AN INVERTEBRATE ASSEMBLAGE FROM THE "MODELO" FORMATION OF REYNIER CANYON, LOS ANGELES COUNTY, CALIFORNIA

Thesis by

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ACKNOWLEDGMENTS

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ABSTRACT

An upper Miocene marine molluscan fauna occurs in basal "Modelo" sandstones that flank Reynier Canyon of the southeastern portion of the Ventura basin, Los Angeles County, California. Here two outliers of these beds are nonconformable on continental siltstones and tuffs of the Mint Canyon formation. The "Modelo" sandstones grade upward into punky diatomaceous shales also of the "Modelo" sequence, that contain *Anadara* cf. *obispoana* and a foraminiferal fauna reported to be Mohnian in age. The shales of one outlier are unconformably overlain by other marine sediments.

The upper Miocene age of the basal "Modelo" beds is indicated by the presence of *Clementia* cf. *martini*, *Dosinia arnoldi*, *Lyropecten estrellanus* ss., *Spisula albaria*, *Tivela diabloensis*, and a large *Ostrea*. Approximately half of the forms in the Reynier Canyon "Modelo" fauna also occur in the Elsmere Canyon fauna; these include *Laevicardium centifilosum*, *L. quadragenarium* var. *fernandoense*, *Lucina nuttallii*, *Cancellaria elsmerensis*, *C. hemphilli*, *C. tritonidea*, *Surculites remondii*, *Murithais eldridgei*, *Nuculana taphria*, and *Turritella cooperi*. The Elsmere Canyon species *Patinopecten lohri*, *Astrodapsis fernandoensis*, and *Dendraster* sp. are missing from the "Modelo" fauna.

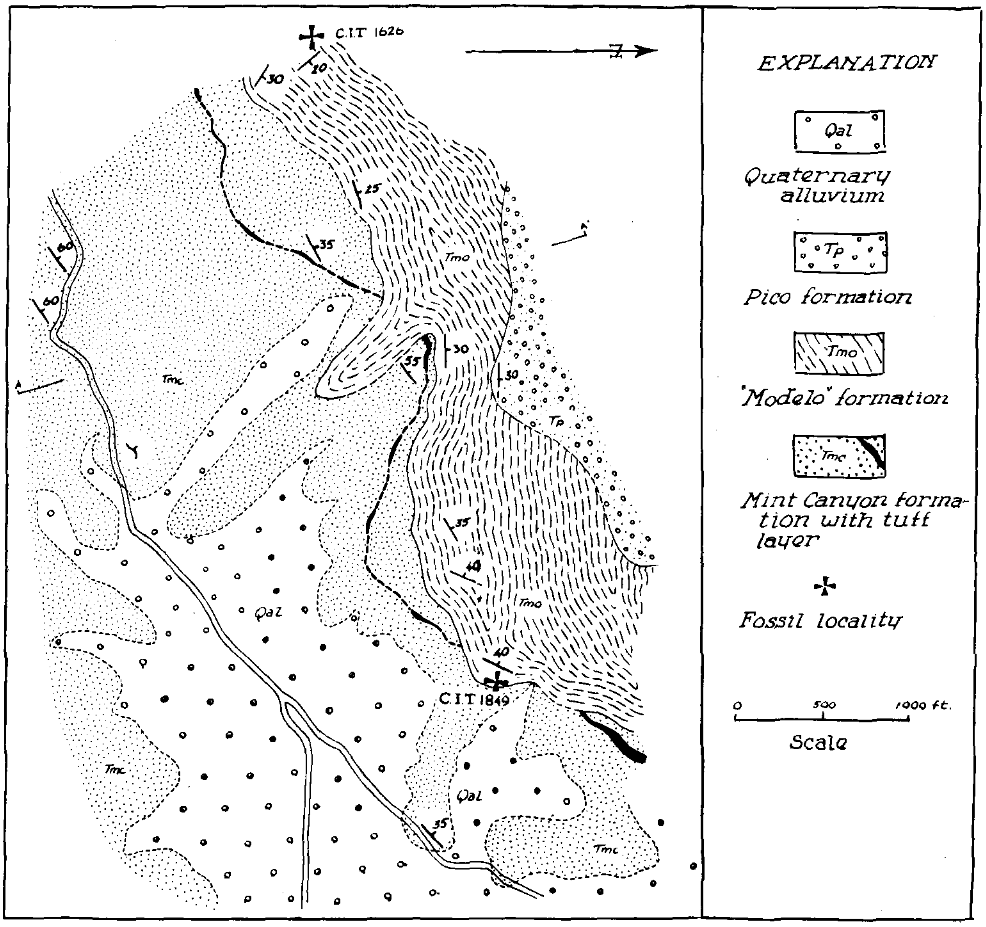

Figure 1. Geologic map of Reynier Canyon area.

AN INVERTEBRATE ASSEMBLAGE FROM THE "MODELO" FORMATION OF REYNIER CANYON, LOS ANGELES COUNTY, CALIFORNIA

INTRODUCTION

The position of the Miocene-Pliocene boundary in the sediments of the eastern Ventura Basin has long been a subject of interest to students of California Cenozoic stratigraphy. In this region, during upper Miocene-lower Pliocene time, both non-marine and marine beds were deposited. The non-marine beds, several thousand feet thick, comprise the Mint Canyon formation, and consist predominantly of sandstone, siltstone and conglomerate. The marine beds, several hundred feet thick, are mostly sandstones, siltstones and diatomaceous shales. These have been correlated with part of the "Modelo" formation.

The contact between the two formations is exposed along a northwest-trending belt about 18 miles long, and extending from the Elizabeth Lake Canyon area on the northwest to the Sand Canyon area on the southeast. For a distance of about 12 miles between Elizabeth Lake Canyon and Bouquet Canyon, the contact's exposures are almost continuous, but southeast of Bouquet Canyon it is overlapped by post-"Modelo" formations. In the Sand Canyon area "Modelo" beds are contained in two erosional outliers resting upon Mint Canyon strata.

Kew[1] in his pioneer reconnaissance study of a part of the eastern

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1/ Kew, W. S. W., Geology and oil resources of a part of Los Angeles and Ventura Counties, California: U. S. Geol. Survey Bull. 753, p. 68, 1924.

Ventura Basin, described the "Modelo" beds exposed in the Dry Canyon, Bouquet Canyon and Sand Canyon areas as unconformable upon sediments of the Mint Canyon formation. In exposures of the contact north of Dry Canyon, however, Clements[2] has reported an apparent lateral and

2/ Clements, T., Structure of the southeastern part of Tejon Quadrangle, California: Am. Assoc. Petroleum Geologists Bull., vol. 21, no. 2, p. 215, Feb. 1937.

upward gradation from Mint Canyon into "Modelo" beds. Jahns[3], in a

3/ Jahns, R. H., Stratigraphy of the easternmost Ventura basin, California: Carn. Inst. Wash. Publ. 514, pp. 145-194, 1940.

later, detailed study of the vicinity of Bouquet Canyon area, concurred with Kew's description of the unconformity in that area.

Reynier Canyon, a small tributary draining northeastward into Sand Canyon, separates the two "Modelo" outliers. The "Modelo" sediments of the larger outlier, which is immediately northwest of Reynier Canyon, are in turn overlain by marine beds of the Pico formation and by non-marine beds of the Saugus formation. The other contains only "Modelo" sediments. The "Modelo" occurrences in both outliers were apparently at one time joined to the more extensive exposures of the formation in the Bouquet Canyon area.

The data contained in the present report were gathered along the southern border of the northwestern outlier. Here the unconformity separating the Mint Canyon and "Modelo" formations, as previously reported by Kew[4], is clearly shown. In this area also, the basal

4/ Kew, W. S. W., op. cit.

"Modelo" beds contain a numerous and hitherto unreported invertebrate fauna.

Invertebrate collections from the "Modelo" beds exposed to the northwest of Bouquet Canyon[5] have been assigned to the uppermost Miocene

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5/ Kew, W. S. W., op. cit.

Woodring, W. P., Age of the "Modelo" formation of the Santa Monica Mountains, California (abstract): Geol. Soc. America Bull., vol. 41, p. 155, 1930.

Maxson, J. H., Miocene-Pliocene boundary: a paper presented before the 15th annual meeting of the Pacific Section, Amer. Assoc. Petrol. Geol., Los Angeles, California, November 3, 1938.

White, R. C., and Buffington, E. C., Age of the Modelo (?) beds in Haskell and Dry Canyons, northern Los Angeles County, California (abstract): Geol. Soc. America Bull., vol. 59, p. 1389, 1948.

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as recognized by invertebrate paleontologists. Stock[6] and Maxson[7],

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6/ Stock, Chester, as cited by Kew, W. S. W., op. cit.

7/ Maxson, J. H., A Tertiary mammalian fauna from the Mint Canyon formation of southern California: Carnegie Inst., Washington, Pub. 404, pp. 77-112, 1930.

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who have studied the vertebrate fauna of the Mint Canyon beds, concluded that these also were deposited in upper Miocene time. The Mint Canyon fauna, however, includes the genus Hipparion, a form that most vertebrate paleontologists consider a guide to the Pliocene. Stirton[8],

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8/ Stirton, R. A., Critical review of the Mint Canyon mammalian fauna and its correlative significance: Amer. Jour. Sci., 5th ser., vol. 26, No. 156, pp. 569-573, December 1933.

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in a critical review of the problem held to this conventional concept and advocated a lower Pliocene age for the Mint Canyon formation.

The age problem, therefore, involves a seeming anomaly in which marine beds with an uppermost Miocene invertebrate fauna unconformably overlie non-marine beds containing a vertebrate fauna recognized by most vertebrate paleontologists as lower Pliocene.

A conciliation of these conflicting interpretations will not be attempted here. It is hoped, however, that additional data will assist in such a conciliation and will lend precision to the correlation of the "Modelo" formation of eastern Ventura basin.

TOPOGRAPHIC AND GEOLOGIC FEATURES OF THE REYNIER CANYON AREA

General features

Reynier Canyon, a feature of moderate relief, has been cut in Tertiary and Quaternary sediments, which, in general, dip northward from the crystalline rocks of the north flank of the San Gabriel Range. The canyon, for most of its course, follows the strike of a relatively unresistant, silty part of the Mint Canyon formation. The canyon walls are underlain principally by the relatively resistant sandy and conglomeratic beds lower in the Mint Canyon formation, and by the cliff-forming sandstone and shaly beds of the "Modelo" formation as exposed in the outliers flanking the canyon (pl. I).

The stream channel is dry except during periods of heavy winter rains. It drains from a common divide (elevation about 2010 feet) with Placerita Canyon, and joins Sand Canyon at a point about one mile to the northeast (elevation about 1800 feet). The cliffs along the southern edge of the larger outlier rise to an elevation of about 2300 feet. The Canyon bottom, for the lower half of its length, has an alluvial filling into which the present stream channel has cut to a depth of as much as 10 feet.

The basal "Modelo" beds containing the fauna herein described are immediately west to west-southwest of the Reynier Canyon-Sand Canyon confluence, and are exposed low along the northwestern wall of Reynier Canyon itself.

The accompanying geologic map (fig. 1) which was prepared on a scale of 600 feet to the inch, covers an area approximately 0.6 mile long and 0.4 mile wide along the Mint Canyon-"Modelo" contact. An enlargement of a part of the 1:24,000 scale Sylmar quadrangle was used as a base. The geologic map is merely a large-scale refinement of a very small part of Kew's reconnaissance map.

Mint Canyon formation

The Mint Canyon formation in the Reynier Canyon area is less well exposed than elsewhere in the eastern Ventura basin. Kew[9] has shown

9/ Kew, W. S. W., op. cit., accompanying map.

that south of Reynier Canyon, the northwest-trending San Gabriel fault has brought Mint Canyon beds against the pre-Tertiary crystalline rocks of the San Gabriel Range. Between this contact and the southernmost

Plate I. View northwestward of north wall of Reynier Canyon showing C.I.T. locality 1849. "Modelo" strata exposed on cliffs; Mint Canyon strata underlie gentle slopes in foreground.

basal "Modelo" exposures of the larger outlier, the Mint Canyon beds dip steeply to moderately northward. This section, totaling about 3000 feet in thickness, is conglomeratic in its lower third and sandy and silty in its upper two thirds. Its general lithologic features are outlined in Table 1.

The beds of the upper part of the Mint Canyon formation, underlying the lower slopes of the northwestern side of Reynier Canyon, are mostly hidden beneath a grass covered surface. Several landslides also have partly obscured the geology, but a tuff bed, within the Mint Canyon formation and close to the "Modelo" contact, is relatively well exposed for the length of the accompanying map. As traced laterally, this bed successively diverges from and converges on the basal "Modelo" beds.

At a point near the western edge of the map the tuff bed and basal "Modelo" beds are separated by a thickness of approximately 300 feet of Mint Canyon sandstone, but at two places within four-tenths of a mile to the east, the basal "Modelo" beds overlap the tuff bed. At the westernmost overlap, the unconformity is marked by an angular discordance of at least 45 degrees.

"Modelo" formation

Beds of the "Modelo" formation in the Reynier Canyon area are best exposed in the cliffs along the canyon's northwest wall. Here a thickness of about 400 feet of marine sandstone, siltstone and diatomaceous shale has been folded in a broad syncline plunging gently northward. An outline of the formations lithologic features is included in

Table 1.

<u>Measured section of Mint Canyon and "Modelo" formations exposed in Reynier Canyon area</u>

Pico formation, unmeasured sandstones and conglomerates.

unconformity

"Modelo" formation

4.	Siltstone and diatomaceous shale, very pale orange; evenly bedded. C. I. T. loc. 1626 200 feet above base of "Modelo" formation.	220 feet
3.	Sandstone, very pale orange to pale yellowish orange; in alternating layers of fine-grained massive and shaly strata; layers about 10 feet thick. Grades laterally westward into diatomaceous shale similar to unit 4. Cliff-forming.	120
2.	Sandstone, yellowish gray, fine-grained; becomes shaly upward; gypsiferous, locally fissile. Abundant iron-stained fractures.	40
1.	Sandstone, yellowish gray, fine-grained, friable, massive to evenly bedded. Contains discontinuous concretionary layers with abundant marine megafossils. Conglomeratic layers common and basal conglomerate prominent in eastern Reynier Canyon area. Boulders mostly granitic; maximum diameter 2 feet. Contains C. I. T. Loc. 1849.	20
	Total thickness of "Modelo" formation	400

unconformity

Mint Canyon formation

11.	Sandstone and siltstone, light gray to pale yellowish orange. Pale olive beds in lower part.	300

10.	Tuff, vitric, pale gray; fine-grained, friable to compact. Continuous.	15
9.	Sandstone and siltstone, similar to unit 8, concretionary layers more abundant. Non-tuffaceous.	80
8.	Sandstone and siltstone, light gray to pale yellowish orange; pebbles become less abundant upwards. Local hard concretionary layers. Highly lenticular vitric tuff beds in lower part. Poorly exposed	750
7.	Sandstone and siltstone, same as below, but with local orange layers.	250
6.	Sandstone and siltstone, same as below, but color predominantly very pale yellow.	150
5.	Sandstone and siltstone, grayish yellow to yellowish gray. Pebbly conglomerate layers common.	125
4.	Sandstone, grayish yellow, medium-grained, pebbly. Conglomeratic layers become more abundant upward; fragments mostly granitic, locally volcanic, better rounded than in lower conglomerate.	90
3.	Siltstone, light gray to light olive gray. Mostly poorly stratified and poorly cemented, but shaly layers are common. Contains a few discontinuous tuffaceous layers several inches thick. Underlies nonresistant grassy slopes.	370
2.	Siltstone, moderate reddish brown to moderate yellowish brown becoming grayish upward, interbedded with conglomerate. Alternating layers several feet thick. Conglomerate very poorly sorted, contains angular to subrounded fragments mostly leucogranite, granite gneiss and mica schist.	500
1.	Sandstone, arkosic, with interbedded pebble-to-boulder conglomerate. Pale to dark yellowish orange. Poorly stratified, poorly cemented. Conglomerate contains angular to subrounded fragments of leucogranite, granite gneiss and biotite schist. Poorly exposed.	500
	Total thickness of Mint Canyon formation	3100 feet

Table 1.

In the eastern half of the area mapped the basal "Modelo" beds contain pebble to boulder conglomeratic lenses as much as 10 feet thick. The overlying beds, in general, grade upward from fine-grained sandstone through siltstone to diatomaceous shales. Prominent in the section, however, are layers about 10 feet thick alternately coarser and finer grained. Also noteworthy is an east to west lateral gradation from sandy to shaly beds.

In addition to the megafossils noted in the following section, the Reynier Canyon "Modelo" beds are reported to contain a microfossil assemblage of Mohnian age[10].

10/ Durham, J. W., personal communication.

FAUNA OF THE "MODELO" FORMATION

Localities

Most of the invertebrate megafossils in the "Modelo" sediments of Ryenier Canyon have been found in strata approximately 15 feet above the basal beds in the eastern half of the accompanying map. The remains are particularly abundant in conglomeratic and calcareous concretionary lenses along this horizon. More than half of the specimens in the collection described in this paper were gathered at the locality C.I.T. loc. 1849. During the early stages of the study, the collections gathered at different places along this fossiliferous horizon were kept separate; but when found to contain virtually identical assemblages, they were combined.

Faunal list, C.I.T. loc. 1849

Pelecypoda

Chione cf. pabloensis (Clark)

C. cf. fernandoensis English

Clementia cf. martini Clark

C. (Egesta) cf. pertenuis (Gabb)

Cryptomya cf. ovalis Conrad

Cyathodonta cf. pedroana Dall

Dosinia arnoldi Clark

Laevicardium (Nemocardium) centifilosum (Carpenter)

L (Trachycardium) cf. quadragenarium (Conrad) var. fernandoense (Arnold)

Lucina (Myrtea) cf. acutilineata Conrad

L. (Here) cf. excavata Carpenter

L. (Myrtea) nuttallii Conrad

L. (Miltha) xantusi (Dall)

Macoma indentata Carpenter

M. secta (Conrad)

M. cf. planiscula Grant and Gale

Marcia (Compsomyax) cf. subdiaphana Carpenter

Nuculana taphria (Dall)

Ostrea cf. titan Conrad

Panope generosa Gould

Paphia tenerrima Carpenter

Pecten (Lyropecten) estrellanus (Conrad)

Sanguinolaria cf. nuttallii Conrad

<u>Solen</u> cf. <u>perrini</u> Clark

<u>Spisula</u> <u>albaria</u> Conrad var.

<u>Tellina</u> <u>idae</u> Dall

<u>Tivela</u> (<u>Pachydesma</u>) <u>diabloensis</u> Clark

<u>Venus</u> (<u>Chione</u>) cf. <u>elsmerensis</u> (English)

<u>V.</u> (<u>C.</u>) cf. <u>parapodema</u> Dall

Gastropoda

<u>Astraea</u> (<u>Pomaulax</u>) <u>gradata</u> Grant and Gale

<u>Bursa</u> sp.

<u>Calliostoma</u> cf. <u>costatum</u> (Martyn)

<u>C.</u> cf. <u>supragranosum</u> Carpenter

<u>Cancellaria</u> <u>elsmerensis</u> English

<u>C.</u> <u>hemphilli</u> Dall

<u>C.</u> <u>tritonidea</u> Gabb

<u>Calcantharus</u> <u>breaensis</u> (Carson)

<u>Calyptraea</u> <u>filosa</u> (Gabb)

<u>Cantharus</u> <u>elsmerensis</u> Carson

<u>Conus</u> <u>californicus</u> Hinds

<u>Crepidula</u> (<u>Crepipatella</u>) cf. <u>charybdis</u> Berry

<u>C.</u> <u>adunca</u> Sowerby

<u>C.</u> <u>aculeata</u> (Gmelin)

<u>C.</u> <u>princeps</u> Conrad

<u>Crucibulum</u> <u>imbricatum</u> (Sowerby)

<u>Ficus</u> (<u>Trophosycon</u>) <u>ocoyana</u> (Conrad)

<u>Kelletia</u> <u>jahnsi</u> n. sp.

<u>Mitrella</u> cf. <u>carinata</u> (Hinds) var. <u>gausapata</u> (Gould)

<u>Murithais</u> <u>eldridgei</u> (Arnold)

<u>Polinices</u> (<u>Neverita</u>) <u>reclusianus</u> (Deshayes) var. <u>callosus</u> (Gabb)

<u>Sinum</u> <u>scopulosum</u> (Conrad)

<u>Surculites</u> (<u>Megasurcula</u>) <u>remondii</u> (Gabb)

<u>Turritella</u> <u>cooperi</u> Carpenter

<u>T.</u> <u>vanvlecki</u> Arnold sub. sp. <u>hemphilli</u> Applin MS

A second fossiliferous horizon, containing a less numerous assemblage and apparently less continuous than the horizon described above, is exposed at locality C.I.T. loc. 1626. This is approximately 200 feet above the base of the "Modelo" formation.

Descriptions of species

Anadara cf. obispoana (Conrad)

Plate II, figure 1.

Arca obispoana Conrad, Pacific R. R. Rept., vol. 7, p. 192, pl. 5, fig. 1.

Anadara (Scapharca ?) obispoana subsp. obispoana (Conrad), Reinhart, Geol. Soc. America, Spec. P. no. 47, pp. 70-72, pl. 10, figs. 6, 7, 11.

Well preserved external casts of Anadara cf. obispoana abundant at C.I.T. loc. 1626, approximately 200 feet stratigraphically higher than C.I.T. loc. 1849. Shell flat, elongate; height about 2/3 length. Markedly inequilateral; beak close to anterior end. Hinge margin straight. Anterior and posterior margins sharply arcuate; ventral margin broadly arcuate. Sculpture of about 28 relatively flat ribs which become wider posteriorly. Interspaces at anterior end equally as wide as ribs. Posterior ribs have no distinct interspaces. Ribs contain poorly preserved fine lines, which are most pronounced near ventral margin. Entire rib pattern slightly concave toward dorsal posterior margin. Faint concentric growth lines most prominent near posterior and posterior ventral margins. Dentition not well preserved. Figured specimen 42 millimeters long, 29 millimeters high.

Compared with previously figured specimens of <u>Anadara obispoana</u>[11/],

<u>11</u>/ Arnold, R., Descriptions of new Cretaceous and Tertiary fossils from the Santa Cruz Mountains, California; U. S. Nat. Mus., Pr., vol. 32, no. 1545, pl. 35, fig. 1, 1908.

Branner, J. C., Newsom, J. F., and Arnold, R., U. S. Geol. Survey, Santa Cruz Folio, fig. 50, 1909.

Grant, U. S. IV, and Gale, R. H., Pliocene and Pleistocene mollusca of California: San Diego Soc. Nat. History, Mem., vol. 1, pl. 32, fig. 49, 1931.

Reinhart, P. W., Mesozoic and Cenozoic Arcidae from the Pacific slope of North America: Geol. Soc. America, Spec. Paper, p. 47, pl. 10, figs. 6, 7, 11, 1943.

the Reynier Canyon specimens have a more extended posterior ventral margin, a beak that is closer to the anterior end, narrower interspaces between the posterior ribs, and an average of 2 additional ribs.

The occurrences of <u>Anadara obispoana</u>, as listed by Reinhart[12/],

<u>12</u>/ Reinhart, P. W., op. cit., p. 71.

are in the Monterey and Temblor formations (both Miocene). A reported Pliocene occurrence of the species[13/], in western Kamchatka, is dis-

<u>13</u>/ Slodkewitsch, W. S., Tertiary pelecypoda from the Far East: Acad. Sci. U. S. S. R., Paleant Inst., Paleontology of U. S. S. R., vol. 10, pt. 3, fasc. 19, pt. II, p. 106, pl. XI, fig. 4,

credited by Reinhart. The species has not been noted in the Pliocene of California.

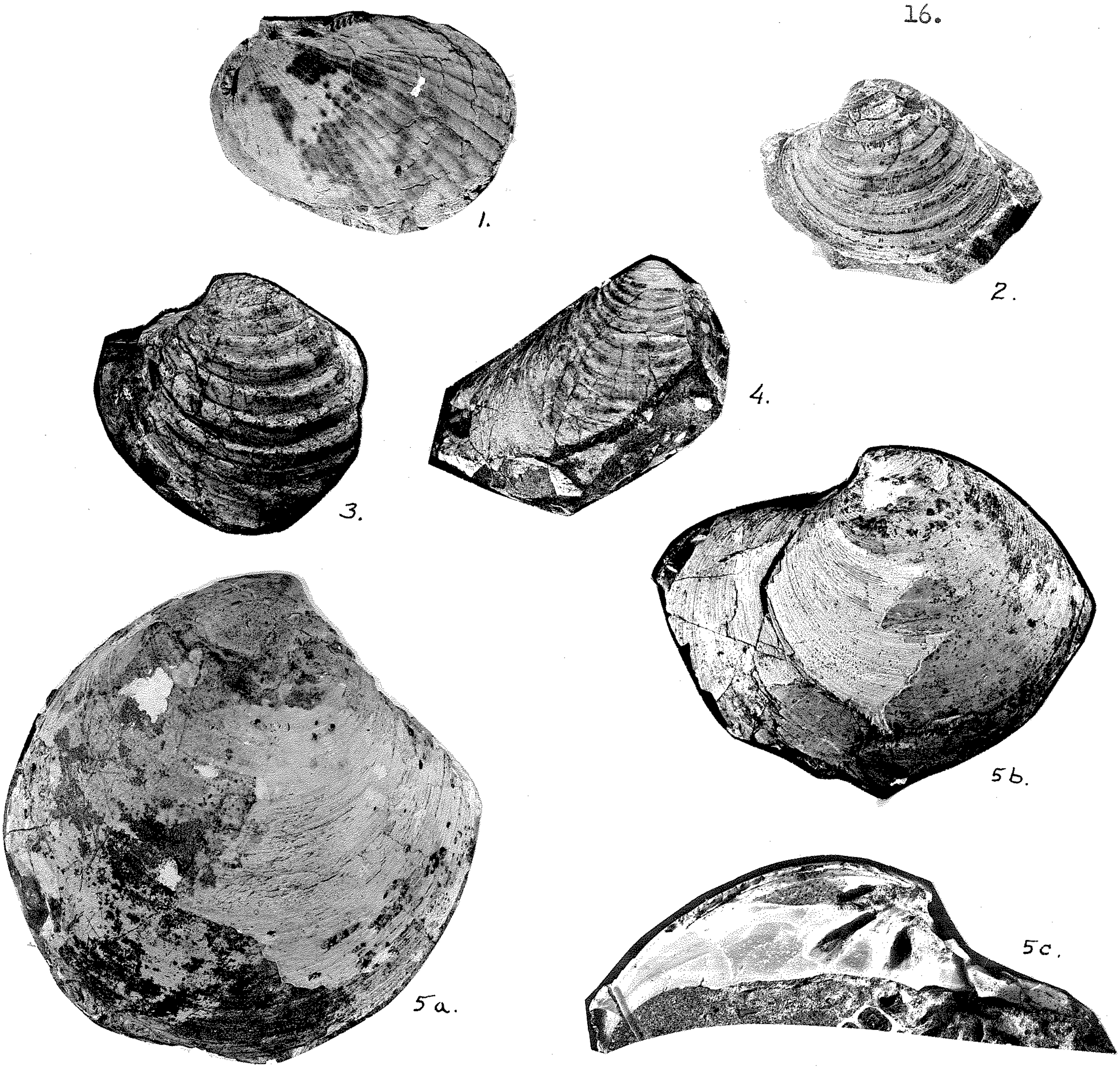

Plate II. Pelecypods from C. I. T. localities 1849 and 1626.
1. *Anadara* cf. *obispoana* (Conrad), right valve, locality 1626.
2. *Chione fernandoensis* English, left valve, locality 1849.
3. *Clementia* cf. *martini* Clark, left valve, locality 1849.
4. *Clementia* (*Egesta*) cf. *pertenuis* (Gabb), right valve, locality 1849.
5. *Dosinia arnoldi* Clark; 5a, exterior of right valve; 5b, exterior ventral part of left valve; 5c, dentition of 5b; x 1 1/2, locality 1849.

Clementia cf. martini, Clark

Plate II, figure 3.

Venus martini Clark, Univ. California Publ. Geol., vol. 8, pp. 470-471, pl. 54, fig. 1, 1915.

A single specimen (left valve) of Clementia cf. martini was collected at C.I.T. loc. 1849. The ventral margin and anterior and posterior extensions are not preserved. Shell thin. Anterior dorsal edge markedly concave, upper part of posterior dorsal edge nearly straight; lower part not preserved. Surface ornamented by heavy, evenly spaced lines. Dentition poorly preserved, but does show remnants of three fairly heavy cardinals.

The pronounced concavity of the anterior dorsal margin of the Clementia martini specimen distinguishes it from C. pertenuis.

Clementia martini has been noted in the San Pablo (upper Miocene) formation east of San Francisco Bay and in the Santa Margarita (upper Miocene) formation north of Coalinga[14/]. It is unreported in Pliocene

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14/ Clark, B. L., Fauna of the San Pablo group of middle California: Univ. California Publ. Geol., vol. 8, pp. 470-471, 1915.

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formations.

Clementia (Egesta) cf. pertenuis (Gabb)

Plate II, figure 4.

Venus pertenuis Gabb, Geol. Surv. Calif., Palaco., vol. 2, pp. 55-56, pl. 5, fig. 37, 1868-9.

Clementia (Egesta) pertenuis (Gabb), Woodring, U. S. Geol. Survey Prof. Paper 147-C, pp. 40-42, pl. 16, figs. 1-6, 1926. Grant and Gale, San Diego Soc. Nat. History, Mem. 1, p. 335, 1931.

Two incomplete right valves of Clementia (Egesta) cf. pertenuis were collected at C.I.T. loc. 1849. The anterior end of each specimen is poorly preserved, but the shape of the shell is apparently only slightly inequilateral. Posterior end elongate; posterior slope convex. Lunular area only slightly depressed. Coarse, concentric waves pronounced on dorsal margin. Fine concentric threads appear lower on shell. Near ventral margin waves disappear and threads become prominent. Hinge of larger specimen, though not perfectly preserved, shows three cardinal teeth: a thin, anterior cardinal; a thicker middle cardinal; and the upper part of a thin, posterior cardinal. The larger specimen, when complete, was probably more than 55 millimeters in length and more than 45 millimeters in height.

Twenty-two of the 23 Clementia (Egesta) pertenuis localities listed by Woodring[15/] are in Miocene formations. The other locality, re-

15/ Woodring, W. P., American Tertiary mollusks of the genus Clementia: U. S. Geol. Survey Prof. Paper, 147-C, pp. 40-42, 1926.

ported by Branner, Newsome, and Arnold[16/], is in the lower part of the

16/ Branner, J. C., Newsome, J. F., and Arnold, Ralph, U. S. Geol. Survey Geol. Atlas, Santa Cruz folio (No. 163), pp. 5-6, 1909.

Purisima, a formation thought by them to be partly upper Miocene, but now considered to be Pliocene.

Dosinia arnoldi Clark

Plate II, figures 5a, b, c.

Dosinia arnoldi, Clark Univ. California Publ. Geol., vol. 8, pp. 459-460, pl. 51, figs 1, 2, 1915.

Several specimens of Dosenia arnoldi were collected at C. I. T. loc. 1849. Most are poorly preserved, but the exterior of one right valve is nearly complete. Another, a left valve, has a moderately well preserved dentition. Shell circular in outline; length and width almost equal. Lunule well impressed. Anterior dorsal shoulder extends prominently from umbo. Posterior dorsal margin slopes away from beak in a relatively broad, uniform arc. Surface near posterior dorsal margin does not have the depression noted in Clark's description of the type[17/]. Surface of the shell covered with fine concentric growth

17/ Clark, B. L., Fauna of the San Pablo group of middle California: Univ. California Publ. Geo., vol. 8, pp. 459-460, 1915.

lines, closely, though irregularly, spaced. Posterior part of hinge plate has a lower margin that is relatively straight, but curves sharply downward near the posterior margin of the shell. Three cardinal teeth in left valve. Anterior cardinal relatively thin and short, inclined slightly toward anterior margin of shell. Middle cardinal fairly heavy and inclined toward posterior margin at angle of about 45 degrees. Posterior cardinal inclined at relatively low angle, but poorly preserved on all Reynier Canyon specimens. Height of figured specimen 73 millimeters; width 73 millimeters.

Grant and Gale[18/] list Dosinia arnoldi, D. jacalitosana, and

18/ Grant, U. S., IV, and Gale, R. H., op. cit., p. 352.

D. Merriami as synonyms of D. ponderosa var. jacalitosana. The Reynier Canyon specimens, however, closely resemble the type arnoldi at the University of California, and differ in several respects from the type merriami.

The prominently extended anterior dorsal shoulder and the large radius of curvature of the posterior margin of arnoldi are not shown in jacalitosana. The anterior cardinal in the left valve of ponderosa var. jacalitosana as figured by Grant and Gale[19/] has a marked posterior

19/ Grant, U. S., IV, and Gale, H. R., op. cit., pl. 15, fig. 3.

inclination which is shown neither in the Reynier Canyon specimens nor in Clark's figured type. The socket separating the anterior and middle cardinals is, therefore, distinctly wider in arnoldi than in jacalitosana. The Reynier Canyon specimens have a larger radius of curvature of the dorsal posterior margin and a more pronounced extension of the anterior shoulder than does merriami.

The type locality of Dosinia arnoldi is in the upper San Pablo (upper Miocene). It has also been found in Santa Margarita (upper Miocene) beds north of Coalinga[20/]; but it not known in the Pliocene.

20/ Clark, B. L., op. cit., p. 460.

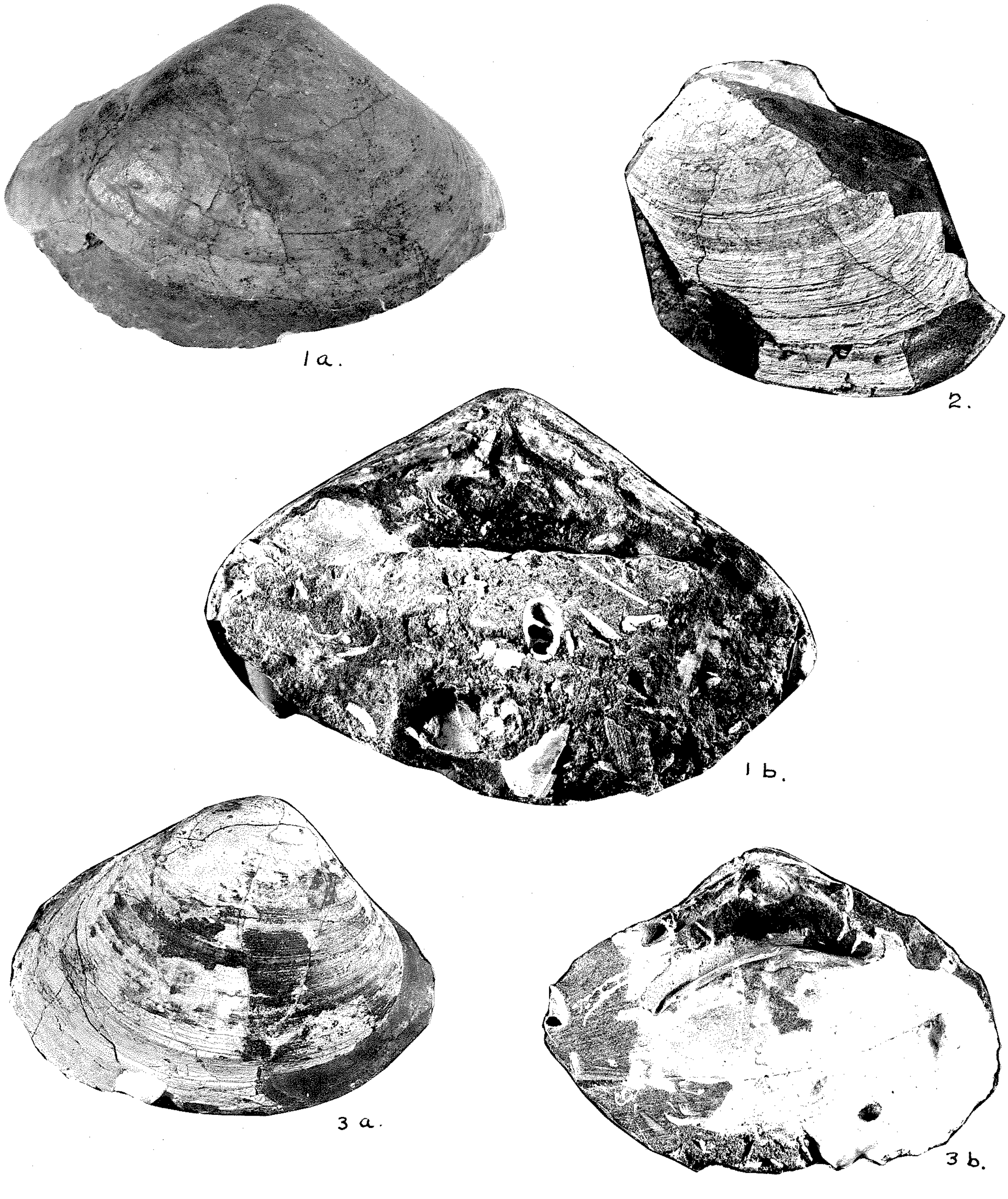

Plate III. Pelecypods from C. I. T. locality 1849.

1. Spisula albaria (Conrad); 1a, exterior of right valve; 1b, interior of same valve, x 1 1/3.
2. Paphia cf. tenerrima Carpenter; right valve.
3. Tivela (Pachydesma) diabloensis Clark; 3a, exterior of right valve; 3b, interior of same valve.

Spisula albaria (Conrad)

Plate III, figures 1 a, b.

Mactra albaria Conrad, Amer. Jour. Sci., Ser. 2, vol. 5, p. 432, fig. 4, 1848, reprinted by Dall, U. S. Geol. Survey Prof. Paper 59, p. 150, fig. 4, 1909.

Spisula (Hemimactra) albaria Conrad, Dall, U. S. Geol. Survey Prof. Paper 59, pl. 130-131, pl. 10, fig. 1, 1909.

Spisula albaria (Conrad), Clark, Univ. California Publ. Geol., vol. 8, p. 420, pl. 60, fig. 8, 1915; Packard, Univ. California Publ. Geol., vol. 9, p. 290, pl. 24, fig. 1, pl. 25, figs. 3-8, 1916.

Mactra (Spisula) albaria Conrad, Grant and Gale, San Diego Soc. Nat. History, Mem. 1, pp. 395-396, pl. 23, figs. 3a, 3b, 1931.

One specimen, a right valve, of Spisula albaria was collected at C.I.T. loc. 1849. It is fairly well preserved but ventral margin is incomplete. Shell medium size, rather heavy, trigonal, nearly equilateral. Height approximately 2/3 length. Dorsal margins slope from beak at about same angle. Anterior dorsal margin slightly convex. Posterior dorsal margin straight. Both dorsal slopes rather wide. Lunular area depressed. Low posterior umbonal ridge. Shell surface smooth except for fine growth lines which are clustered in irregularly spaced concentric waves. Dentition poorly preserved but apparently similar to the albaria dentitions figured by Packard[21].

21/ Packard, E. L., Mesozoic and Cenozoic Mactrinae of the Pacific Coast of North America: Univ. California Publ. Geol., vol. 9, pl. 25, figs. 3-6, 1916.

The Reynier Canyon Spisula albaria specimen is markedly more elongate than the Pliocene and Miocene albaria specimens figured by Packard[22], but closely resembles in shape the Miocene specimen figured

22/ Packard, E. L., op. cit., pl. 24, fig. 1, pl. 25, figs. 3-8, 1916.

Dall[23] and the Miocene (?) specimen figured by Grant and Gale[24].

23/ Dall, W. H., The Miocene of Astoria and Coos Bay, Oregon: U. S. Geol. Survey Prof. Paper 59, pl. 10, fig. 1, 1909.

24/ Grant, U. S., IV, and Gale, H. R., op. cit., pl. 23, figs. 3a, 3b, 1931.

Tivela (Pachydesma) diabloensis Clark

Plate III, figures 3a, b.

Tivela (Pachydesma) diabloensis, Clark, Univ. California Publ. Geol., vol. 8, pp. 462-463, pl. 54, figs. 5, 6, pl. 55, fig. 1, 1915.

One specimen, a right valve, of Tivela (Pachydesma) diabloensis, was collected at locality C.I.T. loc. 1849. Shell heavy, trigonal, nearly equilateral. Height approximately 2/3 length. Posterior dorsal margin slightly concave near beak, anterior extension slightly convex. Posterior dorsal margin strongly depressed, but the carina noted by Clark[25], on the type specimen is missing (not preserved ?) on the

25/ Clark, B. L., op. cit., pp. 462-463.

