments were broken to 2-5 inch pieces with a small sledge on a steel plate and weathered rinds from the outside of each fragment were culled out or chipped off.
3) The remaining pieces were broken with a small sledge to 1/8-1/2 inch fragments.
4) These fragments were rolled on a small steel plate with a 2 inch diameter steel roller, reducing the grain size to less than 1 mm.
5) The crushed rock was split in a seamless aluminum sample splitter; 1/32 or 1/64 was withdrawn for chemical analysis (about 50 grams).
6) The remaining sample was cleaned of tramp iron by passing over an aluminum chute backed with a strong alnico magnet. Analysis of the magnetic concentrate from this procedure showed it to contain negligible amounts of magnetite.
7) The crushed rock was sized by sieving in 37, 100 and 200 mesh bolting cloth. It was found necessary to extract a 300 mesh fraction from sample 4-1 due to the fine grain size of

the biotite. For this purpose a brass sieve was used, possibly adding Cu and Zn contamination to the ilmenite separates.

The -100, +200 mesh size fraction of each sample was then subjected to the following mineral separation procedures:

1) Flotation of feldspar and quartz in tetrabromoethane.
2) Flotation of biotite, etc., in methylene iodide.
3) Flotation of potash-feldspar from all samples except 4-2 in a mixture of tetrabromoethane and acetone. The density of the liquid was continuously adjusted during the procedure so that a specimen of microcline could just float in the liquid.

It was found that yields of biotite concentrate from step 1 above were far below expected yields based on modal analyses. To increase the efficiency of this separation the floats from this step were passed through a Carpco magnetic separator (Carpco Engineering Company) and the magnetic fraction was re-run through step 1. An additional yield up to four times the original amount was obtained.

The floats from step 2 were processed on a Franz Isodynamic Separator. Maintaining a constant slope and tilt of the track, it was found that a 95 per cent pure biotite separate could be obtained in one pass at a single current setting (0.3 amperes) for all samples, indicating a uniform degree of magnetic subsceptibility in all the biotites. Further purification was done on most samples by recycling on the Franz separator and by hand picking of foreign grains. The final biotite separates were in all cases over 99 per cent pure although variable amounts of opaque inclusions (graphite?) were present within the grains. The principal contaminant in all samples was the orange mineral described above. In sample 4-5 this mineral makes up almost 0.1 per cent of the final separate.

Ilmenite was obtained from the sinks collected in step 2 by processing on the Franz Isodynamic Separator and again all samples were uniform with respect to magnetic susceptibility. The ilmenite concentrates from all samples were less than 30 mg. They were 97-99 per cent pure, the major contaminant being biotite. Traces of zircon may be present in all samples.

Garnet was obtained as a non-magnetic phase on the separation of the ilmenite from sample 4-7. The garnet separate was further purified by hand picking, yielding about 3 mg. of pure garnet for optical and x-ray analysis.

Grain mounts of the biotite and polished grain mounts of the ilmenite were made to check for purity and amount of inclusions.

B. MAJOR ELEMENT CHEMICAL ANALYSIS

1. Total Rock

Chemical analysis of the total rock samples was performed by x-ray fluorescence spectroscopy and in part by classical wet-chemical means. The procedure is tabulated below:

1) Sample preparation:

The 1/32 or 1/64 aliquots split from the whole crushed rock were further split and aliquots of about 10 grams were removed. These splits were ground in an agate mortar on a Fisher automatic grinding machine for 2 hours.

2) Standards:

Two of the whole rock aliquots, 4-1 and 4-7, were split to yield further 10 gram samples which were then sent out for wet-chemical analysis. In addition, five analysed granites, supplied by L. T. Silver, were prepared in the same fashion as the arkose samples. It had been inferred from their similarity to the arkose in modal composition that the granites would span the range of chemical composition of the arkose samples and this was confirmed for most elements in practice.

3) Equipment:

X-ray source: Norelco, tungsten target, run at 50 KV, 35 ma.

X-ray spectrograph: Norelco

Counter: flow proportional, P-10 gas

Helium path for analysis of lighter elements

Preamplifier: Baird-Atomic Model 219A

Linear amplifier: Atomic Instruments, Model 218

Pulse height analyser: Atomic Instruments, Model 210

Scaler: Baird Atomic Model 134

High voltage power supply (for counter): Hamner Model N401

Analysing crystal: Ammonium dihydrogen phosphate, 2D = 10.648 Å

Sample holders: Norelco polyethylene powder holder; sample cavity: 35 x 22 x 1 mm. Sample packed to uniform thickness.

4) Analysis procedure:

The following fluorescence lines were used for analysis:

Fe Kα	1.934 Å	K Kα	3.735 Å
Mn Kα	2.100	Si Kα	7.111
Ti Kα	2.745	Al Kα	8.320
Ca Kα	3.353	Mg Kα	9.869

The second order reflection of the Mn line was used due to Fe L interference in the first order. 2θ values for each element were determined empirically modifying them slightly from the theoretical Bragg angles due to slight errors in orientation of the analysing crystal. The pulse-height analyser was used for all elements except Ca and Fe. Window and base-line settings were determined empirically for each element, using pure salts of the element as samples.

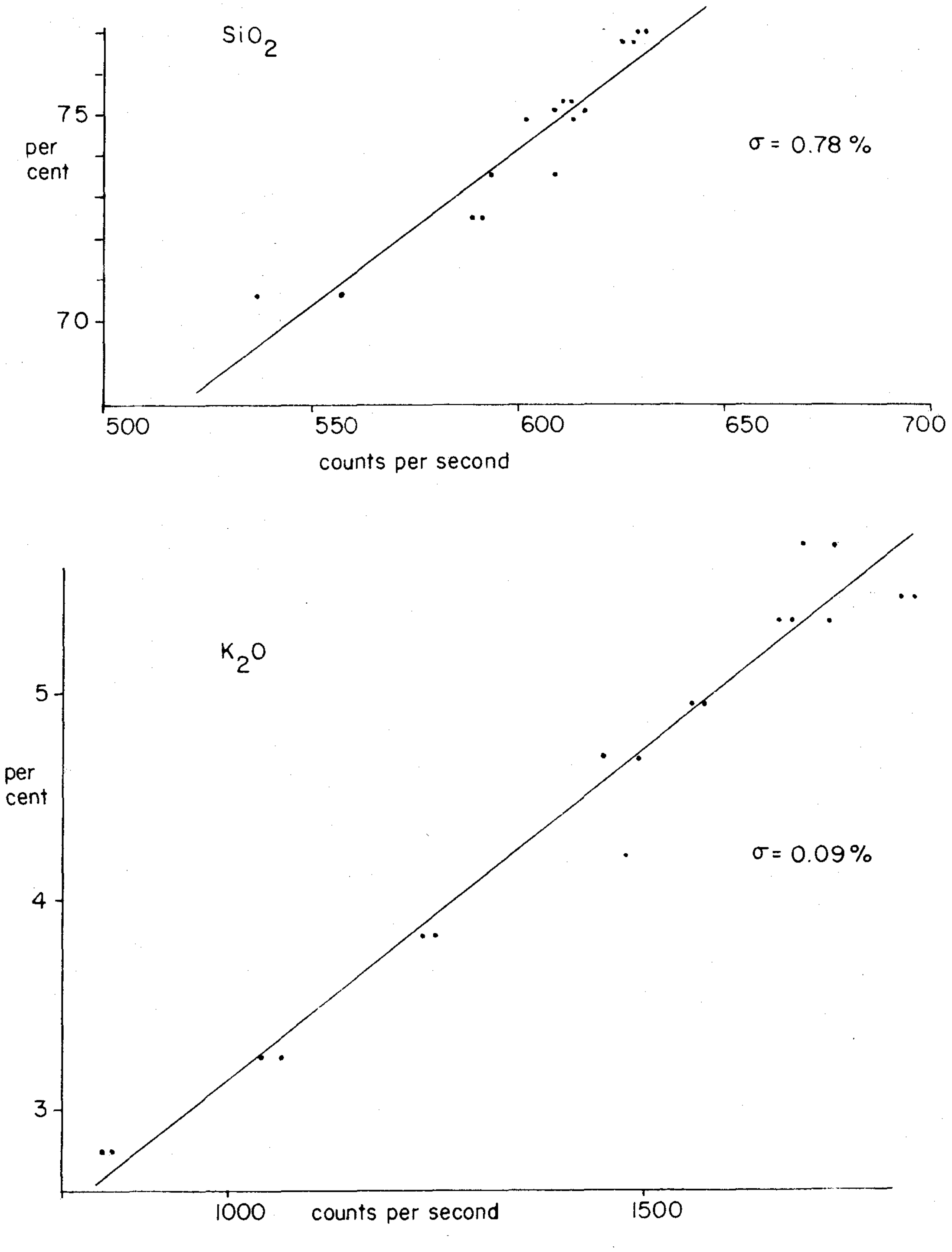
The parameter of analysis here was total intensity of the characteristic radiation from a sample of standard thickness and shape. The intensity was measured by making a fixed number of counts, measuring the time and calculating the counting rate. Background was measured at 1.0 to 1.5 degrees 2θ away from the peak on either side (a fixed location for measurement being chosen for each element). Except for Mg the theoretical error due to uncertainty in the background was small compared to other errors (Zingaro, 1958). For Mg the peak/background ratio was usually of the order of unity and presumably extraneous effects (fluorescence of the analysing crystal, interfering lines) might have a marked effect on the accuracy. Where background intensity was asymmetric due to the proximity of a strong peak (e.g. Fe Lα near the Mn Kα) the decline of background intensity with 2θ was assumed to be linear and the values on either side of the peak were averaged. In all cases background was subtracted from the peak intensity to obtain net intensity of the characteristic radiation.

The seven analysed specimens (2 meta-arkoses and 5 granites) were analysed by the above procedure and least-square adjusted lines were fitted to their counting rate vs. concentration values. The wet-chemically analysed arkoses in general fell within one standard deviation of the best-fitted line. This is taken to indicate that no great errors were introduced due to difference in matrix effects between the granites and meta-arkoses. The least-squares adjustment yields a value for the standard deviation of the population of analyses obtained for a given true concentration. For all elements except Mg the standard deviation thus obtained was small compared with the range of values among the standards and samples. Typical working curves are shown in fig. 27a.

The remaining five meta-arkoses were analysed by the above procedure, converting counting rates into concentration by use of the regression equations derived above. Two analyses were made of each sample (on a single load of a powder holder). The analyses are of course incomplete, lacking values for Na_2O, H_2O and the FeO/Fe_2O_3 ratio. An attempt was made to determine Na_2O using a gypsum crystal as an analyser but the peak/background ratio for all the standards (containing about 5 per cent Na_2O) was too low to permit reliable counting statistics.

As a check on the accuracy of the analytic procedure, values for the two analysed arkoses were obtained using the counting rates obtained while setting up the working curves and converting to concentrations with the regression equations. These analyses are compared with

FIG.27a Typical working curves for X-ray fluorescence analysis
A. Total rock



the wet-chemical analyses posted in Table 12. For the purposes of the comparison ferrous and ferric iron as reported by the analyst were converted to equivalent total Fe_2O_3.

	4-1		4-7		
	x-ray	chemical	x-ray	chemical	s
Fe_2O_{3t}	1.24	1.08	1.70	1.63	0.07
TiO_2	0.33	0.27	0.24	0.31	0.02
CaO	0.29	0.70	0.76	1.03	0.51
K_2O	3.85	3.84	3.26	3.26	0.09
MnO	0.02	0.01	0.04	0.04	0.01
SiO_2	76.33	77.00	76.07	76.76	0.78
Al_2O_3	12.16	12.33	12.13	12.23	0.57
MgO	0.29	0.26	0.48	0.49	0.37
Na_2O	n.d.	3.71	n.d.	3.47	
Remainder		0.24		0.43	
	94.57	99.48	94.68	99.53	
Na_2O + Rem.	3.95		3.90		
	98.46		98.58		

Remainder = H_2O, P_2O_5; s = standard deviation from least-squares adj.

The x-ray analyses sum a little low, it is observed, although four of the unknowns analysed by this technique summed close to 100 per cent without adding Na_2O, etc., indicating that they must be too high. K_2O and MnO are in excellent agreement. The agreement between the MgO values is probably fortuitous considering the standard deviation given above.

Table 10 illustrates the reproducibility of the counting rate data for re-runs of the same sample-pack. A re-run of one of the rock samples was also made using a second powder-pack (the data are not shown

TABLE 10. X-ray Fluorescence Counting Rate Reproducibility

I. Successive Counts on the Same Powder Pack in Counts per Minute

a. Rock Analyses

	Sample 4-3			Sample 4-6		
	2/24/59 (cps)	3/4/59 (cps)	error (%)	2/24/59 (cps)	3/5/59 (cps)	error (%)
Fe	3256.3	3232.2	0.7	4262	4195	1.6
Ti	274.98	278.84	1.4	408.4	404.2	1.0
Ca	519.35	486.64	6.2	473.9	458.8	3.2
K	968.42	935.70	3.4	1114.3	1109.9	0.4
Mn	19.17	17.47	8.9	19.81	20.19	1.9
Si	726.18	714.98	1.5	619.76	654.20	5.3
Al	40.01	39.17	2.1	49.43	33.85	31.5
Mg	0.73	0.71	2.7	1.19	0.94	21.0
average error			3.4			8.2

b. Biotite Analyses

	Sample 4-4			Sample 4-6		
	3/29/59 (cps)	3/30/59 (cps)	error (%)	3/29/59 (cps)	3/30/59 (cps)	error (%)
Fe	4541	4557	0.4	4666	4725	1.2
Mn	1191	1125	5.5	962.5	980.2	1.8
Ti	4859	4757	2.1	4325	4382	1.3
Ca	127.2	125.6	1.3	43.0	41.8	2.8
K	528.7	524.1	0.9	586.5	593.8	1.2
Si	473.0	456.5	3.5	445.6	448.51	0.6
Al	218.0	213.52	2.1	219.4	224.6	2.3
Mg	15.33	15.61	1.9	14.43	14.25	1.2
average error			2.2			1.5

II. Repacked Biotite Sample

	pack A (cps)	pack B (cps)	error(%)
Fe	4587	4660	1.6
Mn	1232	1456	15.4
Ti	3931	4000	1.7
Ca	38.23	34.69	9.3
K	4739	4803	1.3
Si	450.13	453.68	0.8
Al	253.68	387.72	34.6
Mg	n.d.	n.d.	
		average error	9.2

Explanation: The table lists duplicate analyses of samples of rock and biotite.

$$\text{error (\%)} = \frac{\text{(difference between counting rates observed)}}{\text{(higher counting rate)}} \times 100$$

The average error (= (sum of errors)/8) is taken to be an estimate of the reproducibility of analyses of a given sample.

here). This measurement gave about the same deviations for each element as re-runs on the same powder-pack.

2. Biotite

An aliquot of about 2 grams was split from each biotite separate using a small aluminum splitter. Each sample was ground for two hours in an agate mortar on a Fisher automatic grinding machine. Even with such lengthy grinding it was observed that flakes up to 0.01 mm across were preserved which in turn tended to lie flat on the packed surface of the powder mount. This tendency toward particle orientation, although roughly uniform from sample to sample, may possibly contribute to errors in reproducibility of the analyses for differently prepared samples of biotite.

The writer was furnished with four analysed biotites by A. E. J. Engel. In addition, two biotites from the meta-arkose sequence, 4-2 and 4-7 were chemically analysed by classical procedures. The two analysed biotites were split from the same fraction as that used in the x-ray procedures but were not ground.

The analytic equipment was the same as that used for rock analysis with the following exceptions:

Analysing crystal: EDDT (Ethylene diamine ditartarate) for all elements except Mg; ADP for Mg.

Sample holder: Lucite slabs with milled depressions 20 mm. x 20 mm. x 0.3 mm. Sample packed to uniform thickness.

ADP was found unsatisfactory as an analysing crystal because of the high background intensity of P Kα radiation excited by the Fe Kα radiation from the sample. ADP was used to analyse for Mg because the geometry of the machine did not permit reaching the 2θ value for Mg Kα with the EDDT crystal as an analyser.

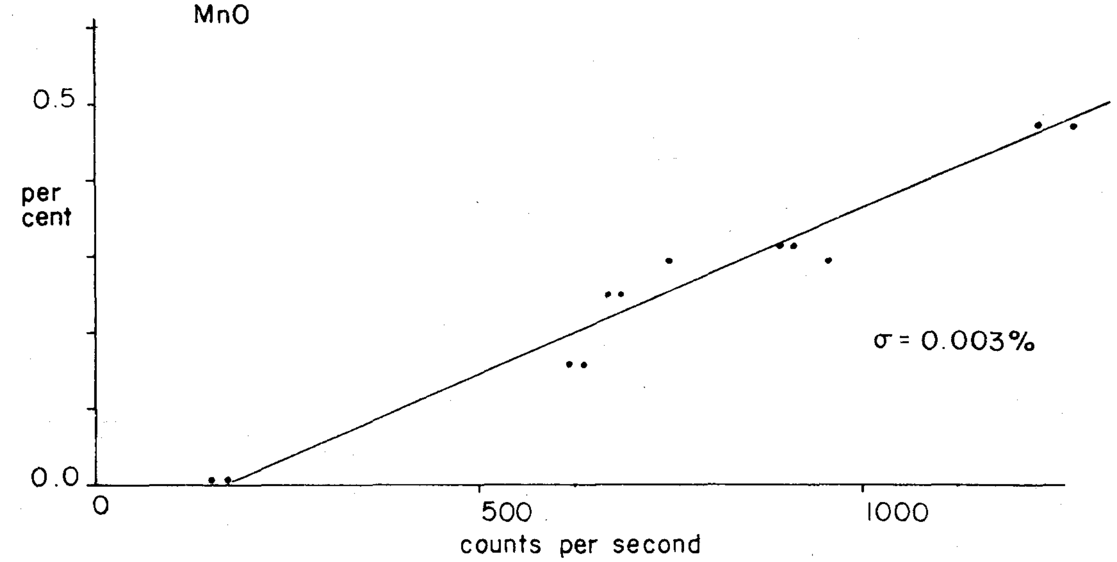
The same emission lines as in rock analysis were used for the biotites. Because of Fe interference the second order of K Kα was used. It was found that, using the EDDT crystal, background was negligible more than 0.5 degrees 2θ away from the peak for all elements except Ca, Mn and Mg. The Fe Kα intensity was so high that, to avoid coincidence errors, it was found necessary to analyse this line with the ADP crystal, but values obtained by either technique are about the same. The ADP calibration values have slightly less dispersion about the best-fitted working curve. The ADP values have been used to evaluate Fe_2O_3 concentrations in the unknown biotites.

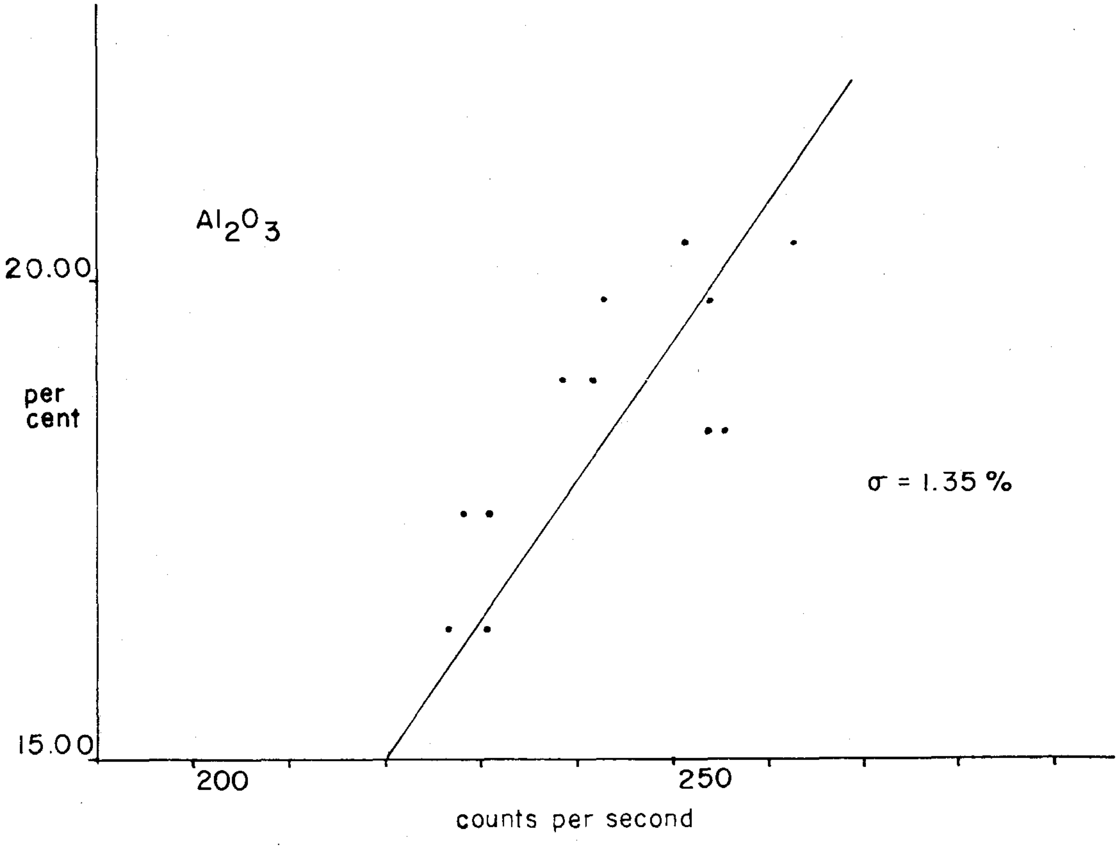
All other features of the analytic procedure are as described for rock analysis.

A working curve for each element was obtained by a least squares adjustment of the counting rate vs. concentration values for the standards. A standard deviation was obtained from the dispersion of points about this line, as before. Typical working curves are shown in fig. 27b.

The remaining five biotites from the meta-arkose were analysed by this procedure. As a check on the accuracy of the analytic procedure the compositions of the two chemically analysed biotites from the meta-arkose were calculated using the observed x-ray intensities. A comparison of the two analyses together with the least squares relative standard deviation is given below:

FIG.27b Typical working curves for X-ray fluorescence analysis
B. Biotite





	4-2		4-7			
	x-ray	chemical	x-ray	chemical	s_{rel} (%)	s_{abs}
MgO	8.82	8.53	8.64	7.62	5.6	0.51
Fe_2O_3	23.15	21.85	22.59	21.91	4.7	0.72
MnO	0.31	0.31	0.45	0.47	0.8	0.003
CaO	0.14	0.10	0.06	0.08	–**	(0.08)
TiO_2	2.89	2.68	3.70	3.42	7.3	0.27
SiO_2	35.30	37.41	33.82	36.33	11.1	0.70
Al_2O_3	16.21	16.35	20.22	18.40	8.0	1.35
K_2O	8.84	8.96	9.75	9.06	7.0	0.65
	95.63	96.19*	99.23	97.29*		

*Partial sum, not including H_2O, P_2O_5, F, Cl, Na_2O

**see discussion of statistics at end of this chapter

s_{rel} = relative standard deviation from least-squares adjustment

s_{abs} = ($s_{rel.}$) x (average amount present in biotites 4-1 through 4-7)

The analytical errors inherent in this technique are too great for most elements to permit evaluation of variations in composition between samples. It is believed by the writer that the failure to grind the biotite to a homogeneously fine grain size may be a significant factor in determining the accuracy of an analysis. Possibly fluxing of the samples in a borax or sodium carbonate pellet may yield more consistent results.

C. TRACE-ELEMENT CHEMICAL ANALYSIS

1. The total rock

Trace element analysis of the total rock was performed according to the spectrographic technique described in Engel, Engel, Godijn and Chodos (1958). This technique involves the measurement of the total intensity of the emission line from the sample. The conditions of analysis are summarized below:

Spectrograph: Jarrell-Ash 3.4 m. grating instrument, Wadsworth mount, dispersion 5.2 Å/mm. in the first order.

Excitation: 19 ampere short circuit D.C. arc from a Jarrell-Ash Varisource. Sample as anode. Analytical gap, 4 mm., focussed by spherical lens on aperture to mask glowing electrodes. Image from aperture focussed by vertical cylindrical lens on slit. Central 2 mm. used with a slit width of 25 microns; 25 mg. samples burned to completion (90-120 seconds).

Electrodes: High purity one-quarter inch graphite rod as the anode. U.C.C. No. 3417 pointed one-eighth inch cathode.

Wave length range: 2300-4800 Å in the first order.

Plates: Eastman Kodak III-0.

Processing: 4 minutes in DK 50 developer at 20°C. 20 sec. shortstop, 10 min. in acid fix, 20 min. wash.

Plate calibration: Selected iron lines after method of Dieke and Crosswhite (1943). Each plate is calibrated.

Densitometer: Jarrell-Ash Model 2100.

The analytic lines and their sensitivities are given in Table 10. The rock pulp was diluted 4 rock: 1 graphite. Two splits of each rock were prepared and diluted and each split was run in duplicate. Two samples each of G-1 and W-1 were run as a check on the accuracy relative to results obtained by this technique by other workers in the same laboratory. The comparison indicated good agreement in general. The values quoted are the arithmetic mean of the four analyses.

As a control on the amount of contamination introduced in preparation of the samples, a sample of Brazilian quartz known to be quite pure was run through the same series of processes as the rocks being analysed, starting with breaking on a steel plate and not including magnetic benficiation. Spectrographic analysis of this sample indicated that the following elements might be expected as contaminants:

Ni	2.0 parts per million
Cu	1.3
Ag	2.0

All other elements listed in table 10 were looked for and not found. The sensitivities for this control sample are the same as for the rock, as listed in table 11. In addition, appreciable iron contamination would be expected though not enough to influence the major element determinations for the total rock.

2. Biotite

The trace element concentrations in the biotite samples were obtained using the same technique as in total rock analysis. However, in order to make the samples more "rock-like" the biotites were diluted 1 graphite: 2 biotite: 2 (9 quartz + 1 $NaCO_3$). The Na-quartz mixture was prepared from spectrographically pure materials. Concentrations obtained from the working curves were multiplied by two to give the true concentrations. Each sample was run in triplicate. The values quoted are the arithmetic means of the three analyses.

3. Ilmenites

The ilmenite separates were analysed by the same technique as used for total rock. They were diluted 1 graphite: 1 ilmenite: 3 Na-quartz and observed concentrations were multiplied by 4. Sample sizes

TABLE 11. Emission Lines Used For Spectrographic Analysis

	λ (Å)	σ_{est} (%) rock	biotite	ilmenite	sensitivity* (ppm)
Pb	2833	21	-	23	10
Mn	2949.2	15	36	9	2
Ti	3261.6	8	14	-	10
Zr	3273.0	24	22	8	10
Zn	3345.0	-	26	12	20
Ga	2943	24	19	-**	2
Ni	3414.8	38	20	-	1
	3050.8	-	-	24	2
Cu	3274	15	9	8	1
Co	3453	38	15	9	1
Yb	3289.4	20	25	22	1
Y	3327.9	23	15	13	10
V	4379.2	30	15	34	1
Sr	4607.3	21	-	-	2
	3464	-	33	-	100
Ba	4554	19	18	16	2
La	4337.5	34	14	9	15
Sc	3353.7	35	19	18	3
Cr	4254	24	11	22	1
Ag	3280.7	-	-	6	0.6
Nb	3163	-	31	18	10
Sn	2840	-	-	24	4

*Sensitivity is given for rock analyses:

For biotite analyses, multiply by 2;
For ilmenite analyses, multiply by 4.

**Fe interference too high.

permitted only duplicate runs on each mineral separate. Values quoted are the arithmetic means of the duplicates.

Due to the high iron concentration in these samples precision is considerably less for many elements than in the biotite or rock analyses.

