Geology of the Cerro Gordo Mining District Inyo County, California

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CONTENTS

ILLUSTRATIONS

[Plates are in pocket]

TABLES

GEOLOGY OF THE CERRO GORDO MINING DISTRICT, INYO COUNTY, CALIFORNIA

By C. W. Merriam

ABSTRACT

The Inyo Mountains near Cerro Gordo comprise strongly folded and faulted sedimentary rocks ranging in age from Ordovician to Middle Triassic. These were intruded by granitic bodies, aplite dikes, and by innumerable andesitic and dacitic dikes of later age. Though largely nonfoliated, the sedimentary rocks have undergone varying degrees of contact and hydrothermal metamorphism productive of hornfels, calc-hornfels, phyllite, and quartzite.

Tertiary basaltic rocks and tuffs cover older rocks at the southern tip of the range, but do not enter the area of the present map.

Paleozoic rocks of the Cerro Gordo area are more than 11,000 feet thick and include all systems from Ordovician through Permian. Mapped units which have wide distribution in the Great Basin are the Pogonip group, Eureka quartzite, and Ely Springs dolomite of the Ordovician and the Chainman shale of Mississippian age. Silurian and Devonian rocks are represented by the Hidden Valley dolomite and the Lost Burro formation, the former being largely Silurian, but embracing Lower Devonian strata at the top. The Lost Burro includes the *Stringocephalus* zone near the base and is of late Middle and Upper Devonian age. This unit is largely a nondolomitic marble in this area and is especially important as host rock of the principal ore bodies.

The Mississippian system is represented by two principal formations: Tin Mountain limestone below and Chainman shale above. A third unit, the Perdido formation, wedges in to the east between Tin Mountain and Chainman. Being less than 100 feet thick near Cerro Gordo, it has not been differentiated from the Chainman in mapping.

Pennsylvanian and Permian strata are divided into two formations: Keeler Canyon formation of Pennsylvanian to Early Permian age and Owens Valley formation of Permian age. These strata are predominantly impure carbonates, with subordinate shale, siltstone, sandstone, conglomerate and chert. Stratigraphic division of the Pennsylvanian and Permian was accomplished mainly by study of the abundant fusulinids.

Some 4,000 feet of Lower and Middle Triassic rocks are exposed on the west side of the Inyo Mountains. Of marine origin is the lower 1,800 feet which comprises shale, thinly-bedded limestone and thick-bedded lenticular reefy limestone. The upper part of the Triassic section comprises volcanic rocks and land-laid deposits in which reddish coloration is characteristic. The Triassic succession is incomplete, for on the west side of the Triassic belt the volcanic rocks are in fault contact with fusulinid-bearing Permian beds of the Owens Valley formation.

Intrusive rocks of the Cerro Gordo area include the older granitic and aplitic rocks of possible Cretaceous age and younger andesitic and dacitic porphyry dikes. The younger porphyry dikes occur in large numbers and for the greater part strike northwest. In the Cerro Gordo mine such dikes, in fractured condition, seem to have served as avenues of ascent for mineralizing solutions.

Rocks of the Cerro Gordo area are extensively folded and faulted. Most significant structural feature is the large asymmetrical south-plunging Cerro Gordo anticline which forms a sort of backbone to the Inyo Range. On its flanks and crest are irregular subsidiary flexures. Bordering the major anticline are many smaller folds with northwest axial trend. These range greatly in magnitude and tightness, partly in response to varying competency of strata involved. Some of the folds are related to reverse faults or thrusts. The Cerro Gordo mine is situated in the axial zone of the anticline which carries its name.

Faults having a northerly trend are characteristic of the region. Among these is the important Cerro Gordo fault, master fault of the Cerro Gordo mine. Northwestward-trending normal faults greatly complicate geologic structure in the Cerro Gordo mine, where certain of these offset ore bodies.

Silver, gold, lead, zinc, and in minor amounts copper are the metallic commodities of the Cerro Gordo area. The Cerro Gordo mine formed by consolidation of the Union mine, San Felipe, and the Santa Maria far exceeds in production all others of the area combined. Estimates of total Cerro Gordo mine production show about 4,400,000 ounces of silver, 37,000 tons of lead, and 12,000 tons of zinc from zinc carbonate ore. Year of peak production was 1874. More than half the lead and three-fourths of the silver were produced in the years 1869 through 1876.

Ores of the Cerro Gordo mine occur in Devonian marble of the Lost Burro formation on the east or footwall side of the northward-trending Cerro Gordo fault. This fault is seemingly normal and carries Chainman shale down on the west against marble of the Lost Burro formation. Largest ore bodies were found in two channels which rake steeply to the south, which is the plunge direction of the Cerro Gordo anticline. The two principal ore channels known as the Union chimney on the north and Jefferson chimney on the south occur in fractured marble close to the master Cerro Gordo fault. They were fed by fissures which formed in sympathy to movement on the master fault. Major ore bodies occurred also in the sheared Jefferson diabasic dike. Quartz veins with northwest strike yielded siliceous ores of silver, lead, and copper. Carbonate-zinc ores are secondary, derived by leaching of sulfide ores in the Union chimney vicinity. Supergene zinc-carbonate ores replaced unmineralized Lost Burro marble along bedding. In the lower part of the Union chimney, primary sulfide replacement was also controlled in part by bedding.

The Union ore channel was bottomed near the northwestward-trending San Felipe siliceous vein where the vein lies against a dacite porphyry dike. The very steep Jefferson chimney extended to a much greater depth, but was cut off below the 900 level by northwest-trending normal faults. Ore in the Despreciada section of the mine may represent faulted deeper parts of the Jefferson chimney.

South of Cerro Gordo, the Morning Star mine, the Charles Lease tunnel, and the 8,100-foot low-level Estelle tunnel were opened to explore the Castle Rock siliceous vein and ground beneath gossans in the Tin Mountain limestone. The Estelle also provided means of searching for inferred deep continuations of the rich Cerro Gordo ore channels. Morning Star and Estelle production was small. That of the Morning Star came principally from the Gold stope in Lost Burro marble. Estelle ore was mined near the tunnel level from upper Hidden Valley dolomite east of the Cerro Gordo anticline axis.

Among lesser mines the Ella, the Perseverance, and mines in Belmont Canyon yielded siliceous silver-bearing ores used as fluxing material in the Cerro Gordo furnaces. These mines are in a wide zone of northwest shearing which includes northwestward-trending quartz veins of the tetrahedrite-galena-barite type characteristics of the region.

West of Cerro Gordo the now inaccessible Ignacio mine lies in altered and intruded Chainman shale near the boundary with overlying silicated limestone of the Keeler Canyon formation. Principal Ignacio silver production seems to have come from a fissure zone along the northeastward-trending Ignacio fault. Westernmost mine of the Cerro Gordo area is the Sunset, which lies in partly silicated limestone of the Keeler Canyon. A small amount of lead and silver came from two narrow intersecting veins.

INTRODUCTION

The Cerro Gordo mining district derives its name from a limestone peak (alt, 9,184 ft) near the south end of the Inyo Mountains (fig. 1). Together with a lofty northern prolongation known as the White Mountains, the Inyos occupy a position near the west margin of the Great Basin. Across Owens Valley rises the impressive Sierran scarp (fig. 2) culminating in Mount Whitney. On the east lies the rugged Panamint terrane and beyond it Death Valley. Roughly parallel mountain ranges and basins having a northwesterly trend characterize the Inyo-Death Valley region. Northward-trending geomorphic features, though evident, are less obvious than in the more typical Great Basin territory to the north and east.

At the foot of the southern Inyo Mountains is Owens Lake (alt, 3,570 ft), now practically dry (Gale, 1915) because of diversion of Owens River. Saline Valley, a smaller dry lake basin (Gale, 1914), flanks the range on the east, its lowest point (alt, 1,059 ft) some 2,500 feet below the Owens Valley floor (fig. 3). Difference of altitude from the Sierran crest near Mount Whitney to the Owens Valley floor is about 10,000 feet, and therefore commensurate with that which separates the higher Inyo summits from the bottom of Saline Valley.

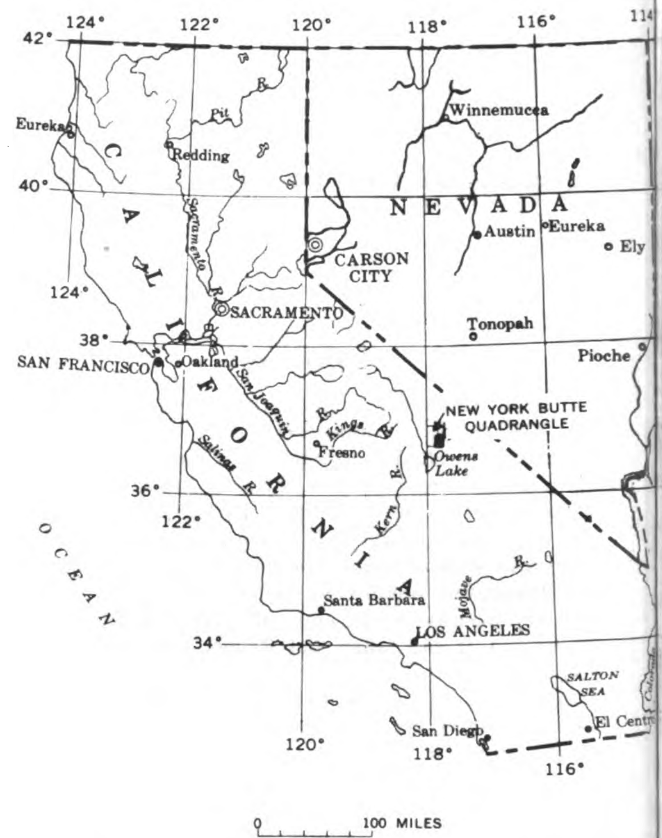

FIGURE 1.—Index map showing the location of the Cerro Gordo mining district (shaded) in New York Butte quadrangle.

Desert climate of the Inyos resembles that of other mountain ranges in the southern Great Basin. At Keeler, on Owens Lake, precipitation averages about 3.15 inches per year (Lee, W. T., 1906, p. 18; Lee, C.H., 1912, p. 23) but is considerably greater on the higher mountain slopes. During winter the range is sometimes snow covered above 6,000 feet; in fact, work stoppage at the Cerro Gordo mine by reason of heavy drifting snow is a not infrequent entry in the mine records. Cloudbursts in Keeler Canyon during the summer have many times partly destroyed the mine road, threatening vulnerable Keeler itself.

Persistent streams are few and small in the arid Inyo Mountains, unlike the Sierra Nevada which borders Owens Valley on the west. Normal runoff appears to be greater on the east slopes of the Inyos than on the west, where the disparity in moisture between this range and the opposing Sierra is patently manifested by differences in vegetation and geomorphic development. Except during storms the east flank of the Inyo Range is drained by a few minor spring-fed streams like that in Hunter Canyon 11 miles northwest of Cerro Gordo. Normally these do not exhibit surface flowage

FIGURE 2.—View from a point near Hart Camp looking west across the west slope of the Inyo Mountains and Owens Lake basin to the Sierra Nevada. In the middleground are westward-dipping resistant Triassic volcanic rocks and underlying marine Triassic shaly beds. Near the middle of the photograph at the lake margin is Smelter Hill, which is underlain by Ordovician and Silurian rocks. Photograph by L. G. Henbest.

far beyond the fan heads near the canyon emergence.

Vegetation of the lower Inyo slopes includes such desert types as sagebrush, greasewood, rabbit brush, desert holly, and salt grass. Near habitations the introduced phreatophyte salt cedar has taken hold and continues to spread. Up to altitudes of about 7,500 feet are scattered stands of Joshua tree. Above this altitude the juniper and piñon pine are dominant, with thickets of mountain mahogany in favored situations. Much of the pine was cut for smelter charcoal during boom days of the Cerro Gordo mine.

A famed and once spectacular producer of silver and lead, the Cerro Gordo mine is near the summit of the range at Cerro Gordo Peak. The terrane is extremely precipitous in this vicinity, especially along the east slope; north of Cerro Gordo it rivals the bold east face of the Sierra itself (fig. 4). This factor of ruggedness, coupled with scarcity of readily obtainable water, contributes immeasurably in this area to the difficulties of access and mine operation.

Springs are few, small, and unreliable in the higher parts of the Inyo Mountains. Close to the range summit, 3½ miles north of Cerro Gordo are several small springs which emerge from the Chainman shale. These springs are named Belshaw, Cronn, and Mexican spring (pl. 1); water from them has in the past been piped to Cerro Gordo, providing an inadequate but welcome supply. The three springs are known collectively as the Cerro Gordo springs. Water from better springs in granite areas of upper Hunter Canyon and Craig Canyon once reached Cerro Gordo briefly through an expensive and elaborate pumping system and a 12-mile pipeline.

Nearest important center of population is Lone Pine, 15 miles northwest of Cerro Gordo beyond the north end of Owens Lake. At the foot of the range on the former lakeshore is the small village of Keeler, southern terminus of the Southern Pacific narrow-gage railroad. Keeler Canyon, with its mouth 1½ miles northeast of the village, leads upward to Cerro Gordo Peak.

When the narrow-gage railroad was completed, about 1883, Keeler became a shipping point for the Cerro Gordo mine and other lead-silver mines at Darwin. Of late years the town has been kept alive by a fairly active Inyo County talc industry. Owens Valley is served by a broad-gage branch of the Southern Pacific Railway from Mojave, which connects with the narrow-gage line at Owenyo northeast of Lone Pine. The narrow-gage line today operates only from Keeler to Laws Station near Bishop, Calif. It continued formerly over Montgomery Pass to Nevada. Owens Valley is connected with Mojave by an excellent paved highway which passes through Lone Pine to Bishop and beyond. From a junction south of Lone Pine the paved highway to Death Valley now bypasses Keeler.

A very steep and not infrequently washed-out road follows part way up Keeler Canyon, rising 4,600 feet in a distance of about 7½ miles to reach Cerro Gordo. From the mine it descends along San Lucas Canyon on the east slope to connect with a road from Lee Flat to the Bonham talc mines. The Cerro Gordo area is otherwise largely inaccessible to conventional wheeled vehicles. During the course of this work, jeeps and other four-wheel-drive vehicles made it possible to reach the range summit at Burgess mine following the old salt company wagon road up Swansea Wash. (pl. 2). A road along the mountain crest from Burgess mine to Mexican Spring was likewise usable. In days of active mining and prospecting many pack trails existed; most of these have fallen into a state of disrepair, though used occasionally by hunters and cattlemen. Elaborate aerial tramways have twice been constructed from Keeler to Cerro Gordo. That now in existence operated successfully from about 1915 to fairly recent times. A remarkable 13-mile tram (Gale, 1914, p. 416), now long abandoned, was constructed about 1912 by a salt com-

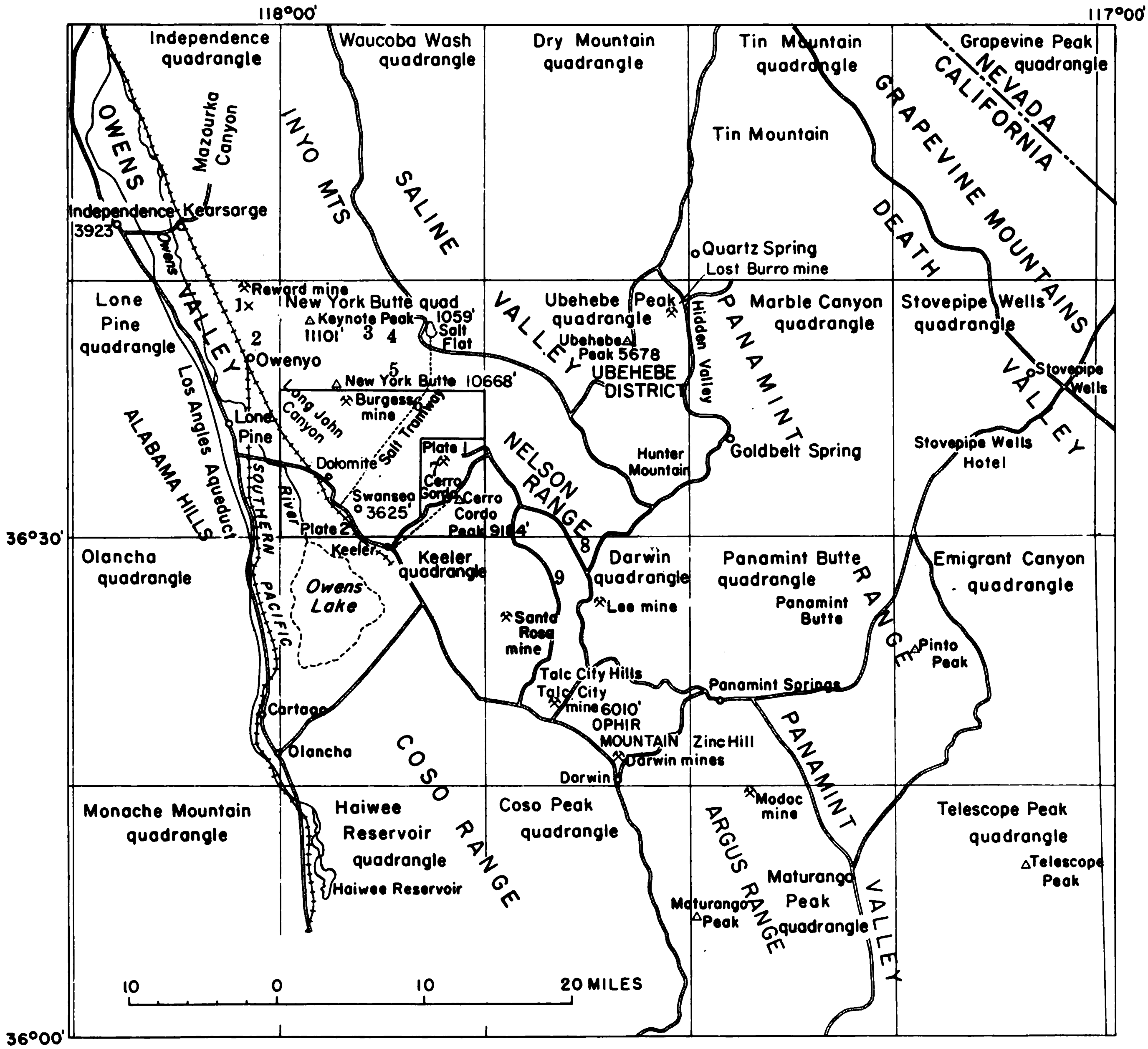

FIGURE 3.—Index map of the Inyo-Panamint region showing localities to which reference is made. 1. Fossill Hill, 2. Union Wash, 3. Beveridge Canyon, 4. Hunter Canyon, 5. Craig Canyon, 6. Daisy Canyon, 7. Bonham Talc mines, 8. Lee Flat, 9. Santa Rosa Flat.

FIGURE 4.—Oblique aerial photograph facing east slope of the Inyo Mountains just north of Cerro Gordo; San Lucas Canyon (S) at extreme left on range top. Dark rocks on range crest at left are Tin Mountain limestone overlying cliff-forming banded Devonian marble of the Lost Burro formation. Carboniferous strata underlie a large part of steep slopes in foreground. Note recent fault (R) in foreground at Saline Valley margin. H, Hart mine; BC, talc mines in Bonham Canyon; B, Belshaw spring; M, Mexican Spring; T, Summit station on salt tram; D, Daisy Canyon. Aerial photograph by J. H. Maxson.

pany. It led from a plant northwest of Swansea across the Inyo Mountains to Saline Valley.

The Cerro Gordo mining district is known above all for its yield of silver and lead, which reached a peak in 1874. From 1911 to 1919 carbonate zinc was likewise an important product. Gold and copper, recovered especially from certain of the siliceous ores in this district, were actually minor commodities and byproducts of lead and silver extraction.

Nonmetallic products of the region include salines and talc. For 60 years the salines were produced in various evaporating works near Keeler (Goodyear, 1888, p. 227; Gale, 1915, p. 253–264). Until about 1950 the Natural Soda Products Co. on the lakeshore south of Keeler produced soda ash and other byproduct salines. Salt shipments from Saline Valley via the 13-mile aerial tram were discontinued 20 or more years ago, and the tram and salt mills allowed to decay. Talc has been extensively prospected for in the southern Inyo Mountains. In recent years this commodity became the principal export (Page, 1951). The mill of the Sierra Talc Co. is at Keeler.

Other commodities for which prospecting has been done with indifferent success near Cerro Gordo are tremolite asbestos, beryllium, and tungsten. Whereas, like Darwin, the geologic environment of the southern Inyos appears favorable for tungsten, no such minerals have with certainty been recognized at Cerro Gordo. There is an unconfirmed report of scheelite in the Union tunnel.

Worthy of mention, though not in connection with gold, is the so-called Keeler gold mine and mill southeast of Keeler. During World War II this mill was reconditioned for concentration of tungsten ore from the Darwin district.

Since construction of the narrow-gage railroad in 1882–83, building-stone quarries have from time to time been worked in dolomite and marble along the west foot of the range. According to Knopf (1918, p. 123), stone from these quarries was used in construction of the Mills Building in San Francisco. Several Los Angeles buildings are said to have been faced with it (Merrill, 1903 p. 206–207; Hill, 1912). On the whole, the rock is too strongly fractured to provide good dimension blocks. Silurian dolomite at selected localities near Dolomite Station yields an attractive snowy-white product. After pulverizing, this material is currently shipped for use in terrazzo. Devonian limestone quarried at the Cerro Gordo mine has in recent years been transported by the tramline for commercial uses. The Union tunnel was used in the quarrying operation.

In past years a poor quality of red and green Triassic "slate" was quarried in Slate Canyon. It may have possible value for roofing granules.

PREVIOUS GEOLOGICAL STUDIES IN THE CERRO GORDO AREA

Study of Cerro Gordo geology may be said to have begun with a visit by Goodyear in 1870. Published in 1888 his results are accompanied by a crudely sketched geological cross section of the Inyo Range north of Cerro Gordo. The ore deposits and their geologic occurrence are first dealt with by Raymond (1873, p. 17–21), after an examination of the mine by him in 1872. Geology and ore deposits of the Inyo Mountains region are discussed in general terms by Fairbanks (1894, p. 473–475; 1896a, p. 150), and Gale (1914, 1915) described the occurrence of salines in neighboring Owens Valley and Saline Valley.

Geological work by Knopf at Cerro Gordo for the Geological Survey led to publication in 1914 of a preliminary paper on geologic structure and relations of the ore deposits, especially the zinc carbonate at that time undergoing active development. A second and more comprehensive report (Knopf, 1918) constitutes a geologic reconnaissance of the Inyo Range. Included is a review of the geology and ore deposits at Cerro Gordo. The work is amplified by very general reconnaissance geologic mapping and a treatment of the stratigraphy by Kirk (1918).

Occurrence and production of ores at Cerro Gordo and nearby mines are discussed from the engineer's viewpoint by Tucker and Sampson (1938, p. 431–433) and later by Norman and Stewart (1951, p. 58) of the California State Division of Mines. In 1951 an economic geology report on the talc deposits on the east side of the Inyo Range near Cerro Gordo was contributed by Page (1951, p. 23).

PRESENT INVESTIGATION

This study began in December 1942, as a U.S. Geological Survey lead-zinc strategic-minerals project. Geological mapping was at that time initiated by the writer in several parts of western Inyo County as a preliminary step in appraisal of lead and zinc resources of the region. Among areas in which work was started are the Ubehebe district, the Lee mine area, the Santa Rosa mine area, and the Cerro Gordo area. At Cerro Gordo the surface near the mine was mapped by planetable, and accessible underground openings were mapped. During 1944, when the mine was temporarily reopened, several visits were made by the writer for purposes of underground mapping.

Beginning in July 1946, a long-term quadrangle-mapping project was undertaken by the U.S. Geological

Survey in cooperation with the State of California. Coverage of a scheduled 15-minute quadrangle (New York Butte quadrangle) which includes the Cerro Gordo area was anticipated. Detailed geologic mapping on aerial photographs and measurement of stratigraphic sections through the southern Inyo Range were begun by the writer, who was joined in this undertaking during the fall of 1946 by W. C. Smith. Owing to numerous interruptions during which both authors were engaged in other work, these field studies were continued through the fall of 1948. Aerial photographs were used as field sheets until 1949, when the New York Butte quadrangle topographic sheet became available.

Geologic maps were first compiled on a planimetric base made from airphotos by the radial-line-strip technique. However, the general geologic map here included (pl. 2) has for its base the New York Butte quadrangle; except for the Cerro Gordo mining district (pl. 1), all geological features were transferred by inspection from the aerial photographs on which the mapping was done.

During July, August, and September 1950, June 1951, and June 1952, detailed geologic mapping was carried out by Merriam on the new topographic base, within the area covered by the accompanying Cerro Gordo mining district geologic map (pl. 1).

Scope and limitations.—The present contribution documents only part of the work carried out by the U.S. Geological Survey in the southern Inyo Mountains. It relates specifically to the geology of that part covered by the accompanying detailed geologic map (pl. 1), and especially to the immediate vicinity of the Cerro Gordo mine itself. Separate studies have been made by the writer on the stratigraphy and paleontology of the Paleozoic rocks (Merriam, 1954). Detailed studies of Triassic rocks and of the areal and economic geology of the entire New York Butte quadrangle were participated in extensively by other members of the U.S. Geological Survey parties, and are subjects of separate reports in preparation. With reference to the details of petrology and metamorphism little has been accomplished beyond the work of Knopf (1918).

Emphasis of this work was placed upon stratigraphy (Merriam, 1947) in relation to structure of a fairly sizeable area surrounding the Cerro Gordo mine in order that these data might be translated into problems of economic geology in the mine itself.

ACKNOWLEDGMENTS

The writer was assisted in the field during the course of this investigation by R. D. Nininger, R. L. Griggs, R. C. Douglass, L. S. McGirk, Jr., and E.M. MacKevett. In later phases of the fieldwork, W. C. Smith also took part in underground mapping at the Cerro Gordo mine and in surface mapping of adjacent parts of the New York Butte quadrangle. Laboratory assistance and consultation in connection with petrographic problems has been provided by E. M. MacKevett, J. G. Moore, and R. K. Hose.

During the course of this investigation several persons interested in Cerro Gordo mining and geological problems were most helpful in providing underground maps of partly inaccessible mine workings. Among these contributors are H. E. Olund, formerly of Imperial Metals Co. at Darwin, Calif., the late J. Percy Hart, J. J. Beeson, of Salt Lake City, Utah and J. B. Stone, of the Golden Queen mine, Mojave, Calif. The operators of the Golden Queen opened parts of the Cerro Gordo mine for examination in 1943 and 1944. Underground mapping of previously inaccessible workings on the lower mine levels was begun by the Geological Survey at that time. From 1946 until 1949 the mine was kept open by W. C. Rigg and associates. Access to the important "China stope" parts of which were reopened by Rigg, provided a rare opportunity to study and map in this section of the mine. Acknowledgment is made especially to W. C. Rigg and to J. B. Stone for use of camp facilities and for making available not only mine maps but earlier reports and pertinent information on history and production. Thanks are due them for permission to publish some of the data in connection with underground geology of the mines.

Among illustrations accompanying this report are photographs taken by L. G. Henbest and one oblique aerial photograph of the Inyo Range by J. H. Maxson.

GEOLOGIC SETTING

The southern Inyo Range comprises strongly folded and faulted sedimentary rocks which range in age from Ordovician to Middle Triassic (table 1; pl. 2). Except for the upper part of the Triassic sequence, these rocks are largely of marine origin and include limestone, dolomite, quartzite, and shale. The upper part of the Triassic sequence is predominantly volcanic with intercalations of terrestrial sediments. Paleozoic and Triassic rocks are penetrated by small- to intermediate-sized granitic bodies, by aplites, and by large numbers of andesitic dikes.

As might be anticipated from the number of plutonic bodies and dikes, the sedimentary rocks of this area have undergone varying degrees of contact and hydrothermal metamorphism. Foliation is, however, not pronounced though developed locally. Black shale like the Chainman shale is here and there phyllitic; carbonate rocks are not uncommonly changed to marble or calc-hornfels.

In terms of paleogeography the Paleozoic section ranging from Ordovician to Permian is quite representative of the Great Basin province. Considered jointly with Cambrian rocks exposed farther north in the range the Cerro Gordo Paleozoic column is one of the more inclusive of the southwestern Cordilleran belt, and as such has in recent years received a considerable amount of detailed stratigraphic and paleontologic study.

At the south end of the Inyo Mountains, the older rocks are largely covered by Tertiary volcanic rocks which are seemingly coextensive with those of the Coso Mountains region to the southwest. The volcanic rocks have been considerably faulted, but have not otherwise been appreciably deformed.

STRATIGRAPHIC GEOLOGY

The idea that Paleozoic sedimentary rocks of the southern Inyo Mountains embraced units quite similar to those in distant parts of the Great Basin has been current for many years and is expressed in an earlier publication (Kirk, 1918). The present study serves to support and to amplify this conception. Work by McAllister (1952, 1955, 1956) in the neighboring Ubehebe district demonstrates intimate stratigraphic and paleontologic relationships with central Nevada, especially with the Eureka district (Nolan, Merriam and Williams, 1956). Whereas in general these studies point up the provincial similarities from place to place in certain parts of the Great Basin column, they also bring out great local facies differences and add support to the existence of major lineal trends or axes on opposite sides of which the conditions of deposition and the faunas were in Paleozoic time quite different. For example, in the Ordovician of the west-central Great Basin, a western graptolitic shale and an eastern carbonate facies may be distinguished (Merriam and Anderson, 1942). In these terms the southern Inyo Mountains lie in the eastern or carbonate facies.

Significant local facies changes are well illustrated in the Silurian system by the nearly complete change in a northerly direction from dolomite at Cerro Gordo to limestone at Mazourka Canyon (fig. 3). Likewise between Cerro Gordo and the Ubehebe district the Devonian strata change from limestone to dolomite in an easterly direction.

Only the Paleozoic rocks of the area have been studied in detail by the writer. Work on the Triassic sedimentary and volcanic rocks was largely undertaken by other members of the Geological Survey parties and will be reported on separately.

Especially important at Cerro Gordo are the Devonian strata which are host rocks for the major ore bodies. During the course of this work, attention has been given the possibility of stratigraphic control of ore deposition, widely recognized in other lead-zinc districts where the ores occur as replacement of limestone.

The Paleozoic column at Cerro Gordo has an approximate total thickness of 11,100 feet. Individual system thicknesses are of the following orders of magnitude.

	Feet
Permian	2,000
Pennsylvanian	2,000
Mississippian	1,550
Devonian	1,800
Silurian	1,500
Ordovician	2,250

TABLE 1.—*Stratigraphic column of the Cerro Gordo mining district*

Age		Formation	Thickness (feet)	Lithology
Triassic		Unnamed rocks	4,000±	Andesite flows and pyroclastic rocks with intercalated red sandstone and shale.
				Marine shale and limestone.
		Unconformity		
Permian		Owens Valley formation	1,800±	Silty and sandy limestone, fusulinid limestone, siliceous conglomerate, limestone conglomerate, shale, siltstone, sandstone, and hornfels.
		Local unconformity		
Carboniferous	Pennsylvanian	Keeler Canyon formation	2,200±	Sandy and pebbly fusulinid limestone, shale, siltstone, and marble.
Carboniferous	Mississippian	Chainman shale	1,000±	Dark-gray silty shale and phyllite. Limestone interbeds.
		Perdido formation	0–200±	Limestone, chert, siltstone, and quartzite.
		Tin Mountain limestone	350	Dark-gray limestone, chert nodules.
Devonian		Lost Burro formation	1,600±	Light- and dark-gray marble, dolomite, quartzite.
Early Devonian and Silurian		Hidden Valley dolomite (Lower boundary difficult to establish in this area.)	1,700±	Massive light- and dark-gray dolomite, quartzite.
Ordovician		Ely Springs dolomite	240–550±	Light- and dark-gray cherty dolomite.
		Eureka quartzite	400±	Light-gray vitreous quartzite.
		Pogonip group (Basal part not exposed in this area.)	1,350±	Saccharoidal dolomite and limestone.

ORDOVICIAN SYSTEM

Strata of the Ordovician system comprise three units: Pogonip group, Eureka quartzite, and rocks provisionally referred to the Ely Springs dolomite. These rocks as exposed in the area under discussion do not include a full representation of the Ordovician system as it is known in this part of the Great Basin, for the lower part of the Pogonip group does not crop out. The boundary with the Silurian is uncertain here, as at many places in the Great Basin, and cannot be given a specific map delineation. For mapping purposes the Upper Ordovician beds assigned provisionally to Ely Springs dolomite are not differentiated from the Silurian and Lower Devonian Hidden Valley dolomite.

POGONIP GROUP

General features.—Rocks of the Pogonip group ranging in age from Early to Middle Ordovician have been redefined in the Eureka district of central Nevada (Nolan, Merriam and Williams, 1956, p. 23–29). In the type region the group embraces three formations which occupy the interval between the Upper Cambrian Windfall formation and the Eureka quartzite or its equivalent. No formational differentiation has been attempted in the southern Inyos, where it is well exposed in stratigraphic continuity only along the western foothills but is quite generally altered and not very fossiliferous. In the Ubehebe area (McAllister, 1952, p. 10) the group is better represented and would probably lend itself to stratigraphic subdivision. Here in the Inyos the Pogonip group is treated for preliminary mapping purposes as a single formation.

Areal distribution.—Only a small exposure of the Pogonip was recognized in the Cerro Gordo mining district (pl. 1); it lies on the south side of Bonham Canyon near the boundary of the older rocks with the heavy fanglomerate. At this place the Pogonip is overlain by Eureka quartzite on the east and is in fault contact with the Hidden Valley dolomite on the west. Along the western foothills of the range, extensive outcrops of the Pogonip are found at Smelter Hill, Front Ridge, the mouth of Brooklyn Canyon, and east of Granite Hill (pl. 2). Strata of this unit are well exposed to the north at Mazourka Canyon (fig. 3) and in the Talc City Hills at the south end of the Inyo Mountains (Hall and MacKevett, 1958, pl. 2).

Thickness and lithology.—Only a small part of the Pogonip is measurable in the fault block at Bonham Canyon. On the west side of the Inyo Range at Front Ridge, a thickness of about 1,350 feet is present, although the base is not exposed.

In the Front Ridge belt the Pogonip consists of thick-bedded blocky-weathering saccharoidal medium-gray dolomite and medium- to light-bluish-gray fine-grained marble and limestone. Interbedded with the limestone and marble is fine-textured dense quartzite. The blocky dolomite underlies the limestone and marble, the two units being separated by an interval of platy crinkly bedded brown-weathering siliceous shale or hornfels. Before alteration the hornfelsic beds were probably a cherty calcareous shale with fine-grained cherty limestone intercalations.

At Front Ridge where the Pogonip section is more continuous, it has been subdivided stratigraphically on an informal basis. The lower blocky dolomite is referred to as Pogonip A, the intermediate shaly horn felsic beds as Pogonip B, and the upper limestone and marble unit as Pogonip C. Locally the dolomitization has affected Pogonip C as well as Pogonip A.

At Bonham Canyon, Pogonip C is an iron-stained medium- to light-gray sugary dolomite with patches which still retain a slightly bluish color. *Receptaculites* and large gastropods of the genera *Maclurites* and *Palliseria* are present.

Age and correlation.—Where the Pogonip group is best shown in the central Nevada region, it includes at least three formations (Nolan, Merriam and Williams, 1956, p. 23–29). The lowest unit, known as the Goodwin limestone, is of Early Ordovician age, the middle or Ninemile formation is also Early Orodovician, and the upper or Antelope Valley limestone is of Early and Middle Ordovician age. In the southern Inyo Mountains all fossils obtained form the Pogonip are from Pogonip C, the upper unit, which correlates with the Antelope Valley limestone of central Nevada. The large gastropods *Maclurites* and *Palliseria*, together with *Receptaculites*, are indicative of the *Palliseria* zone, which is the middle of three faunal zones in the Antelope Valley limestone. A brachiopod fauna, probably representing the lower or *Orthidiella* zone, was found on the west side of the Inyos east of Granite Hill (loc. 40). In the Ubehebe area (McAllister, 1952, p. 11), faunas representing the intervals of the Goodwin limestone and Ninemile formation have been collected. On lithologic grounds it appears likely that Pogonip B of the Inyos is roughly correlative with the Ninemile.

EUREKA QUARTZITE

General features.—The Eureka quartzite of late Middle to Late(?) Ordovician age is one of the more useful stratigraphic keys in the Inyo Mountains and the Great Basin generally. Named for the Eureka district, Nevada (Nolan, Merriam and Williams, 1956, p. 29), this resistant light-colored vitreous quartzite overlies the Pogonip and is overlain by usually dark-gray dolomite or limestone of Late Ordovician (Richmond) age.

Areal distribution.—The Eureka quartzite has a limited distribution in the area under consideration. It rests upon Pogonip on the south side of Bonham Canyon and occupies a larger area between Bonham Canyon and San Lucas Canyon (pl. 1). Good exposures may be seen on the road along San Lucas Canyon at a point 2 miles northeast of Cerro Gordo. The Eureka crops out much more extensively in the western foothills of the Inyo Range between Smelter Hill and Granite Hill (pl. 2), reappearing to the north at Mazourka Canyon (fig. 3).

Thickness and lithology.—On the west side of the Inyo Mountains, the Eureka quartzite has an average thickness of about 400 feet. It is characteristically a clean sugary dense vitreous quartzite in which clear rounded grains are easily recognized. Normally the texture is medium to fine with the grains cemented by siliceous matter. Color ranges from nearly white to light brown and dark grayish brown; weathered surfaces show varying degrees of limonitic staining, depending upon the amount of iron impurity. Bedding ranges from massive to platy, but with heavier bedding predominant. Crossbedding is fairly common, especially in the upper part of the formation. Recrystallization and additive silicic alteration have not uncommonly obliterated original sand-grain borders, giving the rock a finely granular, almost homogeneous appearance.

The Eureka quartzite is usually noncalcareous, unlike the otherwise quite similar Silurian and Devonian quartzite of this region; these contain varying amounts of calcite or dolomite as cement or as actual interbeds. At Mazourka Canyon (fig. 3), however, there are dolomite interbeds in the Eureka, which is exceptional.

Throughout its extent in the Great Basin, the Eureka quartzite rests on rocks ranging in age from Late Cambrian to Middle Ordovician. The basal contact is at some points a disconformity, although no evidence of such break was noted in the Inyos.

The Eureka quartzite of the southern Inyos lends itself to stratigraphic zonation based on lithology and color differences. As in many sections of this formation, the lower part, ranging in thickness from 125 to about 200 feet, is darker, more deeply iron stained, and tends to be less heavily bedded (Kirk, 1933, p. 28, 30), whereas the upper part is more massive, predominantly white or lighter gray and more commonly crossbedded.

In San Lucas Canyon and the west Inyo foothills, the Eureka includes a basal dark-gray angillaceous member varying from 40 to about 75 feet in thickness. It comprises shale, silty shale, and fine sandstone. At Mazourka Canyon (fig. 3) these beds are fossiliferous and have been assigned to the "Barrel Spring formation" (Phleger, 1933, p. 5). As elsewhere in the Great Basin the lower shaly beds are discontinuous, being wholly absent at some Eureka outcrops of the Inyo Mountains and the Ubehebe area. In central Nevada (Nolan, Merriam and Williams, 1956, p. 30) the Eureka quartzite interval is locally more than half occupied by such deposits, which will be designated as a new formation in a forthcoming report.

Age and correlation.—The Eureka quartzite lies between Pogonip beds of Middle Ordovician age and strata which carry a Late Ordovician (Richmond) fauna. Where the lower shaly facies of the Eureka is present, it carries faunas of about middle Trenton age, thus indicating that the Eureka interval represents some part of late Middle and possibly early Late Ordovician time.

ELY SPRINGS DOLOMITE

General features.—Dark-gray, often quite cherty, dolomite overlying the Eureka quartzite in the southern Inyo Mountains is provisionally referred to as Ely Springs dolomite, a name originally given to similar rocks in the distant Pioche region of southeastern Nevada (Westgate and Knopf, 1932, p. 15). The dark-gray, at some places almost black, Late Ordovician Ely Springs band contrasts sharply with the light-gray Eureka where seen from a distance.

Unlike the base, where the lithologic change is abrupt, the top of the Ely Springs is not readily definable in this area. Very similar types of dark cherty dolomite are in fact repeated within the overlying Hidden Valley, and the Ordovician-Silurian boundary appears to be definable only on a paleontologic basis. This factor together with strong deformation and mild alteration of the rocks in question leads the writer to the conclusion that separation of the Ely Springs from the Hidden Valley for mapping purposes is at present impracticable.

Areal distribution.—The Ely Springs dolomite is exposed 1½ miles southeast of Bonham's talc mines (pl. 1). It is best shown, however, in the rugged western foothills of the range between Granite Hill on the north and Smelter Hill on the south (pl. 2). Equivalent strata given local names from one region to another are represented throughout much of the southern Cordilleran belt. Although generally associated with the Eureka quartzite, these strata extend far beyond the limits of even that widespread formation.

Thickness and lithology.—Thickness of the Ely Springs cannot be given accurately because of uncertain position of its upper boundary in this region. At Mazourka Canyon (fig. 3), where the top is placed just beneath a massive 20-foot chert member, it is about 270 feet thick. At Smelter Hill the formation is at least 240 feet thick, but the massive chert bed was not recognized.

At Smelter Hill the lower 150 feet of the Ely Springs is a dark-gray very cherty saccharoidal dolomite, chert constituting ⅕ to ¼ of the entire rock. The chert is usualy gray and occurs as very irregular nodules or as lenses elongated with bedding. Recrystallization of the chert has commonly led to development of felted aggregates and rosettes of lime-silicate minerals. The nodular cherty dolomite is inclined to be rather thinly bedded. Toward the top of the lower 150-foot member, the amount of chert decreases with passage upward into a less dark gray almost noncherty dolomite member about 90 feet thick. This higher member is heavier bedded, rather coarse textured or sugary granular, and blocky weathering. Above this member the blocky dolomite becomes light gray or nearly white in the vicinity of Smelter Hill and is classified for the greater part with the Silurian part of the Hidden Valley.

Age and correlation.—Strata to which the name Ely Springs dolomite is applied at Cerro Gordo are similar lithologically to those which rest upon the Eureka quartzite at many points throughout the central and southern Great Basin. These rocks generally contain Richmond Late Ordovician faunas. To the northeast in Utah the name Fish Haven dolomite is used for virtually the same unit; in central Nevada, where limestone locally takes the place of dolomite, it is the Hanson Creek formation. The Montoya limestone of New Mexico is likewise similar both lithologically and faunally.

Loose streptelasmid horn corals, probably derived from this unit, were collected in the area between San Lucas Canyon and Bonham Canyon (pl. 1). These corals came from the vicinity of talc prospects in the Ely Springs dolomite adjacent to the large Eureka quartzite exposure. Scarcity of good fossils is explained in part by rock alteration. Convincing evidence of Richmond age is found in Ely Springs dolomite of the Talc City Hills (fig. 3) at the south end of the Inyo Mountains. About 2½ miles northwest of the Talc City mine, the writer collected silicified corals and brachiopods at locality 39; representative Ely Springs fossils from this locality are listed as follows:

Halysites (*Catenipora*) sp.
Columnaria cf. *C. alveolata* (Goldfuss)
Streptelasmid corals, several types
Heterorthis sp.
Glyptorthis cf. *G. insculpta* (Hall)
Thaerodonta sp.
Lepidocyclus (at least two species)
Platystrophia sp.
Onniella cf. *O. quadrata* Wang
Zygospira n. sp.
Strophomena sp.
Plaesiomys sp.

SILURIAN SYSTEM

A dolomite-quartzite sequence about 1,500 feet thick is assigned to the Silurian system. These strata occur between Late Ordovician Ely Springs dolomite and rocks of established Devonian age. In the southern Inyo Mountains no objectively mappable contact with the Ely Springs was recognized, and the Silurian-Devonian relation appears to be one of transition. Deterrents to fixing of these systemic limits are such factors as scarcity of fossils, shearing, fracturing, and rock alteration; especially is this true along the Silurian-Devonian boundary zone at Cerro Gordo. Hence the system boundaries both below and above remain indefinite.

Complications noted in connection with the differentiating of Silurian rocks from those of the Ordovician and Devonian are by no means confined to this region; such difficulties are encountered generally in the Great Basin, where the middle Paleozoic is normally dolomite containing few good fossils.

Silurian rocks of the adjoining Ubehebe district are described by McAllister (1952, p. 15) as the Hidden Valley dolomite. However, the typical Hidden Valley as originally defined is not entirely Silurian, but includes Lower Devonian (Oriskany) dolomite at the top. For purposes of this report the term Hidden Valley dolomite is applied, in the original sense, to Silurian and Lower Devonian strata of the southern Inyo Mountains. Later work may well point up the desirability of separating the Devonian part, either as an independent formation, or perhaps by considering it a member of the overlying Devonian Lost Burro formation.

Facies changes complicate the Silurian stratigraphy of this region, as well illustrated by nearly complete change from dolomite at Cerro Gordo to limestone on the north at Mazourka Canyon (fig. 3). Also noteworthy is the introduction of thick local quartzite units west of the Ubehebe district.

For purposes of map representation the Hidden Valley dolomite and the Ely Springs dolomite are not differentiated at Cerro Gordo.

HIDDEN VALLEY DOLOMITE

General features.—The Hidden Valley dolomite, comprising dolomite and quartzite, lies between the Ely Springs dolomite and the Devonian Lost Burro formation. This discussion applies mainly to the Silurian part, which constitutes most of the formation. Details of the Early Devonian part of the Hidden Valley are dealt with more fully under rocks of the Devonian system.

Type section of the Hidden Valley dolomite is in the Ubehebe district, 20 miles northeast of Cerro Gordo (McAllister, 1952, p. 15). At Cerro Gordo the formation differs from the type section by including considerable amounts of quartzite and chert (Merriam, 1951); it has furthermore been subjected to rather widespread hydrothermal alteration, with local development of commercial talc deposits.

Areal distribution.—The Hidden Valley dolomite crops out over a large area on the east limb of the Cerro

Gordo anticline, extending from the north side of Bonham Canyon southward to the Belmont mine area on the east side of the range (pl. 2). The formation is well exposed along the western Inyo foothills from Smelter Hill through Front Ridge. A separate belt extends from the Dolomite Hills along the east side of Brooklyn Canyon to the mouth of Long John Canyon. Lack of outcrop prevents tracing of this dolomite into the seemingly equivalent Silurian limestone facies at Mazourka Canyon, 12 miles to the northwest.

Thickness and lithology.—The Hidden Valley dolomite in this area is roughly 1,750 feet thick. Of this, roughly 1,500 feet is Silurian, the remainder of Early Devonian age.

The formation consists of light- to medium-gray blocky dolomite, dark-gray cherty dolomite, arenaceous dolomite, and quartzite. Heavy-bedded blocky jointed saccharoidal dolomite makes up at least half the unit and ranges from medium and light gray to white on fresh fracture. A fine banding or lamination is noted here and there. Brown iron-stained dolomite lenses attain a thickness of 25 feet and a lateral extent of several hundred feet. Iron-stained fractures and joints are common. In the upper part of the formation, light-gray dolomite contains abundant crinoidal debris. When pulverized some of the purer Hidden Valley dolomite is snowy white and has long been quarried for terrazzo and other commercial purposes.

Dolomite of the Hidden Valley resembles that of the Pogonip group, but on the whole is inclined to be lighter gray.

As exposed in the Salt Mill Hills (pl. 2), the dark-gray cherty dolomite tends to exhibit thinner and more distinct bedding than noncherty phases. The dark-gray to black chert occurs in thin irregular nodules or discontinuous interbeds a few inches thick. Locally the siliceous dolomite has been altered to talcy or tremolitic material, in fact some of the minable talc deposits are an alteration product of this Silurian facies. The dark cherty Silurian dolomite is lithologically indistinguishable from that of the Ely Springs dolomite without benefit of stratigraphy.

Quartzitic rocks of the Hidden Valley formation range from dense vitreous white or light-gray types resembling Eureka quartzite to arenaceous dolomite containing varying proportions of rounded quartz grains. Bedding ranges from heavy in the vitreous quartzite low in carbonate cement to relatively thin, as where limy quartzite and dark gray cherty dolomite are interbedded. The quartz grains are well rounded and medium to fine. Quite characteristic is the non vitreous limy quartzite with white aphanitic carbonate cement. All gradations are found from this type to white dolomite with scattered quartz grains. On the west side of the Inyo Range at Smelter Hill and Salt Mill Hills four lithologic units can be defined in the Hidden Valley dolomite as follows:

TABLE 2.—*Members of the Hidden Valley dolomite, Salt Mill Hills*

Age	Formation	Member	Lithology	Thickness (feet)
Middle and Late Devonian	Lost Burro formation			
		(Fault relation)		
Early Devonian	Hidden Valley dolomite 1,740 ft.	D	Light- to brownish-gray blocky saccharoidal dolomite.	150
Silurian		C	Light-gray and white quartzite, limy quartzite, and arenaceous dolomite.	540
		B	Medium- to dark-gray well-bedded cherty dolomite.	250
		A	Light-gray to white thick-bedded blocky saccharoidal dolomite.	800
		(Transitional relation)		
Late Ordovician	Ely Springs dolomite			

Between Bonham Canyon and the Belmont mine on the east side of the Inyo Range (pl. 1), total thickness of the Hidden Valley is roughly commensurate with that tabulated above for the western foothills. Details of the stratigraphy differ somewhat. Rock types which characterize members B and C to the west appear to be mixed here in a sequence of interbedded vitreous quartzite, limy quartzite, and dark-gray dolomite, part of which is cherty. Quartzite zones range from a few inches to several feet in thickness but reach a maximum of about 45 feet in a few places. This thickness is less than the maximum on the west side at Salt Mill Hills, where the quartzite zones are on the whole more continuous laterally. In the eastern sequence dark-gray dolomite and well-bedded cherty dolomite appear to predominate over quartzite, individual cherty dolomite lenses reaching a maximum thickness of about 60 feet. Blocky light-gray saccharoidal dolomite corresponding to member A underlies the quartzite-cherty dolomite interval, while above is the upper blocky dolomite of member D with its contained crinoidal debris. Details of the stratigraphy are interpreted with difficulty because of the sheared and locally altered condition of these strata in Bonham Canyon. It is quite evident, however, that individual quartzite and cherty dolomite bodies pinch out rapidly in the depositional sense. To the east the quartzites apparently disappear before reaching the Ubehebe area (fig. 3), while to the west toward Owens Valley, these highly siliceous deposits are even more extensively developed. Northwestward in the direction of Mazourka Canyon, they again disappear from the section, as the Hidden Valley dolomite changes to limestone.

Age and correlation.—In Bonham Canyon the cherty dolomite associated with quartzite at locality 32 has yielded an excellent fauna of silicified material including corals and brachiopods which indicate a Middle Silurian age. The highly disturbed beds from which the material came are believed to represent either member B or member C, probably above the middle of the formation. Among the common forms are species of *Halysites*, *Heliolites*, and *Favosites*, together with abundant *Atrypa*, a large smooth pentameroid (possibly *Pentamerus*) and *Schizoramma* sp. The *Schizoramma* is significant for it resembles a species from the Brownsport formation of Tennessee (Amsden, 1949, p. 45), which is believed to be of Niagaran age.

Elsewhere in the southern Inyo Mountains, fossils are extremely rare in the Hidden Valley dolomite. They are locally abundant and are silicified in dolomite of the type area (McAllister, 1952, p. 16). The faunas include *Porpites*, *Diplophyllum cf. D. caespitosum*, *?Catazyga* sp., and large conical dasycladacean algae. These algae, to which the name *Verticillopora annulata* has been given by Rezak (1959), resemble closely those found at several other localities in Great Basin Silurian rocks.

Where the Hidden Valley changes northward to limestone at Mazourka Canyon, the rocks are abundantly fossiliferous, but the material is poorly preserved. The faunas consist very largely of corals, only *Atrypa* and rhynchonellids (*Eatonia bicostata* Stauffer) being at all common among the brachiopods. The large dasycladacean algae (*Verticillopora annulata* Rezak) are most prolific here and provide a tie with the Hidden Valley of the type areas as well as with the Roberts Mountains formation of central Nevada and the Laketown dolomite of western Utah.

Among corals of the limestone facies at Mazourka are many conforming to the general features of *Strombodes*. Others are assigned to *Chonophyllum*, *Rhizophyllum*, *Heliolites*, *Alveolites*, and *Cladopora*. Also present are large cyathophyllids and bushy forms of the *Phacelophyllum* and *Disphyllum* types.

The Silurian part of the Hidden Valley dolomite of the Inyo Mountains probably ranges in age from Early to Late Silurian and seemingly correlates with the combined Roberts Mountains formation and Lone Mountain dolomite of central Nevada (Nolan, Merriam, and Williams, 1956, p. 36). As mentioned above, the Hidden Valley of the original definition (McAllister, 1952. p. 17) also includes Early Devonian. The age of these uppermost beds is dealt with under the section "Devonian system."

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DEVONIAN SYSTEM

Banded limestone and marble of Devonian age make the impressive east-facing cliffs and very rugged upper east slopes of the Inyo Range stretching northward from Cerro Gordo. Rocks of this system underlie roughly one-fifth of the area under consideration and are especially important as host rock of the major ore bodies. The Devonian system is represented by about 1,800 feet of strata, a thickness that exceeds the Silurian by a possible 300 feet (fig. 5). Unlike the Silurian of the Cerro Gordo area only a small part of the Devonian is dolomitized.

Two formations are involved, the Lost Burro formation, which includes most of the system, and an underlying Lower Devonian zone, which occupies the upper part of the Hidden Valley dolomite.

Strata previously assigned to the Devonian in this region (Kirk, 1918, p. 36; Stauffer, 1930, p. 8–91) are largely, if not entirely, Silurian (Nolan, 1943, p. 153) and are considered under treatment of the Silurian part of the Hidden Valley dolomite. Certain other strata at Cerro Gordo previously included with the Carboniferous (Knopf, 1918, p. 110) are actually Devonian.

The Devonian section of the southern Inyo Mountains is about half the thickness of that in the central Great Basin. Fossil evidence as well as lithology suggests that most of the attenuation may well be in the middle part of the system at Cerro Gordo.

Boundary of the Devonian with the Mississippian is marked by an abrupt and readily mappable lithologic

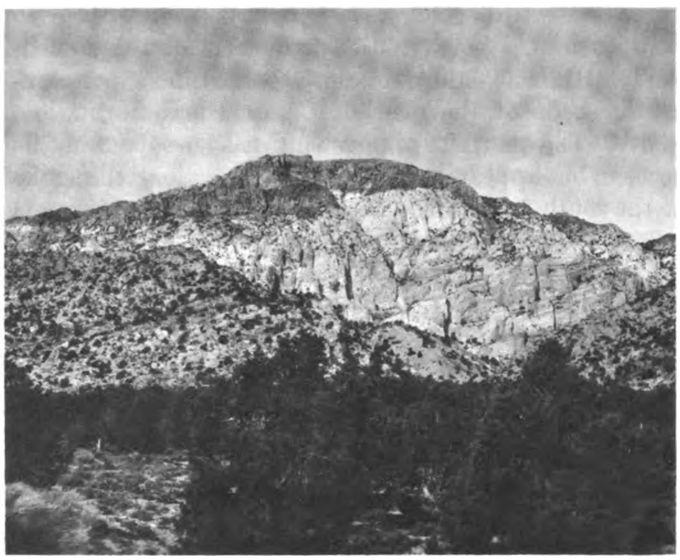

FIGURE 5.—View looking northwest at the top of the Inyo Mountains 2 miles north of Cerro Gordo. Highest dark-gray strata are Tin Mountain limestone resting conformably on cliff-forming Devonian marble of the Lost Burro formation, host rock of the principal ore bodies at Cerro Gordo. Foreground underlain by Hidden Valley dolomite.

change and by introduction of wholly new faunas. On the other hand the boundary with the Silurian remains uncertain, may be transitional, and falls somewhere within the upper 400 feet or so of the Hidden Valley dolomite wherein the Early Devonian Oriskany fauna is to be expected.

Except for a few clean, washed quartz sands or quartzite the Devonian rocks of this region are almost entirely carbonate.

UPPER PART OF HIDDEN VALLEY DOLOMITE (DEVONIAN PART ONLY)

General features.—Fossil evidence shows conclusively in the type area (McAllister, 1952, p. 15, 17) that at least part of the upper 400 feet of the Hidden Valley is Early Devonian. Fossils are few in these rocks at Cerro Gordo, but largely on the basis of lithology the upper 250 to 350 feet of this formation (member D of the Hidden Valley) is likewise provisionally classified as Early Devonian.

Areal distribution.—Rocks classified as member D of the Hidden Valley are shown along Bonham Canyon in the vicinity of the talc mines, and on the east side of San Lucas Canyon near the Perseverance mine. On the west side of the Inyo Range, these strata may be observed in the Salt Mill Hills and in the rugged terrane east of Brooklyn Canyon.

Lithology and stratigraphy.—Upper beds of the Hidden Valley dolomite (member D) may be observed near the trail from San Lucas Canyon to the Perseverance mine. About 800 feet northwest of the mine (loc. 25) these beds with a thickness of about 350 feet appear to be transitional between the main part of the saccharoidal Hidden Valley dolomite below and *Stringocephalus* beds of the overlying Lost Burro formation, above. The quartzite of member C has lensed out in this section, but appears again within a mile along the strike to the south, toward the Belmont mine. The member D or transition beds comprise blocky medium- to light-gray dolomite, partly dolomitized bluish-gray limestone, and near the bottom thinner bedded light-gray-weathering very fine grained dolomite which contains abundant poorly preserved high-spired gastropods of the *Loxonema* type (loc. 25).

Upper beds of the Hidden Valley (member D) are well represented near the talc mines on the northeast side of Bonham Canyon (pl. 1). In this belt they have been strongly sheared and locally involved in the talc mineralization, but give the impression of a transitional interval between the middle quartzite beds of the Hidden Valley and the Lost Burro formation. On the west side of the range at Brooklyn Canyon, the strata of member D are similarly transitional.

Age and correlation.—The upper beds of member D of the Hidden Valley lie between the *Stringocephalus* beds of the lower part of the Lost Burro, which are of late Middle Devonian age, and the quartzite beds of the Hidden Valley (member C), which are of Niagaran (Silurian) age. In the Ubehebe district (McAllister, 1952, p. 17) the upper beds of the Hidden Valley in question have yielded an excellent silicified fauna of Oriskany (Early Devonian) age identical with that from the lower part of the Nevada formation in central Nevada (Merriam, 1940, p. 50). The Oriskany fauna includes "*Spirifer*" *kobehana*, "*Spirifer*" cf. "*S.*" *arenosus*, and *Papiliophyllum elegantulum* Stumm. Whereas member D of the Hidden Valley is thus directly correlated with the lower part of the Nevada formation, there appear to be neither beds nor faunas in the area under discussion which correspond to the overlying middle part of the Nevada.

The gastropod-bearing very fine grained dolomite near the bottom of member D on the Perseverance mine trail agrees lithologically with the Beacon Peak dolomite member of the Nevada formation in the Eureka district (Nolan, Merriam and Williams, 1956, p. 42) and with the Sevy dolomite of western Utah. Aphanitic well-bedded dolomite of this type appears in fact to be rather characteristic of the Lower Devonian in the Great Basin. At Eureka it rests disconformably upon the blocky saccharoidal Lone Mountain dolomite which correlates with at least the middle part of the Hidden Valley. Large *Loxonema*-like gastropods in member D are indeterminate but resemble forms which occur in the lower and middle parts of the Nevada formation.

LOST BURRO FORMATION

General features.—The name Lost Burro has been applied to Devonian strata in the Ubehebe district (McAllister, 1952, p. 18), which occupy the interval between Hidden Valley dolomite and the Tin Mountain limestone of Mississippian age. Type section of the formation is at Lost Burro Gap, Ubehebe Peak quadrangle. In the type area these rocks are largely dolomite, differing in this respect from approximately equivalent beds of the southern Inyos, which are in the main rather pure limestone or marble. The Lost Burro includes most of the Devonian rocks in this area, resting apparently without break upon the lower Devonian uppermost part of the Hidden Valley dolomite.

This important ore host was long and appropriately known to the mining profession as "Cerro Gordo marble", a term now formally eliminated by published adoption of Lost Burro; moreover the name "Cerro

Gordo" has nomenclatorial priority elsewhere, having been published much earlier in application to other formations.

Areal distribution.—The Lost Burro formation has been mapped from the Belmont Mine area on the east side of the range, along the higher more cliffy slopes to Bonham Canyon (pl. 1). Northward from Bonham Canyon the main marble belt of the Lost Burro departs eastward of the Inyo crest to occupy terrane of intermediate and lower altitude through the Daisy Canyon drainage to Hunter Canyon (fig 3). North of Hunter Canyon, near the east foot of the Inyo Range, marble of the Lost Burro formation extends into a marble-dolomite-calc-hornfels complex and is with difficulty distinguishable from the Silurian and older rocks. From Cerro Gordo north for 6 miles, the strikingly banded white, gray, and bluish-gray marble of this division is sculptured in rugged cliffy exposures which face east and rise abruptly above less steep terrane formed by the blocky Silurian and Lower Devonian dolomite.

Marble of the Lost Burro reappears at the Lee mine on Lee Flat, east of the Inyo Mountains (fig. 3).

In the west Inyo foothills a narrow belt of thoroughly marbleized and locally more strongly altered Lost Burro has been mapped for 6 miles along the east side of the Salt Mill Hills through the very steep terrane east of Brooklyn Canyon (pl. 2). At Mazourka Canyon (fig. 3) to the northwest, the formation appears to be entirely absent, having been removed by erosion at the unconformity which separates Chainman shale and Perdido from the Silurian limestone.

Thickness.—The Lost Burro formation, measured in detail on the northwest side of Cerro Gordo Peak, is about 1,600 feet thick. North of Cerro Gordo and in the west Inyo foothills where these beds are highly deformed, an assumed average thickness of 1,500 feet accords with outcrop width. McAllister (1952, p. 18) has measured 1,525 feet of this unit south of Lost Burro Gap in the Ubehebe district. Sections measured elsewhere in that area by McAllister exceed 2,000 feet.

Lithology.—The Lost Burro formation near Cerro Gordo is prevailingly massive craggy cliff-forming marble and limestone of white to bluish-gray color. Locally the color ranges from white to light bluish-gray to dark gray. Viewed from the east the finely sculptured cliffs and crags of Lost Burro north of Cerro Gordo present a striking pattern, with intricate bands and patches of contrasting gray and white, an aspect peculiar to the formation in this area. In the duller gray blocky Silurian dolomite below, the outcrops are more uniform, making on the whole less precipitous staircaselike, slopes.

Although this formation is predominantly thick-bedded, there are subordinate flaggy and platy limestone units. At some points thick-bedded marble shows a varvelike millimeter lamination.

The marble or limestone of the Lost Burro formation tends to be of fairly uniform finely crystalline to subporcellaneous texture and weathers smooth. Small bodies of more coarsely recrystallized marble are present. Except along fissures and fractures, where additive metasomatic activity has taken place, the rock consists mainly of calcite and has been used commercially where pure calcium carbonate is required. With its normally low content of clay, silica, iron, and magnesium, the Lost Burro contains the purest limestone of the region. In the southern Inyo Mountains, unlike the Ubehebe district (McAllister, 1952, p. 18), the Lost Burro is mainly nondolomitic. Dolomitization was noted only near the base in a transition interval between this formation and the normal upper dolomite of the Hidden Valley. Here and there in the transition zone, the bluish-gray smooth-weathering limestone exhibits irregular patches and tongues of probably diagenetic dolomite wherein the bluish gray is lost and the rock becomes saccharoidal, weathering with a rough sandy surface.