Reynier Canyon specimen. Anterior dorsal margin but slightly depressed; has poorly developed carina. Closely spaced growth lines become clustered in crude concentric waves toward ventral margin. Right valve has three cardinals; posterior cardinal elongate and parallels dorsal margin; middle cardinal inclined toward posterior end; anterior cardinal slightly inclined to anterior end. Reynier Canyon specimen 74 millimeters long, approximately 54 millimeters high.

Tivela (Pachydesma) diabloensis has been noted in the Lower San Pablo Group (Upper Miocene), but not in Pliocene sediments.

Kelletia jahnsi n. sp.

Plate IV, figures 3a, b, c.

An undescribed Kelletia species is relatively abundant at C.I.T. loc. 1849. It is of medium size and weight. Shell fusiform. Spire highly elevated; spiral angle about 40 degrees. Five to seven whorls, ventricose. Body whorl has rounded outline with slight angulation about one-fourth of its length in front of suture. Succeeding whorls are distinctly angulated above the middle. Body whorl has numerous, irregularly spaced, fine crenulations; succeeding whorls have 9 to 11 pronounced nodes extending from suture to suture. In front of angulation, nodes are nearly vertical, but have a slight convex curve. Suture distinct and straight. Spiral ornamentation a system or riblets. Spiral whorls have about 14 riblets, all of nearly uniform width except a distinctly wider riblet in front of suture. Body whorl has about 16 riblets in front of angulation, 8 riblets on shoulder. Riblets on body whorl below angulation are wider than those above and have a faint groove slightly in front of the middle. Fine growth lines are only longitudinal sculpure. Aperture ovoid, constricted at canal. Canal relatively long and sharply recurved. Figured specimen 56 millimeters long, 25 millimeters wide at maximum diameter of body whorl.

Kelletia jahnsi n. sp. var.

Plate IV, figures 3b, c.

Several specimens collected at C.I.T. loc. 1849 differ markedly from Kelletia n. sp. in the shape of the body whorl, but in all other features are similar. The body whorl in these variants has as pronounced an angulation and as prominent nodes as the spiral whorls. Some of these specimens are as large or larger than the type Kelletia n. sp., and apparently are not immature individuals.

Conclusions

As shown in tables 2 and 3, more than half of the Reynier Canyon species have previously been noted in the lower Pliocene assemblages of Elsmere and Holser Canyons a few miles to the west. Indeed 16 of the Reynier Canyon species apparently have not been reported below the lower Pliocene. In general, however, these species have a limited geographic range and are undiagnostic. Seven species from Reynier Canyon - Anadara cf. Dosinia arnoldi, Ostrea cf. titan, Tivela diabloensis, and

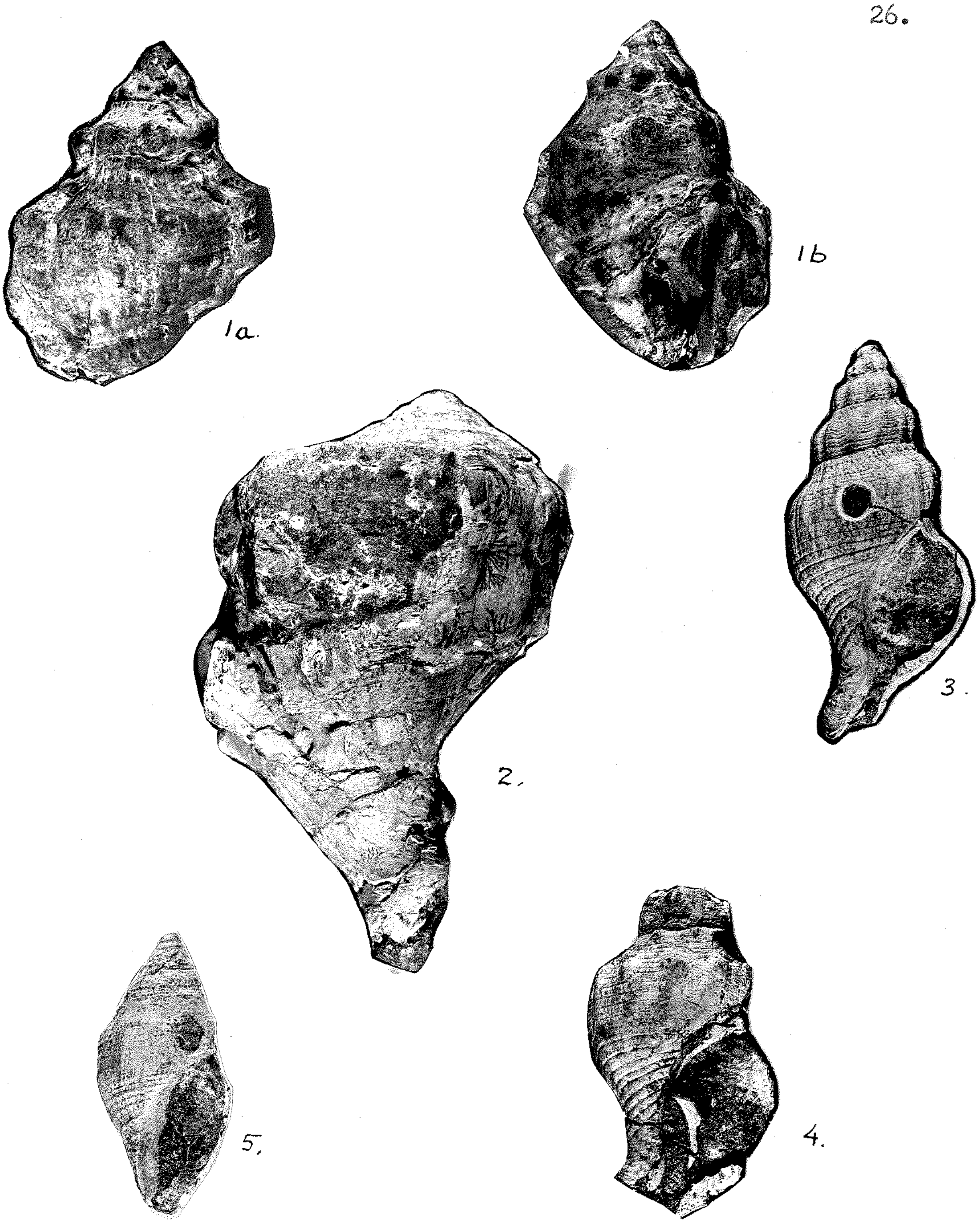

Plate IV. Gastropods from C. I. T. locality 1849; natural size.

1. Bursa n. sp. (?).
2. Ficus (Trophosycon) ocoyana (Conrad), about twice as large as average specimen from this locality.
3. Kelletia jahnsi n. sp., rounded body whorl.
4. Kelletia jahnsi n. sp. var., angulated body whorl.
5. Surculites (Megasurcula) remondii (Gabb).

Spisula albaria - are widely recognized as pre-lower Pliocene forms. The species of this group, a typical upper Miocene assemblage, have not been reported in the Elsmere and Holser Canyon faunas. Conversely in these faunas the forms that are believed diagnostic of the lower Pliocene - *Patinopecten lohri*, *Astrodapsis* sp., and *Dendraster* sp. - are apparently missing in Reynier Canyon. Therefore, the Reynier Canyon fauna, although closely related to the faunas of Elsmere and Holser Canyons, is probably ancestral to them, and belongs to the upper Miocene of the invertebrate paleontologists.

Pelecypoda

Chione pabloensis (Clark)

C. fernandoensis English

Clementia cf. martini Clark

C. (Egesta) cf. pertenuis (Gabb)

Cryptomya ovalis Conrad

Cyathodonta pedroana Dall

Dosinia arnoldi Clark

Laevicardium (Nemocardium) centifilosum Carpenter

L. (Trachycardium) quadragenarium (Conrad) var. fernandoense (Arnold)

Lucina (Myrtea) acutilineata Conrad

L. (Here) excavata Carpenter

L. (Myrtea) nuttallii Conrad

L. (Miltha) xantusi (Dall)

Macoma indentata Carpenter

M. secta (Conrad)

M. planiscula Grant and Gale

Marcia (Compsomyax) subdiaphana Carpenter

Nuculana taphria (Dall)

Ostrea cf. titan Conrad

Panope generosa Gould

Paphia tenerrima Carpenter

Pecten (Lyropecten) estrellanus (Conrad)

Sanguinolaria cf. nuttallii Conrad var.

Solen cf. perrini Clark

Spisula albaria Conrad var.

Tellina idea Dall

Tivela (Pachydesma) diabloensis Clark

Venus (Chione) cf. elsmerensis (English)

V. (C.) parapodema Dall

Table 2. Geologic range of pelecypod species from C.I.T. loc. 1849. Species from closely related, nearby faunas shown by E (Elsmere Canyon), and H (Holser Canyon).

Oligocene	Miocene			Pliocene			Pleistocene	Recent
	L	M	U	L	M	U		
			X					
				E	H		?	
		X	X					
		X	X	?				
			X	X	X	X	X	X
						?	X	X
			X					
			?	X			X	X
				E				
X		X	X	EH	X	X	X	X
				X	X		X	X
			X	EH	X		X	X
	?	X	X	EH				
			X	EH			X	X
			X	H			X	X
		?						X
		?		EH	X		X	X
			?	X	EH	X	X	X
			X					
		X	X	EH		X	X	X
				EH			X	X
		X	X	X				
			X	X			X	X
				X	X			
X			X					
		X	X	X	X	X	X	X
			X					
			?	E				
			?	EH?	H?		?	

Gastropoda	Oligocene	Miocene L	Miocene M	Miocene U	Pliocene L	Pliocene M	Pliocene U	Pleistocene	Recent
Astraea (Pomaulax) gradata Grant and Gale					H				
Bursa sp.									
Calliostoma cf. costatum (Martyn)						X		X	X
Calliostoma supragranosum Carpenter								X	X
Cancellaria elsmerensis English					E				
C. hemphilli Dall						H			
C. tritonidea Gabb					E				
Calcantharus breaensis (Carson)					X				
Calyptraea filosa (Gabb)				X	E	?			
Cantharus elsmerensis Carson					E				
Conus californicus Hinds					E	H	X		
Crepidula (Crepipatella) charybdis Berry									
C. adunca Sowerby				X		X		X	X
C. aculeata (Gmelin)						X		X	X
C. princeps Conrad		X		X	E	H	X	X	
Crucibulum imbricatum (Sowerby)						X		X	X
Ficus (Trophosycon) ocoyana (Conrad)					E	X			
Kelletia jahnsi n. sp.									
Mitrella cf. carinata (Hinds) var. gausapata (Gould)					E				
Murithais eldridgei (Arnold)	X				E				
Polinices (Neverita) reclusianus (Deshayes) var. callosus (Gabb)	X	X	X	X	H				
Sinum scopulosum (Conrad)	X			X	E	H		X	X
Surculites (Megasurcula) remondii (Gabb)				X	E	X		X	X
Turritella cooperi Carpenter					E H	H			
T. vanvlecki Arnold sub. sp. hemphilli Applin MS					X	H			

Table 3. Geologic range of gastropod species from CIT loc. 1849. Species from closely related, nearby fauna shown by E (Elsmere Canyon) and H (Holser Canyon).

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GEOLOGY AND ORIGIN OF TALC DEPOSITS
OF EASTERN CALIFORNIA

Thesis by

Lauren A. Wright

In Partial Fulfillment of the Requirements

for the Degree of

Doctor of Philosophy

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Mr. Rudolph von Huene prepared approximately 200 thin sections, many of which required much more than average attention. Of particular value were three giant sections of White Eagle mine specimens, each showing the entire alteration sequence of adamellite to talc.

Courtesies extended by members of the California talc industry are too numerous to list here. The cooperative interest of this group, was a major factor in furthering the project. Capable and willing field assistance was given by the following men: E. M. Shoemaker, R. S. Orr, M. D. Turner, and W. E. Ver Planck.

Funds for the payment of two rock analyses were generously supplied by the Division of the Geological Sciences, California Institute of Technology. Other analytical data have been furnished by the Sierra Talc and Clay Company, the Western Talc Company, and the Southern

California Minerals Company. The writer is further indebted to Drs. Campbell, Jahns, and Engel for critically reading the manuscript.

GEOLOGY AND ORIGIN OF TALC DEPOSITS OF EASTERN CALIFORNIA

ABSTRACT

A 200-mile northwest-trending belt in eastern California contains more than 100 talc-bearing localities. The belt is divisible into three districts; each with talc deposits in a distinctive terrane.

In the southernmost district, near Silver Lake, tremolitic talc deposits have formed in highly metamorphosed Archean (?) sedimentary rocks extensively invaded by lamprophyre and granitic rocks. The development of talc-tremolite rock probably involved the extensive introduction of MgO and SiO_2 to silica poor dolomite in a complex, multi-stage history. The MgO may have been released in the granitization of high-magnesian sediments.

In the Southern Death Valley-Kingston Range region tremolitic talc deposits occur at or near the borders of a diabase sill intruded near the base of the lowermost carbonate unit in the Algonkian Pahrump series. The deposits have generally altered from dolomite, both siliceous and silica-poor. Other alteration rocks rich in alkali feldspar are associated with the talc deposits and border diabase bodies higher in the member. MgO and SiO_2 have been introduced to form the talc bodies; SiO_2, Al_2O_3, K_2O, and probably Na_2O to form the feldspathic rocks. Diabase magma was probably the source of most of the additive material, but some may have been derived from connate water.

In the Inyo Range area tremolite-poor talc deposits have formed as alterations of Paleozoic dolomite and quartzite, and of Mesozoic granitic intrusives. The White Eagle deposit shows all three types of alteration,

but has formed largely as a replacement of adamellite. Feldspars, quartz and ferromagnesian minerals were decomposed; the talc alteration followed an advance wave of albitization. The additive MgO probably was leached from dolomites at depth.

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GEOLOGY AND ORIGIN OF TALC DEPOSITS OF EASTERN CALIFORNIA

INTRODUCTION

GENERAL STATEMENT

All but a small part of the output of commercial talc[1] in

1/ In commercial usage, the term "talc" is applied to numerous mineral mixtures composed predominantly of magnesium silicate minerals. In such mixtures, the mineral talc is commonly, but not necessarily, a prominent constituent. In this report, the term "talc", unless otherwise qualified, will refer to the commercial material.

California and Nevada has been obtained from deposits within an elongate belt near and roughly paralleling the central part of the border between the two states (fig. 1). Nearly 100 known talc deposits of commercial interest are contained within this belt, which is approximately 200 miles in length and 30 miles in average width. It extends from the vicinity of Silver Lake in north-central San Bernardino County northwestward to the Palmetto-Oasis district of western Nevada, and includes the southern Death Valley and Kingston Range region in California, and the Inyo Range of California.

Although the geographic restriction of the deposits at first suggests a persistent genetic relationship, from place to place within the belt talc has formed in markedly different geologic environments. Even to the casual observer, the talc deposits of California are separable into three talc districts, each within a distinctive geologic terrane.

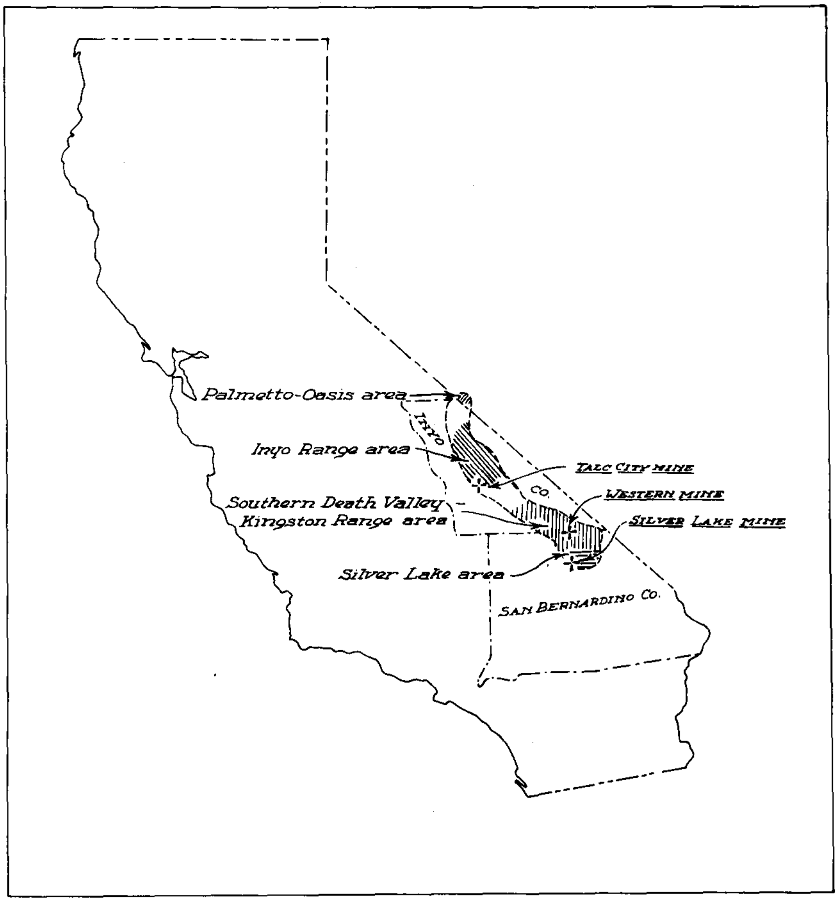


Figure 1. Index map showing location of talc-bearing belt of eastern California and western Nevada, its four component districts, and the location of the three largest mines.

The southernmost talc-bearing area, which is here designated as the Silver Lake talc district, is relatively small. Its principal deposits are centered at two localities, one a few miles north of the Silver Lake playa, and another 12 miles to the north-northeast of Silver Lake near the small settlement of Yucca Grove on the highway of U. S. Routes 91 and 466. At these localities, highly metamorphosed sedimentary rocks are the hosts of the talc bodies. Granitic rocks underlie most of the intervening area and also intimately penetrate the metasediments.

At the Silver Lake locality, several commercial talc bodies have been developed by workings known collectively as the Silver Lake mine. The output of the Yucca Grove area has been obtained from two operations, the Calmasil and Pomona mines. Evidence to be cited later indicates a probable Archean age for both the metasediments and the talc bodies.

A second talc district occupies an area of approximately 1000 square miles north and northwest of the Silver Lake province, and extends from the southeastern part of the Panamint Range eastward to the eastern part of the Kingston Range. This area, which will be referred to as the Southern Death Valley-Kingston Range talc district, contains about 40 known talc-bearing localities. All of the deposits are in the Crystal Spring formation, the lowest of three formations that comprise the pre-Cambrian Pahrump series of Algonkian age. Each deposit is an alteration of carbonate strata near the base of a massive, carbonate member in the middle of the formation. Each is also in contact with or near a diabase sill that ordinarily separates the carbonate member from underlying units of quartzite, argillite and shale.

The mines in this district that have been most actively worked since 1940 are the Montgomery and Warm Spring mines in the southeastern Panamint Range; the Monarch, Pleasanton and Ibex mines near Ibex Springs at the southern end of the Ibex Hills; the Superior, White Cap, and Pongo mines north of Saratoga Springs at the southern end of Death Valley; the Western mine in the Alexander Hills; and the Smith and Excelsior mines on the west and east flanks respectively of the Kingston Range. Because of their economic importance, these are the deposits of the southern Death Valley-Kingston Range region that have been studied most closely by the writer, and will be the ones most frequently referred to in the description of the district.

The deposits of the third talc district of eastern California are most closely grouped near the southern end of the Inyo Range in an area about 13 miles southeast of Keeler, Inyo County. Other talc deposits, however, are also exposed on the west and east flanks of the Inyo Range, in Eureka Valley, and in the northern part of the Panamint Range. Most of the talc in this district has formed as an alteration of Ordovician and Silurian dolomite and quartzite, but most of the talc in one large body, the White Eagle deposit, has formed as a replacement of adamellite (quartz monzonite).

Approximately 20 talc deposits of commercial interest have been noted in this area, which, in this report, will be referred to as the Inyo Range talc district. Most of the deposits contain a variety of talc known to the ceramic industry as steatite[2]. This material is an

2/ Steatite, in its present ceramic meaning, is an essentially pure talc (maximum allowable CaO and Fe_2O_3 contents each 1.5 percent), that is adaptable to the techniques of electrical insulator manufacture.

essential ingredient in the manufacture of certain high frequency electrical insulators. It was in critical demand during World War II and remains so.

Most of the talc produced in the Inyo Range district has been obtained from a single property, the Talc City mine, near the southern end of the range and about midway between the towns of Keeler and Darwin. The White Mountain mine, on the east slope of the Inyo Range and above the south end of Saline Valley, has the second largest talc output of the mines in this district. The other talc properties in the province have been worked less continuously and on a smaller scale than these two. In this report, only the white Eagle deposit will be described.

The deposits of the Palmetto-Oasis district in Nevada are similar to those of the Inyo Range and may well represent an extension of the Inyo Range talc district. The Nevada deposits, which were also studied in the U. S. Geological Survey's wartime steatite program, will not be discussed in this report.

SCOPE AND METHODS OF THE PRESENT INVESTIGATION

From May to December, 1942, the writer, then an employee of the U. S. Geological Survey, assisted Dr. B. M. Page in studies of actual and potential steatite sources in California, Nevada and New Mexico. This program, which centered largely about the Inyo Range and Palmetto-Oasis district, involved plane-table mapping of most of the larger deposits that were believed to contain talc of steatite purity. Included in this group was the White Eagle deposit (pl. 4), mapped in May 1942. Petrographic studies of the White Eagle deposit were begun in the winter of 1947 when the writer was a student at the California Institute of Technology, and were continued at intervals during the winter of 1948 while he was employed by the California Division of Mines.

The investigation of the Silver Lake and Southern Death Valley-Kingston Range talc provinces has been a project of the California Division of Mines. This phase of the study has required five field months and four office months during the period of October 1947-February 1950.

A plane-table map on a scale of 100 feet to the inch was made of most of the Silver Lake talc zone (pl. 1). The Western (pl. 2) and the Superior-White Cap mine areas (pl. 3) were mapped by plane-table on a scale of 200 feet to the inch. Additional data were gathered during the inspection of about 30 other deposits in the two districts. Because observations in the Southern Death Valley-Kingston Range district indicated a stratigraphic control of talc deposition, a better understanding of the pre-Cambrian stratigraphy of the region was deemed desirable. Consequently, detailed sections were measured at ten of the localities at which the lower part of the late pre-Cambrian series appears to be best exposed and least deformed. These localities are: (1) near Warm Springs in the southeastern Panamint Range; (2) on the northern flank of the Owlshead Mountains; (3) north and (4) west of Ibex Springs at the southern end of the Ibex Hills; at the (5) Superior and (6) Saratoga mines north of Saratoga Springs; (7) at the Western mine in the Alexander Hills; (8) at the Rogers mine on the western slope of the Kingston Range; (9) at the type locality of the Crystal Spring formation in the west central part of the Kingston Range; and (10) on the western slope of the Silurian Hills. Most of the stratigraphic features of these sections will be but briefly discussed in this report. Emphasis will be placed on the features that bear most directly on the origin of the talc deposits.

SILVER LAKE TALC DEPOSITS

PHYSICAL FEATURES

The talc deposits of the Silver Lake area are in a group of hills about 10 miles northeast of the Silver Lake playa. The deposits are 14 miles north-northeast of Baker, a settlement on the highway of U. S. Routes 91 and 466 connecting Barstow and Las Vegas.

The Silver Lake mine is 16 miles by road from Baker. It is reached by traveling 8 miles on State Highway 127 north-northwest from Baker, and thence 8 miles east-northeast by a graded dirt road. The two roads join at the site of the old Silver Lake siding on the now abandoned

Tonopah and Tidewater Railroad. This siding which is on the north edge of the playa was formerly the shipping point for the Silver Lake mine. The talc is now trucked to Dunn siding, a point on the Union Pacific Railroad about 23 miles west-southwest of Baker.

The altitude of the Silver Lake playa is slightly less than 1000 feet. The talc-bearing zone lies at an average altitude of 2500 feet, and the nearby hills are as much as 300 feet higher. Most of the talc deposits are exposed low on the southern slope of a west-trending ridge, but the two westernmost deposits underlie low rises on relatively level ground. West of these rises, bedrock is hidden beneath alluvium.

Talc deposits have been worked at 6 localities along the zone's two mile length. The five workings shown on the accompanying map (pl. 1) are spaced at intervals of approximately 1200 feet. From west to east, these are the (1) Western Addenda, (2) Eastern Addenda, (3) Gould, (4) Number Two and One-half, and (5) Number Two workings. At each of these, one or two shafts have been sunk on talc bodies. The deposits have been worked by overhand stopes connected to drifts in talc.

The sixth locality at which talc bodies have been opened is at the eastern end of the zone. The development here, known as the Number Four workings, is mainly a group of irregular drifts, gently inclined stopes, and rooms connected to the surface by a vertical shaft.

The Gould workings have a maximum strike-length of about 1500 feet, and are by far the most extensive of the Silver Lake mine area. In 1919 the Gould talc deposits were opened by a shaft, which subsequently was sunk to a depth of about 250 feet. The talc was hoisted through this shaft until 1925, when the shaft was intersected at about the 150-foot level by an adit driven eastward from a canyon wall. Since then the talc has been trammed through the adit.

Because precipitation in the Silver Lake area averages less than 3 inches annually, the mine workings are dry even to their lowest levels. The water consumed at the mine is trucked from a well about 5 miles to the south.

The region contains a very sparse vegetation; most of the hill-slopes are virtually barren bedrock exposures. Some of the slopes in the mine area, however, are overlain by relatively thick talus slopes and remnants of an earlier alluvial cover.

PREVIOUS INVESTIGATION

Very little is known of the geologic features of the Silver Lake region outside of the immediate area of the accompanying map. In a reconnaissance study Miller[3/] briefly described metasedimentary and

3/ Miller, W. J., Crystalline rocks of southern California: Geol. Soc. America Bull., vol. 57, pp. 498-501, 1946.

granitic rocks exposed in the hills between the mine and the Silver Lake siding. These and similar rocks exposed near Halloran Spring (about 5 miles west-southwest of Yucca Grove) were believed by Miller to be early pre-Cambrian in age and to belong to the Halloran Complex as defined by him. Although the Silver Lake mine area was but briefly mentioned by Miller[4], he noted the presence of both granitic

[4] Miller, W. J., op. cit., pp. 490-500.

and metasedimentary rocks, and tentatively correlated them with the Halloran Complex.

GENERAL GEOLOGY

The Silver Lake talc deposits lie in a series of highly metamorphosed but relatively undeformed sedimentary rocks. Most of the deposits occur in a narrow zone, not more than 50 feet wide and discontinuously exposed for a distance of about 2 miles. In the westernmost 5000 feet (pl. 1) of the talc-bearing zone, the deposits and the enclosing metasedimentary rocks dip moderately to steeply southward and are cut by several cross-faults. At the eastern end of the zone, the deposits are in the southern part of a structural block characterized by broad open folds but bordered on the east, south, and west by highly brecciated and intricately faulted blocks of both granitic rock and metasediments that contain no talc bodies. The relatively simple structural features of both the southward-dipping and the folded parts of the talc-bearing zone have been greatly complicated by the emplacement of lamprophyre and several varieties of granitic rocks.

In the absence of detailed geologic data on the Silver Lake region outside of the immediate area of the mine, neither the age of the metasediments nor the time of formation of the talc bodies can be determined with certainty. Miller's[5/] suggestion of an early pre-

5/ Miller, W. J., op. cit.

Cambrian age for the metasedimentary granitic rocks north of the Silver Lake siding, was based on the presence of granite-free, slightly metamorphosed Carboniferous sediments immediately west of Baker. The presence of relatively unmetamorphosed sediments in the Silurian Hills about 8 miles north of the Silver Lake mine area also suggests an early pre-Cambrian age for the Silver Lake metasediments; but Hewett[6/] has

6/ Hewett, D. F., The geology and mineral resources of the Ivanpah Quadrangle: U. S. Geol. Survey Prof. Paper, in press.

shown that a wide belt of Mesozoic quartz monzonite lies at least as close as 5 miles northeast of the mine area. The possibility that some or all of the granitic rocks of the mine area are genetically related to this rock must not be overlooked. Hewett[7/] has noted, however, that

7/ Hewett, D. F., personal communication.

the quartz monzonite is a structurally homogeneous unit. None of the granitic rocks of the Silver Lake area are as devoid of structure. Indeed both planar and linear features are well displayed in them. Because of the slight degree of regional metamorphism shown by the late pre-Cambrian rocks nearest the mine area, and the contrast in the respective structures of the granitic rocks of the mine area and the Mesozoic quartz monzonite, the writer agrees with Miller's suggestion of an early pre-Cambrian age for the rocks of the talc-bearing belt.

The actual age of the talc bearing metasedimentary rocks of the Silver Lake talc district may well remain in question, however, because they appear to be geographically restricted and are not known to be in contact with other metamorphic or sedimentary rocks of proved age.

Because the western part of the Silver Lake talc zone is the most continuously exposed and the least faulted, the spatial relations and general petrologic features of all of the rock units are best shown here. The following discussion will center mainly about this area, but the same rock types are also in the eastern part of the zone.

ROCK UNITS

METASEDIMENTARY ROCKS

General features

The metasedimentary units shown on the accompanying map are part of a considerably thicker section, exposed for many hundreds of feet both north and south of the talc-bearing zone. The section's continuity has been broken by the emplacement of granitic rocks. These predominate in the west part of the mine area where they enclose numerous metasedimentary "islands". Nevertheless five metasedimentary rock members can be recognized throughout most of the western 5000 feet of the talc-bearing zone. In upward succession, these are here named the hornfels, quartz-biotite schist, quartz-muscovite schist, forsterite marble, and quartzite members. Their petrologic features are outlined

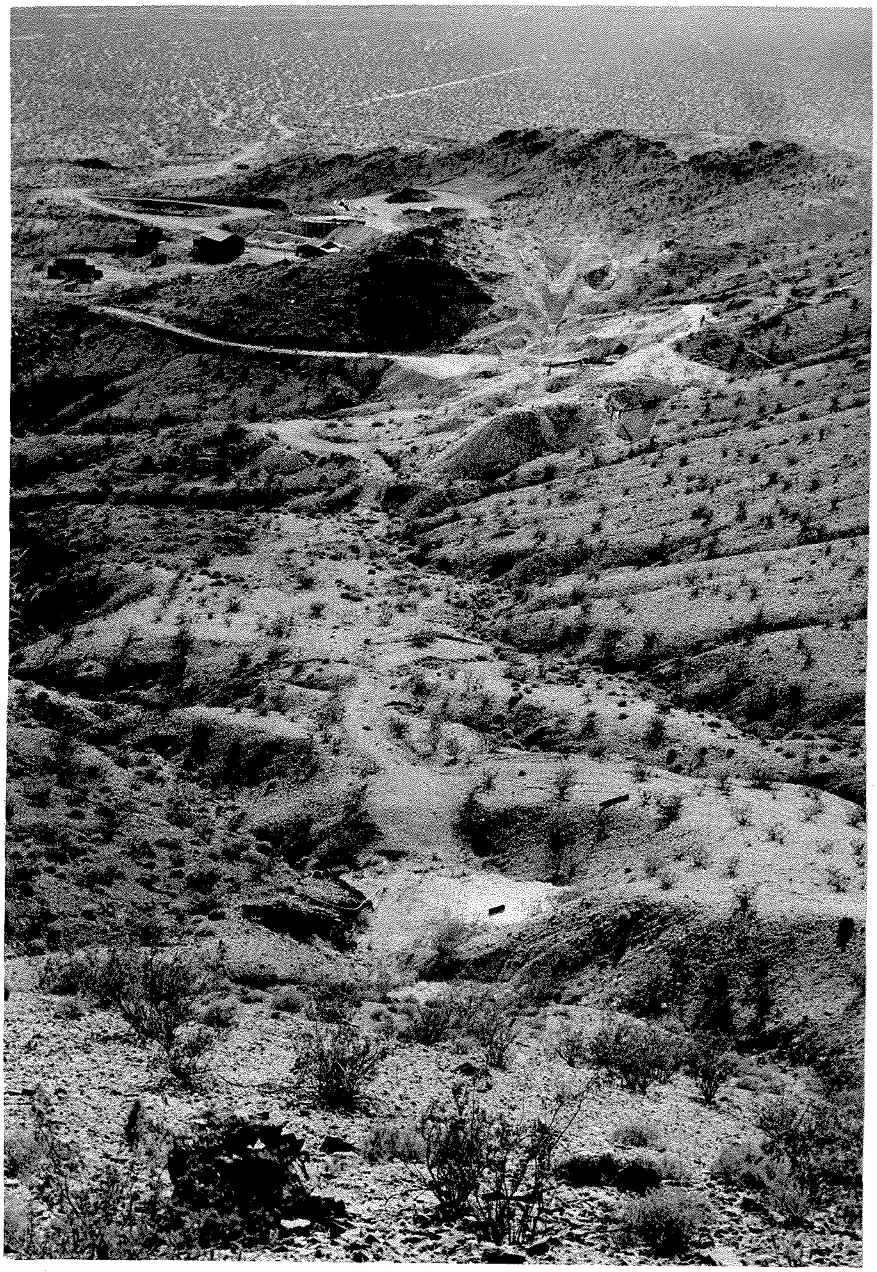


Plate 5. View westward along western part of Silver Lake talc zone. Gould workings in middle distance, Western Addenda workings on flat in upper right corner. Talc bodies dip steeply southward.

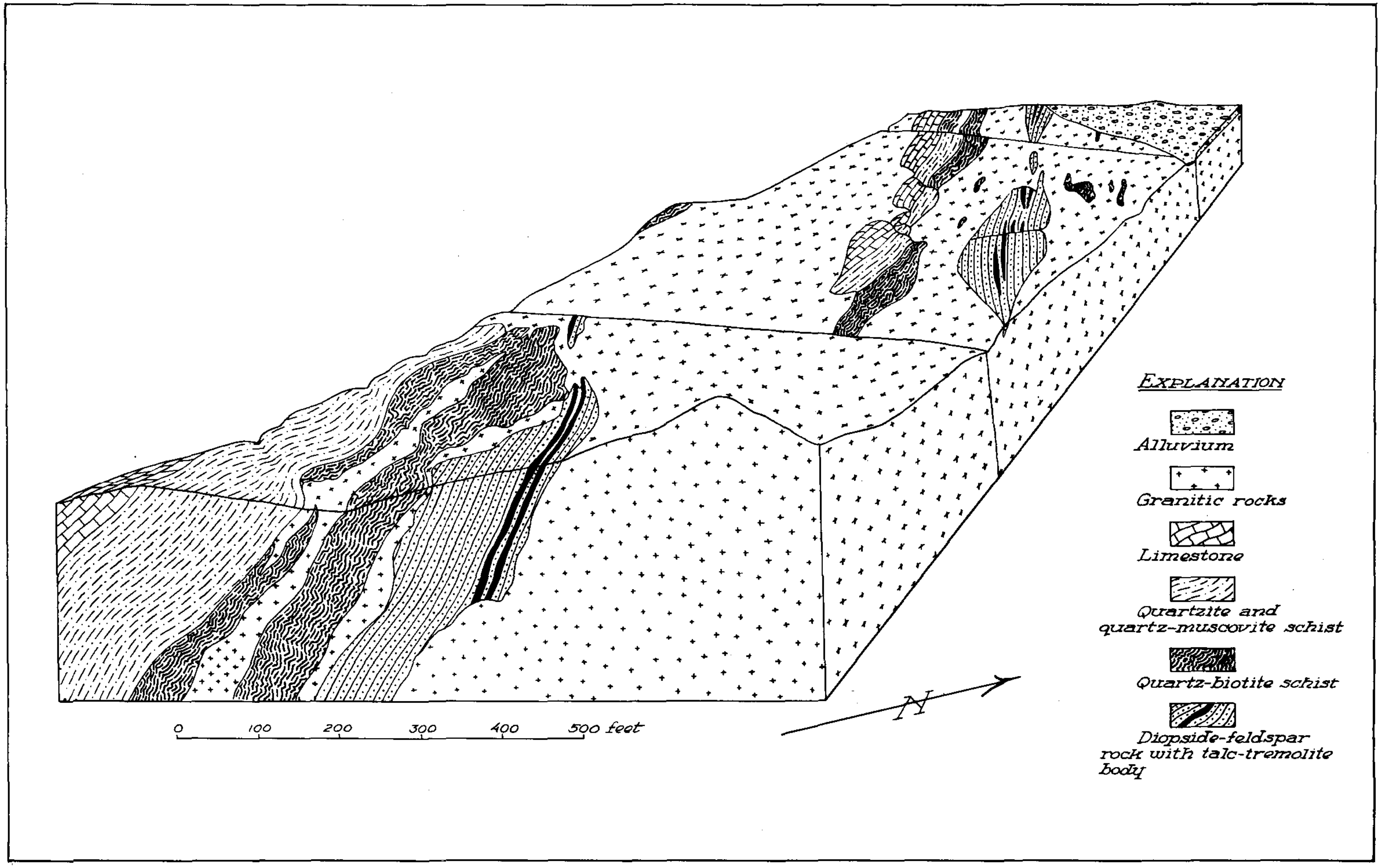


Figure 2. Generalized block diagram of western part of Silver Lake talc-bearing area.

in Table 1. That this is the true stratigraphic succession is shown by their occurrence in the same order upward in the broadly folded rocks at the eastern end of the zone.

The most characteristic structural features shown by the outcrop pattern of these units are (1) a general lack of deformation except for the uniform tilting and broad folding, and (2) a parallelism of the metasedimentary planar textures with contacts of sedimentary origin. Many bodies that clearly were once sedimentary strata, such as marble layers in quartzite and quartzite layers in schist and hornfels are commonly only a few feet thick and several hundreds of feet long. Such layers have nearly straight or broadly curved traces. They are ordinarily parallel with each other and with comparably undeformed contacts that separate the five principal metasedimentary units.

Among the textural features that parallel these sedimentary planes are the schistosity of most of the schist occurrences, foliation planes in the hornfels and tremolitic rock, a planar distribution of forsterite grains in tremolitic rock and in marble, and the flattening of quartz grains in quartzite and quartz-mica schist.

Hornfels member

Distribution

The lowermost and most northerly exposed of the five metasedimentary members is also the most heterogeneous. Because a green diopsidic rock with a hornfelsic texture predominates, the name "hornfels member" is applied. This member also contains all of the large concentrations of talc and tremolite, as well as subordinate amounts of other non-hornfelsic rock types. The northern boundary of the member is mostly in contact with granitic rocks; but large elongate masses of biotite schist and muscovite schist are included in the granitic rocks and appear to be remnants of units stratigraphically beneath the member.

The hornfels member has a maximum thickness of about 150 feet but it is ordinarily much thinner. The thinning is caused principally by the encroachment of granitic rocks and is not, in general, a stratigraphic or deformational feature. The hornfels member appears to have resisted such encroachment more effectively than the bordering mica schists. Consequently, in the western part of the mine area, where the granitic rocks are the most extensive, the hornfels is the principal island-forming rock. The Western Addenda and Eastern Addenda talc deposits are in two such islands.

The rock types within the hornfels member, in general, exist as distinct units in sharp contact with one another, but the textural and mineralogic variants of the diopsidic hornfels commonly have gradational relationships.

Petrology

The hornfels itself is essentially a diopside-feldspar-quartz-calcite rock; but garnet and phlogopite are locally prominent. The rock ranges in color from pale green to grayish olive green, and is generally fine-grained, although some is medium- to coarse-grained.

In most exposures, the rock is dense and tough, but in some it is relatively friable. From place to place within the member, the rock ranges from thinly and evenly layered to massive.

Schist, composed of various proportions of tremolite, actinolite, phlogopite and alkali feldspar, and locally containing an appreciable proportion of biotite, forms conspicuous layers in the hornfels member, but is less extensive than the diopsidic rock. The micas and amphibole grains are in very uniform dimensional alignment. The schist ranges in color from dark through light green to yellowish gray. In sunlight, the phlogopite-rich rock has a lustrous, golden sheen. Layers of tremolite-phlogopite-albite schist, from a few inches to a foot or two thick, ordinarily separate the commercial talc bodies from diopsidic wall rock. Along the hanging wall of one of the Gould talc bodies, the schist is as much as 5 feet thick. Schistose layers within diopsidic rock range from a fraction of an inch to several feet in thickness.

The member contains a few thin quartzite beds, commonly traceable for several hundred feet. Marble is likewise a subordinate unit. East of the Gould workings, a very thin part of the member contains marble as the principal metasedimentary rock, but has been extensively invaded by pegmatite.

The following comments on the talc-tremolite bodies anticipate data to be included in a more detailed section to follow, but bear upon the origin of the member as a whole. The bodies range in width from several inches to about 15 feet, are as much as 800 feet long, and consistently occur near the center of the hornfels member. Some grade laterally into diopsidic rock, others abut against granite, still others are terminated by cross-faults. The talc-tremolite bodies commonly occur in pairs, each body from 5 to 15 feet thick and separated by from 10 to 15 feet of diopside hornfels.