Biotite was observed as a contaminant in some of the ilmenite samples and visually estimated to make up less than 3 per cent of the sample. To check in a semi-quantitative fashion on the degree of contamination by this mineral, analyses were made (spectrographically) for Al and Mg. Up to 0.8 per cent Al_2O_3 and 0.1 per cent MgO are present in these samples. The ratio of Al to Mg is about 10:1 for all samples compared with a ratio of 2:1 in biotite. This suggests that some other source of Al contamination is present unless the ilmenites themselves contain as much as 1 per cent Al_2O_3 which seems unlikely. The maximum amount of biotite which could be present, assuming all the MgO to be in that phase, is about 1 per cent, suggesting that the visual estimates were high. Of course much of the MgO may be in the ilmenite itself (compare analyses in Vincent and Phillips, 1950). This amount of biotite would not affect the trace element concentrations for most elements although Ba and Mn may be contributed in measurable quantities.

D. OTHER ANALYTIC PROCEDURES

The following analytic procedures were also utilized in analysing the various mineral separates.

1. X-ray Diffraction

a. Potash-feldspar

The potash-feldspar separates obtained from the meta-arkoses were analysed by the technique of Bowen and Tuttle to determine their albite content. This technique involves the measurement of the 201 spacing of the feldspar. Two conditions must be fulfilled in order to assure the validity of these measurements:

1) the feldspar must be homogeneous, lacking exsolved albite;

2) the feldspar must have monoclinic symmetry.

The first condition is virtually satisfied by all the samples as observed in thin section. Less than 0.1 per cent by volume of lammelar perthitic albite was observed in thin section. No reflections from this phase were observed in the x-ray patterns. Assuming this phase to have been in solid solution at the time of last recrystallization, this would make the estimate of the albite content of the potash-feldspar low by about o.1 mol per cent.

To test the fulfillment of the second condition, a measurement of the separation of the 130 and $\bar{1}30$ reflections was made on three of the feldspars (McKenzie, 1954). For monoclinic (sanidinized) potash-feldspar these reflections should converge to a single 130 reflection. No peak separation was observed, indicating the feldspars to be monoclinic. The significance of this observation is discussed later.

The 201 spacing of the potash-feldspars was determined on a Norelco diffractometer by counting over the peak at intervals of 0.01 2θ. The time for a fixed count was determined at each reflection angle and plotted. From this plot the angle of maximum intensity can be determined to ± 0.003° 2θ. No internal standard was used. The measurement was reproducible over a period of several weeks to within ± 0.005° 2θ.

b. Biotite

Spacing of the 001 planes in the biotites separated from the meta-arkose was determined by the same method. Again no internal standard was used. The 003 and 004 reflections were observed and their d-spacings were converted to $c_o \sin \beta$. No attempt was made to determine the true length of the unit cell.

2. Optical determinations

Index of refraction measurements was made on the biotites from the meta-arkoses. Plagioclase compositions were determined from the minimum refractive index measured on grains lying on a cleavage by the method of Tsuboi (Winchell and Winchell, 1951, p. 280).

All measurements were made in Na_D light. The immersion oils were calibrated on an Abbe refractometer. All index measurements are considered to be accurate to ± 0.001.

An attempt was made to determine the 2V's of the biotites by measuring the maximum distance of separation of the isogyres with an ocular micrometer. The microscope used for this purpose was calibrated with a pegmatitic muscovite for which it was assumed that $2V = 45^{\circ}$. The optic angle is given by the equation (Wahlstrom, 1943, p. 155):

$$2V = 2 \sin^{-1} \frac{1}{2\beta K} D \qquad (1)$$

where β is the intermediate index of the biotite, D is the separation of the isogyres in scale divisions on the ocular, and K is an empirical constant given by:

$$K = \frac{D(\text{muscovite})}{2 \sin V(\text{musc.})(\beta \text{ musc })} \qquad (2)$$

For the range of 2V's investigated this relation is virtually linear.

E. STATISTICAL CONSIDERATIONS

1. X-ray Fluorescence

Duplicate analyses of the total rock and biotite samples by x-ray fluorescence give essentially identical counting rates for all elements except Mg (see table 7a). It is evident from this fact that the precision of a single analysis may be quite good. However, from the dispersion of points about the working curves it is evident that other factors result in rather low accuracy for most elements. Because only two analysed samples of the particular materials being studied (meta-arkoses and their biotites) were available it is impossible to determine to what extent this dispersion is intrinsic to the method of preparation and analysis of the samples and to what extent it is due to difference in matrix effects between the standards (taken from quite different rocks) and the samples. The only recourse is to accept the variance of individual analyses as obtained by the least squares fitting as an indication of the degree of resolution of the technique. For one element, Ca (in biotites) this results in being forced to discard all the analyses as being indistinguishable since the observed least squares standard deviation (0.08 per cent) is approximately equal to the mean concentration of the samples. In this case it has been decided to accept the x-ray analyses as valid, assuming that for a set of cogenetic biotites of approximately constant major element composition the values thus obtained will be a good estimate of the relative abundance of calcium in each sample, although the absolute amount is indeterminate. It is not unlikely that the standard deviation for wet-chemical analyses of

Ca in biotites would be of about the same magnitude, if it could be evaluated.

2. Trace-element Concentrations

The standard deviations for individual analyses of the trace-element concentrations of samples run in quadruplicate were obtained using the "linear estimate" of σ given by Dixon and Massey (1957, p. 405):

$$\sigma \cong 0.4539(x_4-x_1) + 0.1102(x_3-x_2) \qquad (3)$$

where x_i's are the observed concentrations listed in decreasing order. A similar formula was used for triplicate samples, namely:

$$\sigma \cong 0.5908(x_3-x_1) \qquad (4)$$

and for duplicate samples:

$$\sigma \cong 0.8868(x_2-x_1) \, . \qquad (5)$$

These standard deviations are listed in table 10 with the analytical lines.

For determinations of the fractionation factors between biotite and rock and between ilmenite and biotite, the standard deviation is given by the relation

$$\sigma(F) = \overline{F} \left(\left(\frac{\sigma_x}{\overline{x}}\right)^2 + \left(\frac{\sigma_y}{\overline{y}}\right)^2 \right)^{1/2} \qquad (6)$$

where F is the fractionation factor for a given element E, F = (conc. of E. in phase x)/(conc. of E in phase y), $\overline{F}$ is the mean of the fractionation values for that element, σ_x, σ_y are the standard deviations (obtained as above) for the concentrations in each phase and $\overline{x}$, $\overline{y}$ are the mean concentrations of E in phases x and y.

CHAPTER XII. CHEMISTRY OF THE ARKOSE

A. COMPOSITION OF THE TOTAL ROCK

1. Major Elements

The chemical analyses of the total rock as obtained by x-ray fluorescence are given in Table 12. They are plotted on the accompanying figure (fig. 28a) in sequence of increasing metamorphic grade. CIPW Norms have been computed for the two chemically analysed specimens.

The mean composition of the unit (average of 4-1 through 4-7, given in Table 15) is close to Daly's average granite composition. It is higher in silica and appreciably lower in total Fe, CaO and MgO. The Na_2O/K_2O ratio is somewhat higher. The other discrepancies could be accounted for by dilution of a granitic rock with quartz, especially if the granitic rock were somewhat less silicic than Daly's average.

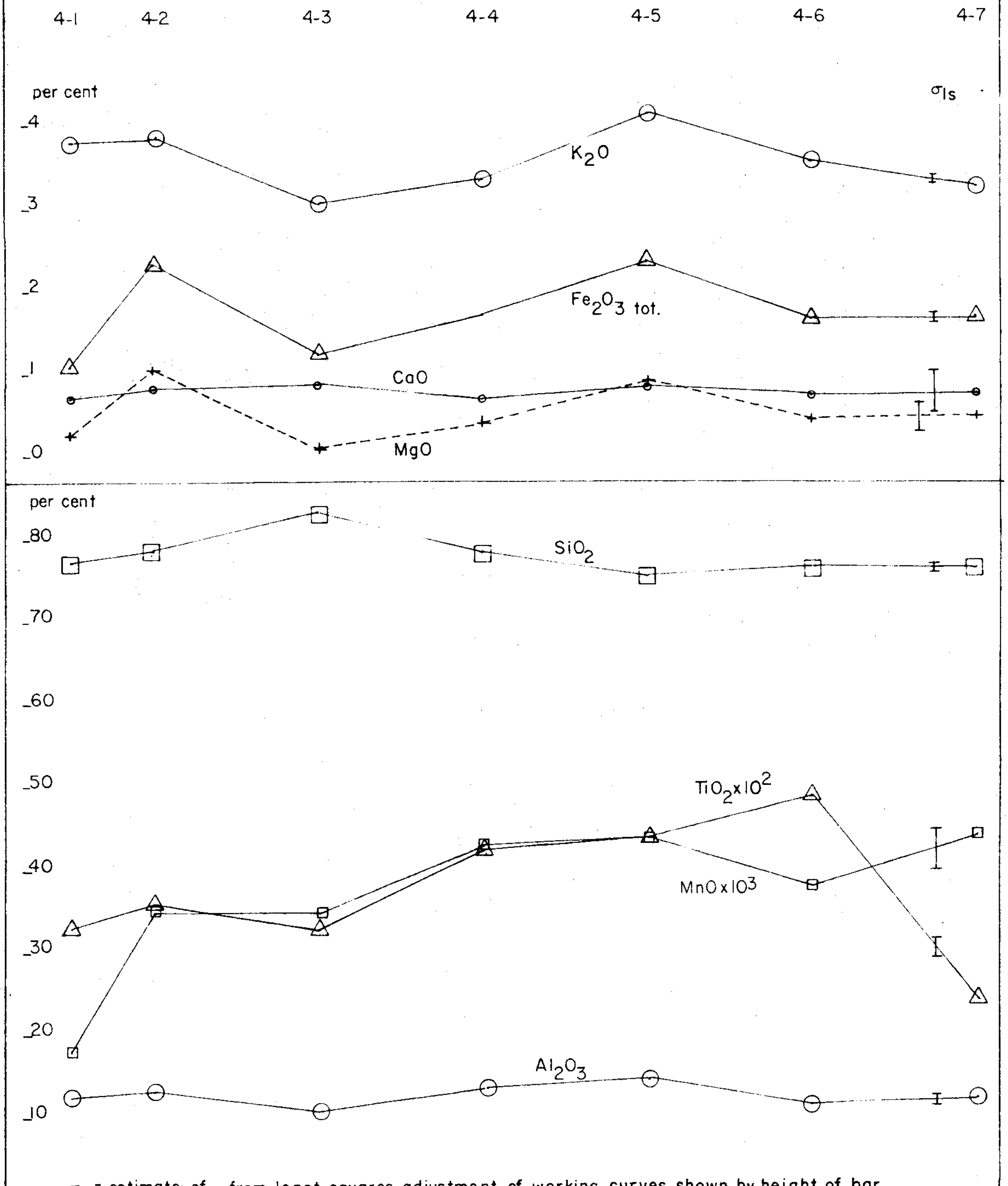
There is evidently no appreciable change in the abundance of the major components of this rock with increasing rank of metamorphism. This was to be expected since the modal analyses showed no consistent trends. What fluctuations are observed appear to be random in character, attributable to minor lateral compositional variation in an approximately homogeneous stratum. Sympathetic variation in Fe_2O_3 and MgO is due to these elements being bound together in biotite. However the systematic parallelism of these two elements shows that they must have been incorporated into the parent sediment in an approximately constant ratio, possibly as a single authigenic or clastic ferromagnesian mineral. This possibility is discussed further in relation to the chemistry of the biotite. Antipathetic variation in K_2O and Al_2O_3 versus SiO_2 corresponds

TABLE 12. Analyses of Major Elements of Meta-Arkose

	4-1	4-1G	4-2	4-3	4-4	4-5	4-6	4-7	4-7G
SiO_2	76.33	77.00	78.92	83.46	78.15	75.41	76.92	76.07	76.76
Al_2O_3	12.16	12.33	13.09	10.92	13.67	14.46	11.49	12.13	12.23
TiO_2	0.33	0.27	0.36	0.33	0.43	0.44	0.49	0.24	0.31
Fe_2O_3	1.24	0.51	2.38	1.28	1.72	2.37	1.65	1.70	0.36
FeO		0.51							1.14
MnO	0.018	0.01	0.035	0.035	0.043	0.044	0.038	0.044	0.04
MgO	0.29	0.26	1.05	0.12	0.41	0.94	0.47	0.48	0.49
CaO	0.29	0.70	0.81	0.90	0.71	0.86	0.77	0.76	1.03
Na_2O		3.71							3.47
K_2O	3.85	3.84	3.87	3.07	3.38	4.14	3.57	3.26	3.26
P_2O_5		0.08							0.09
H_2O^+		0.23							0.35
H_2O^-		0.03							0.00
	94.51	99.48	100.52	100.11	98.51	98.66	95.40	94.68	99.53

G = Doris Thaemlitz, analyst

FIG **28a** Chemical analyses of the meta-arkose (X-ray).



σ_{ls} = estimate of σ from least squares adjustment of working curves, shown by height of bar.

to varying proportions of feldspar and quartz. Presumably Na_2O would show similar antipathetic relations to quartz.

The two minor components analysed by this technique, TiO_2 and MnO, show a parallel but irregular increasing trend from lowest to highest rank assemblages. These may represent accidental changes in composition due to sedimentary facies change. It is possible however that these changes are due to mild metasomatism concurrent with the metamorphism of these rocks. The metasomatic material may perhaps be derived from the adjacent biotite schists. Mn^{2+} and Ti^{4+} ions may have been trapped out of a pervasive pore fluid present during metamorphism, especially if some phase already present in the rock became an increasingly acceptable host to these ions with increasing temperature of metamorphism. This is apparently true of biotite, as will be discussed below.

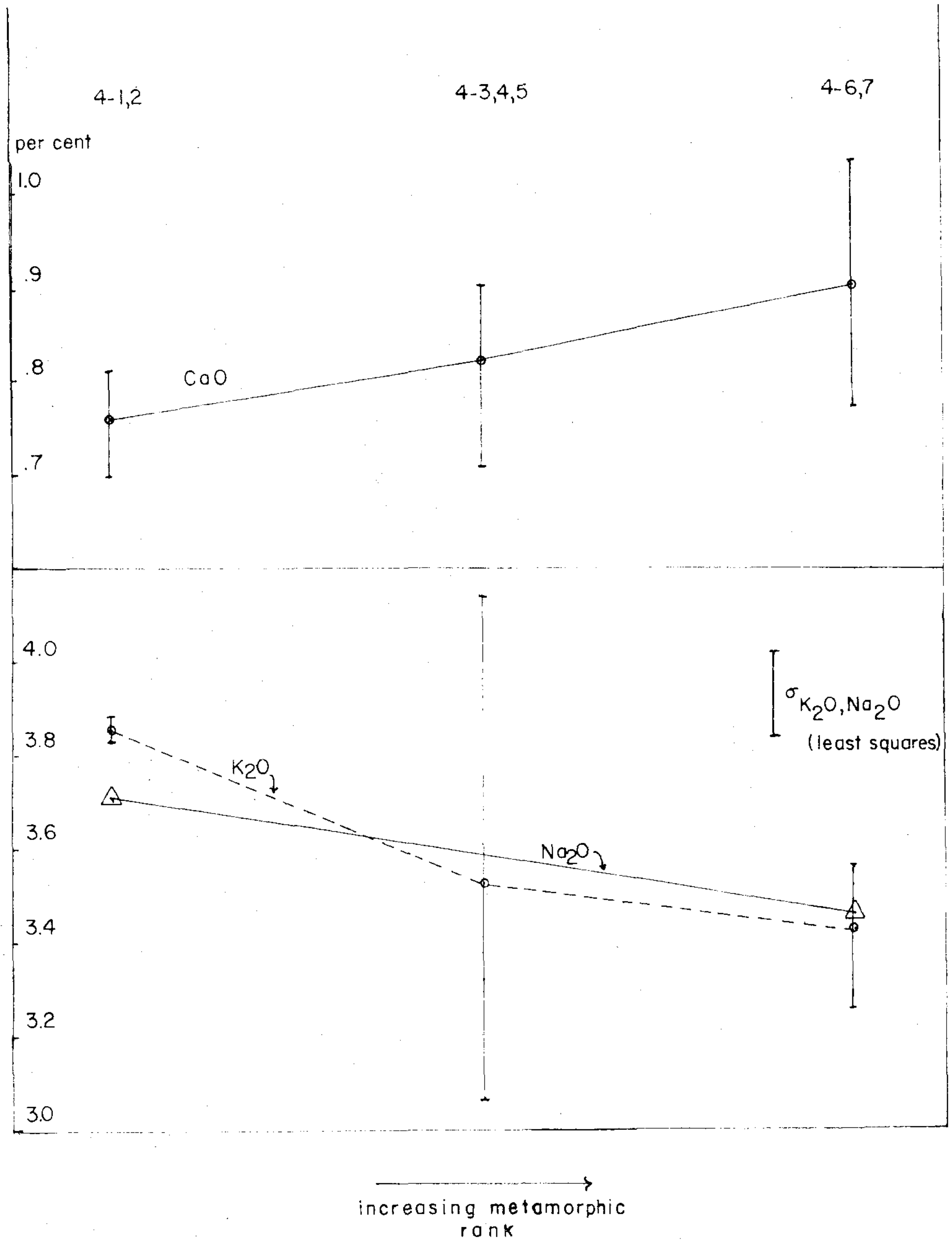
The CIPW norms (Table 8) are interesting in that both rocks contain small amounts of normative corundum. In granitic rocks it has been long observed that rocks with corundum in the norm usually contain muscovite in the mode. In this case the small amount of aluminum in excess of the alkalis has gone into biotite. In both norms the amount of plagioclase is considerably in excess of the modal abundances. This is in part due to the presence of some of the albite molecule in the potash-feldspar structure. In a large part this has been shown above to reflect inability to distinguish untwinned plagioclase from quartz in modal analysis. Hematite which appears in the norm of 4-1 may be partly present as free hematite in the rock, indistinguishable from the other opaque phases. It must also represent Fe_2O_3 incorporated in the biotite, with Al_2O_3 substituting for SiO_2 (or possibly $O^=$ for OH^-) to balance the charge excess.

In view of the rather large standard deviations of the chemical analyses by x-ray fluorescence it is apparent that small metasomatic changes with increasing metamorphism may be overlooked. This is especially true due to the small number of samples studied. The samples have been gathered into three groups, low rank (samples 4-1 and 4-2), middle rank (4-3, 4-4, 4-5) and high rank (4-6, 4-7) and analyses for each group have been averaged. In this way it is possible to detect suggestions of trends in the composition with rank of metamorphism. These trends are admittedly weak and would not be justified on the basis of a stringent statistical analysis of the data, but they are the best estimate one has available of the nature of the chemical variation in these rocks.

The alkalies and lime have been plotted in fig. 28b with such a grouping of the seven analyses. The only data available for Na_2O are from the chemical analyses of the two end-members of the series. The suggestion of this plot is that the alkalis decrease with progressive metamorphism while lime increases. Similar grouping of the analyses of the other elements shows no such consistent trends.

These trends in alkali and lime are comparable to similar variations observed by Engel and Engel (1953) in their study of a monotonous paragneiss subjected to progressive metamorphism in the Adirondack Mountains in New York. Similar behavior was noted by Ramberg (1951). To conclusively test the validity of such a trend would require a sampling procedure sufficiently broad in scope to overcome the effects due to the relict sedimentary variations in composition in the meta-arkose. Such

FIG.28b Grouped means of total-rock analyses.



effects have evidently not been overcome in this investigation.

The possible significance of such a variation has been discussed by Engel and Engel (op. cit.). This phenomenon is in accord with the deductions made earlier in the present study (Chapter IV) in regard to the origin of the migmatites. The progressive heating of the sediments has presumably permitted the selective mobilization of the lowest melting fraction of the rock and concentration of a more basic residue. This process of "degranitization" implies a dispersal of the liberated solute ions (K^+, Na^+) from the parent rock. In this instance one cannot however point to an obvious sink for this granitic material as one could in the case of the veinitic migmatites.

Although the x-ray fluorescence analyses do not yield any data on the oxidation state of iron in the samples the two chemical analyses yield some information on this point. The atomic ratio of Fe^{++}/Fe^{+++} changes from 1.11 in rock 4-1 to 3.42 in rock 4-7. This change is also reflected in the change of oxidation state of iron in the biotite of samples 4-2 and 4-7. Shaw (1954b) and Engel et. al. (1953) both report similar changes with progressive metamorphism. The possible significance of such changes will be discussed in relation to the chemistry of the biotites.

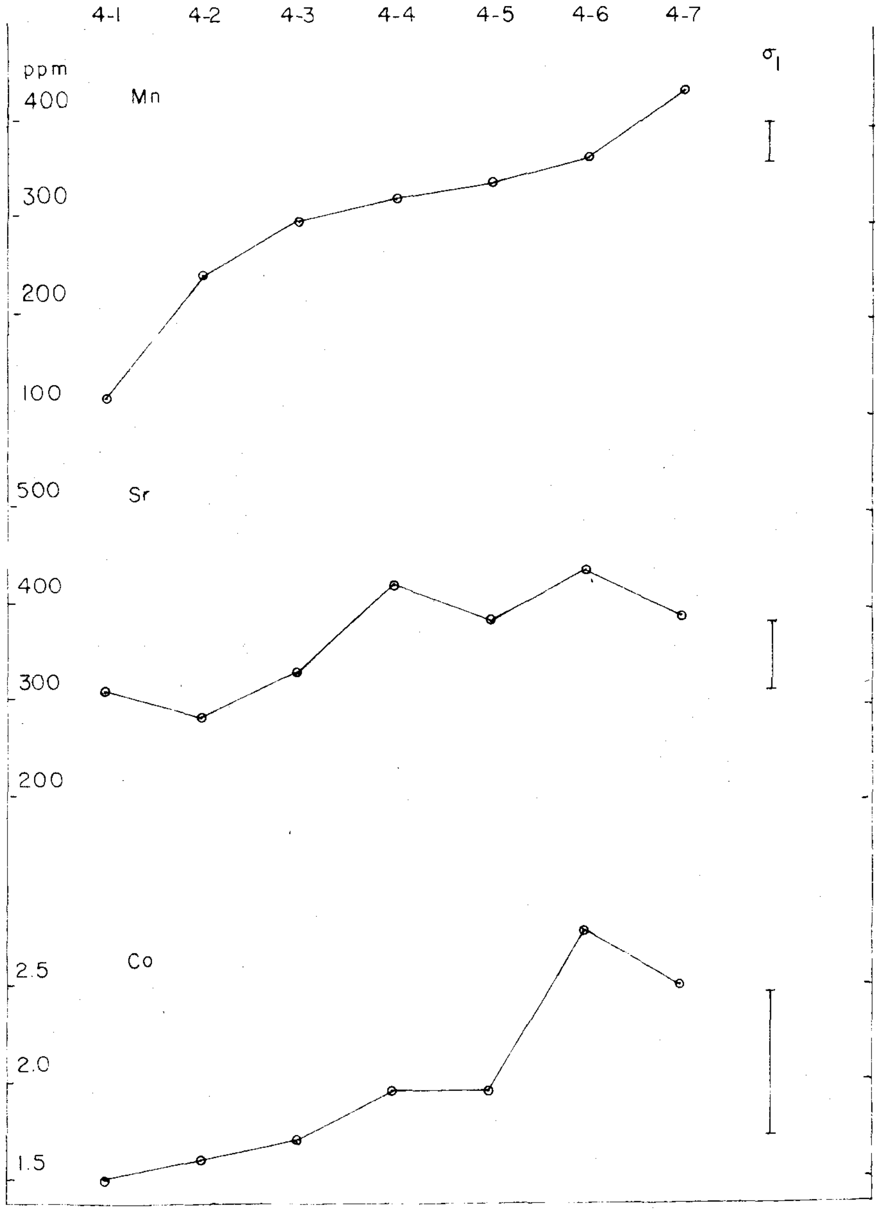
2. Trace Elements

The trace element composition of the meta-arkose is shown in table 13. The average composition is similar in most respects to the averages given by Engel and Engel for a middle rank paragneiss and to Shaw's values for the Littleton formation. Variations in individual elements are in the direction one would expect for dilution of a more biotite-rich metasediment with "neutral" quartz and with feldspar en-

TABLE 13. Trace Element Composition of the Total Rock (ppm)

	4-1	4-2	4-3	4-4	4-5	4-6	4-7	σ (ppm)
Pb	23	21	12	19	26	23	26	4
Mn	109	240	300	320	330	360	443	45
Ti	2250	2600	2555	2700	2500	3900	3400	225
Zr	183	188	233	378	338	607	276	76
Ga	9	9	5	9	11	9	11	2
Ni	2.1	4.5	2.0	4.9	4.0	4.1	3.3	1.1
Cu	6.5	2.9	7.9	5.8	3.7	4.1	5.9	1.4
Nb	10	10	10	10	10	14	10	-
Co	1.5	1.6	1.7	1.9	1.9	2.8	2.5	0.5
Yb	2.1	1.9	2.0	3.0	3.1	3.6	2.8	0.5
Y	19	13	21	30	27	33	28	6
La	38	46	45	35	56	70	61	8
Sc	5.8	7.3	5.2	5.1	7.9	7.5	5.1	2.2
Ag	1	1	0.5	1	1	1	1	1
Cr	10.4	18.3	14.5	16.5	19.0	16.0	16.1	3.8
V	29.2	21.0	15.5	16.3	20.3	16.7	15	5.7
Sr	310	280	330	420	380	440	390	70
Ba	1620	1660	1530	1650	1190	1680	1730	260

FIG.29 Trace element composition of total rock



σ_l = linear estimate of σ for replicate analyses, shown by height of bar.

FIG.29 Trace element composition of total rock.