The Lost Burro includes several siliceous zones in which quartz sand grains constitute the siliceous material. Chert is extremely rare in this formation at Cerro Gordo. On the contrary in the Ubehebe area McAllister (1952, p. 18) finds abundant chert in a lower 155-foot sandy dolomite not recognized in the Inyo Mountains. At Cerro Gordo, hydrothermal jasperization has taken place locally along fractures. Absence of chert is significant, for the overlying Tin Mountain limestone with abundant chert may on this basis be distinguished.

Quartzite and sandy limestone characterize the lower part of the Lost Burro but actually constitute an insignificant part of the unit in terms of volume. The quartzite or sandstone beds range in thickness from a few inches to a maximum of 25 feet. Thin quartzite beds in places can pass laterally into inconspicuous limy quartz sandstone and limestone with scattered quartz grains. The standstone and quartzite are well sorted, showing medium to coarse rounded quartz grains. Northeast of the Cerro Gordo mine the thicker quartzite beds crop out as prominent ribs but vary greatly in density and thickness from point to point. Where dense and vitreous, these deposits are light gray or white, resembling vitreous Eureka quartzite or the denser quartzite of the Hidden Valley formation. Sporadic appearance of these clean quartz sands as lenses or beds in an otherwise rather pure limestone is somewhat enigmatic, suggesting aeolian introduction.

At Cerro Gordo the lower part of the Lost Burro is abundantly fossiliferous. A great part of the limestone is of organic origin as attested by the biohermal masses of stromatoporoids and "spaghetti coral." Not uncommonly these coral-rich lenses are dark gray, owing to their high carbon content and contrast sharply with light-gray barren limestone which surrounds them. Color banding and streaking previously referred to seems to be a matter of carbon distribution related to factors of sedimentation.

At many points in the marble of the Lost Burro, intense compressional deformation is revealed by intricate minor flexures. In the vicinity of Cerro Gordo a steeply dipping fracture cleavage or sheeting shows prominently in the Lost Burro formation. These features, together with jointing, have influenced sculpture considerably, especially as noted in the higher summits and crags north of Cerro Gordo. The sheeting may easily be mistaken for bedding in the massive marbles. True bedding is revealed where these superimposed structural features intersect quartzite interbeds.

Stratigraphy.—The Lost Burro formation is underlain with seemingly transitional relation by the Lower Devonian upper part of the Hidden Valley and is overlain with apparent conformity but abrupt lithologic change by the Mississippian Tin Mountain limestone. The Tin Mountain differs from the Lost Burro by being uniformly dark gray and containing much chert. Whereas no evidence of disconformity was found at this boundary, the sharp lithologic change and the abrupt introduction of a Madison fauna suggests a hiatus.

Stratigraphic control in the Lost Burro is especially important as an aid to underground structural interpretation at the Cerro Gordo mine, largely developed in this formation. Coral beds and quartzite exposed in San Lucas Canyon just north of the mine would be expected to have value as underground structural keys.

A lower lithologic zone (zone A) and an upper lithologic zone (zone B) are recognized in the Lost Burro formation of the Cerro Gordo mine vicinity. Zone A is 575 feet thick, as measured on the east side of San Lucas Canyon; zone B is 1,025 feet thick and embraces the remainder of the formation.

Darker bluish gray marbles and limestone are conspicuous in zone A, predominating over the light gray and white phases. In the lower 200 feet spotty incipient dolomitization causes bleached patches in the normally bluish-gray marble. An important paleontological datum is the *Stringocephalus* bed which immediately overlies the transitional interval showing patchy dolomitization.

Quartz sand and quartzite members characterize the upper 250 feet of zone A, the most conspicuous being a 23-foot member and a higher 6-foot bed at the top of the zone. Sandstone layers ranging in thickness from 1 inch to about a foot are fairly common. Scattered quartz sand grains occur in many of the limestone layers.

Zone B, the upper of the two, exhibits a predominance of the light-gray to white heavy-bedded cliffy marble streaked and banded with bluish gray. Siliceous sandstone layers are rare, though one a few inches thick occurs 640 feet below the top of the formation. "Spaghetti coral" and stromatoporoid beds are less numerous than in zone A, but dark-gray "spaghetti" beds occur sparingly up to an horizon 500 feet above the bottom of zone B.

Possible value of these stratigraphic zones for purposes of structural geology is considered below under geology of the Cerro Gordo mine.

Age and correlation.—*Stringocephalus*, a diagnostic late Middle Devonian brachiopod, occurs near the base of the Lost Burro north of Cerro Gordo. Excellent faunas of Late Devonian age were collected from the upper part of this unit in the Ubehebe district. Thus the Lost Burro formation ranges in age from late Middle to Late Devonian.

Coral and stromatoporoid remains form much of the limestone in zone A, the lower part of the formation. The "spaghetti coral" limestone of this division is made up in part of poorly preserved slender *Cladopora*, but probably includes *Amphipora*, a digitate branching stromatoporoid common in beds of equivalent age throughout the Great Basin.

Stringocephalus serves to correlate the basal Lost Burro with the upper part of the Nevada formation in central Nevada (Merriam, 1940, p. 24), where this genus ranges through about 400 feet of strata.

Some 1,300 feet of the Lost Burro above *Stringocephalus*, roughly three-fourths of the formation, correlates with the Devils Gate limestone of central Nevada, and agrees with it remarkedly well in thickness. In the Eureka district, Nevada, the Devils Gate limestone averages about 1,300 feet in thickness and rests upon the uppermost member of the Nevada formation containing *Stringocephalus*. As at Cerro Gordo, "spaghetti coral" and stromatoporoids are builders of limestone in the lower part of the Devils Gate.

Poorly preserved gastropods collected 2,000 feet northwest of the Newsboy mine (loc. 26) are possibly *Orecopia mccoyi* (Walcott), a characteristic species of the *Spirifer argentarius* zone in the upper part of the Devils Gate limestone. Late Devonian species which characterize the *Cyrtospirifer* zone of the uppermost Devils Gate have not been found in the southern Inyos. Fossils of this zone, to be expected near the top of the Lost Burro, have been found by McAllister

(1952, p. 18) in the Ubehebe area, where they occur in an upper 35-foot sandy dolomite and sandstone bed (bed 5). The typical Lost Burro of the Ubehebe area differs by being to a considerable extent dolomite. Its thickness is, however, comparable to that of the limestone in the Lost Burro formation near Cerro Gordo, and like the Cerro Gordo section the late Middle Devonian *Stringocephalus* zone occurs near its base.

Unrecognized in the southern Inyos and at Ubehebe are the distinctive *Martinia kirki* and *Spirifer pinyonensis* faunas; in central Nevada these occupy zones within the middle and lower parts of the Nevada formation. Comparative thicknesses suggest absence of beds representing these zones at Cerro Gordo, where the entire Devonian is only about 1,800 feet thick, about half that of the central Nevada region. Oriskany Lower Devonian fossils in the upper Hidden Valley dolomite at Ubehebe are nonetheless clear evidence that strata equivalent in age to the lowermost Nevada are represented. Further details relating to paleontology and correlation of the Lost Burro and the Hidden Valley are being dealt with by the writer in a separate paper in preparation.

MISSISSIPPIAN SYSTEM

The Mississippian system is represented in the southern Inyo Mountains by two principal formations: these are the Chainman shale above and the Tin Mountain limestone below. A minor unit, the Perdido formation, lies between the Tin Mountain and the Chainman; it has not been mapped separately.

The Perdido is much thicker to the east in the Ubehebe district where McAllister (1952, p. 22–25) has distinguished it as a separate formation. West to east thickening of the Perdido between Cerro Gordo and the Ubehebe area is believed by the writer to take place at the expense of the lower part of the Chainman shale, of which it is accordingly viewed as a facies.

TIN MOUNTAIN LIMESTONE

General features.—The dark-gray Tin Mountain limestone of Early Mississippian age rests upon the Lost Burro formation and is overlain by the Perdido formation, also of Mississippian age. Named for a peak in the northernmost Panamint Range (McAllister, 1952, p. 20) the Tin Mountain is a prominent cliff maker in its type area. In the Cerro Gordo area it forms a dark-gray band at the top of the cliff-making lighter colored Lost Burro (figs. 5 and 6).

Areal distribution.—Tin Mountain limestone forms the top of Cerro Gordo Peak and composes most of the higher west slope between that point and the Morning Star mine to the south. West of San Lucas Canyon the intermittent dark band of Tin Mountain may be fol-

FIGURE 6.—View looking north along higher east slope of the Inyo Mountains near Cerro Gordo. Left distance shows Chainman shale at crest. Middle distance, crags in sheared Tin Mountain limestone against light-gray Lost Burro formation on east. In foreground, fusulinid-bearing limestone of the Keeler Canyon formation downfaulted on Ignacio fault.

lowed northward for some 5 miles near the top of the east-facing cliff slopes. North of Daisy Canyon on the very rugged east slopes and on the lower west side of the range, it is differentiated with difficulty from the squeezed and altered Lost Burro. The Tin Mountain limestone crops out near the Lee mine on Lee Flat (fig. 3) and at several points between the Lee mine and the Belmont mine. At Mazourka Canyon it has not been recognized beneath the unconformity which separates Perdido from the Silurian limestone.

Thickness.—The Tin Mountain is about 350 feet thick at Cerro Gordo Peak, thinning appreciably as it is followed northward along the crest of the Inyo Mountains. At Mexican Spring it has thinned to about 75 feet. In the type area the Tin Mountain is 475 feet thick.

Lithology.—At Cerro Gordo the Tin Mountain limestone is a medium-bluish to dark-bluish-gray fine-grained limestone and its beds range in thickness from less than 1 inch to 2 or more feet. Weathering platy and flaggy to massive, this formation appears more massive from a distance than it actually is.

Dark-gray and black chert is common as blobs and irregular lenses which are either iron stained or bleached light gray and white. Much organic material is present in the form of crinoid and coralline debris. White calcite of the crinoid fragments contrasts with the dark limestone matrix. Content of argillaceous matter is low, chief impurities being silica of the chert and the dark finely divided carbonaceous material.

North of Cerro Gordo, where the Tin Mountain is affected by fracture cleavage, it forms jagged cliffy slopes. Distribution of chert nodules makes possible the distinction between bedding and cleavage, otherwise difficult.

Near Cerro Gordo the Tin Mountain limestone is thinner and more uniform lithologically than in the Ubehebe district, where, according to McAllister (1952, p. 20, 21), it shows pale-red partings, and in the lower part brownish-gray to pale-red calcareous shale.

Stratigraphy.—Although the Lost Burro-Tin Mountain contact is an important system boundary, no physical evidence of erosion or angular discordance was noted. It is nonetheless well defined lithologically and readily mappable, unlike most system boundaries in the Paleozoic of the Great Basin. The fairly uniform dark-gray band of the cherty Tin Mountain contrasts rather sharply with the streaked and patchy light-gray and darker gray Lost Burro beneath.

In the type Tin Mountain, McAllister (1952, p. 21) recognized two members: (a) a lower unit 275 feet thick with beds 2 to 6 inches thick and including brownish-gray to pale-red calcareous shale; (b) an upper cliff-forming member 200 feet thick with beds a few inches to 2 feet thick and with faint pale-red partings. Except for pale-red partings the Tin Mountain at Cerro Gordo more nearly resembles the upper member. Seeming absence of the lower beds, which include red and brown calcareous shale, suggests erosion or nondeposition, possibly at the Devonian-Mississippian boundary in the southern Inyo Mountains.

At Cerro Gordo the Tin Mountain limestone is overlain with sharp contact by the fine-grained quartzite member which constitutes a westerly tongue of the Perdido formation. Followed northward along the range crest the quartzite of the Perdido disappears at some points, as the Tin Mountain itself becomes thinner. Such behavior suggests disconformity at the Perdido-Tin Mountain contact. In keeping with this reasoning is the observation that in Mazourka Canyon the Perdido is spotty and rests with profound unconformity upon Silurian limestone. In the Cerro Gordo area the depositional change at the Tin Mountain-Perdido contact is one of the more significant of the Paleozoic column. The normal marine carbonate environment gives way here to one of land-derived siliceous clastic materials which include silt, sand, and highly argillaceous shale. Such facies characterize the Chainman-Diamond Peak interval over very large areas in the Great Basin. Land-plant remains in the Chainman bespeak emergence and presence of exposed land at no great distance. To the east, conditions were somewhat different, for in the type Tin Mountain area McAllister (1952, p. 21) finds the relation of this carbonate unit to the overlying Perdido to be one of gradation.

Age and correlation.—Although the Tin Mountain limestone is loaded with crinoid remains and other fossil debris in the Cerro Gordo area, it has yielded few determinable fossils. The crinoid columnals are often of large size reaching a diameter of more than half an inch. Near the Cerro Gordo springs, 3½ miles northwest of Cerro Gordo, the Tin Mountain contains *Syringopora* and horn corals of the *Caninia* type. Near the Lee mine (fig. 3) and in the long ridge to the north this limestone contains *Syringopora*, *Triplophyllites?*, and a distinctive brachiopod referred to here as *Brachythyris* sp. A. Spirifers found here resemble *Spirifer rowleyi* Weller or *S. grimesi* Hall. In the Ubehebe district the formation has yielded abundant fossil material (McAllister, 1952, p. 21). On the southeast side of Perdido Canyon (loc. 36), the writer made collections which include the following:

Aulopora sp.
Syringopora sp.
Ekvasophyllum n. sp. (*Ekvasophyllum* Parks, 1951)
Caninia sp.
Triplophyllites? sp.
Chonetes cf. *C. loganensis* Hall and Whitfield
Schuchertella cf. *S. chemungensis* (Conrad)
Orthothetes inflatus (White and Whitfield)
Productus sp. (small form)
Spirifer cf. *S. centronatus* Winchell
Spirifer cf. *S. missouriensis* Swallow
Brachythyris sp. A (finely ribbed)
Euomphalus cf. *E. utahensis* Hall and Whitfield
Straparollus? cf. *S. ophirensis* Hall and Whitfield
Platyceras sp. (possibly two small species)
Crinoidal material

The Tin Mountain faunas are of Early Mississippian age, indicating correlation with the Madison limestone. Likewise correlative is the Joana limestone of central Nevada (Nolan, Merriam and Williams, 1956, p. 54) which also contains faunas of Madison type.

In central Nevada the conodont-bearing Pilot shale underlies the Joana. Conodonts from the lower part of the Pilot indicate a Late Devonian age (Hass, in Nolan, Merriam, and Williams, 1956, p. 53); the upper part of the Pilot is believed to be Early Mississippian in age. Although the Pilot shale is not recognized in the region under discussion, it is not unlikely that the lower part of the Tin Mountain with red calcareous shale in the Ubehebe district (McAllister, 1952, p. 20) may be of the same age as the lower part of the Pilot shale. Near Eureka, Nev., the lower conodont-bearing shale of the Pilot is reddish at some points.

Further details relating to this part of the stratigraphic column are dealt with by the writer in a paper in preparation.

PERDIDO FORMATION

General features.—The name Perdido formation was given by McAllister (1952, p. 22–25) to a heterogenous and facies-variable sequence of strata with type section near Perdido Canyon, Ubehebe district. Including siltstone, sandstone, shale, conglomerate, chert, and limestone, the Perdido of the type area conformably overlies the Lower Mississippian Tin Mountain limestone and is overlain with conformity by the upper part of the Chainman shale to which McAllister has given the local name "Rest Spring shale."

Comparison of the Inyo Mountains and Ubehebe sections indicates that the Perdido is possibly an eastern complex of lithologic facies which takes the place of lower and middle parts of the Chainman shale. To the east the Perdido of McAllister reaches a thickness of at least 600 feet; to the west near Cerro Gordo it thins to 50 feet and locally pinches out completely at points where black shale of the Chainman rests directly upon Tin Mountain limestone. Because of its thinness in this area the Perdido is included with the Chainman for mapping purposes.

Thickness and areal distribution.—The Perdido formation of the Inyo Mountains ranges in thickness from about 200 feet at Mazourka Canyon (fig. 3) to less than 50 feet at Cerro Gordo. It is spotty and variable in thickness at Mazourka Canyon. North of Cerro Gordo it is unrecognized at some points but reappears as the contact is followed along the strike. Along the Inyo crest (pl. 1) this unit has been traced intermittently from a point south of Belshaw Spring to the area south of the Morning Star mine. Strata assigned to this unit also occur east of the Belmont mine.

Lithology.—With the type Perdido as noted by McAllister (1952, p. 22–23) "heterogeneity is an outstanding characteristic." Siliceous clastic rocks are the most distinctive, ranging "from shale through siltstone and sandstone to conglomerate. Of these, siltstone is the most abundant. It is commonly light gray or pale red and weathers from yellowish to reddish browns." The limestone varies from dark gray with interbedded black and gray chert to medium gray and platy. It is largely fine grained and in part silty to sandy. Certain of the limestones become coarsely clastic, made up of crinoidal debris, shell fragments, and pebbles. Conglomerate in the type area is lenticular and coarse, consisting of reworked rocks of the Perdido.

The greatly thinned Perdido of the Inyo Mountains shows its characteristic diversity in certain sections; in others it is represented solely by a fine sandstone or quartzite. At Cerro Gordo the unit is a fairly uniform dense or flinty fine-grained nearly white or cream-colored tan-weathering quartzite. Appearance of this rock is virtually that of a novaculite. The unit ranges in thickness here from less than 50 feet to about 75 feet. It is rather massive, north of Cerro Gordo forming a conspicuous white cliffy exposure, which contrasts strongly in color with the underlying Tin Mountain limestone. Southeast of Cerro Gordo the resistant quartzite of the Perdido forms a caprock at several points on the range crest. The light-colored rock may easily be confused with Pennsylvanian silicated limestone or "tactite" of the Cerro Gordo area.

Northeast of the Belmont mine fine-grained sands of the Perdido vary from light and medium gray to greenish or pale olive gray, are slightly calcareous at some points, and show faintly a rather fine banding. Pinkish crinoidal limestone with rounded black chert pebbles and brachiopod fragments rests locally upon the sandstone. Fucoidal markings characterize the calcareous sandstone or siltstone. East of Conglomerate Mesa (pl. 1) the Perdido at one exposure is partly calcareous and includes dark-olive-gray impure fine-grained calcareous sandstone of an aspect quite different from the light-colored novaculitelike sandstone of the Cerro Gordo vicinity.

An excellent exposure of highly diverse Perdido about 100 feet thick is found at the Crosscut tunnel, 1¾ miles northwest of Cerro Gordo on the Pipeline trail (pl. 1). Resting with apparent conformity on platy Tin Mountain limestone, the Perdido here comprises calcareous silty shale with lumpy bedding and fucoidal markings, dark-gray siltstone and fine quartzite, silty limestone, and a few dark-gray crinoidal limestone beds. These deposits are partly hornfelsic, showing in places almost a slaty cleavage. Dark-gray Chainman shale conformably overlies the Perdido.

At Mazourka Canyon (fig. 3) 25 miles northwest of Cerro Gordo, the Perdido is lenticular, ranging in thickness from 40 to 200 feet within a few hundred feet along strike. It rests unconformably on limestone of Silurian age. Where the formation is thickest at Mazourka Canyon, it includes three members. At the base a very local 90-foot member consists of medium- to rather coarse grained, heavy-bedded gray quartzite. The basal quartzite is overlain by 100 feet of extremely heterogeneous coarse limestone breccia-conglomerate within which are lenses of coarse gray quartzite; medium- to light-bluish gray medium-grained sandy limestone occurs in lenses or as a general matrix surrounding the conglomerate pieces. Conglomerate cobbles are coarse and angular to subrounded and reach a diameter of 3½ feet. Numerous limestone cobbles have the fine textured light-gray to white appearance of Devonian Lost Burro from which they are believed to have been derived. Other pieces are fine-grained cherty limestone and light-gray saccharoidal dolomite which may have been derived from rocks ranging in age from Ordovician to Devonian.

An upper 20-foot member of the Perdido formation consists of well-bedded silty medium- to dark-gray quartzite with interbeds of medium- to fine-grained light-gray limestone. This zone is seemingly transitional to the overlying plant-bearing black shale of the Chainman.

Stratigraphy.—McAllister's observations (1952, p. 22–23) in the type area of the Perdido with relation to the Perdido-Tin Mountain boundary indicate that "the two formations are somewhat gradational." Where exposures are poor, there is difficulty in mapping the contact. On the other hand the upper contact with "Rest Spring shale" in the type area is sharp. The "Rest Spring," unlike Perdido, normally has no reddish coloration and contains no limestone in that area.

In the Inyo Range, where strata assignable to the Perdido facies constitute much less of the Mississippian column, change from Tin Mountain limestone to sandstone or quartzite of the Perdido is very abrupt and probably a disconformity. These relations may be seen to advantage north of Cerro Gordo. Local absence of the Perdido in this vicinity may be related to an hiatus.

Evidence of pre-Perdido disconformity is convincing at Mazourka Canyon, where thickness differs greatly within short distances. This suggests that coarse Perdido clastic rocks overlie a surface of considerable relief cut in limestone of Silurian age. The Lost Burro formation of Middle and Late Devonian age and the Tin Mountain limestone of Mississippian age are absent, but large marble and limestone cobbles in basal Perdido breccia-conglomerate appear to have been derived from the otherwise missing Devonian unit. To judge from the Cerro Gordo section, as much as 2,000 feet of combined Tin Mountain and Lost Burro may have been stripped away by pre-Perdido erosion.

At Mazourka Canyon the relation of the uppermost 20-foot member of the Perdido to the black platy shale in the Chainman is apparently gradational. This change is in fact less abrupt than that which separates the 20-foot member from the underlying breccia-conglomerate of the Perdido formation.

Disconformity is suggested by westerly stratigraphic thinning and presence of conglomerate within the Perdido of the Ubehebe district (McAllister, 1952, p. 24). Very great overall westerly thinning in the 12 miles which separate Cerro Gordo from the Ubehebe area is best explained as a function of lateral replacing facies relation between the easterly Perdido and the westerly Chainman shale, as outlined more fully below.

Age and correlation.—Direct fossil evidence supporting Late Mississippian age of uppermost type Perdido is the *Cravenoceras* fauna discussed by McAllister (1952, p. 24). Below the *Cravenoceras* fauna in the type Perdido column (loc. 37), the writer has collected a *Triplophyllites* fauna in medium-gray sandy limestone; certain of the species are as follows:

Triplophyllites sp. a
Spirifer cf. *S. brazerianus* Girty, or cf. *S. grimesi* Hall
Spirifer cf. *S. pellaensis* Weller
Spirifer missouriensis Swallow
Echinochonchus sp.
Dictyoclostus sp.
Composita cf. *C. sulcata* Weller
Kaskia cf. *K. chesterensis* Weller and Weller
Deltopecten sp.

The horizon of this assemblage is possibly that of McAllister's (1952, p. 23) unit 10, which lies near the top of a 610-foot Perdido section. This limestone fossil assemblage is similar to that collected from the lower limestone of the Chainman on the Pipeline trail (loc. 22) and resembles the fauna collected at locality 8 south of the Morning Star mine. McAllister's unit 10 is believed to be higher stratigraphically than either locality 22 or locality 8 of the Inyo Mountains section.

CHAINMAN SHALE

General features.—The name Chainman shale is used for strata of Mississippian age which lie between the Perdido formation (or the Tin Mountain where Perdido is absent) and the Keeler Canyon formation (fig. 7).

Type section of the Chainman shale is at the Chainman mine near Lane, Ely mining district, Nevada, where black shale rests upon typical Joana limestone of Early Mississippian (Madison) age (Spencer, 1917, p. 24–26). Edwin Kirk (1918, p. 38), recognizing the similarity between Mississippian black shale of the Inyo Mountains and that of the Eureka district, Nevada, appropriately used the name "White Pine shale," at that time in general use. In recent years the name Chainman shale has been adopted at Eureka in preference to "White Pine Shale" and applies there (Nolan, Merriam, and Williams, 1956, p. 59) to a thick sequence of shale, siltstone, and conglomerate between the Joana

FIGURE 7.—View looking northwest along the west Inyo Mountains slopes. Shows wide belt of dark-gray Chainman shale on right, overlain on left by west-dipping Keeler Canyon formation. Triassic marine beds on extreme left rest unconformably on Keeler Canyon beds where the Owens Valley formation was unrecognized.

limestone below and the Diamond Peak formation above. Among the varied objections to use of "White Pine shale" as defined by Hague at Eureka (Hague, 1883, 1892) is lack of a clearly established type section and inclusion by Hague of Devonian as well as Mississippian strata. Joana limestone and Pilot shale (partly Devonian), now separate formations, were included in the "White Pine shale" of Hague. As discussed below under age and correlation, the upper part of the sequence called Chainman at Cerro Gordo may well include a time-stratigraphic equivalent of some parts of the Diamond Peak formation but does not include conglomerate and arenaceous limestone facies characteristic of the true Diamond Peak.[1]

Areal distribution.—Three principal belts of Chainman shale have been mapped in the southern Inyo Mountains. On the east side of the range this unit crops out at a few points between Black Basin and the first canyon north of Bonham Canyon (pl. 1), but in this belt it is largely beneath fanglomerate. Along the range crest from a point southeast of the Morning Star mine to Cerro Gordo and northward past the Cerro Gordo springs to Daisy Canyon, the Chainman is broadly exposed. A western belt extends along the east side of Salt Mill Hills to the Long John mine vicinity (pl. 2), and a fourth belt in the western foothills may be followed southward from Mazourka Canyon (fig. 3).

From Daisy Canyon northward on the precipitous east side of the range occur highly irregular infolded areas of variously altered Chainman, which have been penetrated by granitoid rocks. Together with the underlying Tin Mountain the altered shale appears as nearly black bands contrasting with the associated light-gray and white marbles of the Lost Burro.

Lithology.—The Chainman shale comprises dark-gray to black carbonaceous clay shale, silty shale, fine sandstone, and limestone. Black noncalcareous clay shale predominates; this ranges from smooth fissile very fine locally almost papery shale to nonfissile dense slaty and blockly-weathering argillite in which the shaly parting is less evident. Fine sands and silty sands occur as interbeds. Limestone is subordinate, occurring in lenses from a few inches to about 70 feet thick. The argillite in places has limonite coatings following the numerous joints and fracture surfaces. Sandstone beds commonly weather brown. The limestone ranges from dense fine-grained dark-gray carbonaceous types to those of a purer variety which are silty or sandy, medium grained, and of medium-gray to bluish-gray color. Coarse-grained crinoidal limestone is also present.

[1] The name "Diamond Peak" was erroneously employed by Kirk (1918, p. 40) for various altered siliceous rocks, some of which are Permian.

Plant remains are locally abundant in the black shale. These are of two principal types: (a) possible algae of vermiform fucoidal configuration and (b) branching stems and flat straight straplike or rushlike impressions of fragmentary nature, probably derived from a coal swamp environment. At some localities marine shells are associated with land-plant remains. The limestone lenses contain marine fossils. Certain of the cleaner, less carbonaceous limestone lenses carry faunas which differ in facies from those of the black shale.

In the southern Inyo Mountains the Chainman shale has been affected widely by low-grade metamorphism. On the whole a structurally weak rock, this formation was subjected to igneous intrusion at many points, and its alteration is related in the main to contact or hydrothermal action. The extent of alteration becomes greater toward the north, where granitoid rocks are much more in evidence at the surface than at Cerro Gordo. Degree of alteration varies from weak silicification or silication through mild slaty or phyllitic change to complete recrystallization, producing in the extreme a fine-grained black rock of almost lamprophyric appearance. Northwest of Cerro Gordo, bordering an intrusive stock, the dense argillite is mildly silicified, assuming a flaggy and blocky-weathering appearance. Pyrite oxidized to limonite is common. More strongly altered phases exhibit secondary silicate crystal growths ranging from minute specks to large knots. In the Mazourka Canyon area chiastolite is reported (Kirk, 1918, p. 38). On the east side of the range, where the road down San Lucas Canyon meets the Lee Flat-Bonham talc mine road (fig. 3), the silty Chainman is partly recrystallized with development of silicate needles. In spite of the metamorphism, impressions of productid brachiopods may still be seen. East of Brooklyn Canyon (pl. 2), strongly altered Chainman ranges from phyllitic to almost schistose locally, showing needlelike crystal growths in a silvery gray sericitic groundmass. Slaty development with cleavage angle differing perceptibly from that of bedding seems rather uncommon, and in general foliation is not well developed in these altered phases.

Thickness.—Exposures of Chainman shale cannot ordinarily be expected to give true stratigraphic thickness because of the faulting, shearing, and drag folding undergone by these incompetent beds. Along the Dolomite Canyon thrust-fault zone in the western belt, apparent thickness ranges from less than 50 feet in fault slivers to more than 1,000 feet in continuous near-vertical sections at Brooklyn Canyon (pl. 2). Near the Crosscut tunnel, 1¾ miles northwest of Cerro Gordo (pl. 1), about 1,100 feet of Chainman was measured, but allowance should be made for possible thickening by drag folds. East of the Inyo Mountains the "Rest Spring shale" of McAllister (1952, p. 26), believed to represent only the upper part of the Chainman, attains a thickness of about 400 feet. The Chainman shale in the Diamond Mountains near Eureka, Nev., reaches a thickness of 3,000 feet, which exceeds the maximum at Cerro Gordo by some 2,000 feet. That the black shale facies of the Chainman thicken appreciably from east to west is evident between the Ubehebe area and Cerro Gordo. Depositional changes involved are considered below under stratigraphy.

Stratigraphy.—The Chainman shale rests conformably on the Perdido formation and is overlain conformably by the Keeler Canyon formation. In discussing these units it is desirable to consider the Chainman and Perdido jointly, for together they occupy an interval whose top and bottom are fixed paleontologically, by the early Mississippian (Madison) Tin Mountain below and the basal Keeler Canyon of Atoka age above. At Cerro Gordo the Chainman and Perdido together are 1,000 feet thick, but only about 100 feet or less of this is Perdido. In the Ubehebe area on the other hand the Perdido thickens to some 600 feet, whereas the Chainman ("Rest Spring") above it thins correspondingly to 400 feet or less. Combined thickness of Perdido and Chainman is nonetheless roughly the same in the two areas, or about 1,000 feet. The lithologic differences involved are explainable by facies interpretation, black Chainman shale occupying the time-stratigraphic interval of most of the Perdido to the west. Subjectively, the Chainman-Perdido boundary may be said to rise in the column eastward with intertonguing pattern, thus transecting imaginary time-stratigraphic planes.

Facies change in a southeasterly direction from black shale of the Chainman at Cerro Gordo to a wholly carbonate Mississippian section is even more notable than northeasterly change into the complex facies of the Perdido. To the southeast in the direction of Darwin (fig. 3), the shale disappears (Hall and MacKevett, 1958, p. 7–9) and the thick Argus Range Carboniferous column (Hopper, 1947, p. 409) appears to be entirely limestone, lacking the Chainman shale facies.

Relation of Chainman to the overlying Keeler Canyon formation of Pennsylvanian and Permian age is seemingly gradational. An intermediate zone with alternating limestone and black shale beds is well shown in upper Soda Canyon and on the east side of the Inyo Range in the first canyon north of Bonham Canyon (pl. 1).

Detailed stratigraphic zonation of the Chainman shale at Cerro Gordo on the basis of lithology and fossils is not yet feasible. Limestone bodies are thickest and most numerous in the lower 250 feet of the formation, being well shown along the Pipeline trail north of Cerro Gordo, where the 3-foot fossiliferous limestone bed is roughly 200 feet above the base of the Chainman. In Mazourka Canyon (fig. 3) the lenticular 70-foot limestone lies with its base 150 feet above the bottom of the Chainman. South of the Morning Star mine, coarse-grained bluish-gray crinoidal limestone rests upon, and is partly lenticular within, sandstones representing a Perdido facies. In the same vicinity partly crinoidal limestone lenses occur in what appears to be lower black shale of the Chainman, but at a higher stratigraphic horizon than that discussed above.

Near the Crosscut tunnel on the Cerro Gordo Pipeline trail, smooth, fissile to almost papery black clay shale occurs in the lower 400 feet of the Chainman, which interval also includes coarser silty interbeds as well as limestone lenses. Marine fossils of the black shale facies are found in these lower beds, where they are associated with land-plant remains. West of the Pipeline trail the upper 600 feet of the formation is poorly exposed. It apparently consists in the main of monotonous dark-gray and black silty shales, passing at the top into a 20-foot zone of platy calcareous shale, overlying which is the basal limestone of the Keeler Canyon with fusulinids.

Age and correlation.—The Chainman shale is older than early Middle Pennsylvanian beds of the Keeler Canyon which overlie it and younger than the Tin Mountain limestone with Early Mississippian (Madison) faunas. Its age is considered to be Late Mississippian.

Several fossil collections have been made in the lower part of the Chainman near Cerro Gordo, but thus far none have come from the 600 feet of strata in the middle and upper parts of the formation in this area. Material collected by Kirk (1918, p. 38–39) on the Pipeline trail north of Cerro Gordo was identified by Girty as follows:

Solenomya? sp.
Caneyella wapanuckensis
Caneyella n. sp., aff. *C. richardsoni*
Orthoceras, several species
Gastrioceras aff. *G. richardsonianum*
Goniatites sp.
Eumorphoceras bisulcatum?
Plant remains
Fish remains

Of this association Girty makes the following comments: "The latter is an interesting and peculiar fauna of the Caney shale of Oklahoma and the related but less well-known fauna of the White Pine shale of Nevada. These faunas I refer to the Upper Mississippian."

During the present investigation, collections were made at locality 22 on the Pipeline trail 1½ miles northwest of Cerro Gordo, where black shale and intercalated limestone are abundantly fossiliferous. Black shale from 200 to 250 feet above the base of the Chainman yielded the following:

Orthoceras sp. (small form)
?*Nautilus* sp. (large form, possibly a *Liroceras*)
Cravenoceras cf. *C. nevadense* Miller and Furnish
Cravenoceras cf. *C. richardsonianum* (Girty)
Cancycla wapanuckensis Girty
Cancycla n. sp., cf. *C. richardsoni* Girty
Shark teeth (coarsely serrated edge, ¾ in. long)
Land plants (abundant, fragmentary; one well-preserved fernlike type)

This assemblage is virtually the same as that studied by Girty. Similar fossil associations were collected at locality 38 on the east side of the Inyo Range.

A 3-foot silty limestone bed only a few feet above the cephalopod-bearing black shale at locality 22 on the Pipeline trail contains the following fauna, and is provisionally correlated with the Brazer limestone:

Triplophyllites sp. A
Dictyoclostus cf. *D. burlingtonensis* (Hall) or cf. *D. fernglenensis* (Weller)
Productus n. sp., cf. *P. semistriatus* Meek
Unlike Meek's type, this form does not have obsolete ribbing on the anterior half of the pedicle valve.
Spirifer cf. *S. pellaensis* Weller
Spirifer cf. *S. brazerianus* Girty or cf. *S. grimesi* Hall
Small goniatite with narrow venter; genus indeterminate

A similar fauna of Brazer aspect is found at locality 8, which lies 3,000 feet southeast of the Morning Star mine, in coarse crinoidal limestone at the top of the sand in the Perdido and near the bottom of the black-shale facies of the Chainman. The following fossils have been determined:

Triplophyllites sp., cf. *T.* sp. A (abundant)
Chaetetes sp.
Spirifer cf. *S. brazerianus* Girty
Spirifer cf. *haydenianus* Girty
Composita cf. *C. lewisensis* Weller
Productus cf. *P. richardsi* Girty
Productus cf. *P. ovatus* Hall
Diaphragmus cf. *D. elegans* (Norwood and Pratten)
Dictyoclostus sp. (large, coarsely ribbed)
Rhynchopora? sp.
Proetus cf. *P. missouriensis* Shumard
fenestellid bryozoans (common)
Pleurotomaria cf. *P. brazeriana* Girty

In connection with lower limestone faunas in the Chainman of the Inyo Mountains, attention is called to a similar assemblage with *Triplophyllites* which occurs below the *Cravenoceras* beds in the Perdido formation of the Ubehebe area (loc. 37). This fauna, elsewhere discussed under the Perdido, probably occupies a higher stratigraphic position than the comparable limestone fauna with *Triplophyllites* sp. A on the Pipeline trail (loc. 22).

Correlation of the Chainman shale is based to a large extent on cephalopods, which occur abundantly in thin highly carbonaceous limestone interbeds of the black shale facies. These cephalopods are the subject of detailed paleontological studies, especially those from the White Pine or Hamilton region of central Nevada (Youngquist, 1949, p. 276, 283). Diverse generically, the cephalopod assemblages are characterized by *Cravenoceras* and *Eumorphoceras* and are regarded as of Late Mississippian age. Little is known of their vertical ranges or stratigraphic zonation. Some of the central Nevada occurrences are believed to be in the upper part of the Chainman. The thick Diamond Range stratigraphic column of Chainman shale near Eureka, Nev., otherwise well suited for stratigraphic zonation, has not as yet yielded these cephalopods.

In the Ubehebe district McAllister (1952, p. 23–24) finds a prolific cephalopod fauna with *Cravenoceras* about 600 feet above the base of the type Perdido, an horizon probably falling, in the time-stratigraphic sense well within the upper half of the Chainman column at Cerro Gordo. Thus while the only *Cravenoceras* occurrence recognized at Cerro Gordo is in the lower part of the Chainman, that at Ubehebe seems to be in a stratigraphic position equivalent to the upper part of the Chainman, conforming perhaps to an upper position in the Chainman of the central Nevada *Cravenoceras* faunas discussed above. Studies of zoned collections with good stratigraphic control are needed to determine evolutionary trends and zonation of these interesting forms within the Chainman sequence.

Failure to obtain fossils in the uppermost part of the Chainman at Cerro Gordo leaves open the possibility that it may conceivably be Early Pennsylvanian rather than Late Mississippian; for the relation of the Chainman to the overlying Keeler Canyon seems to be transitional and the lowermost Keeler Canyon fusulinids indicate an early Middle Pennsylvanian (Atoka) age.

The combined Chainman and Perdido of the Inyo Mountains region occupy a time-stratigraphic interval seemingly coinciding with that filled by combined Chainman shale and overlying Diamond Peak formation in the Diamond Mountains near Eureka, Nev. This conclusion follows if Tin Mountain and Joana Limestone are correlative and the *Fusulinella* zone (or Atoka zone) at the base of the Keeler Canyon correlates with the same zone at the base of the Ely limestone in the Diamond Mountains. The thickness discrepancy is great, however, for that of combined Chainman and Diamond Peak is of the order of 6,200 feet compared with 1,000 feet for the Chainman and Perdido stratigraphic column of the southern Inyo Mountains. Except for shale the upper part of the Chainman at Cerro Gordo lacks characteristic lithologic types of the true Diamond Peak.

PENNSYLVANIAN AND PERMIAN ROCKS

In combination the Pennsylvanian and Permian rocks are about 4,000 feet thick; this thickness constitutes about one-third of the exposed 12,000-foot Paleozoic section in the Cerro Gordo area. These strata are predominantly impure carbonates with subordinate shale, calcareous shale, siltstone, sandstone, conglomerate, and chert. Most of these are mildly metamorphosed; carbonate rocks are in part recrystallized to marble and argillaceous rocks to argillite and hornfels. Commonly within a mile of granitic intrusions the limestone is altered to calc-hornfels and tactite.

Pennsylvanian and Permian strata of the Cerro Gordo mine area are divided in two formations: the Keeler Canyon formation of Pennsylvanian to early Permian age inclusive, and the Owens Valley formation of Permian age (Merriam and Hall, 1957). Previously named units, the Reward conglomerate and Owenyo limestone (Kirk, 1918, p. 42–45), are localized lenticular members of the Owens Valley formation; being local facies these have not been utilized as map units.

In previous discussions of undifferentiated Pennsylvanian and Permian rocks of this region, they have been referred to provisionally as Bird Spring(?) formation (McAllister, 1956). In the Ubehebe Peak quadrangle these strata increase in thickness to more than 5,000 feet (McAllister, 1955). The Bird Spring formation, defined in southern Nevada (Hewett, 1931, p. 21; 1956, p. 42; Longwell and Dunbar, 1936, p. 1202), is believed to embrace strata ranging in age from Late Mississippian to Permian.

TABLE 3.—*Pennsylvanian and Permian sequence in the southern Inyo Mountains, Calif.*

Age	Formation		Thickness (feet)	Lithology	Characteristic fossils
Permian	Owens Valley	Upper part	180–500	Sandy and silty limestone with chert pebbles, calcareous sandstone, siliceous conglomerate, limestone-cobble conglomerate, and quartzite.	*Punctospirifer pulcher* (Meek), *Spirifer pseudocameratus* (Girty).
		Disconformity			
		Middle part	400–700	Calcareous shales with argillaceous silty and sandy limestone intercalations; silty clay shale; fine sandstone; hornfels. Fusulinids abundant in limestone.	*Parafusulina*, *Schwagerina*; *Heritschia*, *Parenteletes*.
		Lower part	1,000±	Silty fusulinid limestone, lenticular fairly pure organic limestone, limestone mud-breccias, platy argillaceous limestone, blocky limestone conglomerate, and hornfels.	*Pseudoschwagerina*, *Schwagerina*, subordinate *Parafusulina*, *Triticites*; *Heritschia*, *Omphalotrochus*.
	Angular unconformity on west side Inyo Mountains				
	Keeler Canyon	Upper part	2,200±	Arenaceous silty and pebbly limestone with shale intercalations; calcareous sandstone and siltstone. Shale intercalations commonly pink or maroon. Marble.	*Schwagerina*, *Triticites*.
Pennsylvanian		Middle part			*Triticites*.
		Lower part		Fine-textured fusulinid-crinoidal limestone with round black chert nodules in "golf-ball beds." Tactite and marble.	*Fusulinella*, *Millerella*.

KEELER CANYON FORMATION

Name and occurrence.—The Keeler Canyon formation of Pennsylvanian to Early Permian age was named by Merriam and Hall (1957, p. 4–7) for exposures in upper Keeler Canyon, where the type section lies east of the Estelle tunnel portal and 2 miles southwest of Cerro Gordo Peak (pl. 1). The formation is widely exposed in the southern Inyo Mountains and also underlies most of the Darwin Hills and the northern Argus Range. It generally forms smooth slopes, but resistant beds protrude locally to emphasize the incompetent folded nature of these strata.

In earlier stratigraphic studies by Kirk (1918, p. 40–41), these beds were in part referred to as "basal Pennsylvanian limestones" and in part as "later Pennsylvanian limestone and shale"; siliceous and silicated limestone southwest of the Cerro Gordo mine previously regarded by Kirk as "Diamond Peak quartzite" are mapped by the writer as Keeler Canyon formation.

Lithology.—The Keeler Canyon formation comprises thin-bedded medium- to dark-gray impure silty and arenaceous to pebbly limestone and limy siltstone, with intercalations of pink or maroon fissile shale. Pebbly limestone in a few places grades into chert pebble conglomerate with limestone matrix. Silicified fusulinids are in some places an important constituent of the pebbly limestone. Clastic texture is shown by all these rocks, including the limestone. Bedding is inclined to be platy or flaggy, with few layers as much as 3 feet thick.

The basal 150 to 200 feet differs lithologically from the overlying beds. These lower strata are purer thin-bedded dark-gray crinoidal and fusulinid-bearing limestone containing near-spheroidal black chert nodules ½ to 2 inches in diameter. The cherty *Fusulinella*-bearing strata which were given the field designation "golf ball beds" constitute a reliable stratigraphic marker, having been recognized widely in the southern Inyo Mountains, the Darwin Hills, and the Argus Range.

Thickness and stratigraphic relations.—Because of its incompetent, highly folded nature the Keeler Canyon formation does not lend itself to accurate thickness appraisal. In the Cerro Gordo area the formation is on the average about 2,200 feet thick, as judged by measurement of several sections which ranged from 1,300 to 2,500 feet. Where the section is greatly thinned, there is evidence that segments have been cut out by faulting.

The Keeler Canyon formation rests conformably upon the Upper Mississippian Chainman shale. In the Darwin quadrangle (Hall and MacKevett, 1958, p. 9) it is underlain by thinly bedded Lee Flat limestone which apparently occupies at least part of the normal interval of the Chainman shale. In the southern Inyo Mountains the Owens Valley formation of Permian age rests unconformably on the Keeler Canyon, at some places with angular discordance of about 15°.

Age and correlation.—Abundant fusulinids make possible a threefold paleontologic zonation of the Keeler Canyon formation as follows:

Upper part	*Schwagerina-Triticites* zone
Middle part	*Triticites* zone
Lower part	*Fusulinella* zone

Fusulinella occurs only in the basal beds, the occurrence being principally that of the "golf ball beds" in the lower 200 feet. Locally *Fusulinella* is associated with *Millerella* in the Santa Rosa Hills, Darwin quadrangle, and at upper Soda Canyon, New York Butte quadrangle (pl. 1). The *Fusulinella*-bearing beds are Middle Pennsylvanian of about Atoka age.

Triticites characterizes the middle and upper parts of the Keeler Canyon formation. Diversity of type from horizon to horizon suggests that these forms might be employed for detailed zonation in a more refined stratigraphic study. Most of the recognized forms of *Triticites* in the Keeler Canyon indicate Pennsylvanian age, but in the uppermost beds on the Cerro Gordo road, early Permian species of *Schwagerina* or *Pseudofusulina* become abundant in association with *Triticites*. That the formation includes strata of both Pennsylvanian and Permian (Wolfcamp) age is therefore indicated.

The Keeler Canyon formation correlates with the Pennsylvanian and the early part of the Permian division of the Bird Spring formation in southern Nevada (Longwell and Dunbar, 1936), which also includes the zones of *Fusulinella* and *Triticites*. The Bird Spring formation restudied northwest of Las Vegas by Longwell and Dunbar is not wholly Pennsylvanian but embraces Late Mississippian and Permian stata.

The Ely limestone of central Nevada (Nolan, Merriam, and Williams, 1956, p. 61–63) is in part at least correlative with the Keeler Canyon, as demonstrated by occurrence of the *Fusulinella*-zone near its base.

PERMIAN SYSTEM

OWENS VALLEY FORMATION

Name and occurrence.—The name Owens Valley formation was proposed (Merriam and Hall, 1957, p. 7) for highly variable marine strata of Permian age which occupy large areas on the western slope of the Inyo Mountains near the Owens Valley border. Type locality of the new formation is in the foothills between Union Wash and the Reward mine (fig. 3), about 9 miles southeast of Independence between Owenyo and Kearsarge.

The Owens Valley formation has been mapped in the Inyo Mountains from the Reward mine, 5 miles north of Owenyo, to Conglomerate Mesa in the Darwin quadrangle. It also underlies the east slope of the Darwin Hills and most of the northern part of the Argus Range in the Darwin and Panamint Butte quadrangles. Northwest of the Reward mine the formation strikes into the alluvium of Owens Valley and is not known to reappear again to the north.

Lithology and stratigraphy.—The Owens Valley formation comprises interbedded silty and sandy limestone, fairly pure biogenic limestone, argillaceous shale siltstone, sandstone, and conglomerate. The formation is highly variable lithologically, both laterally and across the section. Because of the heterogeneity and absence of marker beds, mappable subunits are, for the most part, quite local and of the nature of large tongues or lenses. Nonetheless, on the basis of lithology and fossil faunas certain broad stratigraphic zones are recognized; these are referred to informally as lower, middle, and upper parts of the Owens Valley formation.

In areas of plutonic intrusion, especially on the west side of the Inyo Mountains, the Owens Valley formation has been altered to argillite, a minor amount of quartzite, and hornfels. Rocks previously classified as "Diamond Peak quartzite" (Kirk, 1918, p. 40) are mainly altered middle and lower parts of the Owens Valley. The Diamond Peak formation, a central Nevada Upper Mississippian unit (Nolan, Merriam, and Williams, 1956, p. 60–61), has not been traced to the Inyo Mountains.

The Owens Valley formation rests with local angular discordance upon the Keeler Canyon formation (fig. 8), as shown 2 miles south of Cerro Gordo road on the west side of the Inyo Mountains.

Early Triassic marine beds are unconformable upon the Owens Valley formation, although angular discordance is not generally observable at individual contact exposures. Mapping of the Permian-Triassic boundary southward from the Burgess mine (pl. 2) reveals an hiatus of varying magnitude. To the north Early Triassic *Ussuria* beds rest on the upper or Phosphoria Late Permian part of the Owens Valley formation, while to the south near Cerro Gordo road these Early Triassic strata came to lie on the upper part of the Keeler Canyon formation of Wolfcamp Early Permian age.

Discordant relations mapped in upper Soda Canyon (pl. 1, pl. 2) around the pronounced westerly salient of Paleozoic rocks are explainable as manifestations of the pre-Triassic unconformity.

The lower part of the Owens Valley formation comprises tan-weathering shale, sandy limestone, lenticular bodies of massive fairly pure limestone, conglomerate, and coarse limestone sedimentary breccia. Basal to the formation are the limestone breccias, containing angular blocks as much as 3 feet in diameter. Lenticular masses of this breccia rest with angular discordance upon truncated beds of the Keeler Canyon. Also quite lenticular and localized is the gray clean limestone partly composed of crinoidal fusulinid, molluscan, coral, and other shell materials. Such carbonate lenses form prominent ridges or bold craggy outcrops of medium dark gray color.

The lower part of the Owens Valley formation differs from the underlying Keeler Canyon in lacking pink or maroon punky shale, by presence of the large fairly pure biogenic limestone bodies, and by occur-

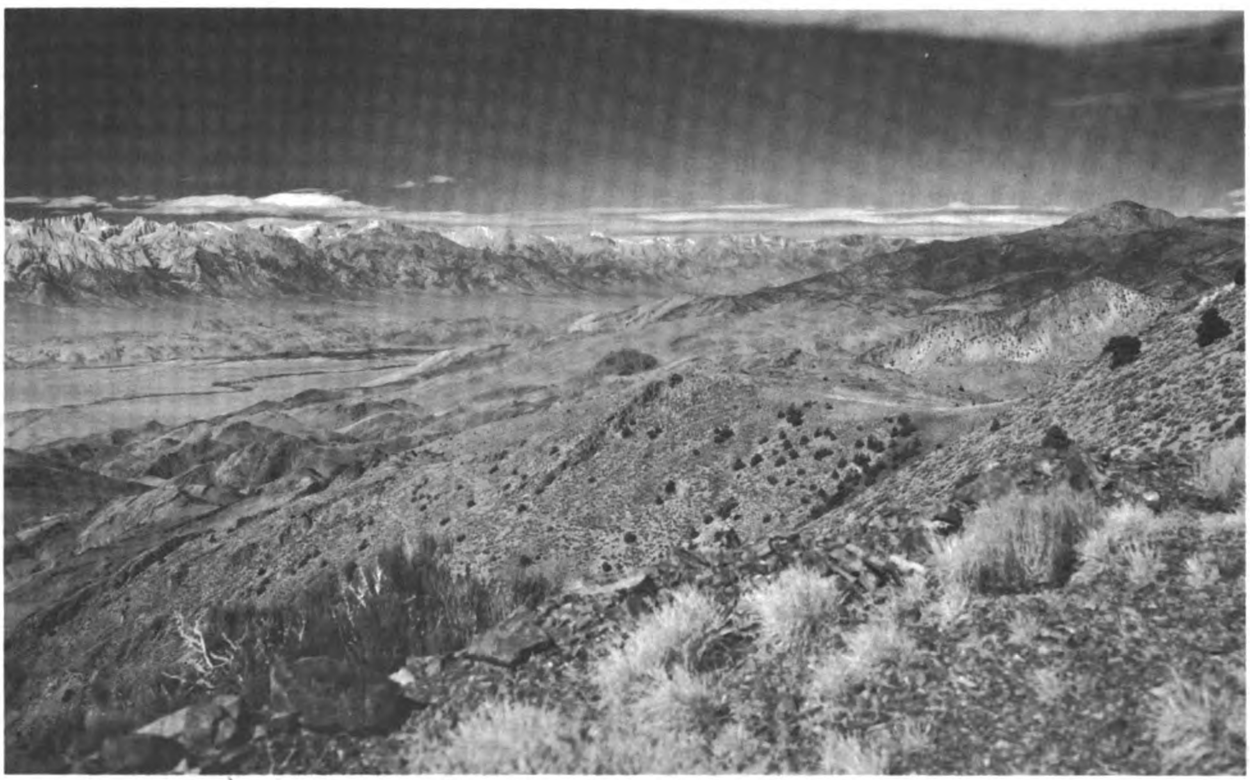

FIGURE 8.—View from a point near Hart Camp looking northwest along the west Inyo Mountains slopes; Owens Valley and Sierra Nevada in distance. Saddle in middleground shows Chainman shale overlain to west by limestone of the Keeler Canyon formation. Light band on right middleground is Permian Owens Valley formation at Permian Bluff. New York Butte in right distance. Photograph by L. G. Henbest.

nce of abundant crossbedding in the sandy limestone, feature rare in the Keeler Canyon beds.

Generally speaking, the middle part of the Owens alley includes much more shale than the lower part, nd fossiliferous limestone is less abundant. At Conomerate Mesa (pl. 1), a 200-foot unit in the upper art of the middle part of the Owens Valley shows nspicuous brick-red, greenish-gray, and yellowishrown fissile shale with thin siltstone beds. The thology changes gradationally, passing from the lower the middle part of the Owens Valley formation.

Conglomerate and sandstone are abundant in the pper beds of the Owens Valley. At the type locality nd southeastward through the western Inyo Mounains this part of the formation comprises calcareous andstone, arenaceous limestone, and conglomerate. ome of the conglomerates are highly siliceous; others re limestone conglomerates.

Upper beds of the Owens Valley in the type area are est shown at Fossil Hill south of the Reward mine. Here the beds are 300 feet thick and consist of fossilearing calcareous sandstone, sandy and silty limetone, and siliceous conglomerate. Beneath the upper eds at this exposure is phyllitic shale and hornfels. A omewhat undulant contact is shown with overlying hale of Early Triassic age.

Southeast through the western Inyo Mountains, the ipper part of the Owens Valley shows more limestone conglomerate and less of the siliceous or chert conglomerate. The intensely siliceous lenticular Reward conglomerate member described by Kirk (1918, p. 42–43) is a facies of the upper Owens Valley, which probably for the most part underlies the 300-foot upper interval at Fossil Hill. However, these Fossil Hill exposures are separated from the typical Reward by a broad wash which obscures the stratigraphic relations. Half a mile south of the Reward mine (fig. 3) the typical Reward is actually in large part quartzite and has a maximum thickness of 500 feet.

The Owenyo limestone member, 125 feet thick, crops out between Union Wash and the Reward mine threefourths of a mile southeast of Fossil Hill (Merriam and Hall, 1957, p. 3). A local member of the Owens Valley formation, it may be correlative with part of the 300-foot upper unit of the Owens Valley at Fossil Hill. The Owenyo includes a lower partly silicated white limestone with chert pebbles overlain by dense hornfels.

At Conglomerate Mesa (pl. 1) the contact between the middle and the upper parts of the Owens Valley is sharp and possibly disconformable. The upper division consists of limestone-cobble conglomerate, calcareous sandstone, and siltstone about 180 feet thick, forming the resistant cap rock of the mesa. The conglomerate contains fragments of gray limestone and silty limestone 1 to 4 inches in diameter embedded in a calcareous sandy matrix. Local patches of secondary silicification in the conglomerate at Conglomerate Mesa resemble the siliceous facies of the Reward conglomerate member.

Thickness.—No precise thickness can be given for the Owens Valley formation because of its folded and faulted nature. At Fossil Hill in the type area the formation is about 1,800 feet thick, whereas from Conglomerate Mesa eastward in the Darwin quadrangle it may increase to about 3,000 feet. This figure is, however, an estimate made in highly folded and faulted terrane where allowance must be made for duplication of beds. North of Cerro Gordo road in the New York Butte quadrangle, the formation locally pinches out between the Keeler Canyon formation and the Lower Triassic rocks (pl. 2).

Age and correlation.—The Owens Valley formation ranges from late Wolfcamp or early Leonard to Word and possibly Capitan age. Correlation and paleontologic divisions are based largely on fusulinids, by far the most numerous fossils. In the upper part of the formation, however, all known faunas represent brachiopod and molluscan facies, the fusulinids being conspicuously lacking except in reworked cobbles. Molluscan, coral, and brachiopod facies are very much localized in the middle and lower parts of the formation.

Three paleontologic zones have been recognized as follows:

3. *Spirifer pseudocameratus* zone
2. *Parafusulina* zone
1. *Pseudoschwagerina* zone

The upper or *Spirifer pseudocameratus* zone corresponds quite well to the upper part of the Owens Valley in the lithologic sense. *Parafusulina* of the middle part of the Owens Valley is not restricted to this part of the column, for it occurs below with *Pseudoschwagerina* in beds regarded as of late Wolfcamp or early Leonard age.

Spirifer pseudocameratus Girty is abundant in calcareous sandstone-limestone conglomerate facies of the upper Owens Valley formation. Beds loaded with this gregarious form have been recognized here and there from the type area at Fossil Hill southward to a point near the formational pinchout north of Cerro Gordo road (pl. 2). *Punctospirifer pulcher* (Meek) is less common and in places occurs in beds with a cephalopod fauna. This form occurs at Fossil Hill from which locality the Phosphoria fauna identified by Girty (*in* Kirk, 1918, p. 44–45) may have come. *Punctospirifer pulcher* is less abundant than *Spirifer pseudocameratus*.

These two forms were not found in association at most of the fossil localities, the beds with abundant *S. pseudocameratus* apparently being for the most part lower in the section than those with the *Punctospirifer* fauna.

Parafusulina occurs abundantly with *Schwagerina* in the middle part of the Owens Valley, locally with coral, molluscan, and brachiopod assemblages. *Heritschia* and *Parenteletes* are among the common forms. The *Pseudoschwagerina* zone is characterized not only by this genus but also by *Pseudofusulina* and a form classed with question as *Triticites*, which genera carry over from the Keeler Canyon. Fusulinids are often abundant in the purer limestone bodies of the lower part of the Owens Valley, occurring here with corals of the genus *Heritschia* and a large *Omphalotrochus* resembling *O. whitneyi* Meek.

The lower part of the Owens Valley formation correlates with the Permian part of the Bird Spring formation in southern Nevada containing the *Pseudoschwagerina* zone. In central Nevada the lower part of the Garden Valley formation with *Parafusulina* and associated *Pseudoschwagerina* is likewise correlative with the lower part of the Owens Valley, whereas the Carbon Ridge formation at Eureka, Nev. (Nolan, Merriam, and Williams, 1956, p. 64–67) with *Omphalotrochus* cf. *O. whitneyi* Meek and *Parafusulina* may be alined with the lower and middle parts of the Owens Valley. *Pseudoschwagerina* and *Omphalotrochus* cf. *O. whitneyi* point to a correlation with the lower part of the McCloud limestone of northern California; the higher part of the McCloud and Nosoni formations with *Parafusulina* may be alined with the middle and possibly the upper part of the Owens Valley. Upper beds of the Owens Valley with *Punctospirifer pulcher* (Meek) are correlative with the Phosphoria formation.

TRIASSIC SYSTEM

Rocks of Triassic age occupy a belt 26 miles long on the west side of the Inyo Mountains, extending northwest from the Ubehebe trail east of Keeler through Union Wash (fig. 3). Two principal divisions are represented, the lower being marine, the upper comprising volcanic rocks and land-laid deposits. The thickest sections and those best suited to stratigraphic study lie south of Union Wash where the marine fossil zones are well shown. So far as known, Triassic rocks occur only on the west side of the Inyo Range.

Along the Cerro Gordo road, some 4,000 feet of Triassic rocks are exposed, of which the lower 1,800 feet is marine and the remainder constitutes the land-laid volcanic rock sequence. In the area under consideration (pl. 2) the Triassic section is incomplete, for on the west side the Triassic belt is in fault contact with the Permian Owens Valley formation.

TRIASSIC MARINE ROCKS

Triassic marine rocks, thus far unnamed, are strongly folded and rest unconformably either upon the Permian Owens Valley formation or upon the Keeler Canyon formation of Pennsylvanian and early Permian age. The marine beds are overlain with apparent conformity by conglomerates classified as the basal unit of the unnamed Triassic volcanic sequence. At Union Wash (fig. 3) these marine beds are of particular interest as the source of significant Lower and Middle Triassic ammonoid faunas (Smith, 1901; 1904; 1914, p. 5–6; 1932, p. 9. Hyatt and Smith, 1905, p. 15, 20). Only a minor part of the Inyo belt of Triassic rocks is covered by the detailed geologic map (pl. 1).

LITHOLOGY AND STRATIGRAPHY

The marine rocks of Triassic age comprise mottled silty and argillaceous limestone, dark-gray to black carbonaceous limestone, lenticular bluish-gray limestone, and a large amount of argillaceous shale. Sandstone is uncommon. On the Cerro Gordo road the marine Triassic may be divided as follows into three stratigraphic units:

TABLE 4.—*Triassic rocks of the Cerro Gordo mining district*

Age	Formation (feet)	Characteristic fossils	Lithology (feet)	
Middle Triassic(?) (Upper part possibly as young as the Dunlap formation of Jurassic age.)	Volcanic sequence 2,200	No fossil evidence	5. Upper volcanic zone, 1,400	Dense andesitic flows, breccia, and tuffs of gray, red, and purple; intercalated red and green shale, slate, and sand, stone.
			4. Lower volcanic zone with tuffaceous sandstone and conglomerate, 800	Dense tuffaceous rocks, shale sandstone, and conglomerate of gray, red, purple, and green.
?				
Early Middle and Early Triassic	Triassic marine rocks 1,800		3. Upper reefy limestone zone, 750	Platy limestone and shale with thick reeflike lenses of massive limestone which weather out prominently.
		?		
		Parapopanoceras zone ? *Meekoceras* zone	2. Middle shale-limestone zone, 1,000	Gray fissile and platy shale with dark-gray limestone interbeds; poorly preserved ammonoids in limestone beds and concretions.
		Ussuria subzone	1. Lower brown-mottled limestone, 50±	Brownish-gray mottled silty nodular poorly bedded limestone; contains *Ussuria* and abundant minute gastropods.

Lower, brown-mottled limestone.—The lowest Triassic beds of this region consist of a variable patchy gray and brown-mottled silty-sandy and phosphatic limestone ranging in thickness from 20 feet to more than 75 feet. Locally it has a nodular appearance with irregularly rounded or ovoidal bodies of medium-gray limestone as much as 3 inches long surrounded by dark-yellowish-brown material. The gray nodules differ in texture from the brown matrix, giving the impression of reworking rather than concretionary origin. As a rule poorly bedded, the gray and brown-mottled limestone has a crinkly appearance in places and often breaks up on weathering to flat jagged lumpy pieces. This zone lacks uniformity of color and lithology. It crops out with a conspicuous ragged aspect contrasting strongly with beds above and below.

Characteristic of the mottled limestone are minute dark-brown bodies averaging about 1 mm in diameter. These prove to be phosphatic fillings of very small gastropod shells. The spiral brown bodies weather in relief and are so numerous in places as to form an important constituent of the rock mass.