In addition to talc and tremolite, the bodies also contain appreciable proportions of forsterite, serpentine, calcite, chlorite (?), and phlogopite. The silicates have formed in the following general sequence: forsterite, tremolite (first generation), chrysotile, tremolite (second generation) and phlogopite, chlorite (?) and talc.

Petrography

Hornfelsic rocks. Microscopic studies of the truly hornfelsic facies of the hornfels member show that, despite a wide range in the relative proportions of minerals, it does have persistent mineralogic and textural characteristics. In thin sections of each of 20 rock specimens collected at widely distributed points within the member, both diopside and feldspar are present. These minerals are generally accompanied by quartz and calcite. Garnet and phlogopite, though much less abundant, are locally prominent in certain layers. In a few sections, the diopside has been partly altered to talc or serpentine, but such alteration is rare.

The grains of diopside, feldspar, and quartz, the principal silicates, are generally of uniform size and average less than 0.3 mm. in diameter. From place to place within the member, these minerals are present in widely different proportions, and locally any one of the three may predominate. The pronounced, thin banding typical of much of the diopsidic facies is caused chiefly by alternating diopsidic and felsic layers.

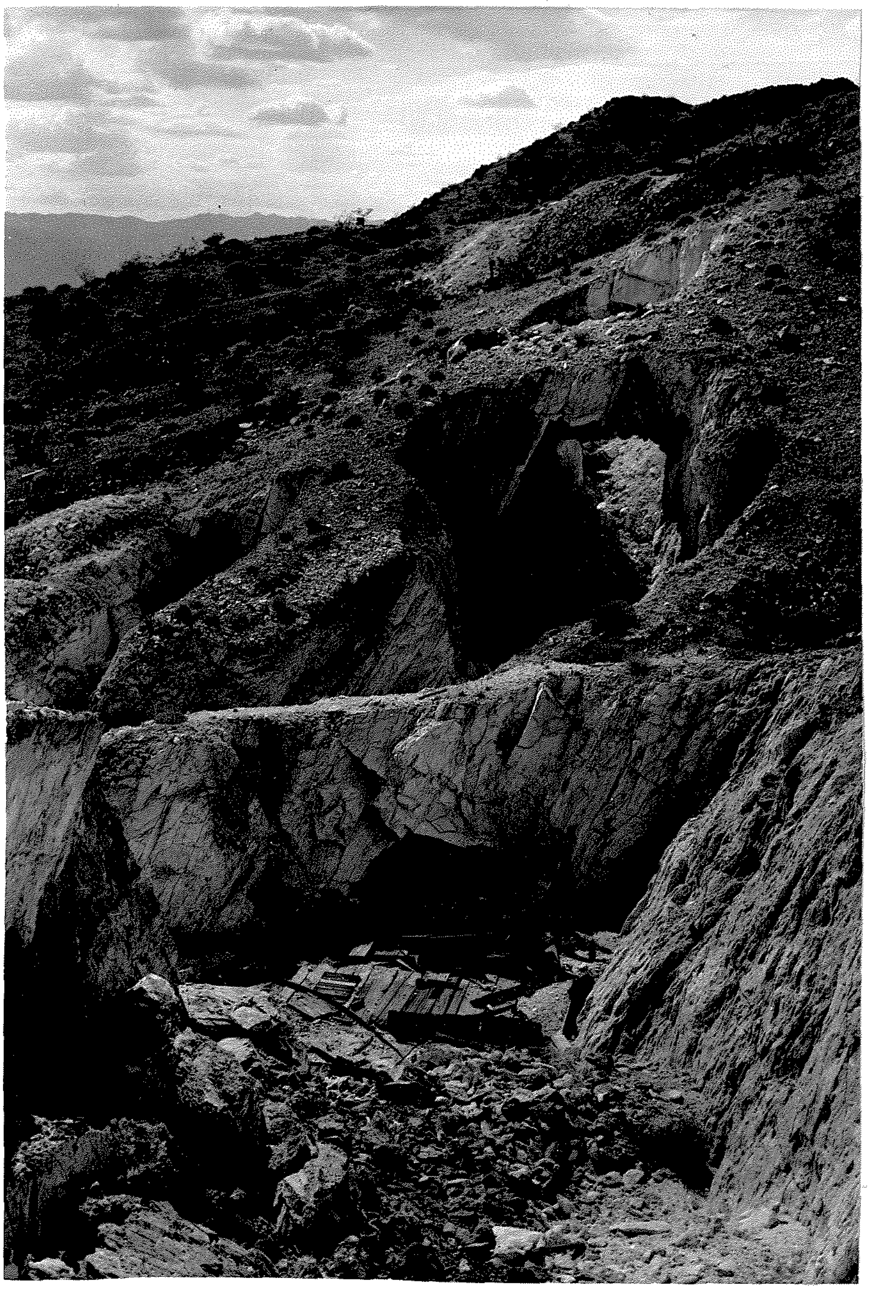


Plate 6. View westward along talc-bearing zone in hornfels member. Stoped material is composed mostly of tremolite and talc. Wall rock is a diopside-feldspar-quartz-calcite hornfels.

Most of the diopside is in subhedral grains forming a mosaic texture with quartz and feldspar; but some of the diopside in a few thin sections is in relatively large, very irregular grains that are deeply embayed and transected by grains of quartz and microcline. In the specimens that were examined microscopically, the diopside content ranges from about 10 to more than 95 percent.

Both plagioclase and potash feldspars are abundant in the diopsidic facies as a whole, although in large masses of the rock, feldspar is very subordinate to diopside. Locally, either potash or plagioclase feldspar composes more than half of the rock, and in a few places layers several feet thick were noted in which quartz and alkali feldspar occur to the exclusion of diopside. Most of the plagioclase is albite or sodic oligoclase, but in some specimens it is as calcic as bytownite. In one section, bytownite grains are as much as 6 mm. in diameter and enclose smaller equant diopside grains.

The garnet is straw yellow in transmitted light, and occurs in small, irregular grains as part of the diopside-feldspar-quartz mosaic. Garnet grain clusters several millimeters in length are common. Such clusters are characteristically elongate parallel with the banding of the rock.

Calcite in the hornfels phase persistently forms the coarsest grains. Although most are less than 1 mm. in diameter, diameters of as much as 1 cm. are not uncommon. The calcite occurs in veinlets and in large irregular poikilitic grains that surround smaller grains of silicate minerals.

Most of the talc, serpentine, and phlogopite in the diopsidic rock apparently occur within a few feet of the borders of the talc-tremolite bodies. Near these borders minute layers of the tremolite-phlogopite schist are commonly interlayered with diopside-feldspar rock. Sparsely distributed tremolite and dark green amphibole, however, were observed elsewhere in the diopsidic rock.

Schistose rocks. In thin section, specimens typical of the schistose rocks of the hornfels member are shown to be uniformly fine-grained, but to have a wide range in mineral composition. Most of the rock is composed of grains from 0.5 to 1.5 mm. in length. Some layers contain only tremolite or actinolite; others contain tremolite with subordinate phlogopite and alkali feldspar; still others are composed principally of phlogopite. Diopside is notably lacking.

A representative specimen of the schist that separates the commercial talc bodies from wall rock contains approximately 60 percent tremolite, 20 percent phlogopite, and 20 percent alkali feldspar. In this rock, some of the tremolite blades exceed 1 mm. in length. These are commonly corroded by aggregates of finer grained phlogopite and feldspar. Some of the feldspar grains show plagioclase twinning, and apparently all have indices less than that of balsam. Some, if not all, of the feldspar, therefore, is albite or sodic oligoclase.

Metamorphism

General statement. Because the metasedimentary rocks of the Silver Lake mine area cannot be traced into less metamorphosed facies, the origin of the hornfels member as well as the other metasedimentary units, is somewhat obscured. The nature of the original rocks, the role of metasomatism, and the influence of stress are each partly speculative. The bodies of hornfels, quartzite and talc-tremolite rock, stratiform in appearance and with contrasting bulk chemical compositions, apparently reflect chemical or physical differences in original sedimentary layers. Metamorphic differentiation seems to have produced the small-scale textural laminations, but most of the gross layering, involving laterally persistent units several or more feet thick, is probably a relict sedimentary feature.

Shearing in non-schistose tremolite rock has produced zones of talc schist; the thin laminations are probably an effect of metamorphic differentiation in a stressed environment. Widespread metasomatism is shown in the alteration of forsterite to tremolite, serpentine, and talc, of tremolite to serpentine, carbonate, and talc, and of diopside to tremolite; but none of the contacts between the large metasedimentary bodies are clearly of metasomatic origin nor are they related in space to contacts with granitic rocks.

The quartzite, carbonate rocks and schist of the hornfels member have more abundant counterparts higher in the section. The following discussion of the member's metamorphism, therefore, will be concerned mainly with the hornfelsic rocks.

Derivation of present chemical composition. Mineral assemblages identical or similar to those of the hornfelsic rocks of the Silver Lake area have been noted at numerous other localities where marly and arenaceous dolomites have been metamorphosed under conditions characteristic of the amphibolite facies. Hornfelses of this origin were described in 1914 and 1915 by Eskola[8/] in his classic descriptions

8/ Eskola, P., On the petrology of the Orijärvi region in southwestern Finland: Comm. géol. Finlande Bull., no. 40, 1914.

------, On the relation between chemical and mineralogical composition in the metamorphic rocks of the Orijärvi region: Comm. géol. Finlande, Bull., no. 44, 1915.

of the rocks of the Orijärvi region of Finland. Similarly derived hornfelses in other parts of the world have been described since by others, including Adams and Barlow[9/], Sugi[10/], and Benson and Bartrum[11/].

9/ Adams, F. D., and Barlow, A. R., Geology of the Haliburton and Bancroft areas, Province of Ontario, Geol. Surv. Canada, Mem. no. 6, 1910.

10/ Sugi, K., A preliminary study on the metamorphic rocks of southern Abukuma Plateau: Jap. Jour. Geol. and Geogr., vol. 12, nos. 3-4, pp. 115-151, 1935.

11/ Benson, W. N., and Bartrum, J. A., The geology of the region about Chalky and Preservation inlets: III; Roy. Soc. New Zealand, Trans. vol. 65, pt. 2, pp. 108-152, 1935.

If isochemical reconstitution is assumed, the diopside-plagioclase-microcline-calcite-quartz assemblage, which composes most of the truly hornfelsic rock of the hornfels member, would have been derived from original impure dolomite rich in alumina, silica, potash and soda. The abundance of albite in some of the sections indicates a Na_2O fraction of about 3 to 5 percent, which is distinctly higher than normal for an impure dolomite, and suggests that much of the Na_2O may have been introduced. A chemical analysis of a specimen typical of the feldspar-diopside hornfels, however, shows 1.26 percent of Na_2O which may well have been present in the original rock. Textures in which microcline and quartz appear to have formed at the expense of diopside suggest at least a local introduction of potash and alumina into other rocks of the hornfels member, an introduction also indicated by the presence in tremolite bodies of late-stage phlogopite, occurring as fracture-controlled veinlets and bordering granitic dikes.

The preponderance of sodic over calcic feldspar in a calcite-bearing rock, however, need not necessarily be attributed to the metasomatic introduction of soda, but could well have been caused by high pressure and shearing stress. Such conditions are known[12/] to restrict

12/ Turner, F. J., The genesis of oligoclase in certain schists: Geol. Mag., vol. lxx, pp. 529-541, 1933.

the lime content of plagioclase. The formation of microcline in preference to muscovite, probably was controlled by the excess of alkalies; whether of sedimentary or hydrothermal origin. As indicated by Turner[13/], in all but highly magnesian rocks, potash feldspar in an

13/ Turner, F. J., Evolution of the metamorphic rocks: Geol. Soc. Am. Mem. 30, p. 94, 1948.

environment of calcite will always crystallize in preference to mica if the weight ratio $\frac{CaO + K_2O + Na_2O}{Al_2O_3}$ exceeds unity.

Although diopside is present in the hornfelsic rocks to the almost complete exclusion of other magnesium-bearing minerals, elsewhere in the mine area it occurs in replacement veinlets in forstertic marble. These veinlets suggest the possibility that at least some of the diopside in the hornfels is of hydrothermal origin. The diopside in the hornfels, however, is not veinlet-forming, and probably was derived mostly from original constituents. The thorough dissemination of quartz suggests that silica also was originally abundant.

Physical conditions and equilibrium relations. Significant to a consideration of the physical environment and equilibrium relations of the hornfelsic rocks are (1) the preponderance of diopside over amphibole and the post-diopside age of most or all of the amphibole, (2) the abundance of quartz and the absence of forsterite, and (3) the widespread quartz-calcite association and the absence of wollastonite.

Bowen[14/], in his classic discussion of the progressive metamorphism of siliceous limestone and dolomite, notes that diopside first appears in the fourth of 13 steps that mark increasing decarbonation with rising temperature. This step, which is indicated by the reaction

14/ Bowen, N. L., Progressive metamorphism of siliceous limestone and dolomite: Jour. Geol., vol. 48, no. 3, p. 245, 1940.

$$\underset{\text{calcite}}{2CaCO_3} + \underset{\text{tremolite}}{3CaMg_3(SiO_3)_4} \rightleftharpoons \underset{\text{diopside}}{5CaMgSi_2O_6} + \underset{\text{forsterite}}{2Mg_2SiO_4} + CO_2$$

is defined by a P-T curve showing the temperature at which the reaction can proceed at a given pressure. At higher temperatures for a given pressure, calcite and tremolite cannot coexist in equilibrium.

Bowen's[15/] schematic representation of this curve suggests, for example,

15/ Bowen, N. L., op. cit., p. 256.

that the reaction could occur at about 470 degrees C. under 200 atmospheres pressure, and at about 600 degrees C. under 2000 atmospheres pressure.

Bowen[16/] has also arranged the following sequence of minerals in

16/ Bowen, N. L., op. cit., p. 260.

the order of their production with rising temperature: (1) tremolite, (2) forsterite, (3) diopside, (4) periclase, (5) wollastonite, (6) monticellite, (7) akermanite, (8) spurrite, (9) merwinite, (10) larnite. He cautions, however, that the presence of any one of these minerals can be used as an indicator of metamorphic temperature only if the original material was a siliceous dolomitic limestone, and only when such a rock was sufficiently immobile to prevent the rapid elimination of carbon dioxide from the system. Bowen[17/] also notes that "the

17/ Bowen, N. L., op. cit., p. 258.

amount of alumina present (or added) might .... be so great as to prevent the formation of some of the reference phases in all stages of metamorphism". He believes, however, that in the contact metamorphism of a siliceous dolomite a high concentration of carbon dioxide is generally maintained, and that the index minerals, as listed, can be used with reasonable certainty.

If, in the hornfelsic rocks the formation of diopside was preceded or accompanied by the formation of tremolite or forsterite, evidence of this was not noted by the writer. Tremolite appears in the first of the thirteen steps which is represented by the reaction

$$\underset{\text{dolomite}}{3CaMg(CO_3)_2} + \underset{\text{quartz}}{4SiO_2} \rightleftharpoons \underset{\text{tremolite}}{CaMg_3(SiO_3)_4} + \underset{\text{calcite}}{CaCO_3} + 4CO_2$$

provided an anhydrous formula for tremolite is used. As indicated by Bowen, the hydrous character of tremolite introduces a complicating factor, because a participating liquid phase must be present both above and below the reaction temperature. "If the rock 'boils dry'[18/] below

18/ The term "dry", as used by Bowen and in the present paper, does not imply that no water is present, but that no liquid water is present.

the reaction, no tremolite is formed; if it 'boils dry' at the reaction point, some tremolite forms"[19/].

19/ Bowen, N. L., op. cit., p. 241.

In the known examples of thermal metamorphism of siliceous dolomites tremolite, as a mineral phase, is commonly absent, and forsterite is ordinarily believed to be the first phase formed. Because forsterite is unstable in the presence of quartz, the forsterite-quartz association is cited as an example of disequilibrium.

The extent of circulating solutions, during the formation of diopside, cannot be demonstrated with certainty. It will be shown in following sections, however, that the mafic and granitic masses, which intimately penetrate the hornfels member, were emplaced in an order of decreasing basicity and that much, if not all, of the metasomatism is related to the granitic rocks. It is possible, therefore, that the highest temperatures were reached in a relatively "dry" environment before the granitic rocks were emplaced, perhaps during the emplacement of the mafic rocks.

If circulating waters were present in negligible amount, or, if abundant, were able to maintain a high concentration of carbon dioxide, during the formation of the diopside, several factors could account for the general absence of tremolite and forsterite. Water may not have been available to enter into the tremolite-forming reaction or an abundance of alumina may have prevented the appearance of tremolite altogether. If tremolite did exist and equilibrium was attained upon the appearance of diopside, the instability of the association calcite-tremolite would have led to the disappearance of tremolite as a mineral phase. The absence of forsterite could be attributed to either an abundance of alumina or to an attainment of equilibrium wherein forsterite was unstable in the presence of quartz.

Wollastonite, like forsterite, is notably absent from the hornfels. The production of wollastonite by the reaction of quartz and calcite, as shown by the equation $\underset{\text{calcite}}{CaCO_3} + \underset{\text{quartz}}{SiO_2} \rightleftharpoons \underset{\text{wollastonite}}{CaSiO_3} + CO_2$, is commonly cited as a reliable indicator of temperature and pressure during metamorphism. The association of quartz and calcite in equilibrium is likewise used to indicate that, at a given pressure, the temperature that would permit the reaction has not been reached. This reaction is the sixth of the thirteen steps mentioned above.

Its P-T curve, as shown by Bowen[20], includes, for example, points at

20/ Bowen, N. L., op. cit., p. 256.

about 650 degrees C. and 200 atmospheres pressure, and at about 780 degrees C. and 2000 atmospheres pressure. But caution should be used in citing of even these temperatures as the highest at which calcite and quartz can coexist at the indicated pressures; and the possibility that the highest temperatures in the metamorphism of the rocks of the Silver Lake area were attained under conditions of high temperature and shearing stress should not be overlooked. Such conditions are ascribed to the metamorphism of rocks of the staurolite-kyanite facies as described by Turner[21]. In this facies wollastonite is believed to be unstable.

21/ Turner, F. J., op. cit., pp. 86, 102.

If the presence of diopside and the association of calcite and quartz can be taken as reliable indicators of metamorphic temperature and pressure in the development of the hornfels, the maximum temperature would lie in the area between the curves of step 4 and step 6 as shown by Bowen. If, for example, a pressure of 1000 atmospheres is assumed, a temperature in excess of 600 degrees C. would be indicated to assure the formation of diopside; and a temperature of 780 degrees C. probably could not have been greatly exceeded without the appearance of wollastonite.

The talc bodies are composed mostly of minerals produced during a period of retrogressive metamorphism and extensive metasomatism, which will be discussed in more detail in a section to follow. Remnant grains of forsterite partly to wholly replaced by tremolite, serpentine, and talc, were noted at several places along the talc-bearing zone. The forsterite grains appear to be the only representatives of the assemblage that antedated the three later minerals. Of the three, tremolite is by far the most abundant. The association of tremolite and forsterite, known only to occur in metamorphosed carbonate rocks, indicates the original sediment to have been carbonate-rich. In contrast with the enclosing wall-rocks of hornfels, the talc bodies contain only one or two percent each of alumina, potash, and soda. That the original sedimentary rock was comparably poor in these constituents, is probably a valid assumption.

If equilibrium can be assumed at the time of the production of the forsterite, a deficiency of silica would thereby be indicated. That the original rock was silica-deficient is suggested by virtually complete absence of quartz in the talc bodies, except in granitic veinlets. It should be recalled, however, that in other metamorphic terranes forsterite and quartz are commonly in close association, howbeit in disequilibrium, and that an original silica deficiency cannot be conclusively demonstrated for the talc bodies.

The lenses of various schistose rock types in the hornfels member generally appear to represent zones of stress that parallel the planar features of the member. The phlogopite and tremolite, so common in the schistose layers, are probably both younger than the diopside in the enclosing rocks. Although diopside is notably absent in the larger schist bodies, it is in association with amphibole (tremolite or actinolite) along the schist borders and along contacts between hornfels and granitic dikelets. Wherever age relations between diopside and the amphibole were observed, the amphibole was consistently later[22/]. That phlogopite commonly occurs in tremolite rock as veinlets and as borders of late-stage granitic dikelets, shows that much, if not all, of the phlogopite in the mine area was late-forming.

22/ A diopsidic border zone along a granite veinlet in tremolite rock provided the only observed exception to this statement.

Quartz-biotite schist member

Distribution

A quartz-rich, commonly schistose metasedimentary unit containing biotite as a distinctive mineral, persists for the full length of the talc-bearing zone. This quartz-biotite schist member (pl. 7a) overlies the hornfels member; but the two are ordinarily separated by bodies of granitic rock or lamprophyre. The quartz-biotite schist member's full thickness is about 150 feet. Invasion of granitic rock has caused the member to be considerably thinner for much of its exposed length, but has nowhere visibly displaced or distorted it.

Other biotite-rich metasedimentary units exist below and above the part of the section occupied by the five members considered here; but, of the five, only the quartz-biotite schist contains biotite as a characteristic mineral.

Petrology

Though easily recognized by its biotite content, the member has a markedly variable mineralogy. Its texture ranges from massive granular to schistose, and from fine- to medium-grained. The more massive varieties, represented by exposures south of the Gould workings are largely quartzites with very subordinate amounts of biotite, muscovite, and feldspar. A crude planar structure in these rocks is produced by layers of contrasting grain size and by layers with a somewhat higher than average mica content.

The quartzitic varieties, which are typically light gray, grade into dark gray, distinctly schistose rocks containing as much as 30 percent biotite and appreciable amounts of muscovite and feldspar. Biotitic layers characteristically alternate with felsic lenticles or bands one-eight inch or less thick. The rocks of this phase commonly grade into migmatites which, in turn, pass gradationally into poorly foliated granitic rocks. Both the migmatites and granitic rocks contain dikelets of leucogranite and aplite that lie across the planar structures.

Migmatitization is particularly well shown in the area south of the two Addenda workings where the hornfels and biotite schist members are separated by a 200-foot belt of predominantly tonalitic rock. Here, as elsewhere, the contacts between hornfels and granitic rocks are sharp. The biotite schist, however, is gradationally separated from the granitic belt by migmatite. In a zone from 50 to 100 feet wide, there is a gradation from schist-free tonalite with a crude planar structure, through tonalite with many schistose layers and highly migmatized schist, to schist relatively free of migmatite. Similar gradations are common elsewhere. Virtually all of the truly migmatitic rock in the mine area is, in this manner, associated with the more micaceous phases of quartz-biotite schist. The quartzitic phases, as well as the area's other non-schistose rocks are generally free of migmatite, but contain granitic material in relatively large sill-like masses.

Petrography

In thin section, a specimen of a biotitic phase of the quartz-biotite schist member is shown to contain about one-half quartz, one-fourth microcline, one-fourth mica, and minor amounts of albite, apatite, and opaque grains. The quartz and microcline form a mosaic with grains as much as 5 mm. in diameter. Most of the microcline, however, is concentrated in what seems to be a migmatitic layer. The quartz grains show undulatory extinction and a tendency toward elongation parallel with the schistosity.

About three-fifths of the mica is biotite; the remainder is muscovite and sericite. Most of the mica shreds are in general alignment and are clustered in discontinuous parallel layers. Biotite grains are as much as 3 mm. in length; the muscovite grains are ordinarily smaller. Some of the muscovite is in large grains, but most is in sericitic aggregates that have partly to wholly replaced biotite. The microcline is slightly sericitized.

Metamorphism

The unmigmatitized parts of the biotite schist member, consisting predominantly of dimensionally oriented, elongate quartz grains and biotite shreds, seem best attributed to the metamorphism of an impure, carbonate-free quartzite under conditions of high directional pressure. That biotite, rather than muscovite, is the principal mica, suggests iron- and magnesia-rich impurities indigenous to the original sediment in the form of chloritic or perhaps mafic tuffaceous material. The abundance of biotite in and near the migmatite zone in the west part of the mine area could, perhaps, be cited as a basic "front" in which iron and magnesia had been introduced in advance of a "granitization wave"[23/].

23/ See, for example, Reynolds, D. L., The association of basic "fronts" with granitization: Sci. Prog., vol. 35, pp. 205-219, 1947.

Several factors, however, mitigate against the presence of biotite-rich basic "fronts" in the mine area. The distribution of biotite in the metasediments appears to be largely, if not entirely, a stratigraphic feature. Well defined stratigraphic planes separate the quartz-biotite schist member from the underlying biotite-poor hornfels member and the overlying biotite-free quartz-muscovite schist members. None of the metasedimentary rocks that normally contain little or no biotite show the development of the mineral in the vicinity of granitic contacts. There is widespread evidence of iron and magnesia metasomatism in the hornfels member; but the two generally have been separated in time (i.e. dolomite (?) replaced by tremolite rock which is in turn altered to actinolite along borders of granitic dikelets) and did not produce biotite. Instead, in the mine area, biotite schists appear to have been most susceptible to migmatization. A degree of impoverishment in iron and magnesia and enrichment in potash is indicated by a progressive decrease in biotite with increasing intensity of migmatitization, a trend shown microscopically by the partial sericitization of the biotite.

The reconstitution of initially solid rocks to migmatites commonly has been described as occurring in the presence of silicate melts. Eskola[24], for example, limits the term "migmatite" to rocks of this

24/ Eskola, P., et. al., Die Entstehung der Gesteine. Springer, Berlin, 1939.

origin. Such melts have been indicated variously as true magmas, as highly fluid magmas, and as liquids produced by differential fusion. The development of rocks with migmatitic textures has also been ascribed to metasomatic replacement involving an aqueous pore solution, and to diffusion in the solid state. Few geologists insist, however, that all migmatitic rocks must develop by any one of these processes. Turner[25] has stated that the presence of a silicate melt during the

25/ Turner, F. J., op. cit., p. 305.

development of a migmatite is indicated by a "general abundance of pegmatitic, aplitic and other igneous veins, lenses, and streaks"; whereas quartzose veins prevail in strictly metamorphic rocks. If this criterion is valid, the migmatite zones of the mine area were of the silicate melt variety.

Time limitations did not permit a detailed petrographic study of the migmatites of the Silver Lake mine area. Megascopic observations and detailed mapping, however, show that the enclosing schists have not been displaced or shouldered aside in the development of migmatite.

Quartz-muscovite schist member

Distribution

The quartz-muscovite schist member is pale orange to dark yellowish orange in color and consequently contrasts with the underlying gray quartz-biotite schist. In the area mapped, the quartz-muscovite schist ranges from 35 to 125 feet in thickness; but farther east, in the vicinity of the Number Four workings, the member is much thicker.

Petrology

The member is composed mostly of very even layers, alternately micaceous and quartzitic, and ranging in thickness from a fraction of an inch to several feet. The schistose rock commonly contains scattered iron oxide grains producing a pepper-sprinkled appearance.

Locally interbedded with the schist and quartzite are elongate lenses of marble similar in appearance to the forsterite marble of the overlying member. The quartz-muscovite schist member also contains a small amount of quartzose amphibolite in thin layers. This member is virtually free of migmatite, but contains sill-like bodies of granitic rocks.

Petrography

A thin section of a specimen, typical of the more schistose facies of the member, is composed of approximately three-fourths quartz and one-fourth muscovite. It also contains one or two percent of ferric oxide and a minor amount of sphene and rutile.

Most of the quartz grains have lengths from 1 1/2 to 3 times their widths and are dimensionally oriented parallel with the schistosity. The grains are generally less than 0.3 mm. in long dimension; only a few exceed 0.5 mm. They show no undulatory extinction. The rutile, which occurs as numerous needle-like inclusions in the quartz, is also oriented parallel with the schistosity.

Some of the muscovite occurs in shreds from 0.1 to 1 mm. in length, but most of it is in aggregates of much smaller sericite shreds. The aggregates are peripheral to the quartz grains, and form thin layers that are the principal cause of the schistosity. The ferric oxide grains are dark red and opaque to translucent.

Microscopic examination of a specimen from one of the marble lenses shows that it is an ophicalcite composed of about three-fourths carbonate, one fourth chrysotile, and a few percent of antigorite, talc and opaque material. The carbonate, mostly if not entirely dolomite, is in grains that average about 0.2 mm. in diameter. These are locally stained with iron oxide, but are generally unclouded. The serpentine is in evenly disseminated grains that average less than 0.1 mm. in maximum dimension. Most are nearly equant grains of chrysotile, probably pseudomorphic after forsterite, but some are antigorite shreds, partly to wholly replaced by talc. The opaque material occurs mostly as minute inclusions in the equant serpentine grains. The presence of both forsterite and clinohumite elsewhere in the marble lenses was noted in the inspection of mineral fragments in immersion media.

Thin section studies show that the amphibolite schist layers are composed almost entirely of elongate grains of quartz and ferruginous amphibole of approximately equal proportion. Both are dimensionally parallel and range mostly from 0.5 to 1 mm. in long dimension. The amphibole (actinolite or hornblende) is strongly pleochroic in shades of pale yellow to medium green. Very minor amounts of sphene and opaque material were the only accessories noted.

Metamorphism

Rocks composed essentially of quartz, subordinate muscovite, and minor amounts of ferric oxide, rutile, and sphene, appear originally to have been impure sandstone notably lacking in magnesian and carbonate material. Although the proportion of muscovite varies from layer to layer, its very even distribution vertically and laterally within individual layers several inches to several feet thick suggests that much, if not all of the muscovite-forming material was present in the original sediment, as sericite, argillaceous material, or a mixture of the two, and that this gross layering is a stratigraphic feature. As in the other rocks that also show minute layering, the thin laminations are probably attributable to metamorphic differentiation. In a less metamorphosed terrane these smaller laminations could easily be interpreted as bedding planes, but the survival of such strata is improbable in the rocks as highly metamorphosed as those of the Silver Lake mine area.

The quartzose amphibolite layers are apparently metamorphosed impure sandstone beds, but it is unlikely that impurities were present in proportions from which amphibole (actinolite or hornblende) alone was formed without the addition or subtraction of material. Perhaps CO_2 was simply removed from an original carbonate fraction, or MgO, SiO_2, or iron oxide may have been added as they were in other rocks of the area. The textures of the rock record little of its metamorphic history except formation under conditions of high pressure and stress producing thin laminations. Because the amphiboles in the other members post date such minerals as diopside and forsterite, the amphibole of the quartzose amphibolite may have formed relatively late.

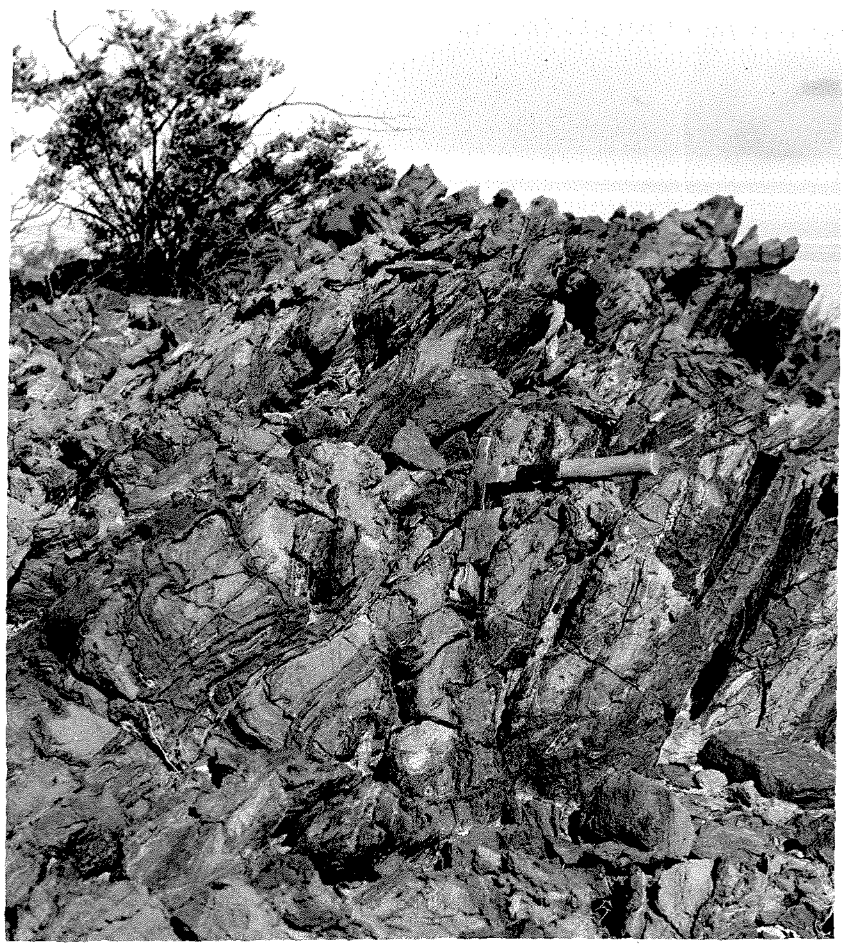
Forsterite marble member

Distribution

The forsterite marble member (pl. 7b), a crystalline unit rich in disseminated magnesian silicates (i.e. forsterite, clinohumite, chrysotile, antigorite, and talc), and containing numerous diopside-chrysotile-antigorite-calcite veins, persists from the area south of the Gould talc bodies westerly for about 4000 feet to the alluvial overlap. In this area, the member ranges from 45 to 65 feet thick. South of the eastern part of the Gould workings it terminates against a granitic mass, but east of the mass a series of thinner marble lenses apparently

Plate 7a. Quartz-biotite schist member (right) containing granitic layers and in contact with crudely foliated tonalite (left). Exposure near Western Addenda shaft.

Plate 7b. Forsterite marble member containing diopside- and serpentine-bearing veinlets (dark).



represents an extension of the unit. The forsterite marble member is the thickest and uppermost of a group of lithologically similar marble layers, smaller bodies of which lie in the quartz-muscovite schist member below.

Petrology

On fresh surfaces the marble is generally light-olive gray, but a dark gray or dark, greenish gray color is not uncommon. It is a medium-grained, dense rock, characteristically massive, but showing a crude planar structure. It weathers to a hackly surface, mostly pale yellowish brown, but showing the veinlets in prominent relief as dark brown-weathering bodies (pl. 7b). These veinlets, mostly one-eighth to one-half inch thick, are much more abundant in the member itself than in the marble of the underlying lenses. Although many of the veinlets parallel the general attitude of the member, others criss-cross it, seemingly at random. The disseminated grains also protrude from the weathered surface, and are commonly clustered in planes parallel to the general attitude of the member.

Petrography

Thin sections of several typical unveined specimens of the marble member contain approximately three-fourths crystalline carbonate and one-fourth disseminated magnesian silicate grains. The carbonate grains, mostly dolomite, average about 0.1 mm. in diameter and are clouded by very fine-grained semi-opaque particles.

Of the disseminated magnesian silicates, forsterite and clinohumite are by far the most common (fig. 3). The relative proportion of

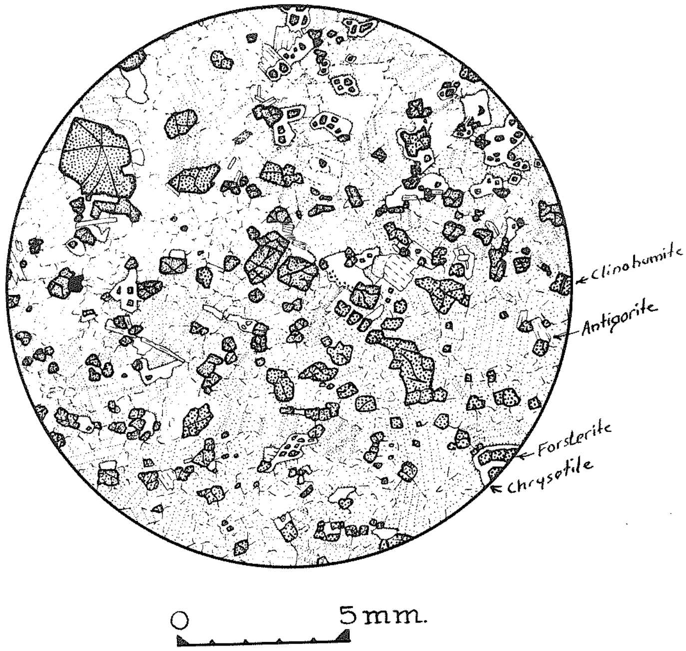


Figure 3. Marble containing forsterite (partly to wholly replaced by chrysotile), clinohumite (unaltered), and antigorite. From forsterite marble member.

the two minerals differs from section to section, but on the average, appear to be equally abundant in the marble of the member itself. In the lenses below, forsterite (or chrysotile altered from it) apparently predominates.

The grains of both the forsterite and clinohumite are equant and sub-rounded. Both have maximum diameters of about 2 mm., but most grains are less than 0.5 mm. in diameter, and distinctly smaller than the forsterite grains in the talc-tremolite bodies. Both are similar in appearance and in optical properties; but the clinohumite is ordinarily distinquishable by a straw yellow pleochroism. The forsterite is best recognized by a partial to complete alteration to chrysotile. The clinohumite, by contrast, is unaltered.

Scattered shreds of antigorite mostly 1 mm. or less in length, compose 2 or 3 percent of each thin section. These have been partly to wholly altered to talc. Minute grains of a steel gray opaque mineral, probably magnetite, are common inclusions in the chrysotile derived from forsterite. Green spinel grains are sparsely scattered through several of the thin sections.

An inspection of numerous powdered specimens of the veinlet material showed that chrysotile is the most abundant mineral in these bodies, but that diopside locally predominates. Where chrysotile and diopside are associated, serpentine is the later of the two minerals. The diameters of the diopside grains average less than 0.3 mm., but some are as much as 3 mm. Antigorite shreds are interstitial to the diopside; both are transected by numerous chrysotile veinlets and by less numerous calcite veinlets. Opaque grains are commonly included in the antigorite shreds and are particularly abundant in the chrysotile veinlets. Extended petrographic studies may well show that the serpentine in all of the veinlets has similarly formed at the expense of diopside and antigorite.

Metamorphism

The silica deficient assemblages that characterize the forsterite marble member and the marble lenses in the underlying schist are typical metamorphic derivatives of low-silica magnesian limestones and dolomites; but the paragenesis is somewhat obscured by the effects of circulating solutions, the full extent of which cannot be completely demonstrated. In the absence of contradictory evidence, it is assumed that the evenly disseminated silicate grains, in large measure, reflect the amount and character of the impurities in the original rock, and that a large part of the vein-forming material has been introduced. That hydrous solutions also penetrated the intervein blocks of the member and the unveined lenses beneath is shown by the widespread serpentinization of forsterite and alteration of antigorite to talc. The introduction of fluorine is indicated by the abundant clinohumite.

That the combined amount of forsterite and clinohumite remains nearly constant throughout these rocks, regardless of the proportions of the two minerals, suggests that the magnesia and silica of both were derived largely from the original sediments. The abundance of clinohumite would thereby seem to be a measure of the availability of fluorine. Because the rock contains no free silica, it can be reasonably assumed that the silica, in the original sediments was consumed in the development of the magnesian silicates. The presence of very thinly disseminated spinel, the only aluminous constituent of the marbles, probably points to the original presence of a correspondingly low proportion of argillaceous material.

The metamorphic significance of the disseminated antigorite shreds is not clear. If formed at the expense of a pre-existing magnesian silicate, they contain no remnants of it, nor is it demonstrably pseudomorphic. Bowen and Tuttle[26/] have cited laboratory evidence that

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26/ Bowen, N. L., and Tuttle, O. F., The system $MgO-SiO_2-H_2O$: Geol. Soc. America Bull., vol. 60, pp. 439-460, 1949.

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pure magnesian serpentine cannot exist at temperatures above 500 degrees C. regardless of pressure. Therefore, the antigorite and chrysotile probably formed at temperatures well below maximum for the metamorphism. It will be remembered that the mineral associations of the hornfelsic rocks strongly suggest maximum temperatures well in excess of 500 degrees C. Higher temperatures are also suggested by the position of chrysotile in the paragenesis of talc-tremolite rock.

The serpentinization of forsterite by merely the addition of H_2O also yields brucite by the equation,

$$\underset{\text{forsterite}}{2Mg_2SiO_4} + 3H_2O \rightleftharpoons \underset{\text{brucite}}{Mg(OH)_2} + \underset{\text{serpentine}}{H_4Mg_3Si_2O_9}$$

a reaction that Bowen and Tuttle[27/] have shown to proceed only at

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27/ Bowen, N. L., and Tuttle, O. F., op. cit., p. 542

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temperatures less than 400 degrees C. The serpentinization of forsterite in the marbles of the mine area, however, was apparently unaccompanied by brucite. If, as many geologists believe, such serpentinization involves no change in volume, silica and water were added to the forsterite while magnesia was removed.