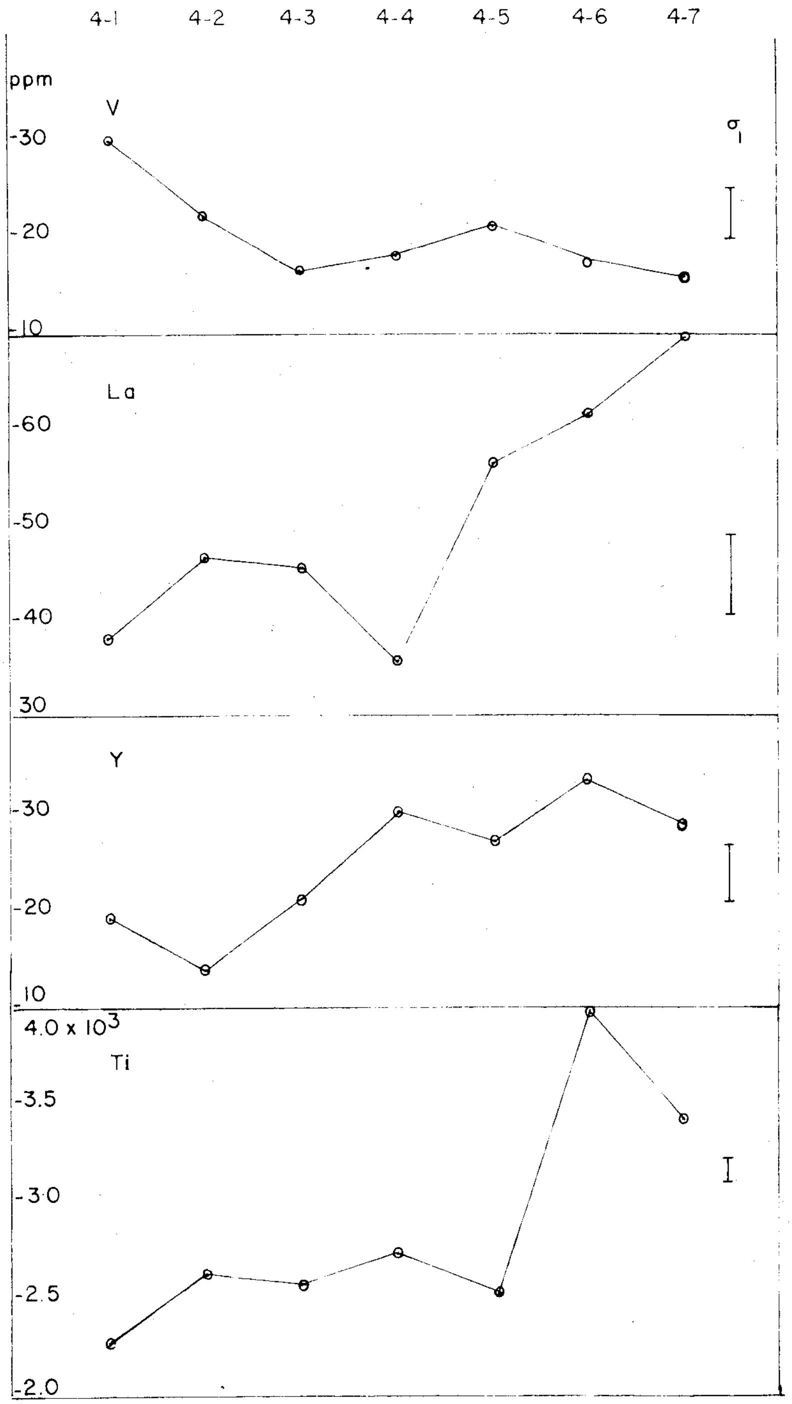


TABLE 14. Average Trace Element Composition of Total Rock

	1	2	3	4	5	6	7
Pb	21	12	15	16	27		
Mn	300	356	341				
Ti	2800	2800	3000				
Zr	315	171	220	191	203	125	150
Ni	3.6	15	21	64	57		
Cu	7	16	12	23	13		
Yb	2.6	3	3				
Y	24	46	58	39	52	15	30
La	24	tr				-	-
Sc	6.3	12	17	11	16	10	10
Ag	1	-	-				
V	19	56	81	125	120	29	75
Cr	16	35	56	113	109	3	15
Sr	335	310	304	524	760	250	250
Ba	1350	612	492			1000	700
Ga	9	11	12	16	20	20	20

1. Average of meta-arkose samples 4-1 through 4-7, Winchester-Hemet area, California. H. P. S. Analyst.
2. Average composition of least altered paragneiss, Emeryville area, Adirondack Mtns., N. Y. A. A. Chodos and E. Godijn analysts (Engel et. al., 1958). (Middle grade metamorphism)
3. Average composition of least altered paragneiss from Colton area, Adirondack Mtns., N. Y. Chodos and Godijn analysts, (Engel et. al., 1958). (High grade metamorphism)
4. Average composition of Littleton formation, middle grade metamorphism. (Shaw, 1954).
5. Average composition of Littleton formation, high grade metamorphism. (Shaw, 1954).
6. Granodiorite west of Lakeview, southern California batholith. (Sen, Nockolds and Allen, 1959).
7. Granodiorite phase of Bonsall tonalite, southern California batholith. (Sen, Nockolds and Allen, 1959).

riched in Sr, Ba and Pb. The latter elements are relatively high while Ni, Cu, Cr, Sc and V, elements readily entering into substitution for Mg and Fe in biotite, are slightly lower than for the pelitic schists.

For further comparison analyses of two granitic rocks of the southern California batholith by Sen, Nockolds and Allen (1959) are given. Again in most respects the values are comparable. Sr and Ba are closer to the concentrations observed in the meta-arkose. The high Zr content of the meta-arkose is a reflection of the slight concentration of clastic zircon in the parent sediment, due to its high resistance to weathering. Vanadium is markedly lower than either the metamorphic rocks or the granodiorites.

For most elements the concentrations are uniform within experimental error (1 σ) throughout the sequence of samples. However, as shown in plots of concentration against position in the sample sequence, selected elements do show trends in variation of composition. La, Y, Co and Sr show a tendency to increase while V, behaving rather erratically, may be inferred to decrease through the sequence. The trend of increasing Mn and Ti concentrations seen in the x-ray fluorescence analyses is borne out by the spectrographic data.

Such trends in composition may again be referred to trapping of mobile ions from the surrounding pore fluid of the rock at higher temperatures of metamorphism. The increase in Sr is sympathetic with the less marked increase in CaO for which SrO may substitute in the feldspar structure. Mn and Ti will be seen to increase in the biotite as their concentration in the rock increases. This mineral evidently serves as a sink within the rock for these two elements. It is not clear what the role of La, Y and V may be in the process of metamorphism.

Engel reports a tendency for Cr, Ca, Ni and V to increase through the action of progressive metamorphism and "degranitization". In the same series Ba appears to increase. These elements are virtually constant in concentration through the seven samples studied here. The apparent decrease in V is inexplicable. It should be noted that the maximum "degranitization" observed in this sequence as measured by loss of K_2O and Na_2O is of considerably smaller extent than that described by Engel et. al. It may be assumed to have involved correspondingly less migration of these trace elements.

Shaw notes increases in Ni and Cu and decrease in Pb with progressive metamorphism of the Littleton formation. No such trends can be observed here.

The behavior of trace elements in magmatic rocks have been largely explained on the basis of principles of ionic substitution by the various trace elements for major elements of similar ionic radius and charge. This results, in magmatic rocks, in parallel trends of trace element and major element evolution. In general the trace element composition of the meta-arkoses is not correlatable with major element concentrations of the total rock. This is reasonable since there is no reason to presume that the primary sedimentary minerals were homogeneous with respect to the degree of trace element substitution for major elements. For example Ni, Cr and Cu would be substituted to varying extents in clastic biotite and chlorite of the parent sediment depending on the sources of these minerals. In fact much of the bulk of these minerals may have been incorporated into the parent arkose as ions adsorbed on clay minerals or carbonaceous matter or as trace amounts of hydroxides or oxides mixed with hydrous iron oxide. One example of a consistent relation between a minor and major element is a correlation shown between Pb and K_2O in fig. 39. The ratio of Pb to K_2O is about 4-6 ppm Pb/per cent K_2O. This constancy of relative abundance presumably reflects the contribution of a single source of potash-feldspar to the primary sediment, a feldspar of fairly homogeneous trace element composition. We are neglecting here the smaller amount of Pb substituting for K^+ in biotite. Sr and Ba do not show such a correlation at all. The rare earths in the rock may be correlated fairly well with the Zr content of the rock (fig. 39). This may be in part a reflection of the fact that the rare earths, especially lanthanum, were brought into the parent sediment as trace elements in zircon. However it is apparent from the high ratio of La:Zr (about 0.2) that a considerable amount of the rare earth content of this rock must be otherwise accounted for. The La analyses of the biotite separates (given below) show amounts sufficient to account for the high La:Zr ratio of the rock. Apparently a large part of the rare earth content of the rock entered the parent sediment in other phases than zircon, perhaps as adsorbed ions or hydrated iron oxide minerals. The correlation between Zr and La + Yb suggests that zirconium itself may have entered the rock in this way and that, consequently, much of the zircon of the rock is authigenic.

3. Summary of the Chemistry of the Total Rock

The seven analysed samples of the meta-arkose unit are strikingly uniform in composition in both major and minor element constituents. The major element constituents SiO_2, Al_2O_3, total iron (as Fe_2O_3) and MgO show no significant variation through the sequence. That is, variation from the group mean (given in table 15) for all samples is not appreciably greater than the error for an individual analysis. For the elements K_2O, CaO and Na_2O (assuming the same relative variance for Na_2O as for K_2O) the differences along the sample sequence are also in most cases less than the error of measurement and only by means of a rather artificial grouping of samples is any suggestion of a trend in composition brought out. For the minor elements TiO_2 and MnO there is a measurable trend towards increase in abundance with rank of metamorphism. This is considered to be indicative of some metasomatic process at work during metamorphism.

With respect to the trace elements the picture of uniformity along the sequence is preserved for most elements studied. In general the composition of the rock is either uniform within one or two standard deviations or shows a consistent trend from one end of the sample sequence to the other. A trend of increasing concentration with metamorphic grade is shown for the elements Mn, Ti, La, Y, Sr and Co. For the last three named the magnitude in shift of concentration from one end of the sequence to the other is less than two times the standard deviation of a single analysis and the significance of the trends is questionable. Vanadium decreases in concentration along the sequence.

Summarizing these observations the unit which has been sampled, part of subunit fvB4 of the French Valley formation, is at present a very homogeneous rock although it has undoubtedly been subjected to a process of metamorphism which, in adjacent schists, has produced extreme changes in mineralogy and texture. To some extent the homogeneity may have been imposed on the rock during metamorphism. However, in view of the preservation of relict sedimentary textures in the low and middle rank specimens it seems unlikely that a great amount of internal redistribution could have gone on. The homogeneity is all the more remarkable in view of the obvious small-scale inhomogeneity represented by the disseminated shale clasts enclosed in the quartzite.

4. The Origin of the Meta-Arkose in the Light of its Chemical Composition

The composition of the meta-arkose studied here is close in most respects to the average of five arkoses given by Pettijohn (table 15, column 3). It is lower in total iron, has a higher Na/K ratio, is lower in CaO and somewhat lower in Al_2O_3. The single distinguishing feature clearly related to its metamorphism is the highly reduced state of the iron. In the meta-arkoses the ratio Fe^{++}/Fe^{+++} has been seen to range from 1.11 to 3.42. In the average arkose this ratio is 0.92. This is a result of a large part of the iron entering the sediment in its most oxidized possible state, having been in equilibrium with the relatively high O_2 concentrations of the atmosphere and sea-water.

In most respects the mineralogy of the meta-arkose is probably identical to that of the parent sediment. The abundances of the three major constituents, potash-feldspar, plagioclase and quartz are about what would be expected in an arenite of this bulk composition. Potash-feldspar is apparently stable in this compositional field at the lowest grades of metamorphism, although coexisting pelitic schists of this grade do not contain potash-feldspar.

The original clay mineral, clastic mica and oxide constituents of this rock have recrystallized to an assemblage of biotite and ilmenite. Only traces of magnetite are present, as would be expected from the very low Fe_2O_3 content. Presumably at a lower grade of metamorphism

TABLE 15. Average Composition of the Meta-Arkose

	1	2	3
SiO_2	76.8	69.9	76.4
TiO_2	0.37	0.39	0.41
Al_2O_3	12.6	14.8	10.6
Fe_2O_3	1.76	1.62	1.22
FeO		1.67	0.25
MnO	0.04	0.13	0.25
MgO	0.52	0.97	0.23
CaO	0.79	2.07	1.30
Na_2O	3.6	3.3	1.8
K_2O	3.6	4.07	5.0

1. Average of the x-ray fluorescence analyses of the meta-arkose, samples 4-1 through 4-7. Na_2O values based on two chemical analyses.
2. Daly's average granite (Johannsen, 1932, p. 193).
3. Average of five arkoses (Pettijohn, 1957, p. 323).

the constituents of the biotite were represented by an assemblage of chlorite and potash-feldspar. This assemblage is found preserved in low-rank arkosic quartzites of the Santa Ana Mountains, together with varying amounts of muscovite. The amounts of the respective constituents were evidently so distributed that all the chlorite reacted leaving an excess of potash-feldspar. The reaction forming biotite must in general use up some hematite or magnetite since biotite commonly has a higher Fe/Mg ratio than coexistent chlorite (Ramberg, 1952a, p. 141).

The primary feldspar component of the arkose would in the most general case have been derived from several sources, both igneous and metamorphic. As will be discussed further below, this feldspar component is very homogeneous throughout the rock except insofar as it varies with metamorphic grade. The Na_2O content of the potash-feldspar from all specimens analyzed, even the lowest rank (4-1) is such as to produce a single sharp 201 reflection in x-ray patterns. Any perthitic intergrowths that may have been present in the potash-feldspar component of the parent sediment have also apparently been re-dissolved in the host feldspar. Plagioclase grains from a single specimen are uniform in composition to within 2 per cent albite molecule, although there is some variability in the degree of albite twinning. This apparent homogenization of feldspar from inferred diverse sources is all the more remarkable in view of the preservation of relict angular clastic grains with a wide range of grain size. In view of these facts it seems more reasonable to attribute this feldspar component to a single fairly homogeneous source. To some extent the homogeneity of the feldspar from inferred diverse sources is all the more remarkable in view of the preservation of relict angular clastic grains with a wide range of grain size. In view of these facts it seems more reasonable to attribute this feldspar component to a single fairly homogeneous source. To some extent the homogeneity of the feldspar may be due to the persistent action of chemical weathering in removing more calcic plagioclase from the parent sediment. However the high total abundance of feldspar in the rock suggests it to have been derived from a rather immature sediment, not subject to prolonged chemical weathering.

An alternative explanation to account for the remarkable homogeneity of feldspar in the total rock is that part or all of the rock may have been derived from a volcanic source. Although this hypothesis has many favorable features the overall stratigraphic and petrologic character of the subunit fvB4 argues against this origin, as was discussed in Chapter II, Part I. Granting that the possibility exists, it is to be noted that the chemical composition of the rock fits a possible igneous composition and extrusive (and pyroclastic) rocks of this composition are known to exist, though none have thus far been recognized in the metasedimentary rocks of the Peninsular Ranges.

B. BIOTITE

1. Introduction

Biotite is present as the major mafic phase in all the arkose samples. It varies in abundance from 7 to 15 per cent by volume of the total rock. This phase serves as a host for most of the MgO and FeO of the rock as well as for much of the TiO_2 and MnO. Study of its crystal structure indicates that it should be a likely host for many trace elements. These may substitute for example for (Mg, Fe) or for K depending on their charge and ionic radius. One of the purposes of this study was to determine the extent to which such substitution is affected by the processes of progressive metamorphism. It is not expected that much variation in the ratio of Mg/Fe would be found since, except for ilmenite, which shares FeO, biotite is the only phase into which these elements would enter in major amounts. Likewise, except for the subtle effects due to balancing in the Al:Si ratio in plagioclase, it was not expected that much change in Al_2O_3 or SiO_2 would be observed since the only other aluminous phases are feldspars (which are observed to change little if at all with progressive metamorphism). The relation of the structure of the biotite to the observed chemical changes is discussed below.

2. Major Element Chemistry

The two chemical analyses and the x-ray spectrographic data are compiled in table 16. The compositional variation of the biotite is small through the sequence of progressively metamorphosed samples. Using the two chemically analysed biotites, the molecular proportions in each coordination position have been calculated (table 17). The octahedral and tetrahedral positions are respectively the sites of Mg^{2+} in the ideal phlogopite structure and Si^{4+} in that structure. Variations in the ratio of Si to Al are reflected in variations in the Al content of both sites. The "large cation" position is that of K^+ in the ideal phlogopite structure, lying between two sandwiches of tetrahedra and octahedra. Similar calculations have not been made for the x-ray analyses due to

TABLE 16. Major Element Composition of Biotites From the Meta-Arkoses

	4-1	4-2	4-2 G	4-3	4-4	4-5	4-6	4-7	4-7 G	σ_x
SiO_2	35.41	37.35	37.41	34.77	36.69	35.97	35.70	35.75	36.33	0.70
Al_2O_3	18.89	16.21	16.35	17.78	14.37	16.72	15.28	20.22	18.40	1.35
TiO_2	3.16	2.89	2.68	3.65	4.64	3.80	4.11	3.70	3.42	0.27
Fe_2O_3	21.31	22.71	6.97	21.23	21.35	22.96	22.46	22.25	2.63	0.72
FeO			13.38						17.33	
MnO	0.25	0.31	0.31	0.33	0.41	0.38	0.34	0.45	0.47	0.003
MgO	8.88	8.82	8.53	10.55	9.47	8.33	8.77	8.64	7.62	0.51
CaO	0.14	0.12	0.10	0.10	0.19	0.08	0.07	0.06	0.08	0.084
Na_2O			0.19						0.14	
K_2O	9.61	8.84	8.96	9.54	8.65	9.53	9.73	9.75	9.06	0.65
Rb_2O			0.07						0.08	
H_2O^+			3.69						3.35	
H_2O^-			0.36						0.10	
F			0.41						0.39	
Cl			0.06						0.02	
P_2O_5			0.07						0.03	
C			pnd						pnd	
	98.32	97.69	98.83	97.14	96.38	97.87	93.45	101.16	99.28	
c sin β (Å)	10.067	10.069		10.068	10.069	10.067	10.062	10.054		
n_γ	1.646	1.643		1.648	1.646	1.642	1.651	1.647		

G = C. O. Ingamells, analyst.

pnd = present but not determined (graphite).

σ_x = standard deviation of x-ray analysis, estimated from least-squares adjustment of working curves.

TABLE 17. Model Biotite Compositions of Chemically Analysed Biotite Calculations Assuming Sum of Negative Charges = 22

	4-2	Net Charge On Layer*	4-7	Net Charge On Layer*
Tetrahedral				
Si^{+4}	2.828		2.838	
Al^{+3}	1.166	−0.16	1.161	−0.27
P^{+5}	0.005		0.001	
	4.000		4.000	
Octahedral				
Al^{+3}	0.290		0.517	
Ti^{+4}	0.152		0.199	
Fe^{+3}	0.396	+0.27	0.153	+0.70
Fe^{+2}	0.845		1.122	
Mn^{+2}	0.021		0.030	
Mg^{+2}	0.960		0.878	
	2.664		2.899	
Large Cation				
Ca^{+2}	0.009		0.007	
Na^{+}	0.027		0.023	
K^{+}	0.813	−0.15	0.892	−0.07
Rb^{+}	0.003		0.005	
	0.852		0.934	
Anion				
$(OH)^{-}$	1.861		1.728	
F^{-}	0.100	(−0.03)**	0.098	(−0.31)**
Cl^{-}	0.007		0.003	
	1.968		1.829	
($O^{=}$	0.032)		(0.171)	
excess charge		−0.04		+0.36

*Charge on structural layers of biotite, per unit cell, assuming anionic (negative) charges to be distributed as in ideal phlogopite structure, namely:

- 15 on tetrahedral layers;
- 6 on octahedral layer;
- 1 around large cation;

the charge due to $(OH)^{-}$, F^{-}, Cl^{-} and residual $O^{=}$ is added to the negative charges on the octahedral layer.

**Excess charge over +2.00 for anion position.

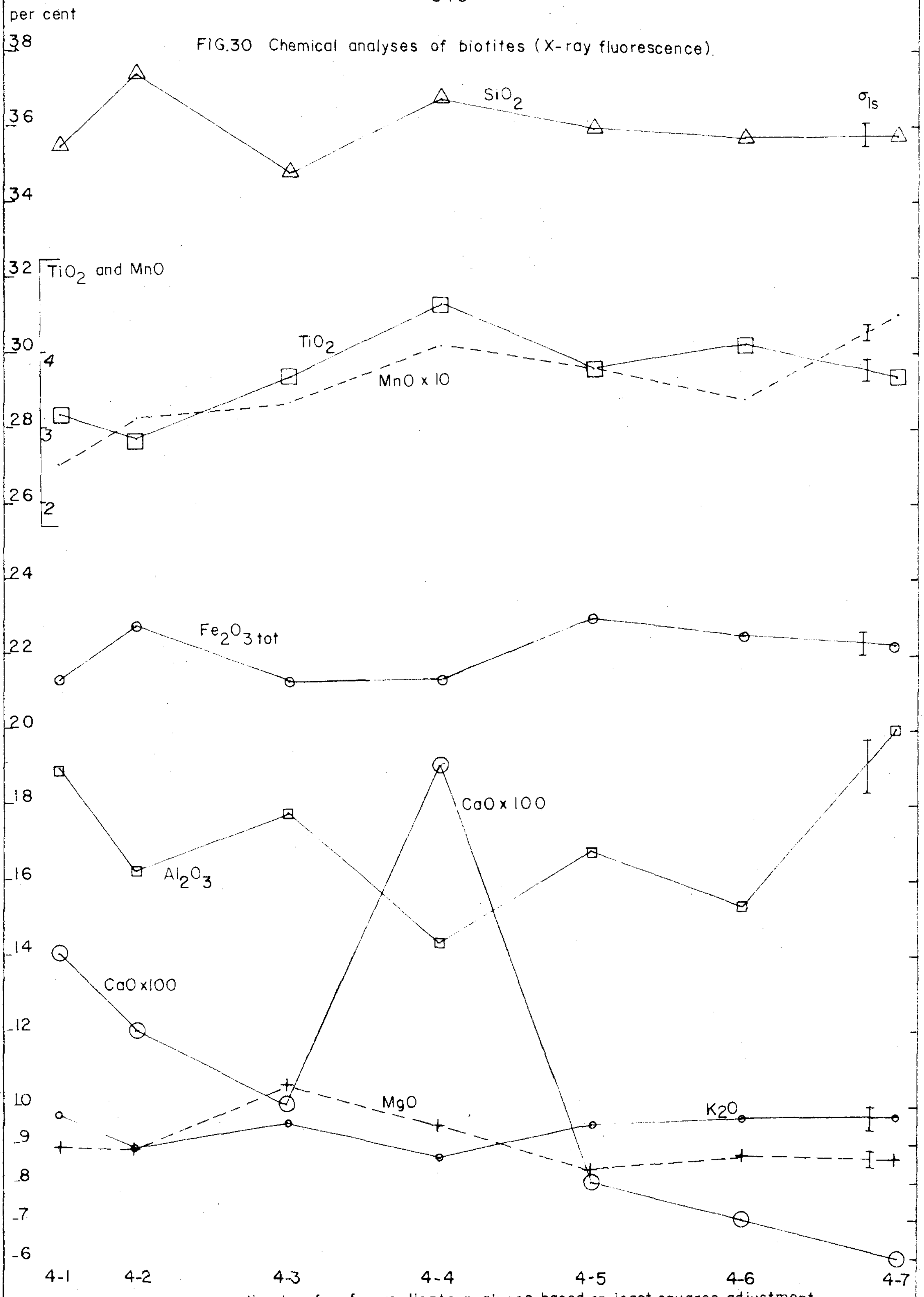
the lack of critical data (Fe^{++}/Fe^{+++}, H_2O, F, Na_2O).

The analyses have been plotted against their position in the sample sequence. The major constituents Al_2O_3, SiO_2, MgO, Fe (total) and K_2O as expected do not vary in any consistent fashion through the sequence. Their deviations from the mean values for the whole sequence are in most cases larger than one standard deviation (of the working curve). However these variations are apparently random and do not define any trend.

For the minor elements CaO, MnO and TiO_2 the situation is somewhat different. MnO almost certainly increases along the sequence (as is shown in the spectrographic data). CaO, with one exception, decreases along the sequence, from 0.14 per cent at the "cold" end to 0.06 (as determined by x-ray fluorescence) at the "hot" end.

Because of the low value of the CaO content there is some uncertainty as to its significance. The P_2O_5 content of one of the analyzed biotites (4-2) is such as to permit most of the CaO to have been present as fine inclusions of apatite. About 0.2 volume per cent apatite could account for the observed phosphate content. Very fine colorless inclusions have been observed in this biotite separate, but they do not appear to make up as much as 1/1000 of the volume of the biotite, though it is admittedly very difficult to estimate this quantity. For purposes of comparison biotite analyses taken at random from the literature (both igneous and metamorphic) show P_2O_5 contents ranging from 0.09 to 0.19 weight per cent. The ratio of P_2O_5 in rock to that in the biotite varies from 1 to 5 as described for various metamorphic rocks. In the two analysed meta-arkose samples it is respectively 1 and 3. However it is

FIG.30 Chemical analyses of biotites (X-ray fluorescence).



σ_{ls} = estimate of σ for replicate analyses, based on least squares adjustment, shown by height of bar.

not unlikely that other analyses of biotite may also err in this respect (due to inclusions of apatite). Because complete analyses are not available for the other biotites in the present suite of samples it is not possible to say if there is a consistent correlation between the phosphate and calcium content of the biotites. It is presumed that the CaO observed is present largely as structural Ca^{++} substituting in the 12-coordinated (K^{+}) position, though the preceding discussion must be accepted as a caution. As noted above (Chapter XI, section E) the x-ray values for CaO should be considered as reliable only in a relative sense, due to the large variance of the working curve.

TiO_2 varies somewhat irregularly through the sequence, suggesting an increase with metamorphic rank, although samples 4-3, 4-4 and 4-5 contain the largest TiO_2 contents, rather than the final two samples, 4-6 and 4-7. This trend is qualitatively confirmed by the spectrographic data. Ilmenite is present in all these rocks, indicating that the system is essentially saturated with TiO_2. (Actually this is not exactly true since a pure TiO_2 phase (rutile) is not present in most samples.) The conclusion one is led to is that, even in the presence of excess TiO_2, the amount of this molecule that can enter the biotite structure increases only to a limited extent with increasing temperatures of metamorphism and may actually decrease somewhat at the highest grade. The amount of CaO and MnO that can enter the biotite lattice at a given temperature would not be specified except for a given rock composition, since biotite coexists with phases of variable composition with respect to these elements (plagioclase and ilmenite respectively).

As to the FeO/Fe_2O_3 ratio in the biotites evidence is available only for the two end members. There is a marked increase in the ratio Fe^{++}/Fe^{+++} from sample 4-2 to sample 4-7. Considering the molecular proportion assignments to the "ideal" biotite molecules (table 17), it is noted that, in spite of the evident decrease in positive charge on the octahedral layer that would be produced by substituting divalent Fe for trivalent Fe to this extent, the actual net charge on this layer increases from 4-2 to 4-7. This is largely the effect of the additional Al^{3+} and Ti^{4+} entering into this coordination layer in the higher rank sample (4-7).

We may estimate the Fe^{++}/Fe^{+++} ratio for the total rock by the ratio in the biotite itself, and in this way obtain a value for rock 4-2 (chemical analyses are available for rocks 4-1 and 4-7). The assumption that all Fe is in the biotite is slightly in error, in that some Fe^{++} is present as $FeO \cdot TiO_2$ (ilmenite) in all the rocks, making the ratio somewhat higher than for the biotite alone.

TABLE 18. Atomic Ratios of Fe^{++}/Fe^{+++}

Rock 4-1	1.11
Biotite 4-2	0.71
Rock 4-7	3.42
Biotite 4-7	2.44

On the basis of this limited evidence there is some indication that the iron in the rock as a whole tends toward a more reduced state with increasing metamorphic rank. This is in agreement with the results of Engel et. al. (1953), Shaw (1954b) and others for pelitic schists. The effect here is more marked than in typical pelitic schists since the initial oxidation state of the iron in arkoses is commonly much more favored toward Fe^{+++} than in shales or greywackes.

Eugster (1959) has studied the equilibrium of annite and magnetite at elevated temperatures and P_{O_2} with $P_{tot} = P_{H_2O} + P_{H_2}$. He shows (fig. 9, p. 413) that for a constant P_{O_2} increasing temperature favors the reduction of iron in an assemblage of magnetite and anhydrous potassium silicates, forming annite. In the present case we are dealing with a system in which most of the iron is already present in the biotite as the annite molecule ($K\,Fe_3Si_3Al\,O_{10}(OH)_2$). It is not clear that the same generalization as to the change in oxidation state with increasing temperature can be made. The system we are dealing with here may be buffered (in the sense used by Eugster), in that the adjacent schists are rich in graphite. The partial pressure of oxygen in the meta-arkose during metamorphism may have been externally controlled by the reaction

$$C + O_2 = CO_2 .$$

3. Trace Elements

The spectrographic data are given for each element in table 19. Plots of some of these elements against position in the sampling sequence are shown in fig. 31.

For comparison the trace element contents of some biotites from igneous and metamorphic rocks are shown in table 20. It is evident that the trace element content is highly variable. In igneous rocks it has been shown (Sen, Nockolds and Allen, op. cit.) that the content of a given element is a function of the concentration in the total rock. In some respects one may note what appear to be general differences between the

TABLE 19. Trace Element Concentrations in Biotite (parts per million)

	4-1	4-2	4-3	4-4	4-5	4-6	4-7	σ(ppm)
Mn	1520	1460	3240	2180	1920	1800	3320	790
Ti	9800	8800	14800	17800	11600	15000	12600	1600
Zr	74	106	100	178	210	152	106	29
Ga	68	60	40	64	70	74	56	12
Ni	15.8	17.8	18.2	29.6	17.6	10.2	21.4	3.7
Cu	66	68	108	92	60	46	58	7
Nb	82	34	72	84	72	84	54	21
Co	14.8	12.0	13.6	15.2	12.6	11.8	16.0	2.1
Yb	8.4	3.0	6.2	11.6	11.4	7.2	6.8	2.0
Y	46	20	36	156	88	48	38	9
La	568	616	532	660	556	696	688	86
Sc	56	44	86	62	46	62	56	11
Cr	268	188	406	332	184	398	266	33
V	374	268	450	368	282	410	434	56
Sr	290	300	412	438	388	360	266	116
Ba	472	518	920	716	738	1086	706	133
Zn	242	208	298	208	252	164	301	62

FIG.31 Trace-element content of biotite and fractionation (F) between biotite and rock.