The gray and brown-mottled zone is exposed 350 feet northwest of the Estelle tunnel portal (pl. 1) where it rests upon the Keeler Canyon formation. Because of unconformity the Owens Valley Permian is unrepresented at this point. No angular discordance was recognized. At this locality the zone is about 50 feet thick and exhibits a 4-foot basal limestone bed with abundant brown gastropod fillings and much fragmentary molluscan shell material. The overlying 45 feet of the zone is largely nodular brownish silty limestone which toward the top becomes more argillaceous and platy in a transition interval with phyllitic shales above.

Middle shale-limestone zone.—More than half of the marine Triassic in this area is argillaceous shale with limestone interbeds in more or less rhythmic alternation. On fresh break the shale is medium- to dark-gray, and pale greenish-gray, but weathers to light gray, yellowish orange, and slightly greenish yellow. The shale ranges in structure from fissile or papery to platy. A velvety phyllitic sheen revealed by parting surfaces gives evidence of mild alteration of these strata. Locally the platy shales are slightly calcareous. Limestone interbeds of dark gray color contrast sharply with light-colored shale layers and vary in thickness from less than 1 inch to more than a foot. The limestones range from medium bluish gray to black. They are of medium to very fine grain and except for carbonaceous matter are fairly pure. Platy weathering and fine lamination are characteristic. Small limestone lenses and concretions in light-gray papery shale near the base of this sequence contain an occasional ammonoid shell.

Upper reefy limestone zone.—The proportion of limestone relative to shale and the average thickness of individual limestone beds increase toward the top of the middle shale-limestone zone. About 1,100 feet above the base of the Triassic, dark-gray limestone comes to predominate over shale. Within the upper 800 feet of the marine Triassic where limestone predominates are several craggy hogback-forming limestone bodies ranging in thickness from 50 to about 200

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feet. To these upper strata the name "Blue Gate lime" has been applied locally by geologists. In the vicinity of Blue Gate on Cerro Gordo road, some five such steep-dipping and geomorphically prominent limestone bodies are distributed through about 900 feet of section. Reefy limestones of this kind vary in structure from fairly well bedded to very thick bedded and massive. Texture is medium to fine, and the color ranges from medium to dark gray or bluish gray. Some of these lenses pinch out completely in a few hundred feet, as shaly limestone seems to replace them laterally. Pink and orange shale interbeds are present in the lower part of the reefy limestone zone. Followed northward from Cerro Gordo road the thick upstanding reefy limestone disappears almost entirely from the section before Union Wash is reached.

Origin of the relatively uniform reefy limestone bodies is obscure, there being few clear indications of organic activity. Traces of possible calcareous algae suggest reef origin on a sea bottom where argillaceous limestone and shale accumulated between sites of vigorous organic growth.

Boundary of the upper reefy limestone zone with overlying Triassic volcanic rocks is poorly exposed, generally occupying an alluviated drainage depression which follows bedding strike. While there is no physical evidence of unconformity where Triassic conglomerate succeeds marine beds at this horizon, dips in the volcanic sequence are generally less steep, which suggests possibility of angular discordance.

AGE OF TRIASSIC MARINE ROCKS

The marine beds range in age from Early to Middle Triassic. Lowest Triassic faunas of the region are characterized by the ammonoid *Ussuria*, which is found in the lower gray and brown-mottled zone associated with the minute gastropod fillings, phosphatic brachiopods, and *Monotis*-like pelecypods. This lowest zone is accordingly referred to as the *Ussuria* zone. Northwest of the Estelle tunnel portal (loc. 15) the zone contains only small gastropod fillings and other shell fragments. Followed northward *Ussuria* and the other characteristic forms appear 1½ miles southeast of the Burgess mine. The minute gastropods are nearly everywhere abundant in this zone, but have not been found in the Union Wash section (fig. 3), where the lowest fossil zone recognized is that of *Meekoceras*.

Poorly preserved ammonoids, which probably represent the *Meekoceras* zone, occur in the lower part of the middle shale-limestone zone between the Estelle tunnel and the Cerro Gordo road. Beds containing these ammonoids are there underlain by the *Ussuria* zone with abundant minute gastropod fillings. In the same section, ammonoids may also be collected from higher beds which occur with pink and orange shale in the lower part of the upper reefy limestone interval. Undetermined pelecypod remains have also been found in the massive limestone of this zone northwest of Blue Gate (pl. 2).

Two distinctive ammonite zones are represented in the marine Triassic rocks of Union Wash. These are the *Meekoceras* zone of Early Triassic age and the *Parapopanoceras* zone of the early Middle Triassic. As demonstrated by J. P. Smith (1932, p. 8; 1914, p. 5, 6) fossils of the *Meekoceras* zone occur in a 12-foot limestone bed near the bottom of the Triassic section; those of the *Parapopanoceras* zone occupy a black limestone 6 feet thick and roughly 800 feet stratigraphically above the *Meekoceras* occurrence. Fossils are otherwise rare in the Union Wash section.

The earliest Triassic fossils of the region were not found at Union Wash (fig. 3) but at a locality 1½ miles southeast of the Burgess mine (pl. 2) where the ammonoid *Ussuria* occurs in abundance at the bottom of the section in gray and brown-mottled limestone beds, just above the Permian-Triassic boundary. *Ussuria* is associated here with the previously discussed minute phosphatic gastropod fillings. *Meekoceras* was not recognized with *Ussuria*, and it seems likely the *Ussuria* occurrence is stratigraphically below the 12-foot *Meekoceras* bed at Union Wash. Smith's faunal lists (1932, p. 9) for the Union Wash *Meekoceras* zone do not include *Ussuria*. Probably the *Ussuria* bed in question represents the *Pseudosageceras multilobatum* subzone, lowest subzone of the *Meekoceras* zone as recognized by Smith in southeastern Idaho, whereas the *Meekoceras* bed at Union Wash belongs in the overlying *Owenites* subzone as tabulated by Smith (1932, correlation table facing p. 14).

Poorly preserved fossils occur in shale above the *Ussuria* bed southeast of the Burgess mine and in about the same stratigraphic position between the Estelle tunnel and Cerro Gordo road. It is not unlikely these represent Smith's middle and upper subzones of the *Meekoceras* zone.

Pinkish colored punky shale above the early Middle Triassic *Parapopanoceras* bed at Union Wash carries poorly preserved ammonoids, as does similar pink shale near the bottom of the upper reefy limestone zone on the Cerro Gordo road. The upper reefy limestone was not found at Union Wash. It is probable these occupy a stratigraphic interval above the *Parapopanoceras* zone and are younger than early Middle Triassic.

TRIASSIC VOLCANIC ROCKS

The Triassic volcanic sequence comprises andesitic flows, tuffs, and breccias, with intercalated sandstone, shale, and conglomerate resting conformably upon marine Triassic beds. Although these volcanic rocks seem to be mainly andesitic, they include quartz-bearing flows. Along the Cerro Gordo road, where andesitic rocks appear to predominate, total thickness of the volcanic sequence is about 2,200 feet. However, the lower 800 feet is largely conglomerate, shale, and sandstone, part of which is tuffaceous. North of the Cerro Gordo road (pl. 2) the volcanic belt widens appreciably toward Union Wash (fig. 3), and although no accurate thickness measurement of this heterogeneous and deformed pile of igneous rock and land-laid sediments has been made, it probably exceeds 6,000 feet near Union Wash. The volcanic rocks are altered in varying degree, are often densely silicified, and the pyroclastics in places possess a sheared or almost foliated structure. Locally the volcanics are penetrated by porphyritic intrusive bodies. Whether these intrusives are of shallow origin, and nearly contemporaneous with volcanism, or much younger and possibly related to the Sierran Cretaceous plutonic phase has not been determined.

Along the Cerro Gordo road two thick sheets of andesitic porphyry are recognizable; these have a combined thickness of about 1,400 feet. It was not possible to determine how much is flow rock and how much pyroclastic. These rocks are conspicuously porphyritic, containing large feldspar phenocrysts. Color is reddish to purplish red on the weathered surface.

The more acid quartz-bearing volcanic rocks seem to be concentrated on the west side of the Triassic volcanic belt northwest of the Cerro Gordo road, where quite probably they lie stratigraphically above the andesitic rocks. Light-gray quartz-bearing porphyritic rocks of this supposedly upper sequence are well shown in Slate Canyon (pl. 2).

Tuffaceous deposits of the Triassic sequence are often banded and laminated, exhibiting fine laminated gray to dusky-red layers in alteration with layers of pale-pink to light-gray volcanic debris having a much coarser texture. Fragments are angular and poorly sorted. Pink and red colors change here and there to purple and green. In the red and greenish-gray volcanic breccia the angular fragments, embedded in a fine-grained matrix, range from sand size to several inches. Both quartz-free and quartz-bearing volcanic breccias are recognized, but those without quartz seem to predominate. Quartz-bearing volcanic breccia in Slate Canyon is associated with red and green slate in close proximity to the light-gray quartz-bearing porphyritic rocks mentioned above.

Red and green shale, slate, sandstone, and conglomerate are associated with the Triassic volcanic rocks. Although these lenticular deposits have yielded no fossils, they pass laterally into water-laid pyroclastic rocks and together with these are believed to be of continental origin. Some of the red sandstone exhibits sweeping crossbedding suggestive of wind action. On the Cerro Gordo road the lowest red beds believed to be land-laid occur with basal conglomerates which overlie the marine Triassic rocks. Similar red deposits recur at intervals through the lower 1,000 feet of the volcanic sequence. Pink colored punky shale occurs several hundreds of feet lower in the lower part of the upper marine reefy limestone. This shale contains marine fossils but suggests a possibility of brief emergence and weathering. Similar red shale intercalations containing ammonoids occur at Union Wash near the top of the Triassic marine sequence

Red and green slate is well shown in Slate Canyon where it has been quarried. Near the west edge of the volcanic belt north of the Cerro Gordo road are coarse red sandstone and conglomerate that contain well-rounded pebbles and cobbles of volcanic rocks as much as 10 inches in diameter.

Age of the volcanic sequence remains uncertain. Ammonoids occur in the upper punky-appearing pink shale between the *Parapopanoceras* zone and the volcanic rocks at Union Wash; specimens collected from these shales are, however, indeterminate. The upper marine pink shale as well as land-laid beds within the volcanic sequence itself may be expected to yield fossil evidence eventually. The volcanic rocks can with assurance be dated no more precisely than post-*Parapopanoceras* Middle Triassic. It is not inconceivable that these strata may be as young as the Dunlap formation of Nevada, which is assigned to the Jurassic (N. J. Silberling, oral communication, 1958).

IGNEOUS ROCKS OF THE CERRO GORDO AREA

Southeast of New York Butte in the Cerro Gordo map area (pl. 1), igneous rocks form about 12 percent of the surface. Of these about half are intrusive rocks and half andesitic volcanic rocks of Triassic age. Northward from New York Butte (pl. 2) the plutonic intrusives are much more in evidence, for erosion has stripped away a large part of the sedimentary roof. Late Tertiary basalts are not present in the area under consideration, but cover a large area of the southernmost Inyo Range, 6 miles south of Cerro Gordo.

Rock alteration and the surface pattern of granitoid rock distribution leave little doubt that the entire southern Inyo Range is underlain by plutonic rocks, at depths ranging from a few hundred to a few thousand feet.

Petrologic studies of Inyo Mountains and Sierran plutonic rocks were made by Knopf (1918, p. 59–60), who demonstrated that quartz monzonite and related granitic types are typical of the region under consideration. Special petrographic investigations which bear in some measure upon igneous rock problems at Cerro Gordo have recently been carried out in connection with U.S. Geological Survey mapping of the eastern Sierra Nevada and the Panamint region (written communications P. C. Bateman, J. F. McAllister, W. E. Hall and E. M. MacKevett, and J. G. Moore). In view of these special studies it is not the intent of this discussion to enter into technical description of the many and variable igneous facies but merely to point out gross relations of these rocks to problems of stratigraphy, structure, and ore occurrence.

Of interest are ages and relative ages of the various stocks and dikes in the Cerro Gordo vicinity, especially with relation to periods of ore mineralization. Fractured dikes in the Cerro Gordo mine clearly served as avenues of ascent for mineralizing solutions. Other pertinent aspects of igneous petrology have to do with widespread changes of the sedimentary rocks to hornfels, calc-hornfels, and garnet rock by contact and hydrothermal activity in the vicinity of igneous bodies. Igneous rocks generally in the Cerro Gordo area are so altered and leached that specific determination of the original rock type is difficult.

INTRUSIVE IGNEOUS ROCKS

Intrusive rocks of the Cerro Gordo mine may be grouped with reference to age and rock character in two major categories. These are: (a) the older granitoid rocks and (b) the later andesitic-dacitic porphyry dikes. The granitoid rocks are believed to have been emplaced in Cretaceous time concurrently with the Sierran plutonic rocks; the andesitic-dacitic dikes cut granitoid rocks intrusively at Cerro Gordo and are therefore younger. A third type of intrusive rock is represented by a single diabasic dike in the Cerro Gordo mine. It is probably older than the andesitic-dacitic porphyry group, but its age relation to the granitoid bodies is obscure.

GRANITOID ROCKS

As noted by Knopf (1918, p. 60) with reference to the Inyo Range generally:

> The average granitic rock is a quartz monzonite composed of plagioclase (andesine), orthoclase, quartz, hornblende and biotite. This rock is the variety most widely prevalent. From this average variety the granitic rocks range on the one hand to varieties that may appropriately be termed granite . . . and on the other hand to quartz diorite, diorite and hornblendite. The dark heavy hornblende-rich varieties are especially prevalent in Daisy Canyon (on the east flank of the range) and along the crest of the range northward from New York Butte.

More detailed study of igneous history in the southern Inyos will no doubt reveal a complex succession of plutonic intrusive events, involving a considerable range of magmatic differentiates. For example, in Granite Hill (pl. 2), 3½ miles east of Lone Pine, a light-colored albite-orthoclase granite was intruded after emplacement of the contiguous hornblende-biotite-quartz monzonite. Aplite dikes and sills are numerous west of Cerro Gordo, where they cut the hornfelsic strata and represent late stages of the plutonic activity. As shown along the Cerro Gordo road, 2 miles northeast of Keeler, the aplitic sills penetrate the late Paleozoic sedimentary rocks some distance away from the quartz monzonite.

The common textural variety of quartz monzonitic rock at Cerro Gordo was appropriately termed "monzonite porphyry" by Knopf (1918, p. 112). This type is shown typically in the small Hart Camp stock which penetrates Chainman shale half a mile northwest of Cerro Gordo (pl. 1). A small stock penetrated the Lost Burro formation 1½ miles north of Cerro Gordo, and somewhat larger bodies invade the Lost Burro and adjacent rocks northwest and northeast of the Bonham talc mines. Three quartz monzonitic stocks are also exposed east and west of Cerro Gordo; these are the Newsboy stock a mile east, the Cerro Gordo stock penetrating marble just south of Cerro Gordo Peak, and the small Ignacio stock in Chainman shale half a mile southwest of Cerro Gordo.

The Cerro Gordo stock has the composition of a syenodiorite at the north edge, where it is cut by an east-trending dacite porphyry dike near the marble boundary. As noted by Knopf (1918, p. 112) garnetization took place here.

The Ignacio stock is allied to hornblende-quartz monzonite. Its southern edge is roughly at the contact of Chainman shale and overlying Keeler Canyon formation. Garnetization is especially noticeable on the north side of this igneous body where the contact rocks are penetrated by the Ignacio mine workings.

Of special importance with respect to geology of the Cerro Gordo mine is the Union dike, a narrow granitoid body resembling the Hart Camp and Ignacio monzonite porphyries. The Union dike was apparently intruded along a roughly north-trending shear zone which later influenced ore deposition and subsequently became a locus of postmineralization fault movement.

LATER ANDESITIC AND DACITIC DIKES

Greenish-gray andesitic and dacitic dikes with porphyritic texture are perhaps the most notable features of igneous geology in the southern Inyo Mountains.

These dikes commonly strike northwestward and occur in large numbers. Dike swarms of this character are well exhibited on the west flank of the range, cutting light-gray dolomite at Front Ridge and the Dolomite Hills (pl. 2). Northwest-trending dacite porphyry dikes cut the Union dike at the Cerro Gordo mine. Other dikes of the andesite porphyry type occur along the east side of the range from the Belmont area north to the Bonham talc mines (pl. 1). These dikes evidently took advantage of the pervasive northwest elements of the shear pattern and are usually closely associated with faults and fissures having the same general trend. Talc mineralization (Page, 1951, p. 24–25) is rather closely associated with them, and in the Cerro Gordo mine some of the silver-lead ores lie in close proximity to a dacite porphyry dike. In most places these dikes are altered hydrothermally.

ROCK ALTERATION

Most rocks of the Cerro Gordo area were altered in some measure by processes related to plutonic intrusion. Purer limestone was either marbleized like the marble of the Lost Burro formation and some of the Pogonip, or silicated to produce calc-silicate rocks or "tactite" like those in the Keeler Canyon formation just west of Cerro Gordo. Dense banded hornfels and argillite were produced from shaly limestone and shale. Hornfelsing is especially well shown in the Permian Owens Valley formation along the lower west slopes of the range, where these strata were extensively intruded by granitic rocks and by aplitic sills and dikes. Around the Hart Camp stock northwest of Cerro Gordo the Chainman shale was silicified and changed here and there to argillite and fine quartzite. In fact, nearly all sandstone of the southern Inyo area was modified by additive silicic processes to dense quartzite like those of the Eureka quartzite and the Hidden Valley dolomite.

The greater part of the dolomite in this vicinity is accounted for by diagenesis rather than by hydrothermal activity long after marine deposition. Dolomite ascribed to diagenesis characterizes the Ely Springs dolomite, the Pogonip, and the Hidden Valley dolomite. East of the Inyo Range the Devonian Lost Burro formation is dolomitized in this way but is mainly nonmagnesian marble near Cerro Gordo.

Whereas most rock alteration in the Cerro Gordo vicinity is mild, there are small patches of strong contact metamorphism, as for example garnet rock bordering the Ignacio and Cerro Gordo plutonic stocks. Many of the smaller igneous bodies have themselves been altered hydrothermally, like the diabase dike and many of the northwest-trending andesitic and dacitic dikes. Such alteration is especially evident where these dikes are either mineralized or associated with talc deposits as in the Bonham Canyon vicinity. Late hydrothermal magnesian alteration took place along a kind of pipe or channel in the Cerro Gordo mine (Knopf, 1918, p. 114).

Intensity of thermal metamorphism increases appreciably northward, where granitoid rocks make up a larger part of the surface and where more strongly altered sediments occur as pendant masses.

Altered shale is not as a rule foliated near Cerro Gordo, but some of the shale in the marine and the land-laid Triassic rocks show phyllitic sheen on cleavage faces and in some places have a kind of incipient foliation. A well-defined fracture cleavage or sheeting developed in marble of the Lost Burro near the Cerro Gordo mine.

Metamorphism was sufficient to destroy most of the finer organic structure, and except where isolated spots escaped serious change, the fossils usually appear as ghosts. But even in the marble and hornfels scattered pods are found wherein fossil structures are fairly well preserved.

GEOLOGIC STRUCTURE

Paleozoic and Triassic strata of the southern Inyo Mountains are strongly folded and faulted. These factors together with plutonic igneous intrusion make the geologic structure and the stratigraphy extremely complex. From west to east across the range are a series of northwest-trending folds which vary greatly in magnitude and tightness. Character of the folds is influenced from place to place by change in competency of the rocks involved and to some extent by relation of beds to reverse faults or thrusts. In the medial and higher part of the range at Cerro Gordo, the broad asymmetrical south-plunging Cerro Gordo anticline is the principal geologic structure, forming a sort of backbone to the Inyo Mountains in this vicinity.

The Cerro Gordo anticline has a multitude of irregularities and is far from an ideal, textbook fold. Superposed on its flanks and nose are innumerable subsidiary drag folds and disharmonic folds, many of which are related in origin to compressional faulting. West and east of the major Cerro Gordo anticline, relatively tight flexures of lesser magnitude occupy the foothill belts. Some of these are especially tight or close, where they occur in more shaly strata of late Paleozoic and Triassic age. Except in drag folds and minor disharmonic flexures, most observed fold axes strike northwestward in conformity with orographic trend of this part of the Inyo Range.

Major northwest-trending faults divide the southern Inyo Mountains into linear blocks. Some of these northwest faults are either reverse faults or thrusts, others are normal faults of later origin. Faults having a roughly north trend are also very characteristic of the region, as is true in the Great Basin generally. Most of these north-trending breaks appear to be normal faults. Faults having a northeasterly trend are relatively uncommon.

Drainage patterns strongly reflect the geomorphic influence of fractures and faults in the southern Inyo Mountains. Especially evident are fault-controlled drainage features with northerly alinement; among these are Sand Canyon, the north fork of Slate Canyon, upper San Lucas Canyon, and Upland Valley. Fractures of the north-trending set show distinct geomorphic expression in the upper Keeler Canyon vicinity. Faults and fault zones with northwesterly trend give evidence of fairly recent movement. Among these is the frontal fault zone 1¾ miles northeast of Cerro Gordo.

CERRO GORDO ANTICLINE

The Cerro Gordo anticline involves strata ranging in age from Middle Ordovician or older to Middle Triassic (fig. 9). It has been mapped from its nose at Upland Valley northward to Hunter Canyon (fig. 3), a distance of 15 miles. North of Hunter Canyon where granitic rocks predominate the anticline loses most of its surficial identity.

Cerro Gordo is situated near the axis of the major anticline at a point where the south plunge is evident. The axial zone, where recognizable at the surface, is close to the crest of the Inyo Range; it can be plotted only approximately because of minor warps and faulting. On top of the range northeast of the Morning Star mine the Chainman shale and Perdido formation lie almost flat in an axial segment (pl. 1).

The important ore deposits of the Cerro Gordo mine occurred in the Lost Burro formation of the west limb, but probably not far west of the assumed axial plane. Reversal of dip from west limb to east limb is indicated in the Charles Lease tunnel, on the 660 raise level of the Estelle mine, and in the low-level Estelle tunnel.

The axial zone of the Cerro Gordo anticline strikes approximately N. 22° W. and at Craig Canyon occupies a position about 4½ miles east of the summit of New York Butte (fig. 3). Geometrical relation of the axial zone, as indicated near Cerro Gordo, to its apparent position in the low-level Estelle tunnel suggests a steep west dip of the axial plane. However, the general configuration of the fold in transverse section does not accord well with this view, as shown in figure 9.

Plotted sections through the range at Cerro Gordo illustrate south plunge and asymmetry of the anticline. Beginning of the plunge is manifested by attitude of Devonian marble and Tin Mountain limestone at San Lucas Canyon and Cerro Gordo Peak (pl. 1). A broad semidomical nose is well shown in the relatively incompetent Carboniferous and Permian strata southeast of the Morning Star mine at Upland Valley. Superimposed on the nose are many highly irregular drag flexures and disharmonic warps, too small to be shown on the map. As elsewhere noted the important ore bodies and ore channels in the Cerro Gordo mine rake southward in harmony with plunge of the Cerro Gordo anticline.

Lack of transverse symmetry of the anticline is well shown at Cerro Gordo by the steeply dipping Carbonif-

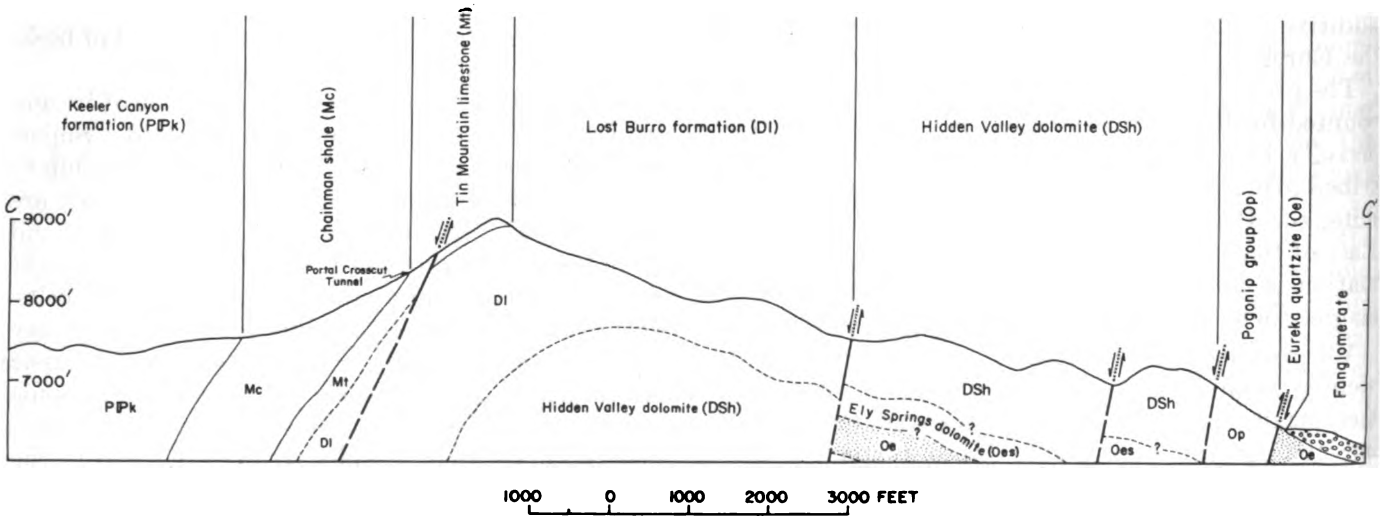

FIGURE 9.—Generalized southwest-northeast structure section (C–C') at Crosscut tunnel north of Cerro Gordo showing the Cerro Gordo anticline. Vertical and horizontal scales same. See plate 1 for location.

erous strata of the west limb as compared with low-dipping Silurian and Devonian beds of the east limb. Steeply dipping Chainman shale and Tin Mountain limestone of the west limb may be observed to advantage along the Pipeline trail to the Cerro Gordo springs. Disposition of the east limb is observable in the more rigid Silurian dolomite from 1 to 2 miles northeast of Cerro Gordo. These relatively low-dipping beds are undulant, with average dip about 20° east. Local steepening takes place in the Devonian marble at San Lucas Canyon near the axial plane, probably as a result of movement on the north-trending Cerro Gordo fault zone. Low dip of strata in the east limb is somewhat puzzling when the supposedly west inclination of the anticlinal axial plane is considered.

SUBSIDIARY FOLDS

On the limbs and crest of the Cerro Gordo anticline are many minor subsidiary flexures. Some of these are clearly drag folds, and some are related to reverse faults and minor thrust faults. The larger of the subsidiary folds are traceable axially for about 2 miles before dying out or being terminated by faults. Some have an echelon distribution.

Anticlinal axes are well shown in sheared Devonian marble 1½ miles northwest of Cerro Gordo and 2 miles northwest of the Bonham talc mines. The latter is overturned to the east. Some secondary folds have a west-dipping axial plane, but in many of the minor drag folds axial plane orientation is seemingly unrelated in any significant way to that of the parent Cerro Gordo anticline.

Northwest-trending subsidiary folds are spectacularly displayed in deeply incised canyons on the rugged east slopes north of Bonham Canyon. In upper Craig and Hunter canyons (fig. 3), the banded Devonian marble shows particularly well the fantastic irregularities which characterize these flexures. Between Hunter and Craig canyons a sharp contact between light-gray Devonian marble and dark-gray Tin Mountain limestone clearly reveals the reversal of dip at the crest of a contorted anticline.

Southeast of Cerro Gordo about 2½ miles the beds of the Keeler Canyon yielded to stress in such manner as to produce many small irregular flexures about the south-plunging anticlinal nose (pl. 1). On the west side of Upland Valley, south of the Cerro Gordo mining district map area, a sharp drag fold in the Permian Owens Valley formation lies contiguous to the north-trending Upland Valley fault, which probably originated as a reverse fault.

FAULTING IN THE CERRO GORDO AREA

A complex network of faults characterizes the southern Inyo Mountains; in no way can this be illustrated more objectively than by mapping the Cerro Gordo mine. Some older faults of the region are thrust or reverse faults; superposed upon these is an intricate pattern of later normal faults.

Faults of compressional origin are especially well shown along the west side of the range. Among these the Dolomite Canyon fault (pl. 2) gives stratigraphic evidence of reverse movement, whereas to the east of it thrust displacement is revealed by an outlier of Silurian dolomite resting upon Permian shale of the Owens Valley formation. Contiguous drag folds suggest that the Upland Valley fault (pl. 1) is also of reverse nature.

Near the Union chimney in the Cerro Gordo mine (fig. 16) are indications of reverse fault movement. However, nearly all the important faults mapped in the mine appear to be normal.

NORMAL FAULTS

Two principal groups of normal faults are recognized: (a) Those which have a northwest trend, and (b) those which strike roughly north. Northeast-trending faults are relatively few. Some of the important faults mapped underground in the Cerro Gordo mine were not identified with assurance at the surface. Among these are the Despreciada fault which strikes northwest and the Omega fault which trends north (block diagram, pl. 4). A possible southeast continuation of the Buena Vista fault at the surface has not actually been demonstrated to join the main Buena Vista fault as mapped underground.

The Bonham Canyon zone of intense northwest shearing and faulting is over a mile wide and comprises numerous faults, fractures, and dikes, only the largest of which are shown on plate 1. Brecciation and crushing is intensified along this zone by intersection with faults of the north-trending Cerro Gordo set. These structural factors no doubt have bearing upon localization of the Bonham Canyon commercial talc deposits (Page, 1951).

Multitudes of northwest-striking fractures, porphyritic dikes, and siliceous veins are related in origin to the pervasive northwest faulting. Among these are dike swarms of the west Inyo foothills as well as the porphyry dikes and quartz veins of the Cerro Gordo mine and its vicinity.

Cerro Gordo fault.—Master fault of the Cerro Gordo mine, this structural element figures importantly in detailed discussion of mine geology. The Cerro Gordo fault has an overall northerly trend and lies in a zone of shearing traceable for 6 miles from Soda Canyon north to Bonham Canyon. Northward continuation along Sand Canyon to Saline Valley (pl. 2) is suggested by physiographic evidence. With normal downthrow on the west, Chainman shale and limestone of the Keeler Canyon formation moved downward in contact with marble of the Lost Burro formation and Tin Mountain limestone of the footwall block. An inlying block of fusulinid limestone of the Keeler Canyon at the head of San Lucas Canyon clearly reveals the nature and magnitude of this displacement (pl. 1). The fusulinid limestones were dropped several hundred feet as a narrow wedge between the Cerro Gordo fault on the east and sympathetic faults with northeasterly strike on the west. On the west side of this downthrown block, the fusulinid limestones abut against Chainman shale (fig. 10). To the east, Chainman shale underlying limestone of the Keeler Canyon forms the hanging wall against marble of the Lost Burro. Mechanics of downthrow suggest those in a graben, with faults on opposite sides dipping inward.

From Cerro Gordo south to the Morning Star mine, alluvium and talus obscure the Cerro Gordo fault. Altered Chainman shale is again exposed to the west near the Morning Star mine, whereas Tin Mountain limestone forms the footwall. The only hanging-wall rock well shown near the fault between Cerro Gordo and the Morning Star is limestone of the Keeler Canyon in somewhat altered condition. A narrow band of Chainman shale is assumed to be present between the Keeler Canyon formation and the fault.

FIGURE 10.—View looking northeast toward Ubehebe through saddle at Cerro Gordo. Union tunnel dump and tramway terminus right of center middleground. Limestone of the Keeler Canyon formation occupies saddle; on left is dark-gray Chainman shale over sheared Tin Mountain limestone in jagged crags. Foreground underlain by altered Keeler Canyon beds.

Alluvial cover in upper San Lucas Canyon prohibits tracing of the inlying Keeler Canyon block northward (pl. 1). The down-dropped inlying block is believed to wedge out between faults of the Cerro Gordo fault set north of the Newtown shaft. Farther north beneath the alluvium of San Lucas Canyon, the Silurian dolomite and Devonian marble of the Lost Burro formation are probably in contact along north-trending faults.

CLEAVAGE OR SHEETING

Knopf (1918, p. 111) calls attention to the conspicuous cleavage or sheeting of the marble and limestone at Cerro Gordo. It shows a prevailingly northwest strike and steep westerly dip. Jagged crags northwest of Cerro Gordo illustrate how this structure has influenced sculpture of the carbonate rocks. In the mine workings especially, the close-spaced cleavage planes are easily confused with bedding. Many surface exposures reveal west-dipping bedding and cleavage in the marble showing no appreciable differences of attitude, except in minor drag folds. Prevailing northwesterly strike and steep westerly dip of the cleavage suggests possible genetic relation to fold axial planes.

AGE OF DEFORMATION

Folding and thrusting took place in the Inyo Mountains region after deposition of the Middle Triassic strata. By analogy with the nearby Sierra Nevada Mountains, these events are assumed to have taken place either in Late Jurassic or Early Cretaceous time. In this connection, dating of the folded volcanic rocks and land-laid beds which rest upon the marine Middle Triassic rocks would be significant. This thick sequence which has thus far yielded no fossils may include strata of Jurassic (Dunlap) age. Normal faulting followed thrusting and reverse faulting. Normal faults have been especially active in late Tertiary to Recent time. As elsewhere noted the Union monzonite porphyry dike was seemingly emplaced very early on a north-trending zone of weakness which later became a locus of movement on the Cerro Gordo normal fault (pl. 1). Normal faults of the Buena Vista and Despreciada sets (block diagram, pl. 4), with northwest strike, cut and offset the older Cerro Gordo fault. It is possible that some north- and northeast-trending high-angle faults could have originated early under compression as tear faults, with later renewal of movement as normal faults under a different stress pattern. Heavy horizontal mullions on the north-trending Omega fault in the Cerro Gordo mine indicate a strike displacement as the last event.

METAL MINES OF THE CERRO GORDO AREA

CERRO GORDO MINE

HISTORY OF THE CERRO GORDO MINE

Owens Valley tradition, unsupported by reliable documentary records, dates the Cerro Gordo strike somewhere in the interval 1861–66. Although actual time and circumstances of this major discovery are quite obscure, Spanish names like San Felipe, Santa Maria, San Lucas, and Ignacio borne by these mines from the earliest days provide testimony of Mexican location. According to one account the discovery was made in 1865 by one Pablo Flores and Mexican companions, said to have located jointly the San Felipe claim (Chalfant, 1933, p. 277–283). Other records (Loew, 1876; Knopf, 1918, p. 108) place the time of discovery by Flores as 1866. The nearby Ignacio and San Lucas silver mines are believed to have been worked by Mexicans during 1865 and 1866 (Chalfant, 1933, p. 278).

Glowing accounts by Mexican prospectors attracted eager and enterprising Americans to Cerro Gordo during the post-Civil War period of vigorous mineral exploitation. Promise of cheaper transportation in 1869 upon completion of the transcontinental railroad further stimulated metal mining. Such was the rapidity of development that by 1870 four lead furnaces were in operation at Cerro Gordo, and plans were already being made to link Owens Valley with the coast by rail. By 1872, 11 mines were active in the Cerro Gordo district, and large shipments of silver-lead bullion ("base bullion") were made regularly to San Francisco via Los Angeles and San Pedro.

From 1869 to 1876 Cerro Gordo enjoyed 7 years of great prosperity. Los Angeles, 200 miles across the desert, profited immensely by the lucrative supply and transportation business, salutary factors in its growth from village to important city (Nadeau, 1948). San Francisco also shared the economic benefits. For a brief period during the decline of metal production at Eureka, Nev., Cerro Gordo became in fact the main source of lead and silver for the Selby Smelter on San Francisco Bay. By 1872, the Union mine owned by Belshaw and Beaudry, the Santa Maria controlled by the Owens Lake Silver Mining and Smelting Co. of New York, and the San Felipe, part of whose stock was ultimately acquired by the New York company, had outstripped all others in importance. The Union mine easily supplied both the Belshaw-Judson furnaces and the independent Beaudry smelter. Furnaces of the Owens Lake Silver Mining and Smelting Co. on the Owens Lake shore near Swansea were fed by ore from the Santa Maria mine.

FIGURE 11.—View of Cerro Gordo looking east. Rugged slopes in distance are south-dipping marble of the Lost Burro formation overlain on extreme right by Tin Mountain limestone. Saddle on left underlain by Keeler Canyon formation. Cerro Gordo fault passes beneath large Belshaw shaft dump (right center).

Notwithstanding richness of bonanza ore, numerous obstacles to successful operation existed. Among these were ruggedness of mountain-top terrane, scarcity of water, remoteness from large centers of population, and excessive transportation costs. Heavy snows are not uncommon at the high altitude of the mines, and the summer months brought cloudbursts to wash out roads and installations at lower elevations.

Water was first transported to Cerro Gordo by pack train and was obtained during winter and spring by melting of snow in tubs. Early in the history of this mountain-top camp a pipeline was laid from the small Cerro Gordo Springs[2] which emerge from Chainman shale near the Inyo crest, 3½ miles north of Cerro Gordo. By pumping to tanks on the mountain top, gravity flow was achieved. This water system, highly susceptible to freezeup, was reconstructed several times and continued to supply water until the 1930's. Pack trains also carried water from the Cerro Gordo Springs and from Miller Springs in a deep canyon, 11 miles north of Cerro Gordo. In 1874, the year of peak production, an elaborate water system was constructed at a cost of $74,000 (Hanks, 1884, p. 365). Hooker steam pumps lifted the water a vertical distance of 1,860 feet in three stages to the summit above Miller Springs, whence it flowed by gravity (Nadeau, 1948, p. 115–117). During the brief period of its operation, water is reported to have been cut to 3 cents per gallon, less than half the previous cost. Good water, sufficient for camp use, was eventually found on the 700 level of the Cerro Gordo mine near the Belshaw shaft. During recent years the pipeline from Cerro Gordo Springs remained

[2] The Cerro Gordo Springs include Mexican Spring, Belshaw Spring, and Cronn Spring (pl. 1). Little more than seeps, they are likely to dry up completely by late summer.

in disrepair, and mine water has been relied upon to supplement that hauled by truck.

To reduce and control freight rates, which ran from 3 to 6 cents per pound (Chalfant, 1933, p. 280), a group of mine owners formed a transportation company known as the Cerro Gordo Freighting Co. It made large profits and remained active until building of the narrow-gage railroad from Nevada about 1883. During the years of Cerro Gordo prosperity, 56 large freight wagons, each drawn by from 16 to 20 animals, are said by Chalfant to have been on the road. To further expedite bullion shipment a small steamer carried the bars across Owens Lake from Swansea to Cartago on the southwest shore. But apparently even these facilities were incapable of moving the metal as rapidly as desired. Visitors to Owens Valley in the 1870's observed great quantities of bullion bars stacked like cord wood on the lake shore, or used to construct temporary shelters for company employees.

Great success in exploitation of the Cerro Gordo bonanza silver-lead deposits was achieved by M. W. Belshaw and Victor Beaudry, whose names appear prominently in the early annals of Owens Valley and Los Angeles mining affairs. During the initial period of development from 1869 to 1871, these extraordinarily enterprising pioneers conducted mine examinations and carried out the metallurgical experimentation which resulted in successful fluxing and smelting on a commercial scale. Thereafter they secured joint control of the Union mine and others which ranked with the Union as the best in the district. Belshaw and Beaudry were unable, however, to acquire title to the San Felipe mine, a factor which later resulted in prolonged litigation.

Mining claims and litigation.—Initial lack of district regulations and a uniform system of mine location at Cerro Gordo virtually assured conflict between operators. There was little respect for claim lines in the early days, and furthermore, occurrence of the richest ores in irregular pipes and pockets did not lend itself well to customary vein or lode methods of location. In the case of the pipelike Union ore body its downward course could not have been accurately predicted in advance of underhand stoping from the discovery pit. On the other hand the San Felipe mine was opened on the clearly defined northwest-trending San Felipe quartz vein (fig. 16). Accordingly the initial Mexican claim is very long and narrow as it follows the vein only a short distance away from the Union discovery pit. The San Felipe owed its importance to siliceous silver ores useful as smelter flux and to the eventual legal aspects of proximity to the rich Union lead chimney, making it a reasonable assumption the latter would be found to intersect the San Felipe vein in depth.

The long narrow Santa Maria, another early Mexican claim, was laid out west of and roughly parallel to the San Felipe along the northwest-trending Buena Vista fault (fig. 16). Its ores were also very pockety, distributed in irregular fashion along the fault zone.

Cerro Gordo was first included in the Lone Pine mining district organized in 1866, but in 1872 a separate Cerro Gordo mining district was formed. Regulations adopted in 1872 brought stricter conformity with new Federal requirements and encouraged greater respect for titles and legitimate claim surveys. By 1873, Deputy U.S. Mineral Surveyers were active in the district.

Encroachment litigation involving the Union Co. and the San Felipe, imminent from the start, began in January 1873 (Knopf, 1918, p. 109; Chalfant, 1933, p. 281; Nadeau, 1948, p. 110–114). The San Felipe Co. claimed discovery title to the Union mine. Leading contestant was the Owens Lake Silver Mining and Smelting Co., owner of the Santa Maria mine and likewise in control of the San Felipe. After lengthy testimony as to whether the Union and San Felipe were on the same lode or separate lodes a verdict was rendered in favor of the San Felipe. The case was appealed to the California State Supreme Court. After prolonged legal conflict it was finally settled in 1876, when the opposing interests combined to form the Union Consolidated Co. Shown on maps dating from the 1880's, the revised Union claim is more nearly of the conventional form, though narrower and longer than a standard lode claim. With side lines bearing a few degrees east of north it crosses both the original San Felipe and Santa Maria claims in such manner as to include much of the better ground worked in the boom days. The Union claim as resurveyed is thus in accord with trend of the principal ore bodies as well as the Union dike and Cerro Gordo fault.

Decline and shutdown in 1878.—Peak of the Cerro Gordo mining activity was reached in 1874 during litigation. At that time most of the rich silver-lead ore bodies were approaching depletion and shortly after consolidation the end of the bonanza period was clearly in view. Production statistics compiled by Belshaw for 1876 manifest a decided fall off in ore grade. In August 1877, destruction of the Union hoisting works by fire foreshadowed shutdown of the Union furnaces, which took place in February of the following year. Beaudry's smelter just west of Cerro Gordo (fig. 16) is said by Nadeau (1948, p. 245) to have continued in operation until 1879.

Mining activity 1879 to 1910.—For 30 years, sporadic mining activity at Cerro Gordo was in the hands of small companies and leasers. Existing underground maps probably dating from the 1880's indicate continued exploration for inferred deeper Union and Santa

Maria ore. A new vertical Union shaft, today appropriately called the Belshaw shaft, was sunk to a depth of 900 feet about 1877 (Nadeau, 1948, p. 245). Drives from the 700-foot shaft level failed to pick up the inferred downward continuation of the Union chimney.

The Jefferson chimney had probably been tapped on the Buena Vista tunnel level (fig. 16) by 1875 (Loew, 1876, p. 63), but this rich ore channel was not followed downward until much later (pl. 4).

Prospecting northward in San Lucas Canyon along the Cerro Gordo fault met with indifferent success. Near the ridge crest south of Cerro Gordo (pl. 1) lay the wide and rather spectacular Castle Rock siliceous vein.[3] The Castle Rock vein was intensively prospected and attempts were made somewhat later to cut this surficially large vein at lower altitudes in the Morning Star mine, the Charles Lease tunnel, and the low-level Estelle tunnel.

About 1883 the narrow gage Carson and Colorado Railroad (Kneiss, 1946, p. 72) reached Keeler, its owners inspired by hopes for a rebirth of Cerro Gordo and opening of mines in the Darwin district. Thereafter renewed efforts to find ore were made at Cerro Gordo (Hanks, 1884, p. 365). Goodyear (1888, p. 228), who revisited Cerro Gordo in 1888, observes that the Union was under lease and shipping small quantities of ore by rail to the Selby Smelter.

Until 1905, records of later day activity at Cerro Gordo are meager. In that year the mine was acquired by the Great Western Ore Purchasing Co. As noted by Knopf (1918, p. 110):

> * * * * A small production was made by this corporation in 1907. Subsequently the property was taken over by the Four Metals Mining Company which erected a 200-ton smelter just east of Keeler and connected it with the mine by an aerial tramway. This company attempted to smelt the old slags from the Cerro Gordo and to work the mine but went into insolvency. L. D. Gordon and associates, who had obtained from the Four Metals Mining Company a lease to extract the zinc ore of the mine, then took over the property by purchase of the bonds of the insolvent corporation * * * *

Mining activity 1911 to 1949.—Important ore discoveries have twice revived the Cerro Gordo. Upon neither occasion did richness or quantity of ore approach that extracted during the bonanza period. Responsible for the revival from 1911 to 1919 were L. D. Gordon and associates, who successfully mined rich zinc carbonate, for which an eastern market existed. During these years Cerro Gordo was a major source of the highest grade zinc carbonate ores produced in this country. In 1914 Gordon and associates acquired title to the properties and effected a reorganization as the Cerro Gordo Mines Co.

An electric powerline was completed to Cerro Gordo in 1917, electricity replacing steam for hoisting, air compressors, and tram operation. The old and inefficient aerial tram was supplanted by a 5½ mile Leschen tram which transported supplies and ore to and from the railway terminus at Keeler. In addition to zinc, new silver-lead ore bodies were discovered and successfully mined during the 1911–1919 period. Among these were ores in the Jefferson diabase dike and the Jefferson chimney (fig. 16 and pl. 4). So far as known there was no significant development of lead ore in the old Union chimney (China stope), doubtless well cobbed by leasers during the 30-year interval following abandonment by the Union Consolidated. For 8 years between 1920 and 1928 the Cerro Gordo Mines Co. followed by various leasers continued to work old stopes in a modest way. By 1925, an important ore body was found in the La Despreciada claim (pl. 4), west of the old Cerro Gordo, at that time under lease to the Estelle Mines Corp. According to T. L. Chapman (written communication, 1930), the Cerro Gordo Mines Co. extracted ore illegally from the La Despreciada but was forced to stop work. As a final settlement about 1928, the Estelle Co. is said to have been given a 30-year lease and bond on the Cerro Gordo property.

The third and last period of significant Cerro Gordo productivity began in 1929 and lasted until 1933. Rich La Despreciada ore was extracted during this 5-year interval, when the Estelle Mines Corp. operated in conjunction with the American Smelting and Refining Co. From 1933 until 1936 the Estelle Mines Corp. continued to operate independently. Since the La Despreciada find no important ore discoveries have been made at Cerro Gordo.

After 1933, small leaser activity continued and further examination of the old Cerro Gordo workings was conducted by geologists. During World War II the mine was opened for examination, sampling and diamond drilling by Goldfields of South Africa in 1944, and by W. C. Rigg and associates from 1946 to 1949. Only small shipments were made before closing.

MINE DEVELOPMENT AND ORE SEARCH

The richest ore channels of the Cerro Gordo district were discovered at the surface and initially mined downward. Ore was feverishly extracted by primitive underhand stoping from open pits and tunnels. Except possibly the Union, operations were at first conducted without apparent plan. After consolidation in 1876, many of these workings were connected to form a com-

[3] The Castle Rock vein is erroneously referred to in mining reports as the "San Felipe vein" (Tucker and Sampson, 1938, p. 438).

plex network ultimately many miles in extent, much of it tributary to the Belshaw vertical shaft.

In 1870, the Union mine, most important of the district, was developed by a vertical shaft sunk partly in Union chimney ore, and by the Union tunnel, which struck the ore channel 175 feet vertically below the discovery pit (fig. 16). The Union tunnel crossed Santa Maria and San Felipe ground to reach the Union ore, and may have been used also by the other companies.

Good accounts of the Union operation are given by Goodyear (1888, p. 254), who visited the district in 1870; by Raymond who examined the mines in 1872; and by Loew (1876). Goodyear calls attention to absence of a definite Union vein. Ore was removed from very irregular "chambers" in shattered limestone filled with cracks in every direction and in some places having large open cavities. Raymond (1873, p. 18) noted that the ore chimney containing these pockets or "chambers" had a northwesterly strike and steep southwesterly dip near the surface, steepening to near vertical at the Union tunnel level and downward therefrom. A low-dipping shoot of exceedingly rich ore extended more than 100 feet west from the main near-vertical chimney. 200 feet below the Union tunnel level. According to Raymond, ore pockets of extraordinary richness and extent were being extracted 210 feet below the Union tunnel downward to a level 275 feet below the tunnel. These ore bodies were in places 40 feet wide, nowhere less than 15 feet wide. This was in 1872; evidently the very rich Union ore channel was soon to be bottomed above the present 550 level. Ore was hoisted in buckets to the Union tunnel, trammed to the surface where fines were sorted from massive galena, and transported thence to the Belshaw-Judson furnaces situated only 450 feet away.

The neighboring San Felipe vein (fig. 16) was mined from pits and trenches. It was reached also by the Union tunnel [4] and the Zero tunnel (pl. 3), the latter driven east near the present Belshaw shaft collar to cut the San Felipe vein and Zero fissure vein (fig. 23).

The Santa Maria (fig. 16) comprised a number of pockety ore bodies distributed for several hundreds of feet along the west side of the Buena Vista fault. These near-surface ores were mined in helter-skelter fashion by open pits, trenches, and tunnels. Among the more extensive Santa Maria workings are those of the Buena Vista and Santa Maria ("86") tunnels, and the Omega tunnel (pl. 3 and fig. 16). The long Omega tunnel was driven south at lower altitude from San Lucas Canyon to explore beneath the Santa Maria near-surface pockets and below the footwall shale of the Buena Vista fault. In later years a connection was made eastward with the Union chimney, possibly by leasers.

[4] Some private mining reports and maps refer to the Union tunnel as the "San Felipe tunnel."

Several of the Santa Maria ore pockets had a footwall of shale on the east. In the early days, before recognition of the Buena Vista fault, the relation of these pockets to the main ore channels was puzzling and is in fact not clearly understood today. Because most of the Santa Maria workings are filled it is appropriate to introduce Raymond's description of 1873:

> The Santa Maria is located lower down Buena Vista Mountain, and nearer to the town of Cerro Gordo than the Union. It runs generally parallel to the latter, and is separated from it by a stratum of clay-shale, which varies in thickness from 30 to 120 feet, and a varying stratum of limestone. The Santa Maria lies closely to the slate, which dips on the surface 64° west. The upper works in the deposit show that it is completely broken down near the surface, so that it is here much wider than lower down, and apparently tipped over, i.e., it dips to the east; but at a depth of about 60 feet the dip changes to the west, becoming conformable with that of the slate. Much work of an irregular character, evidently wanting in a uniform plan, has been done on this deposit. Several tunnels and shafts have developed the following so-called chambers, some of which are now worked out. In going along the vein from north to south they are:
>
> 1. The Santa Maria chamber, 25 feet wide and 50 feet long.
> 2. The front chamber, 15 feet wide and 40 feet long. This ore-mass lies, not on the strike of the vein, but west of it, and lower down the hill. It is in soft, gravelly ground and broken rock, evidently a part of the former outcrop, which is tipped over and has been partly covered up by other detritus.
> 3. The Schneider pocket, on the course of the vein.
> 4. The Buena Vista pockets, 180 feet long and from 5 to 30 feet wide.

The Jefferson ore channel, a surface discovery, was doubtless mined downward in the early days to the Buena Vista tunnel level as a narrow pipe. Evidently the lower and richer ore shoots of this channel were not developed until after the initial boom stage.

Surface pits today mark the Union [5] and Jefferson ore channels. Pits and trenches also remain as evidence of extensive digging on the San Felipe vein. Old engineering maps show a large "Belshaw pit" or "Santa Maria pit" north of the present Belshaw shaft. It is possible that much of the Santa Maria ore smelted at Swansea on Owens Lake came from this pit. A "Union pit," also filled, lay near the Union tunnel entry.[6]

North of the Union discovery pit, the Bullion tunnel (fig. 16) was driven east to the Bullion fissure (fig. 19). A south drift along the fissure probably connected with a carbonate zinc channel just east of the Union lead

[5] The large pit southeast of the Union opening is a limestone quarry (fig. 16). In recent years broken rock was drawn into bins below on the Union tunnel level and transferred to Keeler by tram.

[6] Some private reports refer to the minor ore body extracted from this pit as the "Union ore body." It lay west of the Buena Vista fault and was not connected with the main Union ore channel to the east.

channel. Apparently the Bullion fissure was not especially productive north of the Union, but these workings probably served in later years for extraction and haulage of zinc ore.

In San Lucas Canyon north of Cerro Gordo are the Upper Newtown tunnel (fig. 16) and the Newtown mine. These workings are separate from the Cerro Gordo mine, but fissures explored by them are believed to be continuations of north-trending fissures partly responsible for Cerro Gordo mineralization. There is no record of significant production from the Newtown openings.

After company consolidation, the Belshaw or new Union shaft was sunk vertically 900 feet; it remains the principal mine entry. Levels extend from the Belshaw shaft at 86, 200, 400, 550, 700, and 900 feet below the collar (pl. 3; pl. 4; fig. 18). After 1911 a winze was sunk to an 1,100-foot level. In 1923 the 660 raise level above the Estelle tunnel was driven north to the Cerro Gordo (fig. 18), but is not known to have connected with the overlying Cerro Gordo 1,100 level.

In later years, raises and winzes connected the Cerro Gordo 900 level with the Omega tunnel, providing a required second mine exit. The Omega tunnel was connected with the main Cerro Gordo 200 level.

Exploration for an inferred deeper continuation of the Union channel was the leading objective of the Belshaw shaft. Since the time of Belshaw's operation, loss of the Union ore some 500 feet below the surface was generally accounted for by postmineralization faulting. Exploration headings from the Belshaw shaft levels encountered small ore pockets, especially on the San Felipe vein, the so-called Santa Maria vein and the Zero fissure. Drifts along the Cerro Gordo fault or "contact vein" were wholly unsuccessful. After discovery of new ore in the Jefferson diabase dike about 1911, there was much drifting and raising on this igneous body (pl. 4; fig. 18). The favorable but less fractured marble east of the Cerro Gordo fault zone was explored by long crosscuts without success, and abortive drives were made into the black shale west of the Cerro Gordo fault.

About 1910, upon recognition of the economic value of the carbonate zinc, tortuous stopes were opened just east of and adjoining the old Union lead channel, now known as the China stope (fig. 18; fig. 24). The larger zinc stopes appear to have extended upward roughly on bedding incline from the 550 level to the south end of the Bullion tunnel workings.

The Despreciada ore bodies (pl. 4) were discovered in 1925 by fortunate west drives from the Cerro Gordo 900 and 700 levels. This new ore channel was prospected also from the 550 level. At the same time the Estelle Co. planned to tap Despreciada ore at a lower elevation from the 660 raise level of the Estelle tunnel workings.

Since about 1933 various leasers have driven short and unsuccessful exploration headings from the 200, 400, and 550 levels. These were clearly designed to test geological theories regarding postmineralization faulting of the Union and Jefferson channels. In 1944 diamond drilling was conducted at the south end of the mine from the 900 level, in search of deep inferred fault segments of the Jefferson ore. Although encouraging lead seams and possible Jefferson drag ore were found, no minable ore body was discovered. Search by leasers near the junction of the Union channel and San Felipe vein during the years from 1946 to 1949 was also unsuccessful.

HISTORY OF THE CERRO GORDO SMELTERS

Concentration by direct on-the-spot smelting of Cerro Gordo ore accounts in no small measure for highly successful operation in the years before 1879 (Ingalls, 1908 p. 146). Crude "vasos" first employed by Mexicans may be seen near the Ignacio mine west of town (fig. 16). Primitive methods such as these satisfied Belshaw and Beaudry that the ores were rather easily reduced. After metallurgical experimentation two smelters were set up, that of Belshaw and Judson and the separate unit of Beaudry. Remains of a single masonry stack mark the site of Beaudry's works at the west edge of town. The larger Belshaw-Judson plant lay near the divide, north of the Union tunnel. The Swansea smelter operated by the Owens Lake Silver Mining and Smelting Co. was situated at Smelter Hill on Owens Lake, 8 miles from the mines (pl. 2). Slag piles which long marked the position of all three plants have now been almost entirely removed. A well-preserved stone reverberatory furnace with hillside flue lies at the road bend near the Omega tunnel portal. Unknown is its history, though it may have been designed for slag roasting before that process was abandoned.

Eilers (1873), who visited Cerro Gordo in 1872, describes metallurgical treatment employed by the smelters. His appropriate suggestions for increasing efficiency may well have been adopted in the ensuing years.

Prior to 1871, part of the hand sorted ore was slag roasted in Mexican-type "galenadores," a form of reverberatory furnace. Final smelting was accomplished in small blast furnaces charged with a combination of slag, raw lead ore, siliceous silver ore, limestone, and charcoal. Preliminary slag roasting is said to have been discontinued in 1871 (Chalfant, 1933, p.

279). Loss of lead and silver in blast furnace slag was at first very high; according to Goodyear (1888, p. 256) the slag not uncommonly contained 35 percent lead. Improvement in techniques of smelter charging and the invention by Belshaw of a new water-jacket furnace brought greatly decreased losses.

A considerable proportion of silica was believed necessary at first in smelter charges. This material came from the San Felipe, Ignacio, and Perseverance (San Lucas) as well as various quartz vein mines in the neighboring Belmont district. The siliceous ores contained small amounts of lead, antimony, and copper, together with good silver and gold values. Thus the silver and gold would theoretically enrich the blast furnace product. In a later stage of Cerro Gordo smelter history it was reported that elimination of siliceous ore actually gave better lead and silver recovery.

Silver-lead bullion ("base-bullion") was cast in bars weighing from 80 to 90 pounds (Eilers, 1873, p. 355). So far as known, parting of silver and gold from the lead was not attempted on a large scale at Cerro Gordo.

The very large amounts of charcoal required for smelting were prepared in the vicinity of Cerro Gordo (Eilers, 1873) from piñon pine and mountain mahogany. As at Eureka, Nev., and other early-day western smelting towns, the charcoal industry attained major importance. Little is known of charcoal production at Cerro Gordo, but the product is said to have been exceptionally good for smelting. At Eureka, Nev., the charcoal burners were mainly Swiss-Italians brought in for that purpose, and many of the present inhabitants are descended from them. Within a radius of several miles from Cerro Gordo may be seen the remains of charcoal flats or pits, and occasionally a large carefully prepared wood pile ready for burning. Charcoal pits were usually located where loose gravel was available to cover the wood piles and prevent complete combustion. Permanent beehive stone ovens were seldom employed.

Shipment of Cerro Gordo smelter slag began on a large scale about 1916; over a period of 4 years, 33,000 tons were recovered. Knopf (1918, p. 109) calls attention to increasing metal content with depth in the slag dumps. Outer or upper layers carried not more than $5 per ton in silver and lead, whereas the more deeply buried material was much richer, in 1916 and 1917 becoming a source of considerable profit. Initial high smelter loss in metal was thus confirmed. Slag shipment continued at intervals until recently, with the result that little trace of the smelter dumps now remains, either at Smelter Hill or Cerro Gordo.

CHARACTER AND GRADE OF ORES

Contemporary descriptions of rich Cerro Gordo ores and accounts of their occurrence and extraction are given by Raymond (1873, p. 18–20) and by Goodyear (1888, p. 252–256). A more comprehensive study of Cerro Gordo ore occurrence and mineralogy was made by Knopf (1918, p. 106–116). As the source of relatively uncommon minerals such as caledonite [$Cu_2Pb_5(SO_4)_3(CO_3)(OH)_6$], linarite [$PbCu(SO_4)(OH)_2$] and leadhillite ($Pb_4(SO_4)(CO_3)_2(OH_2)$), Cerro Gordo came to be recognized as an important California mineral occurrence (Murdoch and Webb, 1948, p. 27–28).

Cerro Gordo ores may for convenience be grouped in four categories as follows:

1. Massive silver-lead ores (Union type).
2. Diabase dike silver-lead ores.
3. Siliceous vein ores (San Felipe type).
4. Carbonate zinc ores.

The greater part of Cerro Gordo production was massive silver-lead ore of the Union type, which consisted in the main of argentiferous galena together with the carbonates and sulfates of lead. Raymond (1873, p. 18–19) describes a Union ore body observed by him as follows:

> * * * The ore is very solid, being either reddish-yellow carbonate, or pure gray carbonate, lying in great blodges in the former. The masses of the latter kind have frequently a diameter of from 3 to 6 feet, and always show a concentric arrangement; i.e., every mass of this kind, which has been cut through by the excavations shows concentric rings around an interior nucleus, (generally a small lump of unaltered galena,) the rings being somewhat darker than the main mass. This arrangement presents a beautiful aspect, and though common with gray carbonate of lead, when lying in a ferruginous gangue, it is not often seen on as large a scale as exposed in the Union. The carbonate-ores of the Union, on account of their friability termed "fuse-ores" by the miners average, as delivered to the furnace, about 25 ounces of silver per ton, and the galena from 50 to 80 ounces.

Goodyear (1888, p. 254) noted the presence in 1872 of arsenical and antimonial lead, pyrite, and in some places a small amount of copper. No mention of zinc is made in early reports, either as sulfide or carbonate. It is believed that these massive ores contained very little silica. In 1947 it was noted, however, that quartzy siliceous incrustations were present here and there along walls of old stopes in the lower part of the Union channel.

Knopf (1918, p. 114) observed about 1912 that shoots of lead ore exposed near the surface consisted predominantly of galena but contained some tetrahedrite, sphalerite, and pyrite. The galena exhibited a distinct sheared structure which wrapped around the sphalerite and tetrahedrite. Among the oxidation products was bindheimite, the hydrous antimonate of lead.

A rich lead seam sampled by the writer in 1947 near the bottom of the Union chimney consists of a central mass of finely crystalline galena encrusted with cerussite, anglesite, and limonite.

The diabase dike ore was being extracted from the Jefferson diabase dike during visits by Knopf)1918, p. 114) in 1912 and 1913. His description is in part as follows:

> The ore consists largely of galena but contains some tetrahedrite, zinc blende, and pyrite. Galena is intimately intergrown with the tetrahedrite. Cerusite, anglesite, and bindheimite (the hydrous antimonate of lead) are the most common oxidation products; linarite (the deep azure-blue sulphate of lead and copper), caledonite (the pale-green sulphate of lead and copper), and chrysocolla are present in smaller quantity.

The predominant primary mineral is galena, with subordinate zinc blende, tetrahedrite, and pyrite. Production records for 1914 and 1915 show a recovery of more than 200 tons of copper, which may be accounted for in considerable part by tetrahedrite in diabase dike ore.

Siliceous veins of the San Felipe type are characterized by tetrahedrite and argentiferous galena with malachite, azurite, and other products of oxidation. The metallic minerals are inclosed in a quartz and barite gangue. Veins of this kind are common at Cerro Gordo and in adjoining territory of the Inyo and Panamint Mountains.

Carbonate zinc ores of the Cerro Gordo mine were made the subject of special study by Knopf (1918, p. 106–108; 115–116), who notes that recognition of commercial value of these deposits about 1910 led to revival of mining operations. The zinc ore consisted mainly of smithsonite which was in part quite pure. Also present were limonite, calcite, small quantities of calamine and hydrozincite, and uncommonly aurichalcite and willemite. Knopf's description is as follows:

> The zinc ore consists essentially of the carbonate, smithsonite, with limonite and calcite as impurities. The ore is fine grained in texture and dead white in color, and hence is known as "dry bone." Much of it is characteristically banded or laminated; the laminae range from mere films to half an inch thick. The lamination is curiously convoluted and the patterns produced by the convoluted laminae are not uncommonly irregular-shaped closed ellipses. The occurrence of patterns of this kind proves that the lamination did not result from the replacement of a bedding structure in the marble nor from a sort of photographic "development" by the replacement of bedding structure that was previously imperceptible. The lamination strongly suggests that it originated by the rhythmic precipitation of the zinc carbonate by the calcite of the marble. Vugs are common in the fine-grained smithsonite, and some are lined with more coarsely crystalline smithsonite and with calamine.