The origin of the dolomite, abundant in the carbonate fraction of the rock, is obscure. A survival of dolomite from the original sediment would indicate that the fifth of Bowen's thirteen steps had not been reached. This step marks the decomposition of dolomite, which , in the absence of available silica, introduces periclase as a new mineral phase. The forsterite-calcite-dolomite association is thereby replaced by the forsterite-calcite-periclase association according to the reaction

$$\underset{\text{dolomite}}{CaMg(CO_3)_2} + \underset{\text{calcite}}{CaCO_3} \rightleftharpoons \underset{\text{periclase}}{MgO} + CO_2 .$$

Although periclase does not exist in the rock it may have been completely replaced by some of the serpentine.

If the magnesium in the disseminated silicate minerals is indigenous to the original sedimentary rock, the minerals are abundant enough to contain most, if not all, the magnesium of the original dolomite. The rock's present high proportion of dolomite, therefore, suggests the addition of relatively large amounts of magnesium. Perhaps some of the magnesium was derived from the serpentinization of forsterite; but the abundant veinlets of serpentine and diopside point toward a more distant source for much of the hydrothermally transported magnesia.

The planes along which the silicate grains commonly cluster resemble relict bedding in that they are approximately parallel to planes that are clearly of sedimentary origin. In this rock, as in the others that contain small-scale planar structures, such small strata could hardly have survived the stress and high pressure for which there is such abundant evidence in rocks of the mine area.

Quartzite member

Distribution and petrology

The uppermost of the five members is a vitreous quartzite that in all of its exposures lies between the forsterite marble member and the granitic rock. The quartzite is medium light gray, medium- to coarse-grained, dense, and vitreous. The member also contains very subordinate thinly-layered quartzose amphibolite similar to the amphibolite in the quartz-muscovite schist member. The quartzite was not traced east of the area south of the eastern part of the Gould workings where it abuts against a granitic mass. The exposed thickness of the member ranges from 2 to 25 feet. Because it is nowhere overlain by metasediments, its original thickness may not be indicated in these exposures.

Petrography

A thin section of a specimen typical of the quartzite member contains about 90 percent quartz, 10 percent feldspar, and 2 to 3 percent mica. The quartz grains, although irregular in outline and having sutured borders, are markedly elongate parallel with the overall

Name	Maximum thickness in feet	Description
Upper units	400 +	Complex of metasedimentary bodies, principally quartz-biotite schist; diopside-feldspar hornfels and quartzite less abundant; interlayered with tonalite and granitic gneiss.
Quartzite member	25	Quartzite, light gray, medium-grained, massive, compact, vitreous; contains thin layers of amphibolite.
Marble member	60	Marble, dolomitic, olive gray, medium-grained. Generally shows crude planar structure. Contains abundant disseminated silicate grains (forsterite, clinohumite, chrysotile, antigorite, talc) and small silicate veinlets (diopside, chrysotile, antigorite).
Quartz-muscovite schist member	125	Schist and quartzite composed mostly of quartz and muscovite; orange, fine- to medium-grained. Contains layers of marble and amphibolite.
Quartz-biotite schist member	185	Schist and quartzite composed mostly of quartz and biotite; gray, medium-grained; commonly contains migmatite.
Hornfels member	155	Hornfels composed mostly of feldspar, diopside, and quartz; green, fine- to coarse-grained; thin laminations characteristic; also contains layers of mica schist, tremolite schist, massive tremolite rock, talc schist, quartzite, and marble. A common host for pegmatite and lamprophyre.
Lower units	400 +	Complex of various metasedimentary rocks (schists, hornfels, quartzite) interlayered with tonalite, granite gneiss, and silexite.

Table 1. General features of metasedimentary rock sequence in western part of Silver Lake mine area.

attitude of the member. The lengths of such grains commonly exceed 5 mm. and are ordinarily from 2 to 5 times their widths.

The feldspar and mica grains are much smaller; these occur in the quartz as poikilitic inclusions and are also elongate parallel with the general planar structure of the rock. The lengths of most of the feldspar and mica grains are less than 0.2 mm.; rarely do they exceed 0.5 mm. Although much of the feldspar is highly sericitized, all appears to be alkalic. Many of the less altered feldspar grains are albite. The mica is predominantly biotite, but the section contains a few scattered grains of muscovite. Minute rutile needles, oriented in seemingly random directions, are common in the quartz-grains. Other relatively abundant accessories are sphene, apatite and opaque grains.

Metamorphism

The quartzite apparently is a metamorphosed siliceous sandstone, with very subordinate aluminous material, recrystallized under conditions of high directional pressure, and partly sericitized.

LAMPROPHYRE[28/]

28/ The term "lamprophyre" is used here in its broad compositional and textural meaning and does not imply a diaschistic origin.

Distribution

Lamprophyric rock, mostly hornblende kersantite but gradational through diabase into diorite, is widespread in the Silver Lake mine area, but is much less extensive than the granitic rocks. Most of the lamprophye bodies are dikes or irregular pods less than 50 feet in maximum dimension; but a few are relatively thick, tabular bodies several hundred feet long. In general, both the abundance and size of the lamprophyre bodies increase from west to east within the area mapped. The bodies are particularly numerous within or adjacent to the hornfels member; but, unlike the granitic bodies in the hornfels member, the lamprophyre has not visibly altered the bordering metasediments.

Petrology and petrography

All of the lamprophyre bodies are persistently composed of about two-thirds andesine and one-third mafic minerals. The mafic fraction of the smaller bodies and large parts of the larger bodies consists of biotite and hornblende in nearly equal amounts, to form the hornblende kersantite facies. The larger bodies contain all gradations between this rock and biotite-poor diorite. All facies of the lamprophyre are gray in color. The dioritic facies ordinarily has an ophitic texture. The hornblende kersantite facies is locally ophitic but generally shows a strong secondary schistosity.

The largest lamprophyre body in the mine area lies south of the Number Two and Number Two and One-half workings and separates the hornfels and quartz-biotite schist members. It is approximately 1000 feet in length and 300 feet in maximum width. The bordering metasedimentary units conform in outcrop pattern with the margins of this body, and appear to have been shouldered aside during its emplacement. The lamprophyre of this mass locally grades into diorite and is also cut by dike-like bodies of tonalite, aplite, and pegmatite, each of which has sharp contacts.

A smaller lamprophyre body, approximately 300 feet long and 50 feet wide, lies along the hornfels and quartz-biotite schist contact between the Gould and Number Two and One-half workings and also shows a transition into tonalite. It is most mafic near the hornfels contact, and becomes successively more felsic from north to south toward the quartz-biotite schist contact. The transition is mostly gradational, but it is also marked by a north to south mafic to felsic change in the composition of dikelets that intimately penetrate the rock. Most of the mineralogic and textural variants of the lamprophyre in the mine area, however, grade into one another almost imperceptibly, and are probably closely related in time.

Tonalitic, aplitic, and pegmatitic dikes and dikelets that crosscut the lamprophyre show that it is older than the granitic rocks of the area; but the local gradation of lamprophyre into tonalite suggests that the time of formation of the less felsic phase of the granitic sequence closely followed the emplacement of lamprophyre. Such a gradation, however, may represent a replacement of lamprophyre by tonalite rather than closely related magmatic phases.

An examination of several thin sections of the hornblende kersantite showed only a moderate range in mineralogy. The smaller bodies consistently contain about 70 percent andesine, 15 percent hornblende, and 15 percent biotite. A specimen typical of the medium-grained dioritic parts of the larger bodies is somewhat more mafic in that it contains about 60 percent andesine, 35 percent hornblende, and 5 percent biotite. Sphene and opaque grains are very abundant accessories; apatite is also common.

The rock is holocrystalline and the grains of its principal minerals are generally of comparable size. The minerals of the smaller bodies are mostly in grains that range from 0.2 to 1 mm. long. Average grain sizes of more than 2 mm. are common in the larger bodies. Some of the dioritic rock contains hornblende blades as much as one inch long. Much of the feldspar is partly to thoroughly sericitized. Otherwise, the rock shows very little alteration.

GRANITIC ROCKS

General features

The granitic rocks of the Silver Lake mine area range in composition from tonalitic to silexitic, and in texture from granular to gneissic and schistose. Three rock types comprise most of the granitic material: (1) a granular to poorly-foliated rock, designated as tonalite, but gradational into true granite; (2) a microcline-quartz-mica gneiss; and (3) a microcline silexite. These three persist as relatively well-defined units throughout the area, but in many places the tonalite and microcline-quartz-mica gneiss are too intricately associated to be mapped separately on the scale of the accompanying map (pl. 1). Therefore, the two appear as a single unit.

Dikes and dikelets of pegmatite, aplite, and granite, though subordinate in volume, are widespread. Each is composed principally of microcline and quartz.

Tonalite

Distribution

Under the name "tonalite" is included a structurally homogeneous to moderately schistose rock (pl. 8a) that composes more than half of the granitic material in the mine area. In most of its occurrences this unit is tonalitic, but the granite into which it grades is relatively abundant.

Tonalite together with subordinate amounts of microcline-quartz gneiss, comprises most of the granitic mass that surrounds the large metasedimentary "islands" in the vicinity of the two Addenda workings. Nearly all of the granitic rock south of the Gould, Number Two and One-half, and Number Two workings is likewise tonalite. In this area the rock commonly occurs in sill-like bodies, as much as 50 feet thick and 1100 feet long. The bodies are particularly large and numerous in the quartz-biotite schist member; but several are also in the quartz-muscovite schist member.

In the area south and east of the eastern part of the Gould workings, tonalite occurs in large, irregular masses that contain very elongate metasedimentary inclusions consisting mostly of quartz-biotite schist. Some of the inclusions are several hundred feet in length. The inclusions are ordinarily parallel to each other and have the same attitudes as larger nearby metasedimentary masses.

A schistose granite phase of the tonalite unit exists in two elongate masses, each more than 300 feet long. One is north of the Eastern Addenda workings; the other is north of the Gould workings.

The Eastern Addenda mass is in sharp contact with structurally homogeneous tonalite and appears to be surrounded by it. Much of the other mass is hidden by alluvium and talus; in some exposures, it is bordered by tonalite, and in others by microcline silexite.

As noted above, much or all of the tonalite unit postdates the lamprophyre. Its age relation to the microcline-quartz-mica gneiss and to the microcline silexite is less clear. In many places a narrow zone of tonalite lies between the gneiss or the silexite and metasediments. In the western part of the area, such a zone separates metasediments of the quartzite, forsterite marble, and quartz-muscovite schist members, from microcline-quartz-mica gneiss to the south. This zone is 8 to 50 feet thick and more than 1500 feet long. It is in sharp contact both with gneiss and metasediments; within a few feet of the gneiss it contains numerous large microcline crystals. A zone of the tonalite unit discontinuously separates microcline silexite from the lower border of the hornfels member throughout the eastern part of the area. Here, too, contacts are sharp, but the large microcline crystals are absent.

The pattern of these zones at first suggests an intrusive origin; but the abundance of the microcline crystals near the contact with gneiss, and absence of crosscutting tonalite bodies in gneiss or the silexite do not support the intrusive concept. Indeed, the zones appear to be a less siliceous and less potassic contact phase of the microcline-rich units. At a locality just east of the Gould workings, a dike-like septum of silexite extends into tonalite and apparently postdates it. The tonalite, therefore, appears to be part earlier, and in part contemporaneous with the microcline-quartz-mica gneiss and the microcline silexite. Conversely, much or all of the tonalite is later than the lamprophyre.

Petrology and petrography

The average tonalite is a light to medium gray, medium-grained rock composed of about one-half plagioclase feldspar, one-fourth quartz, and one-fourth potash feldspar, muscovite and biotite. In some places it also contains scattered euhedral microcline phenocrysts as much as two inches in length. In many of its occurrences, it appears to be structurally homogeneous; in others, it has a crude planar structure that approximately parallels the attitude of the nearest metasedimentary masses.

Thin sections of specimens of the tonalite unit gathered at widely spaced localities show that, in spite of the differences in the ratios of potash feldspar to plagioclase, the textural and structural features of the unit are persistent. Most of the plagioclase and quartz are in grains that range from 1 to 5 mm. in maximum dimension. The plagioclase ranges from calcic oligoclase in the granite facies to andesine in the typical tonalite. The potash feldspar is predominantly microcline, the grains of which are commonly much larger (as much as 5 mm. long) than those of plagioclase and quartz. Orthoclase is less abundant and finer grained. The felsic minerals form a typical granitic mosaic, that in some places contains the larger microcline crystals. Microperthitic grains and micrographic intergrowths of quartz in orthoclase are common.

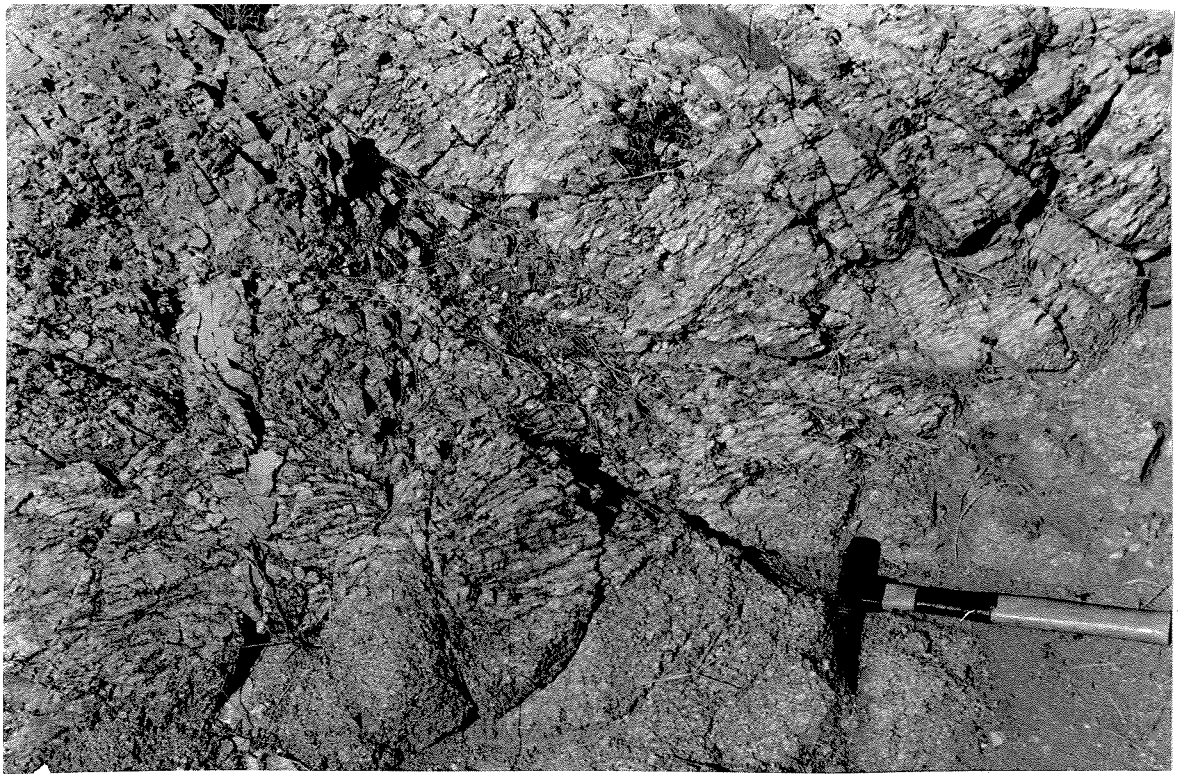


Plate 8a. Contact between tonalite (below) and microcline-quartz-mica gneiss (above).

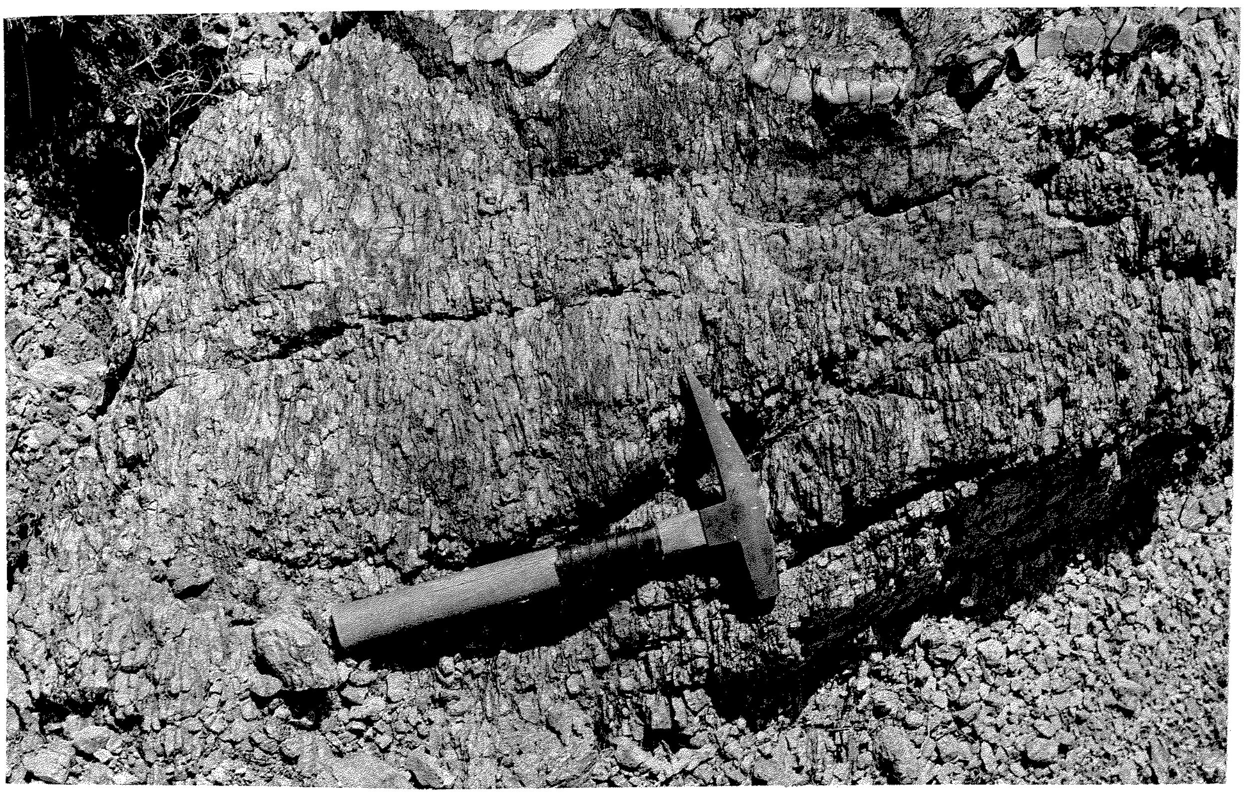


Plate 8b. Detail of microcline-quartz-mica gneiss. Note cross-cutting aplite dikelet at top.

In general, the specimens gathered in the vicinity of large bodies of lamprophyre and in the sill-like bodies within the metasediments, are the richest in plagioclase. The specimens gathered near masses of microcline-quartz-mica gneiss are the richest in potash feldspar.

Mica, consisting of biotite and very subordinate muscovite, ordinarily forms from 3 to 10 percent of the rock. The mica is mostly in shreds from 0.5 to 2 mm. long, and is corroded and transected by quartz and feldspar. Opaque accessory grains are commonly associated with the biotite. In some sections, apatite is also a common accessory. Zircon is present, but less abundant.

Some of the feldspar has been partly sericitized, but the rock as a whole is relatively unaltered. Minute fractures are common, however, and the quartz grains show undulatory extinction.

Microcline-quartz-mica gneiss

Distribution

Microcline-quartz-mica gneiss is a distinctive and prominent unit in the area of the Addenda workings. It commonly occurs in small masses intimately associated with tonalite; but the gneiss composes nearly all of the large granitic mass exposed south of the quartzite member in the area of the two Addenda workings. It is separated from the metasedimentary rocks that border it on the north by the thin zone of tonalite.

In the granitic rock mass exposed north of the quartz-biotite schist member in the Addenda area, the gneiss is subordinate to slightly schistose tonalite. In this area, relatively small bodies of the gneiss are interlayered with tonalite. The two units are generally separated by sharp contacts, but neither rock appears to penetrate the other. East of the Addenda area, very little gneiss was observed.

Petrology and petrography

The gneiss is a light gray, medium-grained rock composed of about one-half potash feldspar, one-third quartz, subordinate albite, biotite and muscovite, and a trace of opaque grains. The rock locally contains microcline crystals as much as 4 mm. long. The foliation (pl. 8a,b) is caused principally by thin layers, rich in aligned mica flakes, alternating with thicker, mica-poor layers composed of a feldspar-quartz mosaic. The micaceous layers are ordinarily spaced about one centimeter apart. Many of the gneiss exposures show a pronounced lineation produced by small crenulations in individual folia and by broad crenulations in folia groups.

In thin section the quartz-microcline mosaic is shown to be composed of irregular grains mostly in the 1 to 4 mm. range in long dimension. The albite and orthoclase grains are much smaller.

The microcline is slightly perthitic. It is also poikilitic and contains numerous rounded grains of quartz and orthoclase. Micrographic intergrowths of orthoclase and quartz are common. Mica shreds are as much as 3 mm. in length. They commonly occur as aligned residua, surrounded and corroded by feldspar and quartz.

Microcline is the least altered of the feldspars, although it is somewhat clouded with minute kaolinite (?) and sericite grains. Many of the albite and orthoclase grains are strongly sericitized. Alteration of biotite to chlorite is also common.

Microcline silexite

Distribution

Microcline silexite, an extensive and widespread unit east of the Eastern Addenda workings, is the most resistant rock in the mine area. The largest silexite body supports the hills along whose lower southern slopes much of the talc-bearing zone is exposed. This body, from 70 feet to more than 400 feet in outcrop width, was mapped for a strike-distance of 3000 feet, and may well continue much farther to the east. Many smaller silexite bodies are exposed in the area that lies south of the talc zone and between the Gould and Number Two workings.

Most of the silexite bodies are elongate lenses parallel with planar elements of the metasediments and of the other granitic rocks; but a few are very irregular bodies. Some are enclosed by tonalite or metasedimentary rocks, but most lie between units of these two. The large body is separated from hornfels on the south by thin, discontinuous tonalite lenses.

Although most of the silexite contacts are relatively sharp, gradational contacts are not uncommon. In several places, for example, the large body grades laterally into quartz-mica schist through several feet. At one locality the writer noted a several-inch gradation of silexite into limestone. Irregular septa of silexite locally extend into tonalite, but apparently none of the silexite occurs in simple, fracture-filling dikes.

Petrology and petrography

The silexite, a white to light gray, medium-grained, equigranular rock, is composed of about 70 percent quartz, 25 percent microcline, and a few percent of muscovite and biotite. Apatite and opaque grains are abundant accessories. Also present are traces of rutile, albite, and sphene. The rock is sufficiently feldspathic to be locally a normal granite. In a few localities the silexite is mica-rich and contains schistose masses, but these are uncommon. In most of its occurrences, the silexite is remarkably uniform, both in texture and composition. The rock is vitreous, compact, and tough. Weathered surfaces show quartz cleavages, so well developed that, upon casual inspection, they give the rock a highly feldspathic appearance. The rock generally has a faint planar structure marked by an alignment of mica flakes, and by a parallel elongation of quartz and microcline grains.

Thin section studies show that most of the quartz and microcline is in grains from 1 mm. to 3 mm. in long dimension. The quartz grains show a marked undulatory extinction and are crisscrossed by numerous rows of minute bubbles. The microcline is locally perthitic. The mica is in shreds that are mostly from 0.5 to 2 mm. long and are corroded and transected by quartz and microcline.

Some of the mica is sericitic and occurs in irregular, vein-like aggregates, of which some are elongate parallel with the planar structure, and others are not. These aggregates apparently record a late-stage sericitization that has proceeded mainly along mineral boundaries and fractures. Sericite shreds also occur sparsely scattered in some of the microcline grains. Much of the biotite is chloritized. Otherwise, the rock is comparatively unaltered.

Observed under the microscope, the gradation from silexite to limestone is complete within about 4 inches, and is marked by grains of quartz, plagioclase and sericite in limestone.

Microcline-quartz dike rocks

Dike rocks, composed mostly of quartz and microcline, but with textures ranging from aplitic to pegmatitic, are widespread throughout the area mapped. From dike to dike within this group, there are all gradations in grain size from fine to very coarse. One type or another has been noted in crosscutting relationship with all of the previously described rock units except the microcline silexite. In this unit, finer grained granitic dikes may well have been overlooked, but their apparent absence may indicate a degree of contemporaneity with the silexite.

The most irregular in outline and generally the largest of the granitic dikes are the pegmatites. These have formed as elongate lenses or irregular pods in most of the rocks of the area. The pegmatites are simple in mineralogy and internal structure. Most are merely bodies of coarse granitic rock in which the grains do not exceed five inches in diameter, and which contain no prominent minerals other than quartz and microcline. Such dikes range from a few inches in width (pl. 10) and a foot or two in length to as much as 30 feet wide and 200 feet long. The pegmatite dikes within the talc bodies rarely exceed 3

Plate 9. Microcline-quartz pegmatite dike cross-cutting diopside-feldspar-quartz-calcite hornfels. Note dark contact zone rich in ferruginous amphibole.

feet in width; many consist entirely of graphic granite.

The granite and aplite dikes generally are narrower and of more uniform thickness than the pegmatites. These finer grained bodies also are as long as 200 feet, but most of them are much shorter.

The microcline-quartz dikes within talc-tremolite rock are commonly bordered by concentrations of phlogopite, chlorite, or ferruginous amphibole (actinolite ?). Borders of phlogopite or chlorite occur as well defined schistose layers that lie close against the dike walls. Actinolite (?) accompanies chlorite, and tremolite accompanies phlogopite in many of these layers. Such schistose borders are as much as 1 inch thick, but most are less than one-half inch thick. The chloritic borders are black to dark green, in striking color contrast to the granitic and tremolitic rocks that they separate. Chlorite-bordered graphic granite dikes locally contain elongate blades of chlorite, as much as 3 inches in length, that extend into the dikes from the dike walls. Some dikes in tremolitic rock do not have schistose borders, but in these the bordering tremolite is ordinarily altered to deep green actinolite (?). Many of the dikes in tremolitic rock contain inclusions of tremolite or actinolite (?).

Microcline-quartz dikes in diopsidic hornfels commonly have clusters of actinolite (?) blades along their margins (pl. 9). These blades, mostly much coarser than the grains of the hornfels, are as much as one-half inch in length. The actinolite has formed at the expense of diopside, but also contains poikilitic inclusions of diopside, feldspar and quartz as remnants of the hornfels assemblage.

Dacite porphyry

A dacite porphyry, the youngest of the dike rocks, is apparently unrelated to the granitic sequence described above. It occurs in elongate bodies, from a few inches to about 10 feet in width, that transect both the metasedimentary and granitic masses. At one place near the center of the mapped area, discontinuous bodies of dacite porphyry have been emplaced along a northwest-trending fault. The fault has displaced both metasedimentary and granitic rocks a horizontal distance of about 15 feet. To the east other dacite porphyry bodies have been explaced along fractures parallel with the fault.

The dacite porphyry is a medium-gray rock, with chalky plagioclase phenocrysts and a very fine-grained groundmass. A thin section of a typical specimen of the rock shows a groundmass composed of approximately 45 percent calcic andesine, 15 percent quartz, 25 percent biotite, 10 percent hornblende, and 5 percent opaque grains. Plagioclase phenocrysts form about 10 percent of the section. These appear to be labradorite, but are highly sericitized and difficult to identify. The andesine and quartz of the groundmass are mostly in grains that range from 0.1 to 0.3 mm. in length; the biotite and hornblende grains are generally 2 to 3 times longer. Most of the plagioclase phenocrysts are 1 to 2 mm. in diameter. The grains of the goundmass are dimensionally aligned, causing a pronounced schistosity.

Emplacement of the large granitic masses

Although critical study of the significance of and mode of emplacement of the granitic rocks of the Silver Lake region requires many more data than could be obtained in the relatively small area of the accompanying map (pl. 1), the mapping did record distributional and structural features that may prove significant in future regional studies, and that may well bear on the origin of the talc-tremolite bodies.

These features indicate emplacement of the large granitic masses without appreciable disturbance of the metasedimentary section. Contacts between metasedimentary units remain nearly equidistant throughout the talc-bearing area, regardless of the amount of granitic material that now exists from place to place in the area. For example, at a locality in the Gould area, a 200-foot thickness of layered rock separates the base of the lowest talc body from the base of the quartz-muscovite schist. Of this thickness, about 150 feet is rock of the quartz-biotite schist and hornfels members and 50 feet is granitic rock. At a locality west of the Eastern Addenda shaft, the same stratigraphic horizons are separated by a 210-foot thickness of which only 40 feet is quartz-biotite schist and hornfels; the remainder is granitic rock.

The tonalite and microcline-quartz-mica gneiss contain numerous metasedimentary inclusions which are characteristically elongate and range in long dimension from a few inches to several hundred feet. Within the granitic masses, inclusions of each metasedimentary rock type are distributed in the same order in which they occur in the metasedimentary section. Each type is confined to a band of granitic rock comparable in thickness to the corresponding metasedimentary member. The inclusions are approximately parallel with each other and with the planar structures of the more complete occurrences of the metasedimentary section. The inclusions, therefore, appear as aligned remnants of the section, and the surrounding tonalite and microcline-quartz-mica gneiss seem to occupy space in which sedimentary rocks once existed. The silexite is generally free of such well defined inclusions but its wispy, schistose inclusions and lateral gradation into limestone lenses and large masses of schist, all apparently undisturbed parts of the metasedimentary section, suggest passive emplacement.

Throughout the area, the surface traces of contacts between metasedimentary units are ordinarily straight or gently curved, whereas contacts between metasediments and granitic rocks are very irregular. The metasediments have not been visibly deformed at the granitic contacts. Indeed, the planar structures in the granitic rocks and the layering and schistosity in the metasediments are everywhere essentially parallel.

The distributional features of the granitic rocks could be attributed to (1) stoping or assimilation in the presence of a magma, (2) to replacement of the metasediments by any of the processes to which granitization is ascribed, or (3) to a combination of fluid invasion and replacement. If the granitic rocks have formed largely by replacement, the process has involved the transformation of widely different metasedimentary rock types into apparently homogeneous granitic bodies, or into granitic bodies whose textural and mineralogic variations seem unrelated to preexisting metasedimentary features. In the Addenda area, for example, the quartz-muscovite schist, forsterite marble, and quartzite members are in juxtaposition to tonalite which, upon megascopic examination, is apparently of uniform composition and texture. Furthermore, the transformation of metasedimentary rocks to mica gneiss would have had to proceed largely along nearly knife-edge contacts. Homogeneity in granitic bodies and the sharpness of their contacts, have commonly been cited as evidence for a magmatic origin.

The well-defined, dike-like bodies of tonalite that cut the larger lamprophyre bodies show no evidence of a replacement origin. Instead their uniform thickness, generally straight traces, and very large length-to-width ratios, strongly suggest that the bodies have been injected along fractures. The numerous aplite and pegmatite dikelets and dikes appear to have originated similarly.

On the other hand, gradational contacts between metasedimentary and granitic rocks do exist. Examples of these are the migmatite zones, and the gradation of silexite into schist and limestone. Furthermore, because it is well-known that sharply defined fronts commonly separate replacement bodies from unreplaced rocks, the sharp-contact criterion, as evidence against replacement should be used with caution. Moreover the planar structures of the granitic rocks, in general, are not attributable to flowage, because they parallel planar structures in the metasediments, regardless of marked irregularities in the contacts between the two rocks. If the granitic rocks crystallized from a magma, post-emplacement stress must have produced their planar structure. If of replacement origin their structures may be relict from the metasedimentary rocks or may have been caused by stress during or after the initial crystallization of the granite-forming minerals. The metasedimentary "islands", arranged in orderly belts in normal stratigraphic sequence and in dimensional and textural alignment, do not have the appearance of xenoliths. Their textural alignment may well be an effect of post-emplacement stress; but, in view of the lack of large-scale deformation in the larger metasedimentary masses, such stress could hardly have drawn all of the inclusions, regardless of size to their present dimensional alignment. If the granitic rock was once mostly a magma, stoping into a roof pendent, of which each "island" was a part, may have caused their present disposition and alignment; but many of the smaller "islands" appear to have been pod-shaped.

With the evidence thus in apparent conflict, and with the lack of proved criteria, the origin of the granitic rocks must remain speculative. Probably both processes, crystallization from a granitic magma and replacement, were effective. But replacement evidence, particularly as shown by the disposition of the metasedimentary "islands", seems most abundant to the writer. Extensive replacement would have involved principally the introduction of potash and soda and the removal of most of the metasedimentary CaO and MgO. The average chemical composition of the five metasedimentary members probably originally lay between 5 and 10 percent MgO, compared with an MgO content for the average granitic rock of less than 2 percent.

TALC BODIES

GENERAL COMPOSITION AND DISTRIBUTION

The rocks mined as commercial talc in the Silver Lake area are composed principally of magnesian silicate minerals. Listed in order of decreasing abundance, these are tremolite, talc, forsterite, serpentine, and chlorite (?). Tremolite forms an estimated three-fourths of the volume of the bodies, and talc is markedly in excess of the others. Calcite is ordinarily present, but in amounts of only 2 to 3 percent.

All of the material currently mined in the area is obtained from bodies within the hornfels member. One relatively small body, which is not being worked, occurs a few hundred feet south of the Number Two and One-half workings. This body, however, is also enclosed in green diopsidic hornfels, apparently a lense separate from the member. Where the hornfels member exists in its full 150-foot thickness, the talc bodies lie very close to the center.

SIZE AND SHAPE

Although largely confined to a narrow, persistent zone, the talc bodies themselves are much less continuous. The mineable bodies range from a few tens of feet to about 800 feet long and from 5 to 15 feet wide. The larger bodies ordinarily occur in parallel pairs (pl. 11a, fig. 4) separated by from 10 to 20 feet of diopsidic rock.

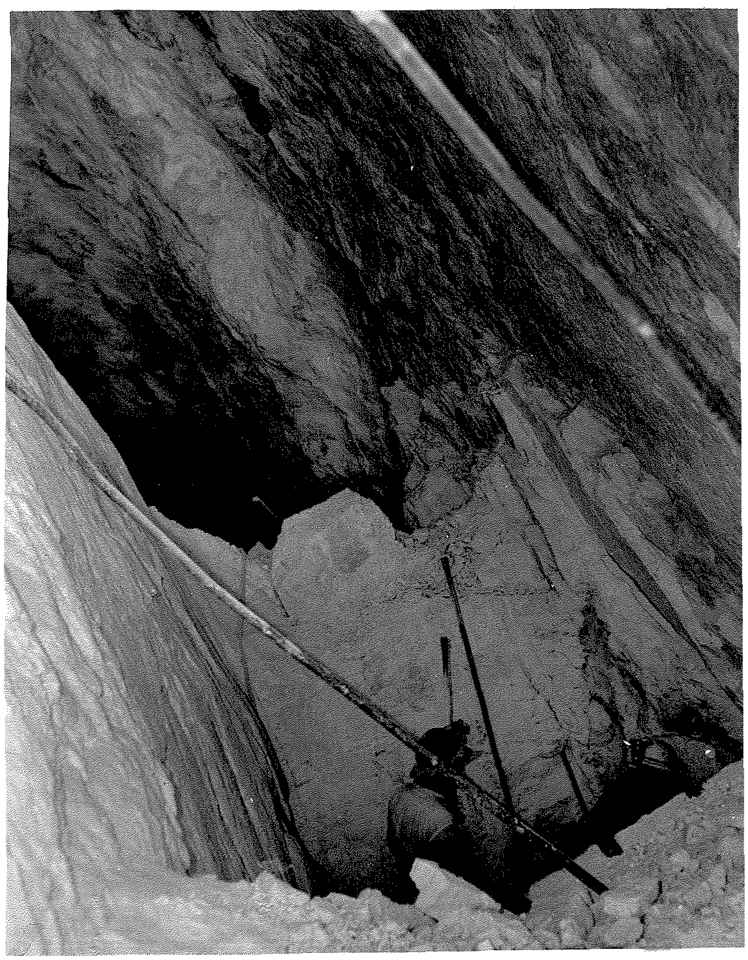
Individual bodies terminate in several ways. Many of them narrow gradually, lensing into diopsidic rock. Others are brought against diopsidic rock by cross-faults. Several of the bodies in the Addenda workings end abruptly against irregular masses of pegmatite. The principal cause of discontinuity, however, has been the emplacement of the larger masses of granitic rocks. One or more of the talc bodies at each of the five workings terminate against rock of the tonalite unit. Microcline silexite transgresses the surface exposures of one body at the eastern end of the Gould workings, and also transgresses the downward extension of much or all of the Gould talc zone at a depth of about 300 feet.

Very pronounced planar structures, in the massive tremolitic rocks as well as the talc schists, parallel the planar structures of the nearby granites and metasedimentary rocks. In the massive tremolitic rocks these structures are caused mostly by variations in grain size, by the concentration of forsterite, calcite, or talc along even layers, and by parallelism of serpentine veinlets.

White tremolitic rock, the most abundant variety in most of the talc bodies, ordinarily occupies from three-fourths to their entire thickness (fig. 5). Layers of talc schist, from a few inches to as much as five feet thick, occur along the foot-walls of most bodies, less abundantly along the hanging walls and within the bodies. The talc schist appears to have formed along shear zones and at the expense of the various types of massive tremolite rock. Tremolite-forsterite-serpentine rock, pre-dating the white tremolitic rock, occurs as residual masses within it and locally forms the entire width of a talc body. Green tremolitic rock, the least abundant of the four, was noted locally in several-foot layers near hanging walls.

Plate 10a. (above). View westward along surface exposures of Gould talc bodies. Two parallel bodies are bordered by diopside-feldspar quartz-calcite hornfels. Tonalite and quartz-biotite schist member exposed on hill to upper right.

Plate 10b. (right). Typical stope in massive tremolitic rock of Gould deposits.



ROCK TYPES AND INTERNAL STRUCTURE OF BODIES

General statement

The commercial talc bodies are composed of several metamorphic rock types characterized by contrasting textures and distinctive magnesian silicate mineral assemblages. All of the types have been marketed as ceramic raw material. They are broadly divisible into (1) schistose rocks composed mostly of the mineral talc and (2) massive rocks predominantly of tremolite. The highly tremolitic rocks are further subdivisible into three varieties: (1) a snowy white rock with subordinate amounts of the mineral talc, (2) a pale bluish green, virtually monomineralic tremolite rock, and (3) a pale yellowish green to pale brownish gray rock rich in disseminated forsterite and numerous serpentine veinlets. The white variety is by far the most abundant, but the other two were noted in relatively large amounts at several places. All gradations exist between the white rock and talc schist containing virtually no tremolite.

White tremolite rock

The snowy white variety of massive tremolitic rock is typically composed of more than three-fourths tremolite, less than one-fourth talc and a few percent carbonate. It is a medium- to coarse-grained rock, and breaks into tough, irregular blocks. Although a decussate texture is most common (fig. 7), layers of contrasting grain size, from one-eighth to one-half inch thick, produce a laminated appearance. Many of the layers consist of tremolite blades lying normal to the planar structure. Less common are layers of tremolite blades dimensionally oriented parallel with the planar structure. Concentrations of micaceous talc within the rock form other schistose layers as small-scale counterparts of the larger bodies of talc schist along the walls of the bodies.

Most of the tremolite blades are between 2 mm. and 2 cm. long. The talc occurs in two principal habits; one equant, the other micaceous. The observed equant grains, as much as 3 mm. in diameter, are disseminated through the tremolite, and represent replacements of it. They commonly contain aligned tremolite residua, but pseudomorphs of talc after tremolite are rare. The micaceous talc forms the schistose layers. In the thinnest layers, aligned talc shreds cross the decussate tremolite needles in a markedly contrasting texture. Tremolite residua that lie askew the schistosity are common.

The carbonate, which forms 2 to 3 percent of the white tremolite rock, occurs in seams and in grains interstitial to the tremolite needles. It occurs in distinct but cloudy grains and appears to have formed contemporaneously with or later than the tremolite. Phlogopite is locally a minor constituent.