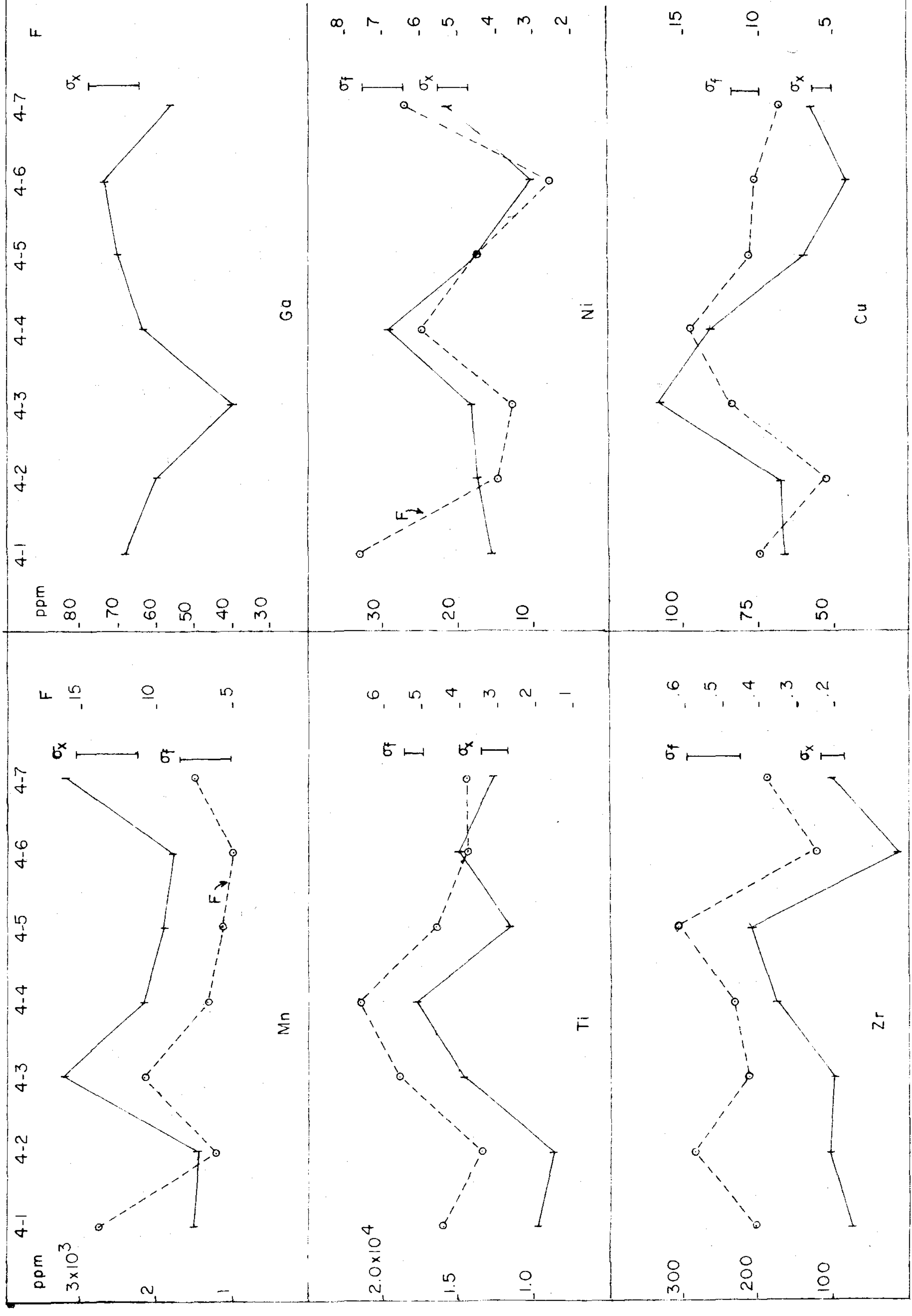
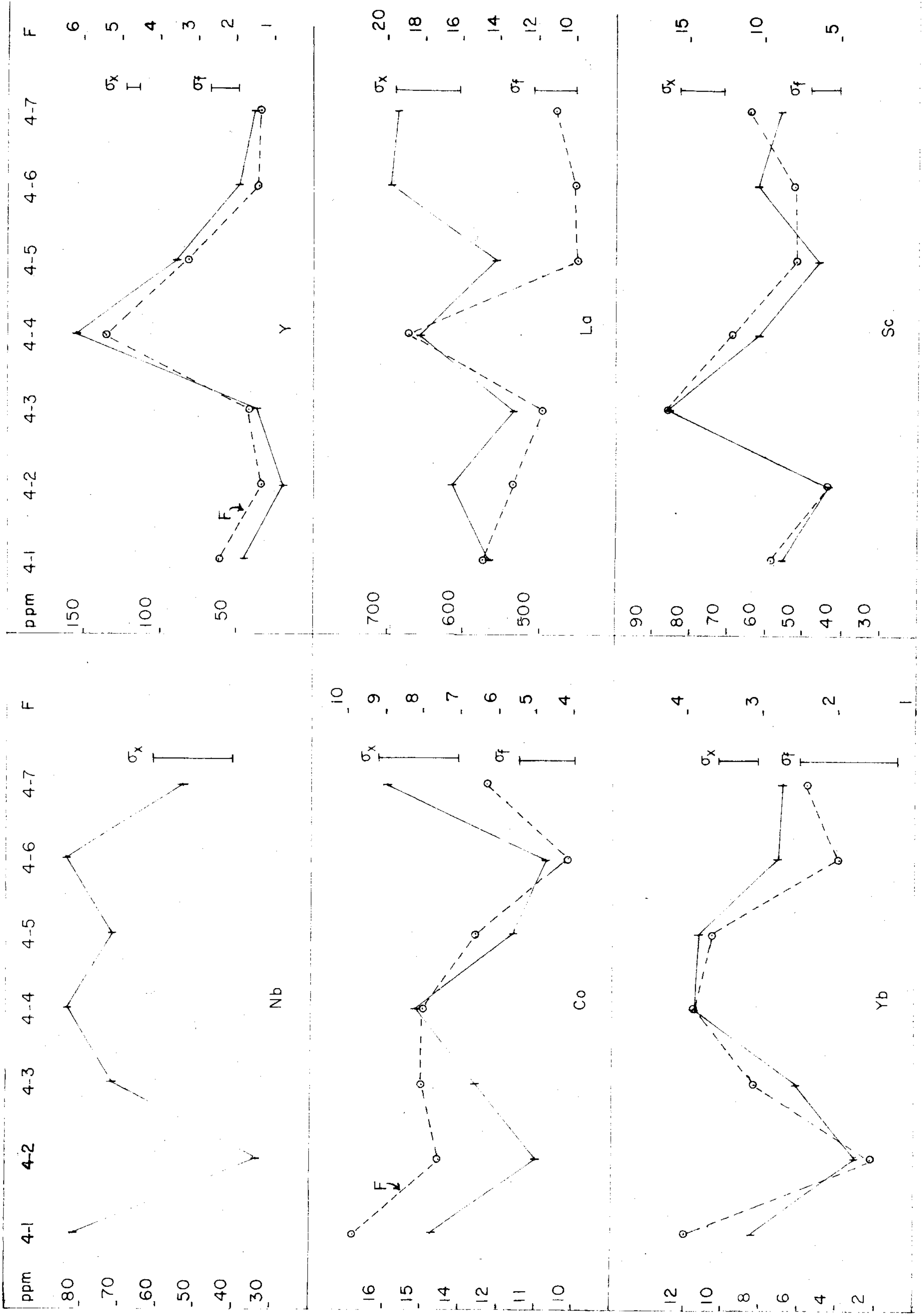


FIG.31 cont'd.


ppm
F
4-1
4-2
4-3
4-4
4-5
4-6
4-7
σx
σf
Y
La
Sc
Nb
Co
Yb
150
100
50
700
600
500
90
80
70
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50
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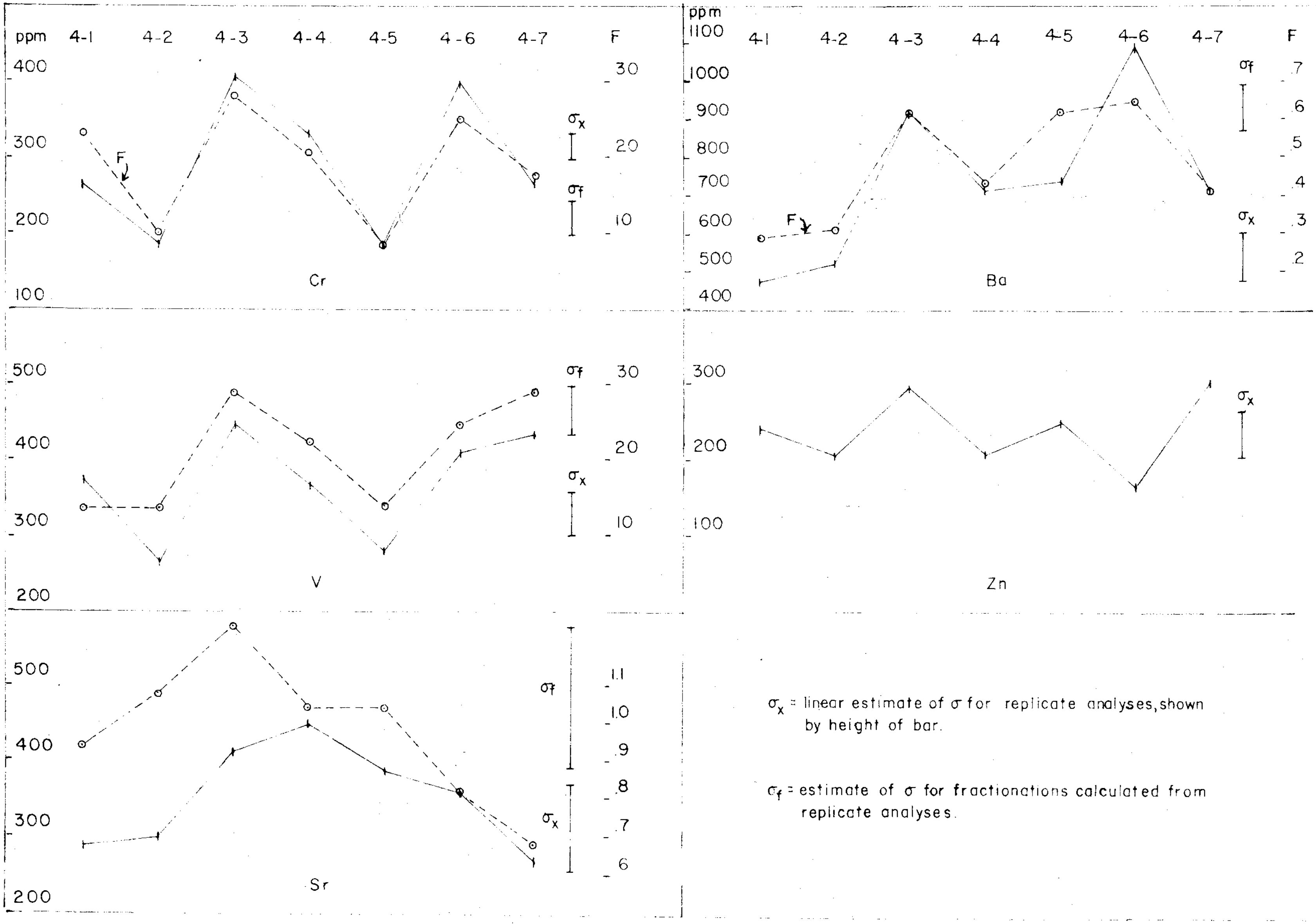
FIG.31 cont'd.



σ_x = linear estimate of σ for replicate analyses, shown by height of bar.

σ_f = estimate of σ for fractionations calculated from replicate analyses.

TABLE 20. Average Composition of Biotites: Trace Elements

	1	1a	2	3	4	5	6	7
Zr	132	0.223	40	45	350	90	100	620
In	50							
Ga	62	7.2	25	15	78	39	63	63
Ni	18.6	5.0	22	30	350	250	320	190
Mn	0.22%	8.0						
Ba	740	0.46	2500	1825				
Cu	76	11.4			360	150	40	88
Nb	68	–						
Co	13.8	7.2	35	50	125	90	90	95
Yb	7.8	3.0						
Cr	300	19	80	80	250	1200	1700	330
V	370	20.6	190	350	690	950	1400	580
Y	62	3.8	–	–	tr	–	–	nd
Zn	250	–			200	230	290	110
La	616	12.8						
Sc	58	10.0	7	5	31	30	40	85
Sr	350	0.98			30	tr	70	tr

1. Biotites from meta-arkose, Winchester-Hemet area.

1a. Fractionation (average) between biotite and meta-arkose host rock, Winchester-Hemet area.

2. Biotite from granodiorite in Bonsall tonalite, Sen, Nockolds and Allen (1959).

3. Biotite from Bonsall tonalite, Sen, Nockolds and Allen (1959).

4. Biotite from muscovite-biotite schist, Wissahickon formation, Philadelphia, George DeVore (1955).

5. Biotite from biotite-garnet-sillimanite schist near Virginia City, Montana, ibid.

6. Biotite from biotite-garnet schist near Gardner, Montana, ibid.

7. Biotite from biotite-garnet schist in Idaho Springs formation, ibid.

Note: All values except fractionations and Mn concentrations are in parts per million.

biotites from the meta-arkose and the samples given for comparison. Sr is higher in the arkose biotites than in either the typical igneous or metamorphic specimens. Ba is lower than in the igneous biotites in general although the Ba content of the meta-arkose is high. Ni, Co, Cr and V as given by Devore are all much higher than in the meta-arkose mica, probably reflecting the higher concentrations of these elements in pelitic schists such as he has sampled. The values for these latter elements obtained in the present study are about average or slightly lower than for biotites from intermediate rocks of the southern California batholith.

One purpose of this study has been to determine the extent to which the trace element content of biotite is a function of temperature of metamorphism. The plot of concentration of each trace element in biotite against its position in the same sequence (fig. 31) shows that for most elements there are large statistically significant variations from rock to rock. In general, however, these variations do not define a consistent trend from one end of the sequence to the other. When the analyses are grouped as before to damp out some of the variation due to local compositional features, certain elements, Cr, Co, V, La and Ba show an increasing trend with metamorphism, but in all but the Ba analyses the overlap between groups is such as to make such trends statistically of very low significance.

In order to study this problem of capturing of trace elements by the biotite more effectively, it would be best to isolate each mineral from the rock and analyse it separately for each component capable of entering the biotite. In this way we could determine the detailed redistribution of elements between the biotite and the other phases, independent of the total rock composition. For the present, lacking such detailed analyses, we can only compare the composition of the biotite to that of the whole rock. In a sense this is equivalent to considering the rock as a neutral matrix of a given trace element composition in which the biotite grains are embedded. The trace element content of the biotite should reflect the availability of each element from this matrix as well as the temperature, pressure and P_{H_2O} of metamorphism. In this respect three factors should be noted:

1) The modal and major element composition of the rock samples from which the biotites have been extracted does not vary appreciably along the sequence. The largest modal variation is in the abundance of biotite itself. The mineral assemblage is virtually constant.
2) The major element composition of the biotite does not vary appreciably along the sequence.

By appreciable variation is meant a consistent trend in composition resulting in differences much larger than the standard error of individual analyses. As noted above, there are randomly distributed variations in all these parameters, always of small magnitude, which could be attributed to variation in the composition of the parent sediment.

It is possible to take in account the effect of the composition of the host rock on the composition of the biotite by calculating the fractionation factor

$$F = \frac{\text{concentration of Element in biotite}}{\text{concentration of Element in rock}} .$$

These have been plotted together with the trace element analyses in fig. 31.

In calculating these fractionations it becomes evident that the errors on spectrographic analysis are sufficient to make the ratios unreasonably large for several elements. The fractionation factor F is given by

$$F = \frac{X_b}{X_b \cdot M_b + X_r \cdot M_r}$$

where X_b and M_b are respectively the concentrations of a given element in biotite and the abundance (weight fraction) of biotite in the rock. X_r and M_r are the trace element concentration and abundance of the non-biotite fraction of the rock. As X_r approaches zero, F approaches $1/M_b$ as a limit. Values of M_b are given by the modal analyses and $1/M_b$ is in general of the order of 10. The limiting values of F are given for each rock in table 21, along with the observed fractionations. For the elements Mn and Sc some F's exceed the limiting values while for V, Cr and La all values of F are too high. Although in fact only negligible amounts of these elements may be present in the rock outside of biotite, this effect is more likely due to the non-linear response of the spectrographic technique used to determine the abundances of these elements, causing lower values to be underestimated and/or higher values to be overestimated.

For most of the elements investigated the fractionation is much greater than unity. Presumably the greater part of the atoms of these elements now associated with the biotite entered that structure at the time that the biotite first crystallized and have not changed much in relative abundance with progressive metamorphism. However even for these

TABLE 21. Trace Element Fractionation Between Biotite and Meta-Arkose

$$F = \frac{\text{concentration (wt. \%) in biotite}}{\text{concentration (wt. \%) in rock}}$$

	σ_{est}	4-1	4-2	4-3	4-4	4-5	4-6	4-7
Cu	1.8	10.1	5.8	12.0	14.6	10.6	10.4	8.6
V	6.9	12.8	12.8	29.0	22.6	14.0	24.6	29.0
Zr	0.145	0.404	0.568	0.430	0.470	0.622	0.250	0.384
Cr	4.7	23.2	10.2	28.0	20.2	9.4	25.0	17.6
Yb	1.26	4.00	1.58	3.10	3.86	3.68	2.00	2.42
Y	0.67	2.42	<1.54	1.72	5.40	3.26	1.46	1.36
La	2.3	15.0	13.4	11.8	18.8	10.0	10.0	11.2
Nb	-	8	>4	>10	9	>8	6	>6
Mn	3.1	13.9	6.1	10.8	6.8	5.8	5.0	7.5
Ba	0.124	0.294	0.312	0.602	0.434	0.620	0.646	0.408
Sr	0.38	0.94	1.08	1.26	1.04	1.04	0.82	0.68
Co	1.5	9.8	7.6	8.0	8.0	6.6	4.2	6.4
Ni	1.1	7.6	4.0	3.6	6.0	4.4	2.6	6.4
Ti	0.4	4.4	3.4	5.6	6.6	4.6	3.8	3.8
Sc	2.0	9.6	6.0	16.6	12.2	5.8	8.2	11.0
Ba_{cor}	0.12	0.28	0.29	0.57	0.41	0.59	0.63	0.37
F_{lim}		11.4	8.3	7.9	8.2	6.0	12.8	9.5

$_{est}$ = estimated standard deviation calculated as shown in Chapter XI, section E.

Ba_{cor} = corrected fractionation for barium as described in text.

F_{lim} = limiting value of F, 1/(weight fraction of biotite).

elements there is some change noted both in the fractionation and the concentration. Vanadium for example, although about 20 times more abundant in the biotite than in the rock increases in abundance from about 330 parts per million to about 420 parts per million, and the fractionation varies from about 12 to 26-28 with increasing metamorphic grade. However for Mn, Co and La, each of which tends to increase in abundance in the biotite with metamorphic grade the trend of fractionation is a decreasing one. This is remarkable since except for minor amounts of these elements in ilmenite, they are largely concentrated in the biotite, as can be seen from the magnitude of the fractionation. If correct, these facts indicate that although concentrations of these three elements in the biotite are rising with grade, concentrations in the rock as a whole are rising even faster, a conclusion which is substantiated by the spectrographic data for the total rock. Evidently although only small concentrations are involved, the other minerals (quartz, feldspar, ilmenite and, in 4-7, garnet) are taking up increasing quantities at higher grades of metamorphism. This entire argument implies that the total rock composition changes as a result of some sort of metasomatism. It is again suspect to the extent that any evidence of metasomatism can be equivalently explained on the basis of relict compositional variation of the parent rock. The question of metasomatism is dealt with again in the conclusion of this chapter.

For those elements which have fractionations of the order of magnitude of unity the pattern is somewhat different. The elements Sr, Ba and Zr are relatively enriched in the rock as whole over the biotite.

As in the case of the P_2O_5 analyses, the zirconium analyses here are suspect in that they may be in part due to the presence of zircon as inclusions in the biotite. Pleochroic haloes are observed around some of the colorless inclusions in all samples of biotite; these inclusions are probably zircon. However they appear to make up much less than 0.01 per cent of the volume of the biotite. Even admitting that zircon is present as inclusions it is possible that they are a product of exsolution from the biotite and that the spectrographic determinations are a true measure of the zirconium concentration in the biotite lattice at the time of last recrystallization. However it is impossible to distinguish this from the alternative that zircon was present as inclusions at the time of crystallization. Admitting this last as a possibility, let us however consider the significance of the Zr values assuming them to be the equilibrium values for the conditions of last recrystallization of each rock sample.

Zr is also present in the rock as zircon, a pure zirconium silicate phase. The fractionation of Zr between biotite and rock is almost constant for all samples, although the absolute concentration varies widely. This is interesting since in theory the concentration of a component in a phase of variable composition in equilibrium with a pure phase of the same component should be only a function of pressure and temperature. Evidently the biotite "sees" not the pure zircon, but rather "sees" a pore-fluid in which the Zr concentration is a function of the total abundance of zircon in the rock. In this case the amount of zirconium available to enter biotite may be limited by the rate at which it could diffuse through the rock to the sites of biotite crystals. In this event, since the rate of diffusion would be proportional to the concentration of zircon, the amount of Zr^{+4} that the biotite would "see" would depend on the rock zirconium concentration.

The concentration of Sr in the biotite varies somewhat erratically along the sample sequence and is in general almost equal to the rock concentration. However the rock concentrations tend to rise in the higher-rank samples (fig. 29) and as a result the fractionation drops in these samples. This is in accord with the earlier observation that the Ca content of the biotite falls with increasing rank of metamorphism. These two elements are similar in chemical properties and ionic radius ($R(Sr^{+2}) = 1.12$ Å; $R(Ca^{+2}) = 0.99$ Å). Although definitive evidence is lacking presumably both elements are entering plagioclase since it is observed (e. g. Engel and Engel, 1958) that plagioclase in equilibrium with other Ca-bearing phases tends to become more calcic with increasing metamorphic rank.

Ba rises in concentration in the biotite with increasing metamorphic rank. This is accompanied by an increase in the fractionation from about 0.3 to about 0.5 since the Ba content of the rock is virtually constant. One must however correct the concentration of Ba in the rock for the amount contained in the biotite since much Ba is in other phases in the rock and the apparent fractionation (F as defined above) does not distinguish these two sites.

The corrected rock concentration is given by

$$C_r = \frac{C_R - (B \times C_B)}{(1 - B)}$$

where C_R is the apparent rock concentration, C_B the biotite concentration and B the modal abundance of biotite in the rock. Such a correction has also been attempted for elements which are highly concentrated in the biotite but the relative accuracy of determination of the rock concentrations is so much less than the accuracy of biotite concentrations that the resultant corrected rock concentrations vary erratically and are in some cases negative numbers. A part of the inaccuracy of such determinations probably lies in the inaccuracy of the modal analyses for biotite content.

When the concentrations of Ba are corrected in this way the trend toward increasing concentrations in the higher rank biotites is still present.

4. Summary of the Chemistry of the Biotite.

The major element analyses of the biotites reveal a degree of homogeneity consistent with the homogeneity of the total rock analyses given above. Since biotite is the only ferromagnesian phase present in the rock (neglecting the traces of ilmenite and garnet) the concentrations of $Fe_2O_{3\,tot}$ and MgO would be expected, in the light of the previous data, to be uniform. They have indeed been found to be so. Likewise, SiO_2, K_2O and Al_2O_3 vary by only a few per cent of the amount present along the sequence. For Al_2O_3 the variation is erratic and, in view of the great uncertainty in individual analyses, quite insignificant.

Titania is present in the biotites in major amounts. It exhibits a poorly defined trend of increasing concentration with metamorphic rank. Lime, present in minor amounts, may decrease by a factor of 2 from the "cold" to the "hot" end of the sequence, although once more the magnitude of error of a single measurement is large compared with the magnitude of the variation.

Trace element analyses of the biotites show erratic fluctuations from sample to sample along the sequence. These fluctuations are in many cases larger than two times the error of a single analysis. For the elements Zn, Nb, Co and Ga the variation is small with respect to analytic error. For no element is there a monotonic trend of concentration through the sequence, either increasing or decreasing. Assembling the samples into three groups of increasing rank and averaging the analyses in each group reveals a tendency toward increase in concentrations of Cr, Co, V, La and Ba. Only the last of these is statistically significant.

Ideally the fractionation of the various minor elements between the biotite and its host rock should be a monotonic function of the degree of metamorphism and therefore of its position along the sample sequence. In fact the elements Mn, Sr, Co, La and Yb display a tendency to become relatively less enriched in the biotite with respect to the total rock with increasing metamorphic grade. Vanadium and barium tend to become relatively more enriched with increasing grade. No element displays a strictly monotonic behavior in this regard. For many elements the variation in fractionation is small compared with the estimated error for a single determination.

In conclusion it is seen that for neither the major nor minor elements are there obvious well defined changes in composition with metamorphic grade. The relative homogeneity of the samples is somewhat less than that of the total rock samples from which they were extracted.

5. Optical Properties of the Biotite

The index of refraction of the biotite from these rocks (fig. 32) does not vary greatly through the sequence of samples. Grouping the samples as before it is seen that there is only a very slight rise in the mean n_γ from 1.645 to 1.649. The index of refraction appears to vary largely as a function of the TiO_2 content in these samples. A plot of n_γ against TiO_2 in mol per cent is shown. Also there is shown a triangular plot of the index vs. the major ions of the rock which may contribute strongly to its refractive index.* The contours of equal index on this diagram tend to be parallel to the Mg-Fe edge and normal to the direction of increasing TiO_2 content.

Values of 2V determined by the technique outlined in Chapter XI tend to scatter widely for a given sample of biotite. This may be due to refraction effects dependent on the thickness of the individual grains. The mean values for each sample lie on a decreasing trend with increasing degree of metamorphism of the host rock.

There is little change in the color of the biotites with advancing grade of metamorphism. Comparing the end members with one another (4-1 and 4-7) one may note a distinct greenish cast in the lowest rank sample, compared with a reddish brown color in the highest rank specimen. The transition along the sequence is so gradual that it is hardly noticeable from specimen to specimen.

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*Also shown on this diagram are values obtained by N. J. Snelling (1957) for biotites from pelitic schists of the Scottish Highlands.