Silver-lead ore grade.—Eilers (1873, p. 356; Knopf, 1918, p. 109) observed that the Union ore contained about 34 percent lead in 1872. At this time, Raymond (1873, p. 18–19) estimated the silver content of oxidized parts only to be about 25 ounces per ton, whereas the galena was much richer, carrying from 50 to 80 ounces per ton (Eilers, 1873, p. 355). Union ore grade seems to have improved with depth, for of 12,171 tons extracted in 1873 and 1874 average assay was 47 percent lead and 87 ounces of silver per ton (Knopf, 1918, p. 109). By 1876, lead ore grade fell appreciably, 9,950 tons as mined averaging only about 21 percent lead. Ores containing 20 percent lead and $60 in silver per ton were reported (Hanks, 1884, p. 365) after the Union Consolidated shutdown.

In 1872, the silver-lead base bullion smelted at Cerro Gordo is said to have contained 140 ounces of silver per ton (Raymond, 1873, p. 21).

Ore shipments totaling 11,610 tons made from 1928 to 1936 are reported[7] to have averaged 38.1 percent lead with 28.6 ounces of silver and 0.073 ounces of gold per ton. These ores were shipped by the Estelle Mines Corp., American Smelting and Refining Co., and by various leasers; in all probability they came largely from the La Despreciada stopes and may be considered fairly representative of the massive type. Evidently much of the primary galena had oxidized to lead carbonate and sulfate, yielding loose easily extractable "sand carbonate" ore. Average metallic content of these ores is given in the following table:

	Assay per ton	
Gold	troy ounces per ton	0.073
Silver	do	28.6
Lead	percent	38.1
Zinc	do	2.8
Sulfur	do	3.8
Iron	do	16.0
Calcium oxide	do	2.5
Insoluble	do	9.4

Zinc ore grade.—According to Knopf (1918, p. 116), zinc ore shipped in 1913 averaged 35 percent zinc, about 10 percent ferric oxide, and 2 percent calcium carbonate, with only a trace of lead and no silver or gold. The Estelle Mines Corp. extracted 1,432 tons of zinc ore from 1933 to 1936. This ore averaged 40.1 percent zinc.

PRODUCTION OF THE CERRO GORDO MINE

The Cerro Gordo mine was California's great silver and lead producer in a decade of active mining which followed the Civil War. Beginning with 1869, total recorded silver yield is of the order of 4,400,000 ounces and total lead production roughly 37,000 tons. Of these totals, more than half of the lead and about three-fourths of the silver were produced in the boom years from 1869 through 1876. Gold was an important smelter byproduct. Beginning with 1911, total zinc production has been about 12,000 tons. Copper was recovered only in the later periods of operation.

[7] Production and grade statistics made available in 1944 by Mr. J. B. Stone, of the Golden Queen mine, Mojave, Calif.

TABLE 5.—*Production of the Cerro Gordo mine, Inyo County, Calif.*[1]

Period (inclusive)	Ore (short tons)	Smelter slag [2] (short tons)	Concentrates [3] (short tons)	Lead (short tons)	Zinc (short tons)	Copper (short tons)	Silver (troy ounces)	Gold (troy ounces)
1869–76	[4] 72, 000	-----	-----	[4] 22, 400	-----	-----	[4] 3, 150, 000	-----
1877–1910	[5] 1, 110	-----	-----	-----	-----	-----	3, 590	10. 64
1911–19	55, 891	33, 012	1, 156	8, 022	11, 194	212	750, 844	638. 47
1920–27	8, 715	883	295	1, 363	-----	41	188, 856	311. 57
1928–34	10, 512	-----	-----	3, 508	540	22	266, 904	858. 83
1935–49	3, 059	1, 377	-----	490	124	11	46, 365	190. 89

[1] Production figures 1906–49 from U.S. Bur. Mines, Mineral Statistics Branch, San Francisco, Calif. Figures 1869–76 are from Knopf (1918, p. 109) as estimated from figures given by Raymond and Belshaw. No production figures available for period from 1877 through 1905.
[2] Reported as "old tailings" by U.S. Bur. Mines.
[3] Concentrates reported by U.S. Bur. Mines in 1912, 1922, 1923 are possibly jig concentrates of material from old mine dumps.
[4] Estimated.
[5] 1906 only.

Production statistics are scanty for the prosperous years 1869 through 1876. Knopf's (1918, p. 109) evaluation for this period is as follows:

> The figures now current in Owens Valley range around $20,000,000, but these estimates surely show the generous influence of time and tradition. The estimates given in contemporary or nearly contemporary reports range from $6,500,000 to $15,000,000. The total output of base bullion from 1869 to 1876, inclusive, obtained by summing up the yearly production given in Raymond's annual statistics of mines and mining in the States and Territories west of the Rocky Mountains,[4] is approximately 22,500 tons. On the assumption that the average value was $300 a ton, as in 1872, the value of the total output of the Cerro Gordo during its most prosperous period was $6,750,000, or in round numbers, $7,000,000.

[4] The production for 1876 is obtained, however, from the statement by Belshaw, previously cited.

During the peak years, Cerro Gordo silver-lead bullion ("base bullion") is reported to have contained about 140 ounces of silver per ton (Raymond, 1873, p. 21). Average value of $300 per ton for the bullion assumes lead at $0.06 a pound and silver $1.29 per ounce. According to Ingalls (1908, p. 146), payment for the lead alone just about defrayed cost of smelting at the mine, together with transportation of bullion to San Francisco for refining. Net profit, after a $25 per ton charge for refining, commonly exceeded $150 per ton of silver-lead bullion.

Cerro Gordo silver-lead ores were on the whole of high quality and easily smelted. Most of the lead ore bodies were fairly uniform in grade and minable to country-rock walls. Hence, assay boundaries of minable grade were seldom indicated at Cerro Gordo and known reserves of marginal grade have never been a significant factor in mine evaluation. In connection with production history, there are unconfirmed reports of very rich gold and silver pockets having been found in mining (Fairbanks, 1896a, p. 150). Unimpeded by records of accuracy, the rumors of such strikes conspired now and again to credit the district with unrealistic or legendary precious-metal wealth.

GEOLOGY OF THE CERRO GORDO MINE

Study of Cerro Gordo geology began with observations on ore deposits by Raymond (1873) and by Goodyear (1888). Knopf (1914) first described the geologic setting of the mine and made significant deductions regarding geologic structure and ore genesis.

During the second period of major production beginning in 1911, underground geologic mapping was practiced successfully as an aid in exploration and mine development. In fact, crudely sketched maps found among the mine records suggest an awareness even earlier of economic values inherent in underground geologic mapping. Large-scale geologic maps of almost the entire mine were prepared for the U.S. Smelting, Mining and Refining Co. by R. T. Walker in 1928. Walker's maps and unpublished report were subsequently much utilized in formulating ore search programs at Cerro Gordo. Underground maps compiled at a later date for the American Smelting and Refining Co. by F. D. Hanson in 1931 and by J. J. Beeson in 1935 are likewise of great value, especially in connection with now inaccessible mine workings. Unpublished engineering reports made by T. L. Chapman in 1930 for the Estelle Mines Corp. provide valuable geologic data on the Cerro Gordo and nearby mines.

During the present study, underground engineering control was provided by the previously mentioned Hanson and Beeson mine-level compilations (pl. 3), supplemented where necessary by tape and Brunton compass traverses. Mine-level maps compiled on a scale of 1 inch equals 100 feet were used as a base from which the isometric block diagram (pl. 4) was prepared.

ROCKS OF THE CERRO GORDO MINE

SEDIMENTARY ROCKS

Stratified rock units of the Cerro Gordo mine vicinity are described earlier in this report. Of these, only the Devonian Lost Burro formation and the Chainman shale of Mississippian age were penetrated extensively by underground workings. Structurally significant nonetheless is the presence of Keeler Canyon formation just north of the mine and occurrence of probable Tin Mountain limestone in the hanging-wall block of the Buena Vista fault. These occurrences shed much light on mechanics of faulting in the mine.

Marble of the Lost Burro formation, where fractured, was evidently a favorable host for sulfide replacement. With one exception, all ore bodies of importance occurred within it. Stratification in the Lost Burro exerted local influence upon ore deposition, as well shown in both lead and zinc stopes of the Union chimney above the 500 level. Bedded ore was mined here from the upper 250 feet of the lower zone (zone A) which contains quartzite interbeds.

The *Stringocephalus* beds at the base of Lost Burro zone A were not recognized underground. So-called "coralline beds" are, however, especially numerous in the mine workings, where they are more characteristic of zone A than zone B. The "coralline beds" are loaded with spaghetti coral (*Cladopora*) and more rarely include stromatoporoids. They do not have diagnostic horizon value in the present state of our knowledge of their stratigraphic occurrence.

Although south rake of the large ore bodies is steeper than bedding, it is important that Lost Burro zone A be explored down dip from these ore bodies, as discussed in connection with the geology of the low-level Estelle tunnel.

IGNEOUS ROCKS

Igneous rocks penetrated by the Cerro Gordo workings are: (a) Union dike monzonite porphyry, (b) diabase of the Jefferson dike, and (c) green porphyry dikes including dacite porphyry and andesite porphyry. Of these the monzonite porphyry, a plutonic rock, is the oldest; green porphyry dikes are the youngest intrusive rocks. The primary massive lead-zinc ores were with certainty introduced after emplacement of monzonite porphyry and diabase; their introduction probably followed intrusion of late green porphyry dikes.

Union dike

Named for the Union tunnel, this monzonite porphyry dike was encountered in several parts of the mine from the surface to the lowest levels. The intrusive is nearly everywhere much altered and leached; where least modified it resembles texturally the monzonite porphyries of the Hart Camp and Newsboy stocks. As described by Knopf (1918, p. 112):

> The dike is of conspicuously porphyritic appearance, owing to the prevalence of large tabular feldspar crystals, and is considerably sheared and deeply stained by oxides of iron and manganese. Specimens from this dike are unsuitable for precise determination; and some material, which at least is not affected by oxidation, taken from the 400-foot level near the intersection of the San Felipe quartz vein, was found to be much altered by the development of dolomite, sericite and pyrite.

The Union dike is 45 feet wide in the Union tunnel. On the 400 level it narrows to 12 feet, but elsewhere widths as much as 60 feet were measured. In spite of alteration this igneous body is recognizable by its texture and provides a valuable key for interpretation of geologic structure. In general, the Union dike trend is that of the Cerro Gordo master fault which it follows rather closely throughout its known extent underground. It has overall northerly strike, bowed somewhat to the west. In the Union tunnel the dike separates Lost Burro formation from Chainman shale; marble forms a footwall against which altered Chainman shale and the dike appear to have moved downward in such manner as to cut out the Tin Mountain limestone which normally intervenes between Lost Burro and Chainman. Part of this displacement may have preceded dike intrusion. At many points from the 900 level to the surface the Union dike occupies the same structural position between Lost Burro and Chainman. Here and there, however, it departs from the immediate fault zone as in the Zero tunnel, where it lies wholly within altered Chainman shale of the hanging-wall block. A narrow wedge of Chainman separates the dike from marble southeast of the Belshaw shaft between the 200 level and the surface. North of this shaft the dike swings east and resumes its normal position against the marble footwall. On the 400 level near the south end of the mine, down-dropped marble of the Lost Burro in the Buena Vista fault hanging-wall block encloses monzonite porphyry resembling that of the Union dike. The altered Union dike was especially susceptible to shearing and mashing. On the 900 level near the Jefferson stope large quantities of clayey gouge were so derived.

Jefferson dike

Named for its intimate relation to the rich Jefferson chimney, this dike consists of altered diabasic rock and has accordingly been referred to in mining reports as the Diabase dike (pl. 4). Its course parallels roughly that of the Cerro Gordo fault in that part of the mine between the 900 and 550 levels. On the 550 level it passes directly into the Jefferson chimney (fig. 22). From the 200 level, where it lies east of the chimney, it has been mapped downward through the chimney to the west side on the 900 level. On the 400 level, the Jefferson dike strikes nearly northward for 600 feet but then swings northeastward in a broad arc. A comparable northeast bend is manifested by the Cerro Gordo fault, as observed on and above the 200 level (pl. 4).

The Jefferson dike meets the Union dike on the 900 level near the Jefferson chimney, where it is believed to cut the Union dike intrusively. Because of the intensely sheared condition between the Jefferson and Despreciada faults, true order of intrusion could not be demonstrated conclusively. As discussed under "Geologic structure," segments of the Jefferson dike have been thrown distributively northwestward in this vicinity. The Jefferson dike is cut intrusively by the younger north green porphyry dike on the 400 and 550 levels. On the 550 level the Jefferson dike is actually offset eastward by movement along the same intersecting green porphyry.

Decomposition nearly everywhere obviates determination of original petrologic character of this dike, but fresher material obtained by Knopf (1918, p. 113) at the face of the Buena Vista tunnel showed diabasic affinities. As host rock of important ore bodies, the Jefferson dike is discussed further under "Ore occurrence in the Cerro Gordo mine."

Dacite porphyry and andesite porphyry dikes

Three quartz-bearing dacite porphyry dikes have been mapped in the Cerro Gordo mine; these strike northwesterly and resemble petrologically others of the late dacite-andesite porphyry dike sets so characteristic of the southern Inyo Mountains. Most of these northwest-trending porphyry dikes appear to be quartzless and are therefore appropriately classed as andesite porphyry, but in many places they are too much altered for precise determination. Chloritization usually imparts a green color, and in the mine these dikes are referred to as green porphyry dikes. Under the name quartz diorite porphyry they are characterized by Knopf (1918, p. 112–113) as follows:

> They are greenish-gray porphyries carrying innumerable small black prisms of hornblende, obscure plagioclase phenocrysts, and corroded quartz crystals. The quartz phenocrysts, although as a rule widely scattered and easily overlooked on casual inspection, are nevertheless peculiarly distinctive of the rock; and they are readily distinguishable even where the hornblende and plagioclase have been obliterated by mineralization.

The three porphyry dikes in the Cerro Gordo mine are referred to as north, middle, and south green porphyry dikes (pl. 4). They were intruded along northwest-trending faults or zones of weakness on which further movement took place after dike emplacement.

The north green porphyry dike was found at several points from the 900 up to the 200 level. On the 700 and 550 levels the San Felipe quartz vein follows its north edge, but upward on the 400 level dike and vein have separated, except for a junction northeast of the Belshaw shaft. The north green porphyry cuts the Jefferson diabasic dike intrusively on the 550 and 400 levels, with eastward offset of the Jefferson on the 550. No such offset of diabase by movement on the green porphyry dike was observed at the intersection on the 400 level. On the 400 level the Cerro Gordo fault is cut and thrown west about 150 feet by movement along the north green porphyry dike.

The middle green porphyry dike passes through the Cerro Gordo fault and the Union dike on the 900, 700, and the 400 level, but apparently without significant offset. It is accordingly evident that the green porphyry dikes are younger than the Union dike and the Jefferson diabasic dike.

MAGNESIAN ALTERATION IN CERRO GORDO MINE

Hydrothermal alteration along faults, fissures, and porphyry dikes is responsible for fairly extensive commercial talc mineralization in the Cerro Gordo vicinity (Page, 1951). In the Cerro Gordo mine there exists a distinct channel or zone adjoining the Union chimney wherein hydrothermal activity has brought about alteration and dolomitization of marble in the Lost Burro formation. Talc was not recognized here. Knopf (1918, p. 114) calls attention to association of this additive alteration with the Union lead channel as having possible value in search for new ore bodies.

The steep channel of magnesian alteration is observable between the 550 level and points above the 400 level south and west of the Union chimney. It extends about 200 feet west of the lead channel on the 400 and 550 levels.

As noted by Knopf the partly dolomitized marble in the channel in some places has a porphyritic appearance, owing to abundant large curved dolomite cleavage surfaces. Disseminated iron oxide gives the rock a yellowish limonitic color. Also present are scattered pockets of iron oxide and a few pods of pyrite. Solutions responsible for this magnesian alteration apparently did not carry lead or zinc. At stations 70 feet and 120 feet above the 550 level near the south edge of the Union chimney, the irony dolomitic rock is separated by faults or fissures from relatively unaltered marble. These breaks strike northwestward and dip southward about 80°.

GEOLOGIC STRUCTURE OF THE CERRO GORDO MINE

Broader aspects of geologic structure in the area surrounding the Cerro Gordo mine have been discussed previously. Features which bear directly upon ore occurrence are: (a) The Cerro Gordo anticline with subsidiary folds, (b) the Cerro Gordo master fault, and (c) numerous fissures and faults with roughly north-south and northwest-southeast strikes (fig. 12).

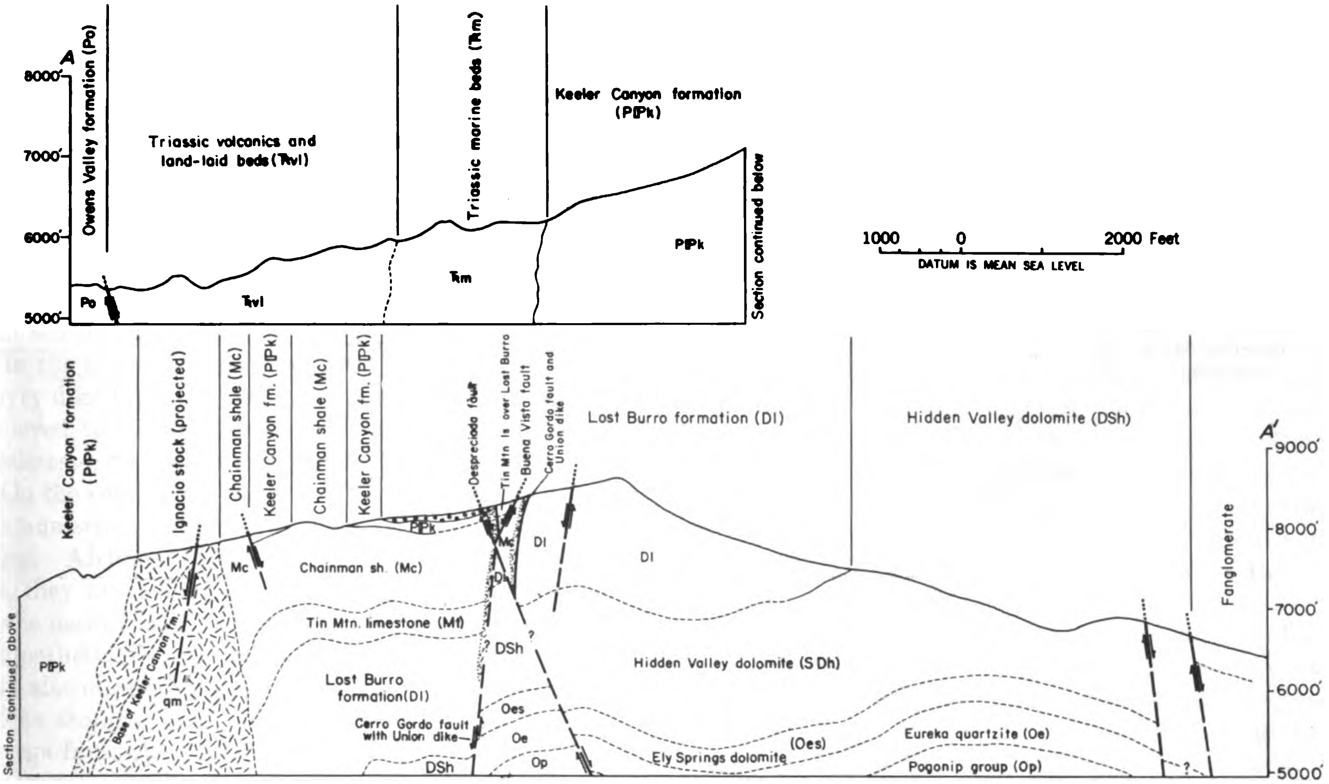

FIGURE 12.—Generalized structure section (*A–A'*) through the higher Inyo Range at Cerro Gordo. Section runs southwest-northeast along line of Cerro Gordo tramline. (See pl. 1 for location of section.) Ignacio stock projected slightly to plane of section. In middle of section the rocks plunge toward observer with south plunge of Cerro Gordo anticline. Vertical and horizontal scales same.

CERRO GORDO ANTICLINE

Localization of lead-zinc ore bodies in marble near the axial plane of a large anticline is more than mere happenstance. Theoretically the stress conditions in this fold segment might well be expected to create exceptionally favorable conditions for sulfide deposition in a carbonate rock, especially in the proximity of monzonite intrusive rocks.

Where bedding and cleavage were observable in the Cerro Gordo mine, the dip is west at high angles except for local reversal in the minor anticlinal warp at the Union chimney. Axial zone of the main Cerro Gordo anticline lies east of the Belshaw shaft, possibly within a distance of 500 feet. Cerro Gordo mine workings accessible during this study lie in the west limb only.[8] At lower altitude the 660-raise level of the Estelle mine (fig. 18, fig. 25) probably enters the east limb because east bedding dips are recorded on that level 1,400 feet south of the Belshaw shaft. The Charles Lease tunnel (pl. 1, fig. 28) also cuts the axial zone.

A small subsidiary anticline is indicated by mapping at the Union chimney. Between the 700 and the 400 levels in this vicinity, local changes in dip and strike of marble bedding outline a rather sharp south-pitching nose resembling that of the parent Cerro Gordo anticline itself.

[8] The long east drive through Lost Burro marble on the 200 level was not accessible. It is possible that this opening entered the axial zone of the main anticline.

CERRO GORDO FAULT

The Cerro Gordo fault (pls. 1 and 2) lies in a belt of roughly north-trending shears traceable from Soda Canyon north to Saline Valley. From the standpoint of structural ore control, the Cerro Gordo fault is appropriately referred to as the master fault of the mine. Operators have long called it the contact vein because of weak copper staining and limonite in the gouge and breccia. Extensive workings were driven along and across it without reward.

Displacement on the Cerro Gordo fault is seemingly normal with Chainman shale downthrown on the west discordantly against marble of the Lost Burro (fig. 13). Although the overall dip is steep and westward, this fault exhibits local steepening to vertical or more rarely changes to steep easterly dip on its downward uneven course from the surface to the deepest mine levels. Later warping may account for some of these dip irregularities.

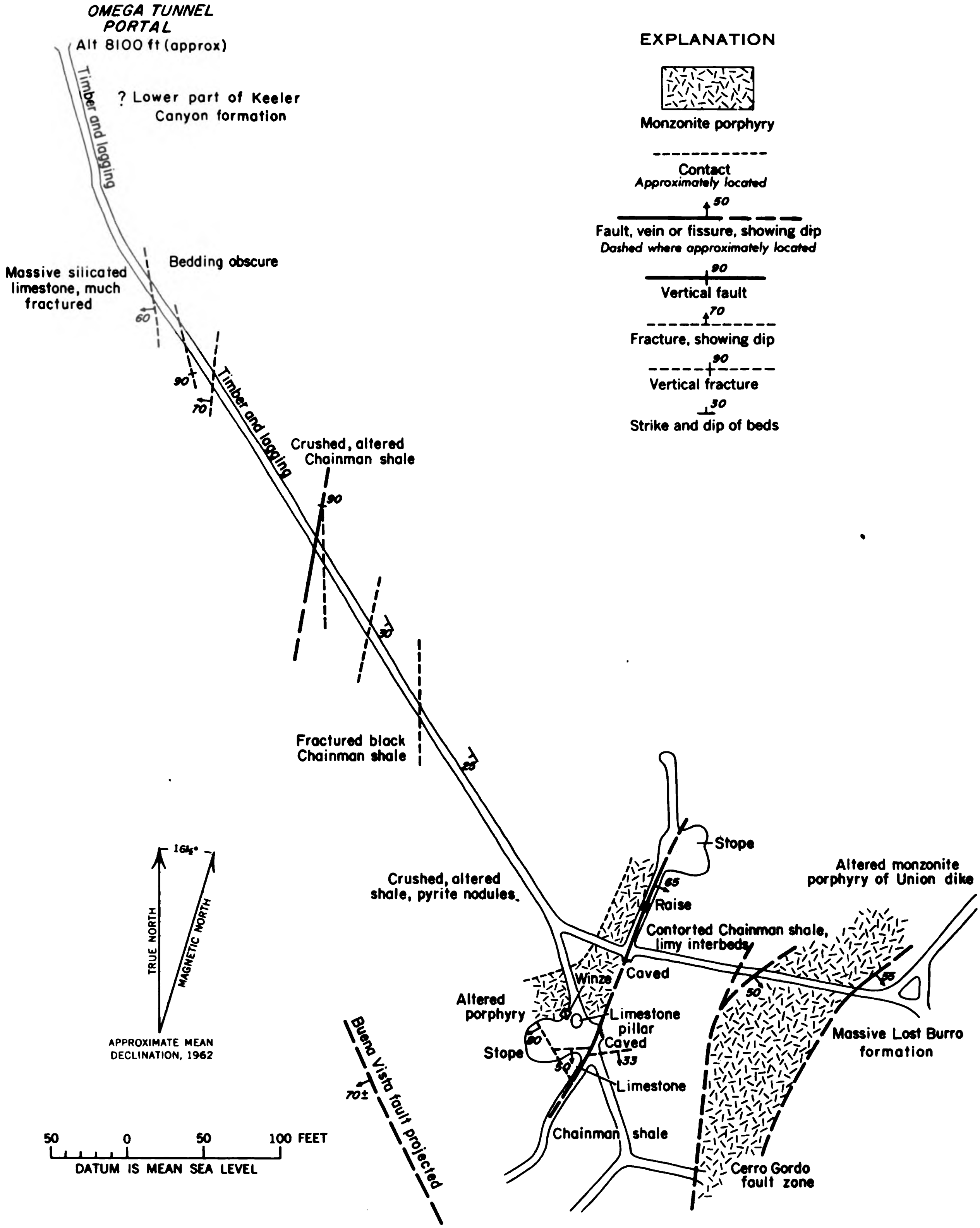

FIGURE 13.—Map of workings of Omega tunnel, Cerro Gordo mine. Shows Union dike in Cerro Gordo fault zone and projection of Buena Vista fault. Small stopes in Chainman shale may have yielded ore formed by replacement of limestone beds in Chainman shale. See plate 3 and figure 16 for location.

Overall trend of the Cerro Gordo fault is nearly north, but in detail it exhibits sinuosities as mapped in the near-surface tunnels. Its trace is bowed westward near the surface, and the major fault is cut by minor cross breaks.

The Cerro Gordo fault zone was traced north of Cerro Gordo along the east side of San Lucas Canyon almost due north to the Newtown shaft (fig. 14). South of the Buena Vista tunnel, the fault presents no good surface exposures, but there is little doubt that it continues southward at least as far as the Morning Star mine, a mile south of Cerro Gordo.

In the Cerro Gordo mine the Union monzonite porphyry dike follows the Cerro Gordo fault zone, which is believed to have existed as a north-trending zone of weakness prior to plutonic intrusion.

On the east or marble side of the Cerro Gordo fault are numerous roughly parallel fractures, fissures, and veins. Although these breaks show no measurable offset, they are believed to be mechanically sympathetic to the main Cerro Gordo master fault. Several of the sympathetic fissures served as avenues of ascent for lead-zinc mineralizing solutions.

The Omega fault.—On the deeper mine levels the Omega fault (pl. 4) parallels the Cerro Gordo fault on the west where it also separates Lost Burro formation from Chainman shale. In spite of connotation to the near-surface Omega tunnel, this fault has not been recognized at the surface nor with assurance on upper tunnel levels. The Omega fault shows heavy near-horizontal mullion structure, but is believed to have originated as a normal fault sympathetic to the Cerro Gordo fault.[9] Several northwest faults, dikes, and veins in the footwall marble of the Lost Burro seem to terminate against the Omega fault, apparently not continuing into the hanging-wall shale of the Chainman to the west (pl. 4).

FIGURE 14.—Looking northeast across San Lucas Canyon from a point north of Cerro Gordo. Middleground shows rugged slopes underlain by ore-bearing marble of the Lost Burro formation. Rounded hills in middle distance underlain by altered and intruded late Paleozoic strata. Distant mountains lie in the Ubehebe district.

[9] The possibility that the Cerro Gordo and associated north-trending faults may have originated very early as tear faults has been considered in connection with folding and thrusting.

NORMAL FAULTS WITH NORTHWEST STRIKE

Faults, dikes, and veins with northwesterly trend cut the older north-south structural features of the Cerro Gordo mine. Offsets on faults of the northwest set are among complications encountered in mine development. Younger faults conceivably responsible for post-mineralization offset of ore bodies include the Buena Vista, La Despreciada (pl. 4), and the Jefferson fault (fig. 22), all with northwest strike. On the other hand, some of these northwest breaks were undoubtedly in existence prior to ore mineralization and probably have a long history of intermittent movement. Intersections of northwest- with north-trending breaks were beyond question important in helping to establish the ore channels or chimneys.

The two most important northwest-trending faults in the mine are the west-dipping Buena Vista and the east dipping La Despreciada. Down-dip projections indicate a meeting of the two in the much broken ground above the 900 level.

Buena Vista fault.—Named for the Buena Vista tunnel (fig. 23), wherein it strikes N. 28° W., this fault passes east of the Belshaw shaft collar and continues northwestward through the Union tunnel. The Buena Vista fault is normal and has been traced downward to the 550 level. Normal displacement is demonstrated by downward shift of hanging-wall Tin Mountain limestone and by downthrow to the west of a segment of the Union dike. Downthrown Tin Mountain is present in the Buena Vista tunnel where rolled fragments of altered igneous rock in the fault gouge appear to be drag material from the Jefferson diabasic dike.

The Buena Vista fault zone is alluvium covered southeast of the Belshaw shaft (fig. 15). Its surface projection probably swings east toward the high gap in dark-gray Tin Mountain limestone, 1,400 feet southeast of the shaft (fig. 16). If this is so downward movement at the gap appears to be less than in the mine. Decreased downthrow southeast of the mine may be accounted for by rotational or scissors movement, or by absorption of movement on cross faults. Another explanation is absorption by distributive offset on several shears which parallel the Buena Vista fault in Tin Mountain limestone west of the gap. Possible role of the Buena Vista fault in cutting and postmineralization downthrow of ore segments in the hanging-wall block is considered in the section on "ore occurrence in the Cerro Gordo mine."

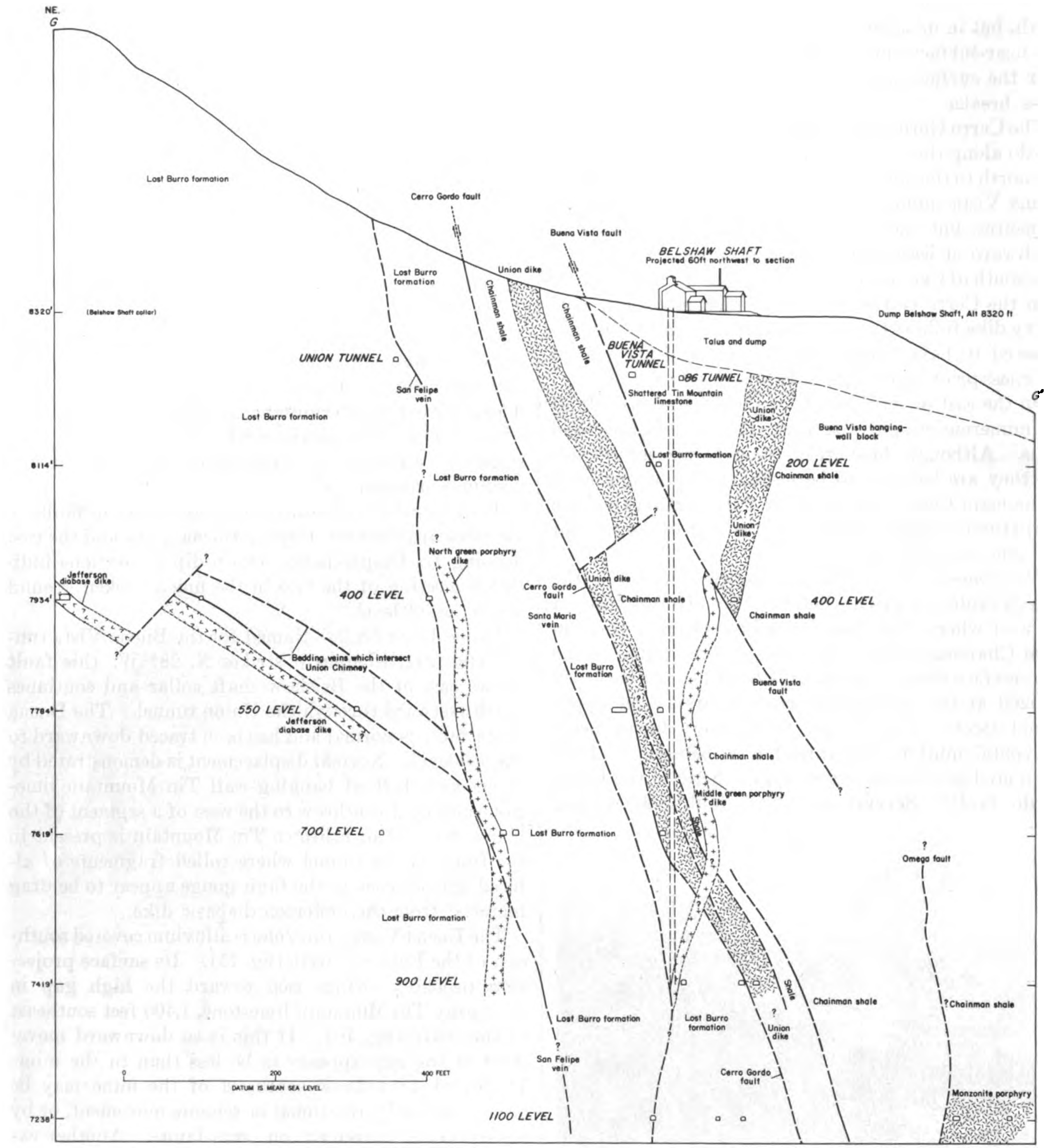

FIGURE 15.—Structure section (*G–G'*) running northeast-southwest (observer facing southeast) through Cerro Gordo mine just north of Belshaw shaft. Shows offset of Union dike on Buena Vista fault, and probable intrusive relation of middle green porphyry dike to Union dike and Cerro Gordo fault. Vertical and horizontal scales same. See plate 3 for location of section.

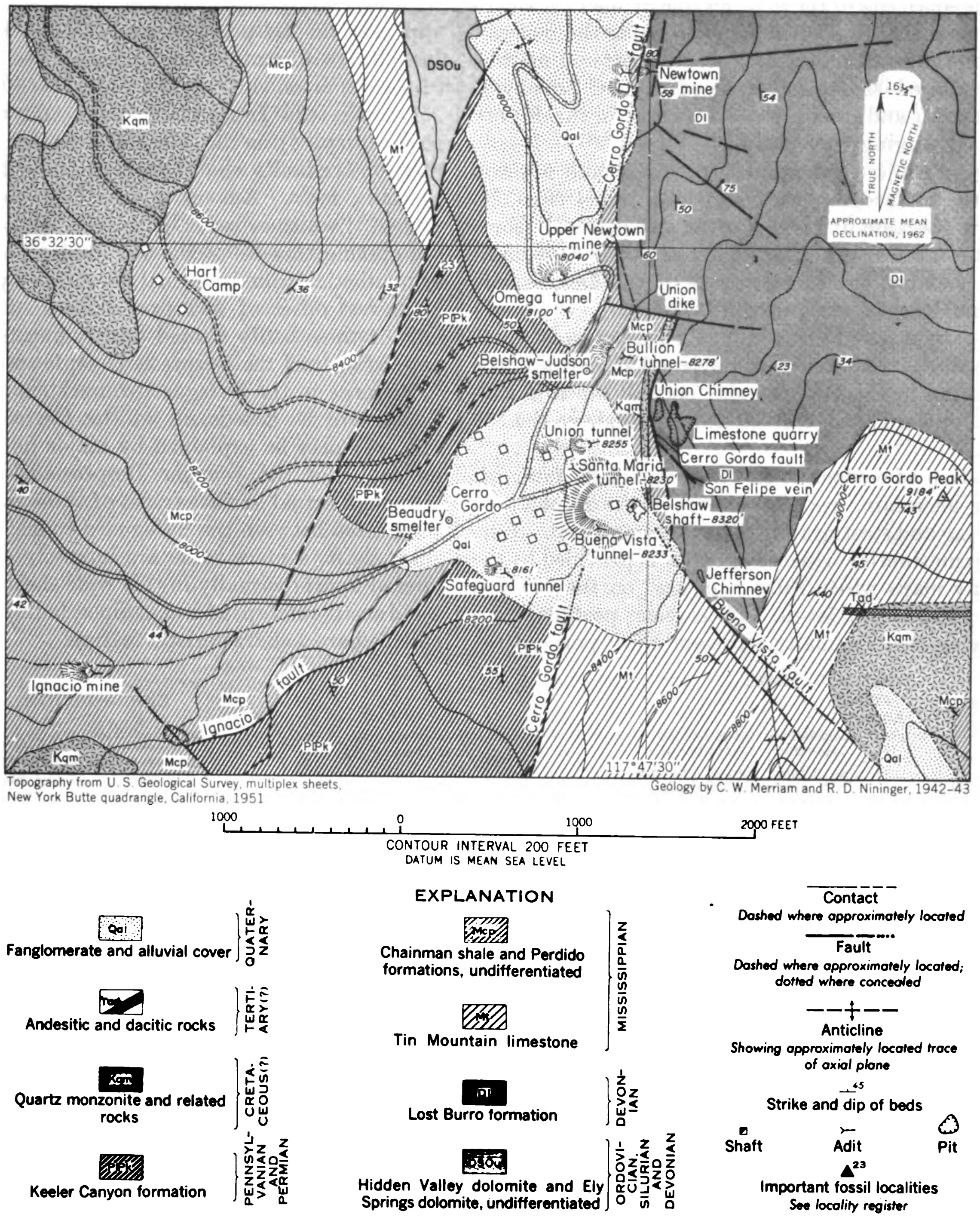

Figure 16.—Enlarged geologic map of area surrounding Cerro Gordo to show location of principal tunnel workings.

The Despreciada fault.—This important northwest-trending fault was observable only on the 900 level at the time of the present study (pl. 4). On that level near the south end of the mine marble of the Lost Burro with "spaghetti coral" (*Cladopora*) makes the footwall block on the west; on the east occurs shattered marble, gougy altered shale, and crushed altered porphyry. Irony matter and local ore pockets occur in this highly disturbed hanging-wall ground.

Upward projection of the Despreciada fault appears to cut the surface near the Safeguard tunnel (fig. 16). However, because of surface cover and generally shattered and altered condition of rocks in the Safeguard tunnel vicinity, the surface trace of the Despreciada fault could not be recognized.

Company maps of inaccessible La Despreciada workings on the 700 and 550 levels indicate that marble of the Lost Burro in the Despreciada footwall block is continuous upward from the 900 at least to above the 550 level.

Direction of movement on the Despreciada fault, although uncertain, appears on the basis of observed fault drag and stratigraphy to have been normal. Further circumstantial evidence of throw is provided by the company maps which show "diabase" on the 700 and 550 La Despreciada levels. If, as seems possible, these igneous segments actually represent offset parts of the Jefferson diabasic dike, it follows that movement on the Despreciada fault was probably normal, but with large strike-slip shift to the northwest. In accordance with this strike-slip interpretation a possibility exists also that the Despreciada footwall ore bodies in the Lost Burro formation are actually offset deeper parts of the Jefferson chimney.

Between the Despreciada fault and the Jefferson chimney (pl. 4) on the 900 level are other east-dipping distributive faults which roughly parallel the main Despreciada fault. Among these is the Jefferson fault (fig. 22), which is reported to have cut off the Jefferson chimney ore below the 900 level. Throw on these sympathetic faults is normal, as shown by distributive northwest offset of Jefferson diabasic dike segments. One of these badly altered dike segments may be examined against the Despreciada fault on the 900 level.

Presence in the Safeguard tunnel (fig. 16) of what appears to be a segment of the Cerro Gordo fault (fig. 17) may be explained by major northwest offset on faults of the younger Despreciada-Jefferson set. Although actual surface traces of these faults were not recognized between the Safeguard tunnel and the Belshaw shaft, surface rocks are intensely shattered and altered hydrothermally in this wide shear zone. Scattered small outcrops of iron-stained fault breccia partly silicified are a characteristic feature.

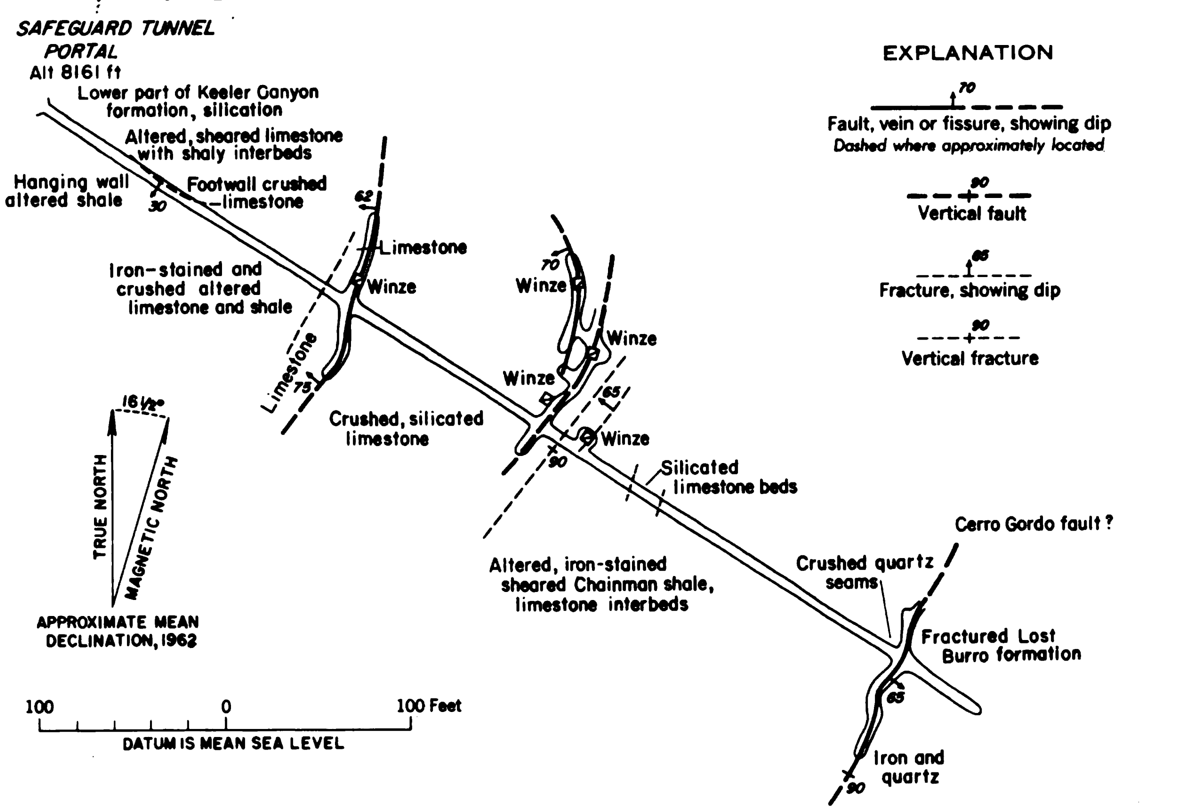

FIGURE 17.—Map of workings of Safeguard tunnel. Fault between altered Chainman shale and marble of the Lost Burro formation, doubtfully an offset segment of the Cerro Gordo fault. For location see plate 3; figure 16.

ORE OCCURRENCE IN THE CERRO GORDO MINE

Ores of the Cerro Gordo mine (pl. 4) occur as small shoots in fissures, as massive bodies in large steeply inclined pipes or "chimneys," in small isolated pockets, and in siliceous veins. Fissure, pocket, and chimney ores are mainly in shattered marble of the Devonian Lost Burro formation, but in addition, rich ore shoots were extracted from fissured parts of the Jefferson diabassic dike. Locally the wallrock is dolomitized hydrothermally, but in the main, wallrock alteration other than marmoritization is moderate. Unlike mines of the Darwin district (Hall and MacKevett, 1958), there is very little silication or change of the carbonate wallrocks to calc-silicates.

Pipelike and bedded carbonate zinc ore bodies are associated with the massive lead chimney ores. The zinc ores are supergene and derived by leaching of primary lead-zinc sulfide ores.

With reference to geologic structure, both premineralization and postmineralization factors must be considered. Premineralization features are the north- and northwest-trending fissures (fig. 18) and fractures which facilitated entry of ore-bearing solutions. Postmineralization structures include northwest-trending faults like the Buena Vista and Despreciada faults which appear to have truncated and offset the lead ore bodies. Rich lead-silver ores mined from hanging-wall blocks of the Buena Vista fault have been explained as postmineralization faulted segments of the large chimneys. This explanation is, however, highly uncertain.

Of special significance in point of structural ore control are the many small north- and northwest-trending fissures and fractures which formed in the footwall marble of the Cerro Gordo master fault. Among structures of this nature which have been mapped are: (a) The Bullion fissure (fig. 19), (b) feeding fissures within and adjacent to the Union chimney, (c) ore-bearing fissures of the Upper and Lower Newtown mines in San

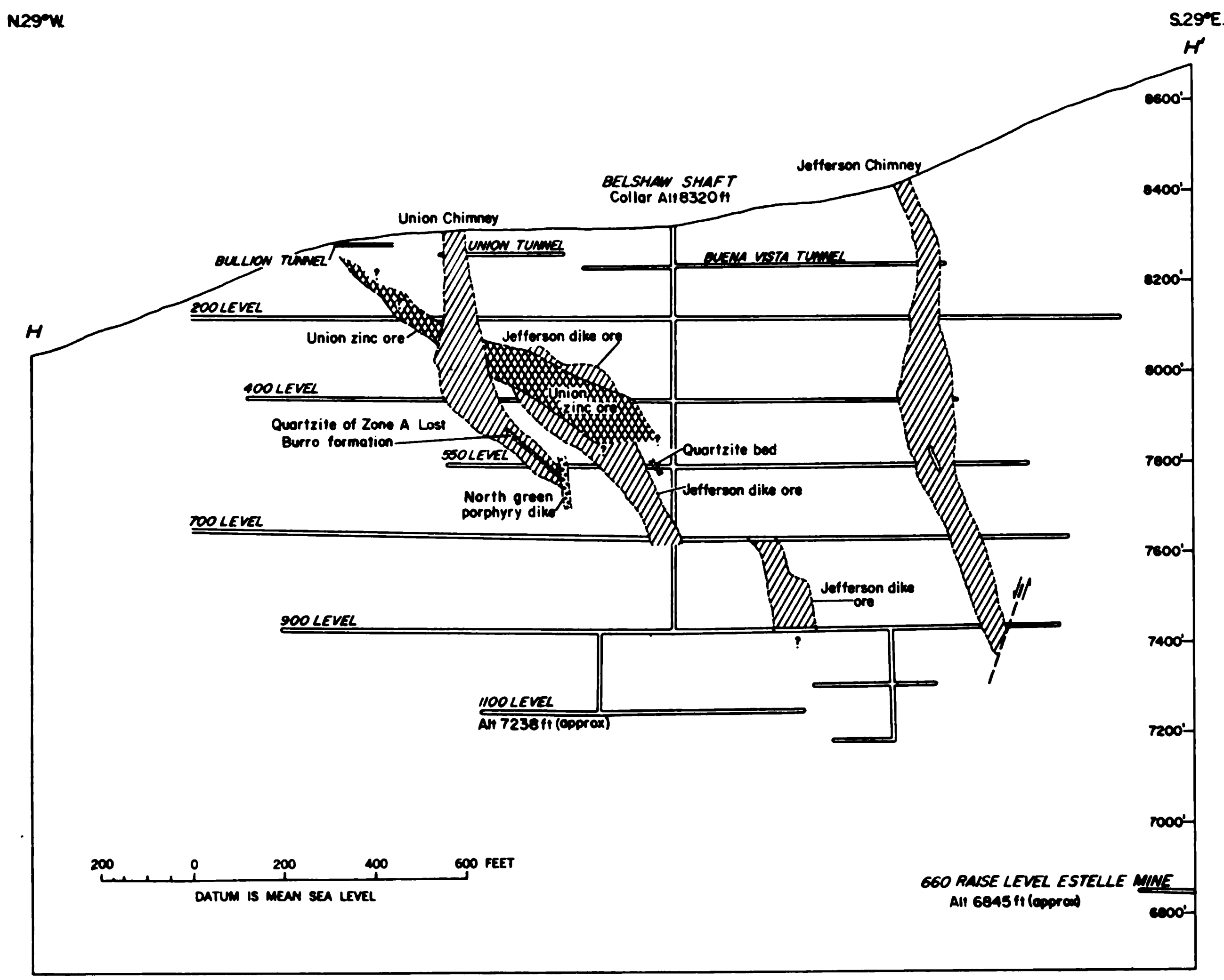

FIGURE 18.—Semidiagrammatic northwest-southeast longitudinal projected section H–H' (observer facing northeast) through Cerro Gordo mine, showing principal ore channels. Horizontal and vertical scales same. See plate 3 for location of section.

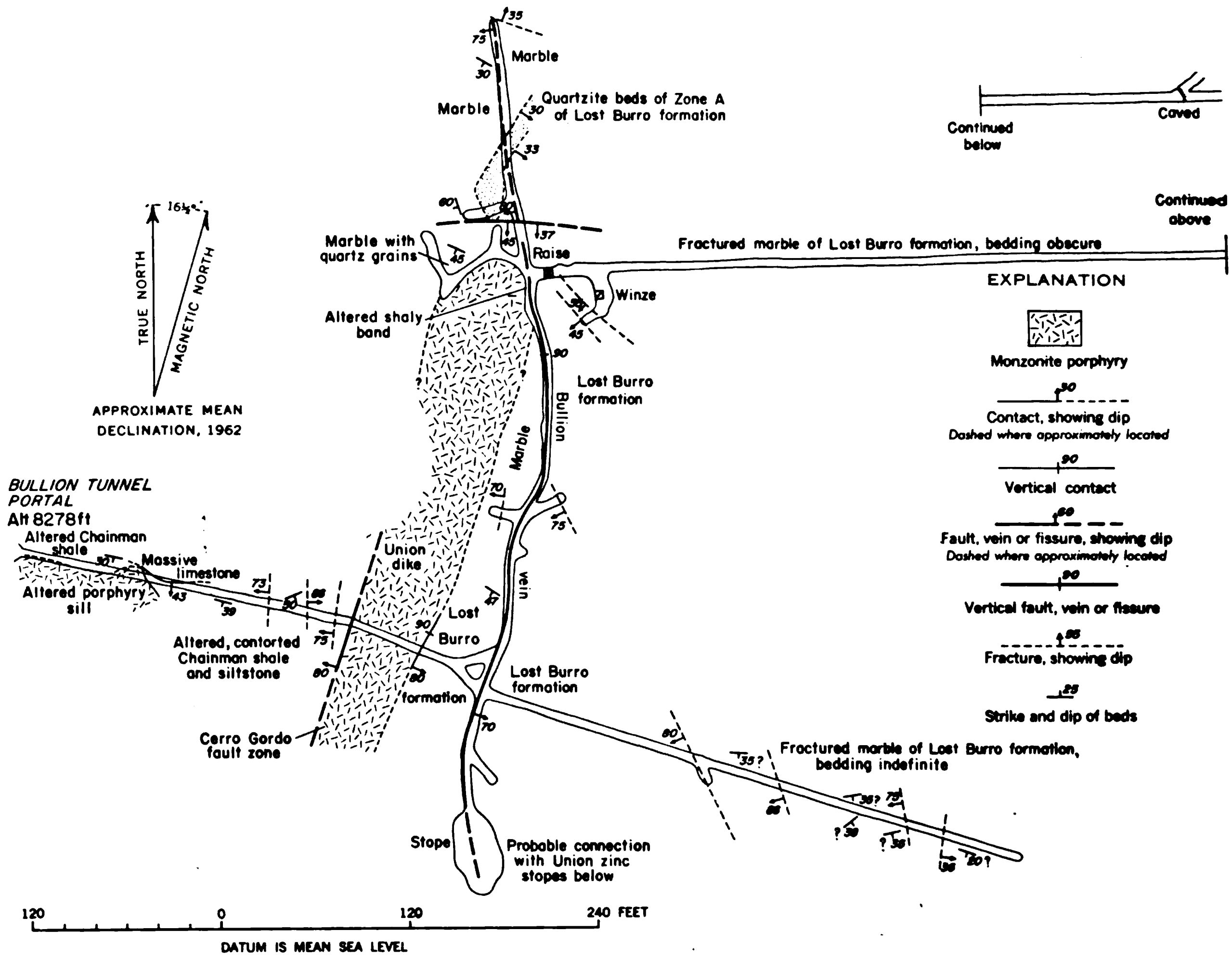

FIGURE 19.—Map of workings of Bullion tunnel. See figure 24 for Union zinc stopes. For location of Bullion tunnel see plate 3; figure 16.

Lucas Canyon, and (d) the Zero fissure vein (fig. 23). The short Zero fissure vein trends a few degrees west of north, paralleling a westward bulge of the Cerro Gordo fault. Fissuring in the Jefferson diabasic dike follows a sinuous course bellied to the west, but on the whole seems to conform in overall trend to the north-south fracture pattern.

Massive primary ore bodies which made the Cerro Gordo a great silver and lead producer were emplaced as shoots in steeply inclined pipes which rake south in the general plunge direction of the Cerro Gordo anticline. There are two principal channels of this kind, the Union chimney (China stope) at the north end of the mine, the Jefferson chimney at the south end. Pinching and swelling on their downward course, these channels expanded locally to exceed 100 feet in greatest horizontal dimension. They were not solid ore from top to bottom, but included ribs of fractured, unreplaced country rock. Some of the stopes which yielded rich galena ore were narrow and tortuous; the walls of these openings are commonly rather clean and little altered.

UNION CHIMNEY

The Union chimney yielded a very large share of the rich Cerro Gordo silver-lead ore. It is localized in marble of the Lost Burro formation near the north end of the mine, just east of the Cerro Gordo master fault (pl. 4; fig. 16; fig. 18). Marble enclosing the pipe is intensely sheared and fractured (fig. 20). On the south and west it has been subjected to additive magnesian alteration. Deposits of supergene carbonate zinc lay directly east of the lead-ore channel (fig. 18, fig. 24). In the early days, ore was extracted from the surface cropping downward to termination of the channel above the 550 level near a union with the north green porphyry dike and associated San Felipe siliceous vein.

FIGURE 20.—Surface pits in marble of the Lost Burro formation at site of Union chimney outcrop. The most important ore discovery at Cerro Gordo was made here by Mexicans about 1865.

The Union chimney [10] (fig. 21) was largely mined out before 1879, but during the present study, parts of the old stopes between the 400 and 550 levels were accessible. Additional data were obtained from old company maps and from published records of Raymond's 1872 examination when the Union mine was in operation.

In northwest-southeast section (fig. 18), the upper part of the Union chimney stands near vertical, but the lower part dips southward more or less with bedding inclination of the Lost Burro formation. Early reports indicate that the upper nearly vertical part had branches and ramifications, one of which led toward the present south drift of the Bullion tunnel (fig. 19). A westerly branch extended toward the Santa Maria workings 200 feet below the Union tunnel. Rich lead carbonate ore in the nearly vertical part occurred as shoots separated by ferruginous matter of lower grade and by septa of partly replaced marble. On the 400 level, the filled stope measures 147 feet horizontally with maximum width of 48 feet. The south-dipping part of the chimney below the 400 level includes several small partly filled stopes which dip south from 30° to 40°.

The south-dipping lower part of the Union chimney follows bedding in the Lost Burro host rock, whereas the upper nearly vertical part cuts across stratification. It is nonetheless probable that, even with the crosscutting relation, variations in size and nature of ore shoots within the pipe are related to change in physical and chemical character of marble beds as the stratigraphic section is transected.

Control of ore deposition by stratification may be observed in old inclined stopes between the 550 level and a sublevel 120 above (fig. 18). A 6-foot white quartzite bed which probably lies stratigraphically near the top of zone A of the Lost Burro strikes near

[10] In later years the name "China ore body" or "China stope" was applied to the Union chimney by leasers, in reference to legendary burial of several Chinese by caving of a stope.

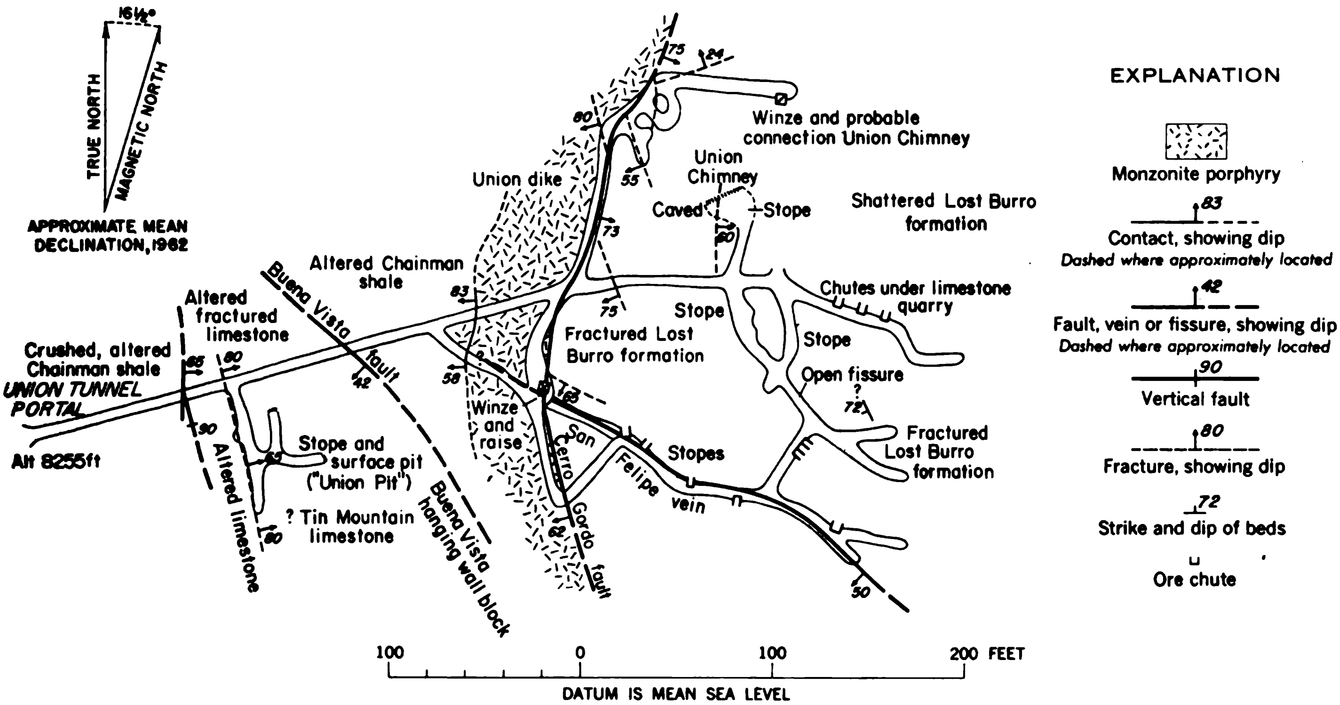

FIGURE 21.—Map of Union tunnel workings (Union mine) showing relation of Union chimney ("China stope") to Cerro Gordo fault and Union dike. Ores of so-called "Union pit" mined west of Buena Vista fault. For location see figure 16 and plate 3.

east and dips 30° to 42° S., in this vicinity. Bedded ore was mined above and below the quartzite. The 6-foot quartzite bed provides a useful key for drill exploration and interpretation of underground geologic structure. Though probably quite local, stratigraphic control of ore deposition in the marble of the Lost Burro is worthy of further consideration in connection with future ore exploration.

East-west width of the Union chimney in the lower bedded-ore part is about 70 feet, but individual stopes are smaller with east-west stope length as much as 35 feet and stope height normal to bedding which ranges from less than 3 feet to more than 10 feet. Ore was mined to rather sharp boundaries with unmineralized or weakly mineralized marble wallrock.

A small south-pitching anticlinal fold is outlined by plotting bedding attitudes in quartzite and marble between the 400 and 700 levels at the Union chimney. Nearly east-west bedding strike suggests that the lower chimney stopes are actually near the axis or nose of this flexure. Small size of the fold suggests that it is subsidiary to the main Cerro Gordo anticline. South pitch of the fold harmonizes with rake of the lower part of the Union chimney.

Host rocks within and bordering the Union chimney are sheared and cut by fissures and minor faults which range in strike from north to northwest. At several points above the 550 level, minor northwest-trending faults offset the 6-foot quartzite marker, in one place with reverse movement.

Iron-bearing fissures striking within a few degrees of north may be observed in mine openings from 80 to 120 feet above the 550 level. At the north margin of the chimney some of the fissures dip east about 40°. They are filled with cavernous limonite containing minor amounts of vein quartz and zinc carbonate, with little indication of lead in association with the iron. The limonitic fissure fillings were evidently derived by oxidation and leaching from primary sulfides.

Relations of mineralization to wallrock may be observed near the south edge of the Union chimney, 100 feet above the 550 level. In this vicinity the 6-foot quartzite marker is overlain by heavy cavernous limonite derived from the primary sulfides which replaced marble along bedding. Carbonate zinc and clayey matter resembling halloysite occur along the limonite-marble boundary below. At one point the topmost 2 inches of the footwall marble is replaced by zinc carbonate. One exposure shows about 6 feet of the limonite overlain by a 12-foot dark-gray marble bed. In the marble bed are many irregular knife-blade seamlets of galena and tabular pods of lead carbonate together with galena nodules. The pods of ore are elongated with bedding. Dark marble enclosing the galena seamlets appears fresh, and the lime-sulfide boundaries are sharp. Pockets and seams of vein quartz containing galena occur in marble beneath the bedded limonite mass.

Fracturing which attended movement on the Cerro Gordo master fault evidently prepared a channel for entry of the ore-bearing solutions. Bottoming of the ore chimney above the 550 level has not, however, been satisfactorily explained. Although postmineralization faulting is the favored explanation, no clear-cut evidence of faulting off has been observed, and no faulted segments have been discovered after many and varied attempts.

Factors to be considered in connection with origin and history of the ore channel are: (a) presence of the north green porphyry dike and associated San Felipe vein near the point where the Union chimney ore was bottomed; (b) lack of clear evidence of faulting off; (c) mineralogical differences between the siliceous San Felipe vein ore and that of the Union chimney.

It was long believed that the Union chimney ore was cut off in depth by a northwest-trending fault in the position of the north green porphyry dike and San Felipe vein. In accordance with this theory, the dike and San Felipe vein were introduced after the massive ore was faulted. Very extensive digging since the 1870's failed to confirm this, and there is no field evidence to show that the porphyry dike or the siliceous vein actually cut the massive Union chimney sulfide ore as earlier inferred.

Speculation regarding the genesis of Cerro Gordo ores has been colored by an hypothesis of successive mineralization stages whereby the low-silica massive lead ores of the Union chimney were first introduced, followed by quartz veins of San Felipe type which are assumed to transect the lead ores.

Production records show that the chimney ores were low in silica. However, vein quartz was observed in the lower Union chimney stopes, but whether this quartz was introduced concurrently with the chimney lead ores or later was not determined.