Green tremolite rock

The pale bluish green tremolite rock, much finer grained than the snowy white variety, is composed of needles , mostly less than 1 mm. long. The specimens observed in thin section contain only tremolite, the decussate texture of which imparts and unusual toughness and compactness to the rock. This rock ordinarily occurs in layers, less

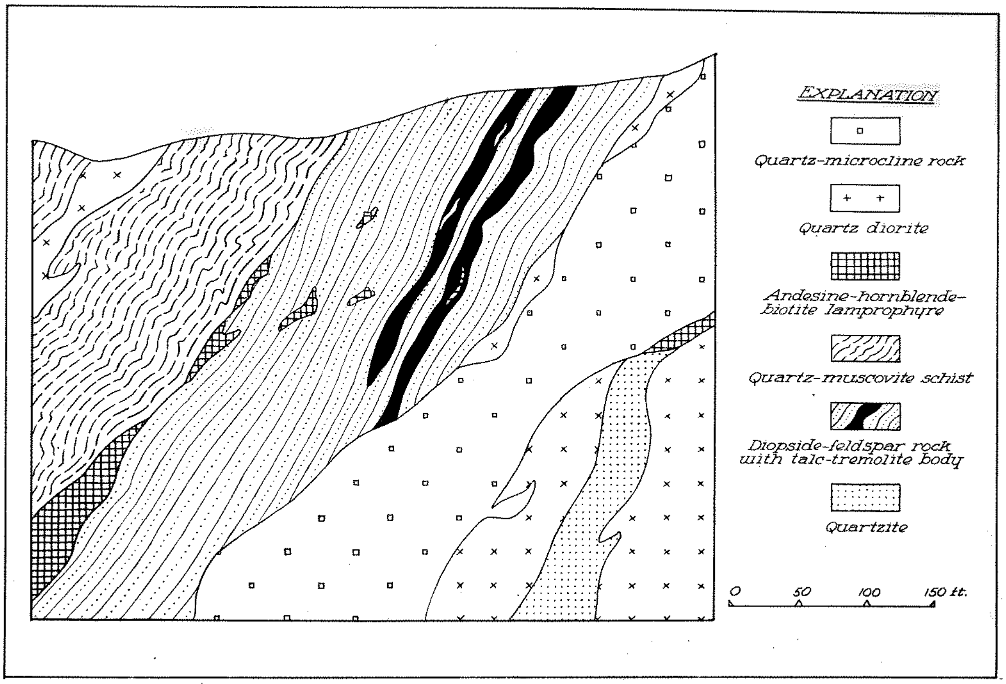


Figure 4. Generalized cross-section through Silver Lake talc-bearing zone.

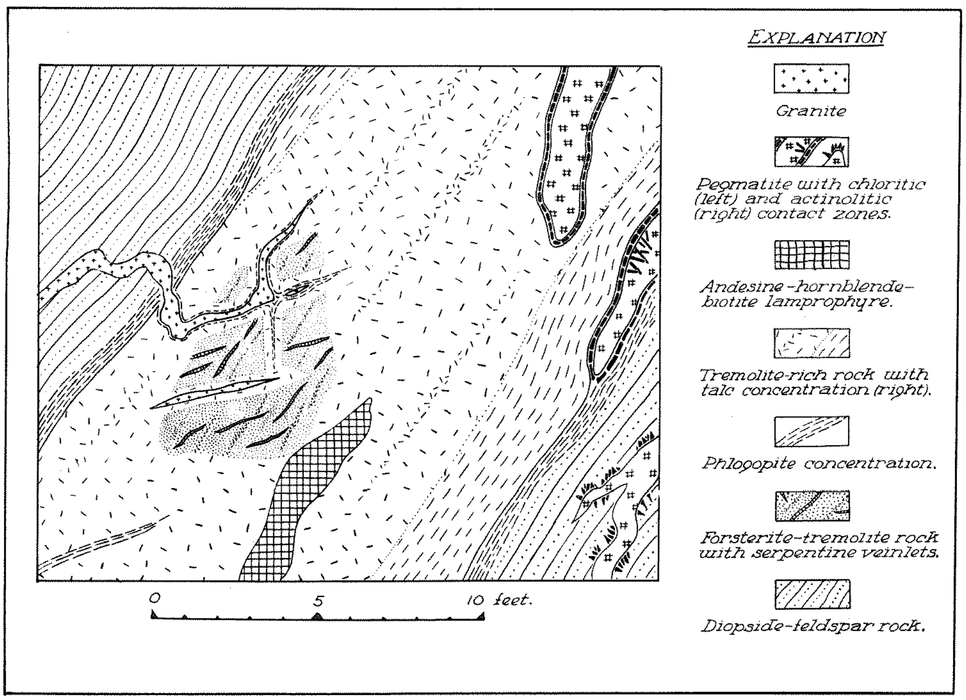


Figure 5. Generalized cross-section showing rock units and internal structure of typical Silver Lake talc body.

than 3 feet in width, parallel to the other planar features of the talc bodies. Such layers were particularly well developed near the hanging wall of the Number Two and One-half deposit, now largely worked out. Where the white and green varieties are in contact, the white tremolite veinlets commonly extend into the green rock; in other places the two types grade into each other.

Tremolite-forsterite-serpentine rock

The massive rock composed mostly of tremolite, forsterite and serpentine and pre-dating the white tremolitic rock, is relatively abundant in the Gould and Eastern Addenda deposits. In the Gould workings it ordinarily occurs as irregular masses, several feet in long dimension, enclosed by white tremolite; but locally in these deposits and extensively in Eastern Addenda deposits, this rock occupies most or all of the entire width of an individual talc body.

The tremolite-forsterite-serpentine rock, colored pale yellowish green to light brownish gray, is composed principally of tremolite. Forsterite and serpentine, in various proportions, form as much as one-third of the rock (fig. 6). It ordinarily also contains an appreciable proportion of talc and two to five percent of carbonate material.

As in the white tremolite rock, decussate tremolite needles, mostly from 2 mm. to 2 cm. long, give the rock an unusual toughness. In hand specimen this tremolite is generally darker colored than the tremolite of the white rock.

Disseminated grains of forsterite or talcose ghosts of forsterite, in planar concentrations, cause a distinct layering. Most of the

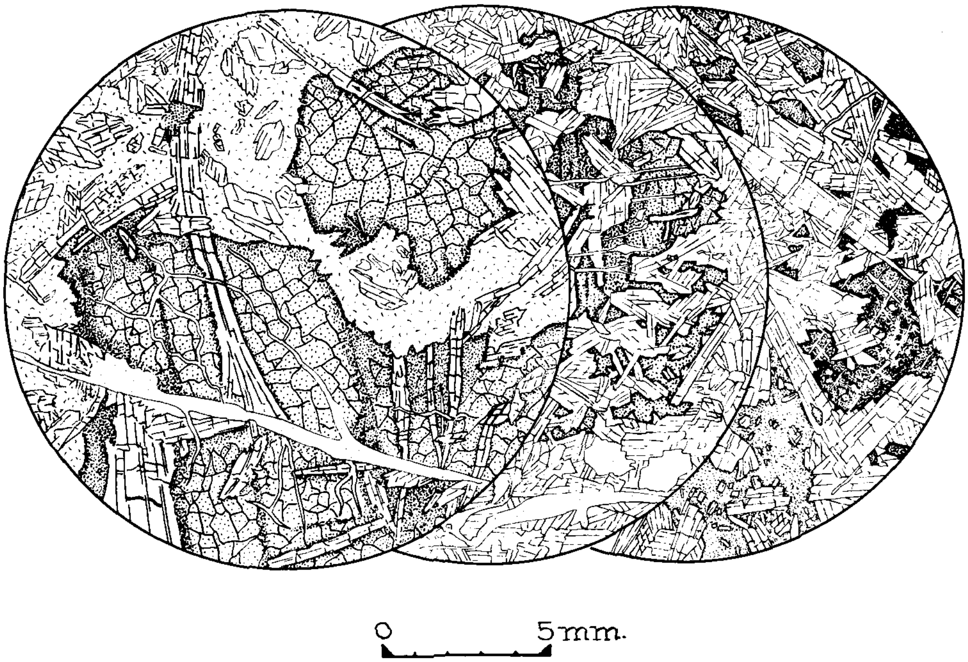


Figure 6. Tremolite-forsterite-serpentine rock showing progressive replacement of forsterite by tremolite. Tremolite partly replaced by talc (upper left) and by carbonate (lower right). Chrysotile veinlets cut all other minerals.

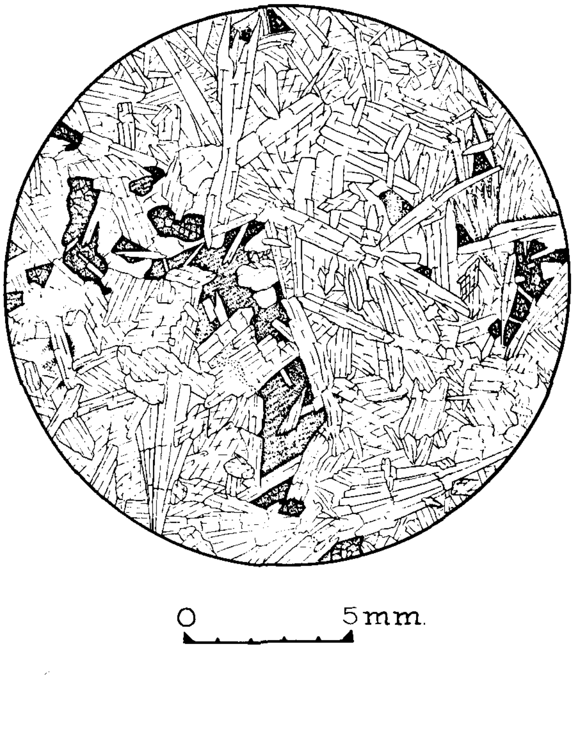


Figure 7. White tremolite rock containing talcose material (center; probably ghost of a forsterite grain) and carbonate.

forsterite is in relatively large, equant grains as much as 5 mm. in diameter. Some of the grains are markedly elongate.

Thin section studies show all stages of a forsterite to tremolite alteration ranging from the marginal corrosion of forsterite by tremolite blades to the existence of aligned forsterite residua in a tremolite mesh. The original outlines of the forsterite grains are generally detectable, and indicate that the mineral formed between one-eighth and one-fourth of the volume of the pre-tremolite rock.

Chrysotile forms alteration rims about forsterite grains, pseudomorphs after tremolite, and networks of megascopic and microscopic veinlets extending across forsterite, tremolite, talc and carbonate grains and grain aggregates indiscriminantly. The veinlets are apple green in hand specimen. They rarely exceed one-fourth inch in width; most of them parallel the other planar features of the nearby rock units. Many, however, form seemingly random, criss-cross patterns. The chrysotile and forsterite, ubiquitous in this earlier metamorphic rock, do not exist in the other rock types of the talc bodies. Chrysotile veinlets extend to, but not beyond, contacts with white tremolitic rock.

The mineral talc occurs in the serpentine- and forsterite-bearing tremolite rock in several habits, but not in the schistose layers so common elsewhere in the deposits. The forsterite has commonly altered to a dark-brown, felty, and very fine-grained talcose material. All stages in this alteration are shown. The completely talcose ghosts of forsterite appear to be most abundant near contacts with the white tremolite rock and a few were noted in the white tremolite rock itself.

Talc also occurs in colorless, fine-grained aggregates that have formed at the expense of tremolite, and, less abundantly, as a replacement of the chrysotile.

The carbonate material is ordinarily very fine-grained and occurs in veinlets and irregular aggregates. An intimate association of carbonate veinlets with chrysotile veinlets, commonly with mutually cross-cutting relations, probably indicates contemporaneity. The carbonate aggregates, largely, if not wholly, formed at the expense of tremolite, and are closely associated with chrysotile that has also replaced tremolite.

That the occurrences of tremolite-forsterite-serpentine rock are older than the white tremolite rock with which they are associated is shown in several ways. In many places the tremolite-forsterite-serpentine rock is cut by phlogopite-filled fractures along which white tremolite needles have grown normal to the fracture walls. Granitic dikes and dikelets that cross this rock are also bordered by zones of white tremolite. Such zones range from a fraction of an inch to several feet in width, and are commensurate in size with the granitic bodies they border. The tremolite-forsterite-serpentine rock that occupies most of the thickness of one of the Eastern Addenda bodies is bordered on both the hanging and foot-walls by one- to three-foot zones of white tremolite rock that appear to have formed along shearing planes localized by the contacts of the deposit with hornfels.

Talc schist

Talc schist, so common in relatively thick layers along the foot-walls of the talc-tremolite bodies, occurring less abundantly along the hanging walls and within the bodies, was the material most sought in the early mining operations. Some is reported to have been marketed as cosmetic talc. Now the schist is mined with massive tremolite rocks and mixed with them.

The talc schist is a snowy white, micaceous, and very friable rock. Although consisting mostly of talc, it contains variable proportions of tremolite and chlorite (?) and a percent or two of carbonate (predominantly calcite). The talc grains, micaceous in habit, are commonly a centimeter or more in diameter. In close dimensional alignment, and characteristically curved, they produce an undulating schistosity and give the rock a lustrous, pearly sheen. Tremolite blades, locally abundant, are less evenly aligned than the talc and are corroded and cross-cut by talc shreds. A mineral with a very low birefringence, tentatively identified as a colorless chlorite, is intimately associated with the talc. Pseudomorphism may be indicated by similarity in habit, but replacement textures were not observed.

METAMORPHISM

Nature of the parent rock

A sedimentary parent rock for the "talc" bodies is indicated by their occurrence in the metasedimentary section as well-defined layers and elongate lenses, paralleling each other and composed of mineral assemblages characteristic of metamorphosed carbonate rocks. That the bodies were strata that contrasted in composition with bordering strata from which the hornfelsic rocks altered, is strongly suggested by

(1) the persistence of the "talc" bodies, ordinarily in pairs, in the same stratigraphic position for the entire 2-mile length of the talc-bearing zone, (2) relatively uniform thicknesses, (3) a marked contrast, in mineralogy and bulk chemical composition, between the bodies and the bordering hornfelses, and (4) a close resemblance in plan with other layers in the section that are unquestionably sedimentary strata (i.e. marble layers in one quartz-muscovite-schist member).

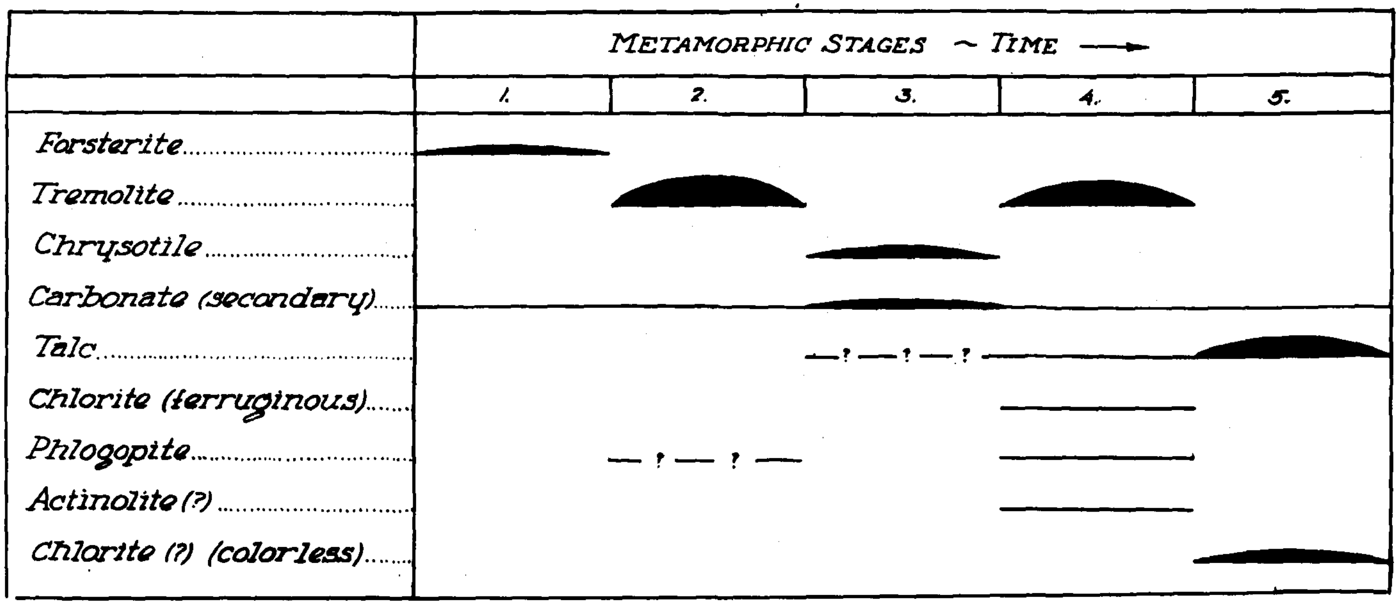
The abundance of forsterite and lack of disseminated quartz in the talc bodies, together with the diopside-quartz association in the hornfels, probably indicates that the original rocks were respectively subsilicic and silica-rich. It will be remembered that in the other metamorphic rocks of the area, the forsteritic units are also quartz-free, whereas the quartz-rich calcareous rocks are forsterite-free and generally diopside-bearing. The forsterite-quartz association, commonly cited as an example of disequilibrium in other metamorphic terranes, apparently does not exist in the Silver Lake mine rocks.

The sparsity of alumina, soda, and potash in the talc bodies probably is also a feature inherited from the original strata. Although much of the potash and soda in the hornfelsic rocks may well have been introduced, these rocks apparently contained silica and alumina by which the alkalies could have been fixed.

The phlogopite within or bordering the talc bodies may have been derived partly or wholly, from original material, the crystallization of feldspar prevented by a high proportion of magnesium. The phlogopite-rich border zones may be effects of metamorphic differentiation involving an outward movement of original material. But even if the potash and alumina in such zones were distributed through the talc bodies associated with them, the bodies would still be much poorer in these materials than comparable thicknesses of bordering hornfels. Moreover the phlogopite borders along granitic dikes, and phlogopite-filled fractures strongly suggest that alumina and potash have been introduced into the talc bodies late in their metamorphic history. The talc bodies, therefore, originally may well have been magnesian carbonate rock, low in silica and nearly free of other impurities.

Physical and chemical environments

The paragenesis of the Silver Lake talc bodies is divisible into five stages, each reflecting a change in physical or chemical environment (fig. 8). In brief, these stages appear to have been as follows. First, dolomite, low in silica, was heated under stress to temperatures in the 600 to 800 degrees C. range, with little or no accompanying metasomatism, yielding a forsterite-bearing carbonate rock. Second, extensive silica and magnesia metasomatism, under non-stress conditions and somewhat lower temperatures, yielded the first generation of tremolite, mostly at the expense of carbonate, partly at the expense of forsterite. Third, the temperature decreased to less than 500 degrees C., producing a partial serpentinization of forsterite and tremolite and a partial carbonatization of tremolite, with little change in bulk chemical composition. Fourth, a rise in temperature concurrent with the intrusion of granitic dikes, and accompanied by the introduction of several percent each of CaO and SiO_2, produced the second generation of tremolite at the expense of all pre-existing



1. Maximum temperatures, probably in 600° C. to 800° C. range; stressed environment; little or no metasomatism; probably concurrent with lamprophyre emplacement.

2. Lower temperatures, probably in 500° C. to 600° C. range; non-stressed environment; extensive $MgO-SiO_2$ metasomatism, genetically related to tonalite.

3. Temperatures below 500° C.; non-stressed environment; metasomatism negligible; post-tonalite.

4. Temperatures probably above 500° C.; non-stressed environment; metasomatism negligible; concurrent with emplacement of granitic dikes.

5. Temperatures below 500° C.; stressed environment; probably moderate MgO metasomatism; follows granitic dike emplacement.

Figure 8. Paragenesis diagram of principal minerals in commercial talc bodies.

minerals. Preceding their alteration to tremolite many of the forsterite residua were altered to a talcose material. Fifth, a renewal of stress, MgO enrichment and probable lowering of temperature yielded abundant talc and subordinate chlorite (?) largely at the expense of second generation tremolite.

That volumes remain essentially unchanged during the second, third, and fourth stages, is indicated by the growth of newly-formed minerals into the pre-existing rocks with no apparent disturbance in the relict textures. The changes in bulk chemistry during these steps, therefore, probably were truly metasomatic.

The silica-magnesia metasomatism of stage two is indicated by (1) the contrast in chemical composition between the inferred low-silica dolomite and the commercial talc rocks (table 2) which consistently contain more than 25 percent MgO and more than 60 percent SiO_2; and (2) the post-forsterite age of the first generation of tremolite. An early production of forsterite at the expense of dolomite and silica, according to the equation

$$\underset{\text{dolomite}}{2CaMg(CO_3)_2} + \underset{\text{quartz}}{SiO_2} \rightleftharpoons \underset{\text{forsterite}}{Mg_2SiO_4} + \underset{\text{calcite}}{2CaCO_3} + 2CO_2$$

would have brought a slight decrease in volume and enrichment in MgO. But such enrichment could not have produced the MgO fraction of the commercial talc rocks. The MgO content of an ideally pure dolomite, it will be recalled, is 19.1 percent. Similarly SiO_2, which probably composed no more than one-eighth of the original rock, apparently increased more than 5-fold in the development of the first generation of tremolite. The formation of tremolite at the expense of calcite-rich carbonate material, would also involve removal of CaO and CO_2. The metasomatism thus related to stage two seems to have produced the most pronounced chemical changes in the metamorphic development of the talc bodies. Why tremolitization in the Silver Lake area was largely confined to the parent rocks of the talc bodies, whereas other calcareous metasediments contain little or no tremolite, is not clear. Perhaps most significant was the degree of access to tremolitizing solutions as determined by the physical nature of the various rocks.

The 600 to 800 degree temperature range postulated above for the formation of diopside in the hornfels may be also claimed for the forsterite, as the two appear to have been coexistant. That the tremolite formed at somewhat lower temperatures is indicated by the inability of tremolite to exist in equilibrium with calcite in a dolomitic rock above the fourth of Bowen's[29/] thirteen steps. Points on Bowen's

29/ Bowen, N. L., op. cit., p. 245.

schematic P-T curve for this step include 550 degrees C. at 500 atmospheres pressure and 630 degrees C. at 2000 atmospheres. It will also be remembered that tremolite precedes both forsterite and diopside when listed in their theoretical order of production with rising temperature.

A further decline in temperature is probably also shown in the appearance in stage three of chrysotile, a mineral which, as noted above, cannot form at temperatures above 500 degrees C. at any pressure. The addition of CO_2, probably at the expense of SiO_2, was needed in the carbonatization of tremolite. Otherwise the changes in bulk chemistry were insignificant during this stage.

The disappearance of chrysotile and reappearance of tremolite in the fourth stage apparently record a rise in temperature above 500 degrees C. This is further indicated by alteration of forsterite to talc near contacts with white tremolite rock. Bowen and Tuttle[30] have

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30/ Bowen, N. L., and Tuttle, O. F., op. cit., p. 452.

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shown that, whereas the serpentinization of forsterite can proceed only at temperatures below 500 degrees C., these workers state that above this temperature "a medium which could add silica or subtract

> magnesia could change forsterite only to talc or to talc and enstatite if the supply of water were deficient. Talc could form from forsterite at low temperatures also but only with intermediate formation of serpentine."

The white tremolite zones that border granitic dikes may well be primarily an effect of heating and suggests that the temperature of the granitic material was well in excess of 500 degrees C.

A replacement by nearly monomineralic tremolite rock of a tremolite-forsterite-serpentine-calcite rock would require principally the subtraction of CO_2 and addition of SiO_2 , both in small proportion.

The fifth and final stage in the metamorphism, as shown in the alteration of the white tremolite to talc, indicates a removal of CaO and an enrichment in MgO. In the formation of the equant talc grains, causing little or no change in volume, an actual addition of MgO is probably indicated. Whether the development of the large talc schist layers caused significant volume changes is not known, but the enrichment in MgO may have been largely or wholly an effect of a shearing out of CaO. In thin section, however, the smaller schist layers appear to have evolved with only a slight disturbance of tremolite. It is probable, therefore, that the development of at least some of the schist-forming talc required additive MgO. At this late stage most of the CO_2 evolved in the earlier stages had probably escaped. This may partly account for the sparsity of carbonate material associated with the tremolite to talc alteration, whereas the earlier tremolite to chrysotile alteration was accompanied by formation of several percent of calcite.

Source of the additive materials

Either by strong inference or by direct evidence the stages in the metamorphism of the talc bodies can be correlated in time with stages in the emplacement of the lamprophyric and granitic rocks. The lamprophyre probably was emplaced during the first metamorphic stage previous to the hydrothermal conditions characteristic of the later stages. The lamprophyre, distinctly more mafic than the granitic rocks, suggests a hotter, dryer environment. The lamprophyre bodies, it will be remembered, show no contact effects, whereas zones of hydrous minerals commonly border the granitic dike rocks.

The tremolitization, marking the pronounced chemical changes of the second stage, appears to have been contemporaneous with the emplacement of the large granitic masses, partly at the expense of magnesium-rich metasedimentary rocks. The hydrothermal activity, shown in the silica-magnesia metasomatism of the talc bodies as well as in the alkali metasomatism indicated for other metasedimentary units, is compatible with the aqueous environment characteristic of granitic rocks.

	1	2	3	4	5	6	7	8
SiO_2	51.17	57.40	58.90	59.25	58.12	59.05	56.29	56.70
Al_2O_3	0.64	1.29	0.57	0.53	0.60	0.78	1.07	9.16
Fe_2O_3 / FeO	0.36	0.86	0.30	0.30	0.28	0.42	0.43	2.04
MgO	29.80	23.91	25.14	25.80	28.65	28.67	28.31	8.55
CaO	11.25	13.55	12.66	11.82	8.11	5.81	9.26	14.49
Na_2O	0.20	0.33	0.26	0.13	0.27	0.41	1.06	1.26
K_2O	0.10	0.11	0.11	0.06	0.11	0.18		5.50
H_2O^+	3.79	2.12	1.60	2.24	2.89	3.71		0.90
H_2O^-	0.67	0.08	0.11	0.13	0.09	0.19	4.02	0.31
CO_2	2.37	None	None	0.02	0.06	0.42		2.30
MnO	0.03	Tr.	Tr.	None	Tr.	Tr.		0.03
Total	100.38	99.65	99.65	100.28	99.18	99.64	104.44	101.74

1. Forsterite-tremolite-serpentine-calcite rock from Gould workings.
2. Green tremolitic rock from No. Two and One-half workings.
3. White tremolitic rock from Gould workings.
4. White talc-tremolite rock from Gould workings.
5. Green talc-tremolite rock from No. Two and One-half workings.
6. Talc-tremolite schist from Gould workings.
7. Commercial talc blend.
8. Diopside-feldspar hornfels wall rock from No. Two and One-half workings.

Table 2. Analyses of representative samples of Silver Lake commercial talcs and wall rock; Alberta J. McArthur, Sierra Talc and Clay Company, analyist.

Moreover the extent of the metasomatism requires a comparably large-scale source.

The third stage, which yielded chrysotile, apparently marks a period separating the end stages of the emplacement of large granitic bodies and the intrusion of the granitic dikes, and accompanied by lower temperatures and decreased hydrothermal activity. The fourth stage is clearly fixed in time by the white tremolitic rock that borders granitic dikes. The fifth or talc-forming stage post-dates the granitic dikes.

The additive material may have been derived from a granitic magma or it may be transferred metasedimentary material. Whether or not the large granitic masses are partly replacement bodies or solidified from a magma, the primary source of the H_2O and SiO_2 was probably mostly magmatic. Silicification and hydration characterize granitization and are common effects of granite intrusion. That granitic magmas can also yield hydrothermal MgO is shown by the common occurrence of magnesian silicate skarns along granite-limestone contacts. But such skarns are ordinarily ferruginous, whereas the Silver talc bodies are not. Iron-free MgO-laden solutions, however, may well have formed during granitization of the magnesium-rich metasedimentary rocks. If MgO was also introduced during talc formation in the fifth stage, the MgO-bearing solutions must have been residual or were derived from an undetermined source.

DEPOSITS OF THE SOUTHERN DEATH VALLEY-KINGSTON RANGE DISTRICT

INTRODUCTION

DELINEATION OF THE DISTRICT

The talc deposits of commercial interest in the Southern Death Valley-Kingston Range district are confined to the Crystal Spring formation, the lowest unit of the late pre-Cambrian Pahrump series. Each deposit is further restricted to the lower part of a massive carbonate member in the middle of the formation and is associated with a diabase sill intruded at or near the base of the member. At every locality visited by the writer, where this part of the formation is exposed, talc mineralization was noted.

The outline of the talc-bearing district (fig. 9), therefore, encloses virtually all of the known exposures of the Crystal Spring formation. The talc-bearing district, a northwest-trending belt athwart the San Bernardino-Inyo county line, is about 70 miles long and averages 15 miles wide. It contains a part of the southeastern Panamint Range, the southern part of the Amargosa Range, the Ibex Hills, the Alexander Hills, a narrow east-trending belt in the central part of the Kingston Range, the western part of the Silurian Hills, and the northern tip of the Owlshead Mountains. The western one-third of the district is largely with in the limits of Death Valley National Monument.

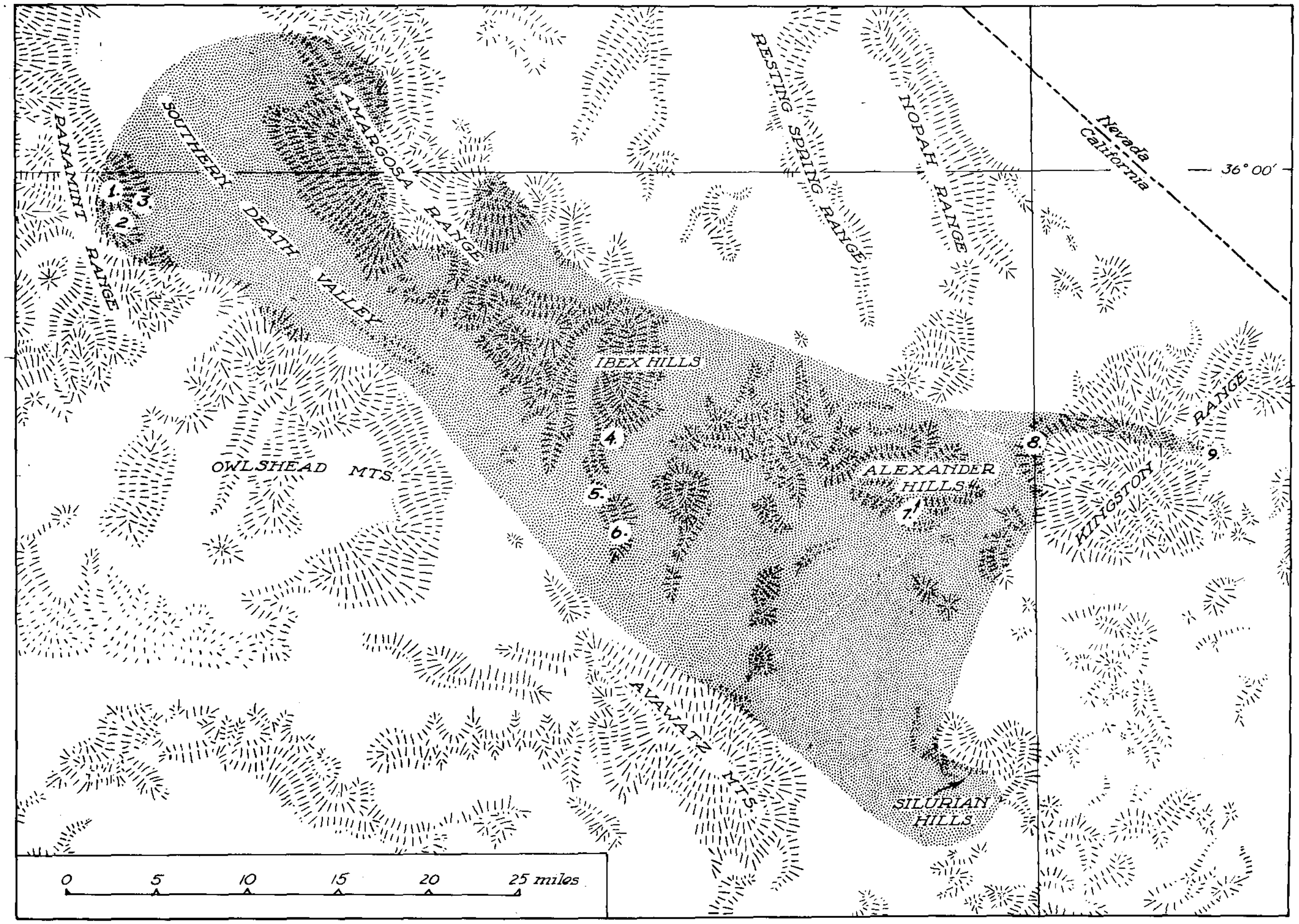

Figure 9. Map of southern Death Valley-Inyo Range region showing locations of principal talc mines; (1) Montgomery, (2) Warm Spring, (3) Death Valley, (4) Ibex and Monarch, (5) Pongo, (6) Superior, (7) Western, (8) Tecopa, and (9) Excelsior. Stippled area contains all known exposures of Crystal Spring formation.

PHYSICAL FEATURES

The Southern Death Valley-Kingston Range talc deposits are in an area famed for its magnificent desert scenery, but likewise noted as one of the most arid and least inhabited regions of North America. Within the district, however, are two settlements, Tecopa and Shoshone, with a combined population of about 300 people. Both are on the Amargosa River east of the Amargosa Range. Tecopa is about 2 miles north of the San Bernardino-Inyo County line. Shoshone is about 9 miles north of Tecopa.

The talc deposits range in altitude from approximately 300 feet near Saratoga Spring at the southern end of Death Valley to nearly 5000 feet in the Kingston Range. Rainfall in the district ranges from an annual average of less than 2 inches on the floor of Death Valley to somewhat more than 5 inches in the higher parts of the Amargosa and Kingston Ranges. All but the easternmost tip of the province is drained by the Amargosa River and its tributaries. The Amargosa, whose headwaters are near Death Valley Junction about 20 airline miles north-northwest of Shoshone, flows south-southeastward to a point about 10 miles south of Tecopa. From this point, the river swings westward and then northward in a broad arc, and ends in the undrained Death Valley basin. The channel carries large quantities of water only during periods of heavy rainfall. Ordinarily it has a continuous, though feeble, flow in the winter months and is dry during the rest of the year.

Except for local marshy areas along the Amargosa River, and in the vicinity of widely spaced springs, the vegetation of the region is exceedingly sparse. Most of the mountain slopes are virtually barren of plants or overburden. Much of the pre-Quaternary geology, however, is hidden beneath depressions containing Pleistocene or Recent continental sediments.

The central part of the province is crossed by State Highway 127, which extends north-northwestward from Baker through Shoshone to Death Valley Juntion. Talc mined in the region formerly was shipped from various points on the Tonopah and Tidewater Railroad which, in general, paralleled the highway. Since 1941, when the rails were removed, talc has been trucked to Dunn siding, a point on the Union Pacific Railroad. Dunn is about ninety miles by road from Shoshone, and 155 miles by rail from Los Angeles.

Roads leading east and west of Highway 127 extend to most of the talc-bearing localities. Autos can be driven to within a mile of nearly every other locality in the district. Most of the mine access roads are graded and fairly well-kept. Others are unimproved and should be used only by persons experienced in desert driving, and preferably with truck-type or four-wheel-drive vehicles.

DISTRIBUTION OF TALC DEPOSITS

As inferred above, the numerous talc-bearing localities in the Southern Death Valley-Kingston Range district are nearly as widespread as exposures of the Crystal Spring formation itself. In the southeastern part of the Panamint Range, deposits have been operated at the Warm Spring mine in Warm Spring Canyon (pl. 16a), at the Death Valley

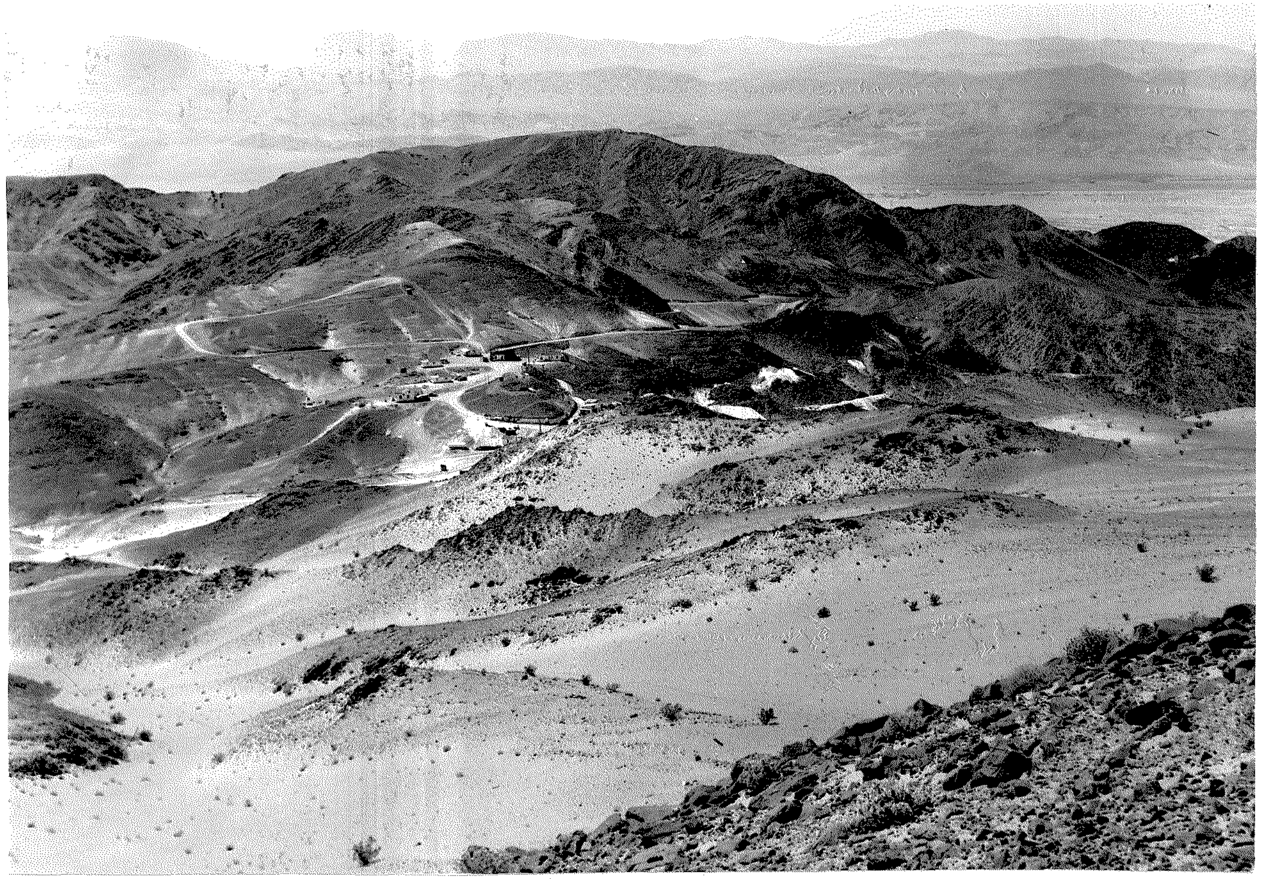

Plate 11. View southward of Superior mine area with southern end of Death Valley in background.

mine in Galena Canyon, and at the Montgomery mine high on the ridge separating the two canyons. Deposits on the Panamint claims, between Anvil Spring and Warm Spring Canyons, were scheduled to be opened in 1951.

Talc-bearing Crystal Spring rocks also exist in the south central part of the Amargosa Range; but these are in relatively small blocks in the chaotic terrane described by Noble[31/] and Curry[32/]. None of these

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31/ Noble, L. F., Structural features of the Virgin Spring area, Death Valley, California: Geol. Soc. America Bull., vol. 52, pp. 941-999, 1941.

32/ Curry, H. D., "Turtleback" fault surfaces in Death Valley, California (abstract); Geol. Soc. American Bull., vol. 49, p. 1875, 1938.

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deposits has been seriously worked. Beds of the Crystal Spring formation are exposed in the southernmost tip of the Amargosa Range, about midway between Ibex Spring and Confidence Mill. Here a talc-bearing zone is discontinuously exposed for a distance of about 3 miles; but this zone , which includes the Brown and Valley deposits, has been only prospected.

Several talc mines and unworked talc deposits are in the north-trending Ibex Hills that lie east of the southern part of the Amargosa Range and extend from the Sheephead Pass area southward to the vicinity of Saratoga Spring. In these hills, the Crystal Spring formation is almost continuously exposed for a distance of 12 miles. Here too, the formation is in comparatively large faulted blocks. The Eclipse mine is on the east flank; the undeveloped Markley deposits are near the crest of the northern part of the hills. The closely spaced Monarch, Pleasanton, and Ibex mines (pl. 12a) are in the vicinity of Ibex Springs and south of the Markley deposits. In the southernmost group of the Ibex Hills north of Saratoga Spring, are the Superior, White Cap, Pongo, and Saratoga mines.

The Grimshaw prospects are exposed low on the west side of a eastward-tilted block of late pre-Cambrian sediments that lie east of the southern part of the Ibex Hills. The original Acme mine is in an exposure of the Crystal Spring formation in the Rainbow Mountain area about 4 miles south-southeast of Tecopa.

In the Alexander Hills, about 9 miles southeast of Tecopa, the Crystal Spring formation is exposed in relatively large, moderately tilted fault blocks. These contain the Western (fig. 11, pl. 15b), Booth mines, and another Acme mine which, to avoid confusion with the original Acme mine at Rainbow Mountain, will be referred to as the "New Acme mine". The Donna Loy mine is in exposures of the formation at the southern end of the Nopah Range north of Tecopa Pass. The Crystal Spring prospect and the Harry Adams, and Excelsior mines are in a narrow belt of the Crystal Spring formation that extends eastwardly through the central part of the Kingston Range. The Tecopa (Smith) and Rogers mines are in Crystal Spring sediments exposed on the west central flank of the Kingston Range.