FIG.32 Optical properties of the biotites

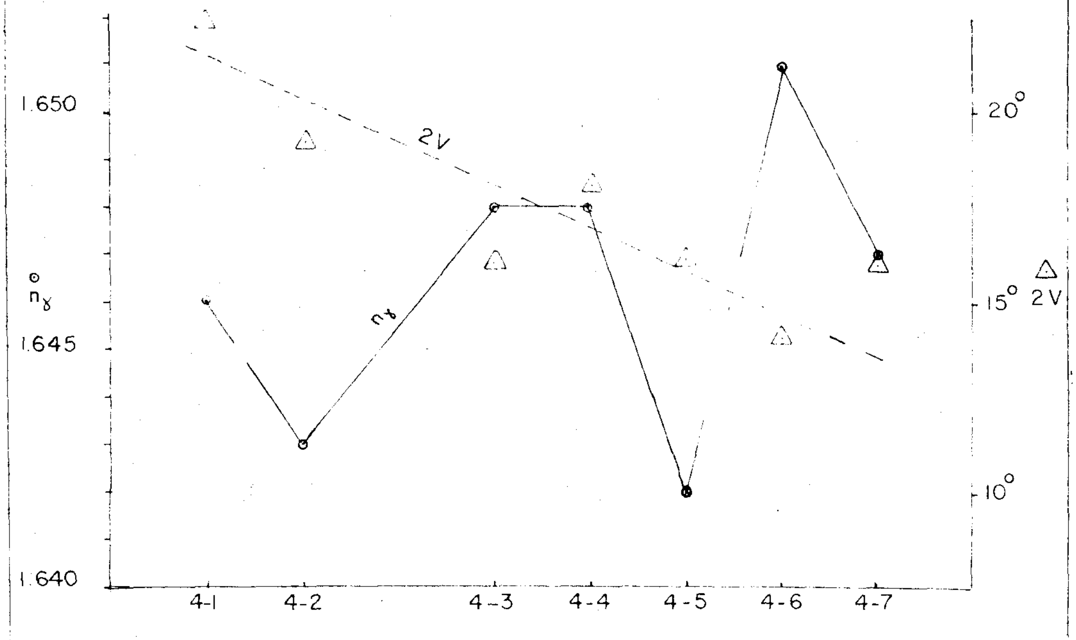
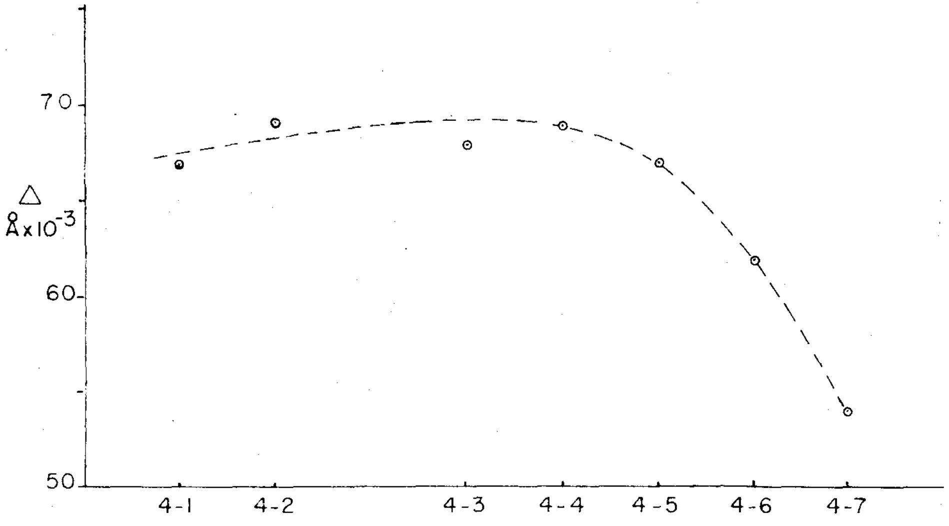
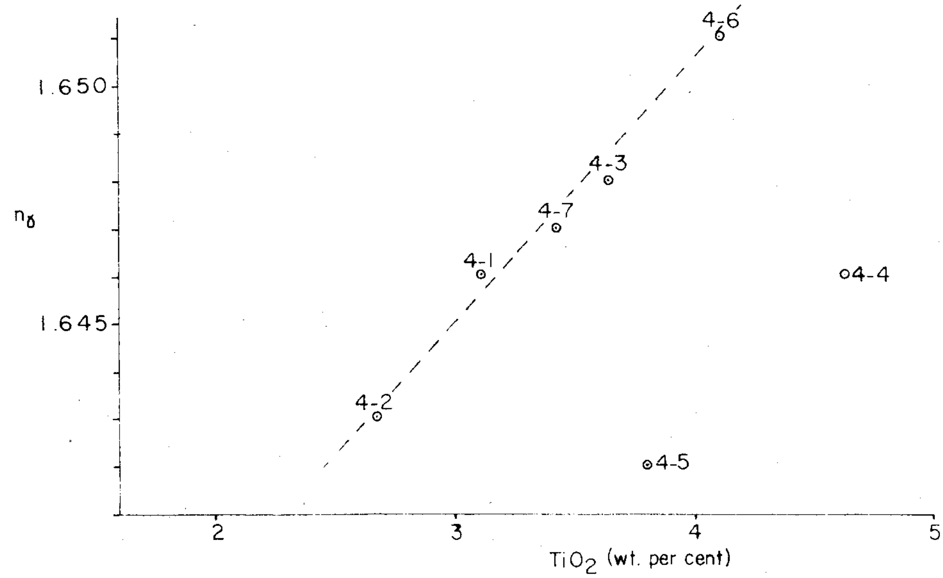


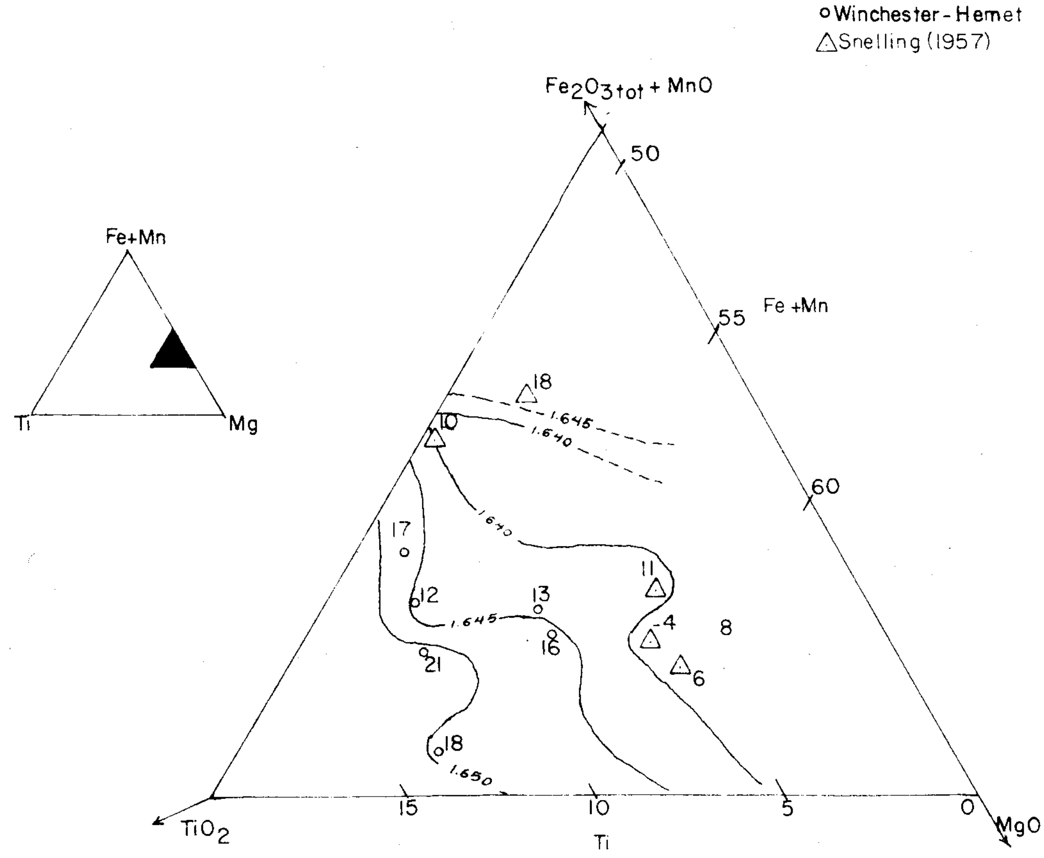
FIG.33 Basal spacing of the biotites



$\triangle = (C_o \sin\beta - 10.000) \times 10^3$

FIG.34 Relation between refractive index and composition of biotite.





number at each point is $n_\gamma - 1.630$

6. X-ray Diffraction Study of the Biotites

The basal spacing, $c_o \sin \beta$, is tabulated for each biotite in table 16 and plotted against its position in the sequence in fig. 33. It is seen that the two highest grade specimens fall off markedly although the deviations from the ideal spacing (about 10.1 Å) are still quite small. It is not clear what this change may be due to. The tetrahedral layers are probably quite rigid with respect to contraction normal to the basal plane. Contraction, if it is a result of change in chemical composition, is effected either in the "large cation" layer, containing Na, Ca and K or in the octahedrally coordinated layer. The changes observed in the large cation layer would tend, it would seem, to expand the lattice. The principal cation in this layer is K^+ (R = 1.33 Å). With progressive metamorphism Ba^{++} (R = 1.34) increases, Sr^{++} (R = 1.12) and Ca^{++} (R = 0.99) both decrease. In the octahedral layer, except for increase in Ti^{+4}, no marked compositional changes occur along the sequence and the ionic radius of Ti (0.68) is so close to those of Mg^{++} (0.67) and Fe^{++} (0.74) respectively that no great change would be expected. The lowering of the oxidation state of the iron ($R(Fe^{+++}) = 0.54$) would similarly expand the lattice.

As an estimate of the extent to which ionic size might be expected to influence the lattice spacing the following calculation has been made. We assume that in any given substitutional position (octahedral, tetrahedral, large cation and anion) an ion is present whose ionic radius is equal to the weighted mean radius of all ions observed in this position by the chemical analysis (table 17). Furthermore it is assumed that the lattice spacing normal to the layers of silicon-oxygen tetrahedra ($c_o \sin \beta$) varies linearly with the sum of these weighted average radii. This assumption neglects the effect of distortion of the lattice due to changing charge distribution as divalent, trivalent and quadrivalent ions substitute for one another. A calculation of the expected difference in basal spacing for the two chemically analysed biotites is shown in table 22. The calculated difference is negligible. This is all the more remarkable in consideration of the differences noted for individual ions (the column "$(a_2 - a_7) \cdot r$"). Any one of these differences if uncompensated would result in a difference in basal spacing as large or larger than the observed difference. And yet the partial differences, as these might be called, are each compensated for. Unfortunately complete chemical data on the other biotite samples are unavailable so that it is impossible to tell whether this is coincidental in this case. However the x-ray fluorescence analyses show differences in Mg, Fe, Ti and Al from sample to sample almost as great as the differences observed between samples 4-2 and 4-7 with no corresponding differences in $c_o \sin \beta$ (observed). Presumably the same sort of compensation may occur in these samples. Assuming that the weakest bonds should be the most easily stretched by substituting ions of different size into the various substituent positions, one would expect that basal spacing would be most sensitive to changes in the contents of the large cation and octahedral positions, the former more than the latter. Relatively little change takes place in the content of the large cation position for the two samples (and in general for metamorphic biotites). Therefore changes in the octahedral position should have the greatest effect in the determination of the lattice

TABLE 22. Theoretical Lattice Parameters of Chemically Analysed Biotites

		4-2			4-7	
	r (Å)	a_2	$a_2 \cdot r$ (Å)	a_7	$a_7 \cdot r$ (Å)	$(a_2 - a_7) \cdot r$ (Å)
Octahedral						
Al^{+3}	0.51	0.109	0.056	0.178	0.091	−0.035
Ti^{+4}	0.68	0.057	0.039	0.069	0.047	−0.008
Fe^{+3}	0.64	0.149	0.095	0.053	0.034	+0.061
Fe^{+2}	0.74	0.317	0.235	0.387	0.286	−0.051
Mn^{+2}	0.80	0.008	0.006	0.010	0.008	−0.002
Mg^{+2}	0.67	0.360	0.241	0.303	0.203	+0.038
			0.672		0.669	+0.003
Tetrahedral						
Si^{+4}	0.42	0.707	0.297	0.710	0.298	−0.001
Al^{+3}	0.51	0.292	0.149	0.290	0.148	+0.001
			0.446		0.446	0.000
Large Cation						
Ca^{+2}	0.99	0.011	0.011	0.008	0.008	+0.003
Na^{+}	0.94	0.032	0.030	0.025	0.024	+0.006
K^{+}	1.33	0.954	1.269	0.963	1.281	−0.021
Rb^{+}	1.47	0.004	0.006	0.005	0.007	−0.001
			1.316		1.320	−0.004
Anion						
$(OH)^{-}$	1.46[1]	0.930	1.358	0.864	1.261	+0.097
F^{-}	1.33	0.050	0.067	0.049	0.065	+0.002
Cl^{-}	1.81	0.004	0.006	0.002	0.003	+0.003
$O^{=}$	1.46	0.016	0.023	0.086	0.126	−0.103
			1.454		1.455	−0.001

Calculated difference in $C_o \sin \beta$ ..... = −0.002 ≅ 0 Å
Observed difference............... = −0.015

r: ionic radius of substituent ion, cations after Ahrens, anions after Zachariasen

a_2, a_7: fraction of total sites in one unit cell occupied by substituent ion in biotites 4-2, 4-7, respectively

[1] ionic radius of $(OH)^{-}$ assumed equal to that of $O^{=}$.

parameters of biotite. Changes in the anion position should have relatively little effect since the anion is essentially free to move normal to the basal plane of the biotite to accomodate the various sized ions in the octahedral position below it. However Gower (1958) has shown that for biotites of diverse chemical composition the relative abundance of fluorine in this position has the greatest effect of any chemical variable investigated in changing the lattice dimensions. This is evidently yet an unsolved problem. Since relatively little change in basal spacing occurs in the samples investigated here it appears to be unprofitable to attempt to explain this effect at this time.

Entry of trivalent and quadrivalent ions (Fe^{+3}, Ti^{+4}, etc.) into the octahedral site may, by increasing the net positive charge on that site, produce a distortion larger than that due to the differences in ionic size alone. This electrostatic effect is however already taken in account in part in the determination of the ionic radius (if determined in oxides or oxygen compounds). A similar effect may be introduced by the substitution of $O^{=}$ for $(OH)^{-}$ in the anion position though the importance of this position in determining the lattice spacing is probably small for the reasons given above. Estimates of the charge balance on the various layers of the two chemically analysed biotites are given in table 17.

All the biotites investigated are either 1M or 3T as indicated by their powder patterns (Yoder and Eugster, 1954). In view of their large 2V's they are probably monoclinic, 1M.

C. ILMENITE TRACE ELEMENT ANALYSES

Only minor element analyses of the ilmenites from rocks 4-1 through 4-7 have been made. Sample 4-1, processed in part in a brass sieve, probably contained some brass contamination and values of Cu, Pb and Zn for this sample should be disregarded. Each of the ilmenite separates was examined in a polished grain mount and found free of inclusions or exsolution lamellae.

The Cr and Ba concentrations, seen to rise in the biotite with increasing metamorphic rank, show a converse relation in the ilmenite (fig. 35). Similarly, Y and Pb show decreasing trends. The apparent decreasing trend in Zr may be in part or entirely due to minute inclusions of zircon in the ilmenite, decreasing in abundance in the higher rank coarser-grained samples. Nickel and manganese alone appear to increase with increasing metamorphic grade.

Fractionation factors between ilmenite and biotite have been calculated for most of the trace elements investigated (table 24). They vary erratically over a wide range from sample to sample and average values do not give a good representation of the behavior of the various elements. If we consider only one rock at a time, for example rocks 4-3 and 4-7, it is seen (fig. 36a) that the fractionation is determined largely by the charge of the ion and perhaps to a lesser extent by its ionic radius. Divalent ions (Cu, Ni, Co and Mn) are preferentially taken into the ilmenite, to a decreasing extent with increasing ionic radius. Trivalent ions (Cr, V, Sc, Yb and La) are taken into the biotite lattice to a slightly decreasing extent with increasing ionic radius. Nb

TABLE 23. Trace Element Composition of the Ilmenites (parts per million)

	(4-1)*	4-2	4-3	4-4	4-5	4-6	4-7	$\overline{X}$	σ (ppm)
Zn	1780*	880	940	1120	1060	1520	1020	1190	143
Zr**	1020	1460	820	960	1040	600	460	910	73
Co	104	28	24	28	28	32	26	28	3
Ba	132	22	36	40	23	12	7	23	4
Mn	5000	7650	9200	8800	9200	8800	9600	8300	750
Pb	312*	72	36	32	36	36	36	41	9
Sn	48	32	28	36	36	32	24	34	8
Ni	56	19	24	28	28	32	32	27	6.5
Nb	720	820	600	820	880	760	600	743	134
La	760	480	600	560	560	660	560	600	54
Cu	2000*	1240	360	480	1480	640	520	990	63
Ag	4.8	5.6	5.6	6.0	6.4	6.0	6.4	5.8	0.4
Yb	4	28	4	0	7	0	0	6	1.3
Y	660	644	216	480	276	144	228	378	49
V	212	84	76	96	80	92	72	83	28
Sc	68	52	56	60	76	68	48	61	11
Cr	760	256	140	28	24	30	28	84	18

*Sample 4-1 contains possible brass contamination. Values for Zn, Pb and Cu are probably in error.

**Possible zircon contamination in all ilmenite samples.

$\overline{X}$ = average of 4-1 through 4-7.

FIG.35 Trace-element content of ilmenite

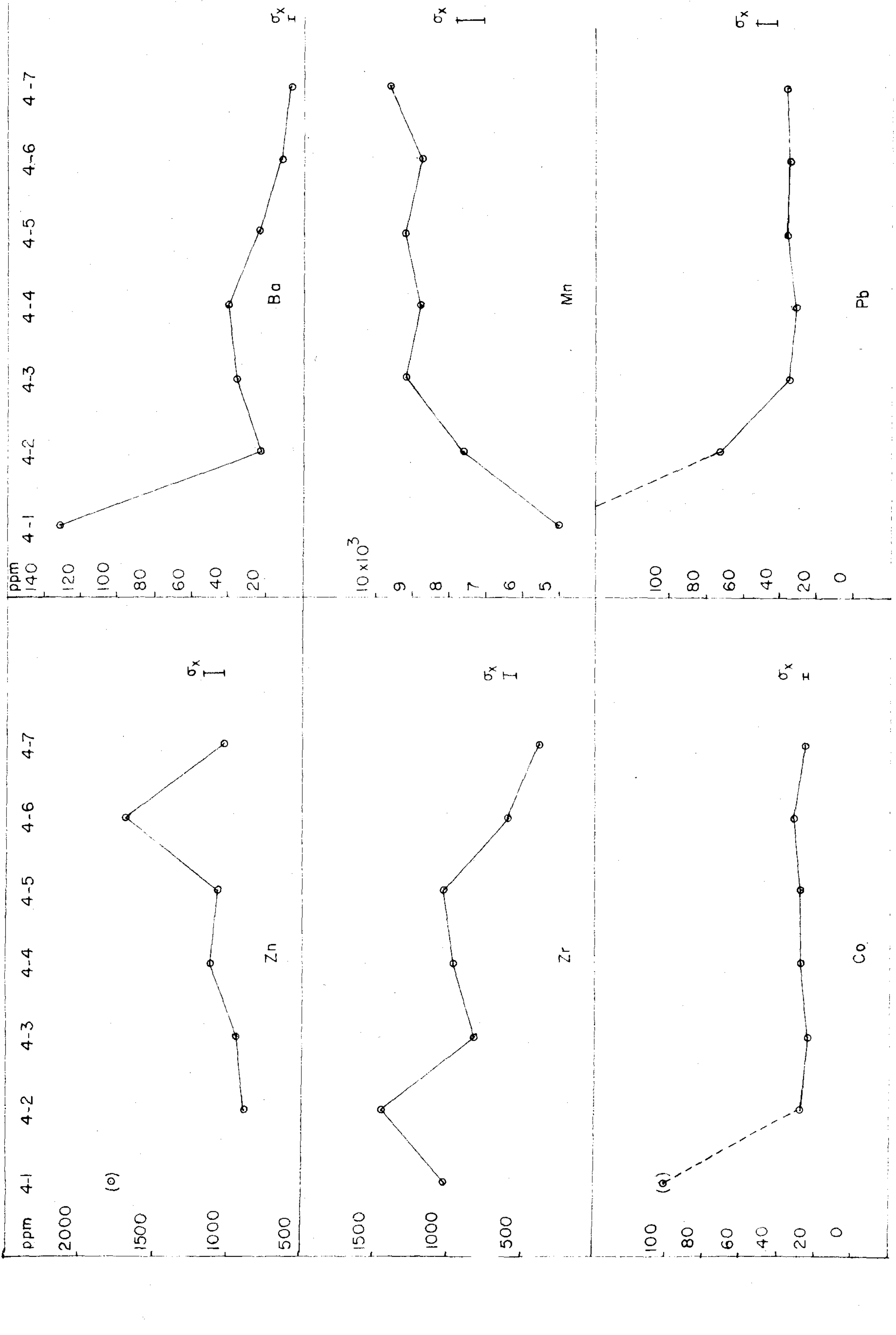


FIG.35 Trace-element content of ilmenites (cont'd.)

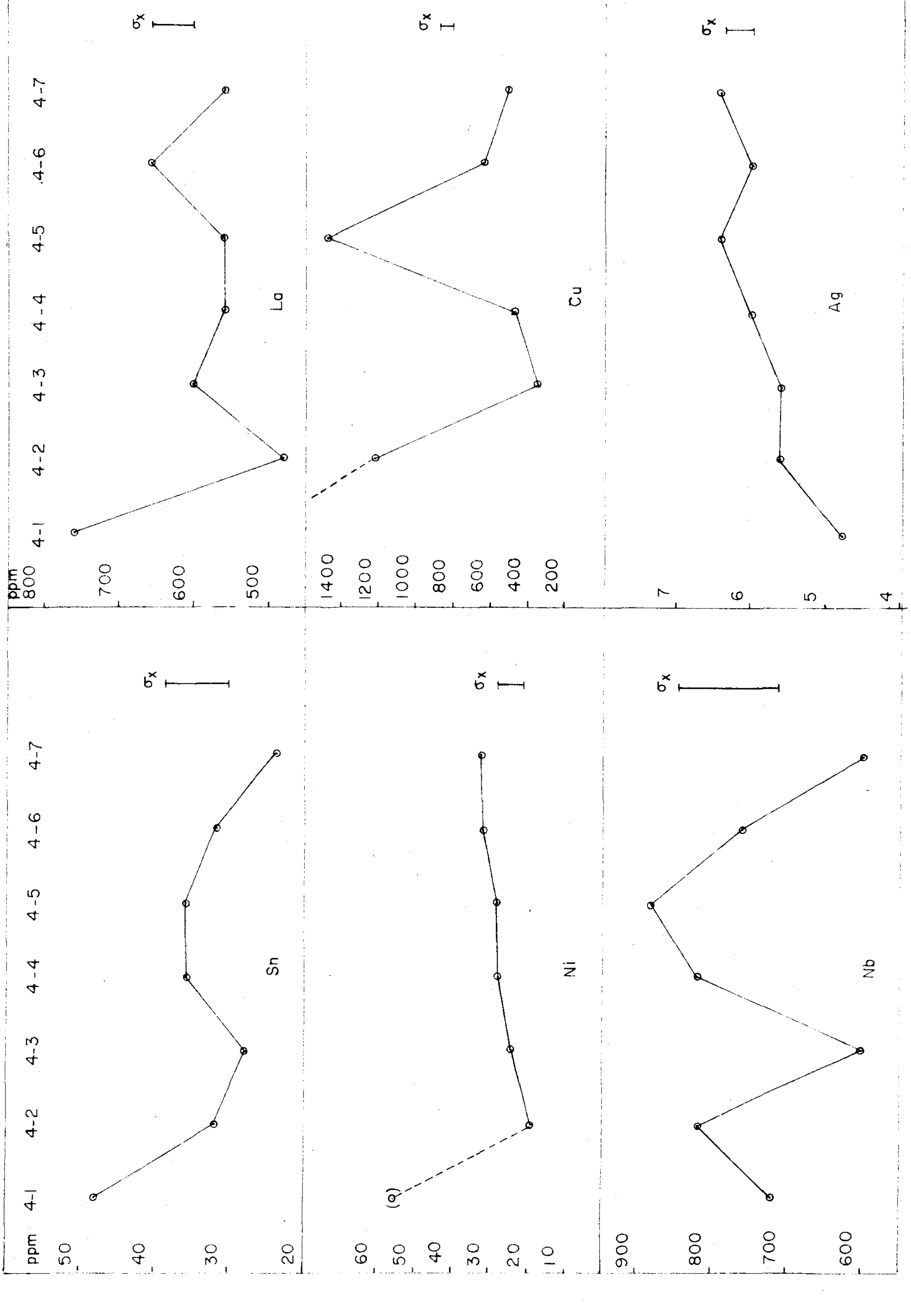
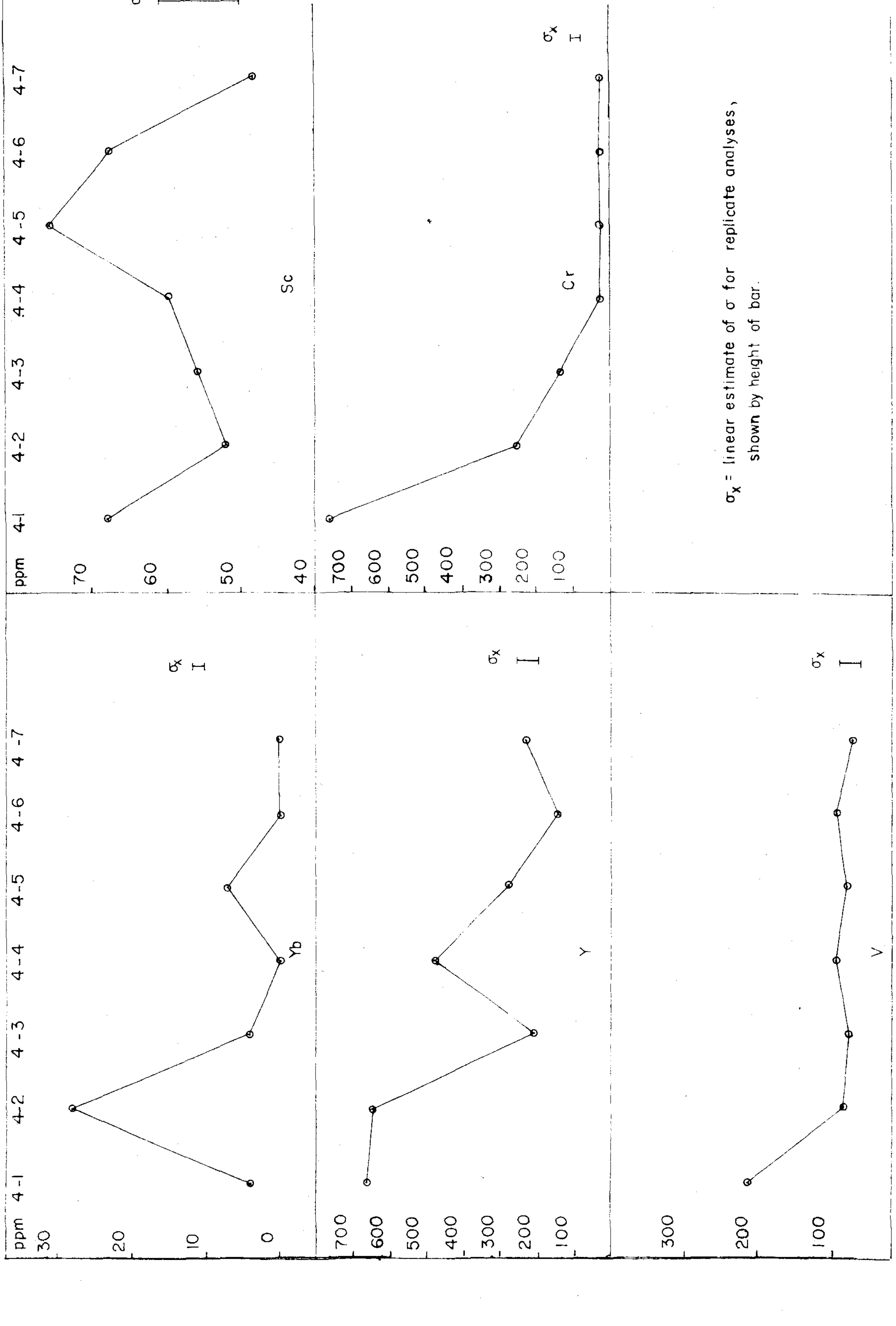


FIG. 35 Trace-element content of ilmenite (cont'd.).



σ_x = linear estimate of σ for replicate analyses, shown by height of bar.