Mineralogical differences between the siliceous San Felipe vein ores and those of the Union chimney have been looked upon as ruling out any possibility of the San Felipe vein having served as a feeder for the Union chimney. Nonetheless it seems quite likely that fissures in the general zone of the San Felipe vein and north green porphyry dike served as avenues of ascent for the mineralizing solutions which formed the chimney ores. Near the bottom of the chimney the footwall marble shows thin knife-blade feeder veins nearly conrdant with bedding. These were called "flat veins" y mine operators. It seems quite probable that e ascending solutions fed off from the steep fissures nd followed stratification planes to form the lower part f the massive lead chimney.

Present studies in the lower bedded part of the Union himney leave the impression that the massive lead ineralization weakened gradually downdip in the pipe s the San Felipe vein zone was approached.

Inferred genetic relation of the Union chimney and he San Felipe vein once had legal as well as practical ignificance. The San Felipe Co., having prior rights, oped for a meeting of the two as basis for a lawsuit. nsofar as known, actual junction was never demontrated as the chimney ore was mined closer to the silieous vein in depth.

JEFFERSON CHIMNEY

The Jefferson chimney lies in fractured marble of he Lost Burro formation close to the Cerro Gordo and Buena Vista faults. It extends about 400 feet deeper han the Union chimney, having been mined from the surface to termination below the 900 level. Except for the small surface pit, all workings in this channel were inaccessible at the time of the present study.

The Jefferson chimney rakes southwest at an overall angle of about 80° (fig. 18). Like the upper part of the Union chimney, it appears to have crosscut bedding in the marble; there is no evidence that ore shoots within the pipe followed bedding trend, as in the deeper part of the Union chimney. In its precipitous downward course the pipe pinches and swells, being highly variable in area and shape of horizontal section from level to level. On one level it is 120 feet long and 40 feet wide. On the 900 level, a large square-set stope had a horizontal area of about 2,000 square feet. The horizontal section of the 400-level stope was roughly 1,000 square feet.

Specific data on ores extracted from the Jefferson chimney are not available. By inference they were largely oxidized and of mineralogic character comparable to those of the Union. Oxide zinc ores as mined adjacent to the Union chimney are not known to have been extracted near the Jefferson. Supergene zinc deposits would be expected in that vicinity also.

The Jefferson chimney is intimately related to the Jefferson diabasic dike which passes through it (pl. 4; fig. 22). On upper levels the dike lies east of the chimney, paralleling the long north-trending horizontal axis of the chimney section. Downward from the 400 level the dike strikes into the chimney, shifting its position downdip to the west of the chimney on the 900 level. Roughly north-trending fractures which determined the position and trend of the diabasic dike also influenced shape and position of the Jefferson chimney, as illustrated by north-south elongation of ore shoots on several levels.

As with the Union chimney, fracturing related to movement on the Cerro Gordo fault undoubtedly prepared the channel for ascent of mineralizing solutions. Fracture and fissure intersections played a part in localizing ore shoots. Fractures of the Zero fissure vein set (fig. 23) and the diabasic dike set strike into the Jefferson chimney. Both carried mineralizing solutions, and important ore bodies were actually emplaced within the dike itself as in the following discussion.

According to reports of mine operators, loss of Jefferson chimney ore below the big 900-level stope was due to offset on the Jefferson fault (fig. 22). Large cross-section of the ore pipe at the 900-level stope seems to favor this structural interpretation. Having a northwesterly strike and easterly dip, the Jefferson fault falls within the Despreciada fault set elsewhere discussed. Ore bodies in the Despreciada footwall conceivably represent faulted-off deeper segments of the Jefferson chimney.

DESPRECIADA FOOTWALL ORE BODIES

The La Despreciada workings off the 550 and 700 levels were not accessible at the time of this investigation. All information on this section of the mine was accordingly obtained from company maps and records. The ore bodies occurred in fractured gray and white marble, quite probably the Lost Burro, and were seemingly disposed in a nearly vertical pipe comparable to the Jefferson chimney, if not actually a faulted deep continuation of it.

Company maps show monzonite porphyry and diabase on the 700 level near the west edge of the La Despreciada ore. Diabase is indicated also on the 550 level above the uppermost Despreciada ore. Assuming that these igneous rocks actually represent segments of the Union and Jefferson dikes, it follows that a northwest normal throw on the Despreciada fault set (pl. 4) could theoretically have shifted a deeper part of the Jefferson chimney upward to the position occupied by these Despreciada footwall ores. It is a significant point that both dikes are also closely associated with the Jefferson chimney where it was bottomed below the 900 level.

JEFFERSON DIABASIC DIKE ORE BODIES

The Jefferson diabasic dike contained the easternmost ore bodies in the Cerro Gordo mine (pl. 4; fig. 18). Ore was deposited in shoots raking steeply southwest or south and was mined from below the 900 level to an undetermined elevation between the 400 and 200 levels. No surface exposure of the dike was recognized. De-

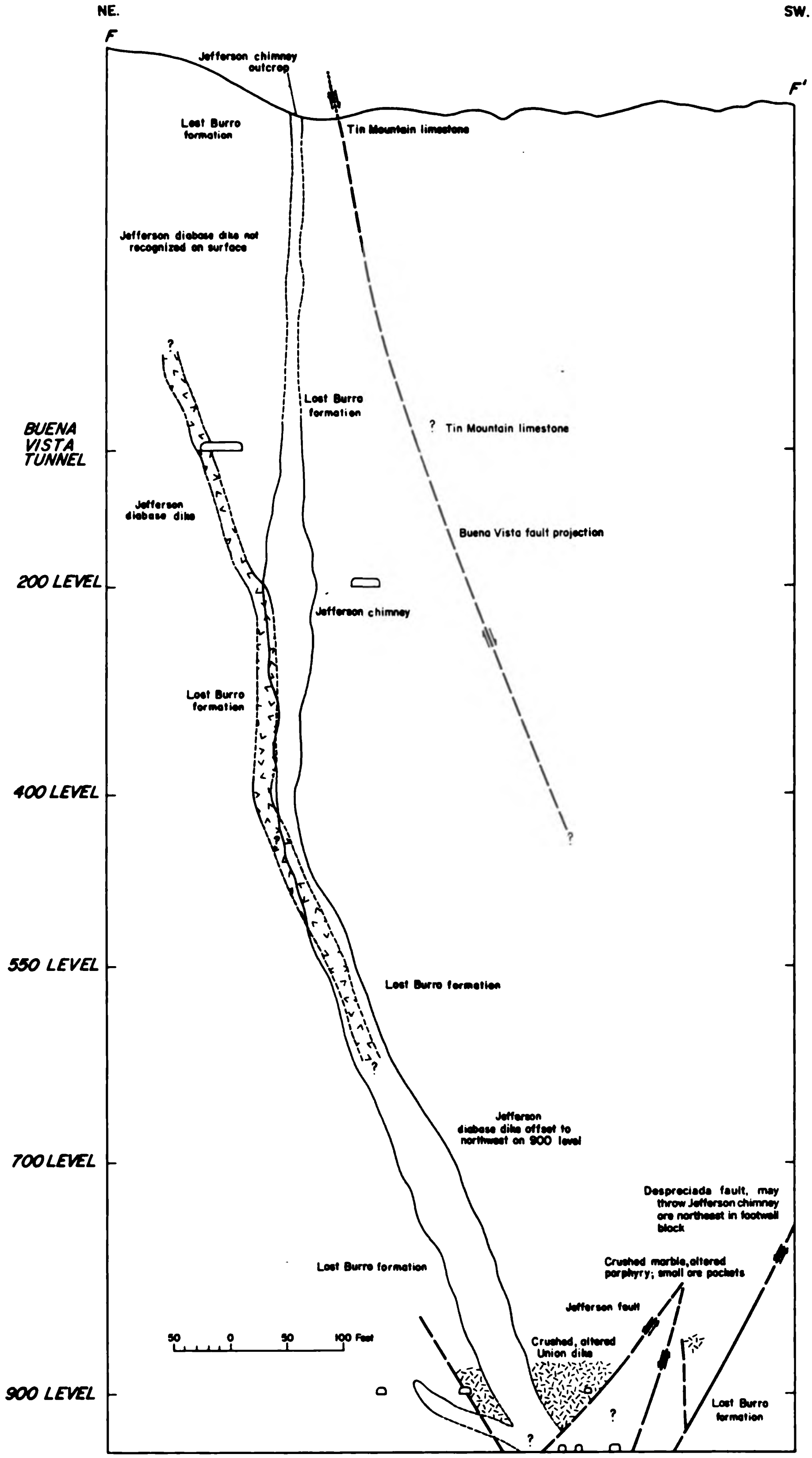

FIGURE 22.—Northeast-southwest section *F–F'* (observer facing southeast) through Cerro Gordo mine at Jefferson chimney. Shows relation of Jefferson diabasic dike to Jefferson chimney, and northwest normal faults of the Depreciada set cutting off Jefferson ore below 900 level. See plate 3 for location of section.

composed diabasic rock in old stopes contains scattered seams and veinlets of galena, lead carbonate, and limonite. According to Knopf (1918, p. 114), the greatest concentration of ore minerals was toward the footwall, and the dike rock itself appeared to be partially replaced by ore at some points. Contrary to expectation there was apparently very little mineralization of the marble walls. Evidently the shearing which permitted entry of sulfide-bearing solutions did not greatly affect the adjacent marble, which would theoretically be more susceptible to sulfide replacement than diabase. Only at the Jefferson chimney, which is intersected by the dike, was shattered marble strongly mineralized.

A large diabasic dike ore shoot was mined from above the 400 level down to the 700 level (fig. 18), where it appears to have terminated abruptly against a fault. On the 400 level the stope is continuous for 200 feet horizontally along the dike. This ore shoot narrowed downward to 60 feet along the dike on the 700 level. Ore was again picked up 150 feet south of its fault bottom on the 700 level and followed on down. Overall rake of this ore channel is southwest. Above the 700 level the rake is about 52°, whereas below the offset it steepens to about 76°. Southward rake of the dike ore agrees with that of other major ore bodies in this mine.

BUENA VISTA HANGING-WALL ORE BODIES

Several ore bodies were extracted in the early days from shattered marbles along the hanging wall of the Buena Vista fault. Most of these lay in the Santa Maria claim and extended from the Union tunnel southeast for about 800 feet.

Before recognition of the Buena Vista fault, it was found that the ores in the Santa Maria claim were abruptly cut off to the east by a footwall of greenish altered shale. For some time the so-called footwall shale was considered normally interbedded with the marble in a continuous stratigraphic sequence (fig. 23). Detailed geology later demonstrated that it was a fault slice of Chainman shale bounded by the Cerro Gordo fault and associated Union dike on the east and by the Buena Vista fault on the west. However, at some points the stratigraphic identity of shattered and altered ore-bearing marble of the Buena Vista hanging-wall blocks remains in doubt. Farther southeast at the Buena Vista tunnel, dark-gray marble occupying the same apparent structural position is evidently Tin Mountain limestone normally downthrown on this fault.

Ore bodies in the Santa Maria claim were mostly small, pockety and lay near the ground surface. Ore was mined in open pits and from the Union and Santa Maria ("86") tunnels underground (fig. 21, fig. 23). A large ore body known as the Santa Maria or Belshaw was removed from a big pit (125 by 140 feet) and by stoping from the Santa Maria tunnel. Much of the Santa Maria Co.'s lead production is probably to be accounted for by this one ore body. A so-called Union pit was opened above the Union tunnel (fig. 16, fig. 21) to extract ore from a small ore body reached also from the tunnel.[11]

Origin of the Buena Vista hanging-wall ore bodies has long been a matter of argument among mining engineers and geologists concerned with the Cerro Gordo mine. Two theories have been advanced: (a) that these ores are postmineralization down-faulted segments of the Union chimney and (b) that they were introduced along a premineralization shear zone which corresponds to the Buena Vista fault zone. At the present time the second theory is favored.

The postmineralization down-faulting hypothesis seems plausible, for an upward projection of the Buena Vista fault might well at an earlier time in geologic history have cut the rich Union chimney. However, the pipe is relatively narrow, whereas the mined-out Buena Vista hanging-wall ore bodies were several in number and were distributed at intervals for hundreds of feet along the Buena Vista fault. Also in this connection the Buena Vista fault should have cut and offset the Jefferson chimney in corresponding manner. Inferred upper segments of the Jefferson have been searched for unsuccessfully by underground drives, testing various theories of Buena Vista fault mechanics.

ZERO FISSURE VEIN

Named for the Zero tunnel, this fissure vein strikes N. 22° W. and dips steeply west (fig. 23). Its southeast projection on the Zero tunnel level heads for the Jefferson chimney, which it conceivably intersects (pl. 4). In a northwesterly direction the fissure vein weakens before meeting the San Felipe vein.

Together with associated minor fissures the Zero fissure vein has been prospected from the surface downward to the 700 level. On the 400 level, iron-bearing fractures in the Zero fissure zone strike N. 25° W. and lead southwest toward the Jefferson chimney, 200 feet away. Several small oxidized ore pockets in the zone of this fissure vein appear to be distributed within a kind of pipe or channel which rakes steeply southward approaching nearer the Jefferson chimney with depth.

[11] The term "Union ore body" has been applied to that mined from the so-called Union pit. To avoid confusion it is advisable to retain the name "Union ore body" for ores of the Union chimney, which was the source of most of the ore extracted from the Union mine.

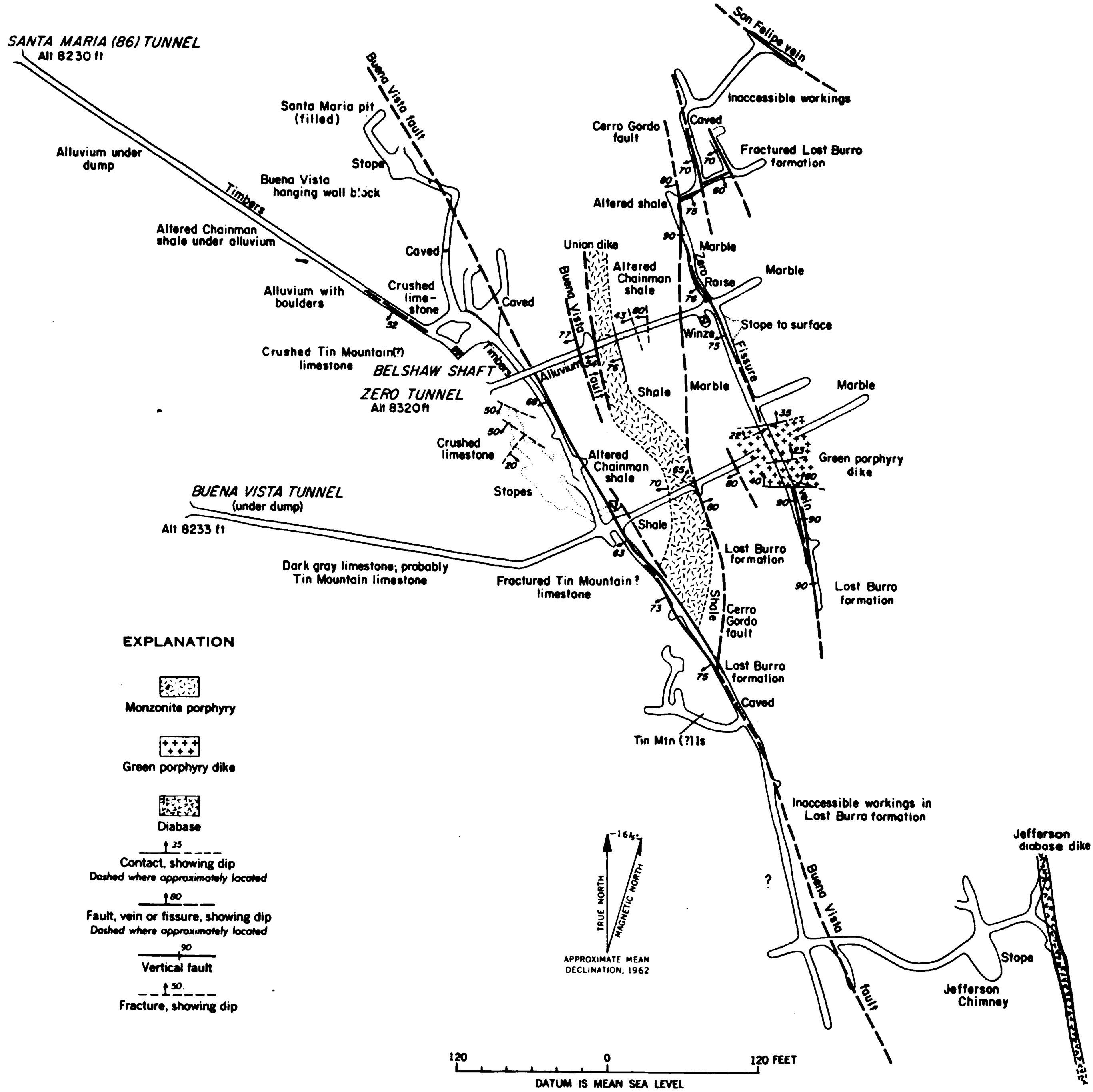

FIGURE 23.—Map of workings of Buena Vista, Santa Maria (86), and Zero tunnels. Shows Buena Vista fault cutting the Union dike and the older Cerro Gordo fault. Buena Vista hanging-wall ore bodies west of Buena Vista fault, Jefferson chimney ore to east. See also Zero fissure vein and San Felipe vein in Zero tunnel workings. For location see figure 16 and plate 3.

Paralleling the westerly bulge of the Cerro Gordo fault and Union dike, the Zero fissure vein appears to be structurally sympathetic to the master fault. On the 400 level the northwest-trending middle green porphyry dike enters the zone of Zero fissure mineralization.

SILICEOUS VEINS OF THE CERRO GORDO MINE

Two quartz veins (the San Felipe and Santa Maria) of the silver-bearing barite-tetrahedrite-galena type have been mined in the Cerro Gordo. These served also as sources of silica for smelter charges. A third quartz vein, known as the Castle Rock vein, lies near the ridge crest south of Cerro Gordo and has erroneously been referred to in reports (Tucker and Sampson, 1938, p. 438) as the "San Felipe vein." The Castle Rock vein has been prospected at depth from the Estelle tunnel.

San Felipe vein.—Surface croppings of this vein were worked by Mexicans about 100 feet south of the Union chimney pit (fig. 16). The vein strikes about N. 48° W. through Lost Burro marble at the outcrop. In the Union tunnel and on the 400 level, the San Felipe vein cuts the Union dike, but so far as known it does not pass west of the Cerro Gordo fault to enter Chainman shale. On the 400 and 550 levels this quartz vein meets the north green porphry dike and at several points is observed to lie either within the dike itself or to follow the dike contact with marble. The San Felipe vein comes very close to a meeting with the Union chimney near the 550 level. No evidence was found that the quartz vein actually cuts the Union chimney as long suspected. There seems to be no compelling geologic evidence to show that the San Felipe vein mineralization is younger than that of the Union chimney as long assumed.

Santa Maria vein.—This quartz vein parallels the San Felipe in the Lost Burro formation, between the 400 and 900 levels. On the 550 level (pl. 4), it lies 175 feet southwest of the San Felipe and like it dips steeply south. Like the San Felipe it appears to transect the Union dike. On the 900 level it is in contact with this monzonite porphyry dike on the west for part of its extent. White quartz with galena and heavy masses of limonitic matter occur within the vein on that level while pockets of cavernous limonite and iron-filled fractures are present in the brecciated marble between the Santa Maria and San Felipe veins. Both veins meet the north-trending Omega fault on the 900 level (pl. 4) but are not known to be offset or to continue beyond it into Chainman shale. Both have been traced 600 feet southeast in marble only to the Jefferson diabasic dike. On the 900 level, the Santa Maria vein and middle green porphyry dike converge northwest of the Belshaw shaft (pl. 4). No actual junction of dike and vein was exposed. Mine workings were driven west across the Cerro Gordo fault into Chainman shale on the 900 level in search for this vein. As these workings are inaccessible continuity of the Santa Maria vein through the Cerro Gordo fault was not ascertained.

The name "Santa Maria" is misleading in application to this siliceous vein. It seemingly bears no genetic relationship to the important near-surface ore bodies in the Santa Maria claim discussed above under ores of the Buena Vista hanging-wall marble blocks.

CERRO GORDO ZINC ORE BODIES

Cerro Gordo supergene zinc deposits were investigated by Knopf (1918, p. 106–108, 115–116) when these ores were being extracted in 1912. Except for a small stope on the 400 level immediately south and southeast of the Union chimney lead channel, all the old zinc workings were inaccessible during the present study in 1946. No primary sulfide zinc ores in minable quantities were discovered at Cerro Gordo.

Evidences of zinc migration and secondary deposition as carbonate are fairly common at edges of the lead channel stopes, where zinc carbonate crusts on marble were observed beneath limonitic matter. White clay of the halloysite type is in some places associated with the zinc and iron. Low zinc content of the smelted Union chimney lead ores is a matter of record. The explanation appears to be leaching of the primary ores (fig. 24) accompanied by redeposition principally as carbonate (smithsonite and hydrozincite) in adjoining marble. Lateral and downward movement of the zinc-bearing solutions was clearly facilitated by bedding structure and by fractures. According to Knopf (1918, p. 115), the carbonate zinc ores extended laterally from the Union lead channel as much as 100 feet. An irregular zinc pipe, observed by Knopf, had a diameter of about 5 feet and extended 150 feet in the direction of the lead channel. The 400-level zinc stope examined in 1946, followed marble bedding with a south dip of 34°. East-west stope length was 35 feet on bedding strike and stope height normal to bedding 8 to 10 feet.

Company stope maps prepared about 1917 show the extent of mined zinc carbonate east of the Union lead channel. The east boundary of the zinc channel is indicated as extending upward on a steep incline from the 400 level to the south drift of the near-surface Bullion tunnel (fig. 18, fig. 24). It is probable the incline is roughly that of marble bedding. Most of the Cerro Gordo zinc is believed to have come from this part of the mine. Figure 24 shows the relation of the Union zinc ore zone to the Union chimney lead channel.

Large volume of the Union zinc ore suggests that it could not all have come from the Union chimney as it exists today. Much of the leached zinc was probably derived from higher parts of the chimney or adjacent sulfide ore bodies which once lay above present ground level. Theoretically these upper sulfide bodies were removed by erosion or by faulting prior to erosion.

Zinc ores of the Union type could well have formed by leaching from the Jefferson chimney and other primary sulfide bodies as well. Thus far these inferred ores await discovery.

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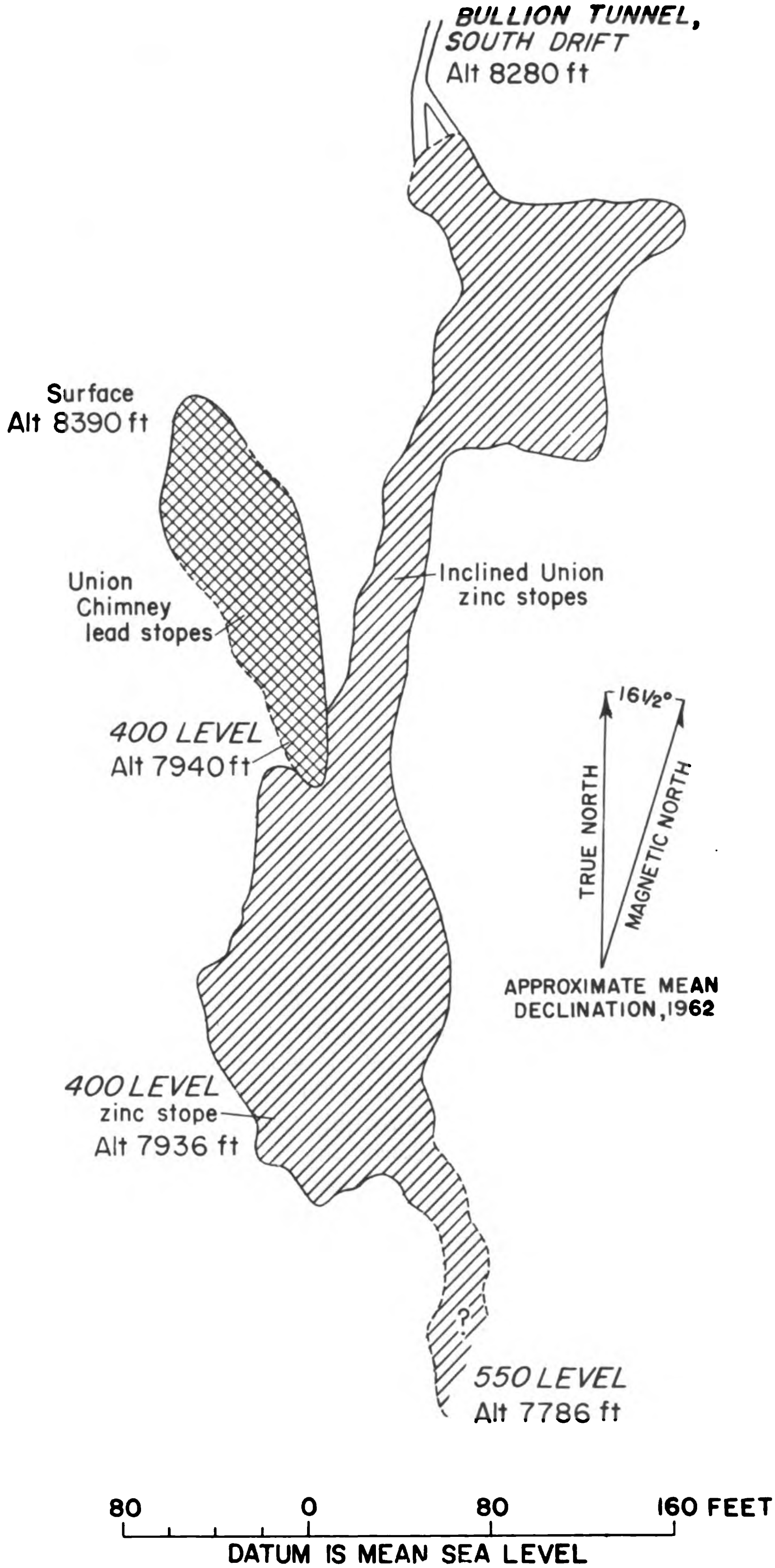

FIGURE 24.—Horizontal projection showing relation of supergene Union zinc carbonate ores to Union chimney lead ores. See also figure 18.

MORNING STAR MINE

The Morning Star mine (pl. 1) is 4,600 feet south of Cerro Gordo at an altitude of 7,725 feet. Most of the workings (pl. 5) are in shattered marble of the Devonian Lost Burro formation. These Devonian rocks do not crop out, being covered by dark-gray cherty Tin Mountain limestone of Mississippian age, which forms most of the surface above the mine. Small exposures of altered Chainman shale and quartzite of the Perdido formation occur at the west edge. No igneous rocks were recognized, the nearest exposed intrusion being quartz monzonite of the Cerro Gordo stock, 3,000 feet northeast. Hydrothermal alteration has nonetheless affected the carbonate rocks, and just south of the mine there is a large area of considerably altered and bleached Chainman shale.

The Morning Star mine is closer to the south nose of the Cerro Gordo anticline than is the Cerro Gordo, and unlike the Cerro Gordo its workings probably pass through the axial plane of the main fold. On the west the Tin Mountain limestone dips moderately to steeply southwest, but in the central part of the mine near the Gold Stope (pl. 5) the underlying Lost Burro formation reveals a swing in bedding strike which doubtless reflects closure of the south-pitching fold. In all probability the axial plane passes somewhere between the Gold Stope and the upper Morning Star tunnel face. This conclusion is supported by observed bedding-strike changes in the Charles Lease tunnel, 1,700 feet north of the Morning Star and by structural relations as mapped in the Estelle tunnel workings about 1,700 feet vertically below the Morning Star.

North-trending faults and fissures are important structural and ore-controlling features in the Morning Star, as they are also at Cerro Gordo. It is doubtful, however, that the Cerro Gordo fault exists as a discrete fault in the Morning Star, but the No. 1 and No. 2 fissures are evidently within the same general fault zone. The Union dike, which follows this fault zone at Cerro Gordo, was not recognized. Minor faults, veins, and fissures with northwesterly trend are numerous in the Morning Star.

The theoretically impervious blanket of Mississippian Chainman shale which once covered the Tin Mountain and Perdido along the anticlinal nose has been veiwed as of possible ore-controlling significance. There is, however, no compelling evidence of postulated damming action or lateral spread of mineralizing solutions in carbonate rocks beneath the impervious shale blanket.

The prominent north-trending faults and fissures in the Morning Star mine are possibly related in origin to stress conditions which existed prior to ore mineralization near the axial plane of the Cerro Gordo anticline.

The Morning Star workings were initially driven to undercut gossans and vein exposures in the iron-stained Tin Mountain limestone and to intersect an inferred south continuation of the Castle Rock siliceous vein, conspicuously exposed 1,900 feet northeast of the upper tunnel entry.

The most extensive mining and exploration were carried out on the north-trending No. 2 fissure (pl. 5) and in the vicinity of the Gold Stope. The north-trending No. 1 and No. 3 fissures were also prospected. A long east drive to cut the north-trending Castle Rock siliceous vein was unrewarded, and it is doubtful that the vein actually reaches the Morning Star from above.

Principal workings of the Morning Star are two tunnel levels driven northeastward to eastward. The upper or 1,700 tunnel at an altitude of 7,725 feet lies about 212 feet above the lower or 1,400 tunnel level. The terms "1,700" and "1,400" were applied by Morning Star operators to express approximate vertical distance above the important low-level Estelle tunnel whose portal lies at altitude 6,080 feet. A winze inclined east at roughly 80° connects the upper and lower tunnel levels on the No. 2 fissure. From the winze, four sublevels were driven on this fissure.

The upper or 1,700 tunnel extends for 1,670 feet through fractured, crushed and iron-stained limestone and marble in which bedding features are for the most part obscured. This condition is especially true of the 400 feet of tunnel from portal to main winze on the No. 2 fissure. The westernmost part of the tunnel appears to be largely in crushed Tin Mountin limestone, whereas east of the fissure only the more massive but commonly shattered marble of the Lost Burro was recognized.

North of the 1,700 tunnel the No. 2 fissure was mined upward to the ground surface.

The Gold Stope was developed on the 1,700 level about 100 feet east of the No. 2 fissure and 500 feet in from the tunnel portal. Judging from unpublished mining reports by F. D. Hanson, this is the most promising section of the mine. Ore distribution was apparently determined here by the bedding in marble of the Lost Burro which possesses a northwesterly strike and southwesterly dip of about 40°. Just east of the Gold Stope the beds appear to strike more nearly eastward, possibly in the vicinity of the axial plane of the south-plunging Cerro Gordo anticline.

On the 1,700 level the No. 1 fissure lies about 60 feet west of the No. 2 fissure and, like it, has a north-south strike and steep easterly dip. The Tin Mountain-Lost

Burro contact may lie either along this fissure or at some undetermined position in the crushed ground between it and the No. 2 fissure. Although nearly 30 feet wide and carrying scattered small ore pockets, the No. 1 fissure has not been productive.

The No. 3 fissure was found at a distance of 1,243 feet from the 1,700 tunnel portal, but no ore was developed on this north-trending structure. It appears to line up with a fissure cut on the Estelle tunnel level some 1,700 feet vertically below, and to which the term "No. 3 fissure" is provisionally applied on that level also.

The 1,700 tunnel was extended due east of the No. 3 fissure to a distance of 1,670 feet from the portal, but failed to meet the Castle Rock vein.

The lower or 1,400 tunnel (pl. 5) was started in altered Chainman shale underlain by a fine-grained quartzite rock having the appearance of novaculite of the Perdido formation. At a distance of 225 feet from the portal, the Perdido is faulted to the east against cherty Tin Mountain limestone. The displacement evidently took place on a bedding fault, with northwesterly strike and dip of 35° SW. Because of the altered and shattered condition of the rock, it was not found possible to fix the position of the Tin Mountain—Lost Burro contact. It appears to lie about 420 feet in from the tunnel entry.

The east-dipping No. 2 fissure was followed for 240 feet. It lies in fractured marble of the Lost Burro formation and contains large pockets of spongy limonite with brecciated marble and a small amount of quartz. There is evidence of postmineralization movement. Streaks of lead are reported in the No. 2 fissure vein on the 1,400 tunnel level, but in general the heavy iron oxides appear barren, and no important lead ore bodies are known to have been extracted here.

An east drive of 120 feet on the 1,400 level from the No. 2 fissure vein to a point beneath the Gold Stope did not encounter mineralization of significance.

Although the Morning Star mine has been active at various times since 1899, the production records are incomplete. Ore shipments totaling 4,127 tons are reported (F. D. Hanson, written communication, 1931) for the years from 1920 to 1931. Value of these shipments at the smelter is said to have been $107,145. Average assay is recorded as about 0.30 ounce of gold per ton, 31 ounces of silver per ton, 5 percent lead, 1 percent copper, and 3 percent zinc. A higher gold assay than would otherwise be expected is accounted for by averaging in production from the Gold Stope. Gold Stope ores are reported to have averaged about 0.80 ounce of gold per ton (F. D. Hanson, written communication, 1931). Ores from other sections of the mine seemingly ran less than 0.15 ounce of gold per ton.

ESTELLE TUNNEL

Portal of the low-level Estelle or Dellaphene tunnel lies 1½ miles southwest of Cerro Gordo at altitude 6,080 feet, or 2,240 feet below the Belshaw shaft collar (pl. 1). The tunnel is virtually straight and bears approximately N. 70° E. toward the Morning Star mine. Begun in 1908 (Knopf, 1918, p. 116), the Estelle tunnel reached its present face 8,100 feet from the entry by 1923. This ambitious and costly exploration drive passes on the tunnel level through almost half the higher Inyo Range just above altitude 6,000 feet, revealing an illuminating stratigraphic and structure section. Mapping of the Estelle tunnel provides data especially pertinent to future interpretation of Cerro Gordo geology. Face of the Estelle tunnel is situated in depth beneath a point 3,100 feet horizontally S. 24° E. of the Belshaw shaft collar. Altitude at the face is about 6,160 feet, or 1,078 feet lower than the Cerro Gordo 1,100 level.

Practical objectives of this drive were threefold: (a) to cut and explore the Castle Rock vein, (b) explore inferred deep continuations of veins in and around the Morning Star mine a mile south of Cerro Gordo, (c) to explore by a northward drive in Estelle ground for downward extensions of the south-raking Jefferson chimney and other Cerro Gordo ore channels.

Production of the Estelle has been small. The record from 1916 to 1926 shows 2,700 tons of lead-silver ore valued at about $80,000 (Hanson, F. D., written communication, 1931). Average metal content reported is 0.016 ounce of gold per ton, 20.00 ounces of silver, 21 percent lead, and 0.7 percent copper (fig. 25).

Rocks through which the Estelle tunnel passes range downward from the upper part of the Keeler Canyon formation at the portal to the upper part of the Hidden Valley dolomite at the face. The Cerro Gordo fault brings Chainman shale into contact with the Hidden Valley; thus cutting out the Perdido, Tin Mountain, and the Lost Burro formations. East of the Cerro Gordo fault, the Lost Burro was encountered in the 800-foot raise above the Estelle tunnel level. A quartzite bed recognized on the 660 raise level is believed to represent zone A of the Lost Burro. Because of very heavy ground encountered in the weak Chainman shale almost continuous timber and lagging were required.

Character of the Cerro Gordo anticline is shown by mapping of the Estelle tunnel which crosses the axial zone of the fold to penetrate calcitic dolomite in the

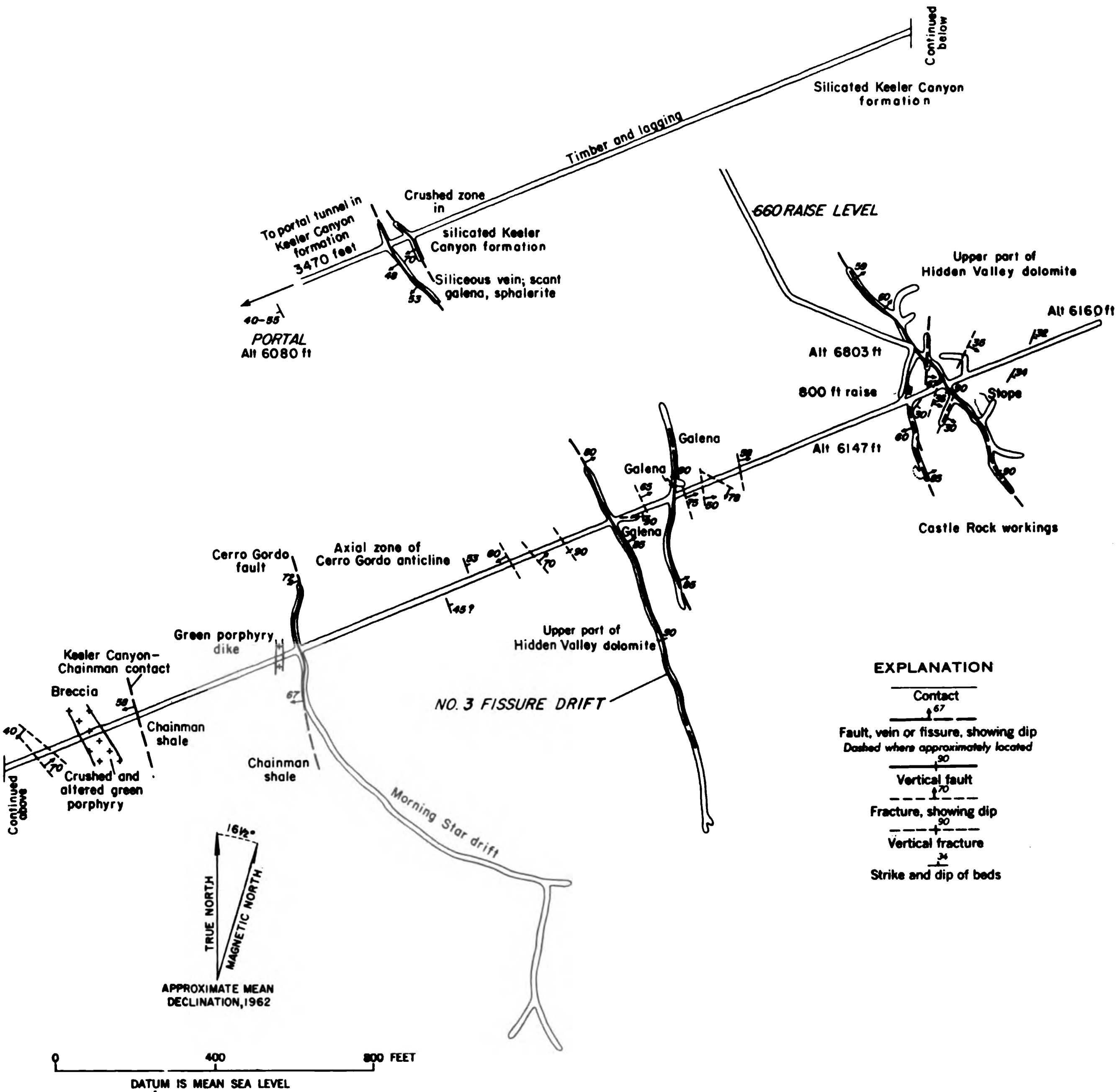

FIGURE 25.—Map of inner workings of Estelle tunnel. See figure 27 for Castle Rock workings. For location see plate 1.

upper part of the Hidden Valley of the east limb. The axial plane apparently lies in shattered dolomite about 6,350 feet in from the portal (or 1,750 feet out from the face). However, the bedding attitude could not be determined in this sheared zone. Silication and silicification have likewise taken place in this vicinity. Upward projection of the assumed axial plane to the supposed position of the axial plane at the Cerro Gordo mine and Charles Lease tunnel suggests a steep westerly dip of the axial plane. This, however, does not accord with plotted structure sections, which even suggest that a primary axial plane may dip eastward at a lower angle. Obviously the asymmetrical Cerro Gordo anticline owes much of its configuration to faulting and is far from a simple fold.

West of the incompetent Chainman shale belt, the steep to moderate westerly dip of the Keeler Canyon formation is observed to advantage (fig. 26). Dip is reversed locally in minor drag folds, which are common in the thinner bedded members of this formation on the surface as well as underground. East of the 6,350 tunnel station, few reliable bedding dips were

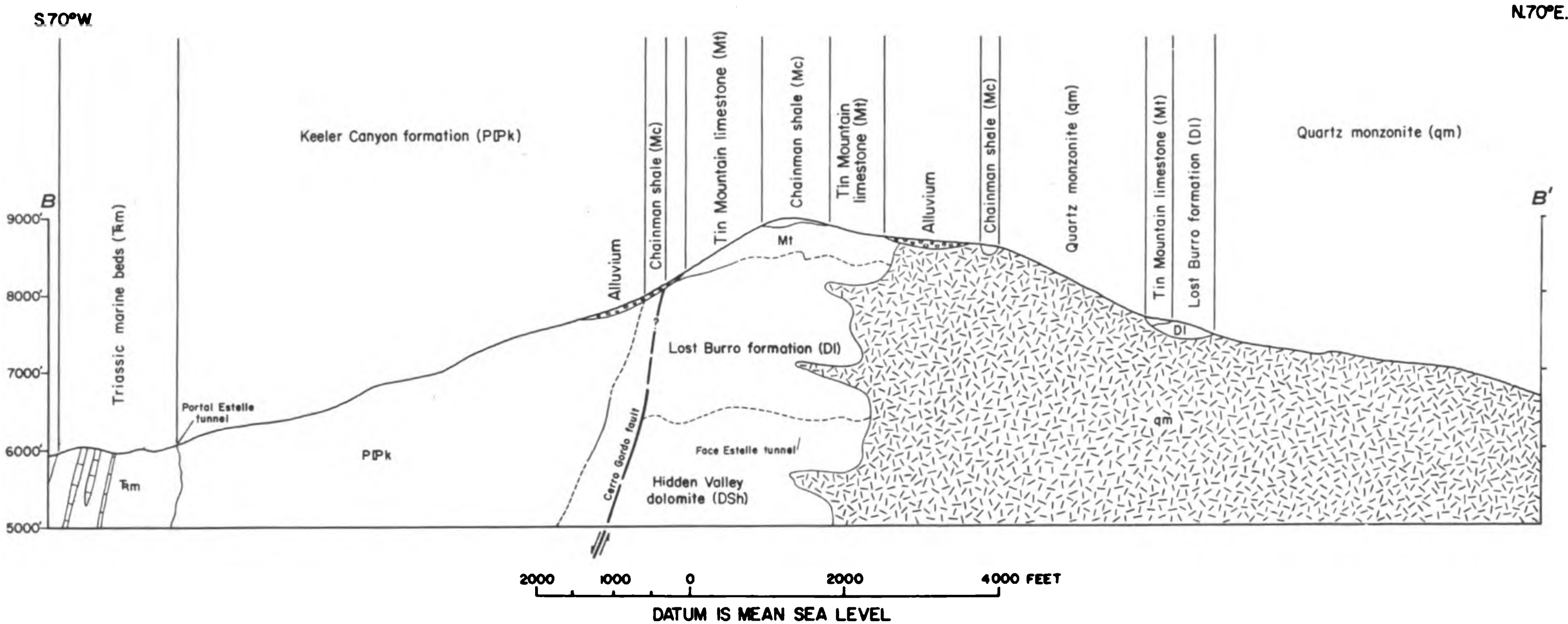

FIGURE 26.—Southwest-northeast structure section *B–B'* through higher Inyo Range along line of Estelle tunnel. Rocks plunge toward observer in axial zone of Cerro Gordo anticline. Horizontal and vertical scales same. See plate 1 for location of section.

observed in the altered and disturbed Hidden Valley until the 800-foot raise is reached. Low dips to the east, averaging about 32°, are recorded between the raise station and the face.

Principal workings of the Estelle mine (fig. 25; fig. 27) are: (a) the Morning Star drift, (b) the No. 3 fissure drift, (c) the Raise workings, and (d) the Castle Rock workings. Only in the Castle Rock workings near the face was minable ore discovered. Favorable showings were reported in marble of the Lost Burro near the top of the 800-foot raise.

Quartz-vein mineralization was found in the Keeler Canyon formation, 3,762 feet in from the tunnel entry (fig. 25). The vein strikes northwestward and dips westward about 70°. Scattered bunches of galena and sphalerite are present here in quartz seams. A weaker vein with lower westerly dip was cut 60 feet west. These quartz-bearing veins were introduced on a northwest-trending shear zone about 120 feet wide and in general accord with bedding attitude of the Keeler Canyon formation. In the vicinity of the shear zone the limestone was altered to calc-silicate rock.

No. 3 fissure drift.—This drift (fig. 25) is so named because it appears to line up with the No. 3 fissure of the Morning Star mine. It strikes a few degrees west of north and stands nearly vertical. White quartz, calcite, and limonitic pockets are present, but no ore mineralization was seen on the Estelle tunnel level.

A north-trending (fig. 25) fissure 160 feet east of the No. 3 fissure where cut by the tunnel was somewhat more encouraging, showing small bunches of galena and other sulfides with calcite at several points. This fissure also stands nearly vertical but has a steep dip eastward at some places. Disseminated sulfides including galena occur with quartz between the two fissures along the main tunnel, about 50 feet east of the No. 3 fissure drift.

Morning Star drift.—These workings (fig. 25) extend southeastward from the main tunnel and pass beneath the Morning Star mine. The drift leaves the main tunnel along the fault which separates Chainman argillite on the west from calcitic dolomite of the footwall block. This major fault dips westward about 70° and is believed to be the Cerro Gordo fault. The footwall dolomite is somewhat siliceous in places. No Perdido formation or Tin Mountain were recognized with assurance, and the footwall dolomite is believed to represent only the upper part of the Hidden Valley. A small amount of quartz and weak copper stain occurs on the fault or "contact vein" as the Cerro Gordo fault is referred to in the Cerro Gordo mine. Small pods of galena are present in the footwall dolomite along the main tunnel, 30 feet east of the fault. One hundred feet south of the main tunnel, the Morning Star drift departs east of the fault to follow weakly mineralized fissuring in the footwall dolomite. The drift cut several barren quartz seams and limonitic pockets, but no ore was discovered. The Hidden Valley dolomite contains open solution caverns at several points along the northwest fissuring.

Raise workings.—Tributary upper level workings connect with the tunnel through an 800-foot vertical raise (fig. 25, fig. 27). This raise was started in 1923 at a station 7,580 feet in from the portal. On a level 660 feet above the tunnel, a station was cut and workings driven northwestward toward the Cerro Gordo mine, eventually extending 2,200 feet in this direction. Specifically this northwest drive was planned to explore the La Despreciada claim on which the Estelle Mining Co. held a long-term lease. Uncertainty now exists as to whether this 660-level drive actually entered the La Despreciada claim. In any case it penetrated ground nearby at a point roughly 400 feet below the Cerro Gordo 1,100-level (fig. 18). So far as known, ore was not found, although in 1925, at about the same time, ore was discovered in the Despreciada claim at a higher altitude by west drives from the Cerro Gordo 700 and 900 levels.

Though inaccessible during the present study, the upper raise workings are known in part through existing maps and reports compiled by mining company engineers and geologists during the last period of operation, about 1931.

While the main objective of the 660-level drive was not achieved, valuable geological data were obtained.[12] According to F. D. Hanson (written communication, 1931), the 660 level is largely in Lost Burro formation. Monzonite porphyry, cut near the north end, is reported to be the Union dike. Of special stratigraphic significance is a 15-foot quartzite bed found just north of the station where the 660 level leaves the 800-foot raise. As this quartzite probably pertains stratigraphically to Lost Burro zone A, a rough measure of southward plunge of the Cerro Gordo anticline can be obtained. It is perhaps significant that bedded ore occurred locally in close proximity to a similar quartzite in the lower part of the Union chimney.

Castle Rock workings.—The Castle Rock workings (fig. 27) are east of the 800-foot raise on the tunnel level, where they lie, 2,850 feet vertically below the south end of the Castle Rock vein outcrop (fig. 27). Drifts were run on northwest-trending fissures which

[12] Recorded in written communications and maps compiled by F. D. Hanson, T. L. Chapman, and J. J. Beeson, supplemented by observations of H. L. Eckloff (oral communication, 1955).

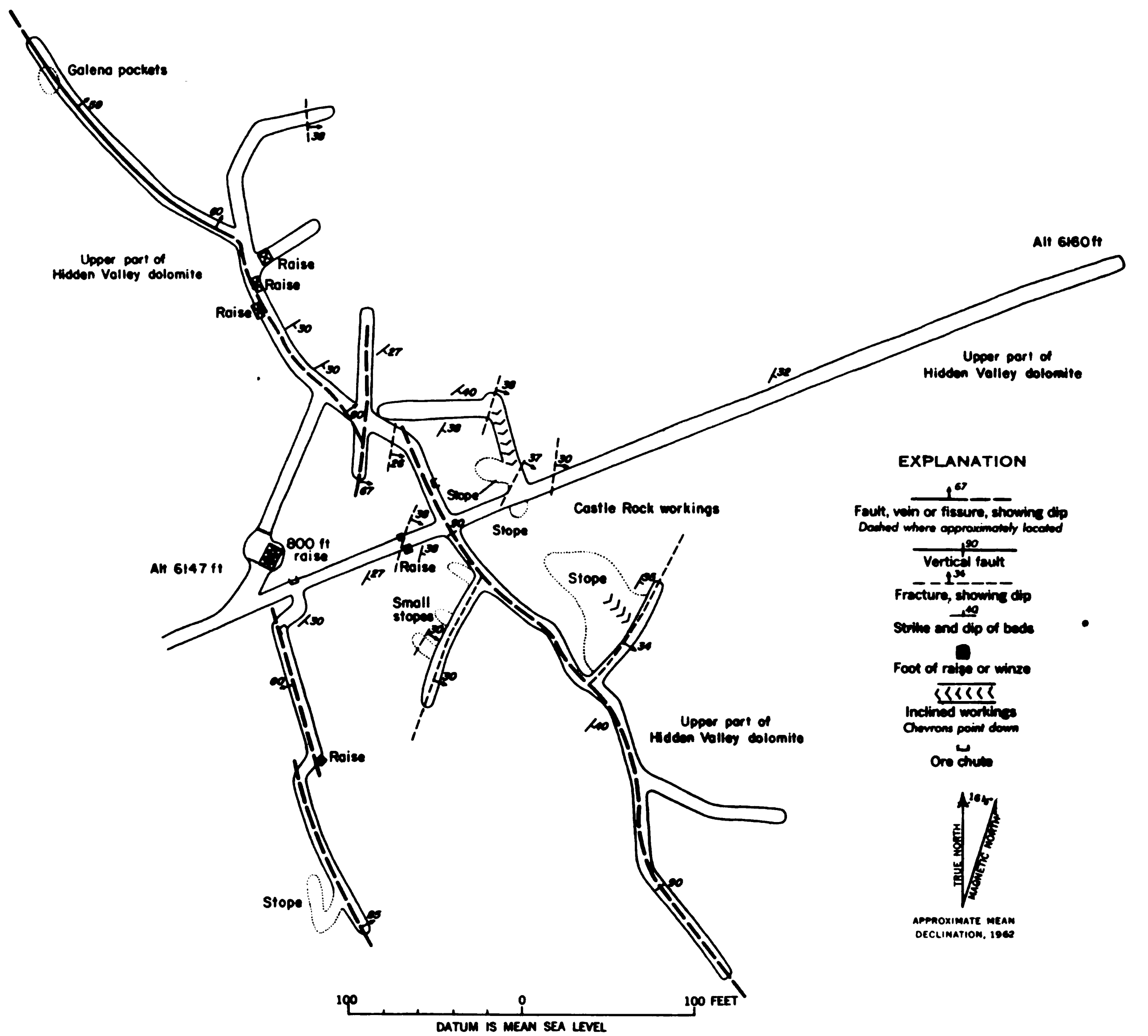

FIGURE 27.—Map of Castle Rock workings, Estelle tunnel. Named for the Castle Rock siliceous vein which crops out at summit of Inyo Range, 2,850 feet above these workings. See figure 25 for location.

carry seams of white quartz containing pockets of galena, with small amounts of limonite, and copper staining. Ore was extracted from a number of small bedding stopes which strike northeastward and dip southeastward with bedding in calcitic dolomite of the upper part of the Hidden Valley. Northerly trend of the Castle Rock vein on the surface is not apparent at this depth. Ore was extracted also above the tunnel level from pockets near the bottom of the 800-foot raise.

Quartzite interbeds of the Lost Burro were not recognized on the Estelle tunnel level. In the future these quartzite key beds might well be sought as stratigraphic markers, and possible indicators of a favorable ore zone. (See fig. 18.)

CHARLES LEASE TUNNEL

The Charles Lease tunnel (fig. 28) is 2,800 feet south of Cerro Gordo and 1,700 feet north of the Morning Star mine at altitude 7,960 feet (pl. 1). It extends eastward for about 1,600 feet, mainly through marble of the Lost Burro and was designed to explore fissured ground east of the Cerro Gordo fault, and especially to prospect the Castle Rock vein. Although this wide siliceous vein is exposed on the ridge crest about 1,550 feet east of the tunnel portal, it was evidently not cut by the tunnel. Faulting may account for absence of the vein where it would be expected in depth.

Exact position of the Cerro Gordo fault is unknown at the Charles Lease tunnel, but the portal is believed to lie in its vicinity. The first 700 feet of tunnel penerates intensely fractured and brecciated limestone and marble of the Tin Mountain and Lost Burro formations within the Cerro Gordo fault zone. Where bedding attitudes were recognizable in this zone of intense deformation, they are from 25° to 35° west (fig. 28) and therefore may pertain to the west limb of the Cerro Gordo anticline.

From the standpoint of geologic structure the tunnel is especially important, for like the Morning Star it passes through the axial part of the Cerro Gordo anticline. At a distance of 670 feet from the portal, or just east of the zone of intense fracturing, the marble of the Lost Burro dips southward; at 1,000 feet the bedding strike swings to northeast and the beds slip southeastward within the east limb of the structure.

No productive veins were found in driving the Charles Lease tunnel (fig. 28). There are numerous discontinuous iron-stained or iron-incrusted seams and fractures, many of which strike northwestward. Fractures with roughly north-south trend are likewise common, especially in a zone about 1,200 feet east of the portal. No definite indication of the Morning Star No. 3 fissure was recognized.

Ore mineralization was cut at a single point 810 feet east of the portal on a nearly vertical vein striking N. 37° W. (fig. 28). At this position there are small pockets of galena and lead carbonate in association with calcite, iron oxide, and weak copper staining. This vein was followed for 70 feet, but pinched down in both directions.

It is noteworthy that the most promising ore discoveries on the Estelle tunnel level were made about 1,880 feet vertically below the Charles Lease tunnel face and, therefore, beneath the Castle Rock vein outcrop.

IGNACIO MINE

The Ignacio silver mine (pl. 1) is in Cerro Gordo Canyon half a mile southwest of Cerro Gordo. This mine was discovered and first operated by Mexicans, who attempted to reduce the ore in small vasos, remains of which may be seen in that vicinity. In later years it was operated in conjunction with the Cerro Gordo mine.

The Ignacio workings (fig. 29) comprise more than 4,000 feet of tunnel, a glory hole, and many pits and trenches. The principal tunnel workings are now inaccessible. Underground data here presented were obtained from stope and tunnel maps found in the Cerro Gordo mine records.

Rocks of the Ignacio mine comprise altered Chainman shale, basal limestone of the Keeler Canyon, and quartz monzonite of the small Ignacio stock (fig. 29). The limestone of the Keeler Canyon has been partly altered to calc-silicate rock (fig. 30). Garnet and epidote are abundant locally. Black shale and siltstone of the Chainman are silicified, having the appearance of argillite and fine-grained quartzite. Some of the wallrocks appear to represent a stratigraphic zone near the top of the Chainman wherein black shale and limestone alternate. Small offshoots of the Ignacio intrusive body are present in the mine workings, but the main body of the stock is to the west.

Some of the Ignacio ore occurred in quartz veins which have a northwesterly strike on the surface. Seams, veins, and pockets of white iron-stained vein quartz may be observed at the mine. On the ridge top 1,200 feet southeast of the main Ignacio tunnel entrance, northwest-trending quartz veins cut the calc-silicate rocks of the Keeler Canyon near a northwest-trending fault separating argillite of the Chainman from the lower part of the Keeler Canyon. Galena and tetrahedrite are present in the vein quartz. From this point west to the Sunset mine, the Cerro Gordo Canyon area was extensively prospected, as evidenced by the many short tunnels and pits.

Localization of the more important Ignacio mineralization appears related to intersections of northwest-trending quartz veins and fissures by the northeast-striking Ignacio fault (fig. 29). On the surface the Ignacio fault dips steeply southeastward and separates argillite of the Chainman from the silicated limestone of the Keeler Canyon. Displacement appears to be normal, with Keeler Canyon downthrown on the south side. Rocks of both footwall and hanging wall are considerably deformed and brecciated near the fault. On the hanging-wall side the Keeler Canyon beds show minor folds along the Cerro Gordo-Morning Star mine road. The Ignacio fault is intersected on the surface by a northwest-trending fault passing through the Ignacio glory hole.

The main Ignacio tunnel extends eastward for 500 feet at the canyon bottom and turns southeastward to run beneath the glory hole. Extensive northwest-trending tunnel workings were driven on quartz veins, fissures, or faults. However, the largest stopes have a northeasterly distribution and appear to be controlled by faulting or fissuring on the downward projection of the Ignacio fault zone. Ore in these shoots was extracted by southeast-inclined winzes and stopes which connect the main tunnel level at an altitude of about 7,580 feet, with higher tunnel levels 190 to 210 feet above the main tunnel. Southeast inclination of these fissure workings ranges from 40° to 60°. Stope maps suggest that ore bunches may have been found where the northeast-trending Ignacio fault intersects the northwest-trending veins and fissures.

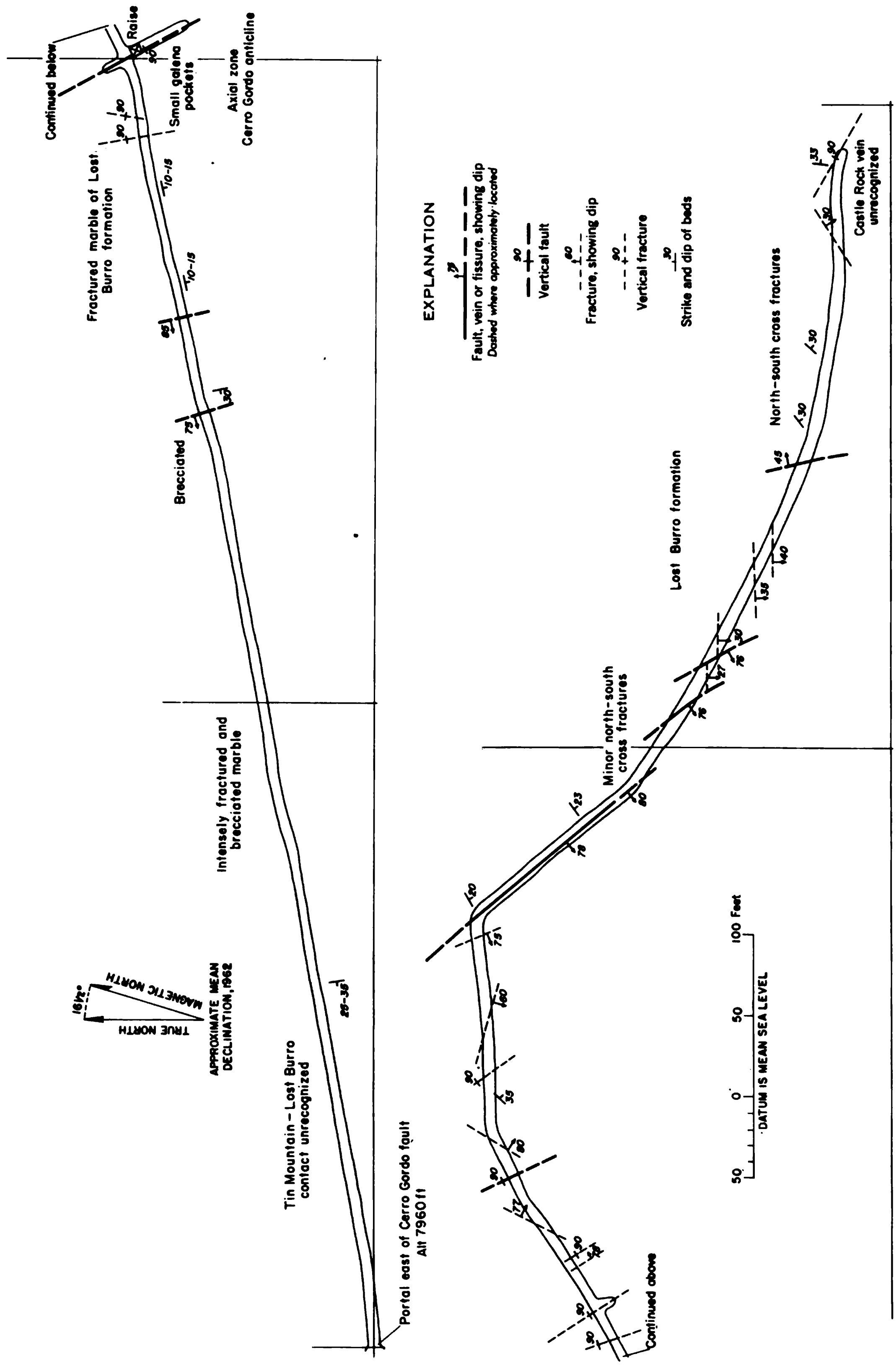

FIGURE 28.—Map of Charles Lease tunnel workings. For location see plate 1.