The southernmost of the known exposures of the Crystal Spring formation are in the western part of the Silurian Hills. Here talc has been obtained from the Annex Number One and Berryhill mines; several other properties have been prospected. The Sheep Creek deposit, near the mouth of Sheep Creek Canyon on the northern slope of the

Avawatz Range, is in a sliver of Crystal Spring rock in the Garlock fault zone. The only other talc-bearing area within the district, and known to the writer, is high on the northernmost tip of the Owlshead Mountains. Here a group of undeveloped deposits are contained in a steeply north-dipping Crystal Spring section.

A description of the geological features of each of these talc-bearing areas cannot be included in the present discussion. As a group, they have many features in common, suggesting that the deposits have formed under essentially the same geological conditions.

THE CRYSTAL SPRING FORMATION

PREVIOUS INVESTIGATION

In the mid-eighteen-seventies G. K. Gilbert[33/], upon observing the geological features of the Saratoga Spring area, was the first to describe the marine sedimentary and intrusive diabasic rocks that were later to be included in the Pahrump series. Gilbert noted their unfossiliferous character, but did not assign an age. In 1902 the same exposures were briefly mentioned by M. R. Campbell[34/] who presumed the

33/ Gilbert, G. K., Report upon the geology of portions of Nevada, Utah, California, and Arizona, examined in the years 1871 and 1872: in Geog. and Geol. Surveys W. 100th Mer. Rept., vol. 3, pp. 34, 170, 1875.

34/ Campbell, M. R., Reconnaissance of the borax deposits of Death Valley and Mojave Desert: U. S. Geol. Survey Bull. 200, p. 14, 1902.

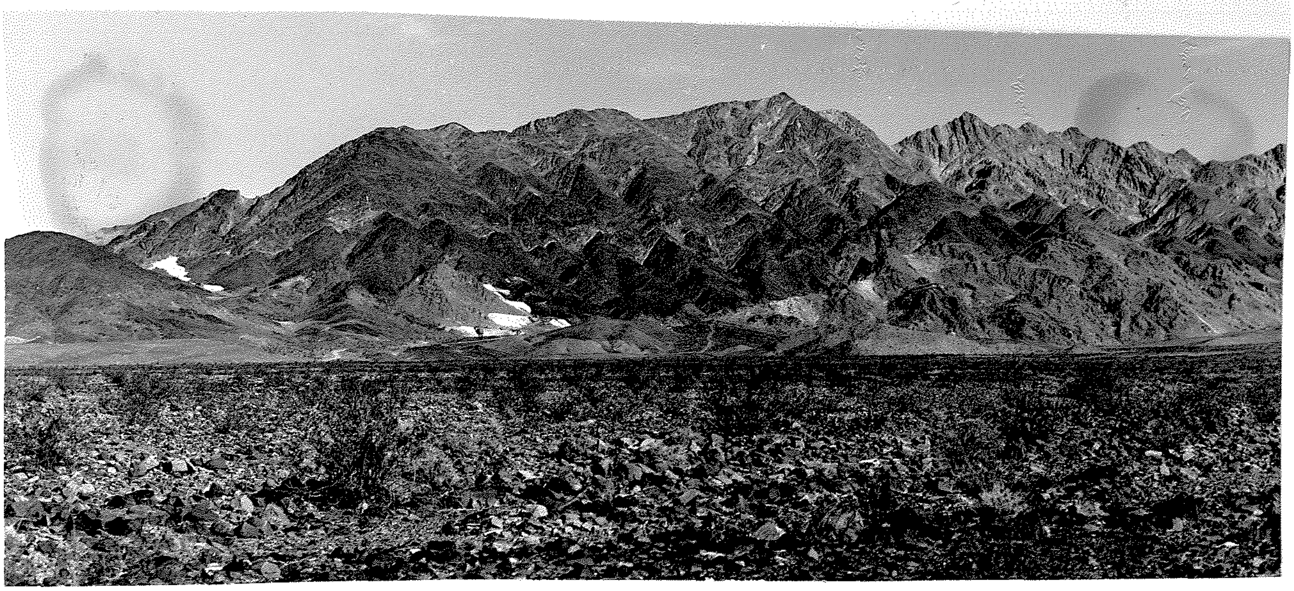

Plate 12a. East face of Ibex Hills, composed of lower part of Crystal Spring formation dipping steeply toward observer. Diabase sill (darkest unit) is overlain by carbonate member (lower slopes and underlain by quartzite and shaly members (higher slopes). Monarch mine at left center; Pleasanton and Ibex mines, far left.

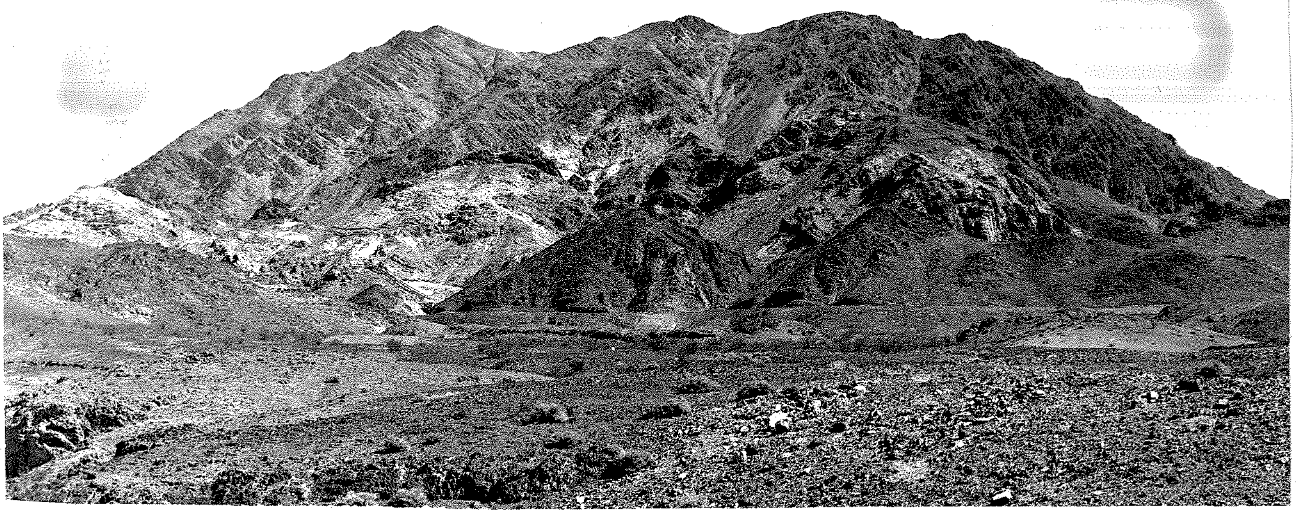

Plate 12b. View northward along the strike of Ibex Hills Crystal Spring section shown in plate 12a. Feldspathic quartzite member exposed on horizon between points 1 and 2, purple shale member between points 2 and 3, and fine-grained quartzite member between points 3 and 4. These lower units are in fault contact with higher units of the formation (mid-foreground) which contain the Ibex deposit (I.).

rocks to be of Cambrian or pre-Cambrian age.

These rocks were described in a general manner by Noble[35] in 1934,

35/ Noble, L. F., Rock formations of Death Valley, California: Science, n. s., vol. 80, no. 2069, pp. 173-178, 1934.

and were briefly mentioned by Hazzard[36] in 1938. Both Noble and Hazzard

36/ Hazzard, J. C., Paleozoic section in the Nopah and Resting Springs Mountains, Inyo County, California: California Jour. Mines and Geology, vol. 33, p. 299, 1938.

assigned to them a late pre-Cambrian age; but the series remained unnamed until 1940, when Hewett[37] proposed the name "Pahrump series".

37/ Hewett, D. F., New formation names to be used in the Kingston Range, Ivanpah quadrangle, California: Washington Acad. Sci. Journ., vol. 30, no. 6, pp. 239-240, 1940.

Hewett recognized a three-fold division consisting, from bottom to top, of the Crystal Spring formation, the Beck Spring dolomite, and the Kingston Peak formation. Representative sections of these formations subsequently have been described in detail by Hewett[38] in an account

38/ Hewett, D. F., Geology and mineral deposits of the Ivanpah quadrangle: U. S. Geol. Survey Prof. Paper, in press.

of the geology of the Ivanpah quadrangle. This quadrangle, however, contains only the narrow, east-trending Pahrump belt in the Kingston Range. In this belt, Hewett noted the presence of diabase sills and talc deposits near the base of the lowermost carbonate beds of the

Crystal Spring formation. He attributed the origin of the talc deposits to an alteration of carbonate beds by diabase.

As noted above, however, most of the exposures of the Pahrump series are west of the Kingston Range in the general region of the Amargosa Valley, Amargosa Range, and southern Death Valley. In 1940, many of the broad stratigraphic and structural features of this region were described by Noble[39] as background for a detailed account of the

39/ Noble, L. F., Structural features of the Virgin Spring area, Death Valley, California: Geol. Soc. America Bull., vol. 52, pp. 841-999, 1941.

geology of the Virgin Spring area in the southern part of the Amargosa Range. Noble also noted that the Crystal Spring formation contains talc deposits as alterations of dolomite at the contacts of diabase intrusions.

A general review of the broad structural features of the region, as conceived by Noble[40], will not be included here. In brief, however,

40/ Noble, L. F., op. cit.

Noble has described the Pahrump occurrences in the Alexander Hills, the southern Nopah Range, Rainbow Mountain, and the central Amargosa Range, as chaos blocks in a gigantic thrust plate that has overridden an autochthanous block composed principally of Archean rocks, but also containing large masses of the Pahrump series. The Pahrump occurrences in the Ibex Hills, the southern Amargosa Range, and the Owlshead Mountains, Noble believes to be part of the autochthanous rock.

The northernmost of the known exposures of the Pahrump series, about 15 miles north of Virgin Spring, have been studied by Curry in unpublished investigations.

Exposures of the Pahrump series at the southern end of the Nopah Range have been briefly mentioned by Mason[41] in 1949. The Silurian

41/ Mason, J. F., Geology of the Tecopa area, southeastern California: Geol. Soc. Am. Bull., vol. 59, pp. 333-352, 1949.

Hills, which contain the southernmost of the Pahrump occurrences have been studied in detail by Kupfur[42].

42/ Kupfur, D. H., Geology of the Silurian Hills, San Bernardino County, California: Unpublished Ph. D. thesis, Yale University, 1950.

GENERAL FEATURES OF THE PRE-CAMBRIAN ROCKS

Archean rocks

At numerous localities the Pahrump series lies with depositional contact upon a complex of much older and much more highly metamorphosed rocks, generally designated as "earlier pre-Cambrian" or "Archean". The complex consists mostly of mica schist, granite gneiss, and micaceous quartzite. Migmatite, small pegmatite dikes, and veins and pods of milky quartz are common. These rocks form a large part of the pre-Tertiary terrane of the southern Death Valley-Kingston Range region. Weathering more readily than the overlying later pre-Cambrian and Paleozoic rocks, they are most commonly exposed in depressions and lower slopes of the mountain ranges.

Pahrump series

Except for the chaos blocks exposed at Rainbow Mountain and in the central Amargosa Range, the Pahrump rocks have, in general, escaped intensive deformation, but occur in large, uniformly dipping fault blocks in which folds are rare. Sections containing all three formations in essentially complete sequence, are known to exist only in the Kingston Range and in the hills north and east of Saratoga Spring. In the Kingston Range the series is as much as 7000 feet thick; near Saratoga Spring it is more than 5500 feet thick. Nowhere else were relatively complete sections of the Kingston Peak formation observed by the writer. Elsewhere, a nearly complete thickness of the Beck Spring dolomite was noted only in the Alexander Hills.

At many places, however, sections of from 2 to 4 thousand feet of the Crystal Spring formation remain virtually intact. As part of the present investigation, detailed sections were measured at the following localities: (1) Warm Spring Canyon in the southeastern Panamint Range, (2) northern Owlshead Mountains, (3) one-half mile north of Saratoga Springs, (4) the Superior mine 2 1/4 miles north of Saratoga Springs, (5) an area due west of Ibex Spring, (6) the Monarch mine one mile north of Ibex Spring, (7) the Western mine in the Alexander Hills, (8) the Rogers mine on the west face of Kingston Range, (9) the type locality of the Crystal Spring formation in the Kingston Range, and (10) the western part of the Silurian Hills.

The sedimentary part of the Crystal Spring formation is composed of a lower 500 to 2200 feet of predominantly arenaceous sediments with subordinate amounts of argillite and shale, a middle 500 to 1000 feet of carbonate beds and massive chert, and an upper 500 to 1000 or more feet of alternating shale, quartzite and carbonate beds. The section has been thickened as much as 1500 or more feet by the intrusion of diabase. Most of the diabase occurs as large sills in the middle and upper part of the formation. Where exposed in its entirety the formation is generally about 4000 feet thick.

Hewett has shown that the Beck Spring dolomite in the Kingston Range is from 1100 to 1200 feet thick and composed predominantly of massive gray dolomite, with sandy and shaly beds in the top 200 feet. The Beck Spring dolomite in the Alexander Hills, although unmeasured is similar in appearance to the Kingston Range occurrence. In the Saratoga Spring section the dolomite is about 1350 feet thick and more thinly bedded than the other occurrences.

The Kingston Peak formation in the Kingston Range, Hewett found to be 1000 to 2000 feet thick and to consist perdominantly of shaly sandstone with local pebble zones, and a middle third composed of a coarse conglomerate with pebbles mainly of quartzite and dolomite. Near Saratoga Spring the exposed part of this formation consists of a lower 400 feet of shaly sandstone and a higher 1500 feet of coarse conglomerate. Here its uppermost strata are hidden beneath alluvium.

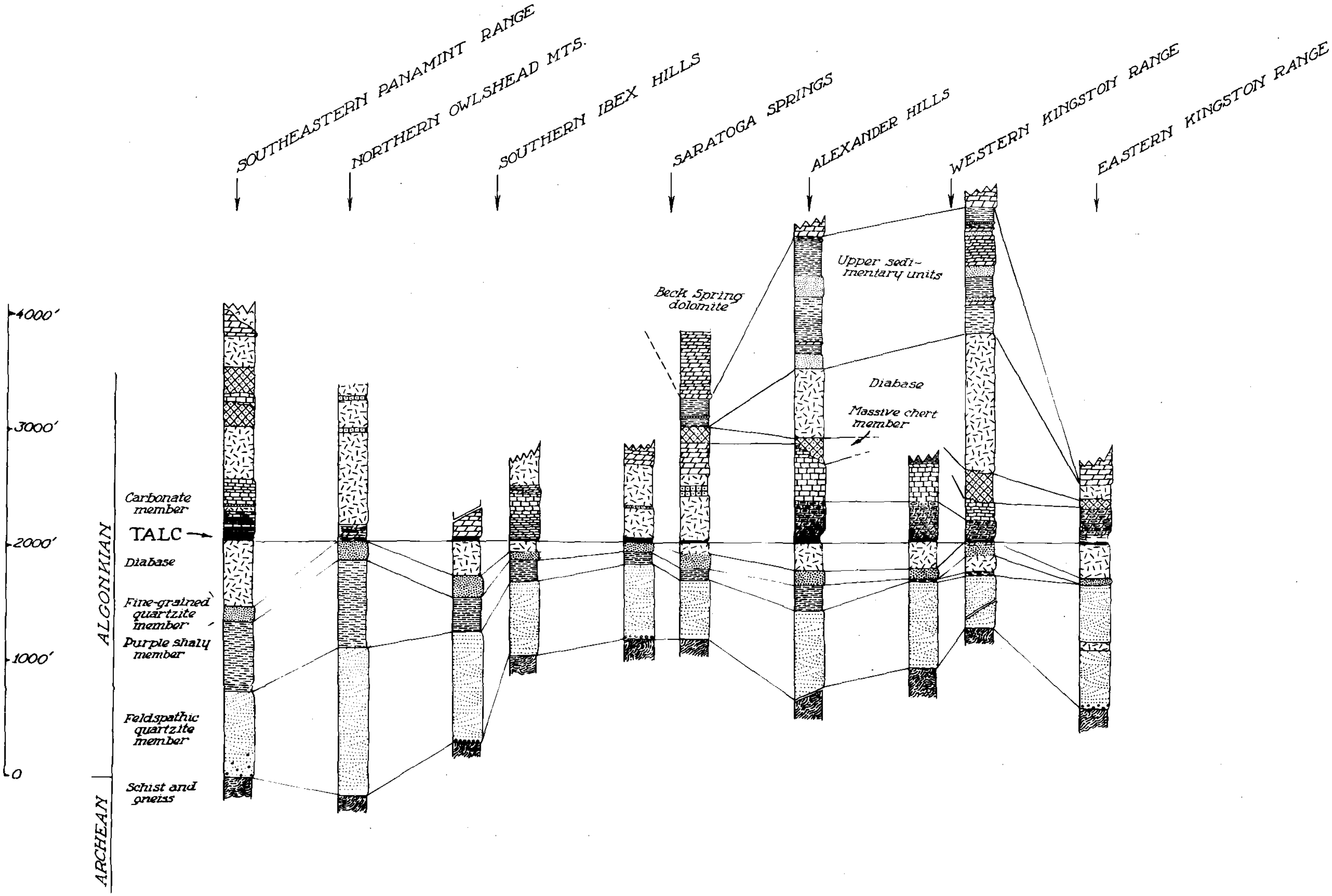

Figure 10. Representative columnar sections of the Crystal Spring formation showing the general restriction of talc mineralization to the lowermost carbonate beds in contact with or near the lower diabase sill.

STRATIGRAPHIC UNITS OF THE CRYSTAL SPRING FORMATION

General statement

Sections of the Crystal Spring formation that apparently are complete, or almost so, occur in the Panamint Range, at Saratoga Spring, and in the Alexander Hills and Crystal Spring areas. The thickest of these sections is at the Warm Spring locality in the Panamint Range where about 4200 feet of sediments are exposed. Here the uppermost of the exposed Crystal Spring strata are overlapped by Tertiary volcanic rocks, and the original section may well have been considerably thicker. The Crystal Spring formation sections measured in the Alexander Hills and at Crystal Spring are respectively 3500 and 3900 feet thick, but several hundred feet of the lower part of each of these sections may be faulted out. The Saratoga Spring section, of which both top and bottom are exposed, and which is intact or nearly so, is only about 2100 feet thick. The sections studied in the northern Owlshead Mountains, the Ibex Spring area, and the western Kingston Range, contain only the lower one-half to two-thirds of the formation.

In most of the occurrences of the Crystal Spring formation, the arenaceous and argillaceous sediments that comprise its lower part are divisible into three units. These will be referred to as "the feldspathic quartzite member", the "purple shale member", and the "fine-grained quartzite member". The carbonate beds in the middle of the formation have a diverse lithology, but are not divisible into units that persist on a region-wide scale. The carbonate beds are, therefore, collectively designated as the "carbonate member". A "massive

Plate 13a. (right) Ripple mark in steeply-dipping, feldspathic quartzite member of the Crystal Spring formation in the Superior mine area. Observer is looking down-section.

Plate 13b. (below) Conglomerate at base of Crystal Spring formation, Superior mine area. Bedding trace slopes gently from upper left to lower right.

chert member" ordinarily overlies the carbonate beds. The sediments that comprise the upper part of the formation, likewise cannot be divided into persistent members, but as a group they will be designated as the "upper sedimentary units".

Feldspathic quartzite member

The most persistent of the Crystal Spring members is the basal quartzite which, in the sections studied, ranges in thickness from about 200 feet in the Silurian Hills to more than 1200 feet in the Owlshead Mountains. It is ordinarily medium to light gray, but in the upper part of the member much of the rock has a light tint of blue or green.

The member has a distinctive, but discontinuous, basal conglomerate (pl. 13b) with a maximum observed thickness of 25 feet and with well-rounded pebbles, cobbles, and boulders as much as 1 1/2 feet in diameter. Much of the conglomerate is composed of the more resistant rocks of the underlying earlier pre-Cambrian complex; but its most abundant and widespread rock type is a dense, vitreous quartzite, not noted in the underlying rocks.

The quartzite above the basal conglomerate is characteristically sub-vitreous and feldspathic. The feldspar, which is commonly present in fractions of one-quarter to one-third, is predominantly microcline. In general, the quartzite grades upward from massive, dark weathering, coarse-grained rock, containing thin conglomerate layers, into more thinly bedded, light-weathering, medium- to fine-grained rock. Some beds high in the member consist of relatively friable sandstone.

Water type cross-laminations are characteristic of the member's entire thickness and ripple marks (pl. 13a) are common in its upper part.

Purple shaly member

A member characterized by a purple to blue color, and composed of shale, argillite and fine-grained quartzite, overlies the basal quartzite member at all of the sections studied by the writer. Its measured thickness ranges from a minimum of 30 feet in the western Kingston Range to a maximum of about 700 feet in the Owlshead Mountains. In the Owlshead Mountains, Panamint Range, and Silurian Hills, the member is composed mostly of fine-grained quartzite. In the other sections, it is chiefly shale.

The shale, and quartzite are ordinarily colored various shades of purple; but some of the shale is blue. Distinctive light green to light blue blotches (pl. 14a) characterize at least part of the member at all of its observed occurrences. Many of the blotches are spheroidal; others are irregular and are commonly elongate parallel with bedding planes. Thin sections show that the blotches are centers about which ferric oxide has been reduced and, in part, removed.

The purple shaly member grades downward into the basal quartzite member. The contact between the two on the accompanying maps and sections is based on a color distinction. Mud cracks are a characteristic feature of the shaly layers.

Fine-grained quartzite member

In all except the Silurian Hills section, the purple shaly member is overlain by a dense, fine-grained quartzite, characteristically green, but locally ranging from gray to light brown and dark red. This member ordinarily has a thickness of from 50 to 100 feet, but is more than 200 feet thick in the Superior mine area. Lenticular carbonate layers from a few inches to several feet thick, though subordinate, are common and increase in abundance upward. These are generally the lowest of the carbonate sediments in the Pahrump series. In a few localities the member also contains subordinate amounts of shale and argillite.

Most of this quartzite is massive, but some is thinly and evenly layered. The member does not show the abundant mud cracks, blotches, ripple marks, and cross-bedding found lower in the section.

Carbonate member

In the sedimentary sequence of the Crystal Spring formation, the lower arenaceous and argillaceous units, comprising the lower third of the formation, are everywhere overlain by a several hundred foot thickness of carbonate sediments. A diabase sill, however, ordinarily separates the carbonate and non-carbonate rocks.

As mentioned above, the carbonate sediments are characterized by marked lithologic variations, but cannot be subdivided into stratigraphic units with a region-wide persistence. In the Kingston Range and Alexander Hills, for example, abundant chert layers (pl. 14a)

Plate 14a. Argillite in the purple shaly member, Superior mine area. Light-colored splotches have formed by reduction of limonitic material. Bedding trace at right angle to pick handle.

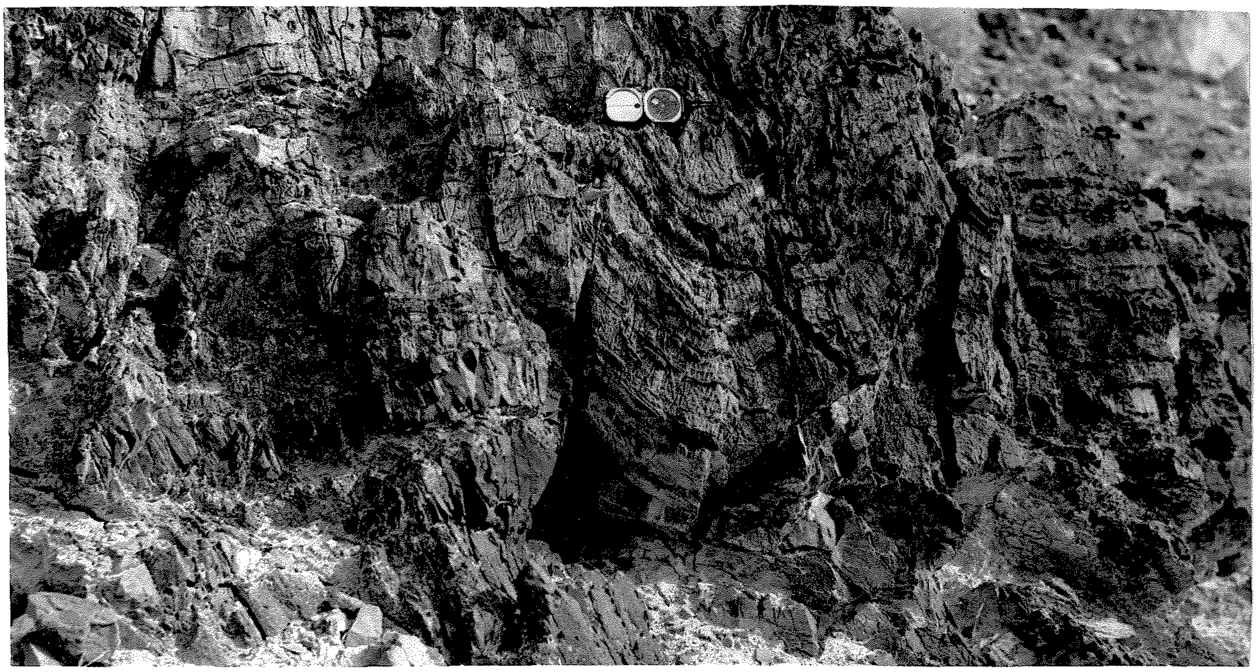

Plate 14b. Cherty dolomite, lower part of carbonate member. Western mine area.

exist in the lower half of the member; exposures of the member, in the Silurian Hills, contain several prominent quartzite beds, each from 15 to 20 feet thick. Yet the carbonate member in the Ibex Spring and Saratoga Spring areas contains very little non-carbonate material.

The character of the carbonate material in the member likewise differs from place to place in the province. The lower half of the member in the western Kingston Range and Alexander Hills is dark-gray and dolomitic, as well as siliceous, and grades upward into light-brown and light gray limestone. In this area the carbonate material is typically fine-grained.

In the Saratoga Spring-Superior mine area, the carbonate member is dolomitic from bottom to top, is pale pink, pale red or light brown on fresh surfaces, and is compact, massive, and fine- to medium-grained.

In the section exposed one mile west of Ibex Spring, the lower part of the member is predominantly a gray, thin-bedded limestone with subordinate siliceous layers. This rock grades upward into silica-free, equally well-bedded limestone. Superimposed upon this limestone are large, irregular dolomite masses, the outlines of which lie athwart the sedimentary layering. Bedding planes, though detectable in the dolomite, are much less distinct than in the limestone. The concentration of MgO, thus indicated by the dolomite masses, probably occurred at a time distinctly later than the sedimentation.

The carbonate member in the Warm Spring Canyon area, unlike its typical occurrences, contains highly tremolitic layers through a thickness of 200 feet in its lower part. The tremolite, mixed with various proportions of fine-grained alkali feldspar and silica, occurs in layers that, in their shapes and disposition, resemble the chert layers common at other localities. The layers are associated with carbonate material that is predominantly calcite. Though not studied in detail, this rock appears to have formed by a low-grade metamorphism of cherty dolomite, a metamorphism unrelated to the diabase intrusives.

The chemical changes indicated by the large dolomite masses in the Ibex Spring area and the tremolitic layers in Warm Spring Canyon, however, are local features. By far the most abundant and wide-spread metamorphic effects shown in the carbonate member are those associated with the diabase sills.

Massive chert member

Massive chert, in thicknesses of as much as 500 feet, lies above the carbonate member in most of the occurrences of the Crystal Spring formation. The chert is ordinarily colored pale-brown to blackish-red, and commonly has a jasperoid appearance. Disseminated chlorite shreds and opaque particles, both in appreciable proportions, are the only other typical constituents. Although massive, the chert commonly shows a distinct, thin, color banding.

The chert ordinarily separates diabase bodies, high in the Crystal Spring formation, from shales of the upper part of the carbonate member. In the Alexander Hills the chert bodies lie beneath a diabase sill and are locally enclosed by diabase. The larger of these thicken and thin markedly, and their lower contacts irregularly transgress strata of the carbonate member. At Crystal Spring and in Warm Spring Canyon chert bodies 100 to 300 feet thick appear to grade laterally and vertically into shales and fine-grained sandstone.

Because the chert bodies are ordinarily in contact with diabase sills, the suggestion is strong that at least some of the chert is genetically related to the diabase and has formed by silicification of both carbonate and non-carbonate strata. In the area of the Superior mine north of Saratoga Spring, however, a very continuous chert layer as much as 150 feet thick is separated from the nearest large diabase body by the underlying carbonate member. The persistent occurrence of chert in essentially the same stratigraphic position, even when unassociated with diabase, and regardless of the amount and position of diabase elsewhere in the section, points toward a sedimentary origin.

Upper sedimentary units

Crystal Spring sedimentary rocks that overlie the massive chert member are much less abundantly exposed than the other units of the formation. In the eastern part of the Kingston Range, Hewett[43] has shown that the Beck Spring dolomite rests directly upon the chert. In many other places the upper sedimentary units have been eroded away or are hidden beneath Quaternary deposits.

If the chert has partly or wholly formed at the expense of sedimentary rocks the original nature of the upper sedimentary units has been somewhat obscured. In the Crystal Spring, Alexander Hills, and Saratoga Spring sections, however, the writer noted well-exposed

43/ Hewett, D. F., op. cit.

sediments between the top of the massive chert member and the base of the Beck Spring dolomite. These sediments range in thickness from about 250 feet in the Saratoga Spring section to more than 1100 feet in the Alexander Hills. This, the upper part of each of the three sections, is composed predominantly of shale, but also contains subordinate amounts of quartzite and dolomite.

Most of the shale is a greenish to brownish gray, but various shades of red and brown are also common. Most is thinly layered, but is also compact and breaks into slabs many times thicker than the individual strata. A characteristic shale in the sections at Saratoga Spring and the Superior mine is grayish orange, strikingly mottled with irregular patterns of iron oxide stain, and containing numerous manganese oxide veinlets.

The quartzite is characteristically fine-grained, massive and compact. Its most common colors are yellowish gray, grayish green, and grayish red. Most of the dolomite is gray, and is interbedded with the shale in layers five feet thick or less. Near the top of the section at Crystal Spring, however, is a 300-foot thick unit composed chiefly of thin-bedded yellowish to medium gray dolomite.

DIABASE

Distribution

Virtually all of the diabase in the Pahrump series is confined to sill-like bodies that lie immediately below, within, or immediately above the carbonate member of the Crystal Spring formation. Small diabase dikes, transecting the sediments of the lower arenaceous and argillaceous part of the formation, were noted at several places; but these contain an insignificant volume compared with the total volume of bodies higher in the formation. At two localities, one in Warm Spring Canyon, the other at the Montgomery mine, the writer observed diabase dikes, more than one hundred feet thick, cutting the carbonate member.

The position most persistently occupied by diabase is at or near the base of the carbonate member. At every locality known to the writer, where this part of the formation is exposed, the diabase is present. Everywhere it forms a sill that is traceable laterally for the length of the exposures of the enclosing sedimentary rocks. These diabase occurrences, therefore, are considered to be parts of a single sill, originally at least as extensive as the 70- by 15-mile belt containing the diabase exposures. This diabase body, here referred to as the "lower sill", ranges in thickness from about 50 feet in the section west of Ibex Spring to more than 1500 feet in Owlshead Mountain section. In the average section, the sill is about 250 feet thick. Its lateral and vertical dimensions place it among the world's largest diabase bodies, comparable in size with the Palisade sill of New York and New Jersey, and with the great Whin sill of northern England.

That it is an intrusive, rather than an extrusive body, is shown by the contact metamorphic effects in the overlying rocks and by continuous, fine-grained selvages, several feet thick, along its upper as well as its lower margins. In its thicker occurrences the sill is ordinarily a multiple body. This is shown by very elongate septa or screens of altered sedimentary rocks, within the sill. These septa occur at various levels, and are themselves bordered, top and bottom, by fine-grained diabase selvages. Internal selvages also commonly exist apart from the septa.

Diabase bodies higher in the formation are commonly as thick or thicker than the lower sill, but are not as persistent. The largest of these ordinarily lie at the top of the carbonate member. In the section at Crystal Spring at the top of the carbonate 1500 feet of diabase separate this member from the upper sedimentary units. In the eastern part of the Kingston Range according to Hewett[44], a diabase body, 120

44/ Hewett, D. F., op. cit.

feet thick, overlies the carbonate member and is overlain by the Beck Spring dolomite. Diabase bodies from 500 to more than 1000 feet thick, also occur high in the Crystal Spring formation in the Alexander Hills, the Ibex Hills and Warm Spring Canyon. The upper diabase bodies are missing in the sections at the Saratoga Spring and Superior mine areas, but are present in both the Warm Spring Canyon and Ibex Spring sections. Thin diabase sills, from a foot or two to several tens of feet thick, within the carbonate member were noted throughout the province. Like the lower sill, the sills within and above the carbonate member have top, bottom, and internal selvages and commonly contain septa.

Previous workers in the region, in noting that the diabase does not intrude the Paleozoic rocks, have assigned it a pre-Cambrian age. The diabase appears to be further restricted to the Crystal Spring formation, as it has not been noted intrusive into the Beck Spring dolomite or Kingston Peak formation. Whether or not an unconformity separates the Beck Spring dolomite from the Crystal Spring formation and the diabase sills, cannot now be stated. The suggestion is strong, however, that the diabase was intruded before the deposition of the Beck Spring dolomite. If so, the intrusion probably occurred at relatively shallow depths and at a time when the Crystal Spring sediments were poorly indurated.

Character of the diabase

Petrology and petrography

The diabase appears originally to have been composed of from 30 to 60 percent plagioclase (mostly labradorite), from 30 to 60 percent mafic minerals (principally hypersthene and augite) and from 2 to 10 percent magnetite and ilmenite; but considerably more than half of the volume of the specimens examined in thin section consists of the secondary minerals uralite, chlorite, sericite and clinozoisite (?). The widespread and pervasive character of the alteration indicates a deuteric origin rather than a hydrothermal effect genetically unrelated to the diabase.

Preliminary megascopic and microscopic examinations of the diabase show a marked mineralogic homogeneity persisting both vertically and laterally. In a few places unusually high concentrations of feldspar were noted; but such concentrations seem to be erratic in distribution and not attributable to a gravitational differentiation.

The diabase generally has a dark-green to greenish-black color. It is mostly medium-grained, but coarse-grained facies are common in the central parts of some of the larger bodies; the selvages are typically fine-grained. An ophitic texture is characteristic, but porphyritic textures with phenocrysts of plagioclase and hypersthene were noted locally in the selvages. In a very few localities some of the coarser diabase contains abundant irregularly angular phenocrysts of plagioclase.

The plagioclase (mostly calcic labradorite), of the ordinary Crystal Spring diabase is in laths, ranging generally from one-half mm. to two mm. long. It has altered principally to sericite and clinozoisite (?), but irregular inclusions of chlorite are common. The plagioclase phenocrysts of the porphyritic rock appear to be sodic labradorite.

Augite, mostly in grains less than 1 mm. in diameter, generally forms 5 percent or less of the rock's volume, and appears to be a remnant of the primary mafic fraction. Uralite, by far the most abundant dark constituent, is partly, perhaps wholly, an alteration of augite. Most of the uralite, however, is unassociated with remnants of earlier mafic minerals. It occurs in irregular grains from 2.5 mm. to 10 mm. in diameter, and also forms felty aggregates. The uralite shows all degrees of chloritization, an alteration particularly well-displayed along cleavage fractures in the larger uralite grains.

The hypersthene, rarely present in proportions greater than one or two percent, is mostly in subhedral grains less than 0.5 mm. in maximum dimension. Unlike the other primary constituents of the diabase, the hypersthene is little altered.

The opaque grains, irregularly angular in outline, are as much as 3 mm. long. Magnetite is the most abundant, but ilmenite is commonly intergrown with it.

Biotite shreds, thinly scattered through most of the thin sections, may be remnants of primary grains. The biotite is partly chloritized. Quartz is uncommon, but locally exists as small interstitial anhedral grains. Apatite is an abundant accessory; sphene is less common.

Chemical composition

As shown in Table 3, a specimen typical of the lower sill, when compared with diabases in general, shows several noteworthy differences in chemical composition. The specimen contains about 7 percent less silica, 3 percent more FeO, and has a distinctly higher Na_2O to CaO ratio. Although the average H_2O figure is not tabulated, other diabases ordinarily contain about 1.5 percent H_2O, or about half as much as the lower sill specimen. Although only one specimen was analysed chemically, thin section studies of this and about 15 other specimens of the lower sill indicate these major chemical differences to be persistent.

The relatively low SiO_2 and high FeO contents largely reflect the abundance of chlorite, formed at the expense of more siliceous and less ferruginous aluminum-bearing silicates. Silica released in such an alteration could well have assisted in the silication of the bordering carbonate rocks. The high H_2O fraction is attributable to the abundance of hydroxyl-bearing alteration minerals.

METAMORPHISM

REGIONAL METAMORPHISM

As previously recognized by both Noble[45] and Hewett[46], the rocks

45/ Noble, L. F., op. cit., p. 950.

46/ Hewett, D. F., op. cit.

of the Pahrump series record a relatively slight regional metamorphism, and are not perceptibly more indurated than the Paleozoic sediments in the region. The carbonate rocks, although crystalline, are mostly fine-grained. In general, temperatures have not been high enough to cause cherty layers to react with the enclosing dolomite. Only the wide-spread tremolitic layers in Warm Spring Canyon area appear to have formed in this manner, unrelated to the diabase intrusion.

Throughout the district the most arenaceous of the Crystal Spring sediments have become quartzites, but sand grains in many of the less pure layers retain their clastic outlines. The maximum degree of metamorphism of the shaly layers is shown by the transformation of some of them to argillites. Although in a few localities the shales and argillites show a scondary schistosity, most of the planar structures throughout the Pahrump section represent bedding. The remarkably good preservation of cross-bedding, ripple marks and mud cracks also indicates a regional metamorphism of low intensity.

	1	2
SiO_2	43.79	51.45
Al_2O_3	16.92	15.64
FeO	10.75	7.93
Fe_2O_3	3.80	3.91
TiO_2	2.87	1.48
MnO	0.35	0.21
C_2O	6.23	9.11
MgO	6.04	5.90
K_2O	1.47	0.98
Na_2O	3.26	3.13
H_2O -150° c	0.23	*
H_2O $+150^\circ$ c	3.62	*
SO_2	Nil	*
P_2O_5	0.38	.26
	99.71	100.00

Table 3. Analysis of a specimen typical of diabase in the lower sill compared with average composition of world-wide diabases.

1. Diasbase near center of lower sill in Western mine area. Analysis by W. H. Herdsman, Glasgow, Scotland.

2. Average composition of 90 diabase specimens (world-wide), given by R. A. Daly in Igneous Rocks and the Depths of the Earth, McGraw-Hill, 1933, p. 406.

*Analyses calculated to water-free.

SILICATED ZONES

Distribution and size

In contrast with the relatively mild effects of regional metamorphism, the carbonate member contains pronounced and widespread zones of silication from a few inches to as much as 200 feet thick, and closely related in space to the diabase sills. The strata at or near the base of the carbonate member are the most persistently silicated and are the ones which by alteration have yielded all of the large bodies of commercial talc. These lower zones ordinarily lie above the lower sill, either in contact with it or separated from it by non-carbonate strata. Locally, where the lower sill intruded above the basal carbonate beds, silicated zones lie beneath it.

Silicated zones also border the sills higher in the member and comprise most of the septa in these and in the lower sill as well. Many of the higher diabase-carbonate rock contacts, however, show little silication or none at all. Conversely the carbonate beds above the lower sill show some degree of silication for virtually its entire length. Lower zones, from 20 to 200 feet wide, are commonly traceable for several thousand feet. The septa are characteristically several tens of feet wide and several hundred to several thousand feet long. The widest and most persistent zones are associated with sills more than 100 feet thick; whereas the smaller sills show thin, discontinuous zones, or none at all. However, at some localities, such as the Superior mine area, where the lower sill is thick and multiple in character, it is overlain by a zone of relatively weak silication.

Non-carbonate strata in contact with diabase ordinarily appear unaltered. As noted above, however, much of the massive chert, commonly in bordering the higher diabase bodies, may have formed by contact silicification of shale and carbonate strata.