TABLE 24. Trace Element Fractionation Between Ilmenite and Biotite

$$F = \frac{\text{concentration (wt. \%) in ilmenite}}{\text{concentration (wt. \%) in biotite}}$$

	σ_{est}	4-1	4-2	4-3	4-4	4-5	4-6	4-7
Nb	2.69	8.78	24.12	6.12	9.76	12.22	9.04	11.12
Mn	1.51	2.20	3.32	2.78	8.48	7.97	2.88	2.89
Ni	1.34	3.80	3.60	3.52	2.56	4.10	9.02	3.56
Cu	1.56	24	15.50	3.00	4.70	35.2	12.30	8.12
Yb	0.067	0.94	9.34	0.64	0.00	0.60	0.00	0.00
Y	1.21	14.34	32	7.66	3.08	2.46	3.00	6.00
V	0.10	0.56	0.32	0.16	0.26	0.28	0.22	0.16
Sc	0.29	1.22	1.18	0.66	0.96	1.66	1.10	0.86
Zn	1.51	7.52	4.24	3.16	5.38	4.20	9.26	3.16
Co	0.49	7.02	2.34	1.96	1.84	2.22	2.70	1.62
Cr	0.18	2.84	1.40	0.34	0.08	0.14	0.08	0.06
Zr	1.80	13.78	13.78	8.20	5.40	4.96	3.94	4.34
Ba	0.010	0.280	0.030	0.040	0.056	0.030	0.012	0.010

FIG.36a Trace-element fractionation between ilmenite and biotite in two samples of meta-arkose.

$$F\left(\frac{I}{B}\right) = \frac{\text{weight per cent in ilmenite}}{\text{weight per cent in biotite}}$$

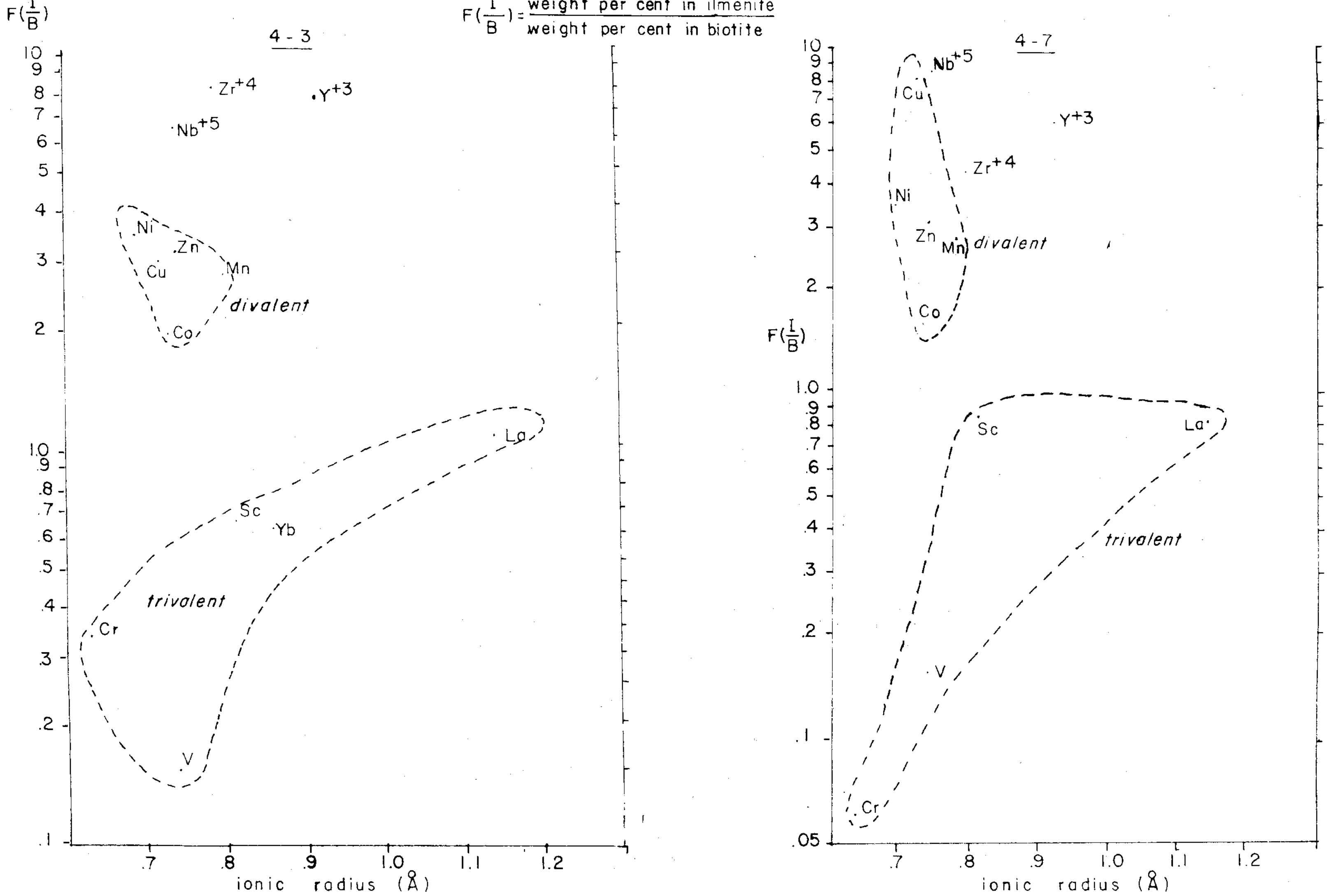
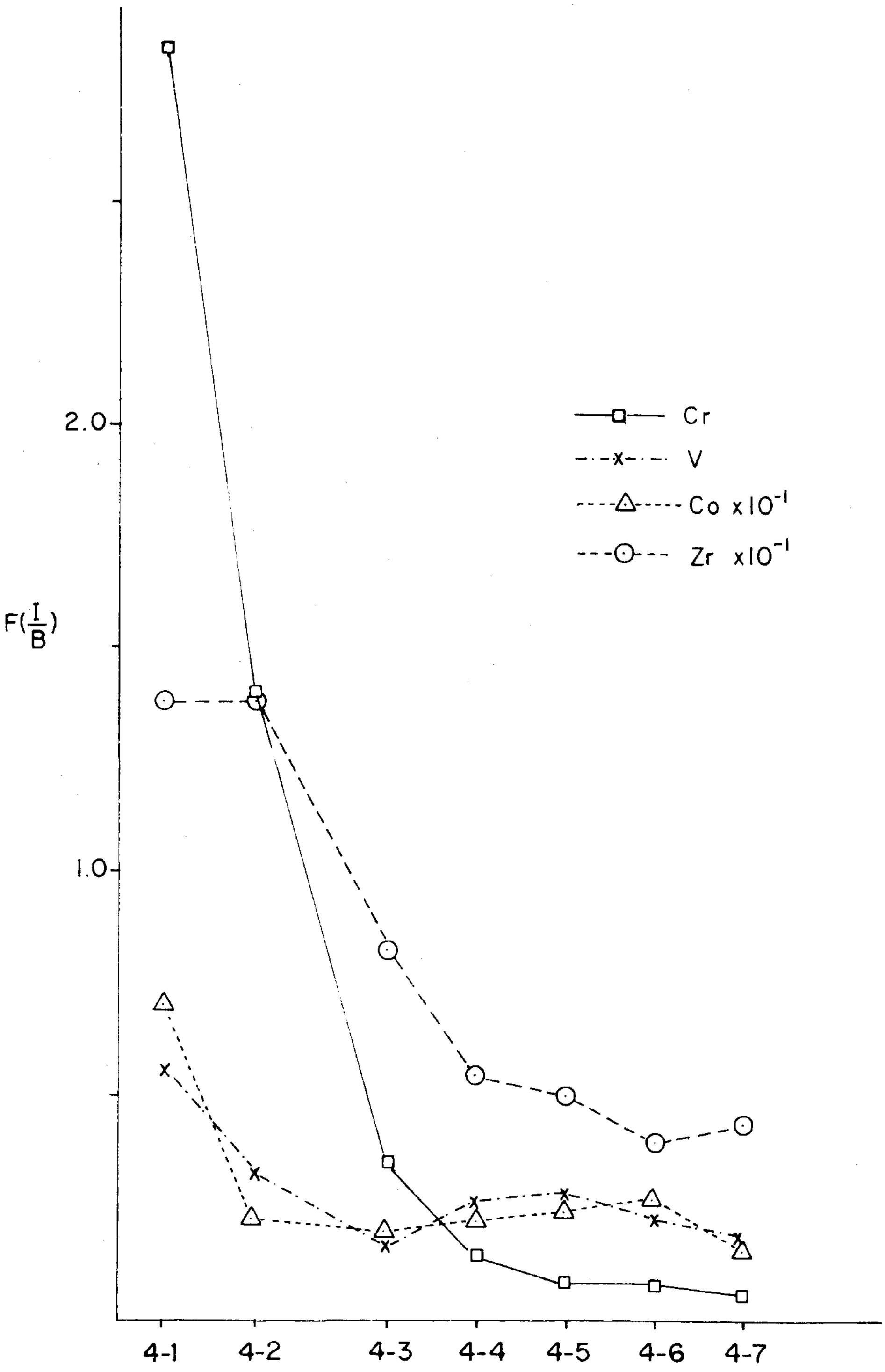


Fig. 36b Fractionation between ilmenite and biotite



and Zr are highly favored in the ilmenite structure, possibly substituting for Ti^{+4}.

The behavior of the divalent ions is controlled by their similarity in charge and size to Fe^{++} as it occurs in the ilmenite structure. These ions are in general about 3 times more abundant in the ilmenite than in the biotite while iron (FeO wt. per cent) is about 3 times higher, indicating that the degree of replacement is not strongly influenced for these ions by the crystallographic environment of the substitution site. In both cases the ions are octahedrally coordinated with oxygen. The trivalent ions on the other hand cannot readily substitute for Fe^{2+} in ilmenite. It is not clear why these ions do not substitute for Ti^{+4} in ilmenite, especially since they must take up essentially the same coordination site in biotite as they would in the Ti^{+4} position in ilmenite, coordinated with 6 oxygen ions. Lanthanum behaves in a fashion intermediate between that of the di- and trivalent ions, but this may be because its large ionic radius ($R(La^{+3}) = 1.12$ Å) forces it to enter preferentially into the larger 6-fold coordination sites of ilmenite. Arguments as to the distribution of ions based on electronegativity (Fyfe, 1951) may be applied here with caution. It is noted that the decreasing sequence of electronegativities for the divalent ions (Cu, Ni, Co, Zn, Mn) roughly parallels the sequence of decreasing fractionation factors for a single rock. For instance, rock 4-7 has the sequence $Cu > Ni > Zn > Co > Mn$. This would indicate that the biotite preferentially takes in ions capable of forming ionic bonds since the degree of covalency of an $O^{=}$ - metal ion bond is a decreasing function of the difference in electronegativity between $O^{=}$ and M^{+n}:

$$C = \exp(-c(x_{O^{=}} - x_{M^{+n}})^2)$$

(Pauling, 1948, p. 69) where x is the electronegativity on an arbitrary scale, $x_{O^{=}} = 3.5$; c is an empirical constant.

As a function of metamorphic rank the fractionation factors vary widely and irregularly. Fractionations for Cr, V, Zr and Co have been plotted in fig. 36b in order of their position in the sample sequence. For these elements the fractionation decreases markedly with metamorphic grade.

For no element in this series of ilmenites is there an obvious increase in fractionation with metamorphic rank.

D. FELDSPAR

1. Plagioclase

Only optical studies of plagioclase have been undertaken in this work, due to the difficulty in separating the plagioclase from the rock. The variation in composition along the sample sequence is slight and irregular. The only trend that is apparent is a slight decrease in the albite content with increasing grade from 85 per cent to 80 per cent. Using the analytical (n vs. Ab) curves of Tsuboi it is impossible to take in account the effect of change in the orthoclase-molecule content of the feldspar. In view of the relatively constant composition of the arkose and the lack of coexistent phases with large CaO content, it is unlikely that much change in the Ab/An ratio should occur.

2. Potash-feldspar

The potash-feldspar separated from the arkoses was examined by x-ray diffraction techniques. No splitting of the 131 reflection was observed in the patterns. In thin section the feldspars are homogeneous or contain less than 0.1 per cent by volume of fine perthite lamellae. They are not twinned.

The soda content (per cent Ab molecule) shows a slight increase from the lowest to the highest rank rock, with some irregularities along the sequence. In general there is a correlation between An content of the plagioclase and Ab content of the potash-feldspar from the same rock although the two do not increase uniformly along the sequence.

3. Evolution of the Feldspars with Progressive Metamorphism

On fig. 37 is shown a plot of the albite content of plagioclase from the meta-arkose samples plotted against the albite content of the coexistent potash-feldspar. We shall assume a closed system and a modal abundance of the feldspars taken from the revised modes of the meta-arkose (table 9). The approximate molar ratio of potash-feldspar (assumed to be pure $KAlSi_3O_8$) to plagioclase (assumed to be An_{10}) is 0.75. Therefore a change of 1 mol per cent Ab in the potash-feldspar should result in an 0.75 mol per cent change in the coexistent plagioclase. The slope of a plot of Ab in plagioclase against Ab in potash-feldspar should be about 0.75. The slope of the best-fitted line on fig. 37 is about 0.3 moles Ab in potash-feldspar/moles An in plagioclase, indicating that one cannot to a first approximation account for the change in the feldspar tie lines by the simple transfer of soda from one phase to the other. Again this may be evidence for loss of alkali (Na_2O in

TABLE 25. Compositions of Co-existing Plagioclase and Potash-Feldspar

Rock	Ab in Kf	An in Plagioclase
4-1	3.1 ± 0.3%	15 ± 2%
4-2	n.d.	10
4-3	5.2	23
4-4	4.2	20
4-5	5.7	21
4-6	4.5	20
4-7	4.9	20

FIG.37 Composition of coexistent feldspars

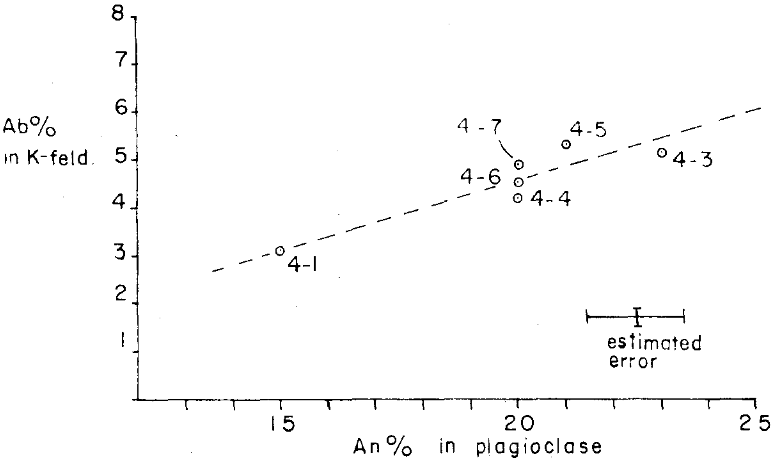
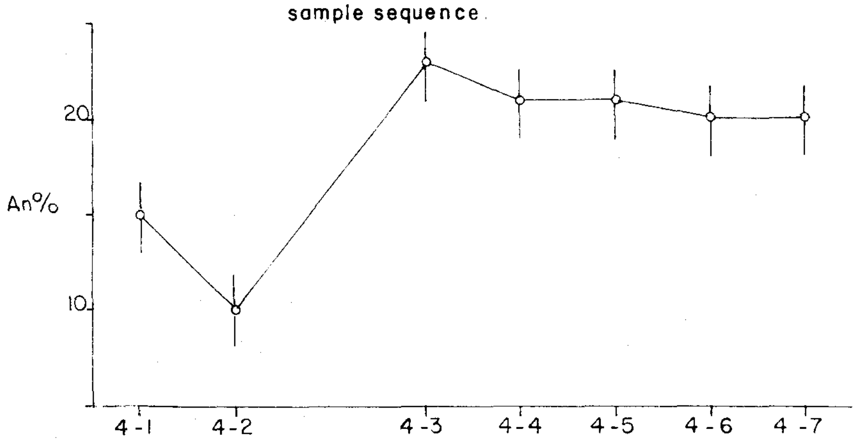


FIG.38 Variation in plagioclase along sample sequence.



this case) by the system, though there is certainly no systematic trend with increasing metamorphism to be seen. The change in volume per cent of plagioclase due to this Na_2O loss would be about 0. 05 x 17. 7 = 1 per cent (vol.). This would be impossible to distinguish from the random fluctuations in feldspar content from sample to sample observed in the modal analyses.

All samples of feldspar tested were shown to be monoclinic. This is remarkable since the inversion from monoclinic to triclinic symmetry has been shown to be complete at 500° C. (Laves, 1952). Geothermometry here indicates that this structural state may be stable down to 250° or less. In view of the uncertainties in temperature determination in this area it would be hazardous to relate this to the experimental evidence, but it may be an indication that the temperature of maximum "sanidization" may be overestimated by the experimental work, due to the sluggishness of the reaction. The work of Heier, on the other hand (1957) tends to confirm the experimental work as applied to natural systems.

E. MAGNETITE

The very small magnetite separates from the arkoses have not been studied in detail. An x-ray (Debye-Scherrer) photograph of the magnetite separate from 4-4 was made. Using the 440 and 553 reflections, a_o was determined to be 8. 427 ± 0. 002 Å. Assuming a linear relation between the a_o of the spinel and the titanium content (for confirmation of this see Vincent et. al., 1957) this magnetite would contain 32 mol per cent Fe_2TiO_4 (ulvospinel), equivalent to 11. 4 per cent TiO_2. This seems an unreasonably high value in comparison with values observed by Buddington et al. (1955) and Engel, et al. (1953) for magnetites obtained from igneous and metamorphic rocks. Engel reports values up to 3.6 per cent TiO_2 in magnetites from a paragneiss estimated to have been metamorphosed at about 600° C. Other ions, e.g., V^{+3}, substituting for Fe^{+3} may be present in sufficient quantity to enlarge the crystal lattice beyond the value that its TiO_2 content alone would require.

F. GARNET

On the basis of the a_o and index of refraction measurements it appears that the garnet occurring in rock 4-7 is a spessartitic almandine. It is shown plotted on fig. 21* in Chapter VI. Because it makes up less than 0.1 per cent by volume of the rock its appearance would not be expected to influence greatly the composition of the existing biotite. In fact, the MnO content of the coexisting biotite is higher than that of the adjacent specimen in the sequence, rather than lower as would be expected if spessartite were forming at the expense of the manganophyllite molecule of the biotite. The fact that the garnet is non-magnetic in response to the Franz Isodynamic separator suggests that it is low in iron.

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*Labelled "12-7" on that figure.

CHAPTER XIII. CONCLUSIONS

A. EVIDENCE BEARING ON MAJOR AND MINOR ELEMENT VARIATION WITH ADVANCING METAMORPHISM

In general the results of this work confirm the evidence of Shaw and earlier workers that progressive metamorphism may occur with essentially no change in chemical composition. The study undertaken is limited in that no great mineralogic changes indicative of progressive metamorphism can be observed to occur in the sequence of rocks sampled. The appropriate metamorphic rank-assignments must be made on the basis of mineral assemblages observed in schist adjacent to the meta-arkose on either side. More importantly, solid-state reactions resulting in the complete break-up of crystal lattices (comparable to the "reconstructive transformations" described by Buerger (1948)) are not found to have occurred in this unit. Such reactions, as for example the reaction of muscovite (plus quartz) to form sillimanite and potash-feldspar, are characteristic of progressive metamorphism and are indeed observed to have occurred in adjacent pelitic and basic igneous rocks. It seems likely that possible metasomatic changes would be most strongly felt by the rock when its minerals were opened up by such drastic reactions. In the present case only quite subtle changes in chemical and crystallographic makeup of the constituent minerals can be inferred to occur and so it is less likely that matter would have been mobilized during such circumstances. However it is noted that considerable recrystallization does take place, as evidenced by increase in grain size and by changes in the distribution of ions between coexisting phases.

It is commonly assumed that during progressive metamorphism the rock is in equilibrium with a pore fluid containing dissolved in it appreciable amounts of the major components of the rock. It is reasonable that, when subjected to a thermal gradient, there should be some tendency for material to diffuse along this gradient through this fluid. Since we are dealing with rocks whose bulk composition is close to that of a granite, the lowest melting fraction (that is, the set of components most enriched in the fluid phase at low temperatures) will be close in composition to the bulk composition of the total rock and therefore loss of these constituents will not be readily observable. Nevertheless there is some indication that in an irregular fashion the alkali content of the meta-arkose decreases and the CaO content increases as it is traced through progressively higher zones of metamorphism. Such effects are evidently very small in this case. They should be subject to further test, involving more complete sampling although it is doubtful, on the basis of the present study, that the effects will be found to be very much larger than the trends displayed here.

With respect to variations in the other major elements of the rock, we may conclude that even the relatively small compositional variations inherited from the parent sediment are sufficient to obscure the record of any possible change in composition with metamorphism. Except for a rise in the Fe^{++}/Fe^{+++} ratio there is no change in mean composition of the grouped samples from one end of the traverse to the other.

The changes in trace element concentrations which have been observed are as follows: increase in Mn, Ti, Co, Sr, Y and La and decrease in V (?). The last may be a spurious effect of spectrographic errors. There is no correlative decrease in the V content of either biotite or ilmenite. The increase in the other elements has been explained by assuming that for each ion a structurally favored site in some mineral is becoming available with progressive metamorphism. This does not require some unobservable diffusion of ions from distant sources of unexplained origin. It may simply represent a varying degree of fractionation with respect to these elements between the rock minerals and the surrounding pore fluid, with perhaps local diffusion extending to adjacent schist units. For Mn, Ti and Co the structurally favored site is in biotite. For Sr, although not independently observed, it is presumed that this site is as a substitute for Ca in plagioclase. Considering only the extreme members of the sample sequence (4-1 and 4-7) for which chemical analyses are available, it is seen that the Sr/Ca ratio varies only from 4.4 to 3.7, well within the error of spectrographic analysis. As for Y and La, although spectrographic errors tend to obscure any trends in composition in the biotite, it is probably into this mineral that these elements enter, in increasing degree with progressive metamorphism.

B. THE RELATION BETWEEN THE TEMPERATURE GRADIENT AND THE SAMPLE SEQUENCE

In the constituent minerals of the meta-arkose, especially biotite and ilmenite for which details are available, significant trends can be noted for only a few of the elements investigated. The tendency in this study has been to take at maximum import evidence of spectrographic data suggesting uniform increase or decrease in composition along the sequence of samples. It has been assumed that such trends will represent the temperature dependence of the various fractionation reactions going on. It has also been implicit in this work that: a) in most or all cases such fractionations should be monotonic increasing or decreasing functions of temperature and b) these functions will be independent of other features of the rock which may vary along the sequence in an irregular fashion. The justification for assumption a), though not shown here, can be generally derived assuming that the various trace elements form ideal solutions with the host minerals and that the heats of reaction forming such solutions are at worst monotonic functions of temperature and at best independent of temperature. The second assumption b) is not independently verifiable due to the complexity of other possible factors which may be active.

One may accept the first assumption and yet insist that the inferred sequence of temperatures is wrong, that is, that there are reversals in the temperature gradient. The superimposed effects of several nearby intrusive masses may have had a significant effect on geochemical equilibria while not affecting the mineralogic phase equilibria which are our key to progressive metamorphism. However, on inspecting the various graphical representations of trends in the fractionations and concentrations in the rock and its mineral constituents it does not appear that any other temperature sequence would yield a unique set of monotonically varying curves. It would appear that the most fruitful assumption is that the sequence of sample stations does follow, as the sequence of isograds which cut it would imply, a sequence of increasing temperatures during metamorphism. Even if this assumption were wrong in some details or if other variables, P_{H_2O}, P_{tot} or local stress conditions had had a differential effect along the sequence it would be impossible to reconstruct such deviations on the basis of seven samples.

C. MAJOR AND MINOR ELEMENT VARIATION IN THE CONSTITUENT BIOTITES AND ILMENITES

We have seen that major element variations in the biotites extracted from the meta-arkose samples are small compared with the analytical errors of the technique used and do not follow any distinct trend. Lacking other abundant ferro-magnesian phases with which to be equilibrated one would indeed expect such variations to be small. Changes in the oxidation state of iron are quite large but cannot be evaluated at all points along the sample sequence.

Minor element trends, although difficult to verify in view of the uncertainties of the analytical data, are suggested for several elements. Mn, Ti, La, Ba, Cr and V show an increase. Co may show an increase, almost completely obscured by scatter in analyses. The rock analyses tend to confirm the idea of an increase in this element. Sr and Ca decrease. It may be that emission-spectrographic data for Ca would be more useful than x-ray fluorescence in the case of such low concentrations.

Ilmenites coexistent with these biotites show some consistent trends in composition with metamorphism. The indication is that the ilmenite structure becomes less hospitable to foreign metal ions with increasing temperature. Except for Mn and Ni both of which increase with advancing grade of metamorphism, the trace element content of the ilmenites is constant or shows a decrease along the sample sequence. These changes are also reflected in falling fractionation factors between ilmenite and biotite and perhaps simply represent transfer of these trace elements (Zr, Cr, La, V, Y and Sn) to the biotite as the trace element tolerance of the biotite increases with rising temperature.

D. BEHAVIOR OF SIMILAR ELEMENTS DURING METAMORPHISM

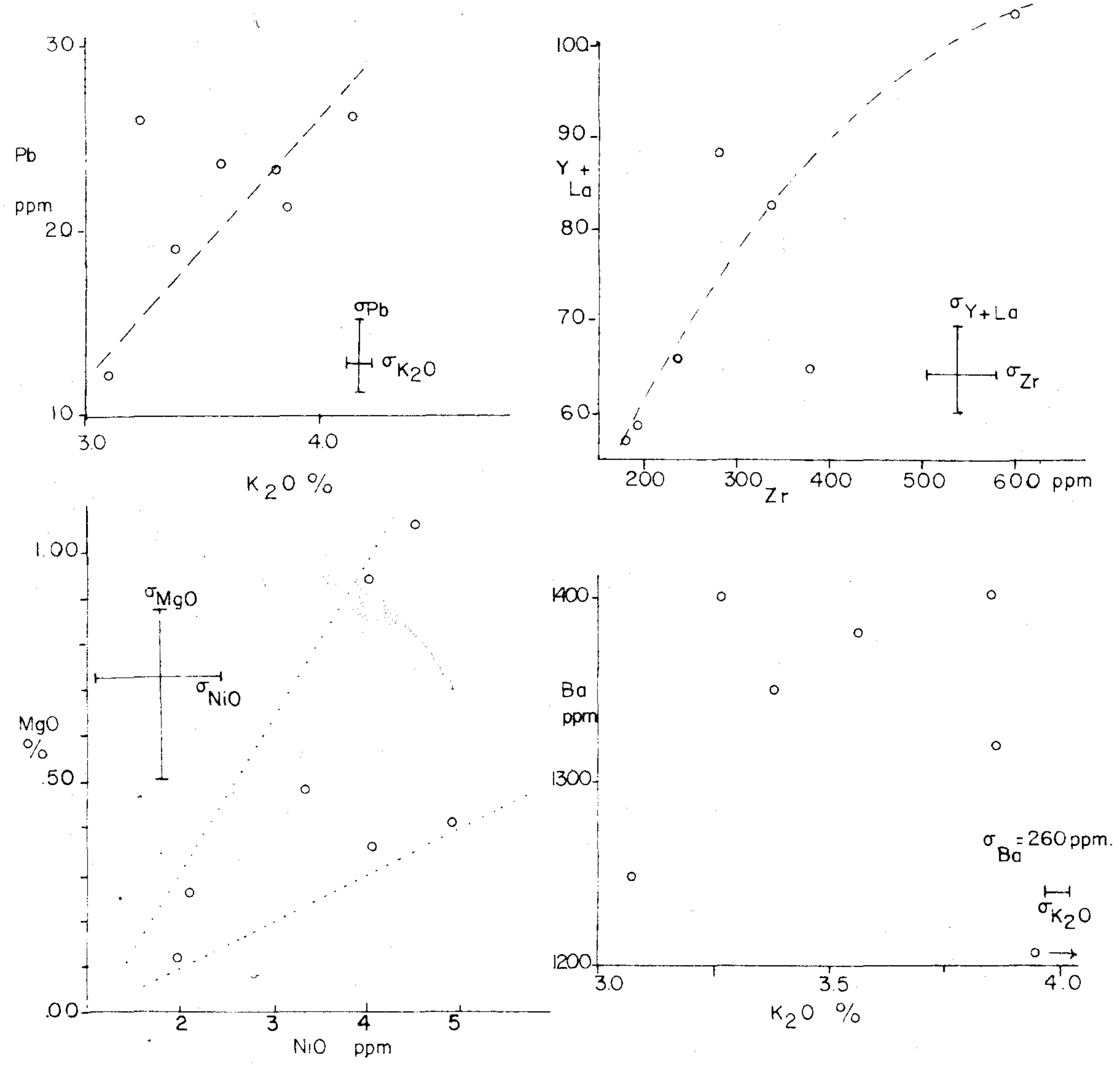
Though not previously discussed in this study, it may be seen that certain elements with similar chemical characteristics show similar variations with metamorphism, as measured in both the total rock and the separate phases. In part these similarities are inherited from the parent sediment since the same elements may have substituted for one another in the minerals of the unmetamorphosed arkose.