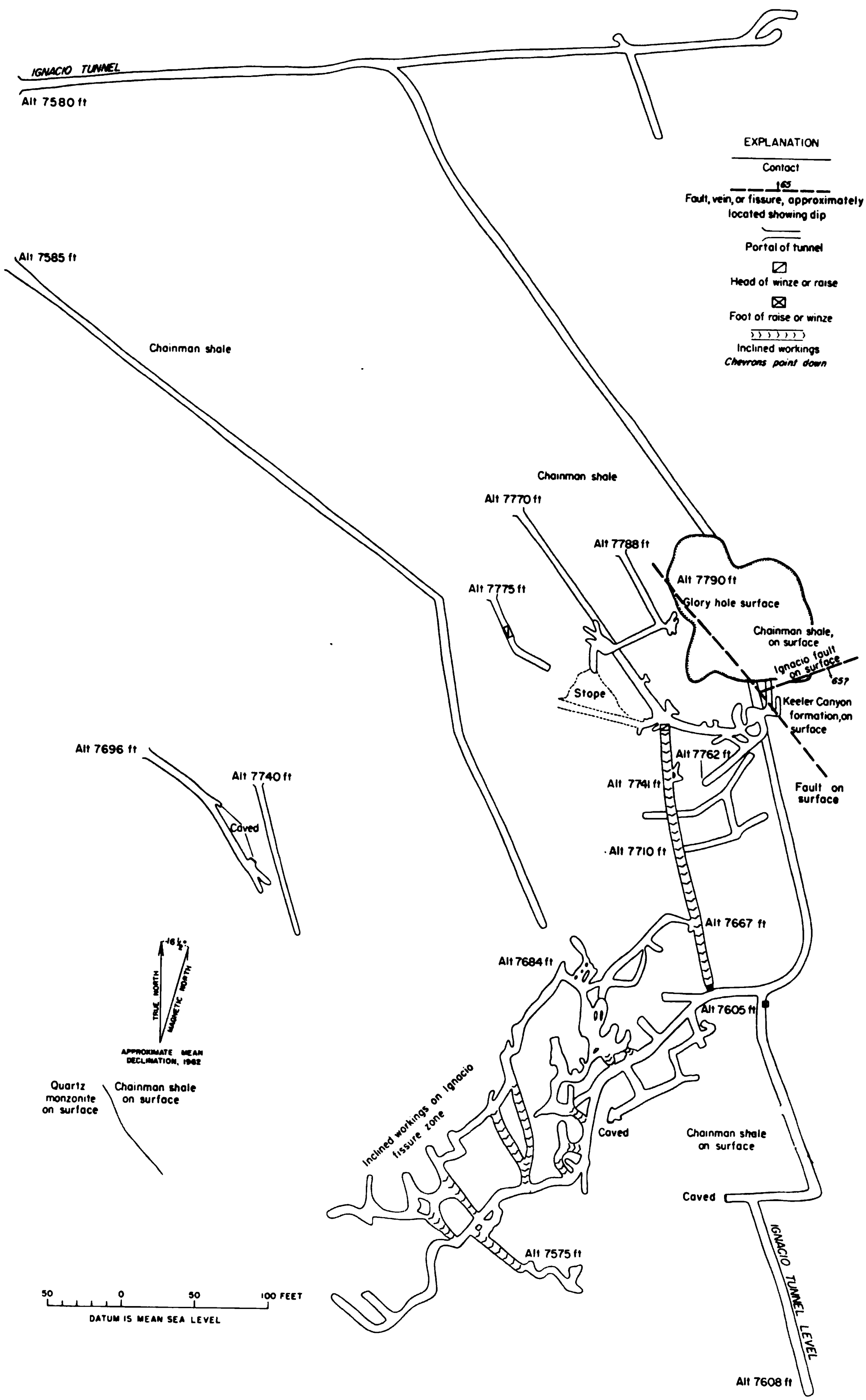

Figure 29.—Map of Ignacio mine workings. For location see plate 1.

FIGURE 30.—View looking southwest along west Inyo Mountains slopes. Chainman shale on left middleground. At the middle of photograph is silicated limestone of the Keeler Canyon formation over altered Chainman shale; Ignacio mine glory hole right of center. Dark-gray basaltic flows in distance lead up to the Coso Mountains. Photograph by L. G. Henbest.

HART MINE

The Hart mine, variously known as the Cerro Gordo Extension and Lead Queen, is near the Inyo crest 1 mile north-northwest of Cerro Gordo (pl. 1). The workings are in uppermost Tin Mountain limestone close to the Perdido contact. An inclined shaft was sunk on a vein whose attitude does not depart greatly from that of the limestone bedding which strikes about N. 30° W. and dips 52° W. Level workings were driven from the shaft at depths of approximately 50, 80, 100, 150, and 200 feet. On the hanging-wall side of the vein the Tin Mountain-Perdido boundary appears to be comformable and a depositional contact.

Only the interval from the surface to the 50-foot level was examined. Within this interval the vein attains a width of 1½ feet and includes limonitic matter containing small galena pockets and lead carbonate together with coarse calcite. There is weak copper staining in places. Shearing has taken place along the vein over a width of 3 to 4 feet. The vein pinches down both north and south of the shaft on the 50-foot level.

Reported distribution of ore pockets from the 50-foot to the 150-foot level suggests a general southward rake.

Contrary to an earlier conception, the ore deposits of the Hart mine bear no direct genetic relation to those of the Cerro Gordo mine and appear to be controlled by bedding trends in strata much higher stratigraphically than those in which the Cerro Gordo ores occur. The vein of the Hart mine is in no way related structurally to the Cerro Gordo fault.

A 40-ton ore shipment made in 1936 by the Cerro Gordo Extension Mining Co. and believed to have been extracted from the Hart mine is reported on the basis of smelter returns to have had the following metal content:

Gold	troy ounce per ton	0.115
Silver	do	15.40
Copper	percent	.55
Lead	do	20.20
Iron	do	6.2
Zinc	do	3.0

Value of the ore after treatment was reported as $18.49 per ton.

CROSSCUT TUNNEL

The Crosscut or Emperor tunnel, 1,070 feet long (fig. 31), is on the Cerro Gordo Pipeline trail 1¾ miles northwest of Cerro Gordo at altitude 8,360 feet (pl. 1). This tunnel was driven N. 81° E. for 870 feet, turning to N. 60° E. for the last 200 feet. The objective is uncertain, but it appears to have been designed to intersect inferred downward continuations of mineral showings in prospect holes along the Inyo summit. Although unsuccessful as an exploratory venture, the Crosscut tunnel provides an excellent stratigraphic and structure section through much of the higher part of the range.

Fifty-five feet in from the portal the Tin Mountain limestone is overlain conformably by the Perdido formation (fig. 31). Contact of the Tin Mountain limestone and Lost Burro formation is a fault seen at 212 feet from the portal. The fault strikes N. 28° W. and dips 65° W. It reveals spongy limonite and a small amount of siliceous matter, but is not known to carry ore. A stratigraphic thickness of only 120 feet of dark-gray cherty limestone of the Tin Mountain is present; this thickness suggests considerable thinning of the formation as a result of the fault displacement. Beyond the 212-foot station, nearly all the tunnel is in Lost Burro marble, which strikes northwestward and dips westward 38° to 50°. Three northwest-trending green porphyry dikes of the andesitic type common to the area were cut in marble of the Lost Burro formation.

SUNSET MINE

The Sunset mine (fig. 32) is in Cerro Gordo Canyon, three-fourths of a mile west of Cerro Gordo at altitude 7,250 feet (pl. 1). An adit was driven 385 feet north-northeastward into limestone in the lower part of the Keeler Canyon formation which is partly silicified and partly altered to calc-silicate rock. Prospect pits on the surface show iron-stained limestone and pockets of limonitic matter. Hydrothermal alteration of the limestone of the Keeler Canyon is similar in character but less intense than that shown at the Ignacio mine a quarter of a mile southeast, where the formation was intruded by the Ignacio quartz monzonite stock.

Ore occurred sparingly in two fissure veins which intersect on the tunnel level 150 feet northeast of the portal (fig. 32). The Wheeler vein strikes N. 17° W. and dips steeply southwestward; the other vein strikes N. 25° E. and stands nearly vertical. The northwest-trending Wheeler vein conforms to bedding strike, but its southwesterly dip is steeper than that of the bedding. In addition to the Wheeler vein which carries lead, short lateral workings were driven on three barren or weakly mineralized fractures also with northwesterly bedding strike. Black chert occurring 36 feet west of the productive Wheeler vein may pertain stratigraphically to the "golf ball" chert zone of the lower part of the Keeler Canyon formation.

Ore was extracted from a small stope at the intersection of the two fissure veins. The ore pocket at this point widened to 4½ feet and was worked vertically for about 20 feet. Seams of galena may be observed here in association with limonite and clayey matter. Most of the primary sulfide was oxidized and leached, leaving masses of honeycombed cavernous limonite and jaspery limonitic matter.

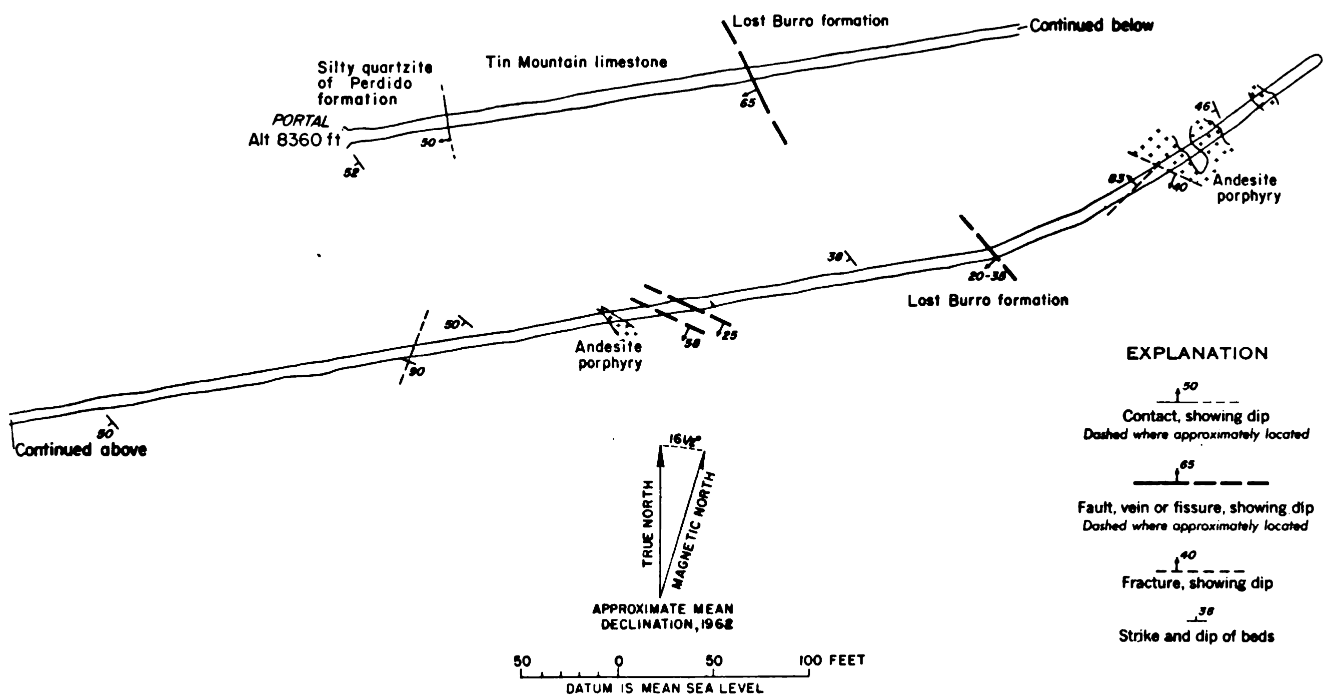

FIGURE 31.—Map of Crosscut tunnel workings. For location see plate 1.

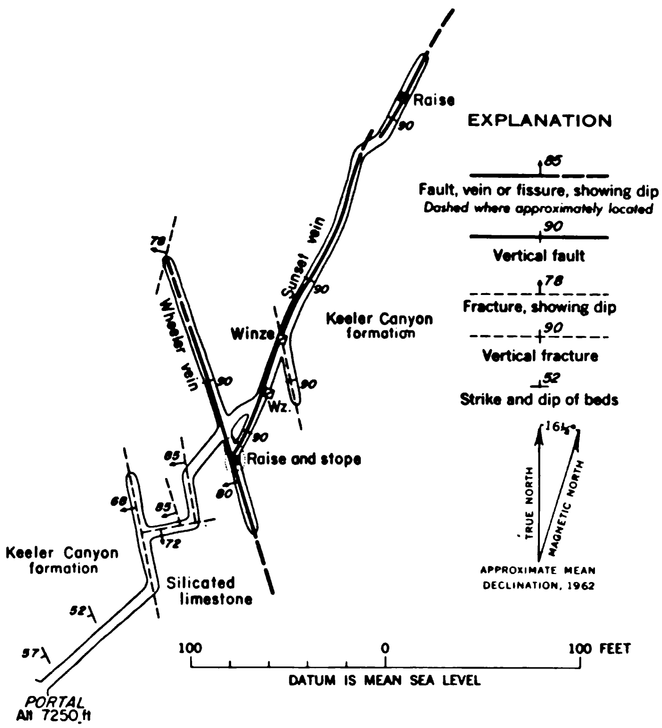

FIGURE 32.—Map of Sunset mine workings. For location see plate 1.

The iron-rich Wheeler vein was followed for 150 feet and shows galena stringers at several points. Toward the northwest face the vein pinches down, and at the southeast face it reveals barren limonitic-clayey pockets in leached chalky vein matter derived in part from sheared calc-silicate wallrock. Postmineralization movement has taken place along this fissure vein.

The northeast fissure vein was followed for 230 feet beyond the intersection of the two veins, being nearly vertical throughout. Small galena bunches are present within 120 feet of the intersection. These have been prospected by winzes and raises.

The Sunset mine differs from most smaller mines of the Cerro Gordo area in that it was not opened on quartz-tetrahedrite mineralization so characteristic of the region.

UPPER NEWTOWN MINE

The Upper Newtown mine is 1,550 feet north of the Belshaw shaft on the east side of San Lucas Canyon (fig. 16) at altitude 8,040 feet. It is 650 feet north of the Bullion tunnel, which is considered part of the Cerro Gordo mine. On the surface the veins in Lost Burro formation were worked by means of trenches. These veins in the footwall of the Cerro Gordo fault range in strike from north to N. 20° E.

The Upper Newtown tunnel (fig. 33) enters sheared and altered green-weathering Chainman shale resting against marble of the Lost Burro on the Cerro Gordo fault. A north-south drift permits examination of the fault for 100 feet. The Tin Mountain limestone has evidently been faulted out, and no trace was recognized of the Union dike which was in fact not found north of the east-west cross fault which cuts off the Bullion fissure (fig. 16). In the Upper Newtown the Cerro Gordo fault dips westward 75°, steepening locally to nearly vertical.

The Upper Newtown vein (fig. 33) was encountered 220 feet east of the tunnel portal. It strikes N. 20° E., dips east 60° to 70°, and was followed underground for 500 feet. Pockets of ore were extracted near the point where the tunnel first intersected the vein. On the south the Upper Newtown vein appears to have been cut off by the east-west cross faulting which also terminated the Bullion fissure on the north. This cross faulting throws the Cerro Gordo fault west at the Upper Newtown. It seems improbable that the Upper Newtown vein and the Bullion fissure were continuous before the offset.

NEWTOWN MINE

The Newtown mine (fig. 16) is on the east side of San Lucas Canyon at altitude 7,900 feet and 2,500 feet north of the Belshaw shaft at Cerro Gordo. It comprises a vertical shaft and two tunnels, of which only the lower was mapped. The shaft, now inaccessible, is estimated to be 300 feet deep and to judge from the size of its dump communicates with extensive lateral workings. For many years this shaft has been covered by a large and conspicuous shafthouse and is equipped with a horse whim.

At the Newtown mine the Cerro Gordo fault is the contact between altered Chainman shale and marble in the lower part of the Lost Burro formation, however, the Union dike was not recognized. This is the northernmost mine which owes its existence to geologic conditions responsible for localization of the important ore bodies in the Cerro Gordo mine itself.

The tunnels were driven eastward to explore veins in the Lost Burro marble of the footwall block. The lower tunnel (fig. 34) extends S. 76° E., 590 feet. Passing for 75 feet through alluvium and Chainman shale it meets the Cerro Gordo fault, beyond which it penetrates the Lost Burro. The fault strikes N. 25° E. and dips 62° W. Overall trend of this master fault is more nearly north-south. Downthrow of the Chainman shale hanging-wall block at this point may be 1,500 feet, for the footwall rocks near the fault appear to be the lower part of the Lost Burro. Marble of the Lost Burro within 100 feet of the fault is somewhat silicified, being very dense and hard. It is partly dolomitic as is

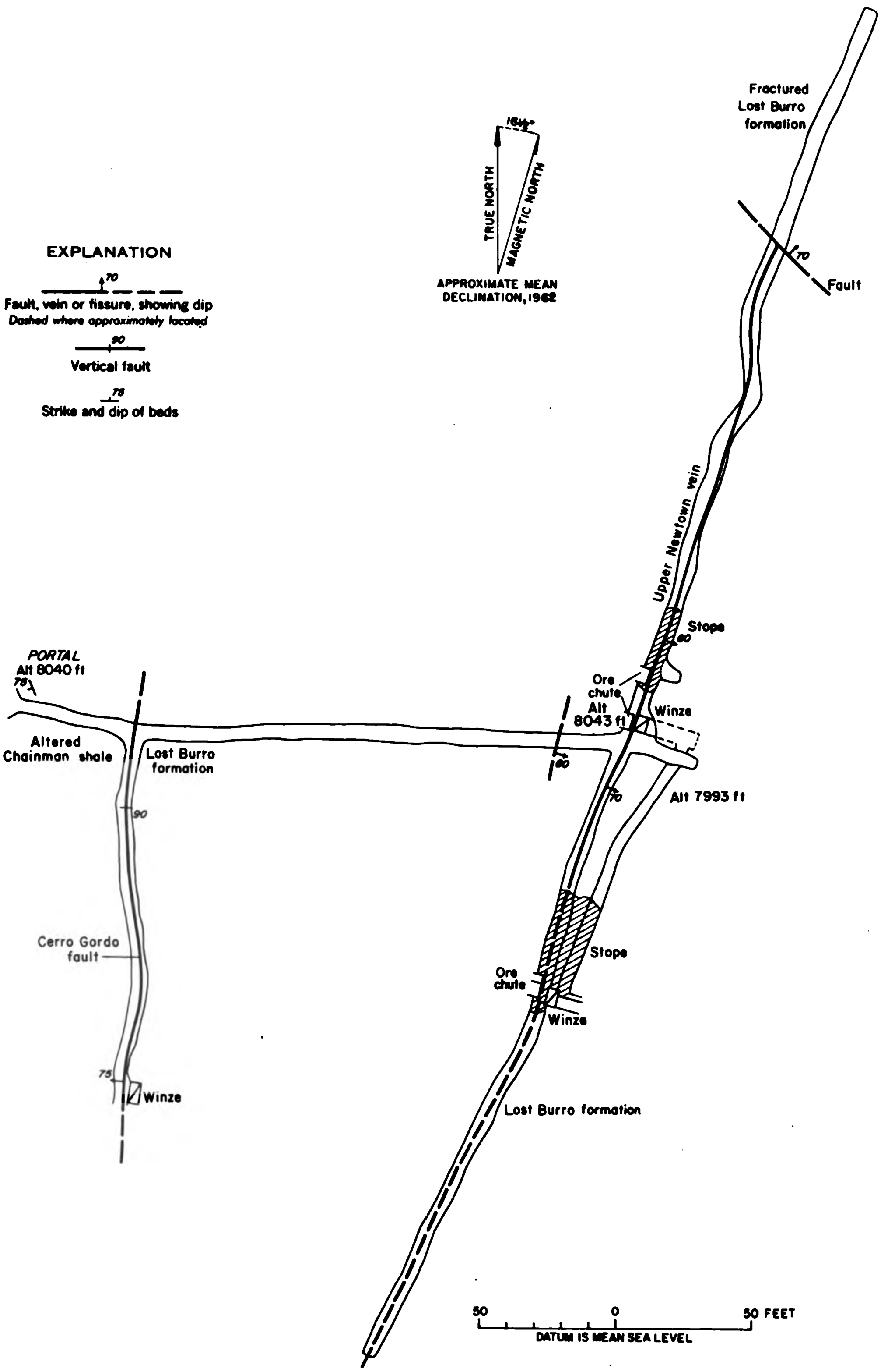

Figure 33.—Map of Upper Newtown mine workings. For location see figure 16.

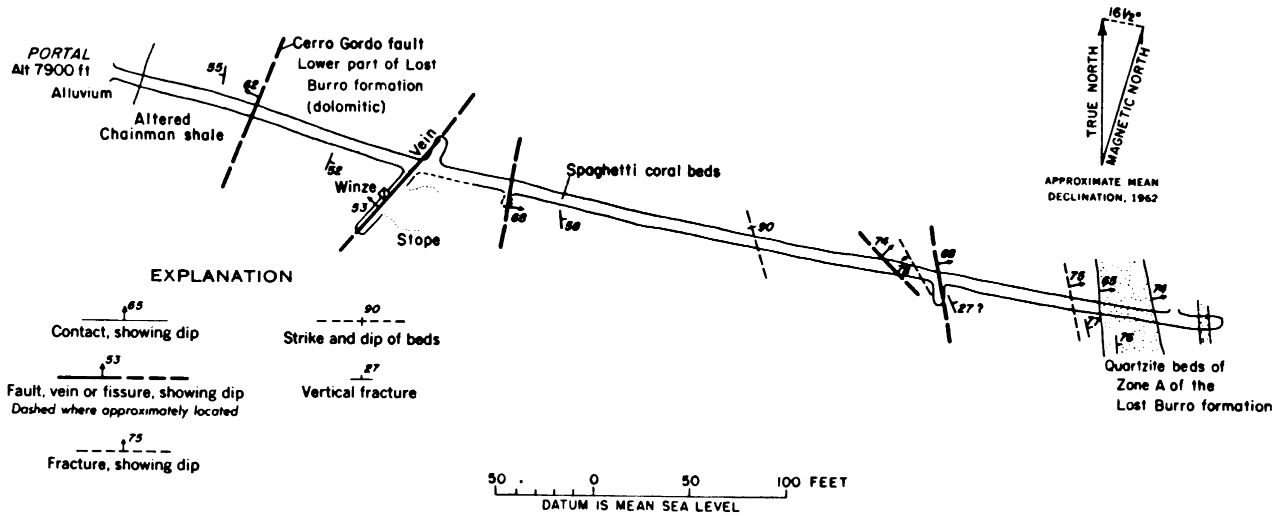

FIGURE 34.—Map of lower tunnel of the Newtown mine workings. For location see figure 16.

characteristic of the lower part of the Lost Burro. Throughout the tunnel the beds dip eastward from 52° to 65°.

Two hundred and forty feet from the portal the lower tunnel passes from the dolomitic lower part of the Lost Burro into nondolomitic marble containing "spaghetti coral." Light-gray quartzite beds in the upper part of zone A of the Lost Burro formation occur near the tunnel face.

Limonitic mineralization is present along the Cerro Gordo fault in the lower tunnel. This is characteristic of the otherwise barren Cerro Gordo fault, referred to in the Cerro Gordo mine as the "contact vein." Several fractures which traverse the Lost Burro northward to northwestward show limonite and calcite incrustations. A minable vein was cut 85 feet east of the Cerro Gordo fault. This vein, which strikes N. 41° E. and dips 53° W., carries scattered galena bunches in a limonite-calcite gangue. Copper stain is present. The galena vein was worked upward to the surface.

ELLA MINE

The Ella mine is 4,200 feet north of Cerro Gordo on the east side of San Lucas Canyon at altitude 7,480 feet (pl. 1). It consists of tunnel workings and trenches on a quartz-tetrahedrite-galena vein in blocky dolomite of the Hidden Valley dolomite.

The main tunnel (fig. 35) was driven southeastward along the Ella vein, which strikes N. 70° W. and dips 65° to 76° N. White quartz and calcite of the vein contain limonite, tetrahedrite, and seams or bunches of galena. Copper staining is derived by oxidation of tetrahedrite. The Ella vein varies in width from a foot to 6 feet, and is crushed locally by late movement. At some points the vein is frozen to dolomite walls, and small quartz stringers containing galena pass outward into the dolomite. Numerous cross fractures intersect the vein. On the tunnel level it is cut off 570 feet southeast of the portal in a fault zone 90 feet wide. In this zone most of the breaks strike northeastward to nearly northward like that at the tunnel face. The north-trending Cerro Gordo fault has not been traced to the Ella mine; its inferred projection beyond the Newtown mine lies some 400 feet west of the Ella tunnel entry.

Fairly large amounts of siliceous ore were extracted from raises on the Ella vein (fig. 35), which widens upward to 6 feet in one stope. Branching of the vein, 120 feet in from the tunnel portal is explainable by fault repetition.

The Ella vein is similar in attitude and mineralogic character to the Perseverance or San Lucas vein, which is 2,200 feet southeast. It is conceivable the two may actually be coextensive. The Ella, Perseverance, Newsboy, and Belmont mines all occupy the same general zone of northwest fissuring; these received quartz vein mineralization like that of the Santa Maria and San Felipe veins at Cerro Gordo. Some of the siliceous fluxing ore used in the Cerro Gordo smelters probably came from the Ella.

PERSEVERANCE MINE

The Perseverance or San Lucas mine lies 1 mile northeast of Cerro Gordo at altitude 7,600 feet. Dolomite of the upper Hidden Valley is the country rock. Bedding has a low dip southward, ranging from nearly flat to 25°.

There are two tunnels in which the principal workings follow a siliceous vein that strikes N. 60° to 70° W. and dips 50° to 65° N. The Perseverance vein attains a width exceeding 2 feet. It consists of white quartz, calcite, and barite with a few small galena

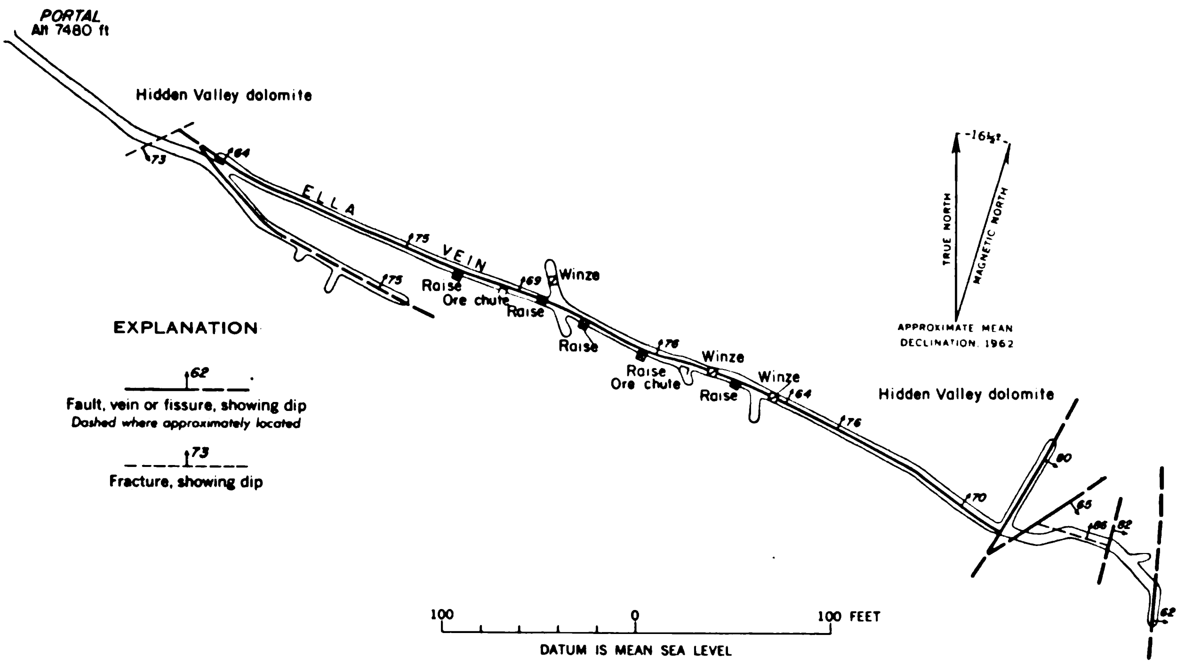

FIGURE 35.—Map of main tunnel, Ella mine. For location see plate 1.

pods and scattered, but rather conspicuous tetrahedrite from which copper staining is derived.

In the eastern or No. 1 tunnel, the vein is discrete for 220 feet, beyond which it becomes less distinct (fig. 36); here and there, in the fractured dolomite, are many quartz seams and pockets. These seams extend in various directions, some being parallel to bedding. The easternmost No. 1 tunnel workings enter a brecciated zone trending a few degrees east of north. This zone appears to cut off the Perseverance quartz vein.

The western or No. 2 tunnel workings follow the Perseverance quartz vein for 285 feet, the vein widening at one point to more than 4 feet. Small galena bunches are present (fig. 36).

Ore was extracted from raises in both tunnels. It is probable that this mine was a source of siliceous fluxing ores employed by the Cerro Gordo smelters. Smelter returns from a 30-ton shipment made in 1917 and believed to have come from the Perseverance gave the following tenor:

		Assay per ton
Gold	troy ounces	0.05
Silver	do	40.95
Copper	percent	2.00

No lead is reported. Value per ton was $31.39 after treatment. This ore evidently contained a large amount of tetrahedrite, to judge from the copper content, and the gold and silver probably occurred in both the galena and the tetrahedrite. Ores of the Perseverance required extensive hand sorting.

The Ella vein and the Perseverance vein may be the same, but their continuity has not been demonstrated by mapping.

SILVER MINES IN BELMONT CANYON

Mines and prospects of Belmont Canyon (pl. 1) lie 1¼ miles east of Cerro Gordo at altitudes ranging from 7,200 to 8,000 feet. Principal mines of this area are the Belmont and the Newsboy. Some of the siliceous ores used as fluxing material in smelting at Cerro Gordo during the 1870's came from this area, mainly from the Belmont. The mines of Belmont Canyon lie in quartz monzonite of the Newsboy stock and in altered Tin Mountain limestone along the intrusive border. The name "Belmont district" has been applied to the vicinity of these mines.

NEWSBOY MINE

Northwest-trending siliceous veins in the intrusive body were mined by means of a tunnel driven westward to northwestward near the canyon bottom. The northwest-trending veins dip 45° to 60° E. Ore was discovered in a marble xenolith intersected by a northwest-trending vein. It was extracted by means of an inclined winze and sublevel on the vein, 40 feet below the tunnel level.

Knopf (1918, p. 117), who visited the Newsboy in 1913, describes the ore as follows:

The ore of the Newsboy mine is a coarse white rather vuggy quartz that carries considerable galena, subordinate tetra-

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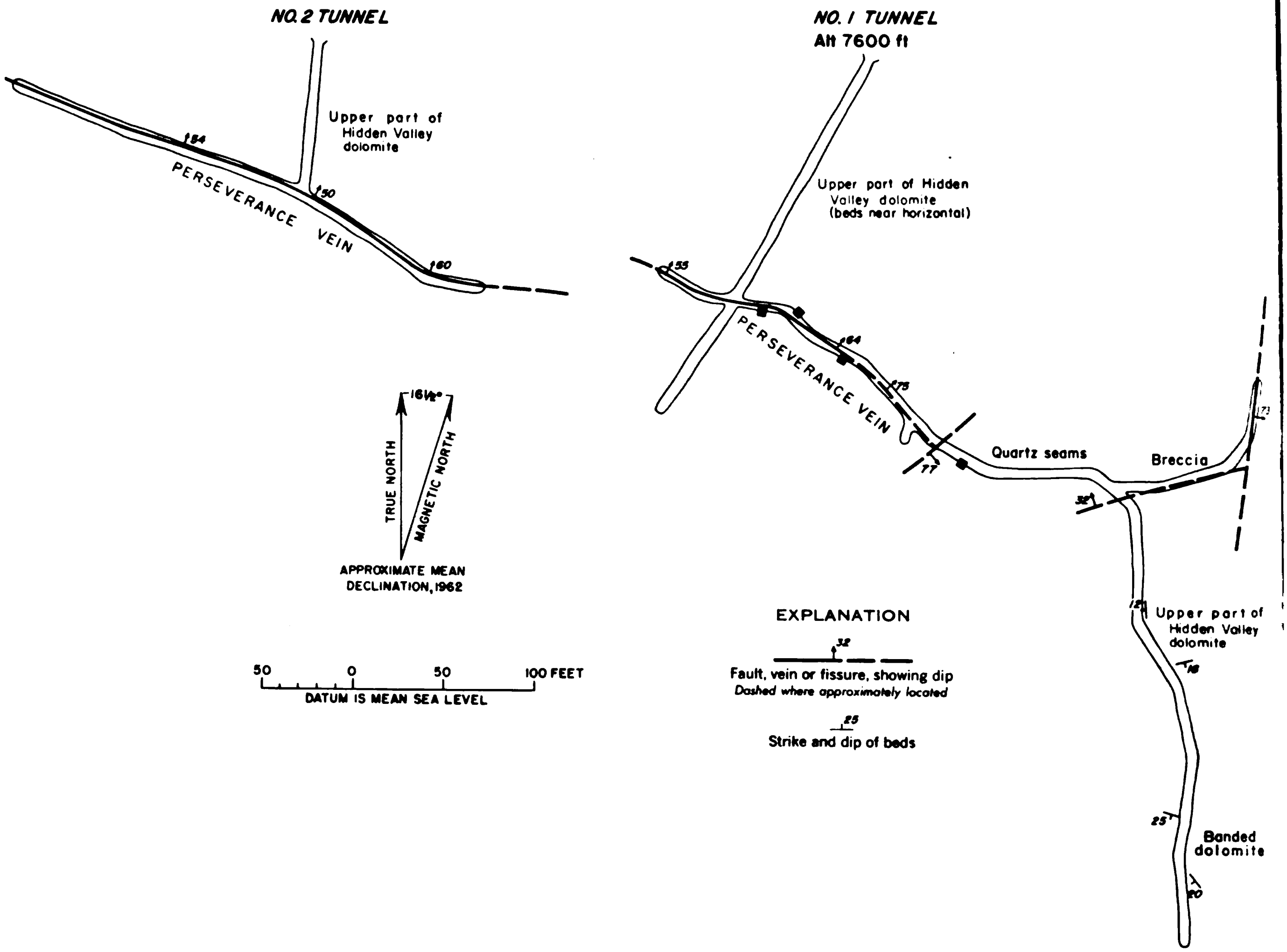

FIGURE 36.—Map of workings of Perseverance or San Lucas mine. For location see plate 1.

hedrite, and rarely chalcopyrite and pyrite. Cerusite and a bright yellow ocher, bindheimite, are common; iron oxide is present; and linarite in places is excellently crystallized in needles. The ocher is rich in precious metals, assays showing 508 ounces of silver and 3.4 ounces of gold to the ton.

The Newsboy quartz-vein ores resemble those of the Perseverance and the Ella, and this mine occurs in the same broad zone of northwest fissuring as the other two.

BELMONT MINE

The Belmont workings, now largely inaccessible, lie on the south side of Belmont Canyon and comprise several tunnels driven southeastward into Tin Mountain limestone and contiguous quartz monzonite. The principal workings enter a body of Tin Mountain limestone, about 750 feet long. This Tin Mountain is largely enclosed by the intrusive body and is partly altered to calc-silicate rock with garnetized seams. Some of the mine openings are near or on intrusive contacts. According to Goodyear (1888, p. 250–254) most of the Belmont ore was mined from quartz veins within the quartz monzonite itself, as was true likewise of the Newsboy mine. The main Belmont vein appears to have had a northwesterly trend and southwesterly dip of 60° to 70°. Argentiferous quartz veins of this mine contain calcite, galena, pyrite, chalcopyrite, tetrahedrite, and copper-bearing minerals derived by oxidation of the primary sulfides. Goodyear (1888, p. 252) reports native silver. Old furnace ruins nearby attest to early reduction attempts at the mine. Most of the ore was hand sorted for transportation by mule to the Cerro Gordo smelters. According to Goodyear, as much as 100 tons per month reached Cerro Gordo in 1870. Better grades of Belmont ore are said to have carried 165 ounces of silver per ton, valued at about $190. Rather large volume of the dumps indicates that the Belmont workings were extensive and suggests that a considerable part of the Cerro Gordo fluxing ores could well have been obtained here (Raymond, 1873, p. 18).

REGISTER OF IMPORTANT FOSSIL LOCALITIES

[See detailed geologic map, pl. 1]

5. Permian Owens Valley formation with fusulinids 8,000 feet south of Estelle tunnel portal.
6. Permian Owens Valley formation with fusulinids on Cerro Gordo road 1 mile southwest of Estelle tunnel portal.
7. Permian Owens Valley formation with fusulinids 6,200 feet south of the Morning Star mine. Owens Valley formation rests on Keeler Canyon formation with angular unconformity at this locality.
8. Chainman shale and Perdido formation with fossiliferous limestone lenses 3,000 feet southeast of the Morning Star mine.
9. Fossiliferous limestone in Chainman shale 3,200 feet south-southeast of the Morning Star mine.
10. Keeler Canyon formation with fusulinids 8,000 feet south of Black Basin, on east side Upland Valley.
11, 12. Keeler Canyon formation with fusulinids on south side Black Basin.
13. Black Chainman shale with fragmentary plant remains 6,200 feet southeast of the Morning Star mine.
14. Keeler Canyon formation with fusulinids at Cerro Gordo tram line.
15. Lower Triassic fossils northwest of Estelle tunnel portal.
16. Fusulinids in upper part of Keeler Canyon formation on Cerro Gordo road. Fusulinids from this locality are assigned to Wolfcamp Early Permian.
18. Permian Owens Valley formation with fusulinids in ravine on northwest side Conglomerate Mesa.
20. Fossils in lower part of the marine Triassic 1,000 feet south of Cerro Gordo road.
21. Coral locality in Tin Mountain limestone 700 feet north of Belshaw spring.
22. Abundant fossils in lower part of the Chainman shale on Pipeline trail 7,000 feet northwest of Cerro Gordo. Invertebrate fossils in shale and limestone; plant remains and shark teeth in shale.
23. Lower part of the Keeler Canyon formation with fusulinids 1,800 feet northwest of Belshaw shaft at Cerro Gordo.
24. Chainman shale with fossils near Hart mine at Inyo crest 4,500 feet north-northwest of Cerro Gordo.
25. Upper part of the Hidden Valley dolomite with gastropods 800 feet northwest of the Perseverance mine.
26. Fossils in Lost Burro formation 2,000 feet northwest of the Newsboy mine.
27. Lower part of the Lost Burro formation with *Stringocephalus* on trail 1,200 feet northwest of the Perseverance mine.
28. Fossils in Lost Burro formation, San Lucas Canyon 900 feet southwest of the Ella mine.
29. Lower part of the Lost Burro formation with *Stringocephalus* 800 feet south of the Perseverance mine.
30. Ely Springs(?) dolomite on ridge 4,200 feet north of the Ella mine.
31. Silurian fossils in Hidden Valley dolomite, south side Bonham Canyon 7,300 feet southeast of Bonham talc camp.
32. Silurian fossils in Hidden Valley dolomite, south side Bonham Canyon 4,800 feet southeast of Bonham talc camp.
33. Pogonip fossils on south side Bonham Canyon 8,000 feet north of the Perseverance mine.
34. Corals in Lost Burro formation, north side of canyon near Bonham talc camp.
35. Corals in Lost Burro formation 6,500 feet north-northwest of the Ella mine.
36. Tin Mountain limestone with fossils. Quartz Spring area McAllister, 1952), Ubehebe district, Inyo County, California; about 2 miles southeast of Quartz Spring on southeast side of Perdido Canyon.
37. Perdido formation with fossils. Quartz Spring area, Ubehebe district, Inyo County, California; type area of Perdido formation between Perdido Canyon and Rest Spring gulch. Upper part of Perdido formation, possibly unit 10 of McAllister (1952, p. 23).
38. Chainman shale with fossils. East of Cerro Gordo mining district map area; 4,700 feet northeast of the Newsboy mine.
39. Ely Springs dolomite with silicified fossils. Talc City Hills at south end Inyo Mountains, Darwin quadrangle; prominent spur in dark gray dolomite west of road to Santa Rosa flat, half a mile southeast of Hard Scramble prospect and 2½ miles northwest of Talc City mine (Hall and MacKevett, 1958, pls. 1 and 2).
40. Pogonip group. Limestone with fossils on west side of Inyo Range about half a mile southwest of the Long John mine and 1¾ miles east of northernmost summit of Granite Hill.

REFERENCES CITED

Amsden, T. W., 1949, Stratigraphy and paleontology of the Brownsport formation (Silurian) of western Tennessee: Yale Univ., Peabody Mus. Nat. History Bull. 5.

Belshaw, M. W., 1884, The cost of mining and smelting; *in* Hanks, H. G., Catalogue and description of the minerals of California: California State Min. Bur., 4th rept. of the State Mineralogist, p. 252.

Chalfant, W. A., 1933, The story of Inyo: Bishop, California; published by W. A. Chalfant, 430 p.

Eilers, Anton, 1873, Contributions to the records of lead smelting in blast furnaces, *in* Raymond, R. W., Statistics of mines and mining in the States and Territories west of the Rocky Mountains for 1872: 5th Ann. Rept. U.S. Commissioner Mining Statistics, U.S. Treasury Dept., p. 354–356.

Fairbanks, H. W., 1894, Preliminary report on the mineral deposits of Inyo, Mono, and Alpine counties: California State Min. Bur., 12th Ann. Rept. of the State Mineralogist, p. 472–478.

———1896a, The mineral deposits of eastern California: Am. Geologist, v. 17, p. 144–158.

———1896b, Notes on the geology of eastern California: Am. Geologist, v. 17, p. 63–74.

Gale, H. S., 1914, Salt, borax, and potash in Saline Valley, Inyo County, California: U.S. Geol. Survey Bull. 540, p. 416–421.

———1915, Salines in the Owens, Searles, and Panamint basins, southeastern California: U.S. Geol. Survey Bull. 580, p. 251–323.

Gilbert, G. K., 1875, Geology of portions of Nevada, Utah, California, and Arizona: U.S. Geog. and Geol. Surveys W. 100th Mer. Rept., v. 3, pt. 1, p. 17–187.

Goodyear, W. A., 1888, Inyo County: California State Min. Bur., 8th rept. of the State Mineralogist, p. 224–309.

Hague, Arnold, 1883, Abstract of report on the geology of the Eureka district, Nevada: U.S. Geol. Survey 3d Ann. Rept., p. 237–290.

Hague, Arnold, 1892, Geology of the Eureka district, Nevada: U.S. Geol. Survey Mon. 20 (with atlas).

Hall, W. E., and MacKevett, E. M., 1958, Economic geology of the Darwin quadrangle, Inyo County, California: California Div. Mines Spec. Rept. 51.

Hanks, H. G., 1884, Catalogue and description of the minerals of California: California State Min. Bur., 4th rept of the State Mineralogist, p. 63–398.

Hewett, D. F., 1931, Geology and ore deposits of the Goodsprings quadrangle, Nevada: U.S. Geol. Survey Prof. Paper 162, 172 p.

———1956, Geology and mineral resources of Ivanpah quadrangle, California and Nevada: U.S. Geol. Survey Prof. Paper 275, 172 p.

Hill, R. T., 1912, Marble deposits of the Inyo Mountains: Min. and Sci. Press, v. 105, p. 86–87.

Hopper, R. H., 1947, Geologic section from the Sierra Nevada to Death Valley, California: Geol. Soc. America Bull., v. 58, p. 393–432.

Hyatt, Alpheus, and Smith, J. P., 1905, The Triassic cephalopod genera of America: U.S. Geol. Survey Prof. Paper 40, 394 p.

Ingalls, W. R., 1908, Lead and zinc in the United States: New York, Hill Publishing Co. (McGraw-Hill), 368 p.

Kirk, Edwin, 1918, The stratigraphy of the Inyo Range, *in* Knopf, Adolph, A geologic reconnaissance of the Inyo Range and the eastern slope of the Sierra Nevada, California: U.S. Geol. Survey Prof, Paper 110, p. 19–48.

———1933, The Eureka quartzite of the Great Basin region: Am. Jour. Sci., 5th ser., v. 26, no. 151, p. 27–44.

Kneiss, G. H., 1941, Bonanza railroads: Stanford Univ., California, Stanford Univ. Press, 148 p.

Knopf, Adolph, 1914, Mineral resources of the Inyo and White mountains, California: U.S. Geol. Survey Bull. 540, p. 81–120.

———1918, A geologic reconnaissance of the Inyo Range and the eastern slope of the Sierra Nevada, California: U.S. Geol. Survey Prof. Paper 110, 130 p.

Lee, C. H., 1912, An intensive study of the water resources of a part of Owens Valley, California: U.S. Geol. Survey Water-Supply Paper 294, 135 p.

Lee, W. T., 1906, Geology and water resources of Owens Valley, California: U.S. Geol. Survey Water-Supply Paper 181, 28 p.

Loew, Oscar, 1876, Report on the geological and mineralogical character of southeastern California and adjacent regions, *in* Wheeler, G. M., 1876, Ann. Rept., U.S. Geog. Surveys W. 100th Mer. Rept., App. JJ, p. 62.

Longwell, C. R., and Dunbar, C. O., 1936, Problems of Pennsylvanian-Permian boundary in southern Nevada: Am. Assoc. Petroleum Geologists Bull., v. 20, no. 9, p. 1198–1207.

McAllister, J. F., 1952, Rocks and structure of the Quartz Spring area, northern Panamint Range, California: California Div. Mines Spec. Rept. 25, 38 p.

———1955, Geology of mineral deposits in the Ubehebe Peak quadrangle, Inyo County, California: California Div. Mines Spec. Rept. 42, 64 p.

———1956, Geology of the Ubehebe Peak quadrangle, California: U.S. Geol. Survey Geol. Quad. Map GQ–95.

Merriam, C. W., 1940, Devonian stratigraphy and paleontology of the Roberts Mountains region, Nevada: Geol. Soc. America Spec. Paper 25, 114 p.

———1947, Stratigraphy of the Cerro Gordo area, Inyo County, California [abs.]: Geol. Soc. America Bull., v. 58, p. 1255–1256.

Merriam, C. W., 1951, Silurian quartzite of the Inyo Mountains, California [abs.]: Geol. Soc. America Bull., v. 62, p. 150.

———1954, Rocks of Paleozoic age in southern California, in Chapter 3, Historical Geology: California State Div. Mines Bull. 170, p. 9–14.

Merriam, C. W., and Anderson, C. A., 1942, Reconnaissance survey of the Roberts Mountains, Nevada: Geol. Soc. America Bull., v. 53, p. 1675–1728.

Merriam, C. W., and Hall, W. E., 1957, Pennsylvanian and Permian rocks of the southern Inyo Mountains, California: U.S. Geol. Survey Bull. 1061–A, p. 1–15.

Merrill, G. P., 1903, Stones for building and decoration: New York, John Wiley and Sons, Inc., 3d ed., 551 p.

Murdoch, Joseph, and Webb, R. W., 1948, Minerals of California: California Div. Mines Bull. 136, p. 1–402.

Nadeau, R. A., 1948, City makers: Garden City, N.Y., Doubleday & Co., Inc., 270 p.

Nolan, T. B., 1943, The Basin and Range province in Utah, Nevada, and California: U.S. Geol. Survey Prof. Paper 197–D, p. 141–196.

Nolan, T. B., Merriam, C. W., and Williams, J. S., 1956, The stratigraphic section in the vicinity of Eureka, Nevada: U.S. Geol. Survey Prof. Paper 276, 77 p.

Norman, L. A., Jr., and Stewart, R. M., 1951, Mines and mineral resources of Inyo County: California Div. Mines, California Jour. Mines and Geology, v. 47, no. 1, p. 17–223.

Page, B. M., 1951, Talc deposits of steatite grade, Inyo County, California: California Div. Mines Spec. Rept. 8, 35 p.

Phleger, F. B., Jr., 1933, Notes on certain Ordovician faunas of the Inyo Mountains, California: Southern California Acad. Sci. Bull., v. 32, pt. 1, p. 1–21.

Raymond, R. W., 1873, Statistics of mines and mining in the States and Territories west of the Rocky Mountains for 1872: 5th Ann. Rept. U.S. Commissioner mining statistics, U.S. Treasury Dept., p. 1–550.

Rezak, Richard, 1959, New Silurian Dasycladaceae from the southwestern United States: Colorado School Mines Quart. Jour., v. 54, no. 1, p. 117–129.

Smith, J. P., 1901, The border line between Paleozoic and Mesozoic in western America: Jour. Geology, v. 9, no. 6, p. 512–521.

———1904, Comparative stratigraphy of the marine Trias of western America: California Acad. Sci. Proc., 3d ser., v. 1, p. 323–430.

———1914, The Middle Triassic marine invertebrate faunas of North America: U.S. Geol. Survey Prof. Paper 83, 254 p.

———1932, Lower Triassic ammonoids of North America: U.S. Geol. Survey Prof. Paper 167, 199 p.

Spencer, A. C., 1917, The geology and ore deposits of Ely, Nevada: U.S. Geol. Survey Prof. Paper 96, 189 p.

Stauffer, C. R., 1930, The Devonian of California: California Univ., Dept. Geol. Sci. Bull., v. 19, p. 81–118.

Tucker, W. B., and Sampson, R. J., 1938, Mineral resources of Inyo County: California Div. Mines, California Jour. Mines and Geology, v. 34, no. 4, p. 368–500.

Westgate, L. G., and Knopf, Adolph, 1932, Geology and ore deposits of the Pioche district, Nevada: U.S. Geol. Survey Prof. Paper 171, 79 p.

Youngquist, W. L., 1949, The cephalopod fauna of the White Pine shale of Nevada: Jour. Paleontology, v. 23, no. 3, p. 276–305; supp. no. 6, p. 613–616.

INDEX

[Where several page numbers appear, major references are in italic]

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Geology and Petrography of Volcanic Rocks of the Truk Islands, East Caroline Islands

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By J. T. STARK *and* R. L. HAY

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A structural, stratigraphic, and petrographic study of the rocks of a highly dissected, partly submerged inactive shield volcano

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CONTENTS

ILLUSTRATIONS

[Plates are in pocket]

TABLES

GEOLOGY AND PETROGRAPHY OF VOLCANIC ROCKS OF THE TRUK ISLANDS, EAST CAROLINE ISLANDS

J. T. STARK and R. L. HAY

ABSTRACT

The Truk Islands are a near-atoll in the North Pacific Ocean at about 7°20′ north latitude and 151° east longitude. They consist of 12 volcanic islands and many low coral reef islands in a lagoon approximately 30 by 40 miles, enclosed by a coral reef. The volcanic islands range from 5 by 2 miles to islands less than a quarter of a mile in diameter. Several peaks on the volcanic islands rise more than 1,000 feet; the highest altitude is 1,453 feet, on Tol. Dense, jungle vegetation covers slopes and crests of the volcanic islands and many of the reef islands.

The Truk Islands are remnants of a large shield volcano, now inactive, which has been partly submerged. Lava flows predominate although pyroclastic deposits are locally interbedded with the flows. The Truk volcano extended about 16,000 feet from the ocean floor to the surface. No evidence of crater walls now exists, but geologic evidence indicates a central crater once erupted large volumes of pyroclastic ejecta. Most of the lavas issued from fissure vents now represented by dikes and dike swarms.

The petrography of the volcanic rocks of Truk is relatively simple except for breccia blocks in the pyroclastic deposits of the central islands. The lava flows and dikes consist of olivine-rich basalt, melilite-nepheline and nepheline basalt, nepheline basanite, andesite, and trachyte. The breccias of central Udot and Eot consist of angular fragments of rock in a fine-grained tuff matrix of crystal and rock fragments. Andesite, trachyte, and basalt blocks predominate in the breccias. Phaneritic blocks of gabbro are locally common, and blocks of hornfelsic, recrystallized gabbro, basalt, and breccia contain veins of fine-grained monzonite and quartz monzonite. Several blocks of monzonite have been found. Gabbro blocks contain dikes of andesite and basalt and inclusions of recrystallized basalt. A small number of limestone xenoliths were found.

The lavas of the Truk volcano clearly represent the alkalic-olivine-basalt-trachyte association common in the Pacific Ocean basin east of the andesite line. The undersaturated lavas of Hawaii are similar to those of Truk, but hypersthene-bearing tholeiitic lavas of Hawaii have no visible counterpart on Truk, either as lavas or pyroclastic ejecta. Unlike Hawaii, quartz trachyte occurs on the central islands of Truk. Most lavas and dikes of the central islands have been hydrothermally altered to some extent. Secondary chlorite and albite partly or entirely replace primary mafic minerals and plagioclase, respectively, and quartz and pyrite have been introduced into some of the rock.

Ejecta from the central crater include a number of rock types not found otherwise at the surface and supply information about rock types and processes at depth in the interior of the volcano. Gabbro was probably emplaced within the volcano, possibly as a stock. The gabbro is undersaturated and fundamentally similar to the basalt flows at the surface and was probably derived from the same magma. Monzonite and quartz monzonite form veins in the gabbro blocks and occur as individual xenoliths. The monzonite crystallized from a hydrous magma, apparently at temperatures high enough to melt and assimilate adjacent gabbro. The quartz monzonite may be the hypabyssal equivalent of the quartz trachyte which assimilated some gabbroic material during intrusion.

Foraminifera in limestone fragments from the central crater indicate a late Tertiary *e* age with a slight possibility that they could be early Tertiary *f*. In terms of the standard time scale, these Foraminifera are very likely early Miocene. Consideration of several factors suggests that this limestone was deposited when the volcano had grown approximately to sea level, prior to the development of the subaerial shield volcano of large size. After the growth of the shield volcano, erosion dissected the cone, and several flows of nepheline-bearing lavas were extruded. After these last eruptions the volcano subsided sufficiently to submerge most of the dissected subaerial shield. A barrier reef subsided relative to sea level as the volcano subsided and formed a lagoon enclosing unsubmerged remnants of the Truk shield volcano.

INTRODUCTION

This study of the geology and petrography of the volcanic rocks of the Truk Islands is based upon fieldwork and petrographic examinations of specimens collected during a survey of the islands for the purpose of a military geology report, made as a part of the Pacific Geological Mapping Program of the U.S. Geological Survey and Corps of Engineers of the U.S. Army. Introductory material here has been summarized from the military geology report (Stark and others, 1958).

LOCATION

The Truk Islands are between north latitudes 7°08′ to 7°41′, and east longitudes 151°26′ and 152°2′. They lie 2,100 miles southeast of Tokyo, 3,450 miles west and slightly south of Pearl Harbor, and 1,000 miles northeast of New Guinea (fig. 1).

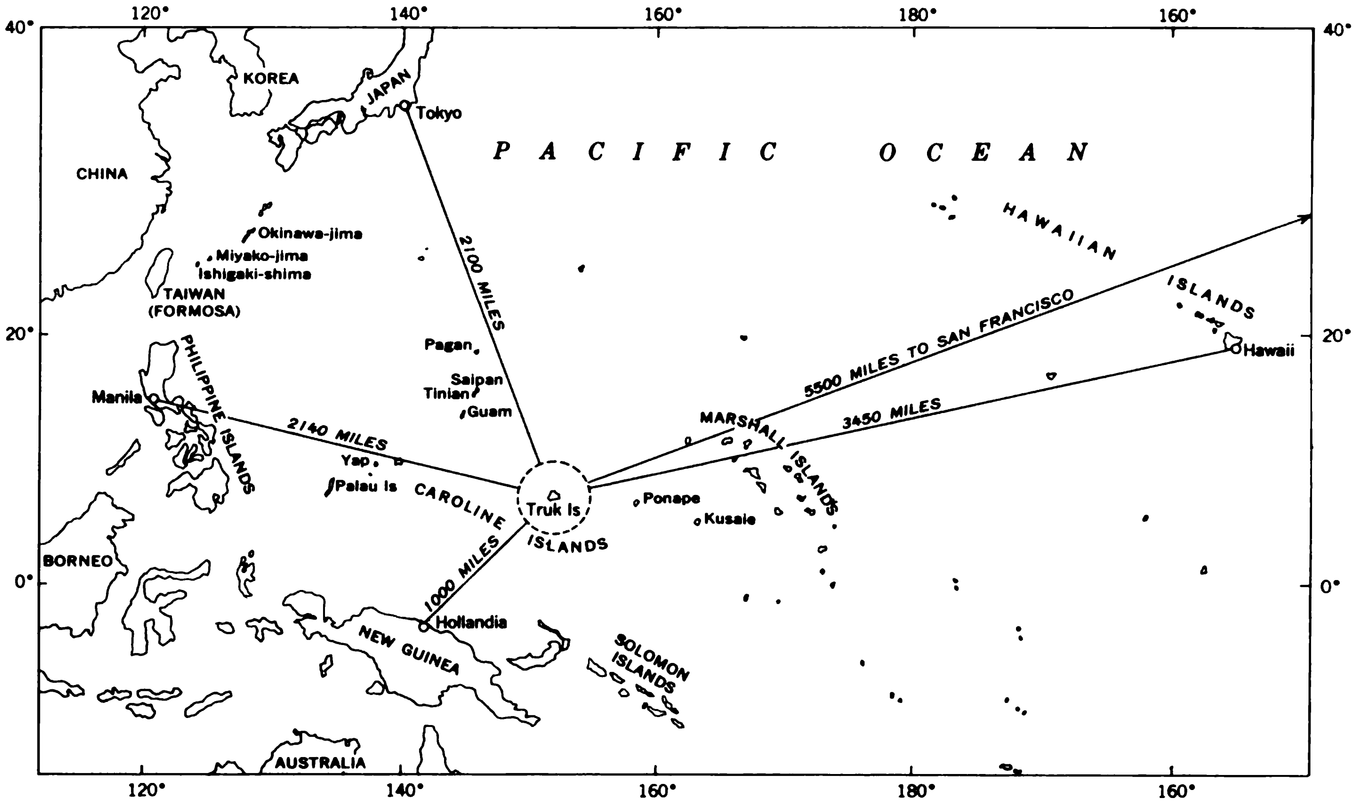

FIGURE 1.—Index map, showing location of Truk Islands.

TOPOGRAPHY AND DRAINAGE

Truk comprises small volcanic and reef islands in a lagoon, Truk Lagoon, formed by a coral reef (fig. 2). It has been termed a "near-atoll." The volcanic islands are separated more or less arbitrarily according to their location into an eastern group, including Moen, Uman, and Faneu (pl. 1); a central group, including Udot, Eot, and Eiol (pl. 1); and a western group, including Tol, Fala-beguets, and Ulalu (pl. 2). The high volcanic islands range in size from the two largest islands, Tol 5 by 2 miles and Moen 4 by 3 miles, to the two smallest islands, Eiol and Faneu, both less than one-quarter of a mile in diameter. The highest altitude is the crest of Mount Tumuital on Tol at 1,453 feet; Mount Tolomen on Dublon Island and Mount Teroken on Moen are both slightly more than 1,100 feet above sea level. Other prominent peaks, rising to nearly 1,000 feet, are on Fefan and Uman. The slopes in general average more than 30 percent, and slopes greater than 60 percent are common at higher altitudes.

Twenty-four low coral-reef islands are within the lagoon. The barrier reef is surmounted by 41 low coral islands. The largest is about 2 miles long and a quarter of a mile wide; most are much smaller. Altitudes on a few of these islands are as much as 8 feet, but most of the low islands are only 5 or 6 feet above mean sea level. Fringing reef flats surround most of the high and low islands.

The high volcanic islands are drained by many small streams during the rainy months, from April through December, but only a few streams persist throughout the dry period. The streams are not actively eroding the high islands, and little sediment is being carried to the lagoon.

Vegetation on the high volcanic islands is commonly more luxuriant than on the low reef islands. The upper slopes of such mountains as Mount Tumuital on Tol, Mount Teroken on Moen, and Mount Tolomen on Dublon Island are covered with dense forests of banyan and pandanus trees entwined with vines and shrubs. These forests were originally more extensive, but they have been replaced on the lower and middle slopes by breadfruit and coconut groves. A large percentage of the coconut groves, as well as additional forest areas, was cleared during the war because of the need for greatly increased food production and for lumber. These areas are now overgrown with grasses. Most of the low islands are covered with coconut palms, pandanus, and breadfruit trees.

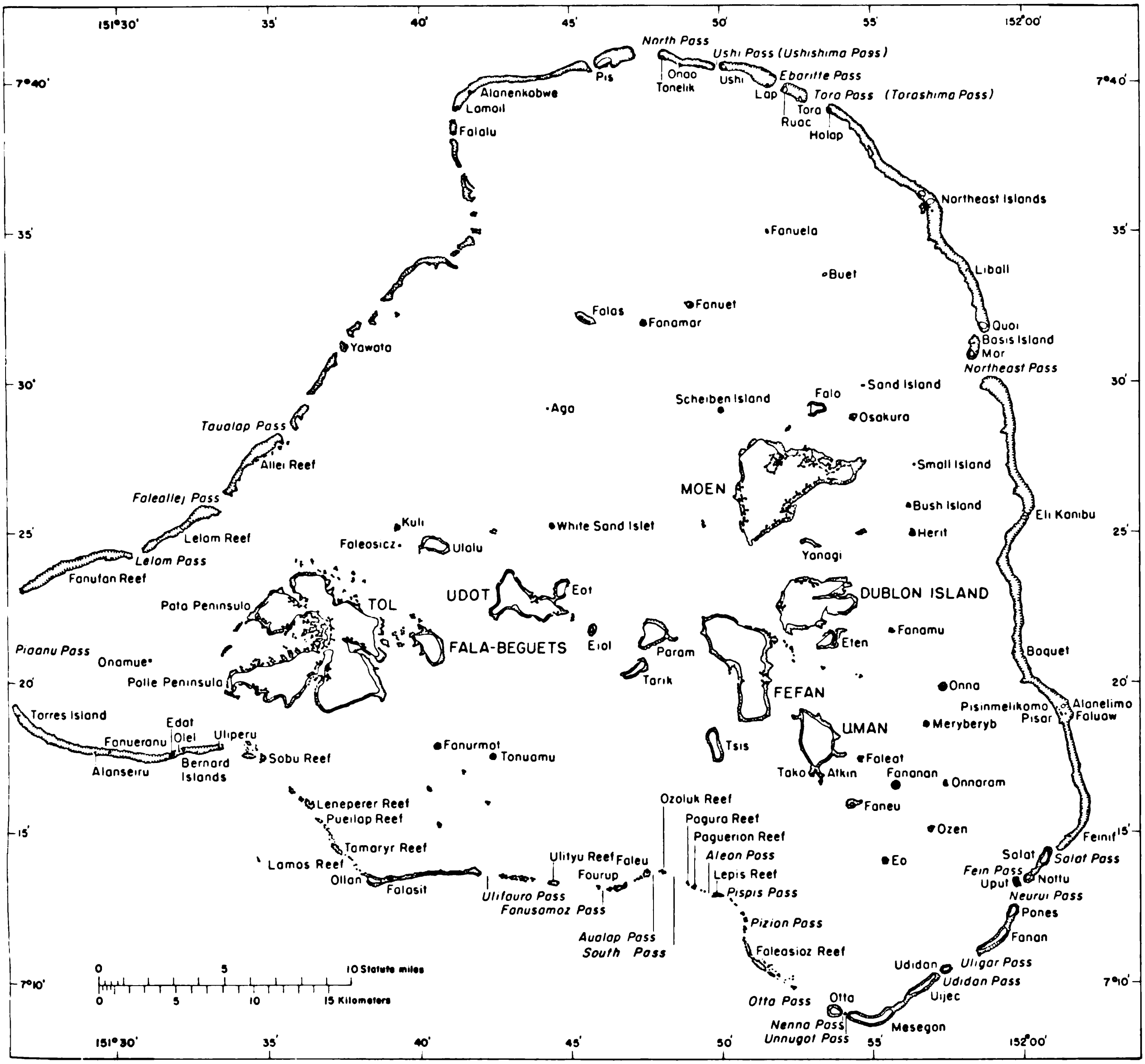

FIGURE 2.—Index map of Truk Islands.

ACKNOWLEDGMENTS

The field party was on the Truk Islands from early in October 1954 until the end of July 1955. Logistic support was furnished by the Corps of Engineers, U.S. Army. Fieldwork was supervised by C. G. Johnson, chief of the Pacific Geological Surveys section, Tokyo, on his several visits to the islands and throughout the progress of the work. Microscopic study of the rocks was started in Truk and completed in offices of the U.S. Geological Survey, Washington, D.C., and the Louisiana State University, Baton Rouge, La. Thanks are expressed to all members of the Truk field party, to W. C. Muller, district administrator for Truk, and to all members of the Trust Territory administration on Moen for their interest and helpful cooperation during the mapping of the islands. Dr. D. B. Stewart is thanked very much for X-ray study of a feldspar in several xenoliths from the islands.

GEOLOGY

The volcanic islands of the Truk Islands are the erosional remnants of a partly submerged large shield volcano, which has been long inactive. Although these remnants consist largely of lava flows, pyroclastic deposits are locally interbedded with the flows. The Truk volcano extends from the ocean floor, about 16,000 feet below to the water surface and was formerly at least several thousand feet above sea level.