General features of the silicated rocks

Petrology

All of the silicated rocks are characteristically fine-grained; all contain an abundance of magnesian silicate minerals. They are, however, divisible into rather clearly defined lithologic types distinguished either by mineralogic or textural differences or by both. The name "talc-tremolite rock" is applied to a variety composed principally of magnesian silicate minerals and notably poor in alumina, alkalies and iron. It is white or very pale shades of gray, green, or pink. Its principal minerals are talc, tremolite, chlorite and calcite. These exist in various proportions and combinations. The rock contains traces of feldspar and quartz, and less than one-half percent, by weight of iron oxide. Textures range from blocky to thinly laminated to schistose. Classified as talc-tremolite rock is all of the material of commercial interest, but much is made sub-commercial by a carbonate content in excess of 15 percent.

Other sub-commercial silicated rocks have distinctly darker colors. They ordinarily contain from 10 to 60 percent alkali feldspar (both albite and orthoclase) and from 5 to 25 percent quartz. The most abundant of these rocks is a massive to platy, pale green to pale bluish gray tremolitic type, conveniently called "green tremolitic rock". Its other common constituents are calcite, diopside, chlorite, and opaque material.

A less abundant, but very widespread sub-commercial rock is a crudely foliated, pale yellowish orange to reddish brown type, named "transition rock" for its restriction in occurrence to the borders and outer parts of zones consisting largely of talc-tremolite rock. This rock, like green tremolitic rock, ordinarily contains appreciable fractions of albite, orthoclase, calcite, and quartz. In general it is distinguished from the green rock by a much lower tremolite content, an abundance of sericite (?), the presence of other hydrous, magnesium-rich minerals such as talc, chlorite, serpentine or phlogopite, and a pervasive limonitic stain.

Well-defined planar features, paralleling the margins of the sills and the sedimentary bedding, characterize each of the three rock types both in gross and in textural detail. Such parallelism is shown by the layers and elongate lenses of the various alteration rocks of individual zones, and by the schistose or laminated textures of most of these rocks. Some of the bodies, however, have a schistosity discordant with their margins.

Distribution of rock types

The zones associated with the lower sill, and altered from the basal beds of the carbonate member, contain all but a small part of the district's talc-tremolite rock. Some of these lower zones consist almost entirely of talc-tremolite rock; but most also contain layers and lenses of the more feldspathic alteration rocks. The transition rock is confined almost entirely to the lower zones. It most commonly occurs as layers separating bodies of talc-tremolite rock and unaltered rocks of the carbonate member. It also interfingers with and forms septa in the outer parts of these bodies. Less common in the lower zones, but by no means rare, are lenses of green tremolitic rock.

The silicated zones that border the diabase bodies higher in the carbonate member, and compose septa in these as well as in the lower sill, consist predominantly of green tremolitic rock. Zones consisting largely or wholly of talc-tremolite rock locally border the higher diabase bodies, but such zones are discontinuous and rarely exceed four feet in width.

At the Montgomery mine and at a locality on the north side of Warm Spring Canyon, the spatial relations between the three varieties of silicated rock are especially well shown. At each place are exposed silicated rocks beneath a large diabase dike. Each dike, lying at a moderate angle with the bedding, can be traced for a lateral distance of about 1000 feet and crosses the lower part of the carbonate member through a stratigraphic thickness of approximately 300 feet.

The silication has produced talc-tremolite rock and transition rock in the lower 50 to 100 feet of the member and green tremolitic rock in the higher exposures. Layering in the silicated rock is parallel with the bedding; the altered rocks grading imperceptibly into unaltered dolomitic strata. The alteration seemingly produced no change in volume. Nor do the major compositional differences in the three types of alteration rocks appear to be attributable to variations in the composition of the dolomite.

Talc-tremolite rock

Gross structural features of bodies of talc-tremolite rock

Most of the talc-tremolite rock associated with the lower sill occurs in the silicated zone directly above the sill or separated from the sill by shale or quartzite a few tens of feet in maximum thickness. A relatively small proportion lies beneath the sill or in the altered septa within it.

Included among the talc-tremolite bodies of commercial interest that lie next to the upper margin of the lower sill are those at the Excelsior, Harry Adams, Western (pl. 15b), Acme, Eclipse, Monarch, Pleasanton, Warm Spring and Death Valley mines. The Crystal Spring, Smith, and Ibex deposits are separated from the upper margin of the sill by non-carbonate strata.

The Superior (pl. 11), White Cap, Sheep Creek and Owlshead Mountain deposits are in zones beneath the sill. The Saratoga and Panamint claims contain concentrations of talc-tremolite rock in the septa within the sill.

The Excelsior, Sheep Creek, White Cap, and Warm Spring Canyon (pl. 15a) deposits are the notable examples of commercial talc-tremolite bodies that form parts of much wider alteration zones composed largely of sub-commercial rock (pl. 15a).

In a few places deposits at the base of the carbonate member have localized faulting with displacements of several hundred or more feet.

Lubricated by talc, blocks of the member have moved over Archean rocks or lie with a transgressive contact against other members of the Pahrump series. Examples of such deposits include the Booth deposit and some of the unexploited deposits in the Western mine area (fig. 11). Brecciated country rock within these zones commonly lessens the commercial value of the talc.

Dimensions of commercial bodies

In all of its observed occurrences the silicated zone above the lower sill can be traced laterally on the surface to points where it is faulted off or lies hidden beneath post-Pahrump rocks. Although the zone is composed predominantly of talc-tremolite rock, individual bodies of this material may lens laterally into other silicated rocks, or may be terminated by faulting within the zone. The talc-tremolite rock occurs mostly in bodies with lengths of several hundreds to several thousands of feet, and widths in the range of 5 to 80 feet; but by no means are all of these bodies of commercial interest.

In their average development the talc-tremolite bodies that have been mined are from 10 to 30 feet wide. These have been explored for several hundred feet along-strike to their termination against faults, to places where they are too narrow to be profitably worked, or to places where they lens into non-commercial alteration rocks.

The largest proved continuous zone of marketable talc-tremolite rock, the principal deposit in the Western mine area, is about 5000 feet long and 80 feet in maximum width. The deepest workings in this deposit are about 300 feet below the surface outcrop.

Plate 15a. View of south wall of Warm Spring Canyon showing Big Talc workings (left) and Kennedy workings (right). Talc lies above diabase sill at base of Crystal Spring carbonate member. Another sill caps the ridge.

Plate 15b. View northward along strike of Western talc deposit. On ridge crest the units of the Crystal Spring formation are exposed from left to right in the following order: (a) feldspathic quartzite, (b) purple shale, (c) fine-grained quartzite, (d) thin talc zone, (e) lower diabase sill, (f) thick talc zone, (g) siliceous dolomite, (h) limestone, (i) higher diabase sill.

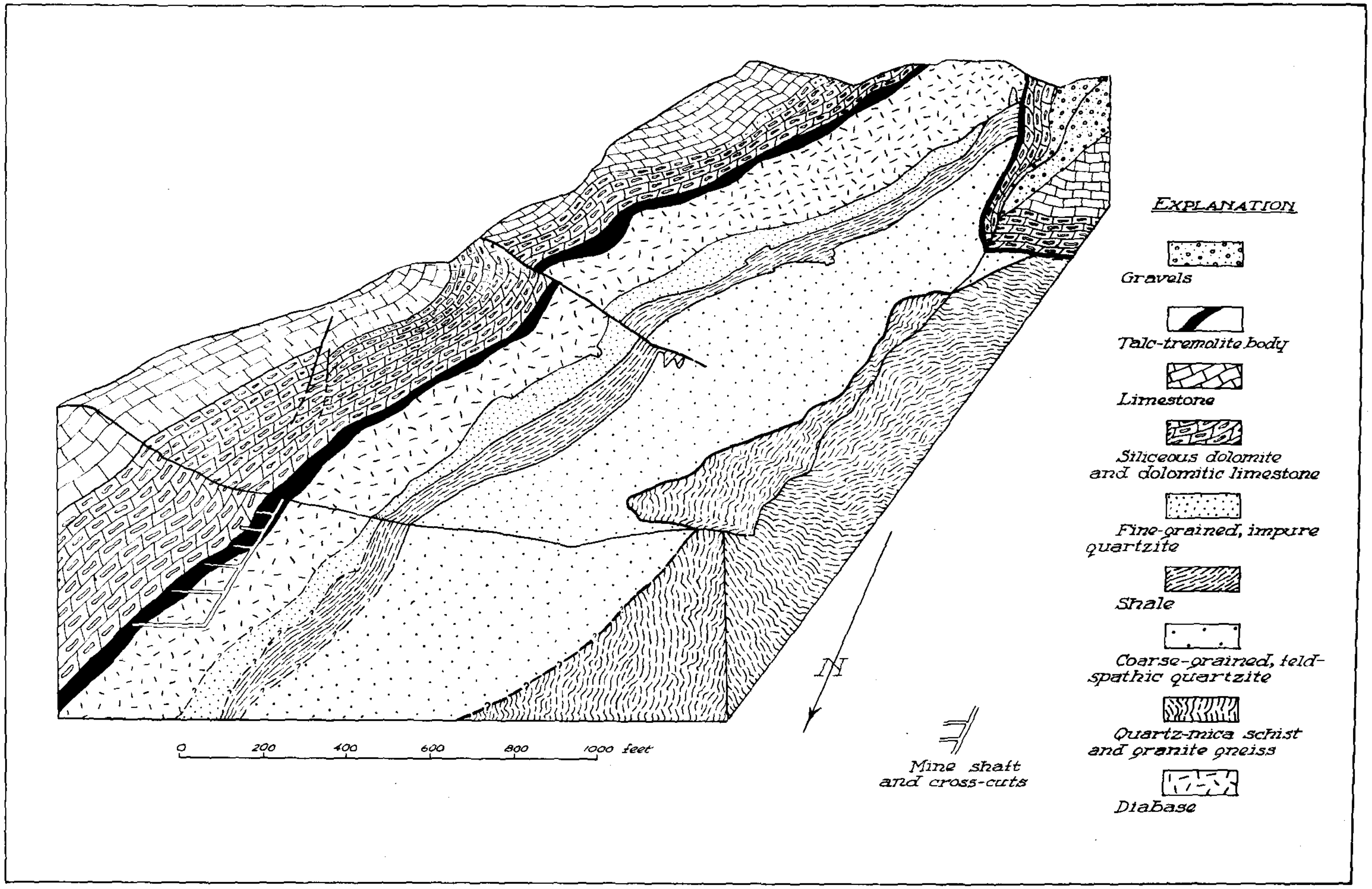

Figure 11. Generalized sketch block-diagram of the central part of the Western talc deposit.

The largest deposit in the Warm Spring Canyon area may prove to be of comparable size, but is poorly exposed at the surface and has much less extensive underground workings. The surface exposure of the longest proved commercial deposit in the Superior mine area is about 1000 feet long. To date none of the district's other commercial deposits have been shown to be more than 1000 feet long, although future development may show that several exceed this length.

Several of the smaller bodies have bottomed within 200 feet of the surface and have been largely worked out. But most of the bodies of commercial interest have not been mined to their down-dip limits.

Internal structure of commercial bodies

Although the zones containing large bodies of talc-tremolite rock have broad similarities, by no means do they everywhere duplicate each other in mineralogy or textures of their contained rock varieties or in the distribution of these varieties within individual bodies. Not only is there a wide range in the proportion and disposition of the sub-commercial alteration rocks associated with the commercial deposits, but the deposits themselves show marked differences in internal structure (fig. 12). Some of the commercial bodies are composed of talc-tremolite rock that is essentially homogeneous from wall to wall, but most consist of two or three layers of talc-tremolite rock distinguished from one another largely by textural differences; but differences in mineral composition are also common.

Textures of the talc-tremolite rock range from thinly-laminated to massive to schistose. The principal mineralogic distinction is the relative abundance of talc and tremolite, one of which generally is present almost to the exclusion of the other; rarely do the two exist in comparable proportions. In either the massive or thinly-laminated rocks, talc or tremolite may predominate, but in the schistose variety, talc appears to be invariably the more abundant. In view of these distinctions, the talc-tremolite rock is separable into the following five varieties: (1) laminated talc-rock, (2) laminated tremolite-rock, (3) massive tremolite-rock, (4) massive talc-rock, and (5) talc schist. Rocks of each of these varieties are currently marketed under the commercial term of "talc". In general, an individual layer contains only one of these varieties, but in a few places the writer noted lateral gradations from one variety to another.

Of the deposits that do not show this layering, some have localized such intense fault movement that any original structure would have been obscured. The relative homogeneity of most of the unlayered deposit, however, appears to be a primary feature.

In most of the layered deposits, the layer next to the diabase consists of the thinly laminated rock in which either talc or tremolite predominates (fig. 12b, c, d). Ordinarily from 5 to 15 feet wide, the laminated layer generally occupies from one-half to one-fifth of the width of the deposit. Upon casual inspection, the laminations may seem to be inherited from a thinly-bedded sedimentary rock. But pointing against such an origin are the seeming rarity of thinly-bedded rock in the unaltered parts of the carbonate member, and lack of any trace in the talc-tremolite rock of even such pronounced pre-alteration features as the chert layers. The close spatial relationship between

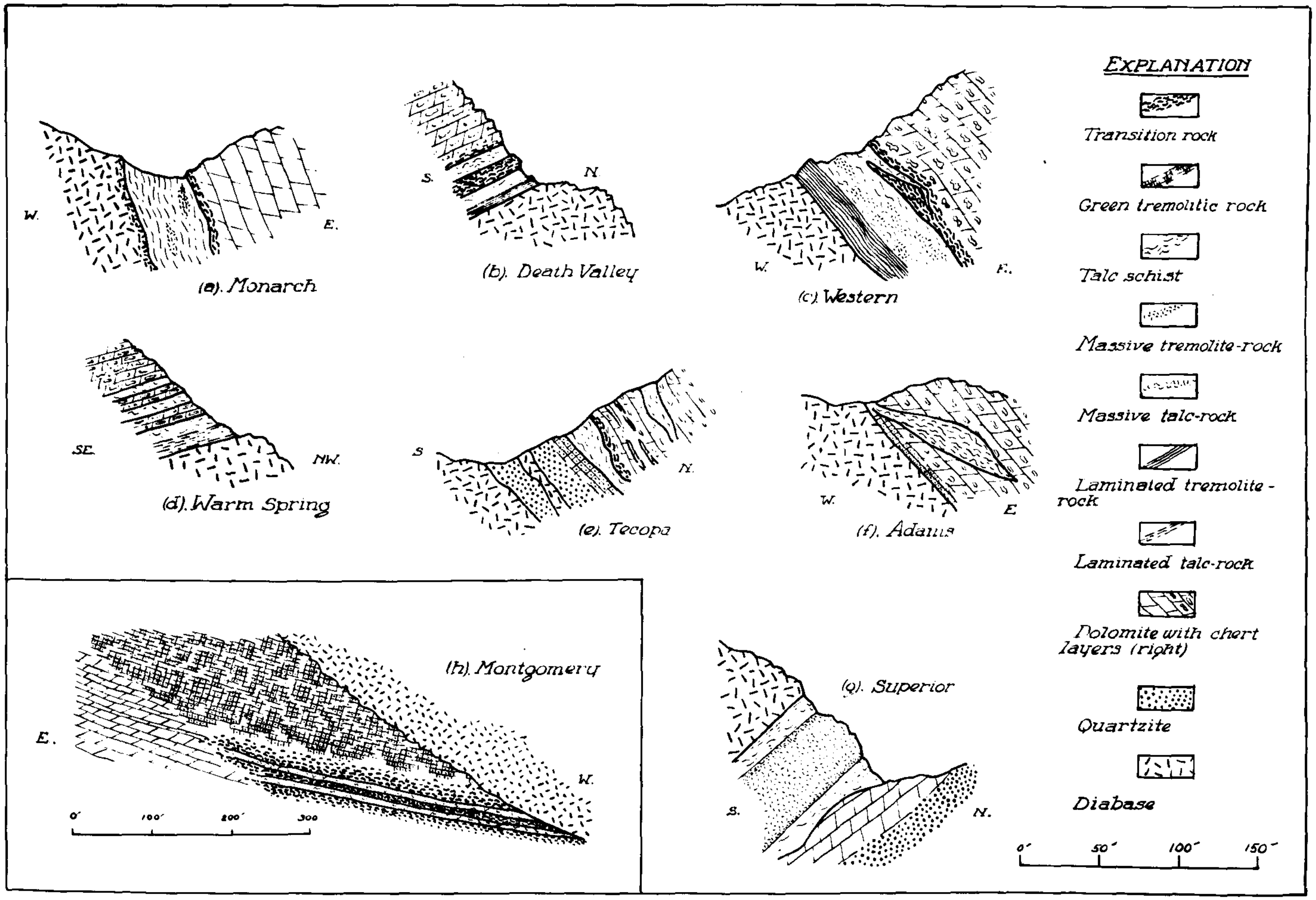

Figure 11. Diagrammatic cross-sections of representative talc deposits in the Southern Death Valley-Kingston Range region.

laminated layers and diabase indicates a metamorphic origin for them. The lamination is perhaps an effect of diffusion adjacent to the diabase margins. Outward from the diabase contact, the laminated layer is ordinarily bordered by a layer of schistose or massive rock containing talc as the predominant silicate mineral. In a few deposits, however, the layer contains a relatively large proportion of massive tremolite rock. A characteristic feature of this layer is a marked pinching and swelling caused, in large part, by faulting localized by the rock's inherent weakness. In several deposits it contains abundant inclusions of country rock, a feature not shown by the laminated layer.

Most of the more tremolitic deposits including the Excelsior, Western (fig. 12c), Acme, and Valley deposits show the two-layered structure for much of their lengths. In these, the thinly laminated layer next to the diabase is highly tremolitic in contrast with the characteristically schistose and talcose outer layer. In such deposits, even though tremolitic masses exist in the outer layer, the outermost several feet are markedly tremolite-poor.

Some of the talc-rich, tremolite-poor deposits also show the two-layered structure. In these, the thinly laminated inner layer, as well as the outer layer, is highly talcose and contains a very subordinate, or even negligible tremolite fraction. The Eclipse deposit and parts of the Warm Spring talc-bearing zone (fig. 12d) are of this type.

The Superior deposit (fig. 12g) is unusual, not only because it is the largest of the known commercial bodies that lie beneath the lower sill, but also because for a distance of about 700 feet it has a three-layered structure. In their surface exposures the inner and outer layers each average more than 15 feet wide, and consist mostly of a fine-grained, talc-rich, tremolite-poor schistose rock. Much of the rock in the layer next to the diabase, like that of the inner layer of the two-ply deposits, is thinly laminated.

The middle layer, which averages about 30 feet thick, is essentially a tremolite-calcite rock. For thicknesses of three or four feet along both margins of the middle layer is a nearly monomineralic tremolite rock of commercial value. In the center of the layer the tremolite is less abundant.

Many of the talc-rich and tremolite-poor deposits contain but one variety of talc-tremolite rock. This is true of such deposits as the Crystal Spring, Tecopa (fig. 12e), Ibex, White Cap, and Owlshead Mountain, which are separated from diabase by non-carbonate, sedimentary strata or by non-commercial alteration rocks. Most of these are composed predominantly of talc schist; locally they contain massive talc rock. Several deposits that are in contact with diabase also are composed mostly of talc schist. These include the Monarch (fig. 12a) and Pleasanton, and parts of the Markley and Brown deposits. The Montgomery deposit (fig. 12h), lying beneath a diabase dike that cuts the carbonate member, is composed mostly of massive talc rock. The Pongo deposit, also in contact with diabase, is composed of laminated rock for its entire 15-foot thickness.

Petrology and petrography

Laminated rocks. The thinly laminated rocks, common in the inner parts of numerous deposits that lie next to diabase, are persistently compact, white and fine-grained. Both the tremolitic and talcose varieties ordinarily contain from 5 to 15 percent of carbonate material. One or more minerals of the chlorite group, the only other relatively abundant silicates, rarely form more than a few percent of the rock, and appear to be absent in a large proportion of the thin sections examined.

Individual laminations have uniform widths, mostly within the 0.1 to 2.0 mm. range. Parting planes are very flat and are commonly traceable for many feet. The laminations are well-shown on weathered surfaces of the tremolitic rock (pl. 16a), where some of the laminae are accentuated by a brownish stain. Weathered specimens of the tremolitic rock characteristically spall into paper-thin layers; but, where unweathered, both the talcose and tremolitic varieties break into relatively thick, platy slabs.

Thin section studies show that the laminations are caused mainly by vertical variations in grain size. Some of the layers contain concentrations of carbonate grains or opaque particles. Both talc and tremolite occur in grains that range in outline from stubby and shred-like to acicular. Most are between 0.01 mm. and 0.2 mm. long; grains from 0.2 mm. to 2.0 mm. long commonly compose individual laminations. The similarity in habit between the talc and tremolite suggests that one is pseudomorphic after the other. The intermediate stages of replacement were not observed, but a tremolite-to-talc alteration noted elsewhere in the deposits suggests that, should such pseudomorphism exist, talc is the later.

Most of the talc and tremolite grains occur, with no apparent preferred dimensional orientation, in mesh-like aggregates; but some of the most pronounced laminations contain relatively large grains that lie normal to the foliation planes. Laminations, containing talc grains dimensionally aligned with the planes, are not uncommon.

Carbonate material (probably calcite with subordinate dolomite) forms from about 2 percent to as much as 10 percent of the volume of the laminated rocks observed in thin section. The carbonate grains range in outline from equant to markedly elongate. Most have maximum dimensions in the 0.02 to 0.4 mm. range. Some of the specimens contain grains that are evenly disseminated through the rock, either singly or in mosaic-like clusters. Others contain disseminated grains that are most abundant along individual laminations. Much of the carbonate material, however, occurs in minute veinlets and lenticles, many of which are also quartz-bearing. Most of these lie parallel with the laminations, but others are cross-cutting features. The disseminated carbonate grains are ordinarily corroded or transected by aggregates of talc or tremolite. The veinlets appear to post-date talc and tremolite.

The chlorite seems to be less abundant in the laminated varieties than in the schistose or massive varieties of talc-tremolite rock. But in a few thin sections it is present in amounts of as much as 20 percent. Some specimens show a relatively even distribution of chlorite grains; in others the mineral is largely concentrated along individual laminations. The chlorite grains show two habits, one plumose or feathery, and another very irregular to equant. Maximum grain dimensions in the range of 0.05 mm. to 0.5 mm. are characteristic of both habits. Both have formed at least partly at the expense of tremolite and carbonate and have been partly replaced by talc.

Some of the specimens are abundantly clouded with pale brown, submicroscopic material. This material also appears to vary in amount from lamination to lamination. Other constituents were noted in proportions less than one percent. Opaque particles, consistently less than 0.01 mm. in diameter, are ubiquitous. Although present only in traces, they are much more abundant in some specimens than in others. Quartz, present in traces, occurs only in lenticles or late-stage veinlets, and is commonly associated with carbonate material. In most specimens quartz was not observed. Sphene was noted in very sparsely scattered grains. Although feldspar was noted in none of the thin sections, should it occur in fine grains and in traces, it easily could have escaped detection.

Massive rocks. The massive varieties of talc-tremolite rock do not show the thin and even layering nor the close spatial relation to diabase contacts characteristic of the laminated varieties. Neither are they composed predominantly of mineral grains in dimensional alignment as are the schistose varieties. The massive rocks break into tough, blocky fragments, rather than into plates or friable, schistose slivers. But they commonly do exhibit a crude planar structure featured by lenticles and layers, in subparallel arrangement, that contrast with each other in texture and mineral content.

In thin section the silicate grains are shown to be ordinarily in mesh-like aggregates or in radial clusters, but relatively common are layers in which grains lie roughly normal to the planar structure. As in the laminated rocks, the carbonate material is disseminated in single grains or in grain clusters, or occurs in minute veinlets, variously oriented with respect to the planar structure. In specimens intermediate between massive and schistose, most of the talc shows a dimensional parallelism, and most of the tremolite lies athwart the schistosity.

Microscopic textural patterns typical of the massive rock are shown in the following examples: (1) very fine-grained, decussate aggregates of talc, of tremolite, or of a mixture of both; (2) aggregates similar to the above in grain size and mineralogy but composed largely of radiating grain clusters; (3) matrices that resemble 1 or 2, but containing knots or elongate lenses of distinctly coarser grained tremolite, or interfingering with irregularly-shaped aggregates of chlorite; (4) tremolite (?) in cloudy, brownish masses embayed and traversed by talc-chlorite veinlets, and (5) brownish masses composed mostly of radiating serpentine blades, and occurring in a much finer-grained talc-tremolite-chlorite matrix.

Except for the differences in fabric, most of the microscopic features of the massive rocks are similar to those of the laminated rocks. This is true of the habit and grain sizes of the talc, serpentine, carbonate, and much of the tremolite. The occurrence and proportions of such minor constituents as the opaque grains, quartz, and sphene are likewise comparable. In individual specimens of the massive rock, however, the average tremolite grain is generally several times longer than the average talc grain. As in the laminated rocks, most of the disseminated carbonate grains or grain clusters are corroded or cross-cut by aggregates of talc or tremolite, which are in turn cut by later carbonate veinlets. A few of the tremolite grains are themselves partly replaced by carbonate. Textures that clearly show age relations between the silicate minerals are not abundant. Talc locally appears to be pseudomorphic after tremolite; and to have replaced tremolite in fine-grained aggregates. That only a small proportion of the talc is similarly pseudomorphic, is suggested by the contrast in average grain sizes of the two minerals. In the massive rocks also, at least part of the chlorite has replaced tremolite, and part of the talc has replaced chlorite.

Schistose rocks. The schistose variety of talc-tremolite rock differs from the laminated and massive varieties mainly in that most of its contained talc grains are in dimensional alignment, indicating formation under stress . Also peculiar to the schistose rock is the persistence of a very high talc to tremolite ratio. In most of the schistose specimens examined in thin section, tremolite was missing altogether. Where present, the tremolite blades ordinarily lie in various positions transverse to the schistosity and appear to be residual from a pre-schist fabric. These features cause most of the schistose rock to be soft and fissile; whereas the other varieties are characteristically tough and platy or blocky.

The schistose rock, however, does share many of the features of the laminated or massive varieties. Like them, it is white or nearly so, and is generally fine-grained. Its principal constituents are talc, tremolite, chlorite and carbonate; but, of these, only talc was observed in all of the thin sections. It also contains minor to very minor proportions of the opaque grains, quartz, sphene, and the sub-microscopic, cloudy material previously noted in comparable proportions in the other talc-tremolite rock varieties.

Talc in some large masses of schist occurs in pearly, curved, micaceous flakes as much as one cm. in diameter; but in its average development the talc is much finer-grained, and the schist is chalky in appearance. Thin section studies show that the finer-grained talc occurs in grains, mostly less than 0.2 mm. long. In aggregate these grains extinguish nearly simultaneously over large parts of single thin sections, and give the appearance of much larger individual grains.

Carbonate material composes at least 2 percent of most of the thin sections examined, and as much as 50 or 60 percent of some. As in the other talc-tremolite rock varieties, the diameters of individual carbonate grains are generally in the 0.02 to 0.4 mm. range, and the grains are distributed singly, in clusters, or in veinlets. Most of the clusters are markedly elongate parallel with the schistosity. The veinlets too generally are parallel with the schistosity, but some lie at high angles across it. The disseminated grains and clusters are commonly corroded and traversed by talc aggregates; the veinlets appear to post-date the talc.

The tremolite in the schist occurs as irregularly scattered blades, either individually or in clusters. The blades ordinarily are several times longer than the aligned talc shreds with which they are associated.

The chlorite, like the tremolite, is rare or absent in most of the thin sections examined; but it comprises as much as 20 percent of some. It occurs in irregular knots and lenticles, that appear to be corroded by talc and are commonly cross-cut by streaks of aligned talc shreds.

Green tremolitic rock

Distribution

Unlike the talc-tremolite rock, the large bodies of which are restricted to alteration zones associated with the lower diabase sill, the green tremolitic rock occurs in alteration zones along the higher sills as well. It also composes most of the material in the septa within the multiple sills. In some localities, including the Excelsior, Superior, and Warm Spring mine areas, a relatively large proportion of the contact zone bordering the lower sill is of the green tremolitic rock; but in most of its occurrences the lower sill is associated with little or none of this rock. Conversely, almost all of the silication associated with the higher diabase bodies has yielded green tremolitic rock.

Where the green tremolitic rock and talc-tremolite rock exist in the same alteration zone, the two generally form layers from several inches to several feet thick, with contacts ranging from very sharp to gradational through several inches. These layers, like those within the individual talc-tremolite bodies, are essentially parallelwith the bedding of the nearest unaltered sedimentary rocks.

The contacts between green tremolitic rock and unaltered carbonate strata also range from sharp to gradational through several inches. Most of these contacts also parallel the sill margins and the bedding, but locally the alteration to green tremolitic rock appears to have been guided by fractures at high angles to the sill margin. These replacement veins extend into carbonate strata that upon megascopic inspection seem to contain little, if any, original argillaceous, or siliceous material.

Petrology and petrography

The green tremolitic rock is compact, fine-grained, and characteristically breaks into evenly-layered slabs from a fraction of an inch to several inches thick. The pale green color is most typical, but the variations to pale bluish gray are quite common. The rock appears persistently to maintain these general megascopic features regardless of the proportions of such other constituents as alkali feldspar, carbonate, quartz, diopside, or opaque particles.

Each of ten thin sections of representative specimens of green tremolitic rock contains an appreciable, but variable, proportion of tremolite. Locally tremolite comprises almost all of the rock. Alkali feldspar (albite and orthoclase) generally is present in fractions of from one-fourth to three-fourths.

Each section also contains from a trace to several percent of finely divided, opaque and semi-opaque material. Carbonate grains form as much as 30 percent of the sections; but in two sections no carbonate was noted. Diopside occurs in only two sections, but forms about 20 percent of the rock in these. Quartz, though difficult to distinguish in thin sections from untwinned feldspar, probably constitutes no more than 5 percent of any section. Relatively rare constituents are actinolite and chlorite. The latter appears to have replaced tremolite.

Silicate and quartz grains generally are from 0.01 mm. to 0.2 mm. in diameter. The carbonate grains range from about 0.01 mm. to about 2.0 mm. in diameter. The opaque and semi-opaque particles rarely exceed 0.01 mm. in diameter.

Each section shows a relatively homogeneous texture featured by an even distribution of tremolite needles. Some of the needles are in jack-straw arrangement; others are in radial patterns extending outward from common centers, or inward from the margins of feldspar or quartz grains. The feldspar and quartz are uniformly disseminated through the rock, or, where in excess of tremolite, form an even-grained mosaic.

The carbonate grains are disseminated and are very irregular in outline, an effect of their corrosion and transection by aggregates of tremolite, feldspar, and quartz. These grains appear to be remnants of an original crystalline carbonate rock. The diopside, as observed in thin section, is largely confined to irregular masses and shows no well-defined replacement textures with the other constituents.

The opaque and semi-opaque grains are rather evenly distributed through the rock, but commonly cluster in clots interstitial to grains of the other minerals.

Chemical analyses of two representative specimens of green tremolitic rock (table 4), compared with analyses of talc-tremolite rock show that the former are distinctly more aluminous and potassic and

	1	2	3	4	5	6	7	8	9	10	11	12
SiO_2	51.09	49.77	51.30	56.42	55.90	56.62	54.77	56.88	35.12	60.90	57.75	62.85
Al_2O_3	0.69	1.84	1.08	0.58	1.24	1.14	1.08	0.45	9.22	8.85	11.54	14.60
Fe_2O_3 } FeO }	0.33	0.16	0.88	0.28	0.31	0.18	0.38	0.63	1.11	1.91	2.22	1.43
MgO	25.36	28.71	29.52	28.11	28.77	27.75	26.05	31.29	12.24	9.88	6.02	6.12
CaO	9.32	9.20	6.20	5.63	4.77	5.91	5.81	4.92	13.95	5.23	7.48	1.32
Na_2O	2.18	1.80	0.16	0.12	0.10	1.13	0.27		1.86	1.46	2.30	2.16
K_2O	1.70	0.61	0.05	0.02	0.44	0.44	0.21		4.27	4.66	7.67	10.37
H_2O+	2.84			4.12	4.40	3.94	4.58		0.95	2.12	3.18	1.17
H_2O-	0.11	7.90	10.65	0.09	0.10	0.10	0.23	5.53	0.14	0.16	0.17	0.17
CO_2	5.21			4.46	4.35	4.40	4.56		21.00	5.30	2.98	0.22
Totals	98.83	99.99	99.84	99.83	100.38	101.61	97.94	99.70	96.86	100.47	101.21	100.41

1. Laminated tremolite rock from Western deposit.
2. Laminated tremolite rock from New Acme deposit.
3. Laminated talc rock from Pongo deposit.
4. Massive talc rock from Montgomery deposit.
5. Talc schist from Tecopa deposit.
6. Massive talc rock from Warm Spring deposit.
7. Talc schist from Ibex deposit.
8. Talc schist from Superior deposit.
9. Transition rock from Western deposit.
10. Transition rock from Tecopa deposit.
11. Green tremolitic rock from Warm Spring mine area.
12. Green tremolitic rock from Warm Spring mine area.

Table 4. Chemical analyses of commercial talcs and other silicated rocks of the Crystal Spring formation. Analyses 2, 3, and 8 furnished by Southern California Minerals Company, analyst undetermined. All others furnished by Sierra Talc and Clay Company, analyst A. J. McArthur.

less magnesian. Other less marked, but probably persistent differences, are higher iron, soda and water percentages in the green tremolitic rock.

Transition rock

Distribution

Transition rock, in its characteristic position at or near the outer margins of the larger bodies of talc-tremolite rock and commonly separating them from unaltered carbonate rocks, is widespread in the silicated zones low in the carbonate member. Among the deposits showing well-defined layers of transition rock are those at the Rogers, Western, Acme, Ibex, and Pleasanton mines. Layers of the rock range in thickness from a few inches to as much as 12 feet. Their contacts with talc-tremolite rock are very sharp, whereas contacts with carbonate strata are gradational.

The transition rock is particularly well developed at the Western mine where it forms an outer border with an average thickness of nearly 10 feet. Bodies of the rock as much as eight feet thick and several tens of feet long interfinger with talc schist and form elongate inclusions within the schist. Veinlets, as much as a foot wide, composed of white talc-schist commonly extend into the transition rock. Some of the inclusions are rimmed with an irregular and discontinuous border of light gray calcite as much as one foot thick. The calcite rims are persistently thickest along the bottoms of the inclusions.

Petrology and petrography

The transition rock is fine-grained and ranges from very hard and tough to punky. It has a typical mottled appearance and is colored from moderate orange pink to pale brown, with the latter by far the more common. Characteristic is a wispy foliation marked by thin, discontinuous, wavy laminations. The laminations, although parallel with planar features in nearby sedimentary rocks, do not appear to be relict bedding.

As previously noted, the transition rock, like the other silicated rocks, is rich in carbonate and hydrous magnesian silicates. Unlike talc-tremolite rock it generally also contains appreciable fractions of quartz, alkali feldspar, and probably sericite. Unlike the green tremolitic rock, it is generally poor in tremolite and rich in one or more non-calcic hydrous magnesian silicate minerals (talc, chlorite, serpentine, or phlogopite). Limonitic material, a distinctive though minor constituent, occurs in pseudomorphs after pyrite, and as a pervasive stain particularly heavy along fractures.

In its characteristic texture, the rock is composed of minute lenticles and veinlets generally less than 2 mm. wide, and of contrasting mineralogy. Some are composed mainly of magnesian silicate minerals, others of very fine-grained quartz, still others of carbonate, albite and quartz. Age relations are not well shown. Some of the chlorite appears to be pseudomorphous after tremolite; much of the talc, albite, quartz and carbonate occurs in late-stage veinlets.

Chemical analyses of two specimens of typical transition rock (table 4) are notably higher in CO_2, Al_2O_3 and K_2O, and lower in MgO than analyses of typical talc-tremolite rock. The former are also somewhat richer in iron and Na_2O. The high Al_2O_3 and K_2O percentages suggests that much of the untwinned alkali feldspar is orthoclase, and that the micaceous material, at first believed to be entirely talc, is largely sericite.

Origin of silicated zones

General statement

Briefly reviewed, a contact metamorphic origin is indicated for the silicated zones by (1) their close spatial relation to diabase bodies, (2) by thicknesses roughly commensurate with the thicknesses of the diabase bodies with which they are associated, and (3) by a recurrence of similar sequences of alteration rock layers outward from the margins of diabase bodies.

The influence of metasomatism is shown in the marked contrast in chemical composition between the altered rocks and the adjacent unaltered carbonate sediments. These differences, considered in more detail below, indicate the abundant addition of MgO and SiO_2 to produce talc-tremolite rock, and SiO_2, K_2O and Al_2O_3 to produce the average green tremolitic rock and transition rock. Large quantities of CaO and CO_2 were removed in the development of all three rock types. The hydrothermal nature of the alteration is shown in the abundance of the hydroxyl-bearing minerals. The extensive hydration of the diabase, probably a deuteric process, appears to relate the metasomatism to the end stages of the diabase emplacement.

Previous studies of contact metamorphism related to mafic igneous rocks

Previous studies of rocks that border mafic igneous bodies ordinarily have shown that metasomatism was not active or was of minor significance. The minerals most characteristic of such border zones are anhydrous and have formed principally through isochemical reconstitution of the wall rocks in a temperature range somewhat higher than can be attributed to the hydroxyl-bearing mineral assemblages of the Crystal Spring alteration zones. To the writer's knowledge the only diabase sills bordered by extensive contact metamorphic bodies composed mostly of hydrous magnesian silicate minerals are those that have intruded Algonkian carbonate strata. In several regions throughout the world asbestos-bearing serpentine bodies have formed in this manner; but such an environment for commercial quantities of talc may well be unique to the Death Valley-Kingston Range region.

Among the reported contact rocks of other mafic bodies are those of the great Whin diabase sill in northern England, as described by Hutchings[47] in 1898. These appear to represent an essentially

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47/ Hutchings, W. M., The contact rocks of the Great Whin sill: Geol. Mag., vol. 5, pp. 69-82, 123-131, 1898.

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isochemical reconstitution of sedimentary rocks. Only soda was described as a derivative of the diabase magma. In the contact zones, which are commonly several tens of feet wide, the siliceous, nonargillaceous limestones were merely recrystallized; whereas the metamorphism of argillaceous limestones and calcareous shales produced such minerals as garnet, augite, idocrase, wollastonite, epidote, hornblende and feldspar.

Lewis[48/], who in 1907 described the contact metamorphism of the

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48/ Lewis, J. V., Petrography of the Newark igneous rocks of New Jersey: New Jersey Geol. Survey, Ann. Rept. of 1907, pp. 97-167, 138-147, 1907.

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shales and arkoses that border the Palisade diabase sill of New York and New Jersey, noted the development of high-temperature assemblages through zones several hundred feet thick. These assemblages, which include the minerals cordierite, pyroxene, plagioclase, quartz, orthoclase, and biotite, form several types of hornfelses. The differences in the hornfels, Lewis states "are dependent only on original differences in the composition of the shales themselves".[49/]

49/ Lewis, J. V., op. cit., p. 139-140.

In 1924, Schwartz[50/] in a description of the comparative effects

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50/ Schwartz, G. M., The contrast in the effect of granite and gabbro intrusions on the Ely greenstone: Jour. Geology, vol. 32, p. 134, 1924.

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of granite and gabbro intrusives on greenstone in northeastern Minnesota, found a thorough dehydration of the greenstone along contacts with gabbro, an effect he attributed to the dryness and high temperature of the gabbro magma. Schwartz[51/], in a later paper, concludes that

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51/ Schwartz, G. M., Metamorphism of extrusives by basic intrusives in the Keweenawan of Minnesota: Geol. Soc. America Bull., vol. 54, p. 1212, 1943.

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metamorphism involving dehydration is "peculiar to the effect of basic intrusives which in general form anhydrous minerals . . . . ".

Ross[52/], also impressed with the lack of metasomatic effects along

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[52/] Ross, C. S., Physico-chemical factors controlling magmatic differentiation and vein formation: Econ. Geol., vol. 23, p. 873, 1928.

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the borders of mafic bodies states as follows:

> "The almost complete absence of recognizable contact-metamorphic effects in the wall rock of even the relatively hot magmas that have formed basalt dikes or diabase sills is noteworthy. Hypabyssal intrusive magmas, even those that are known to have had high temperatures, were in general able to exert little effect on most inclosing rocks by heat energy unassisted by hydrothermal or gaseous emanations . . ."

Evidence of metasomatism in the anhydrous contact zones next to mafic bodies, however, is not altogether lacking. Buddington[53/], for

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[53/] Buddington, A. F., Adirondack igneous rocks and their metamorphism: Geol. Soc. America, Mem. 7, p. 61, 1939.

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example, in his Adirondack Mountain studies has described a gabbro that locally contains limestone lenses partly altered to pyroxene-garnet-scapolite tactite; but he notes that, in this area, contact metamorphism related to gabbros is rare.

Tilley[54/] in describing the high temperature mineral assemblage

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[54/] Tilley, C. E., On larnite (calcium orthosilicate, a new mineral) and its associated minerals from the limestone contact zone of Scawt Hill, Co. Antrim: Mineralog. Mag., vol. 22, p. 86, 1929.