Graphs are shown in fig. 39 of the concentrations of various elements which, on the basis of similarity in charge and ionic radius are presumed to behave similarly with respect to capture and admission by minerals of the meta-arkose. Relations between major and minor element with similar properties are in general poorly defined. Although presumably Mg and Fe are substituted for by the di- and trivalent ions Cu, Ni, Co, Cr, etc., there is no evidence of correlation between any of these ions and either Mg or Fe. This is in marked contrast to the evidence from igneous rocks for which the rules of ionic size and charge can predict quite faithfully the degree of substitution of any of these elements for Mg or Fe. One possible exception to this rule is a weak correlation between Ca and Sr in biotite, shown in fig. 39. Of course Ca is acting almost as a trace element in this case.

Paired trace element analyses on the other hand show in some cases moderately well-defined correlation. This is best exemplified by the rare earths. Others are shown in fig. 39.

A more indirect measure of similarity of behavior is seen in the correlation of fractionation factors for similar elements. As an example there are plotted in fig. 39 fractionation factors F(I/B) for the distribution ratio of V between ilmenite and biotite versus the distribution ratio of Cr between the same two minerals. These two ions have identical charges (+3) and similar ionic radii ($R(V^{+3}) \doteq 0.65$; $R(Cr^{+3}) \doteq 0.63$). Less well defined correlations between the fractionation factors are seen among the elements Zr, La, Yb and Y, all occurring as highly charged ions in the rock. The divalent cations Co, Zn, and Ni also show poorly defined correlations between their F(I/B) values.

FIG.39 Relations between various trace and major elements. a: Total rock



b: Biotite

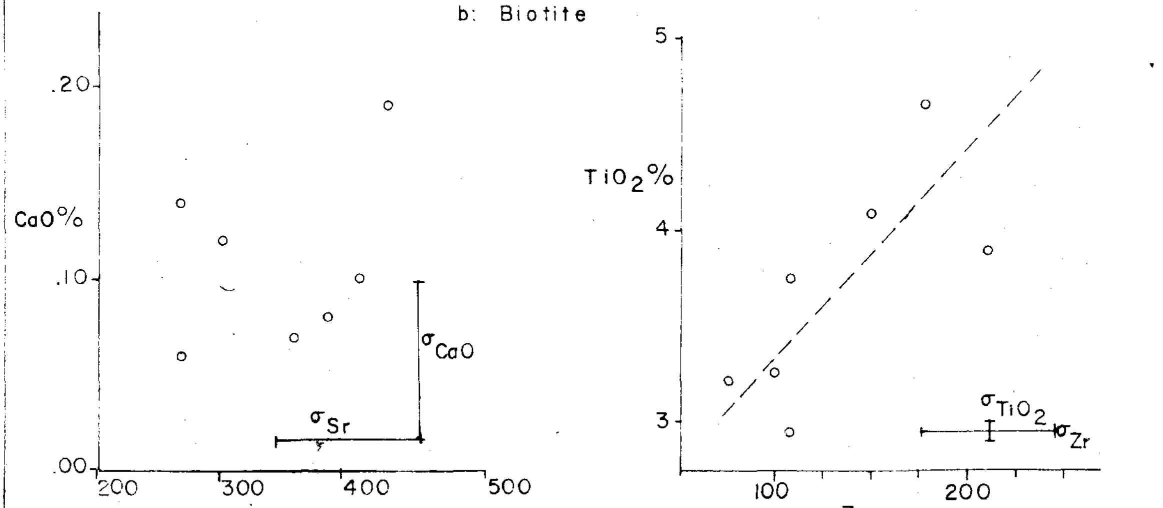
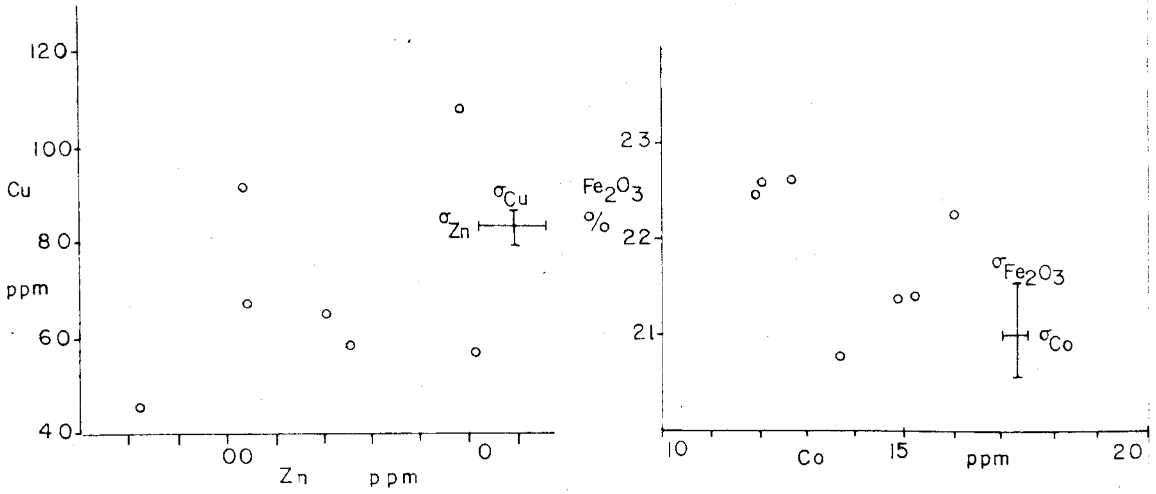
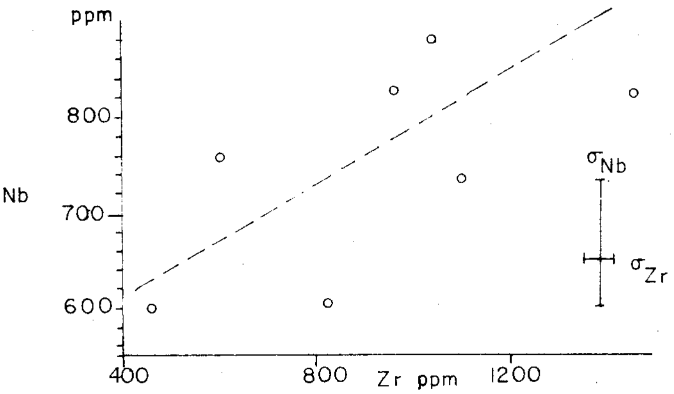


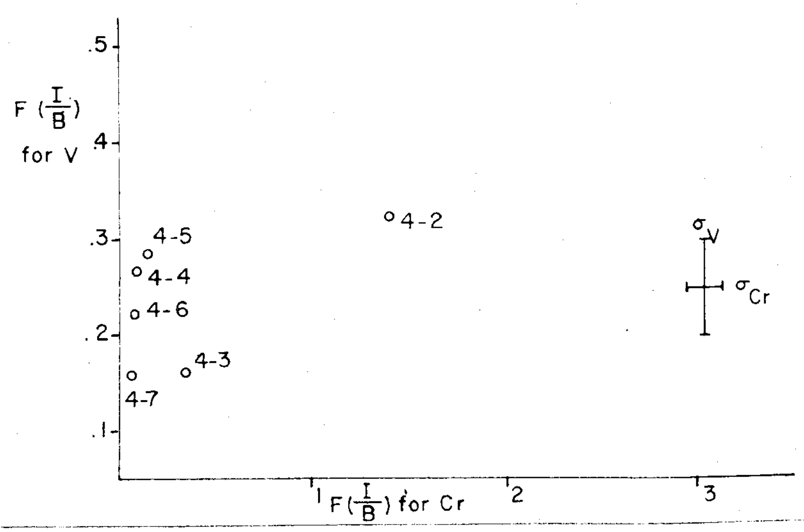
FIG.39 continued : b: Biotite, cont'd.



C. Ilmenite



d. Fractionation between ilmenite and biotite



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APPENDIX

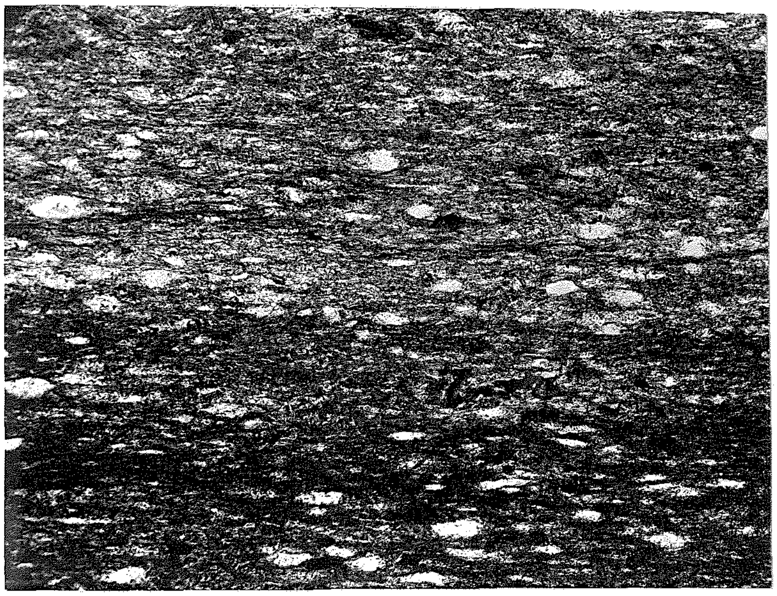
PHOTOGRAPHIC PLATES

Plate 2. Bedford Canyon Formation

Outcrop of Bedford Canyon formation in road cut along Winchester Road, showing interlayered quartzite and phyllite layers.

Plate 3. Bedford Canyon Formation

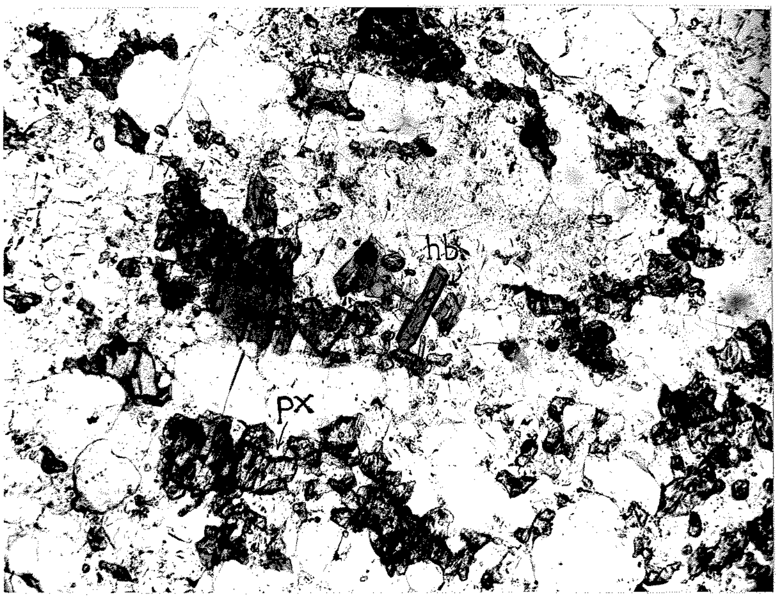
a) Phyllite derived from sandy shale (wacke). Plane light, X 35.



b) Meta-chert, veined with quartz, densely crowded with graphite. Plane light, X 40.

Plate 4. French Valley Formation, Member A

a) Calcareous quartzite, subunit fvA2. hb = hornblende, px = diopside; matrix composed of quartz and oligoclase. Plane light, X 100.



b) Conglomerate from fvA4: coarse clasts of shale embedded in arkose matrix. Porphyroblasts in shale (schist) are andalusite. Plane light, X 3.

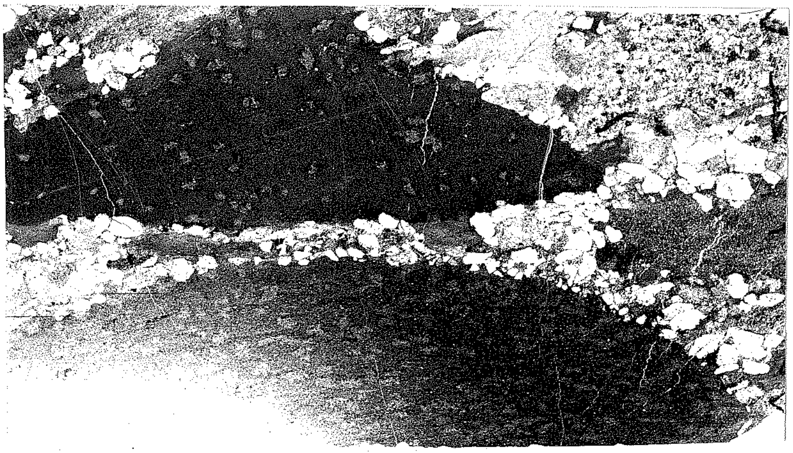


Plate 4. (continued) French Valley Formation, Member A

c) Shale clasts (now schist clots) in conglomeratic arkose of fvA4; boulders in stream channel.

d) "Wacke" from fvA5. Crossed nicols, X 28. The matrix is composed of sericite and biotite.

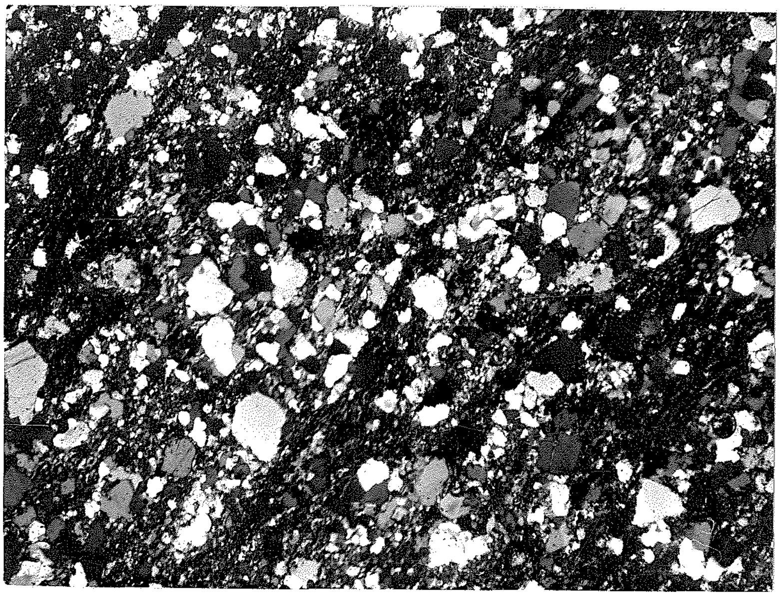
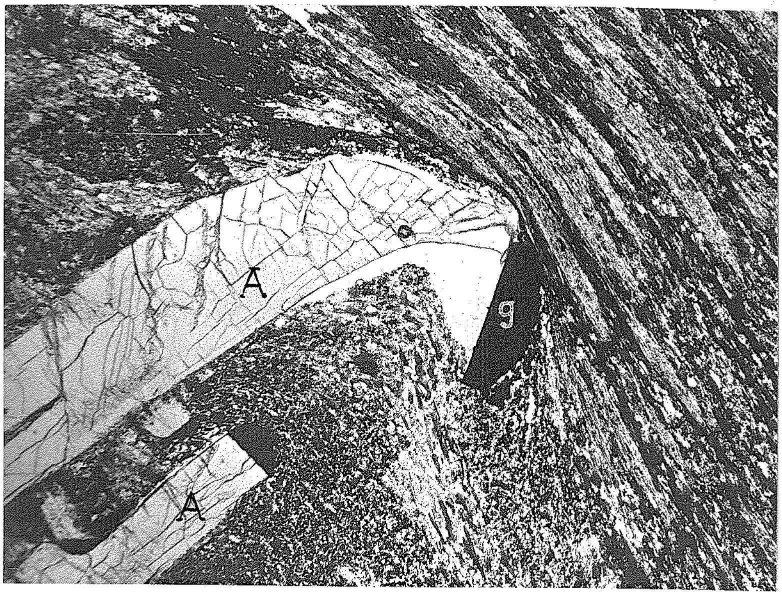


Plate 4. (continued) French Valley Formation, Member A

e) Cross bedding in subunit fvA5. Bedding is right-side up.

Plate 5. French Valley Formation, Members B and C

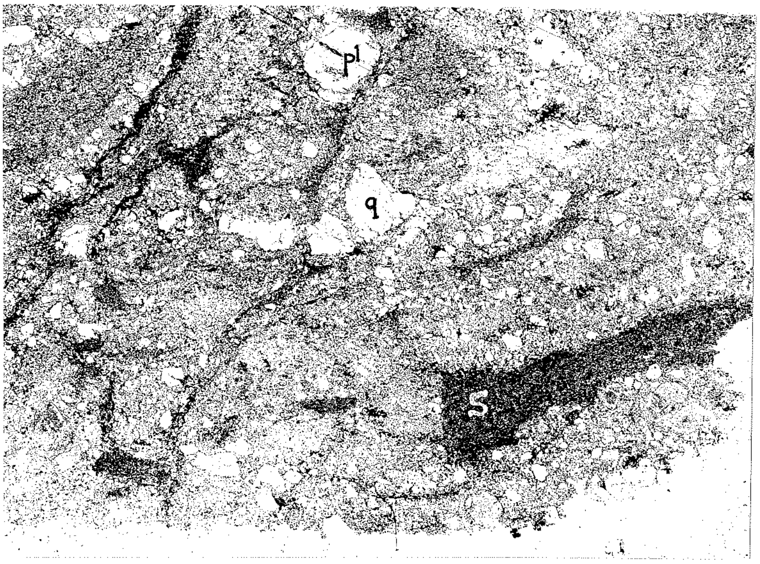
a) Andalusite-biotite schist in fvB2. A = andalusite, q = quartz, g = graphite. Plane light, X 35.



b) Knotted schist in fvC7h, south side of Searls Ridge, showing weathering pits formerly occupied by andalusite porphyroblasts.

Plate 5. (continued) French Valley Formation, Members B and C

c) Conglomerate from fvC4, east flank of Domenigoni Valley syncline. Coarser clasts are vein quartz (q), shale (s) and oligoclase (pl). Matrix is arkose. Plane light, X3.



d) Banded gneiss, fvC7d, showing pinching and swelling of layers. Outcrop on Searls Ridge.

Plate 6. Banded Gneiss

Thin section of banded gneiss from fvC7d, south of Domenigoni Valley. Note concentration of biotite into discrete layers and development of sillimanite (s), possibly pseudomorphous after andalusite. Plane light, X3.

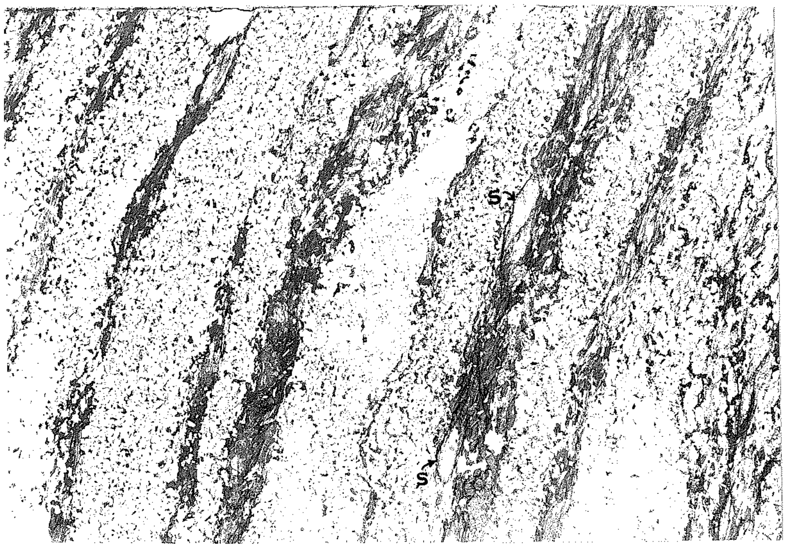


Plate 7. Amphibolite

Banded amphibolite along Rawson Canyon north of Crown Valley. h = hornblende-plagioclase rock; d = diopside-plagioclase rock; pl = plagioclase (An_{40}) vein material.

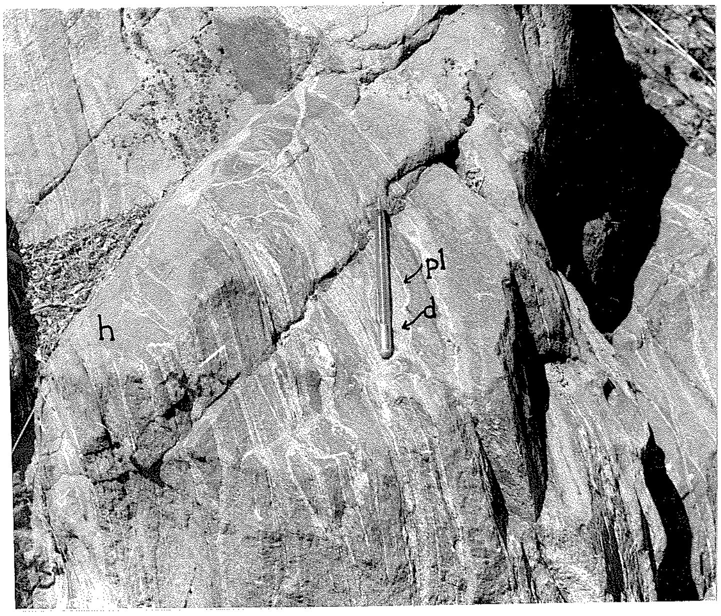
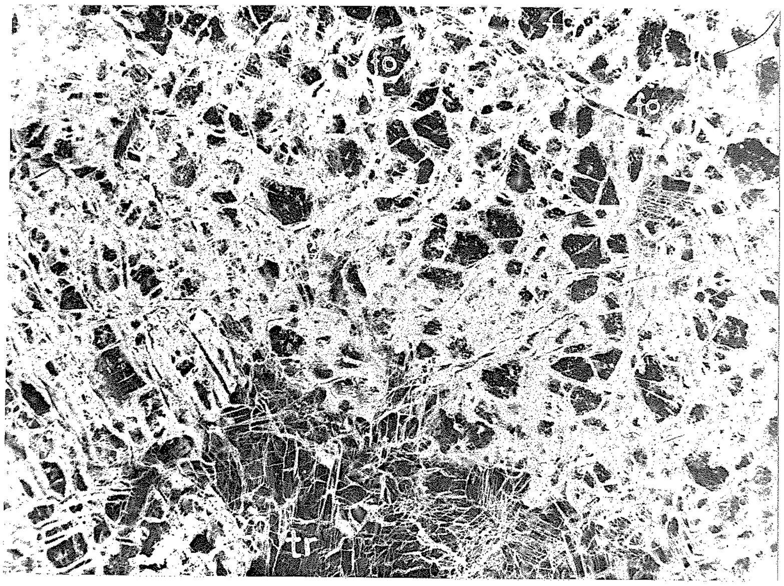


Plate 8. Ultrabasic and Associated Rocks of the Hemet Magnesite Deposit

a) Meta-peridotite: tr = tremolite-actinolite; fo = forsteritic olivine. Fractures are filled with chlorite and serpentine. Plane light, X 35.



b) Porphyroblastic cordierite-biotite gneiss. bi = biotite, co = cordierite, q = quartz, kf = potash-feldspar. Crossed nicols, X 35.

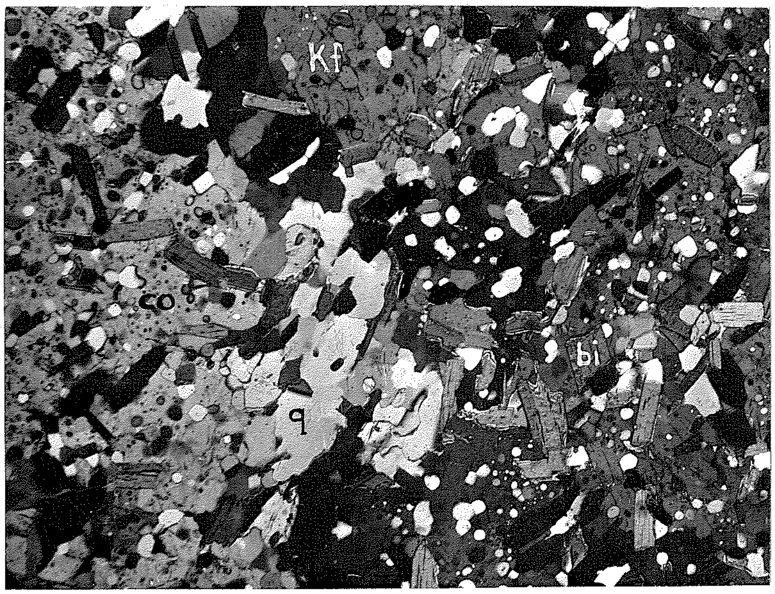
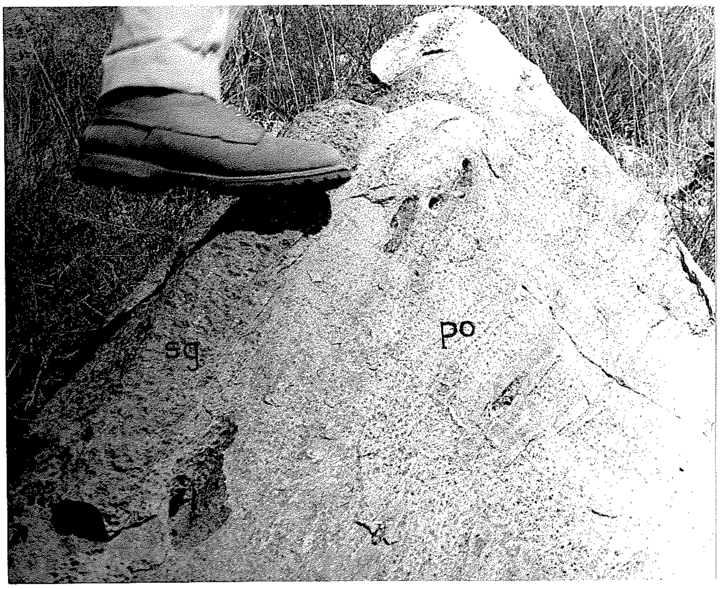


Plate 8. (continued) Ultrabasic and Associated Rocks of the Hemet Magnesite Deposit

c) Contact zone in gneiss: po = porphyroblastic cordierite-biotite gneiss, sg = de-silicified sillimanite-biotite gneiss. Metaperidotite lies to left of sg.



d) Calc-silicate marble: di = diopside, cal = calcite, mte = magnetite, fo = forsterite. Plane light, X 35.