Coral and reef rock now form a surrounding barrier, and fringing reefs border the islands of volcanic rock in the lagoon. Igneous rock of the old volcano almost certainly underlies the low reef islands within the lagoon and presumably forms the basement on which the coralline barrier reef was built. If so, the encircling barrier roughly marks the near-surface shape of the Truk volcanic mass.

At no place on any of the islands have traces of crater or caldera walls or remnants of a central volcanic vent been found. However, geologic evidence indicates that a central crater once erupted large volumes of pyroclastic ejecta, now consolidated as a volcanic breccia. Most of, if not all, the lavas very likely issued from fissure vents now represented by dikes and dike swarms. Many of the breccias were probably transported as mudflows. Some lavas and breccias were subjected to stream action and now form conglomerates, which consist largely of rounded boulders.

The lavas range in composition from olivine-rich basalts through andesites to trachytes and to extremely silica-deficient melilite-nepheline and nepheline basalts. Although the melilite-nepheline and nepheline basalt flows generally appear to have been the last to be erupted, the other types of flows, olivine basalt, andesite, and sodic trachyte, do not occur in any apparent sequence.

A detailed stratigraphic sequence of lava flows and pyroclastic breccia is difficult to establish on the larger islands because of the wide extent of dense vegetation and the thick soil cover that separate exposures of bare rock. Outcrops can be traced only for short distances horizontally or along the strike of the beds. A sequence of flows and breccias occurring in one vertical section may differ radically from the sequence in a section only a hundred yards away. Part of this is due to tongue-like projections of lava and to valley-filling flows and pryoclastic deposits that, in general, conform to the surface slopes existing at the time of extrusion. A further complicating factor in the western group of islands is the strong possibility that most of the flows issued from many separate fissures rather than from one or two well-developed craters. The general similarity of the flows mineralogically and texturally and the absence of distinguishing criteria, with the possible exception of nepheline basalt on Tol and Ulalu, make correlation of individual flows between islands impossible or extremely uncertain. There are, however, suggestive differences between groups of islands, which are discussed under headings of eastern, central, and western island groups.

CLASSIFICATION

In this report rocks are basically classified by color index of the rock, excluding phenocrysts; that is, by the sum of the normative femic minerals (wo+en+fs+fo+fa+mt+hm+il) as calculated from the chemical analyses of nonporphyritic rocks. Basalts will be considered those rocks having a color index greater than 37, and andesites are those rocks having an index between 10 and 37, following the usage of Kuno (1950, p. 958). Trachytes are those rocks consisting largely of alkali feldspar and having a color index less than 10. Rocks not chemically analyzed were classified by estimating the sum of the modal mafic minerals and comparing thin sections with those of texturally similar analyzed rocks. The basalts and andesites intergrade so completely that without chemical analyses about 10 percent of the borderline rocks may arbitrarily be placed in either category. Trachytes do not seem to intergrade with andesites, from which they generally can be separated without difficulty.

The lava flows of the Truk Islands are mineralogically rather simple. Most are undersaturated lavas consisting almost entirely of plagioclase, monoclinic pyroxene, olivine (or pseudomorphs after olivine), anorthoclase, and magnetite. Basalt is the dominant rock; of the nearly 500 specimens collected, 70 percent are basalt and 25 percent are andesite. The remaining 5 percent are sodic trachyte, nepheline basalt, melilite-nepheline basalt, and vitrophyric basalt containing normative nepheline. Nearly all the basalt flows contain more than 5 percent modal olivine and are olivine basalts in the classification of Macdonald (1949, p. 1544); a few of the andesites contain more than 5 percent olivine and are olivine andesites. The proportion of olivine in the finer grained rocks cannot be determined without microscopic study, and rocks will be discussed in the following text as basalt and andesite unless olivine is conspicuously abundant in the hand specimen. None of the olivine basalts contain sufficient mafic minerals (that is, 70 percent) to be classified as the picrite basalt of Macdonald (1949, p. 1544).

The basalts and andesites are almost uniformly dark gray. Both porphyritic and nonporphyritic basalt flows are common, but most of the andesites are nonporphyritic. Many basalts, particularly the finer gained ones, have a platy fracture that is due to planar orientation of plagioclase laths. Most of the andesites have a similar platy fracture and are impossible to distinguish in the field from the fine-grained platy basalts. The trachytes are medium to light gray and generally have a well-developed platy fracture that superficially resembles metamorphic foliation. A trachyte flow on Moen was, in fact, earlier misidentified as schist (Bridge, 1948, p. 217).

EASTERN ISLANDS

MOEN

Moen, largest of the eastern islands, is approximately 7.19 square miles in area (pl. 1). It is a mountainous mass of lava and indurated pyroclastic rocks bordered generally by unconsolidated organic marsh sediments, filled areas (reclaimed land), beach deposits, and mangrove swamps. The summit of Mount Teroken, the highest altitude on Moen, is approximately 1,223 feet.

The volcanic rocks consist of gently dipping lava flows and volcanic conglomerates. Locally, the volcanic conglomerates contain abundant angular debris and may be termed "volcanic breccia." The proportion of breccias and conglomerates to flows in the entire sequence cannot be estimated with accuracy because of the vegetation and soil cover, but the breccias and conglomerates probably form between 5 and 10 percent of the whole.

The flows consist largely of compact columnar-jointed lavas. Most are 35 to 100 feet in thickness and average an estimated 55 feet. The two thickest measure 150 to 200 feet. In general, flows can be traced only for short distances in continuous outcrop, but in southern and central Moen several flows are exposed almost continuously for slightly more than a mile. Olivine basalt is the predominant rock type, forming about 60 percent of the lavas. Moderately consolidated autobrecciated lava, generally from 3 to 10 feet thick, occurs at the top of many flows and less commonly at the base. The complete stratigraphic sequence is nowhere well exposed. The best exposures are on the south slope of Witipon, where nonprophyritic andesite is the only rock exposed from sea level to approximately 600 feet altitude. This andesite sequence is capped by a flow of sodic trachyte that measures at least 150 feet in thickness. Exposures on the jungle-covered slopes of Winifourer are basalt flows overlain by a thick sequence of nonporphyritic and slightly porphyritic andesite flows, which are capped in turn by flows of andesite and extremely porphyritic olivine basalt. As outcrops are separated by vegetation cover, it is not possible in many places to determine whether one or several flows are represented.

The sequence on the north spur of Mount Teroken differs from the exposures on Winifourer, less than half a mile away. Coarse conglomerate and breccia are interbedded in the lower part of the section with basalt lavas. Trachytic, andesitic, and basaltic debris occurs in these beds. They are overlain by several nonporphyritic andesite flows, which are in turn overlain by approximately 600 feet of extremely porphyritic olivine basalt flows, similar to those capping Winifourer. The porphyritic lavas on Mount Teroken extend downward from the crest to an altitude of approximately 600 feet; the base of those on Winifourer is between 900 and 1,000 feet. This difference in altitude between outcrops of similar lavas suggests that from 300 to 400 feet of andesite was eroded from the Mount Teroken area prior to the extrusion of the uppermost lava flows.

Three basalt flows, totaling at least 160 feet (tops are eroded and base not exposed), appear to be filling a valley cut in pyroclastic breccia in southwestern Moen. The slope of the valley wall ranges from 30° to 45°. The rarity of faulting elsewhere suggests that this contact is due to erosion and valley fill rather than to structural displacement.

Small unconformities occur between lava flows and breccia in a number of places. For example, on the north side of Mount Teroken at an altitude of approximately 350 feet, a lenticular body of breccia 20 to 30 feet thick fills a steep-walled narrow valley cut into an andesite flow. None of these unconformities are angular, rather they represent disconformities where valleys were eroded by streams between eruptions.

Volcanic sedimentary deposits are widespread on Moen. They consist predominantly of unstratified pyroclastic breccia in which angular blocks are enclosed in a finer grained unsorted matrix of volcanic detritus. In places, the breccias grade into conglomerates that contain cobbles aand boulders showing the effect of rounding by water. The thickest deposits are in the northeast and central parts of the island. On the northeast slope of Tonaachau, the cobble and boulder conglomerate is 50 feet thick and pinches out laterally to the east and south. In the valley of the Wichen River, the breccia is 300 feet in thickness. In general, the breccia is more thoroughly indurated than the conglomerate. Many beds are only from 2 to 8 feet thick. Fine tuffaceous material is estimated to form between 20 and 50 percent of the beds. The coarse fragments, ranging from a few inches to 6 feet in diameter, are composed of all types of the Moen lavas: basalts, andesites, and trachytes.

The upper and lower contacts of the flows are nearly horizontal, as far as can be determined from the small areas of exposures, except for the few irregularities due to channeling and the uneveness that might be expected from varying thickness of the flows. An attempt was made to measure the altitude of the flows exposed in steep escarpments with aneroid readings along the outcrops. A general dip to the northeast and possible east is indicated. The sodic trachyte flow capping Witipon appears to strike northwest and dips approximately 2½° E. One extremely porphyritic olivine basalt flow was mapped for a distance of 1½ miles on the north side of Moen with strike of N. 45° W., and dips of 3° to 3½° NE. The andesite series of flows exposed on the southern slopes of Mount Teroken and Witipon dips 2° to 2½° E. Based on correlation of the conglomerate exposed at an altitude of approximately 500 feet on Tonaachau with the thick breccia unit 1 mile to the south, a solution of the three-point problem suggests that the top of the conglomerate strikes northward and dips 3½° E. As the breccia may not be equivalent to the conglomerate, the inferred northly strike of the pyroclastic unit is probably not as reliable as the northwest strike measured on the porphyritic olivine basalt flow.

No conclusive evidence of faulting was found on Moen. Either faulting or filling of a steep-walled valley is, however, indicated by relations on the northern corner of Moen. A 160-foot flow of oligoclase andesite, which lies adjacent to the airfield on the north corner of Moen, forms a narrow band terminating abruptly against a flow of olivine basalt along the southern margin of the andesite outcrops. This abrupt termination may represent faulting or the lateral margin of a valley-filling flow. The unusually great thickness of this flow, the thickest andesite flow measured in Truk, can be explained by origin of the flow as a valley filling. However, fault displacement along the contact of the basalt and andesite is a possible explanation of lithologically similar, coarse volcanic conglomerate north of the contact at an altitude of about 6 feet and on the hillside south of the contact at an altitude of about 350 feet. Present data are inadequate to demonstrate faulting or to prove cut-and-fill origin of the steep contact. Similarly, three basalt flows on the southwest part of Moen lie against volcanic breccia along a contact having a dip of 30° to 45°. As volcanic breccia has been clearly observed elsewhere to fill a steep-walled canyon cut in a lava flow, it is likely that the flows in question fill a steep-walled valley cut in breccia. The possibility of faulting, however, cannot be eliminated at this time.

The source of the Moen lavas can only be inferred as no dikes or vents have been found on the island. The general northeast dip suggests a source to the southwest. This suggestion is supported by the northeast trend of the valley-filling lava flows at the southwest end of Moen. The inferred unconformity between nonporphyritic andesite and porphyritic basalt on Mount Teroken and Winifourer indicates a southern source for the upper porphyritic lavas.

FALO

Falo, a small island 1¾ miles north of Moen, is an upland plateau of approximately 140 feet altitude, consisting of several olivine basalt flows surrounded by patches of fresh-water marsh deposits and calcareous beach sands. The total surface area is 0.13 square mile (pl. 1).

The olivine basalt flows are well exposed in cliffs on the north side of Falo and in separate outcrops on top of the plateau. They range in texture from massive and even grained to porphyritic with abundant olivine and pyroxene phenocrysts.

YANAGI

Yanagi, 500 feet long and 250 feet wide, lies approximately half way between Moen and Dublon Island (pl. 1). It rises about 45 feet above its fringing reef and is composed entirely of coarse pyroclastic breccia. The fragments are andesite and olivine basalt similar in composition to the flows on Moen to the north and Dublon Island to the south. No indication of the attitude of the breccia beds on Yanagi was observed.

DUBLON ISLAND

Dublon Island, third largest of the islands in the eastern group (pl. 1), consists of lava flows and pyroclastic deposits. Detailed traverses across Dublon Island's four prominent uplands—Mount Tolomen and Foukenau and the northeast and southeast peninsulas—show no agreement in sequence of flows and interbedded pyroclastic deposits that might suggest specific correlations.

There are only isolated outcrops of olivine basalt and andesite in the western half of the island, and more than three-fourths of the area is mapped as undivided volcanic rocks (pl. 1). On Mount Tolomen the single mapped flow of andesite appears to dip about 3½° E. Olivine basalt and andesite form most of the northeast and southeast peninsulas; however, a nepheline basalt flow is interbedded with olivine basalt at the southeast end of the northeast peninsula and is the only nepheline basalt identified in islands of the eastern group. This flow on Dublon Island is of especial interest in its occurrence as it is interbedded near the base of a series of olivine basalt and andesite flows, in contrast to the nepheline basalt flows on Tol, which occur as the uppermost and last phase of volcanic activity recorded in the Truk Islands.

A thick porphyritic flow of olivine basalt underlies the upper surface of the northeast peninsula. This flow appears to strike N. 30° W.; the angle of dip varies but averages 4° E.

Fragmental deposits were mapped at several places on Dublon Island. A bed of conglomerate from 8 to 10 feet thick, penetrated by manmade caves, lies beneath the two flows capping Foukenau. The conglomerate is poorly sorted, and the coarsest debris is about 4 feet in diameter. Locally, the fragments have been rounded by stream action. Lapilli tuff crops out in a 12-foot-high roadcut, stratigraphically about 100 feet below the conglomerate, and a 4-inch-thick bed of well-laminated fine-grained tuff is interbedded with the lapilli tuff. Unstratified, unsorted breccia forms several lenticular beds in the southwest part of Dublon Island, the thickest of which is at least 75 feet thick. Blocks in the breccia are of olivine basalt and andesite of various textures, accompanied locally by blocks of laminated tuff. There is no obvious indication of a nearby vent, and these breccias may have been deposited by a mudflow.

A dike of andesite cuts an olivine basalt flow in a roadside outcrop on the west side of Dublon Island. The dike is 15 feet wide, strikes N. 40° W., and dips 52° S. It is one of the very few dike exposures in the eastern islands.

Olivine basalt is the predominant rock type on the islands, averaging about 65 percent of the flows; it is nearly twice as common as the andesite. Nepheline basalt forms far less than 1 percent of the island.

ETEN

Eten, 3,500 feet long and 1,500 feet wide, is a small northeast-trending island just half a mile south of Dublon Island (pl. 1). Reclaimed land forms a large part of the island, but volcanic rocks compose a hill about 180 feet high on the south side, and a small knoll, 15 to 20 feet high, at the east end of the island.

The hill consists of olivine basalt flows, conglomerate, and breccia. The conglomerate is made up largely of basalt boulders from 1 to 4 feet in diameter. It crops out at the southwest corner of the hill, where it forms the basal unit. This conglomerate is separated by a steep-walled erosional unconformity from a valley-filling flow of olivine basalt ranging in thickness from 50 feet to about 5 feet where it overlaps the conglomerate. Rudely stratified volcanic breccia and lapilli tuff with interfingering flows of olivine basalt, about 130 feet thick, form the upper parts of the hill. Complexly contorted lower pyroclastic beds and lava flows, overlain by pyroclastic breccias and lapilli tuff that are nearly horizontal and undisturbed, suggest that the lower beds were deformed by a mudflow prior to deposition of the undisturbed beds.

The volcanic rocks of Eten appear to dip gently to the north or northeast. A 20-foot-thick flow, cropping out for approximately 300 feet on the north-central side of the large hill, dips 5° NE. The thinness of the lava flows and the coarseness of the breccia constituents suggest a nearby source. The lack of water stratification in the pyroclastic rocks indicates an eruptive rather than a detrital origin. Their thickness also is suggestive of a local source of eruption. The northeasterly dips of one flow and the pinching out of the lower olivine basalt flow toward the west and the northwest, that is, against the west side of a north-south valley, argue for a local source to the south or southwest. The probable major center of eruption for the eastern group of islands is shown in figure 3. All flows on Eten are olivine basalt.

FEFAN

Fefan consists of a mountainous ridge with four prominent peaks (pl. 1), a mangrove swamp that fringes most of the volcanic upland, and locally, a few areas of lowland marsh and sand. Porphyritic olivine basalt greatly predominates over andesite flows in ratio of about 9 to 1. Maximum thickness of the lava flows is best exposed on the northwest side of Mount Iron, where there is 1,000 feet of interbedded basalt and andesite. The thickest single flow observed is porphyritic olivine basalt, which forms a cliff 80 feet high.

Pyroclastic deposits are rare on the island, having been found only on the middle west slope of Witunumo. Here a 1-foot-thick bed of volcanic tuff rests on the brecciated tops of the undivided flows, and grades upward into 8 feet of pyroclastic breccia containing angular blocks as much as 4 feet in length.

The attitude of the Fefan flows shows a persistent dip to the south. The outcrops of slightly porphyritic basalt on the east and west sides of the central part of the island strike eastward and dip from 1° to 1½° S. The thick porphyritic olivine basalt flow on the northwest corner of the island apparently dips more steeply southward.

PARAM

Param consists of deeply weathered lava flows bordered by fresh-water marsh deposits and calcareous beach sands (pl. 1). On the southern side of the island, reclaimed land was used by the Japanese as an airfield. Scattered exposures of fine-grained andesite and porphyritic olivine basalt can be seen in roadcuts and in small excavations on the north side of the island. Four flows are exposed on the east end of the island in borrow pits, up to an altitude of about 140 feet. At the base of the sequence is weathered porphyritic lava, probably olivine basalt. Slightly porphyritic platy basalt crops out at altitudes between 40 and 50 feet and is overlain by 40 feet of weathered porphyritic olivine basalt. No outcrops are exposed above the olivine ba-

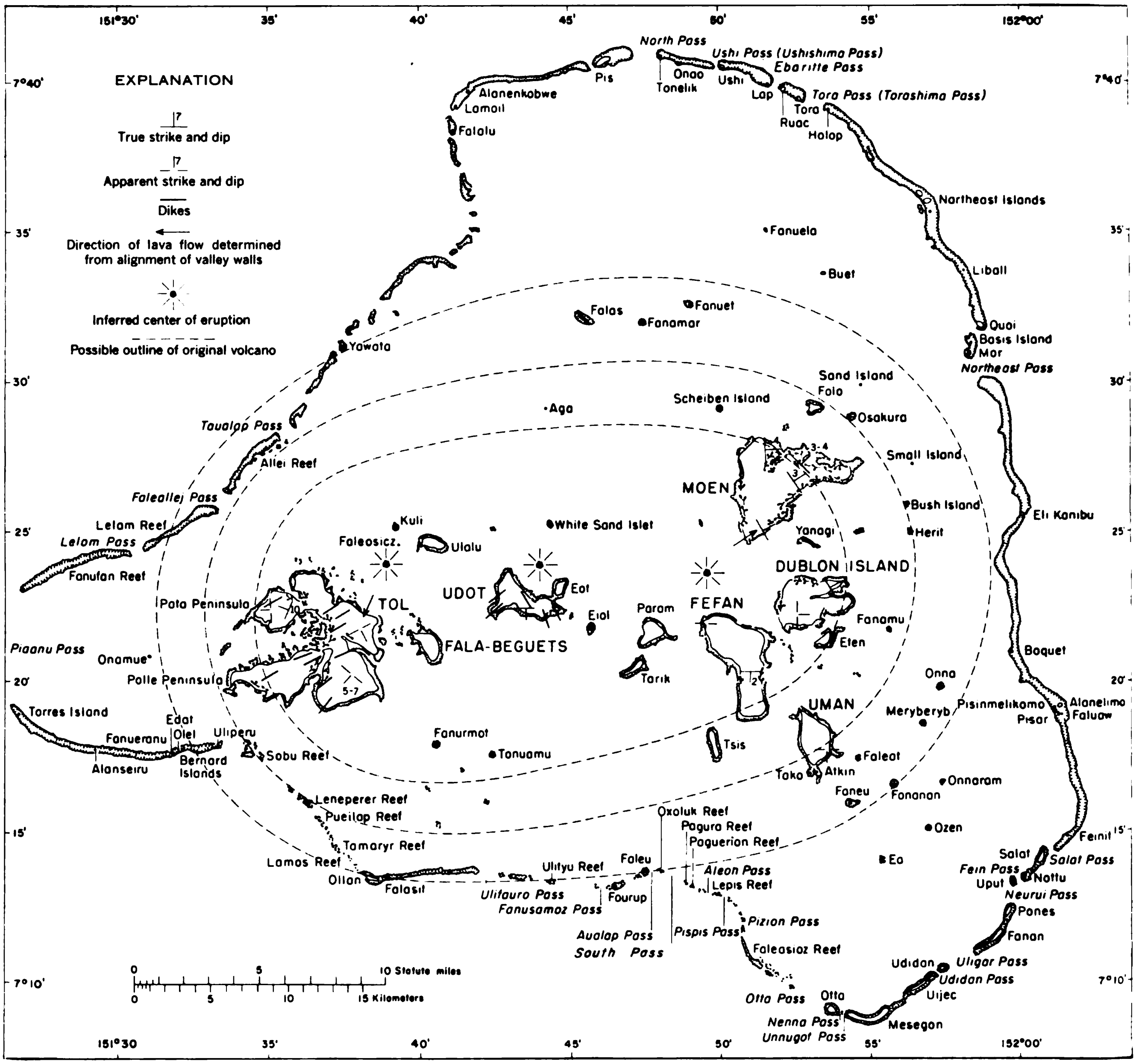

FIGURE 3.—Probable centers of eruption, Truk Islands, and possible outlines of original volcano.

salt to the highest point on the island, about 180 feet; however, abundant float blocks of fine-grained andesite above the olivine basalt suggest that an andesite flow almost certainly formed the uppermost unit on Param.

Rock types on Param are andesite and olivine basalt which occur in almost equal amounts.

TARIK

The surface slopes and crests of the eastern, central, and western peaks of Tarik are covered with loose blocks of olivine basalt. Although bedrock exposures are rare (pl. 1), two outcrops are exposed near sea level in the east-central part of the island: a highly weathere flow crops out at the water's edge, and 10 feet upslop is a small ledge of fine-grained olivine basalt. The to and bottom of the flows are not exposed, and erosio of the water-level outcrop obscures any indication o dip. No andesite was found on the island.

TSIS

Tsis, an island southwest of Fefan, is composed en tirely of porphyritic olivine basalt flows and is fringe on the north by fresh-water swamp deposits and calcare ous beach sands. Exposures are not continuous enoug o map separate flows. The apparent direction of dip of one small outcrop is S. 25° E.

UMAN, TAKO, AND ATKIN

Uman is a conical mass of volcanic rock rising to a central peak about 850 feet above sea level, with several smaller peaks around the lower slopes (pl. 1). Only flows of olivine basalt were observed in outcrop. One bed of volcanic breccia, 4 feet thick, is the only pyroclastic rock exposed. A thick olivine basalt flow on the middle eastern slope of Uman appears from topographic relations to be dipping very gently to the south. A fault with a vertical displacement of 5 to 10 feet on the southeast side of Uman is suggested by an unusually straight valley with a difference in altitude of bench-land surfaces on either side.

Tako and Atkin are small islets of volcanic rock, 900 feet south of Uman. Both are surrounded by coral reefs that are continuous with the fringing reef that surrounds Uman. Tako and Atkin are composed of olivine basalt flows that show well-developed columnar jointing. The flows are believed to be continuous, beneath the reef rock, with the olivine basalt flows of Uman.

FANEU

Faneu is a nearly circular island 2 miles south-southeast of Uman (pl. 1). It is approximately 750 feet in diameter and consists of a hill of volcanic rock 75 feet high, fringed by a sand beach and older beach deposits. Brecciated olivine basalt is exposed several feet above sea level in the base of an old Japanese torpedo cave on the south side of the island. This autoclastic breccia is overlain by a columnar-jointed olivine basalt flow. The lowermost 5 to 10 feet are sparsely porphyritic and grade upward into highly porphyritic rock which extends 30 to 40 feet higher. The hill is capped by sparsely porphyritic lava that may be the same flow.

CENTRAL ISLANDS

Numerous dikes and predominance of pyroclastic breccia over lava flows characterize the central islands of Udot, Eot, and Eiol (pl. 1). The volcanic rocks here differ significantly from those of other Truk islands and deserve detailed description.

PYROCLASTIC DEPOSITS

Volcanic breccia is the most abundant type of pyroclastic rock. In places the tuff matrix (that is, fragments less than 4 mm in diameter) is sufficiently abundant to form tuff-breccia facies of the volcanic breccia. Most coarse fragments in a few beds are less than 3.2 cm in diameter, and the rock is termed "lapilli tuff." All the pyroclastic deposits consist of angular to subangular, rarely rounded fragments of rock in a finer grained tuff matrix. Generally the blocks are between 1 and 12 inches in diameter, but blocks as long as 6 feet have been found. Some breccias consist almost entirely of blocks from 1 to 3 feet in diameter. The tuff matrix ranges from 5 to 75 percent, but generally averages about 25 percent. The pyroclastic deposits are uniformly well indurated and break across fragments and matrix alike. Nearly all the breccia is unstratified, and it commonly forms smooth, rounded cliffs as much as 50 feet high. The breccia usually weathers into blocks and boulders, some 20 feet in diameter, that locally are difficult to distinguish from bedrock. Boulders of breccia as much as 50 feet long occur along the south-central seacoast of Udot.

Fragments in the pyroclastic breccia include a wide range of rock types. Both porphyritic and nonporphyritic andesite, basalt, and trachyte are common. Many blocks are either vesicular or amygdaloidal. A small percentage of the trachyte blocks lack flow texture and are somewhat coarser than the trachyte forming dikes, flows, and autoclastic breccias. Also, some blocks of andesite and basalt are texturally dissimilar to the andesite and the basalt of dikes and flows exposed in the central islands. Blocks of fine- to coarse-grained gabbro are widely distributed in the breccia on Udot. Individual crystals in the gabbro are generally 1/8 to 1/4 inch in diameter, but 1/2-inch crystals characterize pegmatitic zones. Plagioclase forms between 50 and 80 percent of the gabbro, which is commonly pale gray. Many of the gabbro blocks contain veins of coarse-grained latite and quartz latite. A few of the gabbro blocks contain dikes of andesite and basalt. Other blocks contain inclusions of recrystallized volcanic rock. Blocks of hornfelsic, recrystallized gabbro, basalt, and volcanic breccia have also been found in the breccia. Several small blocks of quartz monzonite and fragments of limestone were collected from the breccia on Eot and Udot.

LAVA FLOWS AND AUTOCLASTIC BRECCIAS

Lava flows are widely distributed on Udot and Eot, and one flow occurs on Eiol. Andesite and basalt flows are about equally common, and a few trachyte flows are present. Single lava flows do not form prominent scarps as in the eastern islands, and although flow thicknesses could not be accurately determined, most flows are probably 20 feet thick or somewhat less. In a number of places the lava is intermixed with pyroclastic breccia in irregular tongues and stringers and does not form well-defined flows.

Autoclastic breccia is abundant and widespread on Udot and Eot. Some is vesicular and occurs at the top and base of flows as on the other islands; most, however, forms entire flows that consist largely of angular fragments of nonvesicular lava in a lava matrix. Angular fragments range from a fraction of an inch to more than 6 feet in diameter but most commonly are several inches in diameter. The matrix enclosing the angular fragments is generally nonvesicular lava, but a well-indurated tuff matrix occurs in places. Several autoclastic breccias grade, owing to the increase of tuff matrix and the addition of accessory volcanic blocks, into breccia which appears truly pyroclastic in all respects.

DIKES

The central islands are characterized by numerous dikes, generally between 1 and 5 feet thick. Trachyte and andesite dikes are about equally abundant; basalt dikes are less common. Texture ranges from nonporphyritic to extremely porphyritic. Pyrite is commonly disseminated throughout the dikes as small grains and forms coatings on fracture surfaces. Most dikes have a general uniform thickness and attitude, but some are irregular in shape and vary both in strike and dip. Small dikes, about an inch thick, commonly branch from irregular-shaped dikes and penetrate the country rock for distances of several feet. Multiple dikes, side by side, are exposed in several places. Where two dikes of different attitude intersect, the older dike is commonly irregular in width and sinuous in strike; the younger dike is generally uniform in trend and thickness. Dips of dikes range from 10° to vertical, and strikes represent all points of the compass. Statistically, however, the dikes form a radial pattern having a projected center north of Udot and west of Eot (fig. 3).

UDOT

Udot consists of three major upland segments connected by low narrow divides and is widely fringed by fresh-water swamps, mangrove swamps, and calcareous beach deposits. The three uplands are at altitudes of about 439, 472, and 793 feet, from east to west, and several small peaks rise from the lower slopes of the upland areas. The large conical mass forming the western, highest upland segment is composed largely of andesite flows, basaltic flows, and autoclastic breccias. Elsewhere, breccia and related pyroclastic deposits predominate over the lava flows and autoclastic breccias.

The proportion of different types of blocks in the pyroclastic breccias varies throughout Udot but fragments of andesite generally predominate over other types although trachyte is more abundant in places. Vesicular and amygdaloidal fragments are widespread, and scoriaceous fragments are abundant in some beds. Gabbro fragments have been found locally in breccia from sea level to altitudes of about 750 feet and from the east to the west end of Udot, but much breccia contains no gabbro. The gabbro fragments in most beds range from ¼ to 3 inches in diameter. Gabbro fragments are largest and most abundant in a bed in central Udot where they locally form as much as 10 percent of the breccia. Many gabbro blocks in this bed are a foot in diameter, and a few have average diameters of as much as 2 feet. The gabbro-rich breccia forms an 80-foot-thick bed on the south side of the island but thins northward, pinching out at one place. Blocks of recrystallized basalt and quartz monzonite were found only in the bed richest in gabbro blocks.

Several poorly exposed lava flows have diverse dips of as much as 30° over outcrops of several tens of feet. A small exposure of rudely stratified tuff beds on the west side of the eastern segment, at approximately 400 feet altitude, appears to be nearly horizontal. Much more reliable is the attitude of the bed of gabbro-rich breccia on the central segment; this unit strikes east and dips from 5° to 10° S.

Dikes are numerous in all parts of Udot. They differ considerably in attitude; but dips are generally steep, and most of the strikes fit into the radial orientation pattern with a projected center a mile northeast of east-central Udot (fig. 4). The dikes average approximately 40 percent andesite, 40 percent trachyte, and 20 percent basalt.

EOT

Eot is a small island one-quarter of a mile north of the east end of Udot. It is approximately 3,000 feet long and 1,500 feet wide and trends northward. Fresh-water swamps and calcareous sands fringe parts of the west and east sides of the island.

Limestone at sea level and 4 feet above sea level forms a thin plaster on pyroclastic bedrock at the north point of Udot. Both coral and rounded, waterworn volcanic detritus are cemented by fine-grained calcite. This limestone may be a remnant of a coral reef that formed during a period of slightly higher sea level.

The upland of Eot consists of pyroclastic breccia and one outcrop of an andesite lava flow, on the west-central point of the island. The pyroclastic breccia contains a few small fragments of light-gray and reddish-brown tuffaceous limestone having abundant organic detritus, including Foraminifera of Tertiary age (W. S. Cole, written communication, July 14, 1955).

The breccia and lava flows are cut by many dikes as on Udot. They differ in attitude, but are generally steeply dipping, and their strikes conform to the same radial pattern exhibited by the dikes on Udot. The dikes are of basalt, trachyte, and possibly andesite.

EIOL

Eiol is about 1,200 feet long and 600 feet wide and trends northward (pl. 1). The island consists of a small hill of volcanic rock approximately 90 feet high, bordered by calcareous beach sand at altitudes of less than 5 feet. The hill is composed of breccia overlain by 15 to 20 feet of fine-grained andesite, probably representing a single flow. The breccia is moderately well indurated where unweathered and consists largely of angular fragments of lava ranging from less than an inch to 15 inches in diameter. The tuff matrix forms from 20 to 50 percent of the breccia.

WESTERN ISLANDS

TOL

Tol is an island formed by four upland blocks of volcanic rock, separated by deep embayments (pl. 2), two of which are joined by narrow channels, which are crossed by a wooden bridge and a rock causeway. Mangrove swamps are more extensively developed around Tol than around any other of the Truk Islands, and muck and peat deposits occur in isolated patches between mangrove swamps and the base of the uplands. Calcareous beach sands form a few narrow strips along the north and south shores of the island. The highest altitude in Truk Lagoon is on southeast Tol where Mount Tumuital rises to 1,453 feet.

The geology of the Tol uplands differs from that of both the eastern and the central islands. Tol consists almost entirely of olivine basalt and andesite flows that are cut by steeply dipping fractures and dikes whose dominant strike is northeast. The flows are generally thinner and more vesicular than those of the eastern islands. This series of flows is unconformably overlain on northeast Tol by a thick unfractured flow of melilite-nepheline basalt and one of nepheline basalt.

Most olivine basalt and andesite flows are between 20 and 60 feet in thickness, but flows as thin as 2 feet have been observed. The tops and bottoms of flows are commonly characterized by vesicular texture and scoriaceous autoclastic breccia. A high proportion of the flows is vesicular or amygdaloidal throughout, and many flows consist largely of vesicular and amygdaloidal autoclastic breccia. Owing to the relative thinness of the flows, the lack of columnar jointing, and possibly, the less resistant character of the porous lavas, single lava flows do not make prominent scarps such as are found in the eastern islands.

Plagioclase, augite, and olivine are commonly recognizable as phenocrysts. The basaltic and andesitic flows are generally very similar in appearance, and field distinctions between them are difficult. Except for the plateaulike summits on northeast Tol, exposures are so limited that no flows were mapped singly. Pyroclastic beds are rare, volcanic breccia having been found only in one place.

Dikes have been observed on all the volcanic uplands and are undoubtedly much more abundant than shown on the geologic map (pl. 2). The most prominent exposures are on the southeast half of northeast Tol. Andesite dikes predominate, but olivine basalt dikes are common, and several sodic trachyte dikes were also found. The width of the dikes ranges from 15 inches to 8 feet. Dike swarms show separate intrusions side by side with a total thickness of as much as 10 feet. The dikes are vertical or nearly so and show a uniform strike; the most prevalent direction is N. 60° E. Fractures in the lavas paralleling the dikes are conspicuous in most outcrops in which dikes are present.

It is very likely that these dikes were feeders for lavas of the western islands. The northeast trend of the fractures and the dikes is probably responsible for the dominant northeast orientation of valleys and scarps on northwest and southwest Tol. Faulting may have been an additional factor in controlling valley trends, but fault displacement was nowhere observed along the fractures.

Two dikes, one 3 and one 12 inches thick, may have originated in a way different from the thicker dikes previously described. These dikes of olivine basalt cut the basal part of an andesite flow. The olivine basalt is lithologically similar to the underlying basalt flow and may represent lava squeezed upward from the unsolidified basalt flow during emplacement of the overlying andesite.

Two lava flows on northeast Tol form relatively flat upland surfaces and unconformably overlie the dikes and fractured lavas. The two flows are exposed in palisades around the east and southeast sides of this upland area. The southern flow, of melilite-nepheline basalt, trends 1½ miles northwest and averages a quarter of a mile in width. The surface slopes about 2° SW. The exposed thickness ranges from 150 feet at the south end to 50 feet at the north. The northern flow, of nepheline basalt, is now a relatively small erosional remnant (a mesa) and averages 100 feet in thickness. At one exposure the nepheline basalt flow unconformably overlies a steeply dipping surface eroded in fractured dike-intruded lavas. This surface strikes N. 20° E. and dips 40° to 50° W. It is probably part of one side of a former valley filled by the lava flow.

Unusual pegmatitic zones characterize the upper part of the nepheline basalt flow In the exposed upper 5 feet of the outcrop, pegmatitic segregations are extremely abundant, and a few segregations occur as much as 30 feet below the uppermost exposures. On the eroded surface of the upland, pegmatitic areas range from a few inches to tens of feet in length and are of extremely irregular and varying width. They form slightly more than 50 percent of the eroded surface outcrops of the flow. Small dikelike veins and stringers occur, but in general the pegmatitic areas are very irregular. Some crystals are as much as ½ inch long but average from ⅛ to ¼ inch in their longer dimension. The pegmatite areas grade into the fine grain size that characterizes most of the flow. The lava surface is weathered into fluted pinnacles ranging from several inches to several feet in height (fig. 4).

Most of the volcanic rocks on Tol appear to dip westward, but definite measurements are possible only in a few outcrops because of poor exposures. On northwest Tol interbedded lava and autoclastic breccia strike north and dip 15° W. Autoclastic breccia beds in the

A

B

FIGURE 4.—*A*, Fluted pinnacles on nepheline basalt flow, northeast Tol (1956). *B*, Detail of fluted pinnacles on nepheline basalt flow, northeast Tol (1956).

eastern part of this northwest block strike N. 25° E. and dip 10° W. The uppermost flows on Mount Tumuital, observed from the southeast, show apparent dips of 5° to 7° SW. The melilite-nepheline basalt flow dips 2° SE., nearly at right angles to the older lavas. This flow may fill a valley eroded in the older lavas.

The volcanic rocks of Tol are predominantly olivine basalt, estimated at 75 percent of the total—the andesites are about 20 percent; nepheline basalt, melilite-nepheline basalts, and trachyte dikes form the rest.

FALA-BEGUETS

The elongated dome-shaped island of Fala-beguets lies 1½ miles east of Tol; it is 1½ miles along its northwest axis and averages slightly less than half this in width (pl. 2). Mangrove swamps border the west shore of the island. Calcareous sandy beach deposits extend half a mile westward from the north tip and are represented by two narrow bands on the mangrove-free east shore.

The volcanic rocks of the upland are largely olivine basalt flows. Pyroclastic breccias, if present, are not exposed. The slopes are covered with lava blocks; outcrops are few and so limited in extent that at no place could separate flows be mapped.

ULALU

The small oval island of Ulalu is 3.3 miles northeast of Tol (pl. 2). It trends east and is 4,000 feet long and averages 2,000 feet in width. The altitude of a small peak at the east end is 190 feet, but most of the island is less than 100 feet above sea level. Calcareous beach sands surround all but the northeast and east shores.

The volcanic rock on Ulalu is a single flow of nepheline basalt. The flow is unfractured and is not cut by dikes. Because of its similarity to the nepheline basalt flow on Tol, Ulalu is geologically included with the western rather than the central islands.

WEATHERING

Most volcanic bedrock is intensely weathered to depths ranging from a few inches to an observed 30 feet in one lava flow and to as much as 50 feet in the pyroclastic breccia and volcanic conglomerate. Bedrock exposures comprise only about 5 percent of the volcanic terrain; the remainder is covered by soil from weathering of underlying flows. Fresh, unaltered specimens are obtainable chiefly from ledges, palisades, and stripped surfaces on gently dipping lava flows.

In the deeply weathered flows, the residual centers of joint blocks are of hard, unaltered olivine basalt and andesite in a matrix of limonite-stained clay formed from weathering of the igneous rock. Boulders and cobbles of lava in the pyroclastic breccia and conglomerate show exfoliation and alteration to limonitic clay.

Plagioclase, augite, and magnetite are unaltered in most specimens of basalt and andesite. Fresh olivine occurs in a few weathered flows, but more commonly it is altered in varying degrees to serpentine and saponite. Alteration of interstitial glassy material to chlorite and saponite(?) is common in some flows.

Bauxite is developed in soil overlying the summit flow of sodic trachyte on Witipon, Meon. The bauxite consists of nodules ¼ to 4 inches in longer diameters and superficially resembles dried sponges in shape. The nodules are pale brown on the fresh surface and weather to moderate brown (fig. 5A and 5B). Where most completely developed, the soil profile is about 3 feet thick. The nodules form a 6-inch layer about 4 inches below the surface (fig. 6). Where soil is absent, the nodules locally lie scattered on the bedrock surface.

This occurrence of bauxite was described by Bridge (1948), who mentioned "200 acres" of bauxite exposures. This is evidently a misprint as the exposures are less than one-third of the total summit area of 63 acres of Witipon. Chemical analyses showed the bauxite to be so low in Al_2O_3 (42.78 percent in the Bayer process) that Bridge concluded it was of no commercial importance at the present time. There is nothing in the present survey that modifies his conclusions.

Chemical analyses of the weathering products of the Truk Islands are shown in table 1.

Concentrations of limonite in the form of concretionary nodules (fig. 5–B) cover the surface of olivine-rich basalt lavas on several of the volcanic islands and are especially well developed on Fefan and Ulalu. The nodules are generally ¼ to 1 inch in diameter. Locally the nodules are cemented by a limonitic matrix and are exposed as ledges and blocks. Position of the nodules in the soil profile is illustrated in figure 6–B.

TABLE 1.—*Chemical analyses of weathering products of Truk Islands*

	1. *Paseur 7–1*	2. *Ul. 10–1*	3. *Moen*
SiO_2	0.78	13.58	----------
Al_2O_3	16.47	23.00	53.08
Fe_2O_3	51.63	39.80	7.26
FeO	1.22	.60	----------
MgO	.23	.07	----------
CaO	.39	.24	----------
Na_2O	.28	.24	----------
K_2O	.06	.07	----------
TiO_2	5.36	5.63	.66
P_2O_5	3.88	..85	----------
MnO	.14	.07	----------
[illegible]O_2	.13	----------	----------
[illegible]O_3	.05	----------	----------
H_2O-	----------	4.82	----------
H_2O+	----------	11.71	----------
Insoluble	----------	----------	9.37
Loss on ignition	----------	----------	29.68
	80.62	100.68	100.05

1. Concretionary ferruginous laterite, Fefan. Analysis does not include H_2O. This material is similar to that forming upper part of Ulalu profile, fig. 6–B. Analysts, Asari and Ikawa.
2. Soft ferruginous laterite from center of profile shown in fig. 6–*B*, Ulalu. Analysts, Asari and Ikawa.
3. Bauxite nodule from Witipon, Meon, given by Josiah Bridge, Pacific Sci., v. 2, no. 3, July 1948.

A

B

FIGURE 5.—*A*. Profile of soil on trachyte flow near summit of Witipon, Moen (1956). Nodules of hydrated aluminum oxide form layer in upper part of profile. *B*. Gravelly surface of soils overlying nepheline basalt flow on Ulalu (1956). Ferruginous nodules form surface layer visible in photograph.

ORIGIN, AGE, AND PHYSIOGRAPHIC DEVELOPMENT

ORIGIN

The lava flows and pyroclastic beds of the Truk Islands dip gently away from the central part of Truk Lagoon and are remnants of either a single large volcano or several volcanoes. The volcanic flows on Moen dip northeastward from 2½° to 3½°. Dips of the flows on Dublon Island are generally eastward from 3½° to 4°. The dips of the flows on Fefan are southward from 1° to possibly 5°. A breccia unit on Udot dips 5° to 10° S. Most of the volcanic rocks on Tol dip west and southwest from about 5° to as much as 15°. An exception to these attitudes on Tol is the flow of melilite-nepheline basalt, which dips south or southeast. No evidence favoring deformation was discovered in the course of the fieldwork, and these dips very likely represent the original attitudes of the lava flows and pyroclastic beds. The gentle dips and great preponderance of lava flows over pyroclastic deposits suggest that the volcanic islands of Truk are erosional remnants of a shield volcano. Marine deposits are absent within the volcanic sequence, supporting lithologic analogy with other subaerial shield volcanoes such as those of the Hawaiian Islands.

No traces of crater or caldera walls have been found on any of the Truk Islands; however, field data provide a basis for postulating three main centers of eruption on the shield volcano (fig. 3). The presence of coarse, unstratified pyroclastic breccia and autoclastic breccia flows on Udot and Eot suggest a nearby source. The breccia contains a high proportion of accessory and accidental blocks and was probably explosively erupted from a large crater rather than from the lava-filled dikes of this area. The dikes of Udot and Eot have a radial pattern outward from a point in the lagoon about a mile north of Udot and half a mile west of Eot. By analogy with the radial dikes surrounding the central craters of many volcanoes, the central crater of the Truk shield volcano was probably north of Udot and west of Eot. The southward dip of the breccia beds on Udot is additional evidence that the crater lay to the north of Udot. The innumerable dikes on Tol are petrographically similar to the basalt and andesite flows of Tol and were probably the source for most of, if not all,

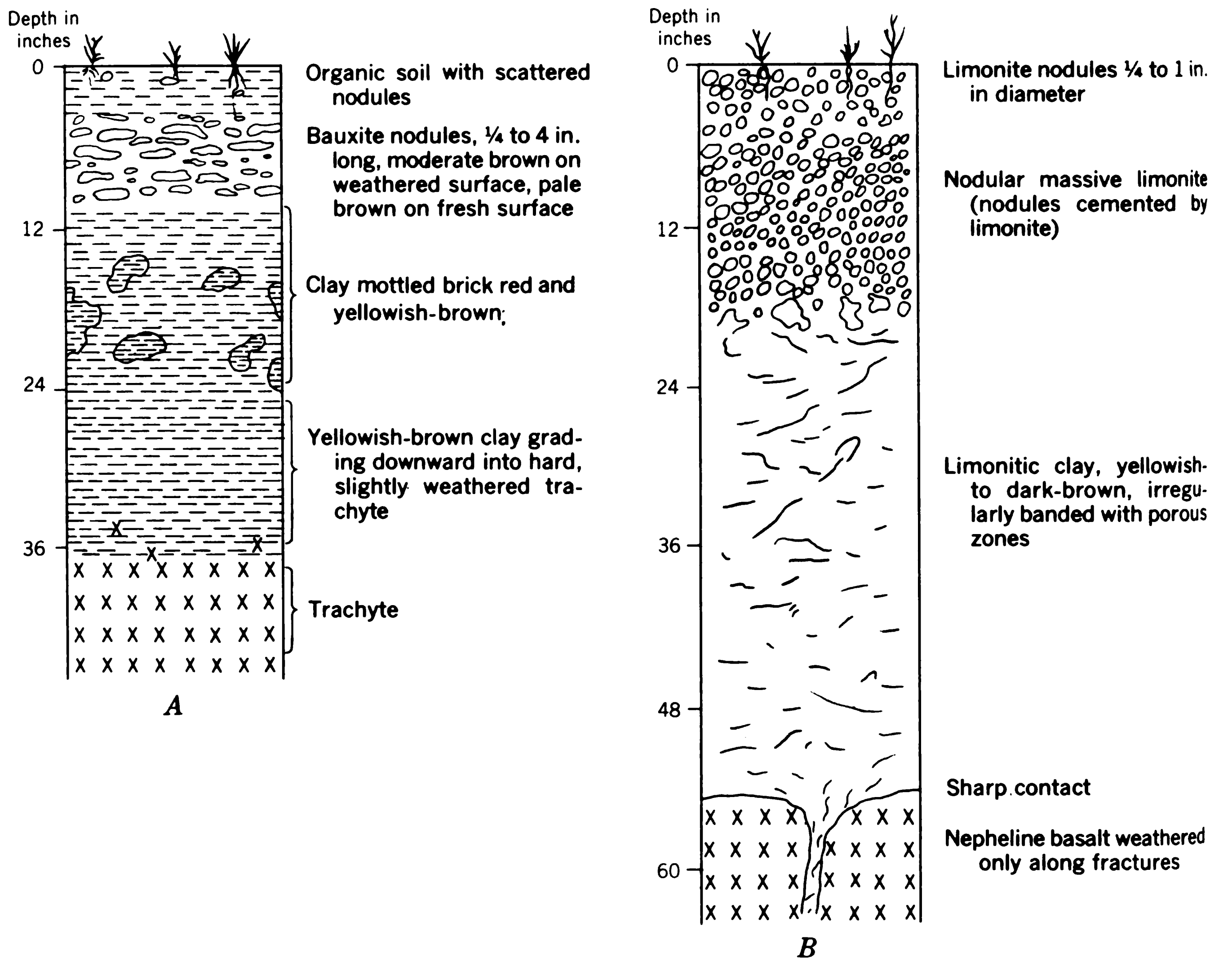

FIGURE 6.—Lateritic soil profiles on Moen and Ulalu illustrating relation of bauxite nodules to trachyte bedrock and limonite nodules to nepheline basalt bedrock.

hese lavas. An analogous dike complex is believed o have been the source for the thick sequence of lava lows forming the primitive shield volcanoes of Oahu (Stearns and Vaksvik, 1935, p. 95–96). Updip proections of the flows of Moen, Dublon Island, and Fefan suggest another center of eruption between Fefan and Moen.

The northeast alinement of valley-filling flows within he volcanic sequence on Moen suggests that the source ay southwest of Moen. Dikes are rare in the eastern slands, and these flows may have been erupted from a large crater rather than from a linear fissure system. The flows of the eastern islands are generally thicker, less vesicular, and more continuous in outcrops downdip than those of the western islands; further suggesting a relatively distant vent rather than a source at fissures of a nearby dike complex.

Other vents (that is, parasitic volcanoes) were undoubtedly present. The thin lava flows and abundance of pyroclastic material on Eten, in the eastern islands, is evidence for a parasitic volcano in the eastern part of the Truk shield volcano.

The nepheline basalt and melilite-nepheline basalt flows on Tol unconformably overlie the dikes and fractured lavas of the shield volcano and originated from another source. Lava flows forming the crest of Mount Tumuital, highest point in the Truk Islands (1,453 feet altitude), are fractured and cut by dikes that suggest that 1,400 feet or more of shield lavas was eroded prior to extrusion of nepheline basalt and melilite-nepheline basalt flows, which filled valleys cut in the shield volcano. These flows were apparently analogous to nepheline basalt and melilite-nepheline basalt flows of Oahu, which represent the latest phase of volcanic activity on that island. Two thousand feet of relief was developed on the Koolau shield volcano of Oahu before the valley-filling flows of nepheline basalt and melilite-nepheline basalt were extruded (Winchell, 1947, p. 3). The melilite-nepheline basalt flow on Tol dips to the south and southeast, and the nepheline basalt flow appears to fill a valley trending N. 20° E. Thus the vent(s) which erupted the nepheline basalt and melilite-nepheline basalt of Tol—and probably the nearby nepheline basalt on Ulalu—most likely lay to the north or northeast of Tol.

AGE

The most exact information about the age of the shield volcano is provided by fossiliferous limestone xenoliths of Udot and Eot. Tayama (1952, p. 84) reported that these xenoliths contain *Cycloclypeus* and *Miogypsina,* which indicate a Miocene age. About a handful of limestone xenoliths was collected from Udot and Eot in the course of the present study. They were examined by W. Storrs Cole, who identified the larger Foraminifera and kindly supplied the following information (written communication, July 14, 1955).

Lepidocyclina (*Nephrolepidina*) *sumatrensis* (Brady), which ranges from Tertiary *e* into Tertiary *f.*

Flosculinella sp., a poor specimen. Range of this genus is upper Tertiary *e* through Tertiary *f.*

Miogypsina (*Miogypsina*), whose range is same as *Flosculinella.* "* * * Certain features of the vertical section compare favorably with *M.* (*M.*) *eccentrica* Tan, an upper Tertiary *e* species; therefore, I am inclined to believe that these samples represent upper Tertiary *e* with the slight possibility that they could be Tertiary *f.*"

In terms of the standard time scale, these Formaminifera are probably early Miocene (Irving, 1952, p. 448).

The fossiliferous limestone xenoliths are interpreted to indicate that the volcano had grown approximately to sea level by early Miocene time. The limestone was probably deposited in shallow water near a coral reef. It must have been deposited on or near the central part of the volcano in order for the xenoliths to be ejected later from the central crater. Thus, the limestone was probably deposited when the volcano had grown from the ocean floor, at a depth of about 15,000 feet, to sea level and was temporarily covered by a coral reef. The fossils do not date the later subaerial growth of the shield volcano, and it is not known when the shield volcano reached its maximum size.

PHYSIOGRAPHIC DEVELOPMENT

Physiographic data and comparison with other volcanoes affords some basis for inferring the later history of the Truk volcano. Modification of the shield volcano to the present number of small islands scattered within a large lagoon is most easily explained by subsidence of the extinct volcano after dissection by erosion. Oahu, Hawaiian Islands, and Ponape, Caroline Islands, provide analogies to earlier stages of dissection and submergence. Oahu consists largely of a deeply dissected pair of shield volcanoes that have subsided several hundred meters since volcanic activity ceased (Stearns, 1946). Ponape, a single dissected shield volcano, has subsided even farther to drown a number of valleys, which are now fiordlike bays. Numerous islands have been formed adjacent to Ponape by partial submergence of coastal hills. Additional subsidence of Ponape would produce a number of rugged islands, the larger of which would be deeply indented by irregular bays and similar in shape to the island of Tol, the highest of the Truk Islands. The present concentration of volcanic islands in eastern and western parts of the Truk Lagoon suggests that the former shield volcano was elongated eastward. The barrier reef enclosing the Truk Lagoon probably originated as a reef fringing the shield volcano, as visualized by Darwin (1909). The reef continued to grow upward as the volcano subsided and gradually became the barrier reef that exists today.

The amount of subsidence can be estimated only between broad limits. Sinking of 1,000 to 2,000 feet is probably necessary to account for the individual Truk islands, formerly peaks on a single dissected shield volcano, but now widely scattered throughout the lagoon. Submarine slopes on the outside of the barrier reef are steep to depths of at least 700 to 800 fathoms, probably reflecting the reef structure. The volcanic platform has not necessarily subsided this much, however, for the lower third or half of the reef edifice may be reef-flank talus below the level of the reef itself.

The volcano may have commenced sinking either during or after active volcanism and may have subsided slowly enough for the reef growth to keep pace with subsidence. Analogy with Eniwetok atoll, which overlies a submerged Eocene or pre-Eocene volcano of the western Pacific (Ladd and others, 1953), suggests that the Truk Islands may continue sinking until the volcanic islands are submerged and only an atoll remains.

The present diameter of the barrier reef suggests a shield volcano roughly the size of Oahu. The outline of the Truk barrier reef probably does not coincide exactly with the margin of the former shield volcano. If the reef grew only upward and the volcano subsided uniformly, then one should expect the reef to be about equally distant from the outermost volcanic islands in the lagoon; yet, the reef lies only 2½ miles from Tol, largest single land mass in the Truk Lagoon, whereas the northern part of the barrier reef lies as much as 18 miles from the nearest volcanic island. The western part of the barrier reef, thin and discontinuous, may have grown inward towards the volcano during subsidence. The reef bordering the northeastern part of the lagoon is thicker and more luxuriant than the western part and may have grown outward as the volcano sank. Asymmetric development of the reef would be expected in this climatic zone, where the northeast trade winds predominate during most of the year.

Most of the coral islands and reefs in the Truk Lagoon probably originated as fringing reefs on volcanic hills, now submerged. Seventy-five feet of subsidence would submerge the reef-fringed island of Faneu and form a coral islet similar to the nearby islets of Fananan and Faleat. Assuming that most lagoonal coral reefs in the lagoon grew on volcanic platforms, then the distribution of reefs and coral islands should afford some indication of the area once covered by the former shield volcano, probably after some dissection. The area in which lagoonal reefs and coral islands are concentrated is indicated in figure 3. This evidence supports the earlier mentioned inference that the reef has grown outward towards the east and northeast beyond the margin of the volcano.

Evidence of Pleistocene change in sea level was sought throughout the Truk Islands. Japanese observers (for example, Tayama, 1952 p. 206) have reported marine terraces as occurring widely in Truk at altitudes of 1 to 2 meters, 30 to 50 meters, and 100 meters. These observers also noted that the presence of "mushroom rocks" of dead coral at places 3 to 5 feet above sea level implies erosion of preexisting reef rock caused by a drop in sea level of 1 to 2 meters. The "mushroom rocks" were studied briefly in the course of the present study and would indeed seem to indicate a relatively recent drop of 3 to 5 feet in sea level. Furthermore, wave-cut notches at an altitude of about 3 feet above sea level were found on the southeast coast of Udot, and a small exposure of reef limestone cemented to volcanic bedrock was found about 4 feet above sea level on the north point of Eot. The present study found no reliable evidence for widespread marine terrace surfaces, however. Flattish surfaces are widespread, but they occur at many horizons and almost certainly represent the upper surfaces of massive lava flows stripped by stream erosion. Most of the flattish surfaces have the same inclination as the base of the underlying flows, a feature not to be expected of marine abrasion terraces. Tayama (1952) reported the occurrence of limonite pebbles and basalt gravel on the terrace surfaces. The present writers believe the "pebbles" are limonitic nodules produced by weathering; the gravel may represent residual cores of spheroidal weathered basalt.

PETROGRAPHY

The Truk Islands lie east of the andesite line (fig. 7) the zone that follows the highly deformed island-arc belt along the oceanward side of the Aleutians, the Japanese Islands, and the Mariana group, through the Caroline Islands, southward around Fiji along the Tonga Islands to New Zealand (Betz and Hess, 1942; Hobbs, 1944). This zone is a major structural and physiographic boundary that separates the region of continental-type rocks—the basalt-andesite-dacite-rhyolite kindred of the Pacific margin—from the Pacific basin region of oceanic type rocks—"predominantly olivine basalts, nepheline basalts, and smaller amounts of their differentiation products" (Hess, 1948). The

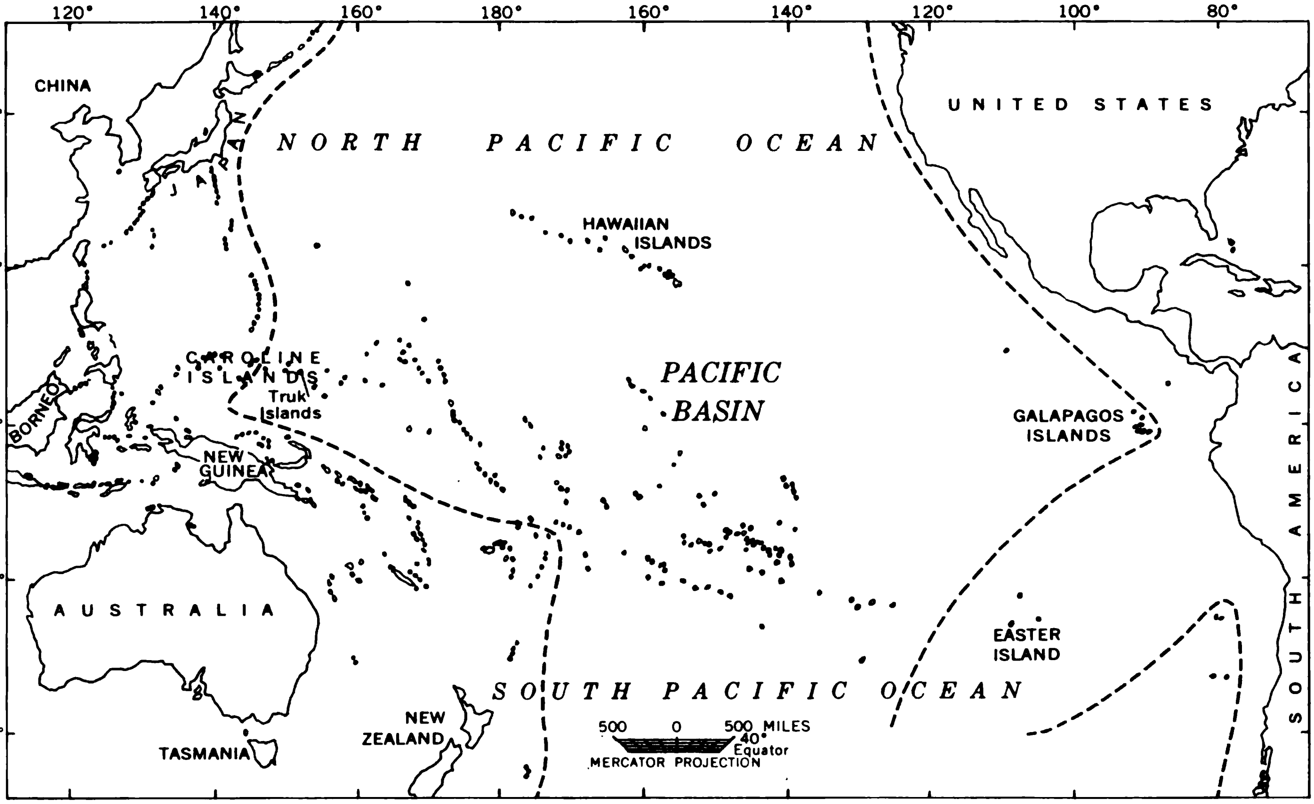

FIGURE 7.—Map of the Pacific Ocean showing location of the Truk Islands. Boundary of the Pacific basin (dashed line) characterized by oceanic crust, is taken from Macdonald (1949, p. 1590, fig. 11). Western boundary of the Pacific basin corresponds to the andesite line.

volcanic rocks of the Truk Islands are chiefly oceanic in type with olivine basalt forming approximately 70 percent, and andesite approximately 25 percent, of all the lava flows. The remaining 5 percent includes nepheline basalt, melilite-nepheline basalt, nepheline basanite, glassy basalts, and sodic trachytes.

This classification, as previously stated, is based on the color index of the rocks excluding phenocrysts; that is, upon the sum of the normative femic minerals (wo+en+fs+fo+fa+mt+hm+il) as calculated from chemical analyses of nonporphyritic rocks. In the absence of chemical analyses, the rocks are classified on the basis of comparison to analyzed rocks and the percentage of modal mafic minerals. In general the results of the classification closely approximates that used by Macdonald (1949, p. 1544) for the Hawaiian petrographic province, which is based primarily on the composition of the feldspars. The coarse blocks in the pyroclastic breccia of the central islands are called gabbros on the basis of mineral composition although their color index is generally less than percentages used to distinguish basalts. The petrography of the central islands of the Truk Lagoon is sufficiently different from that of the eastern and western islands to justify treating the groups separately.

PETROGRAPHY OF EASTERN AND WESTERN ISLANDS

OLIVINE BASALT

The most abundant rock type, both as flows and dikes, is olivine basalt. On fresh surfaces it is black to dark gray and varies through shades of brown and red to light gray, according to the degree of oxidation and weathering. Many outcrops are of hard, unaltered rock; others are reduced to a soft, claylike rubble with only residual boulders fresh enough to be recognized. Textures range from aphanitic to medium-coarse phaneritic and from nonporphyritic to extremely porphyritic, where phenocrysts form as much as 40 to 50 percent of the whole. The tops and bottoms of the flows are commonly vesicular, amygdaloidal, and slightly finer grained than the interior of the flow. Pronounced chilling at the edges is present in narrow selvages of some of the dikes and at the base of a few flows. Generally, however, the difference in grain size between edges and the interior of the flows is slight. Scoriaceous and amygdaloidal basalt characterize many of the autoclasitic breccia zones at the tops and bottoms of flows. Vitrophyric olivine basalt forms a small proportion of basalt flows and a few small dikes in the western islands. Most of the olivine basalts have an intergranular texture, but ophitic to subophitic textures occur. In many flows the lath-shaped plagioclase crystals show typical trachytoid orientation. Such rocks break readily into slabby blocks and present shining surfaces that are due to reflections from parallel crystal and cleavage faces. This rock cleavage, due to flow orientation of elongate and platy minerals, is especially common in andesite flows, but microscopic examination is often necessary before distinction between basalt and andesite can be made.