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in the dolerite-chalk contact zone at Scawt Hill, Antrim concludes that

> "the purity of the chalk outside the thermal zone clearly denotes that solutions from the dolerite magma enriched the contact-zone in silica, magnesia, iron oxides, and alumina, so enabling the formation not only of calcium silicates, but also of merwinite, melilite (gehlenite), and spinel."

The contact metamorphic effects of a diabase sill at Safe Harbor, Pennsylvania, as described by Chapman[55/] in 1950 indicate a two-stage

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55/ Chapman, R. W., Contact-metamorphic effects of Triassic diabase at Safe Harbor, Pennsylvania: Bull. Geol. Soc. America, vol. 61, pp. 191-220, 1950.

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history; the first characterized by high-temperature metamorphism, the second by lower temperature, hydrothermal alteration. In the first stage cordierite and orthoclase formed at the expense of quartz and mica in a quartz-feldspar-biotite-muscovite schist, and forsterite was produced by the reaction of quartz and dolomite in a dolomite containing phlogopite, quartz and feldspar. In the second stage water, ferrous iron and magnesia, and possibly alumina and silica were added to the schist to produce chlorite, clinozoisite, chlorophyllite and specularite; and water, magnesia, silica, iron, zinc, and sulphur were added to the dolomite to produce tremolite, diopside, talc, antigorite, magnetite and sphalerite. Chapman concludes that the magnesia and silica were probably not of magmatic origin and that the magnesium may have been derived from dolomite at depth.

Of the asbestos-bearing serpentine bodies that border diabase sills examples are known in northern and south-central Arizona[56/], in

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56/ Diller, J. S., Mineral Resources U. S., 1907, pt. II, pp. 720-721, 1908.

-------------, Mineral Resources U. S., 1908, pt. II, pp. 705, 1909.

Noble, L. F., The Shinumo quadrangle, Grand Canyon district, Arizona: U. S. Geol. Survey Bull. 549, pp. 57-60, 1913.

Bateman, A. M., An Arizona asbestos deposit: Econ. Geology, vol. 18, pp. 663-690, 1923.

Wilson, E. D., Asbestos deposits of Arizona: Arizona Bur. Mines Bull. 126, pp. 28-33, 1928.

Transvaal[57/], and in southern India[58/], In each area the rocks intru-

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57/ Hall, A. L., Asbestos in the Union of South Africa: S. Africa Geol. Mem., no. 12, pp. 145-156, 1930.

58/ Coulson, A. L., Barytes and asbestos in the ceded districts of the Madras Presy.: India Geol. Survey Mem., vol. 64, pts. 1 and 2, 1933-34.

Murphy, P. B., Genesis of asbestos and barite, Cuddapah district, Rayalaseema: Econ. Geol., vol. 45, pp. 681-695, 1950.

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ded by the sills closely resemble the Crystal Spring formation in both lithology and sequence. These other sedimentary rocks - the Grand Canyon series[59/] in northern Arizona, and the Apache group[60/] in south

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59/ Noble, L. F., op. cit.

60/ Ransome, F. L., Geology of the Globe copper district, Arizona: U. S. Geol. Survey Prof. Paper 12, pp. 28-40, 1903.

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central Arizona, the Transvaal system[61] in Transvaal, and the Cuddapah

61/ DuToit, A. L., Geology of South Africa: Oliver and Bond, Edinburgh, pp. 101-135, 1939.

system[62] in southern India - are also probably Algonkian in age.

62/ Wadia, D. N., Geology of India: McMillan Co., Ltd., London, pp. 86-92, 1939.

In brief, the zones of serpentinization range from a few feet to as much as 200 feet in width and have replaced carbonate sedimentary rocks. Chrysotile asbestos occurs in veins, generally less than 3 inches thick, in massive serpentine. All students of these zones have agreed upon a contact-metamorphic origin; most have believed the diabase to be the source of additive silica, magnesia, and water. But Diller[63] and Noble[64] in early descriptions of the Arizonan deposits,

63/ Diller, J. S., Mineral Resources U. S., 1908, pt. II, p. 705, 1909.

64/ Noble, L. F., op. cit.

ascribed a wholly sedimentary source to the magnesia. Coulson[65] suggested a connate origin for the water of the Indian deposits.

65/ Coulson, A. L., op. cit.

Plate 16a. Laminated tremolite-rock near foot-wall of Western deposit.

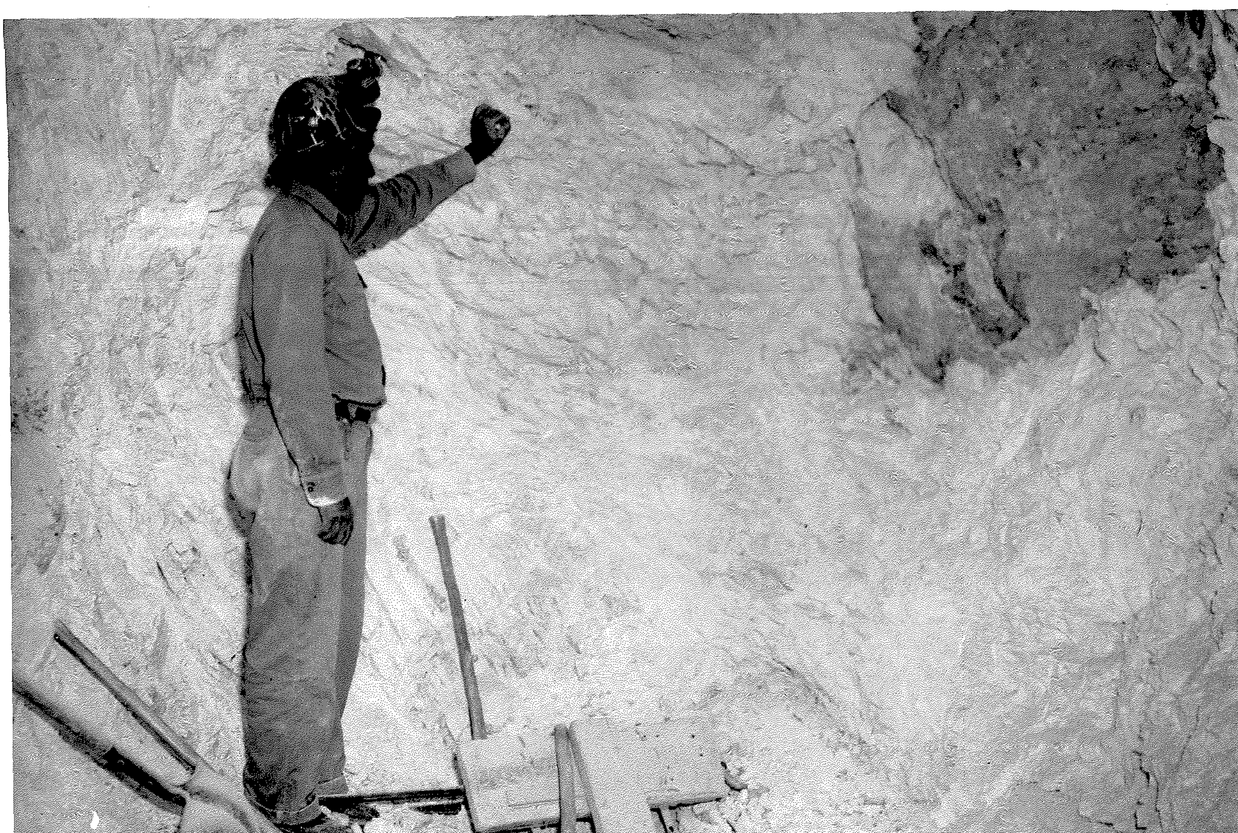

Plate 16b. Brecciated talc schist at hanging wall of Superior deposit.

Paragenesis of the silicated zones

Character and influence of the parent rocks. Because the carbonate member ordinarily shows little lithologic variation within thicknesses of several tens of feet, the parent rocks of the silicated zones are assumed to have been essentially the same as the unaltered carbonate rocks in contact with or nearest the zones. If so, the zones associated with the lower sill consistently have altered from dolomite, either cherty or chert-free. The zones bordering or enclosed by the higher sills have altered from limestone and dolomite, both characteristically chert-free. Of the large concentrations of talc-tremolite rock, only the Ibex deposit appears to have altered from limestone.

A comparison of bodies of talc-tremolite rock formed from cherty dolomite with those formed from chert-free dolomite, shows that the proportion of chert has exercised little, if any, control upon their ultimate mineralogic or chemical composition. For example, the highly talcose Smith deposit, and the highly tremolitic Western deposit have altered from similar cherty dolomites; whereas chert-free dolomite has altered to tremolitic rock at the Valley deposit and talcose rock at the Brown deposit.

In many of the bodies gross layering is a metasomatic effect unrelated to differences in the original strata. This is shown by the common recurrence of the same layered sequence - laminated tremolite rock, talc schist and transition rock - outward from diabase contact regardless of the character of the parent rock. Layering within other zones, however, probably is attributable to such factors as differences in permeability or in the proportion of impurities of the original strata. Although the textural laminations generally do not appear to be relict strata, the altering solutions were probably guided, in large part, by bedding planes, thereby producing the parallelism with bedding in the unaltered, more massive sedimentary rocks.

Talc-tremolite rock and transition rock. The close association of transition rock with talc-tremolite strongly suggests a genetic relationship between the two; the transition rock produced by metasomatism that permitted the coexistance of alkali feldspar, quartz and ferric hydroxide, as well as carbonate and various hydrous magnesian silicates; the talc-tremolite rock a product of metasomatism that yielded hydrous magnesian silicate - carbonate assemblages generally free of ferric hydroxide and feldspar and containing but a minor proportion of late-stage quartz.

Although the transition rock is locally cut by veins of white, micaceous talc, for the most part, it appears to have developed as a concurrent, generally marginal phase of the metasomatism that produced the talc-tremolite rock. The formation of transition rock may have been favored by a decreasing temperature gradient.

As shown in table 4 the commercial talc deposits in the Southern Death Valley-Kingston Range region average about 54 percent SiO_2, 28 percent MgO, 5.5 percent CaO, 4.5 percent CO_2, 4 percent H_2O, 1 percent Al_2O_3, less than 1 percent each of K_2O, Na_2O, and less than 1/2 percent iron oxide. If, as the field relations at the Montgomery mine indicate, the silication involved little or no change in volume, the reconstitution of the carbonate rocks to talc-tremolite rock and transition rock required much additive material and the removal of most of the CaO and CO_2. Even if ideally pure, the dolomites contained but 19.1 percent MgO. With a high chert content so common, the parent rocks probably averaged between 10 and 15 percent MgO. Because few, if any, of the original carbonate rocks, contained as much as 54 percent SiO_2 (many are nearly silica-free), the abundant introduction of SiO_2 is also indicated. The introduction of K_2O and Al_2O_3 in the formation of transition rock is indicated in its high percentage of these two oxides, whereas the unaltered carbonate rocks, as observed in hand specimen and thin section, are ordinarily free of potash- or alumina-bearing minerals. The abundance of albite in much of the transition rock suggests that Na_2O has also been introduced.

From foregoing petrographic descriptions it will be recalled that the major chemical changes in the development of talc-tremolite rock are recorded in the following mineralogic sequences; the replacement of disseminated carbonate by aggregates of tremolite, of chlorite, or of talc; the replacement of tremolite by aggregates or pseudomorphs of chlorite or of talc; and replacement of chlorite by aggregates of talc. The general sequence, therefore, is carbonate, tremolite, chlorite, talc, each of the silicates having formed at the expense of each mineral listed before it. Carbonate later than that in the disseminated grains has locally replaced tremolite and occurs with subordinate quartz in veinlets.

If, as seems likely, a deuteric hydration of the diabase was at least partly contemporaneous with the silication of the bordering rocks, the final stages of the silication occurred near the end of a period of declining temperatures produced by the cooling diabase. The suggestion is strong, therefore, that the sequence tremolite to chlorite to talc formed under successively lower temperatures. This is also suggested (1) by the common occurrence next to the diabase of highly tremolitic layers which are in turn bordered by layers of chlorite-bearing talc schist, and (2) by the occurrence of talcose borders around chert layers as the outermost evidence of silication.

A significant clue to the maximum temperature at which the tremolite could have formed is given by the persistent tremolite-calcite association in all three varieties of silicated rock. These two are diagnostic minerals of the fourth of Bowen's[66] thirteen steps in the

66/ Bowen, M. L., Progressive metamorphism of siliceous limestone and dolomite: Jour. of Geol., vol. 48, no. 3, p. 257, 1940.

progressive metamorphism of siliceous dolomite. This step, above which calcite and tremolite can no longer exist in equilibrium, is shown by the following equation:

$$\underset{\text{calcite}}{2CaCO_3} + \underset{\text{tremolite}}{3CaMg_3(SiO_3)_4} \rightleftharpoons \underset{\text{diopside}}{5CaMgSi_2O_6} + \underset{\text{forsterite}}{2Mg_2SiO_4} + 2CO_2 .$$

In a schematic P-T curve Bowen suggests that for all pressures lower than 1500 atm. the reaction proceeds only at temperatures lower than 650 degrees. Although tremolite does occur in higher temperature assemblages, these occurrences are limited to associations with minerals more magnesian than dolomite[67]. The tremolite-dolomite-calcite

67/ Bowen, N. L., op. cit., p. 245.

association, apparently the earliest shown by the talc-tremolite rock, permits the application of "step four" with reasonable certainty.

The carbonate-tremolite-chlorite-talc sequence, each mineral replacing each of those listed before it, indicates the enrichment in MgO and SiO_2 and removal of CaO and CO_2 to have been continuing processes. The greatest chemical changes, however, appear to have been during the early phase of silication when the highly tremolitic rocks were formed. This also was probably the period of formation of much of the talc and chlorite that has directly replaced carbonate in the outer parts of the bodies. A rapid evolution of CO_2 probably caused maximum pressures during this phase. The replacement of tremolite by chlorite and talc, and of chlorite by talc, therefore, apparently proceeded under successively lower temperatures and pressures and required smaller amounts of MgO and SiO_2.

That stress was also a factor in the later phases of the formation of talc-tremolite rock is shown in the widespread development of talc schist, partly at the expense of rock composed largely of tremolite blades with no pronounced dimensional alignment. It will be recalled, however, that much of the talc rock is not schistose, and that the laminated and schistose layers in a given deposit are generally sharply separated from one another. The masses of laminated rock appear to have undergone little change attributable to stress.

Green tremolitic rock. The paragenesis of the green tremolitic rock consisting, in general, of an apparently contemporaneous development of tremolite and alkali feldspar at the expense of carbonate, is simpler than that of the talc-tremolite rock. Moreover the formation of the green rock has considerably higher proportions of K_2O, Al_2O_3 and probably Na_2O, and correspondingly lower proportions of MgO and H_2O. If, as investigations to date suggest, the green tremolitic rock has formed partly from limestone and partly from dolomite, both cherty and chert-free, but very poor in potassic, sodic, and aluminous materials, these three have been introduced.

In many occurrences of green tremolitic rock, the MgO content may well have been derived entirely from the original sedimentary rocks; but the formation of highly tremolitic varieties, even from ideally pure dolomite, would require an enrichment in MgO. This is also true of the ordinary tremolite-feldspar rock formed from the more calcareous strata. Because the higher carbonate beds ordinarily contain very minor proportions of SiO_2, much of the SiO_2 in the green tremolitic rock also has been introduced.

Cause of petrologic differences between lower and higher zones and sources of additive materials. The localization of talc-tremolitic rock and transition rock to the lower part of the carbonate member, and the preference shown by green tremolitic rock alteration for septa and for zones high in the member, cannot be correlated with original variations in the lithology of the member. Neither can this stratigraphic disposition of alteration rocks be caused by inherent differences in the diabase bodies with which they are associated, a conclusion drawn from the nature of the previously noted silicated zones along diabase dikes that cross the member.

Of probable significance to the problem is the evidence that the lower zones were subject to continuing hydrothermal activity; which produced progressive chemical and mineralogic changes; whereas the tremolite, albite, and quartz in the alteration rock of the septa and higher zones, appear to have formed simultaneously and in equilibrium and to have undergone little, if any, later alteration. Also significant is the likelihood that the diabase intruded soon after the Crystal Spring strata were deposited and that the strata, then poorly consolidated, contained an abundance of connate water. If so, the lower sill, intruding at or near the top of a sequence of water-saturated non-carbonate strata, probably crystallized in a more aqueous environment than did the higher diabase bodies bordered by carbonate strata. Similarly the lower zones along the cross-cutting dikes would have been closer to such a source of water than would the higher zones.

The localization of talc-tremolite rock and transition rock, therefore, is perhaps attributable to the greater availability of connate water to alteration affecting the basal beds of the carbonate member. Indeed, intrusion into water-rich strata may well account for the apparent, unusual wetness of the diabase magma. Most of the additive SiO_2, MgO, K_2O, Na_2O, and Al_2O_3, however, was probably magmatic in origin. It will be remembered that mafic bodies at other localities have been described as sources of one or more of these components, and that in the bodies bordering the diabase of the Crystal Spring formation all five have been introduced in the higher as well as the lower silicated zones. In the Crystal Spring diabase, the abundant alteration of labradorite to chlorite and sericite and of augite to uralite and chlorite suggest that SiO_2

and Na_2O were released in the late stages of diabase emplacement.

Connate water may have diluted hydrothermal solutions emanating from the diabase, or, if strongly saline, may have effectively enriched the solutions in soda, potash, chlorine or magnesium. That additive MgO has traveled upward from other dolomites, as suggested by Chapman[68/] for the contact rocks at Safe Harbor, is unlikely in the

68/ Chapman, R. W., op. cit.

silicated zones of the Crystal Spring carbonate member beneath which dolomite occurs in insignificant volume.

THE WHITE EAGLE DEPOSIT

GENERAL STATEMENT

The White Eagle and the other talc deposits of the Inyo Range district differ in several respects from the talc deposits in the Southern Death Valley-Kingston Range and Silver Lake areas. Mostly alterations of Paleozoic sedimentary rocks, less commonly of Mesozoic granitic rocks, the Inyo Range deposits are either Late Mesozoic or Tertiary in age. They are ordinarily pod-like or irregular in shape, and tremolite-free, thereby contrasting with the highly tremolitic, very elongate pre-Cambrian deposits described in preceeding sections. Moreover, the Inyo Range deposits, apparently showing no genetic relationship with nearby igneous bodies, were farther removed from the source of their altering solutions than were the others.

Most of the Inyo Range deposits occur at or near contacts between dolomite and quartzite and have replaced both. Part of the White Eagle deposit has formed in this manner, but it is largely a replacement of adamellite, which, in the Inyo Range region, is an uncommon parent rock for talc.

PHYSICAL FEATURES

The White Eagle mine, high on the east flank of the Inyo Range (pl. 17), overlooks the north end of Saline Valley, and lies about one mile west of Willow Creek Camp, a small group of permanent buildings at the base of the range. The camp is at an elevation of 2000 feet; the mine is about 1500 feet higher. Annual rainfall probably averages less than 3 inches, but a small perennial stream, Willow Creek, flows southeastward through the camp.

The camp is about 16 airline miles due east of Independence, but it is reached by a 52 mile, fair to poor, dirt road extending southeast from Big Pine. The deposit, about 1 1/2 miles by trail from the road, is exposed on a slope of nearly barren bedrock, locally covered by talus.

GENERAL GEOLOGY

Regional features

As shown by Knopf[69/] in his classic reconnaissance studies

69/ Knopf, Adolph, A geologic reconnaissance of the Inyo Range and the eastern slope of the southern Sierra Nevada, California: U. S. Geol. Survey Prof. Paper 110, 1918.

Plate 17. View westward of east face of Inyo Range showing location of White Eagle deposit (arrow). Slopes beneath and north of deposit underlain principally by granitic rocks; Paleozoic sedimentary rocks are exposed on slopes above and south of deposit.

of the Inyo Range area, a pluton of adamellite underlies an area of about 100 square miles in the central part of the range. The northern boundary of the pluton lies about 2 1/2 miles north of Willow Creek, its southern boundary about 10 miles to the south. Low on the east flank of this part of the range, however, is a belt of westerly dipping Paleozoic sedimentary rocks, extending from the vicinity of Willow Creek southeastward for a distance of eight miles. The belt is bordered by Quaternary alluvium on the east, elsewhere by adamellite. In plan, it has a maximum width of about 2 miles at its northern end. Here the White Eagle talc deposit has formed at a locality where dikes and irregular apophyses of adamellite extend into Paleozoic crystalline carbonate rocks and quartzites.

Geology of the mine area

General features

The geological features of the immediate area of the mine have been described by Page[70/], and will be but briefly reviewed here.

70/ Page, B. N., Talc deposits of steatite grade, Inyo County, California: California Div. Mines, Spec. Rept. 8, in press, 1951.

Three major sedimentary units exist in the mine area. As exposed in upward sequence, these are (1) a unit composed partly of crystalline limestone, partly of dolomitic marble, and containing lenticular, siliceous, algal (?) bodies, (2) a quartzite layer from 75 to 150 feet thick, and (3) a predominantly dolomitic, locally limy, marble, free of siliceous bodies. The carbonate units were not measured, beyond the area of the accompanying map, but are considerably thicker than the quartzite. Although extensively invaded by adamellite, the larger masses of sedimentary rock extend southward into a relatively intact section. Commonly included in the adamellite of the mine area, however, are angular blocks of carbonate rock as much as 200 feet in maximum dimension. Contacts of adamellite with the sedimentary rocks are sharp and ordinarily show a one- to two-inch selvage.

On the basis of lithology, the sedimentary rocks are tentatively correlated with Silurian[71/] units, widely exposed in the Inyo Range area,

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71/ Merriam, C. W., Silurian quartzites of the Inyo Mountains, California: Geol. Soc. Am. Bull., vol. 61, no. 12, 1951.

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and common as parent rocks for talc. No fossils other than the algal (?) remains were noted.

Petrology

Carbonate rocks. Both carbonate units are massive and rarely show bedding features. They range in color from medium gray to very pale orange, and from fine- to medium-grained. In general the dolomitic facies are lighter colored and coarser-grained than the limestones. In thin section the carbonate is seen to form a mosaic of clear, equant grains with diameters ordinarily less than 0.5 mm. in the limestone, and in the 1 mm. to 5 mm. range in the dolomite. The color range is caused mostly by differences in the abundance of opaque, carbonaceous (?) particles. Except for the siliceous algal (?) bodies the only other impurities noted were traces of disseminated tremolite and quartz.

Quartzite. The quartzite, a white, compact, vitreous rock is massive, but locally shows well-developed cross-laminations. The thin sections examined show sutured quartz grains, flattened parallel with the bedding, generally from 1 mm. to 3 mm. long and less than 1 mm. wide in cross section. Rows of minute bubbles criss-cross the quartz grains, seemingly in random directions. The rock contains less than one percent impurities, principally muscovite and alkali feldspar; these occur in grains averaging less than 0.1 mm. in diameter and evenly disseminated through the rock.

Adamellite. The adamellite, a gray, medium-grained, porphyritic rock, appears to be structurally homogeneous. Its average composition is about 45 percent plagioclase, 25 percent potash feldspar, 20 percent quartz, 10 percent biotite, and a percent or two of hornblende. Sphene, magnetite and ilmenite are relatively abundant accessories.

The phenocrysts, generally between 5 mm. and 10 mm. in diameter, are of microcline microperthite. Oligoclase, orthoclase, and microperthite grains of the groundmass are mostly from 1 mm. to 3 mm. in diameter. The feldspars are anhedral with sutured borders. Microperthite and orthoclase grains are equant, oligoclase grains are somewhat elongate. Oscillatory zoning is common in the oligoclase grains and orthoclase grains. Quartz grains are generally round or ovoid in cross-section and have diameters within the 0.1 mm. to 5 mm. range. Biotite shreds and hornblende blades are as much as 5 mm. long.

Sericite, the most abundant mineral, has altered from both feldspar and biotite. In general, the oligoclase is more extensively sericitized than the potash feldspars.

Talc bodies

General features. Most of the talc in the White Eagle mine area is contained in an irregular body (pl. 4) which in plan is a crude obtuse triangle nearly 600 feet long and 200 feet in maximum width. The body trends north-northwest; surface exposures suggest a moderate westerly to southwesterly dip. It is in contact with all three sedimentary rock units and has partly altered from each. But it is bordered mostly by adamellite, both acute angles of the triangle terminating in this rock. That adamellite is the principal parent rock is shown by the existence, throughout the main talc body, of abundant inclusions, all showing various stages in the alteration of adamellite to talc.

The mine area also contains numerous smaller replacement bodies of talcose rock. The largest of these, about 150 feet long and 5 to 10 feet in average width, is in the lower carbonate unit, and is near and roughly parallel to the lower margin of the quartzite. Numerous smaller talcose lenses and veins exist within the quartzite and carbonate rocks.

Character of the talc. Pale green talc is characteristically associated with adamellite, white to very pale green talc with quartzite, and white to medium gray talc with the limestone and dolomite. Irrespective of color and associated rocks, the White Eagle talc is very fine-grained, and is ordinarily semi-translucent and blocky. Some of it has a poorly developed schistosity; friable masses are not uncommon.

Plate 18. Inclusion broken from face of principal White Eagle talc body; originally surrounded by talc. Note bleached transition zone separated from unaltered core by dark, limonitic rim.

Thin section studies show the talc to be in stubby, ovoid to shred-like grains, mostly less than 0.3 mm. long, with no pronounced dimensional alignment. The most common impurities (feldspar in the green talc, quartz in the brighter green to white talc, and carbonate in the gray talc) are residual from pre-existing rocks.

Character of the alteration

Alteration of adamellite. The alteration's preference for adamellite is probably attributable to a higher degree of pre-talc fracturing in this rock than in the sedimentary rocks. The most completely altered parts of the principal talcose body seem to have formed in the most fractured adamellite; whereas the partly altered adamellite inclusions represent relatively unfractured, less permeable remnants of the original rock.

Many of the inclusions contain a core of unaltered adamellite separated from nearly monomineralic talc rock by a transition zone ordinarily less than one foot thick (pl. 18). Other inclusions, more completely altered, contain no unaltered core, but are partly talcose for their full thickness. Where the entire transition is shown, the alteration has produced, from the unaltered rock inward (1) a reddish brown outer zone, less than one inch wide, showing the disintegration of biotite and hornblende, precipitation of the iron as ferric hydroxide, and incipient alteration of feldspar and quartz to talc; (2) a pale grayish green, "bleached" middle zone, free of ferromagnesian minerals and ferric hydroxide, showing pronounced albitization (at the expense of potash feldspar and quartz) and extensive alteration of albite and quartz to talc; and (3) an inner zone consisting almost entirely of talc. Zones of similarly "bleached" adamellite, several tens of feet wide are peripheral to the main talc body. In all of the many exposures showing, within a several inch thickness, the adamellite to talc transition, this zonal sequence is repeated. The alteration, therefore, appears to have been confined to a single period and to have been produced by solutions essentially uniform in composition.

Chemical analyses (table 5) together with modal analyses of thin sections cut across the transition zone, permit a diagrammatic representation of these mineralogic changes (fig. 13). Thin section examinations show that the ferric hydroxide, formed from the disintegration of biotite and hornblende, occurs as a reddish brown stain, filling fractures and coating feldspar and quartz grains. Incipient alteration of feldspar and quartz to talc also characterizes this outer zone.

That virtually all of the feldspar in the middle zone is albite is shown by the rock's very low K_2O content and an Na_2O to CaO ratio of about 13 to 1. Albite pseudomorphs after microcline are cloudy and untwinned. Other albite grains which are euhedral, unclouded and show polysynthetic twinning (pl. 19b), have grown into quartz grains, a replacement causing a progressive increase in albite and decrease in quartz through the transition zone to talc-rock.

The talc which evenly increases in abundance from unaltered rock through the transition zone, has formed principally at the expense of albite, less commonly of quartz. Talc shreds, singly or in aggregate, have developed interstitial to or within albite and quartz grains (pl. 19b), and along fractures. Talc replacement is particularly

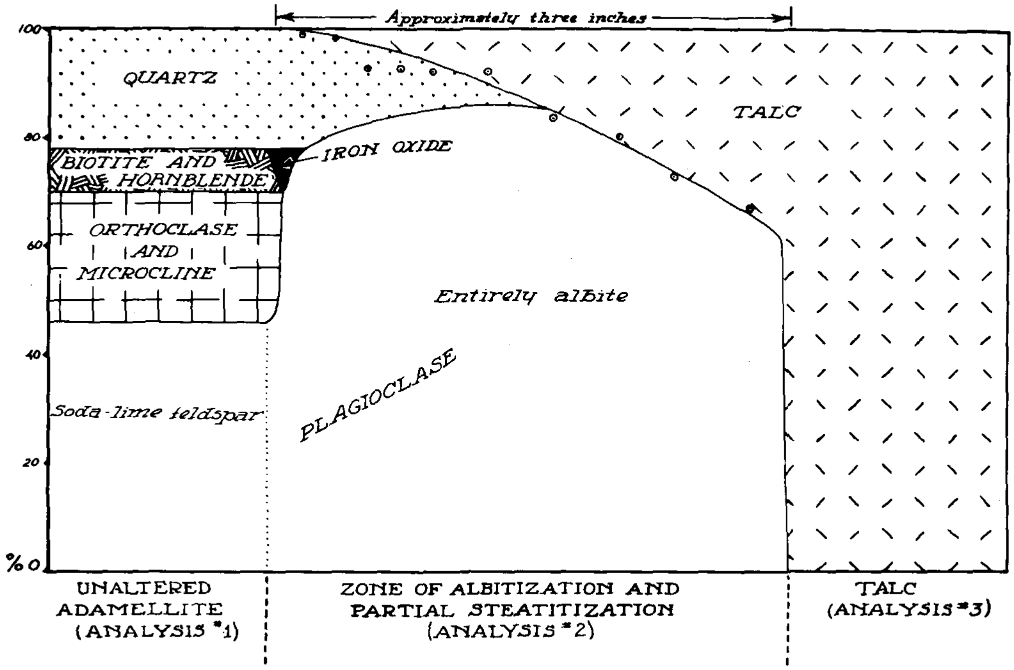

Figure 13. Diagram showing progressive mineralogic changes across the zone of transition between adamellite and talc.

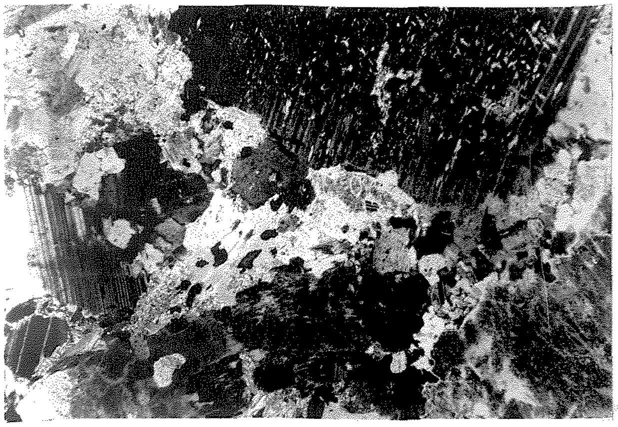

Plate 19a. Photomicrograph showing incipient alteration of albite by talc. Crossed nicols. Field 10 mm. long.

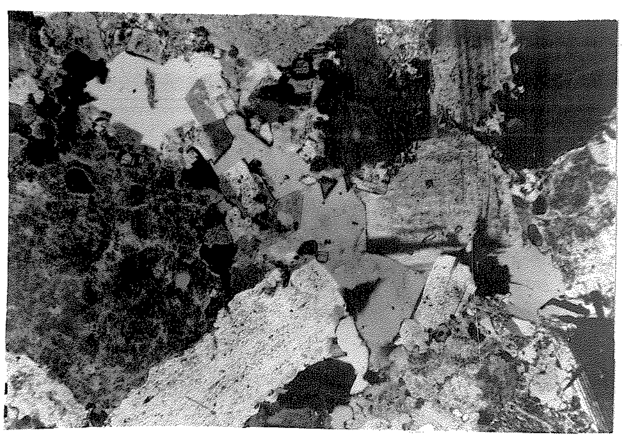

Plate 19b. Photomicrograph showing corrosion of albite by talc. Note quartz grains (center) into which extend subhedral to euhedral albite grains. Crossed nicols. Field 10 mm. long.

pronounced along cleavage cracks and twin lamellae in albite.

The contact between the transition zone and relatively pure talc rock is ordinarily sharp, but in some places there is no perceptible lithologic break. The sharp contacts mark fronts of nearly complete replacement of albite by talc leaving only traces of feldspar in the talc rock. Such fronts are characterized by a band of talc rock less than one-fourth inch thick in which a large proportion of the talc grains lie normal to the contact. Along the gradational contacts numerous albite "islands" exist in a "sea" of talc (pl. 20a).

The alteration of adamellite to talc rock, therefore, has involved the ultimate removal of most of the Al_2O_3, Na_2O, CaO, and Fe_2O_3 originally contained in the adamellite, it has advanced behind a wave of intensive Na_2O and moderate MgO and H_2O enrichment, a wave enclosed by two simultaneously advancing fronts, ordinarily showing closely spaced (one foot or less), sharp boundaries, but in some places more widely spaced and gradational.

<u>Alteration of quartzite</u>. The quartzite to talc alteration is most pronounced at the termination of the quartzite layer against the main talc body. Several talcose veins, all less than 4 feet wide and 100 feet long and apparently guided by fractures extend into the quartzite, other talcose masses irregularly embay it. Contacts between quartzite and relatively pure talc range from very sharp to gradational through several inches.

Partly talcose quartzite, when observed in thin section, shows an association of albite with the talc. The albite grains, identical in habit with the newly-formed euhedral albite in the partly altered

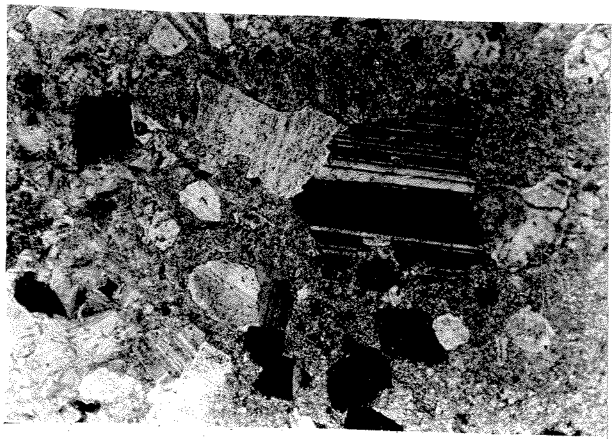

Plate 20a. Photomicrograph of impure talc-rock showing residual albite "islands" in a "sea" of talc. Crossed nicols. Field 10 mm. long.

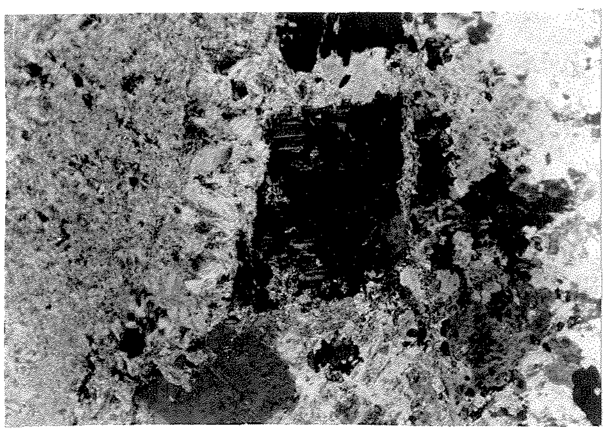

Plate 20b. Photomicrograph of principal replacement front showing corrosion of albite by talc. Talc rock (nearly monomineralic) to left, albite-rich transition rock to left. Crossed nicols. Field 10 mm. long.

adamellite, are much more abundant than the alkali feldspars of the unaltered quartzite. Albite appears to have formed, with talc, interstitial to quartz grains and along fractures, but is consistently earlier than the talc which has replaced both albite and quartz with no marked preference for either. The albite apparently records a wave of Na_2O and Al_2O_3 metasomatism, related to, but preceeding, the MgO (talc-producing) metasomatism. The alteration has involved removal of SiO_2 and addition of MgO and H_2O.

Alteration of carbonate rock. Gray talc, formed at the expense of both dolomite and limestone, occurs in zones, less than 10 feet thick, along the southern and eastern borders of the principal talc body, and in layers and lenses within the carbonate masses. The alteration has largely followed the pre-existing contacts between carbonate rocks and adamellite. Less extensively it has followed bedding planes and fractures within the carbonate rocks. Many of the siliceous algal bodies show a partial to complete selective replacement by talc.

Thin sections of partly altered carbonate rock show that in the incipient stages of alteration talc and chrysotile developed as disseminated flakes at the expense of carbonate. Also present are tremolite grains, thinly scattered, but more abundant than in the thin sections of non-talcose rocks. The chrysotile has been largely replaced by talc, and is less abundant in the partly altered rock than in the completely altered rock. Much of the talc associated with carbonate rock has replaced chrysotile as an intermediate stage. Some of the talc, however, is acicular and probably pseudomorphic after tremolite. Much talc may also have directly replaced carbonate material.

In the carbonate to talc alteration shown at the White Eagle mine, as at the deposits in the Southern Death Valley-Kingston Range and Silver Lake areas, the addition of MgO, SiO_2 and H_2O, and the subtraction of CaO and CO_2 is indicated.

Conclusions

The occurrence of zones between adamellite and talc, everywhere showing the same mineralogic transition inward from unaltered rock, strongly suggests a continuous reaction between adamellite and a solution whose composition remained essentially unchanged during the alteration. That the same solution also altered the quartzite and carbonate rocks is indicated by the close spatial relation of talc bodies altered from each of the three, and by the albite-talc association shown in alteration of both adamellite and quartzite.

The solution was capable of breaking down calcite, dolomite, quartz, potash feldspar, plagioclase, biotite, hornblende, and limonitic material. In so doing it carried away CaO, SiO_2, K_2O, Na_2O, Al_2O_3 and Fe_2O_3. It was also capable of adding MgO and H_2O to each of the three parent rocks and, in addition, SiO_2 to the carbonate rocks. In the siliceous rocks, however, an envelop of albite-forming Na_2O, was pushed ahead of the principal alteration front. The albite, formed principally by the displacement of K_2O by Na_2O in potash feldspar, less abundantly by replacement of quartz, was metastable in the partly talcose rock, but was soon almost completely replaced by talc where the alteration was most intense.

The unstability of the minerals of the parent rocks may be partly attributable to a temperature gradient but the marked compositional changes could only have been produced by a chemically potent hydrothermal solution rich in MgO, and poor in the subtracted materials, but containing sufficient SiO_2 to deposit it in the carbonate rocks.

The Na_2O and Al_2O_3 added in the transition zone probably were derived from nearby adamellite in which talc alteration is nearly complete.

The source of the MgO is less clear. The thorough pre-talc fracturing in the adamellite records a deformational period separating the adamellite emplacement and the talc alteration. Moreover, elsewhere in the region, post-adamellite dikes also show fracture-controlled talc alteration. The MgO, therefore, apparently did not originate in an adamellite magma. At other talc-bearing localities, such as the White Mountain mine about 15 miles to the south, a sedimentary origin for MgO is indicated in an abundant alteration of dolomite to a punky lime-rock, an alteration common along the borders of individual talc bodies. Although lime-rock of this type was not noted near the White Eagle mine, the MgO may well have been leached from dolomitic rocks at depth.

	1. Unaltered adamellite Sp. G. == 2.65	2. Transition rock Sp. G. == 2.62	3. Talc rock Sp. G. == 2.82
		Chemical analyses	
SiO_2	69.09	64.20	61.59
Al_2O_3	15.36	13.79	1.38
Fe_2O_3	.92	.04	2.16
FeO	1.55	.92	
MgO	.86	10.03	29.71
CaO	2.31	.56	.05
Na_2O	4.06	7.36	.17
K_2O	4.03	.17	
H_2O+	.34	1.85	4.66
H_2O-	.05	.11	
CO_2	.02	.16	.28
TiO_2	.34	.35	
P_2O_5	.15	.15	
MnO	.06	.00	
	99.15	99.69	100.00
		Gains and losses per 100 cc. of rock	
SiO_2		−15.88 gm.	−10.40 gm.
Al_2O_3		+ 4.68	−38.61
Fe_2O_3		− 1.99	− 3.55
FeO		− 2.65	
MgO		+23.80	+81.31
CaO		− 4.66	− 5.98
Na_2O		+ 8.53	−20.95
K_2O		−10.23	
H_2O		+ 4.10	+12.90
H_2O		− 3.66	+ 14.72

Table 5. Chemical analyses and changes in bulk chemical composition of suite showing alteration of adamellite to talc. Analyses 1 and 2 by L. C. Peck, University of Minnesota. Analysis 3 by A. J. McArthur, Sierra Talc and Clay Company.

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