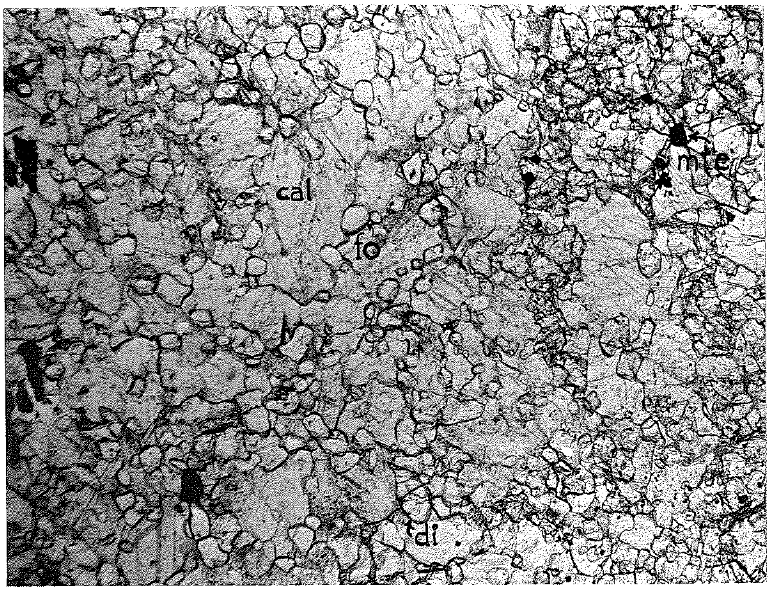


Plate 9. Woodson Mountain Granodiorite

a) Horizontal face of outcrop south of Rawson Ranch. Note lack of foliation.

b) Vertical face of same outcrop as above showing pronounced linear fabric.

Plate 9. (continued) Woodson Mountain Granodiorite

c) Contact between French Valley Formation (banded gneiss) and Woodson Mountain granodiorite. (B29(1,1)).

d) "Ring dike" of San Marcos gabbro partially surrounding Woodson Mountain granodiorite intrusive south of Rawson Ranch. View looking northwest toward Winchester from hill in B20(1,1). Ksm = San Marcos gabbro, Kwm = Woodson Mountain granodiorite. Unmarked areas are underlain by schist and gneiss of French Valley formation.

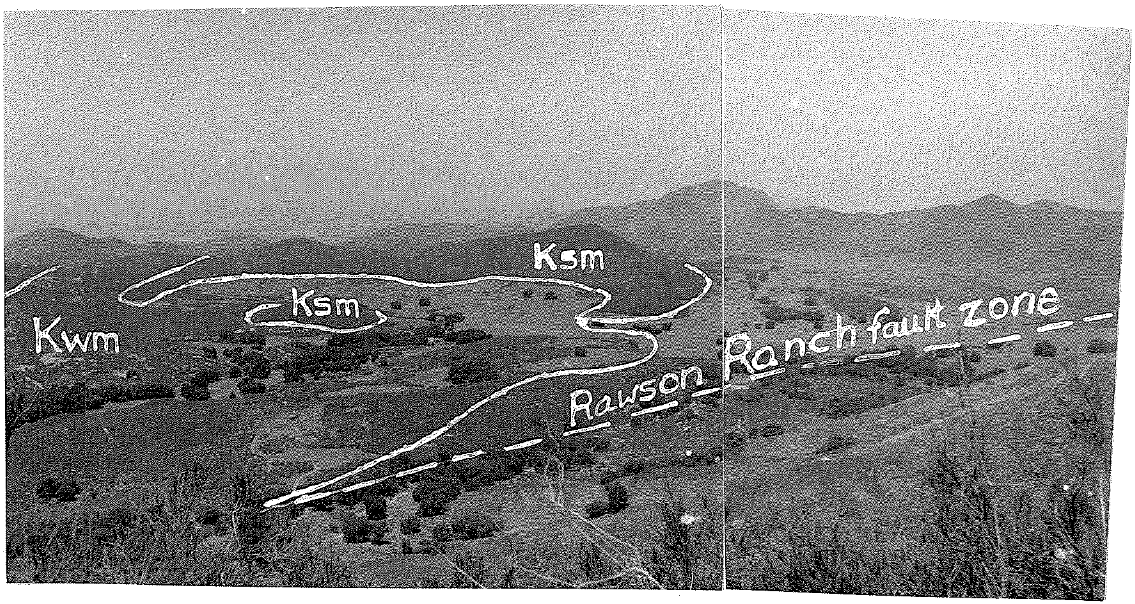
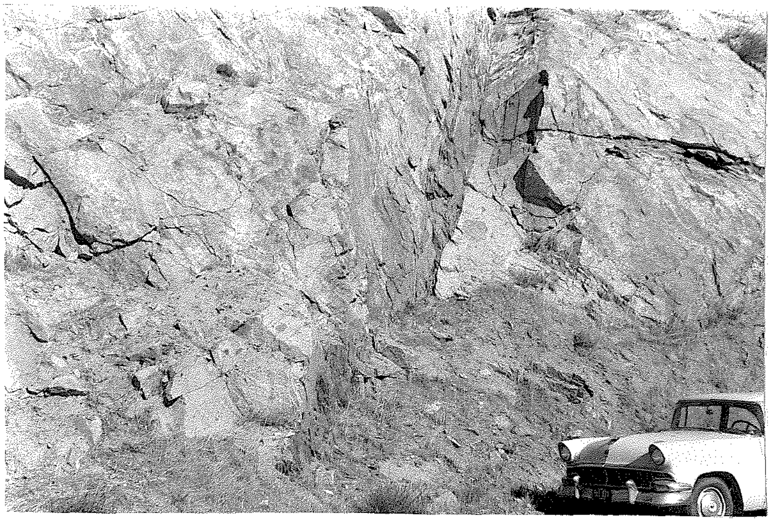


Plate 10. Igneous Rocks

a) Quartz latite dike cutting Domenigoni Valley granodiorite along Winchester Road (A4(4, 3)).



b) North side of Double Butte, showing contact between Domenigoni Valley granodiorite and French Valley formation.

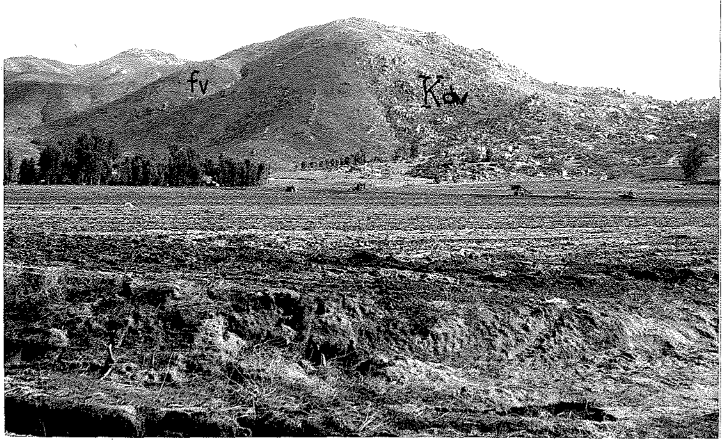


Plate 11. Migmatites

a) Migmatite formed near contact between French Valley formation and Woodson Mountain granodiorite. More typical migmatite facies of French Valley formation too poorly exposed to be photographed.

b) Migmatite breccia, X 0.5.

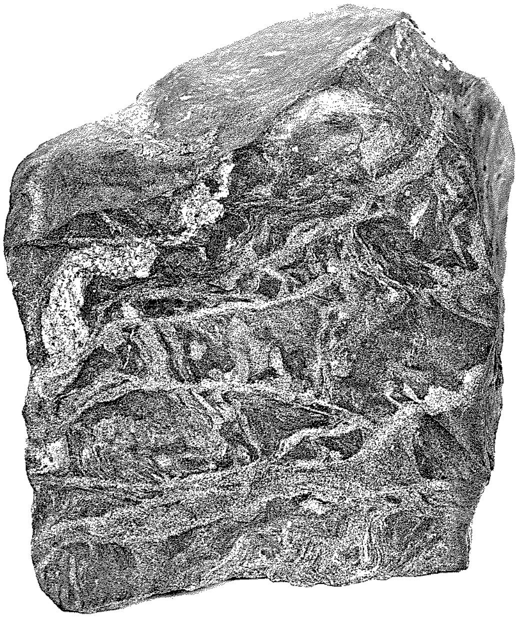


Plate 12. Structural Geology

a) Muscovite graphite schist from French Valley formation (fvC7h) showing shear planes intersecting foliation at high angle. Plane light, X 100.

b) Feldspathic quartzite ($fvBq_2$) north of "ring-dike" of San Marcos gabbro (A19(2, 3)). Note lineation marked by grooves in outcrop. This parallels direction of elongation of quartz grains in rock.

Plate 13. Metamorphism

Biotite being replaced by sillimanite in a biotite-sillimanite-quartz-2-feldspar gneiss, French Valley formation, Member C. Plane light, X 35.

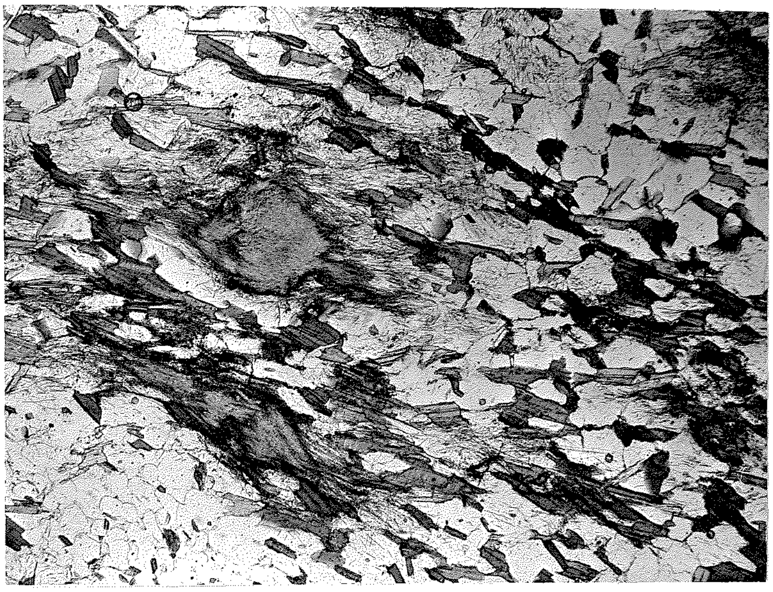
Plate 14. Textural Changes During Progressive Metamorphism

a) Phyllite from lowermost French Valley formation (subunit fvAl). Plane light, X 100.

b) Andalusite schist from French Valley formation. a = andalusite, q = quartz. Mica is intergrown biotite and muscovite. Plane light, X 25.

Plate 14. (continued) Textural Changes During Progressive Metamorphism

c) Biotite-sillimanite gneiss, French Valley formation. Plane light, X 30.



d) Migmatitic gneiss, French Valley formation. Plane light, X 30.

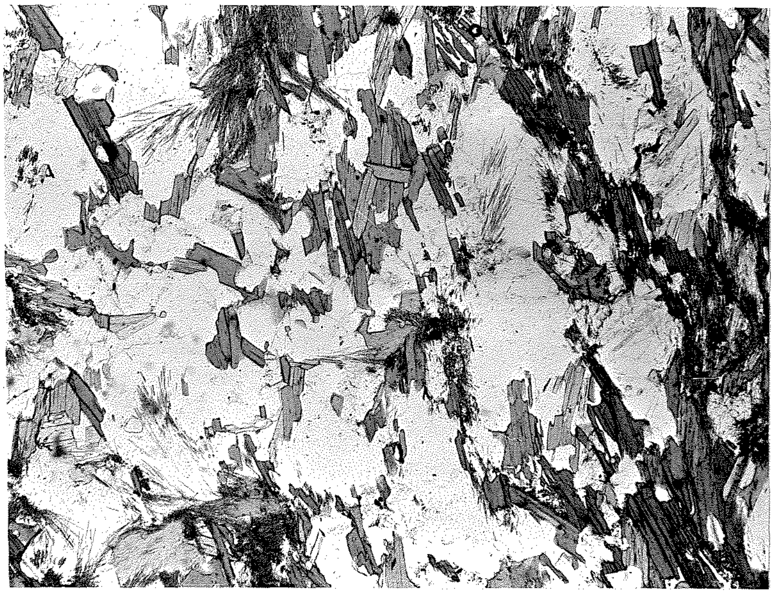
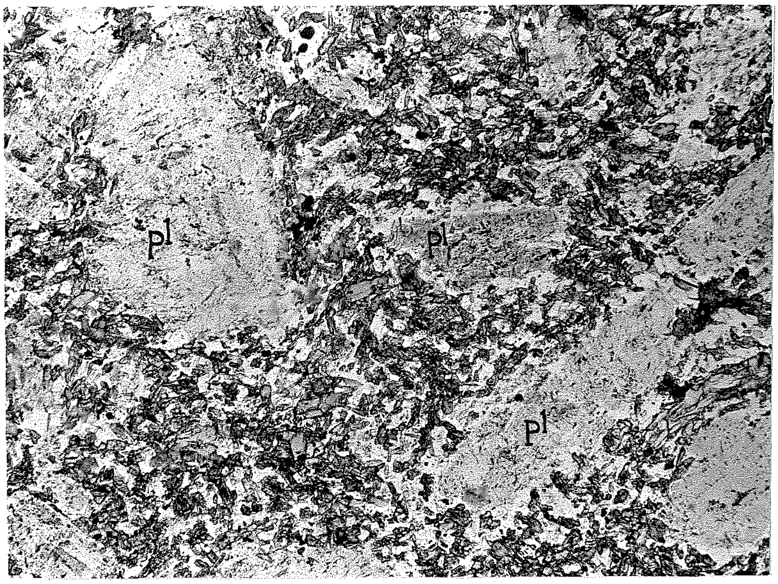


Plate 15. Metamorphism of the Amphibolites

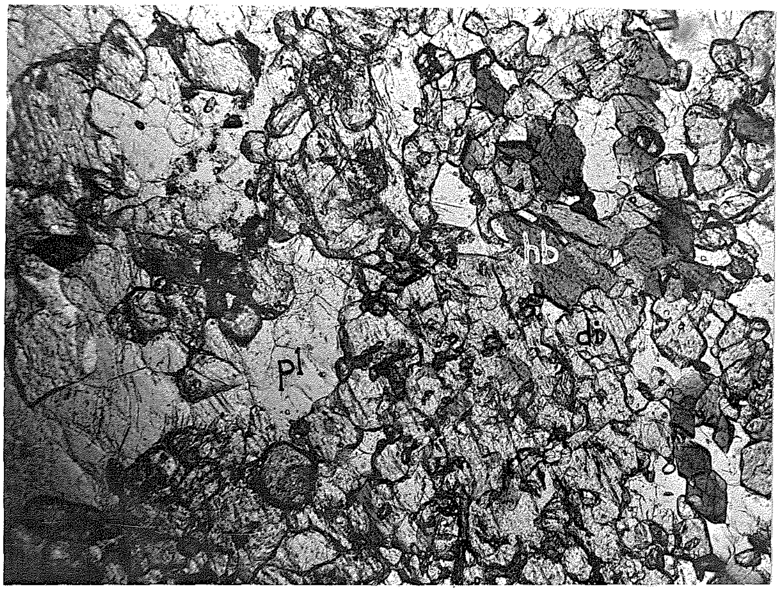
a) Meta-basalt flow, French Valley formation, Member B (Bam). Andalusite zone. Plane light, X 40. (Note relict zoned plagioclase phenocrysts, marked "pl").



b) Same, crossed nicols.

Plate 15. (continued) Metamorphism of the Amphibolites

c) Diopsidic amphibolite, garnet zone. di = diopside, pl = plagioclase, hb = hornblende. Plane light, X 100.



d) Pyroxene-garnet granulite from C 28(1, 4). px = greenish clinopyroxene, gt = garnet, pl = plagioclase. Plane light, X 35.

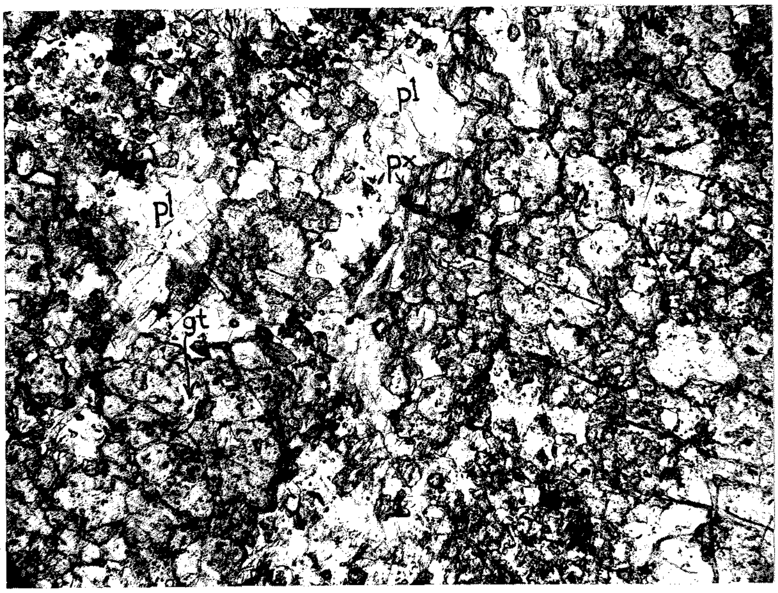
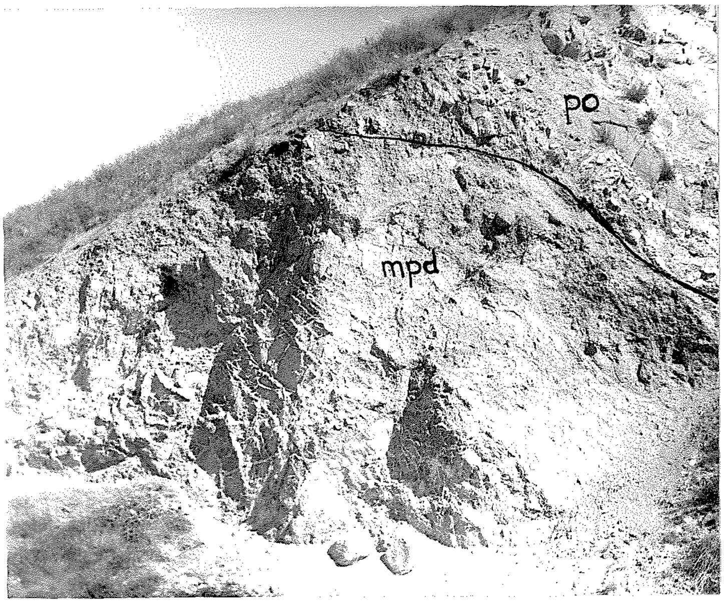


Plate 16. The Hemet Magnesite Deposit

a) Quarry wall, showing uniformly distributed, unoriented veinlets of magnesite cutting altered metaperidotite (mpd). Country rock consists of cordierite gneiss (po).



b) Close-up of typical ore, showing irregular, unoriented veinlets of siliceous magnesite in earthy metaperidotite.

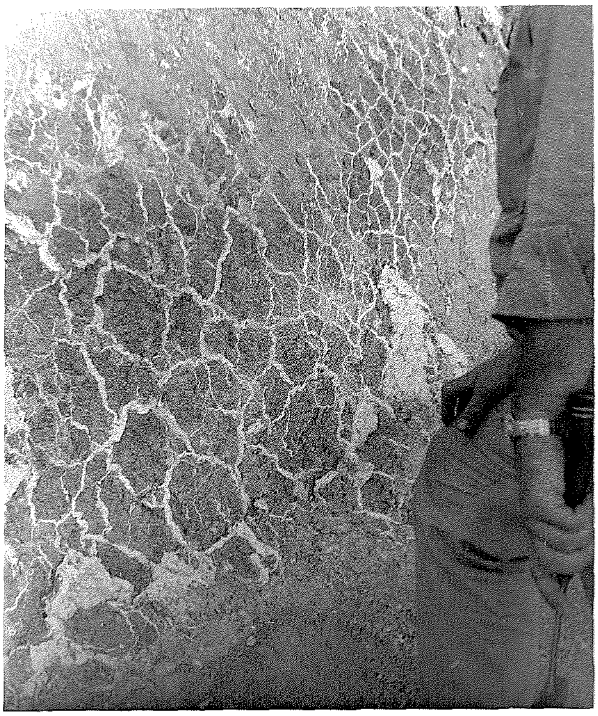
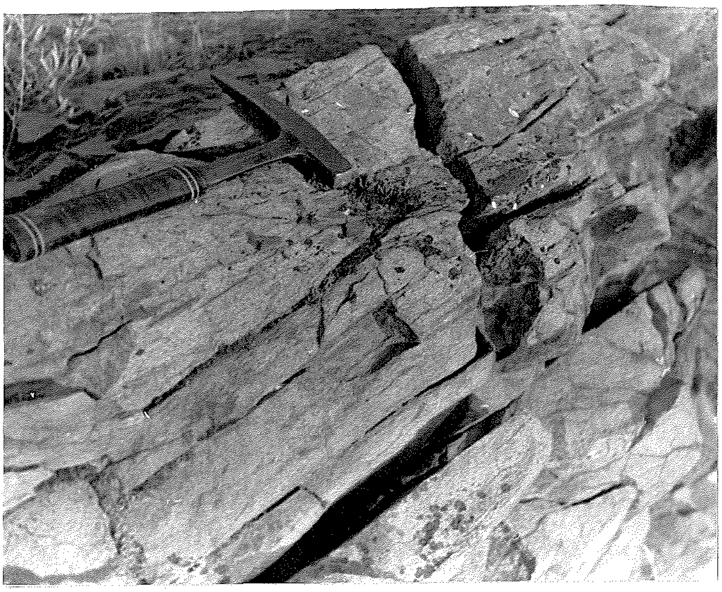


Plate 17. Meta-Arkose of the French Valley Formation, Subunit fvB4

a) Sample locality 4-2. Dark spots are lichens.



b) Sample locality 4-7. Note granite-like outcrop character. Bedding dips almost vertically and strikes away from observer.

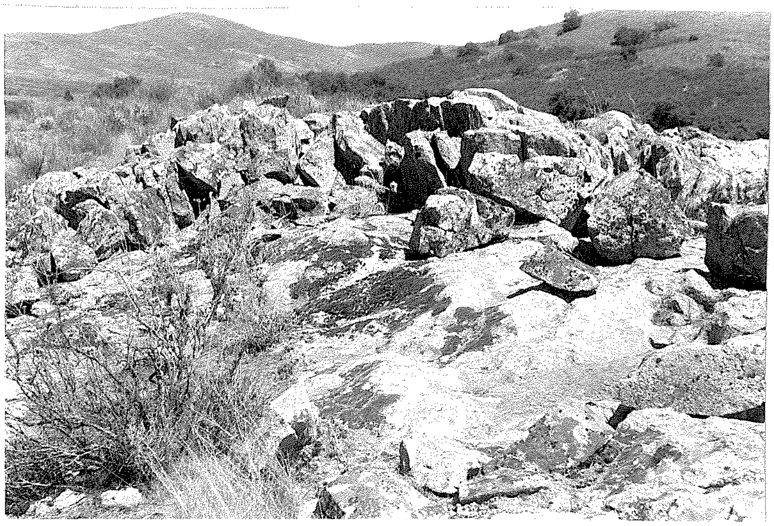
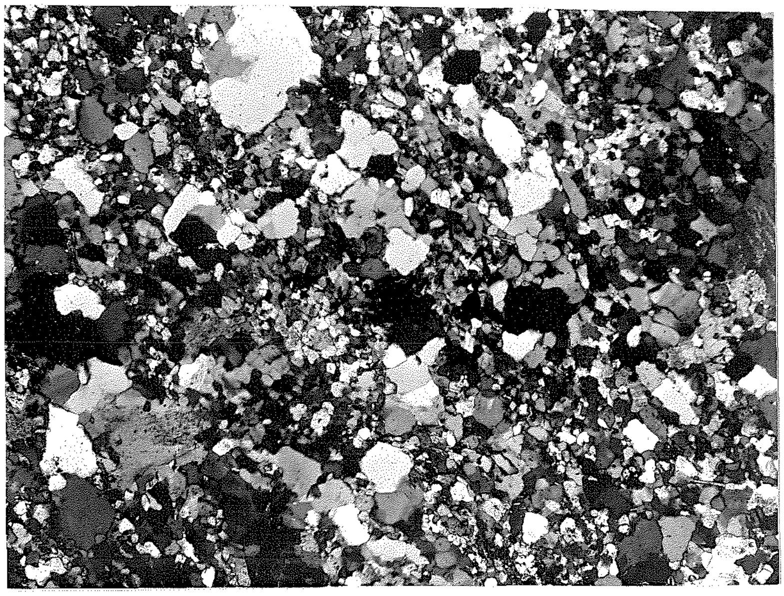


Plate 18. Metamorphism of the Meta-Arkose

a) Specimen 4-1, showing poorly sorted character of primary sedimentary texture. Crossed nicols, X 30.



b) Sequence of photographs showing progressive coarsening and homogenization of texture. b = biotite, pl = plagioclase (altered), gt = garnet, il = ilmenite. Plane light and X 30 in all photographs.

1) Specimen 4-1

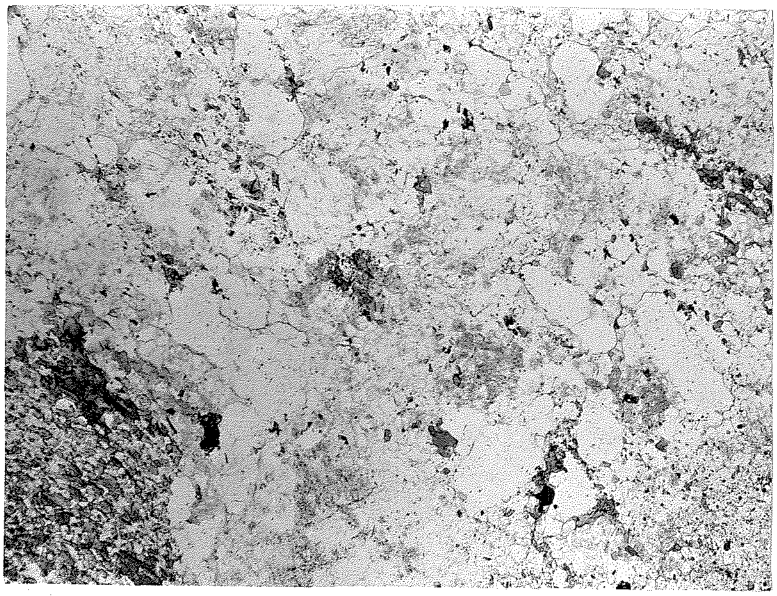
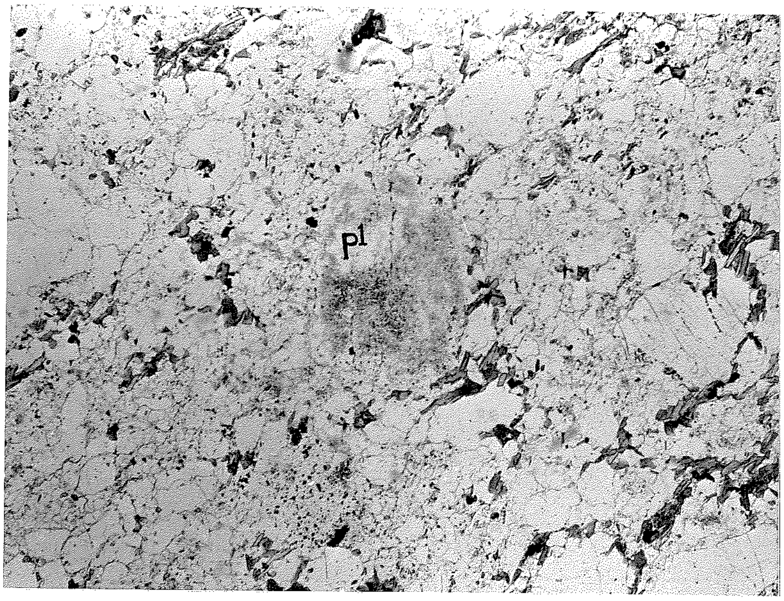


Plate 18. (continued) Metamorphism of the Meta-Arkose

b) continued

2) Specimen 4-4



3) Specimen 4-7