The conspicuous phenocrysts of the olivine basalts are pyroxene, plagioclase, and olivine. Magnetite occurs as a phenocryst in a few of the extremely porphyritic flows. The predominance of any one or any combination of phenocrysts differs widely from flow to flow, and within a few flows the proportion of phenocrysts differs strikingly from bottom to top. The lower 6 to 7 feet of one flow is nonporphytic and grades upward through a vertical distance of several feet into extremely porphyritic lava, which forms the remaining 30 feet of the flow. In many outcrops the olivine is so completely altered that its presence is detected only under the microscope, where it is seen as very small relicts of fresh olivine or as alteration pseudomorphs of serpentine with euhedral outlines of the original olivine preserved. In some flows, pyroxene is the only megascopic mineral; in others olivine is the only recognizable phenocryst. More rarely plagioclase is the only apparent phenocryst, but there are very few of the porphyritic basalts in which plagioclase is wholly unrecognizable in hand specimens. The megascopic minerals range in size from tiny laths of plagioclase up to phenocrysts three-quarters of an inch long. Two generations of phenocrysts are apparent in many sections under the microscope, and, less commonly, three generations occur. Flow orientation is generally evident. The phenocrysts are fractured and torn apart and show embayments and rounded edges that are due to resorption. In many of the flows, there are glomeroporphyritic clusters that may have formed from floating together of anhedral crystals. Most generally they are formed from the same mineral, plagioclase, pyroxene, or olivine, although some clusters contain mixtures of all three.

Plagioclase in the olivine basalts averages between 45 and 50 percent of the rock and ranges from bytownite to andesine. Sodic labradorite is the dominant feldspar in most of the flows, but several flows contain chiefly calcic andesine. The phenocrysts in some flows have bytownite cores with peripheral zones of less sodic plagioclase. Although euhedral and subhedral phenocrysts are common, many are anhedral owing to fracturing and resorption. There is no evidence of saussuritization, but in some flows resorption embayments filled with fine-grained groundmass leave only small, irregular remnants of the original crystals. Many feldspar phenocrysts contain small inclusions of pyroxene and olivine. Zoning is not especially common, and the bands where present are only faintly outlined. Plagioclase laths of the groundmass show random orientation in a few flows, but these are greatly subordinate to textures showing some degree of parallel orientation that is due to flow. In many flows small laths are closely matted together and are molded by flow around the larger phenocrysts.

Alkali feldspar is present in small amount in some of the olivine basalts. It occurs as fine interstitial material having lower refractive index than the plagioclase. As much as 8.5 percent of normative orthoclase is present in some basalts, a further suggestion that this interstitial feldspar is potassic (probably anorthoclase).

The pyroxene content of the olivine basalts ranges from 25 to about 40 percent. It is monoclinic and ranges from colorless augite, with maximum extinction angles ($C \wedge Z$) between 39° and 45°, to faintly tinted pink and stronger shades of violet and reddish tan, in varieties of augite rich in titanium. The titaniferous augites are by far the most abundant. They show strong birefringence and faint pleochroism. Optic angles measured by universal stage range from 50° to 58°, average less than 54°, and α in a typical specimen is 1.695. These data indicate an augite having the composition of Wo_{45}, En_{21} Fs_{34} (tables of Hess, 1949). No orthorhombic pyroxene has been recognized.

The pyroxene phenocrysts are angular to rounded owing to absorption, and commonly form glomeroporphyritic clusters. Deep resorption embayments have reduced many of the larger crystals to extremely irregular relicts. They are generally fresh although in a few flows both phenocrysts and groundmass grains are completely changed to fibrous pseudomorphs of chlorite. In the groundmass the augite occurs in slightly elongated flakes and equigranular grains between the plagioclase grains.

Olivine occurs in all the basalt flows as phenocrysts and in small grains in the groundmass. It generally ranges from about 4 to 20 percent and averages 5 to 10 percent of the basalts. No completely olivine-free flows were observed, and in flows where no olivine is visible its former presence is detected from pseudomorphs of serpentine or saponite(?). Some show typical euhedral outlines, but more commonly the phenocrysts are subhedral to anhedral, and many are deeply embayed. Pyroxene was nowhere found to form reaction coronas around olivine crystals. Rims of serpentine surround fresh relicts of olivine, and larger grains show fibrous alteration to antigorite and chlorite along partings and around grains. Pseudomorphs of these alteration minerals are common. Some flows on Tol are rich in fresh olivine, and its alteration is limited to narrow peripheral fringes and discolorations along partings that are due to incipient development of serpentine. Iddingsite is prominent as an alteration product of olivine, and in many of the flows on Tol, numerous crystals are completely altered to pseudomorphs of iddingsite. Other olivine grains show bands of yellowish-green antigorite surrounding centers of iddingsite. In some places this relation is reversed, with cores of serpentine and chlorite surrounded by peripheral zones of iddingsite.

Magnetite is always present in the olivine basalts. It occurs most commonly as small octahedra and irregular grains from 0.01 to 0.05 mm in diameter that are distributed more or less evenly throughout the rock. In a few flows elongated rods up to 1 mm long appear to be of later growth than the equidimensional grains. In other flows large grains 1 to 2 mm long show extremely irregular vermicular borders. Only rarely, in specimens fresh enough for thin sectioning, do the grains show oxidation to hydrous iron oxides.

Iddingsite is conspicuous in many flows as alteration after olivine. It ranges from light yellow and orange, where it is incipiently developed around edges and in fractures of olivine grains, to deep reddish brown, reddish orange, and bright red in many of the pseudomorphs. A faint pleochroism is discernable in some of the iddingsite, but more generally pleochroism is absent or masked in the darker varieties.

Biotite occurs in small amounts as rods and flakes, generally less than 0.05 mm in length. It ranges from yellow to olive to dark brown and shows faint to strong pleochroism. In a few flakes it is well enough developed to show the typical "bird's eye maple" appearance. It occurs as isolated interstitial crystals of yellow to orange brown.

Serpentine, largely yellow and greenish-yellow antigorite, is a common alteration product of olivine and of some pyroxenes. In a few specimens it forms as much as 10 percent of the rock. Chlorite is a common associate of serpentine in the alteration of olivines and pyroxenes. In some specimens it is the chief constituent of fibrous pseudomorphs after augite. Light-green chlorite is present as an alteration of glass in the glassy groundmass of some aphanitic and scoriaceous flows. Calcite is common in a few flows as amygdules and cavity linings and is irregularly developed interstitially throughout some rocks. It occurs sparsely in the alteration of the plagioclase, but none of the feldspar shows typical saussurization. Zeolite occurs as vesicle fillings in a few glassy and aphanitic basalts. It is generally in the form of radiating rosettes. The fibers show parallel extinction and are length slow. Apatite forms swarms of elongated acicular needles, commonly included in the feldspars. It is generally more abundant in flows near the basalt and andesite borderline but occurs to some degree in all flows.

A light-yellow to dark-brown and black mineral occurs in small amounts in many of the olivine basalts. It forms poorly developed rods and flakes and occurs interstitially in angular areas between plagioclase laths. Other grains appear to be discrete mineral flakes showing an incomplete hexagonal outline and a complete gradation from subtransparent brown to opaque according to thickness of plates. It is similar to ilmenite and to the isomorphous mixture of hematite and ilmenite described by Kuno as being in the basalts of the Hakone volcano (1950, p. 983).

NEPHELINE BASALT

A flow of nepheline basalt occurs interbedded with olivine basalt and andesite flows on Dublon Island and as a capping flow on the upland plateau of northeast Tol. The small island of Ulalu, just northeast of Tol, is entirely composed of nepheline basalt. These rocks differ from the olivine basalts by having feldspathoid in place of feldspar. They are generally porphyritic with small phenocrysts of pyroxene and olivine, which in some specimens form as much as 50 percent of the rock. The holocrystalline groundmass consists of nepheline, magnetite, pyroxene, apatite, and biotite.

The nepheline basalt on Tol is characterized by conspicuous pegmatitic segregations in the upper part and to a lesser extent throughout the flow. The coarsest grains are half an inch long; most average between one-eighth and one-quarter of an inch. The pegmatitic areas are extremely irregular, have gradational boundaries with finer grained parts of the flow, and in places are connected by pegmatitoid veins. Numerous vugs occur in the pegmatitic segregations. Minerals of the pegmatite are the same as in the country rock, predominantly nepheline, pyroxene, olivine, and magnetite. In general the nepheline is subordinate to the pyroxene and olivine. It occurs as poikilitic phenocrysts half an inch in diameter, and as anhedral, irregularly rounded grains in the groundmass. Very small grains of magnetite and olivine are commonly present as inclusions.

Augite is extremely abundant in some of the specimens, ranging from 20 to 50 percent. It occurs commonly as very irregular to euhedral prismatic crystals arranged rudely in radiating pattern. The optic angle measured on the larger grains ranges from 45° to 52°; 2 V of the smaller crystals studied is 48° to 52°. The coarse pyroxene of the pegmatite area is light to dark pink in thin section and is probably highly titaniferous. The pink titanaugite grades outward into a peripheral zone of pleochroic green aegirine-augite. Small amounts of light-green aegirine-augite with lower extinction angles ($C \wedge Z$) than augite also occur as discrete crystals in the groundmass.

Olivine is always present, averaging between 20 and 25 percent. It is generally fresh with only incipient discoloration that is due to serpentinization along partings and around edges of the grains. Some crystals are euhedral, and others show deep resorption embayments. A few grains show the beginning of alteration to iddingsite.

The amount of magnetite ranges considerably, from 1 to 5 percent. It occurs as small octahedral grains sprinkled throughout the groundmass and as a few microphenocrysts. Small grains are commonly enclosed in the nepheline. Ilmenite occurs as small irregular grains in the groundmass and as subhedral grains similar to its occurrence in the olivine basalts.

Mica is present as discrete light- to dark-brown crystals of biotite and as partly formed flakes closely associated with olivine and magnetite. It is generally more abundant in the nepheline basalt than in the olivine basalt and andesite flows but never forms as much as 1 percent of the rock. Specimens from the nepheline basalt flow on Ulalu contain a few grains of pink to pale-brown garnet, probably andradite (var. melanite). In other specimens, highly refractive isotropic grains suggest garnet but are too poorly formed to be positively identified. Apatite occurs in small acicular needles embedded in other minerals.

Small irregular brown opaque areas of ilmenite occur interstitially in discrete flakes. They are similar to the ilmenite in the olivine basalts.

MELILITE-NEPHELINE BASALT

Melilite-nepheline basalt was recognized only on Tol, as a flow capping the southern part of the upland plateau of northeastern Tol. It is similar in most respects to the nepheline basalt described above and differs chiefly by containing melilite as well as nepheline. The melilite is largely confined to the groundmass, but a few microphenocrysts are present in some specimens, always subordinate in amount to the nepheline. The flow displays excellent examples of clathrate texture.

NEPHELINE BASANITE

One flow of nepheline basanite is known to occur on Moen. This lava is similar to the nepheline basalt previously described except that a small amount of feldspar takes the place of some nepheline and a groundmass of dark-brown glass is present.

VITROPHYRIC BASALTS

A few flows consist largely of olivine and pyroxene in a groundmass of brown glass. These petrographically resemble the nepheline basalts except that glass forms the groundmass in place of nepheline. These flows may be nepheline basalts chilled before the nepheline crystallized.

ANDESITE

The andesites of the Truk Islands are mineralogically similar to the olivine basalts. Much of the foregoing description of the olivine basalt applies equally well to the andesites, and only the differences will be stressed.

The andesites generally show trachytoid orientation of the plagioclase laths and are nonporphyritic. The flow structures are so well developed that many andesites fracture into platy slabs with glistening surfaces that are due to parallel crystal faces and cleavages of tabular minerals. As some basalts also exhibit this platy fracture, accurate field determinations are often impossible.

The andesites are composed primarily of the same four types of minerals that are dominant in the olivine basalts: plagioclase feldspar, monoclinic pyroxene, olivine, and magnetite. All four occur as phenocrysts, but generally the andesites are fine grained, and megascopic minerals are relatively few and widely scattered. Microporphyritic textures are seen in thin sections, and some specimens show three generations of mineral growth. The phenocrysts are most commonly between 1 and 2 mm, but may be as much as 3 or 4 mm in diameter. Glomeroporphyritic clusters of phenocrysts and fragments of phenocrysts are common. Euhedral outlines of crystals are partially preserved in some flows but commonly they are subhedral to anhedral and may show extremely irregular edges that are due to resorption.

Andesine is the most common plagioclase although oligoclase is the predominant feldspar in some of the andesites. Zoning in the plagioclase crystals is less obvious than in those of the olivine basalt. Peripheral bands of some of the zoned crystals show extinctions that indicate a slightly more sodic composition than the central cores. The parallel orientation of the groundmass laths and the fragmentation of phenocrysts, owing to flow, are strikingly developed in many specimens. Plagioclase of the groundmass averages between 60 and 70 percent. Small amounts of anorthoclase occur as fine interstitial crystals in the groundmass.

Augite is the most abundant pyroxene. It ranges from typical colorless grains in the groundmass to pink and violet-tinted varieties of titaniferous augite. The darker augite is conspicuous in rocks transitional from andesite to basalt. Optic angles in the pyroxene range from 48° to 53° and average about 50°; α is 1.698 to 1.699. These data indicate an augite of composition Wo_{40}, En_{33}, Fs_{27}, somewhat more iron rich than that of the basalts. Small equigranular grains in the groundmass appear fresh in most rocks. In other rocks the pyroxene has weathered to fibrous masses of chlorite and antigorite.

Olivine is generally present in the andesites in amounts as much as 8 percent. It occurs as phenocrysts and in the groundmass and commonly shows relicts of fresh olivine surrounded by zones of granular and fibrous antigorite, chlorite, iddingsite, and more rarely, small flakes of saponite(?). In some places these alteration minerals form pseudomorphs with euhedral outlines of the original olivine crystals. In the more sodic andesites, fresh olivine is not recognized although its former presence is suggested by the shape of the pseudomorphs of chlorite and other alteration products. Pyroxenes may give similar alterations, but there is little doubt that olivine was once present in all the andesites examined.

Magnetite is ubiquitous and ranges from 3 to 6 percent of the andesite flows. It occurs as very small octahedra and irregular grains, more or less widely distributed throughout the rock, and there are a few larger grains as much as 1 mm in diameter. Around the edges the larger crystals commonly show vermicular extensions, similar to vermicular grains in the olivine basalts. In a few specimens the magnetite occurs almost entirely as small slender rods as much as 1 mm long. Apatite rods ranging from 0.1 to 0.5 mm long are fairly abundant in the andesite flows. Some are enclosed by the feldspar.

Antigorite, chlorite, and olive-green saponite(?) occur as alteration products after olivine and pyroxene. Biotite is more common in the andesites than in the basalts. It occurs as discrete plates scattered throughout the groundmass, rarely forming as much as 1 percent of the rock. Biotite is most common in the basalts of Dublon Island. Later deposits of natrolite and of calcite are present in a few flows as vesicle fillings and irregular deposits along fractures.

Small irregular opaque areas of interstitial ilmenite occur in the groundmass of many of the andesites and appear similar in every way to the ilmenite of the basalts.

TRACHYTE

The trachyte forms only a small part of the volcanic rocks of the Truk Islands, probably less than 2 percent judging from present exposures. On Witipon, a hill on Moen, 150 feet of relatively flat lying trachyte is the youngest flow and forms the caprock. Other exposures occur in a small isolated outcrop on the west slope of Tonaachau (Moen) and in dikes on Tol. The trachyte on Moen was reported as amphibolite schist by Kramer (1908); Bridge (1948) made a restudy of the outcrops, identifying them correctly as trachyte. On the slope of Tonaachau the small exposure is interbedded with andesitic and basaltic flows, and on Tol the trachyte dikes cut both basalt and andesite flows.

The trachyte in hand specimens shows only a few phenocrysts, but most sections under the microscope are microporphyritic with a few oligoclase microphenocrysts in a groundmass of lath-shaped anorthoclase(?). α is 1.527 to 1.528; β is 1.530 to 1.532. The optic angle is estimated between (−) 50° and 60°. The normative plagioclase calculated from one analysis is $Or_{22}Ab_{77}An_1$. Most of the specimens show trachytic textures with closely matted anorthoclase(?) forming approximately 80 percent of the rock. It is this parallel flow structure that gives the megascopic schistose appearance and a rock cleavage resembling that of a metamorphic schist. In a few specimens the trachytic orientation gives way to more intergranular textures. In general, the crystals show much less resorption than those of the andesites and basalts, and in spite of the weathered appearance in outcrops, the trachyte appears remarkably fresh when seen in thin sections.

The pyroxene content is low and rarely forms as much as 1 percent. It is pale-yellowish-green augite with maximum extinction angles averaging around 40°. It occurs interstitial to the feldspar laths and only rarely as phenocrysts.

Small phenocrysts of fresh olivine are present in most specimens. They are generally surrounded by thick rims of alteration products. Serpentine and chlorite are recognized in small patches, but mostly the character of the rims is masked by iron oxide staining.

Magnetite is always present as widely scattered octahedra throughout the groundmass and as a few microphenocrysts. It is never as abundant as in the andesite and averages only between 1 and 2 percent of the rock.

Light-yellow to light-brown biotite is present in small amounts in many of the trachyte specimens. It occurs as poorly developed flakes, interstitial to the feldspar laths and commonly parallel to them. Amphibole also occurs chiefly as pleochroic green and brown hornblende. A few grains show small extinction angles and may be oxyhornblende. Both biotite and amphibole are in small amounts of less than 1 percent.

Apatite occurs as small acicular needles enclosed in other minerals of the groundmass. Small amounts of brown, opaque ilmenite are also present in the groundmass.

The trachytes differ from the andesites by the smaller amounts of dark minerals, by the dominant feldspars being albite and oligoclase rather than oligoclase and andesine as in the andesites, and by the distinct color contrast between the light-gray trachytic flow and the dark-colored andesites. Flows transitional between the andesite and trachyte have not been found, in contrast with those showing the transition from andesite into basalt.

PETROGRAPHY OF THE CENTRAL ISLANDS

LAVA FLOWS AND DIKES

The basalts and andesite flows of the central islands are fundamentally similar to those of the other Truk Islands but are more highly altered. The least altered flows consist largely of plagioclase (oligoclase, andesine, and labradorite), augite, chlorite pseudomorphs after olivine, anorthoclase, and magnetite. Poikilitic hornblende crystals, pinkish brown to brown in pleochroism, are common in one andesite flow; small rodlike crystals of dark-brown hornblende were identified in several other andesites.

Plagioclase and pyroxene are altered to varying extents in most basalts and andesites. Plagioclase phenocrysts of many basalts and andesites are albitized along fractures and througout irregular areas. In some flows the plagioclase is entirely albitized and has a dusty, altered appearance. No slides of basalts and andesites were stained for identification of potassic feldspar. The pyroxene is commonly unaltered in andesites and basalts in which the plagioclase is albitized; conversely the plagioclase is largely unaltered in several andesites in which the pyroxene is entirely replaced by secondary minerals. The pyroxene is extensively replaced by chlorite or calcite or both. Epidote and actinolite less commonly replace the pyroxene. Interstitial chlorite and small irregular crystals of sphene are abundant in the completely altered rocks. Amygdules generally consist of chlorite and calcite. Quartz is present in a few amygdules, and pyrite occurs in some dikes.

All the trachyte flows and dikes are highly altered and consist largely of albite and potassic feldspar having a dusty, altered appearance. Many contain several percent of quartz. The optic angle of the albite is estimated to lie between +70° and +80°; this angle suggests that the albite does not contain potassic feldspar in solid solution. The albite is widely mantled by and intergrown with potassic feldspar, as shown by staining of slides with sodium cobaltinitrite as described by Chayes (1952). In some slides potassic feldspar is concentrated in the marginal part of the albite laths; elsewhere entire crystals consist largely of potassic feldspar. Plagioclase is locally sericitized in a few thin sections. Quartz is present in most trachytes and occurs as anhedral interstitial crystals and as polycrystalline aggregates that form as much as 5 percent of the rock. Some polycrystalline aggregates of quartz, as much as 1.5 mm in diameter, appear to fill pores in the trachyte. Euhedral outlines or "ghost crystals" are common in the crystals of a few pore-filling quartz aggregates, probably representing euhedral nuclei about which later growth of quartz took place. Quartz replaces some small irregular areas of adjacent feldspar crystals, and quartz pseudomorphs after plagioclase phenocrysts are present in one specimen studied. Textural relations indicate that most quartz is primary, but evidence of secondary growth and replacement relations suggest that some quartz is secondary (probably hydrothermal). No primary, unaltered mafic minerals other than magnetite are present in the trachytes, but secondary chlorite is abundant, probably replacing the primary mafic minerals. Sphene is disseminated throughout most trachytes. Pyrite occurs commonly as disseminated crystals and as coatings on fractured surfaces.

PYROCLASTIC BRECCIA

The breccias on Udot and Eot consist of fragments of rock in a fragmental finer grained matrix of rock, crystals, and fine tuff. Locally, these breccias appear to grade into flows of autoclastic breccias in which the matrix is composed of both lava and tuff. In thin section some of the rock fragments can be seen surrounded by smaller chips derived from larger fragments. These fragments were immobilized in the process of autobrecciation, probably during mudflow or some other type of mass movement.

Blocks of andesite, trachyte, quartz trachyte, and basalt predominate in the breccias. Phaneritic gabbro blocks are locally common, and blocks of hornfelsic recrystallized gabbro, basalt, and breccia occur in the bed in which gabbro blocks are most abundant. Several small quartz monzonite blocks were found, and several gabbro blocks contain dikes of monzonite and quartz monzonite. Other gabbro blocks contain dikes of andesite and basalt and inclusions of recrystallized basalt. A few limestone xenoliths were also collected.

ANDESITE, BASALT, AND TRACHYTE BLOCKS

Most of the blocks of andesite, basalt, and trachyte are lithologically similar to the lava flows and dikes, and the pyroxene and plagioclase in most blocks are similarly altered.

One block of quartz trachyte similar to the quartz trachyte of the dikes and flows was studied in detail to determine the nature of the alterations. The rock consists largely of feldspar, which occurs as oriented laths of plagioclase peripherally mantled and partly replaced by potassic feldspar. Both feldspars have a dusty, altered appearance (fig. 8A). Refractive indices indicate that the plagioclase is nearly pure albite. The large optic angle suggests that albite is not significantly potassic. The X-ray diffraction pattern indicates that albite is structurally similar to "low" plagioclase (D. B. Stewart, written communication, 1956). Quartz occurs as anhedral crystals forming an estimated 5 percent of the rock (fig. 8A–B). No primary mafic minerals are present but a chloritic-appearing mineral forms angular interstitial fillings which may replace primary hornblende. Chlorite(?) also replaces plagioclase.

Some trachyte and quartz trachyte blocks are coarser than the trachyte of flows and dikes observed at the surface. The plagioclase of the coarser blocks is albite which lacks the trachytic orientation of dikes and flows. The albite has a turbid appearance, however, and appears similar to the plagioclase of the dikes and flows now exposed. Irregular areas of orthoclase are present within most plagioclase crystals. Brown hornblende that is zoned outward to blue-green riebeckitic amphibole occurs in some blocks. Chlorite, sphene, and, less commonly, epidote form interstitial fillings and locally replace plagioclase. Amygdules consist of calcite, chlorite, and quartz, either singly or in combination. These trachyte blocks probably represent trachyte magma that crystallized below the surface and was altered by hydrothermal solutions and gases(?) similar to those which altered the dikes and flows.

Several blocks of andesite and basalt, coarser in texture than surface flows and dikes, were found. Pyroxene in these blocks is generally altered to pseudomorphous actinolite(?) in which small crystals of sphene are embedded. Less commonly, epidote replaces the pyroxene. The plagioclase is replaced by albite and by potassic feldspar. Quartz occurs in some amygdules of these blocks. One relatively unaltered block of olivine basalt (or olivine dolerite) porphyry was found. The block consists of subhedral to euhedral phenocrysts of olivine, generally 2 to 4 mm long, in an ophitic groundmass of augite (poikilitic crystals, generally 3 mm long), calcic labradorite or bytownite laths, generally 0.4 to 1.5 mm long, and magnetite. Olivine forms 40 percent of the rock, feldspar 30 percent, augite 28 percent, and magnetite 2 percent. The olivine is partly altered to chlorite, and the plagioclase is locally sericitized.

Two blocks of vitrophyric olivine-rich basalt containing amygdules of nearly pure albite were collected. Much of the albite is clear, but more commonly it has a dusty, altered appearance similar to that of the albite in the altered flows and dikes. The albite is structurally "low," according to D. B. Stewart (written communication, 1957). Olivine phenocrysts, formerly abundant in these blocks, are now represented by pseudomorphs of chlorite, serpentine, and, less commonly, albite.

GABBRO BLOCKS

The gabbro blocks are fine to coarse grained and hypidiomorphic granular and consist largely of labradorite, augite, brown hornblende, magnetite, apatite, and both magnetite-chlorite and actinolite-magnetite pseudomorphs after olivine crystals. Reddish-brown hornblende occurs in some of the gabbro. The proportion of mafic minerals (including pseudomorphs after mafic minerals) ranges greatly but is generally between 20 and 35 percent of the total. The grain size of most blocks averages about 5 mm, but it is 1 to 2 mm in some blocks and as much as 1 cm in a few others. The grain size is rather uniform in most blocks but ranges considerably in some. The minerals most commonly lack orientation, although the plagioclase displays excellent planar orientation in a few blocks. The plagioclase in one block has linear as well as planar orientation. The labradorite has an average composition of An_{60-65} but is commonly zoned to more sodic margins.

The pyroxene is uniformly augite. α ranges from 1.688 to 1.696 and 2 V was found to be 45° to 53°. The composition of the pyroxene ranges from wo_{43}, fe_{17} en_{40} to wo_{40} fe_{25} en_{35}, on the basis of tables of Hess (1949). Some augite is pink and is probably titaniferous.

Brown hornblende is present in most thin sections. It occurs as crystals both intergrown with augite and mantling augite and magnetite.

Brown biotite having an altered appearance fringes magnetite in several thin sections. It is most common in blocks which display evidence of thermal recrystallization and may be largely, if not entirely, metamorphic in origin.

A few gabbro blocks have been locally to completely recrystallized to a granoblastic rock similar in composition to the parent gabbro except for olivine which was not found. The gross texture of primary gabbro is preserved in some blocks in which the individual primary igneous crystals are represented by fine-grained monomineralic aggregates of the same composition. The recrystallized plagioclase crystals are commonly 0.4 to 0.7 mm long and about 0.2 to 0.4 mm wide. The recrystallized augite is more variable in size, and in some rocks the small crystals are optically continuous. The augite is recrystallized in some blocks in which the plagioclase is clouded with abundant dustlike inclusions but not otherwise recrystallized. The dusty appear-

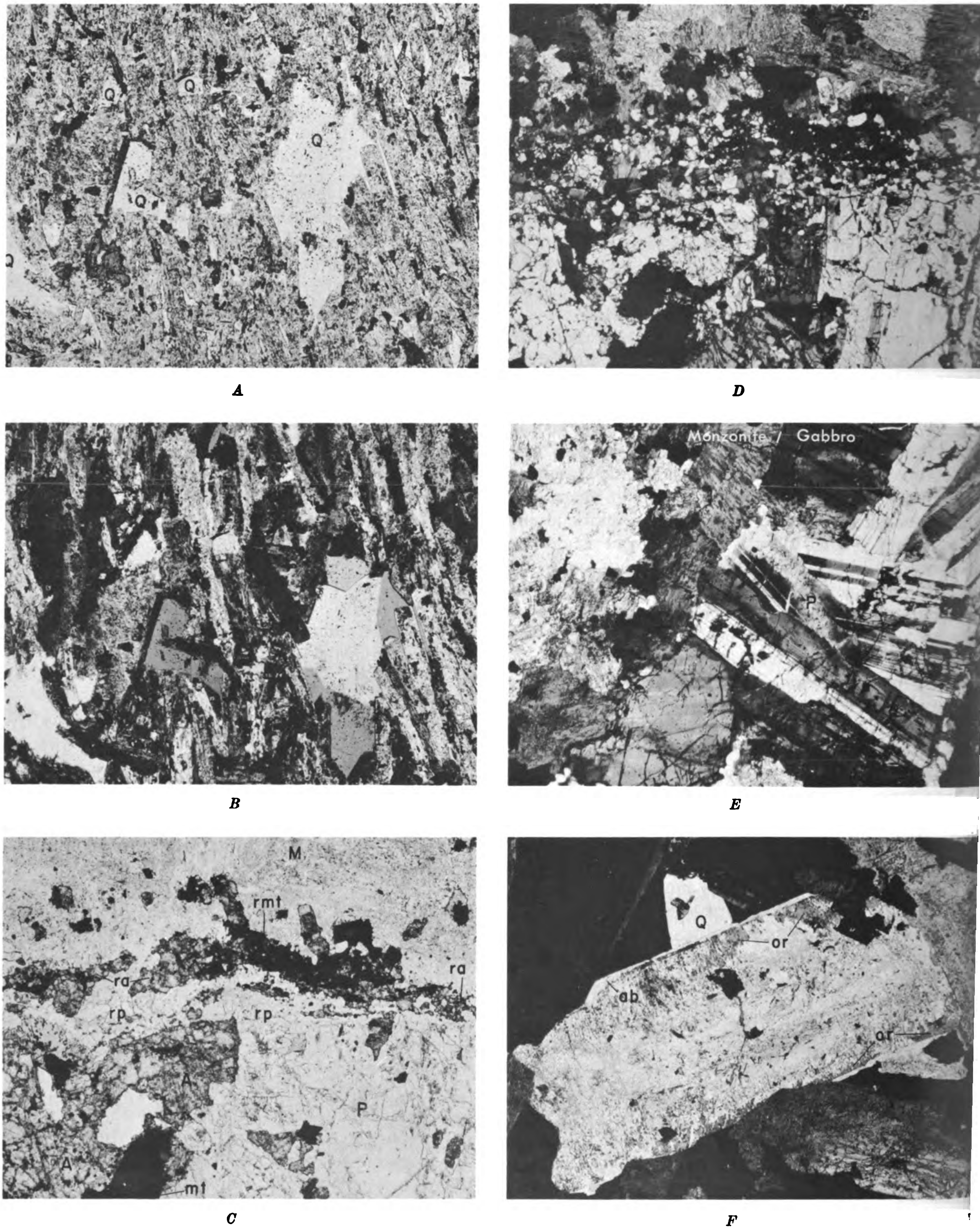

FIGURE 8.—(See legend on opposite page.)

ance of the labradorite was probably caused by heating below the melting point of the plagioclase (MacGregor, 1931). Veinlets of recrystallized augite that cut dusty labradorite crystals in some blocks suggest that the augite was melted and flowed into fractures in the plagioclase crystals. Clouded plagioclase crystals are the only evidence of heating in a few blocks.

All the gabbro blocks are altered to some extent. In the freshest blocks only olivine is altered (to pseudomorphs of magnetite-chlorite and actinolite-magnetite). Augite is partly to wholly altered in many blocks to pale-green actinolite, in which small crystals of sphene are embedded. Epidote replaces some augite, and much of the biotite is partly altered to chlorite. Brown hornblende is replaced by pseudomorphous actinolite and less commonly by epidote in the more altered blocks. Plagioclase of most blocks is partly replaced by alkali feldspar. Albitization has occurred along narrow fractures in some gabbro, but some labradorite of other blocks is entirely replaced by sodic plagioclase and potassic feldspar. Augite is commonly recrystallized but otherwise unaltered in some blocks in which the labradorite is entirely replaced by alkali feldspar. Calcite and quartz are present in many altered blocks, and a veinlet of quartz is contained in one block of gabbro. Some gabbro has been converted by these alterations to an aggregate of epidote and actinolite; other gabbro has been altered to a composition approaching quartz monzonite.

RECRYSTALLIZED BASALT BLOCKS

Blocks of recrystallized basalt as much as 8 inches in diameter occur both as inclusions in the gabbro blocks and as individual xenoliths. The degree of recrystallization varies. The more recrystallized basalt is a plagioclase-augite-hornblende-magnetite hornfels in which biotite may be present. The hornblende is brown and forms poikilitic crystals enclosing the magnetite. All the plagioclase is clear and unzoned (that is, recrystallized) except for zoned relict cores of cloudy appearance in some larger crystals. In less recrystallized blocks, clouded relict plagioclase phenocrysts are commonly embedded in a granoblastic groundmass. Only augite is recrystallized in some blocks. Coarse-grained seams or veinlets are present in one of the hornfels blocks. The granoblastic and clouded appearance of relict plagioclase suggests thermal recrystallization of these basaltic blocks.

BRECCIA BLOCKS

A small proportion of breccia blocks occur within the breccias on Udot and Eot. Many of these blocks consist of fragments of gabbro and volcanic rock in a fragmental matrix. In a few blocks, fragments of gabbro are more abundant than fragments of volcanic rock. Half of one block is gabbro that abuts with a sharp contact against the pyroclastic breccia. The gabbro in this block is sheared adjacent to the breccia, which contains fragments of gabbro, suggesting primary emplacement of the pyroclastic breccia against the gabbro.

A few breccia blocks are recrystallized, and individual lithic fragments display granoblastic textures and grade into the recrystallized tuff matrix. Most breccia blocks, including those recrystallized, are extensively altered to epidote, alkali feldspar (chiefly albite), chlorite, actinolite(?), and calcite.

DIKES OF BASALT AND ANDESITE IN GABBRO BLOCKS

A few gabbro blocks contain dikes of fine-grained basalt and andesite. Nearly all these dikes are between 1/8 and 2 inches wide; the thickest is 6 inches wide. They are commonly finely crystalline to vitrophyric oligoclase andesite. A few dikes of crystalline basalt and andesite having granoblastic textures also cut the gabbro of several blocks. Poikiloblastic biotite crystals as much as 1 mm in diameter are common in one granoblastic dike, and smaller biotite crystals are common in another. A narrow selvage of gabbro adjacent to one basalt dike is recrystallized in a fashion that suggests the gabbro was melted by the invading basalt. Later thermal recrystallization probably accounts for granoblastic textures and biotite poikiloblasts in the dike rock.

VEINS OF MONZONITE IN GABBRO BLOCKS

Veins and veinlets of fine-grained monzonite and quartz monzonite are contained in several gabbro blocks. A similar vein cuts a dike of recrystallized basalt contained in a gabbro block. The veins and veinlets range in width from about 0.04 mm to about 2.5 cm but are most commonly between 2 and 10 mm wide. The thin

FIGURE 8.—Photomicrographs of quartzose rocks from Udot, Truk Islands. *A*, quartz trachyte block chemically analyzed (analysis 23, Ud 187). Note quartz crystals (Q) and dusty, altered appearance of feldspar laths. (Plane-polarized light, × 43). *B*, same as *A* under crossed nicols. Note polycrystalline quartz aggregate at right × 43. *C*, selvage of recrystallized plagioclase (rp), augite (ra), and magnetite (rmt) between vein of monzonite (M) and gabbro. The gabbro consists of plagioclase (P), augite (A), and magnetite (mt). Note augite of selvage partly engulfed in monzonite. (Plane-polarized light, × 29). *D*, same as *C* under crossed nicols × 29. *E*, contact of gabbro inclusion with vein monzonite. Outer zone of one labradorite crystal (P) is partly albitized (crossed nicols, × 31). *F*, feldspar crystal in quartz monzonite block. Core of crystal in An_{10-15} plagioclase, mantled by albite and perthitic(?) orthoclase (or). Outer zone of albite (ab) encloses orthoclase. Note quartz crystal (Q). (Crossed nicols, × 25).

veinlets branch from the veins. The veins are pale gray (almost white) to medium gray. Average grain size is generally between 0.5 and 1.0 mm, and crystals in the veinlets average between 0.04 and 0.5 mm. Single crystals commonly extend the full width of the thinner veinlets. Marginal chilled zones are lacking, but average crystal size is somewhat smaller on one side of several veins than on the opposite side.

The vein rock is largely a mosaic of irregularly shaped sodic oligoclase crystals, but albite (An_{0-5}), orthoclase, augite, magnetite, and zircon are usually present. Labradorite, quartz, biotite, chlorite, epidote, actinolite, pyrite, and calcite occur in one or more veins.

Most of the vein oligoclase is An_{10-15} in composition, but An_{25} oligoclase is common in one vein. The oligoclase is clear where unaltered but contains irregular patches and veinlets of orthoclase having a dusty, altered appearance. The orthoclase may be a result of exsolution or replacement. Dusty albite (An_{0-5}) and orthoclase extensively replace the primary plagioclase of most veins. X-ray diffraction patterns indicate that the sodic plagioclase is structurally similar to the "low" plagioclase of plutonic rocks; the bulk composition of the orthoclase in the veins is not more sodic than $Or_{85}An+Ab_{15}$ (Stewart, written communication, 1956).

Labradorite xenocysts and inclusions can be seen in several thin sections of the veins (fig. 8E). Most labradorite crystals are albitized marginally and along a network of veinlets.

Augite is the dominate mafic mineral. In one vein the optic angle of the pyroxene is 53°, and α is 1.691, indicating a composition identical (within limits of accuracy measurements) with the augite of the gabbro in the same block. Proportion of quartz ranges from 1 or 2 percent to as much as 10 percent. The quartz commonly occurs as interstitial anhedral crystals, some of which are elongated and irregular in shape. Quartz forms a micrographic intergrowth with the oligoclase of one dike which resembles granophyre in petrographic appearance. Quartz occurs in one vein as lenticular crystals between cleavage lamellae of biotite.

Magnetite, apatite, and zircon are common accessory minerals. Biotite is the principal mafic mineral in one dike. Epidote, calcite, chlorite, and less commonly pyrite are present in altered dikes.

Contact relations of the veins generally appear sharp to the naked eye but are more complex in detail. The gabbro is commonly recrystallized in a discontinuous selvage 0.15 to 0.75 mm wide adjacent to the veins (fig. 8*C–D*). Recrystallized plagioclase crystals in the selvage are clear and generally between 0.01 and 0.03 mm in diameter; they seem to be more sodic (An_{45}) than the plagioclase of the gabbro (An_{60}), which is clouded as if heated (MacGregor, 1931). Augite is also recrystallized in the selvage. Distribution of pyroxene and plagioclase crystals in the recrystallized selvage indicate that the selvage represents melted gabbro. The recrystallized plagioclase of the selvage grades outward into the coarser plagioclase of the veins. Magnetite in the gabbro adjacent to the recrystallized selvage is commonly fringed by reddish-brown biotite that extends the width of the selvage. The gabbro is not appreciably recrystallized adjacent to a few veins and most veinlets. Where the selvage is absent, plagioclase crystals of the gabbro commonly extend as much as 0.05 mm out into the veins. These extensions of the gabbroic plagioclase are generally albitized and have a dusty, altered appearance characteristic of the vein albite. Recrystallized selvages indicate that most veins were emplaced at temperatures high enough to melt gabbro; labradorite xenocrysts and augite similar in composition to most of the gabbro indicate that the veins incorporated some gabbroic material.

The thin veinlets consist almost entirely of albite and orthoclase and generally lack a border zone of recrystallized gabbro. Where a recrystallized selvage is absent, the veinlets lack augite. Where cut by a veinlet, labradorite crystals are albitized and form part of the veinlet. Veinlets of albite fill fractures in pyroxene crystals of the gabbro and indicate that albite is not solely a replacement of labradorite.

MONZONITE BLOCKS

Three small blocks of fine-grained quartz monzonite, 2 to 3 inches in diameter, were found in the breccias on Udot. All blocks consist largely of sodic plagioclase and orthoclase but also contain quartz, hornblende, zircon, apatite, and opaque ores. Quartz forms between 2 and 5 percent of the monzonite.

The finest grained block is somewhat coarser but is otherwise similar to the monzonite veins cutting the gabbro. The average grain size is about 1 mm. The block consists largely of dusty-appearing albite, in which a few ragged unreplaced cores of oligoclase remain. Orthoclase is abundantly distributed throughout the albite as fine veinlets and small areas of irregular shape. The plagioclase is structurally similar to the "low" plagioclase of the plutonic rocks (D. B. Stewart, written communication, 1956). Quartz occurs both as interstitial anhedral crystals and as thin finely crystalline veinlets. Partly assimilated inclusions of volcanic rock are present in the block.

Another block is biotite-hornblende-quartz monzonite, with a grain size of 1 mm, in which largely assimilated inclusions of volcanic rock are common. The hornblende is pleochroic (from pinkish brown to green) and locally grades into marginal hornblende of deep bluish green. Most of the augite and biotite is altered to chlorite. Albite having a dusty, altered appearance is the dominant feldspar, but larger crystals have clear cores of more calcic plagioclase (one found to be labradorite) that are veined with albite, particularly adjacent to quartz, which occurs as anhedral interstitial crystals and polycrystalline aggregates. A zone of dusty albite forms an outer mantle enclosing orthoclase and inner plagioclase.

The third and coarsest block is biotite-augite-hornblende-quartz monzonite having a porous, somewhat miarolitic character and an average grain size of about 2 mm. The larger feldspar crystals consist of clear, unaltered cores of oligoclase (An_{10-15}) veined and enclosed by a thick mantle of albite (An_{0-5}) having a dusty, altered appearance. Orthoclase, which forms about 10 percent of the feldspar, occupies irregular areas within the mantle of dusty albite and is most abundant adjacent to a thin outer mantle of albite about 0.015 mm thick (fig. 8). The quartz occurs chiefly as interstitial crystals and forms an estimated 2 to 3 percent of the rock. Euhedral feldspar and quartz crystals protrude into small miarolitic cavities. Most hornblende crystals are pleochroic from brown to brownish green, but marginal zones and some whole crystals are deeply pleochroic from dark violet to blue green and are probably sodic. Most of the biotite and much of the augite are altered to chlorite.

LIMESTONE FRAGMENTS

A handful of limestone fragments were collected from the breccia in the course of the present study. Most are pale reddish brown, and a few are pale gray. Stylolites are present in most limestone fragments, which are bioclastic (that is, largely organic detritus) and consist largely of sand-size debris. Most detritus is algal, individual fragments of which are as much as 1.5 cm long. Both larger and smaller Foraminifera are common. The larger Foraminifera, identified by W. Storrs Cole, comprise the genera *Lepidocyclina*, *Flosculinella*, and *Miogypsina*. (See earlier discussion on p. 15.) Bryozoan and gastropod detritus has also been identified, and echinoderm(?) spicules are common. One fragment of limestone consists largely of intergrown algae and colonial coral and may be a fragment of a reef. Abundance of bioclastic debris, particularly coarse algal detritus, suggests that this limestone was deposited in shallow water, very probably near a reef.

Sand-size grains of volcanic rock are scattered throughout most of the limestone; the proportion of the volcanic detritus ranges from a few percent in most limestone to about 25 percent in one. Most of the volcanic rock fragments consist of calcite pseudomorphs after plagioclase and primary mafic minerals in a matrix of dark-brown glass. It is not known whether the volcanic detritus is pyroclastic or whether is was washed in as derital sediment. The latter possibility is favored by the similar grain size of volcanic and organic debris, and the dissemination of the volcanic detritus throughout the limestone.

PETROGENESIS

CHEMICAL COMPOSITION AND VARIATION DIAGRAMS

Chemical analyses and norms of representative rocks of the Truk Islands are shown in table 2, together with two previously unpublished analyses of basalt from Ponape. They fall into four main classes: olivine basalt, nepheline and nepheline-melitite basalt, andesite, and trachyte. The transition from basalt to andesite is so gradual that sharp distinctions are not possible. No such transitional zone between andesite and trachyte is apparent. The melilite-nepheline and nepheline basalts are characterized by the presence of feldspathoids. These relations are well shown on the variation diagram (fig. 9) where the weight percent of various oxides are plotted as ordinates against the weight percent of silica. The melilite-nepheline basalts with silica about 39 percent show at the low silica side of the diagram. Olivine basalts and andesites fall closely together near the middle part of the diagram, between 42 and 48 percent silica. In general, there is a close association between silica percentages and color index distinction between basalt and andesite. There are exceptions, however, such as specimen 5 (Du-105), which appears to be a typical olivine basalt with a silica percentage of 44.35 and a color index slightly below 37. There is a wide gap between andesites with silica between 45 and 50 percent and the trachyte with 65 percent silica.

The low silica content of the analyzed andesites and basalts in relation to other oxides is again shown in figure 10.

Figure 10, a triangular *SKM* diagram, shows the composition of the rocks in terms of three components where S, K, and M, are molecular amounts: $S = SiO_2 - 2CaO$, $K = 6(Na_2O + K_2O)$, and $M = MgO + FeO - Fe_2O_3 - TiO_2 - CaO - Na_2O - K_2O + Al_2O_3$. The diagram is a modification of the von Wolff (*OLM*) triangle by Prof. J. B. Thompson, of Harvard University (Schmidt, 1957, p. 152). The median horizontal line represents silica saturation; all rocks that fall above this line contain an excess of silica (oversaturated

TABLE 2.—*Chemical composition and norms*

	Melilite-nepheline basalt		Olivine basalt						Andesite						
	1 To-315	2 To-316	3 Mo-134a	4 Fe-118	5 Du-105	6 Mo-153	7 To-18	8 Mo-8	9 Fe-10	10 Mo-9	11 Mo-160	12 Mo-159	13 To-5	14 Ud-135	15 Mo-13
Bulk															
SiO_2	38.62	38.80	42.00	44.24	44.35	44.70	46.03	46.39	46.80	47.47	47.56	48.10	48.59	48.59	55.55
Al_2O_3	13.03	10.00	11.12	13.23	18.05	11.80	15.22	13.93	15.96	14.73	15.87	14.40	14.82	14.84	16.76
Fe_2O_3	3.21	3.70	4.42	5.10	1.92	1.80	1.84	2.07	1.76	3.31	1.48	4.30	2.89	4.50	3.10
FeO	7.93	7.50	10.31	9.74	7.05	11.10	9.77	10.60	9.02	8.25	8.00	8.00	8.49	6.88	4.41
MgO	10.72	12.50	11.44	8.68	5.59	12.00	8.55	11.22	5.04	4.57	4.34	6.40	4.23	3.90	2.67
CaO	13.80	14.70	9.68	8.82	10.14	10.20	9.91	9.11	9.48	8.52	8.28	9.50	8.39	8.45	4.64
Na_2O	3.16	3.30	2.65	2.48	6.36	1.90	3.64	2.44	4.99	5.39	3.88	3.20	5.93	5.39	5.99
K_2O	1.19	1.20	1.35	.90	1.12	.54	.90	.38	1.12	1.17	1.19	.69	1.18	1.13	2.13
H_2O-	1.14	2.50	.24	.40	.56	2.30	.69	.44	.99	.76	.88	2.20	.88	.66	1.09
H_2O+	2.84		2.90	3.20	1.85		1.30	1.11	1.73	1.72	3.90		2.40	1.61	.76
TiO_2	3.58	3.10	2.86	2.57	1.20	2.60	2.00	1.70	2.30	3.15	3.33	2.50	2.21	2.91	1.83
CO_2	----	.05	----	----	----	.05	----	.12	----	----	----	.05	----	----	.03
P_2O_5	.64	1.10	.82	1.03	1.13	.42	.53	.21	.07	.71	.90	.51	.34	.62	.75
MnO	.18	.20	.19	.17	.23	.16	.29	.19	.30	.24	.19	.19	.30	.24	.24
S	----	----	----	----	----	----	----	----	----	----	----	----	----	----	----
Total	100.04	98.65	99.98	100.56	99.55	99.57	100.67	99.91	99.56	99.99	99.80	100.04	100.65	99.72	99.95
Niggli															
Q	----	----	----	----	----	----	----	----	----	----	----	----	----	----	1.3
or	----	----	8.5	5.5	6.5	3.0	5.5	2.5	7.0	7.0	7.5	4.0	7.0	7.0	12.5
ab	----	----	12.3	23.5	12.8	17.0	24.8	22.5	25.5	35.5	37.0	29.5	32.5	38.0	54.5
an	18.3	9.3	15.3	20.8	17.5	23.0	22.5	26.0	18.0	13.0	23.3	24.0	10.7	13.2	12.5
lc	6.0	6.0	----	----	----	----	----	----	----	----	----	----	----	----	----
ne	17.4	18.0	7.1	----	26.5	----	5.5	----	12.0	9.0	----	----	13.2	7.2	----
di wo	13.8	16.3	11.9	7.6	10.2	10.6	7.5	7.2	10.0	10.4	6.6	9.0	11.9	10.4	.6
di en	11.6	14.2	9.2	5.6	6.8	7.6	5.1	5.0	6.0	6.5	4.1	6.6	6.9	8.0	.4
di fs	2.2	2.1	2.7	2.0	3.4	3.0	2.3	2.2	4.0	3.9	2.5	2.4	5.0	2.5	.2
hy en	----	----	----	10.0	----	4.8	----	4.6	----	----	1.3	9.6	----	----	7.0
hy fs	----	----	----	3.4	----	1.9	----	1.7	----	----	.7	3.9	----	----	2.4
ol fo	14.3	16.1	17.8	7.5	10.3	16.5	21.0	16.5	10.8	4.9	5.5	1.6	5.5	3.5	----
ol fa	2.6	2.3	5.1	2.7		6.2		7.0			3.4	.6			
mt	4.1	5.1	4.6	5.7	2.0	2.0	----	2.2	1.8	3.6	1.5	4.6	3.1	4.8	3.3
il	3.5	3.9	4.0	3.6	1.6	3.6	2.8	2.2	3.2	4.6	4.8	3.4	3.2	4.2	2.4
hm	1.3	2.4	----	----	----	----	1.9	----	----	----	----	----	----	----	----
ap	5.0	4.4	1.6	2.1	2.4	.8	1.1	.5	1.6	1.6	1.8	.8	.8	1.3	2.9
C	----	----	----	----	----	----	----	----	----	----	----	----	----	----	----
Ru	----	----	----	----	----	----	----	----	----	----	----	----	----	----	----
Pr	----	----	----	----	----	----	----	----	----	----	----	----	----	----	----
Total	100.1	100.1	100.1	100.0	100.0	100.0	100.0	100.1	99.9	100.0	100.0	100.0	99.8	100.1	100.0

1. Melilite-nepheline basalt flow south side of north Tol, near top of mesa. Analysts, Drs. Ikawa and Asari, Japan Insp. Co., Tokyo.
2. Melilite-nepheline basalt flow south side of north Tol, near top of mesa. Analysts, P.L.D. Elmore and K. E. White, U.S. Geol. Survey.
3. Olivine basalt flow 60 ft. altitude at southeast edge American base, Moen. Analysts, Drs. Ikawa and Asari, Japan Insp. Co., Tokyo.
4. Olivine basalt flow 200 ft. altitude, west Fefan. Analysts, Drs. Ikawa and Asari, Japan Insp. Co., Tokyo.
5. Biotite-bearing olivine basalt flow 150 ft. altitude, Mount Tolomon, Dublon Island. Analysts. Drs. Ikawa and Asari, Japan Insp. Co., Tokyo.
6. Olivine basalt flow; thick flow at east end Moen. Analysts, P. L. D. Elmore and K. E. White, U.S. Geol. Survey.
7. Olivine basalt flow, south coast of south Tol. Analysts, Drs. Ikawa and Asari, Japan Insp. Co., Toyko.
8. Olivine basalt flow, west side Wichap village, Moen. Analyst, Lois Trumbull, U.S. Geol. Survey.
9. Andesite flow, base of trail, Feurep village, Fefan. Analysts, Drs. Ikawa and Asari, Japan Insp. Co., Tokyo.
10. Andesite flow, Elelup village, Moen. Analysts, Drs. Ikawa and Asari, Japan Insp. Co., Tokyo.
11. Andesite flow, Mount Tonaachau, Moen. Analysts, Drs. Ikawa and Asari, Japan Insp. Co., Tokyo.
12. Andesite flow, east end of old Japanese airfield, south end Moen. Analysts, P. L. D. Elmore and K. E. White, U.S. Geol. Survey.
13. Andesite dike, trail above Wonip village, Tol. Analysts, Drs. Ikawa and Asari, Japan Insp. Co., Tokyo.
14. Hornblende-bearing andesite flow. Udot. Analysts, Drs. Ikawa and Asari, Japan Insp. Co., Tokyo.

rocks) and all rocks below it are deficient in silica (undersaturated rocks) and contain one or more minerals with less silica than is required for saturation. The end points of this line represent normative alkali feldspar and total normative pyroxene of the von Wolff triangle. The *SKM* triangle has the advantage that lines connecting normative minerals become valid phase boundaries with removal of the calcic feldspars and pyroxenes. The diagram becomes a partial graphic solution of the norm, and the composition of any rock can be expressed in terms of a combination of three normative mineral components depending on where it falls in the diagram. All the analyzed Truk rocks fall below the silica saturation line. All but the melilite-nepheline basalts, specimens 1(To-315) and 2(To-316) fall into subsidiary triangles, alkali feldspar-olivine-hypersthene or alkali feldspar-olivine-nepheline.

Figures 11 and 12 are triangular diagrams or–ab–an and wo–en–fs to show, respectively, the composition of the normative feldspar and the normative pyroxene of the volcanic rocks of the Truk Islands, plotted with normative feldspar and normative pyroxene of Daly's average rock types. Both diagrams reflect the low silica percentage of the Truk rocks. The apparent undersaturation in the diagrams, plus the abundance of modal olivine in all the basalts (averaging from 10 to 15 percent), and the presence of olivine in all the andesites and in many of the trachytes indicate that the Truk flows were deficient in silica. The lavas exposed at the surface in Truk represent only the upper *of volcanic rocks of Truk Islands and Ponape*

Sodic trachyte	Hornfelsic block	Gabbro block	Basalt in gabbro block	Gabbro block	Granophyre gabbro block	Monzonite block	Trachyte block	Pumice	Basalt	
16 Mo-208	17 Ud-141a	18 Ud-122a	19 Ud-139c	20 Ud-125	21 Ud-126	22 Ud-146	23 Ud-187	24 Ud-138	25 Po-18	26 Po-15
analysis										
65.00	41.24	45.62	46.32	48.50	58.22	63.60	67.61	69.16	42.42	45.70
17.50	16.50	19.08	15.75	22.70	19.59	16.90	15.68	13.53	13.64	15.45
2.10	7.51	5.44	6.68	3.90	1.00	3.09	3.02	.97	3.60	1.82
.93	8.31	4.93	7.11	3.50	2.65	1.72	1.53	2.83	7.88	9.36
.52	6.03	4.63	5.30	3.10	1.12	.61	.85	.39	11.06	6.48
1.20	12.54	13.51	10.40	11.40	3.20	1.30	.36	2.04	11.71	9.46
8.10	1.41	2.30	3.11	3.50	5.62	5.54	5.42	4.92	2.64	2.84
3.60	.15	.25	.34	.35	4.36	3.58	2.51	2.99	.82	1.23
} .38	.18	.06	.04	} 1.00	.26	.48	.34	.14	.82	.42
	1.66	2.71	2.10		1.51	1.91	1.51	1.90	3,49	3.53
.79	3.68	.77	2.40	1.70	1.27	.49	.42	.43	1.79	3.05
.05	--	--	--	.86	--	--	--	tr	--	--
.23	.03	.24	.61	.30	.96	.03	.05	.24	.73	.79
.12	.19	.12	.22	.13	.07	.06	.06	.09	.19	.18
--	--	--	.10	--	--	.03	--	--	--	--
100.52	99.43	99.66	100.48	100.94	99.83	99.31	99.36	99.63	100.79	100.31
Norms										
2.4	0.5	2.7	2.1	0.2	1.7	13.9	23.9	22.0	--	--
20.5	1.0	1.5	1.5	2.0	26.0	21.5	15.0	18.0	5.0	7.5
70.5	13.0	22.5	28.5	31.5	50.5	50.5	50.0	45.0	8.1	27.0
.8	41.0	32.0	29.0	46.0	10.0	4.3	--	6.2	22.7	26.8
--	--	--	--	--	--	--	--	--	--	--
--	--	--	--	--	--	--	--	--	9.8	--
1.2	9.7	15.9	8.0	3.7	--	--	.2	1.1	10.7	6.8
1.2	8.3	13.0	7.2	3.3	3.2	--	.2	.2	13.8	4.6
--	1.4	2.9	.8	.4	1.7	--	--	.9	3.0	2.2
.2	9.5	1.2	7.8	5.3	--	1.6	2.2	1.0	--	2.2
--	1.6	.3	.8	.7	--	.4	--	3.7	--	1.2
--	--	--	--	--	--	--	--	--	15.5	9.0
--	--	--	--	--	--	--	--	--	8.0	4.5
.9	8.4	6.3	7.2	4.1	1.1	3.5	3.0	.9	3.9	2.0
1.0	5.4	1.2	3.4	2.2	1.8	.6	.6	.6	2.4	4.4
.8	--	--	--	--	--	--	.2	--	--	--
.6	.2	.6	1.3	.6	1.9	.5	.3	.6	1.6	1.6
--	--	--	--	--	2.1	2.6	--	--	--	--
--	--	--	--	--	--	--	4.5	--	--	--
--	--	--	2.4	--	--	.7	--	--	--	--
100.1	100.0	100.1	100.0	100.0	100.0	100.1	100.1	100.2	100.0	99.8

15. Andesite, quarry east end of north Moen, airstrip. Analyst, Lois Trumbull, U.S. Geol. Survey.
16. Trachyte flow, top Mount Witipon, Moen. Analysts, P.L.D. Elmore and K. E. White, U.S. Geol. Survey.
17. Hornfelsic block in breccia, Udot. Analysts, Drs. Ikawa and Asari, Japan Insp. Co., Tokyo.
18. Gabbro block in breccia, Udot. Analysts, Drs. Ikawa and Asari, Japan Insp. Co., Tokyo.
19. Basalt in gabbro block, breccia, Udot. Analysts, Drs. Ikawa and Asari, Japan Insp. Co., Tokyo.
20. Gabbro block from breccia bed, south side of central Udot. Analysts, P.L.D. Elmore and K. E. White, U.S. Geol. Survey.
21. Granophyre block, breccia, south side Udot. Analysts, Drs. Ikawa and Asari, Japan Insp. Co., Tokyo.
22. Monzonite block, breccia, Udot. Analysts, Drs. Ikawa and Asari, Japan Insp. Co., Tokyo.
23. Trachyte block, breccia, Udot. Analysts, Drs. Ikawa and Asari, Japan Insp. Co., Tokyo.
24. Recent pumice, beach on north side Udot. Analysts, Drs. Ikawa and Asari, Japan Insp. Co., Tokyo.
25. Basalt flow, Ponape. Analysts, Drs. Ikawa and Asari, Japan Insp. Co., Tokyo.
26. Basalt flow, Ponape. Analysts, Drs. Ikawa and Asari, Japan Insp. Co., Tokyo.

part of the shield volcano and do not necessarily prove that the principal mass of the volcano, now submerged, consists of undersaturated lavas. The flows in Truk are nearly horizontal and represent but a small fraction of the volcano during its final stages of growth.

COMPARISON WITH ROCKS OF OTHER AREAS

Powers (1955, p. 80) in discussing the composition and origin of the basaltic magmas of the Hawaiian Islands points out that olivine basalt is a product of the declining phase of the shield-building eruptions and that the volume of such flows is insignificant compared with that in primitive shields. He concludes that the concept that the parent magma of the primary Hawaiian magma was undersaturated is not justified and that the composition of the olivine basalt in the primitive shields indicates that the primary Hawaiian magma is approximately saturated with silica and should not be called an olivine basalt magma (as recognized earlier by Tilley, 1950, p. 40). The lavas of the Truk Islands are petrographically and chemically similar to the lavas of the declining phase on Hawaii. Whether tholeiitic lavas form the primitive shield of Truk is unknown, as only the upper part of the shield is now above sea level.

Previous studies of the volcanic rocks of Saipan (Schmidt, 1957, p. 153–163) and of Guam by Stark show comparisons of the compositions of the lavas of these islands with those of Palau, Bonin, and Volcano Islands, the Hakone volcano and Izu Peninsula region

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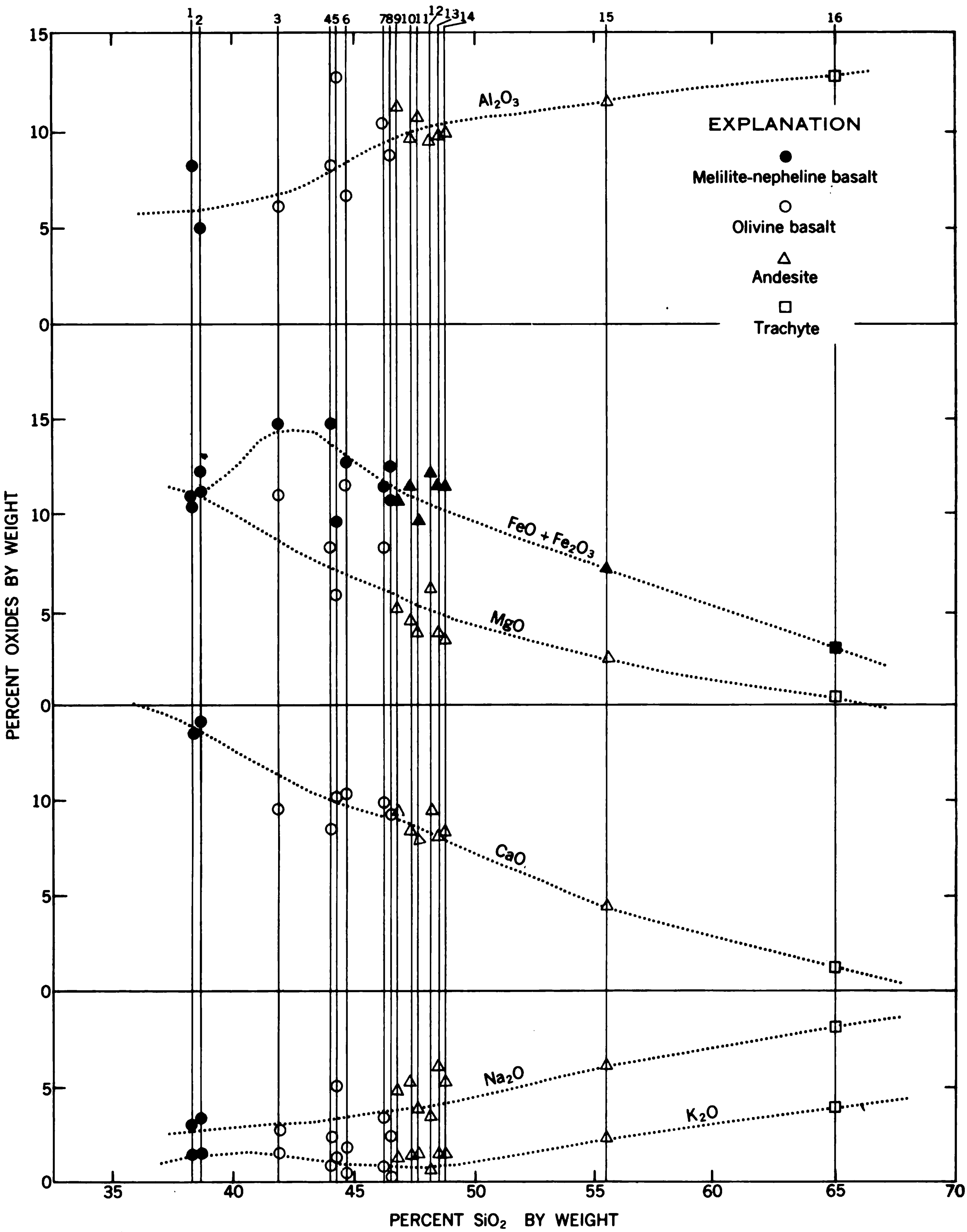

FIGURE 9.—Variation diagram for volcanic rocks of Truk Islands (Numbers refer to analyses and corresponding rock specimens in table 2).

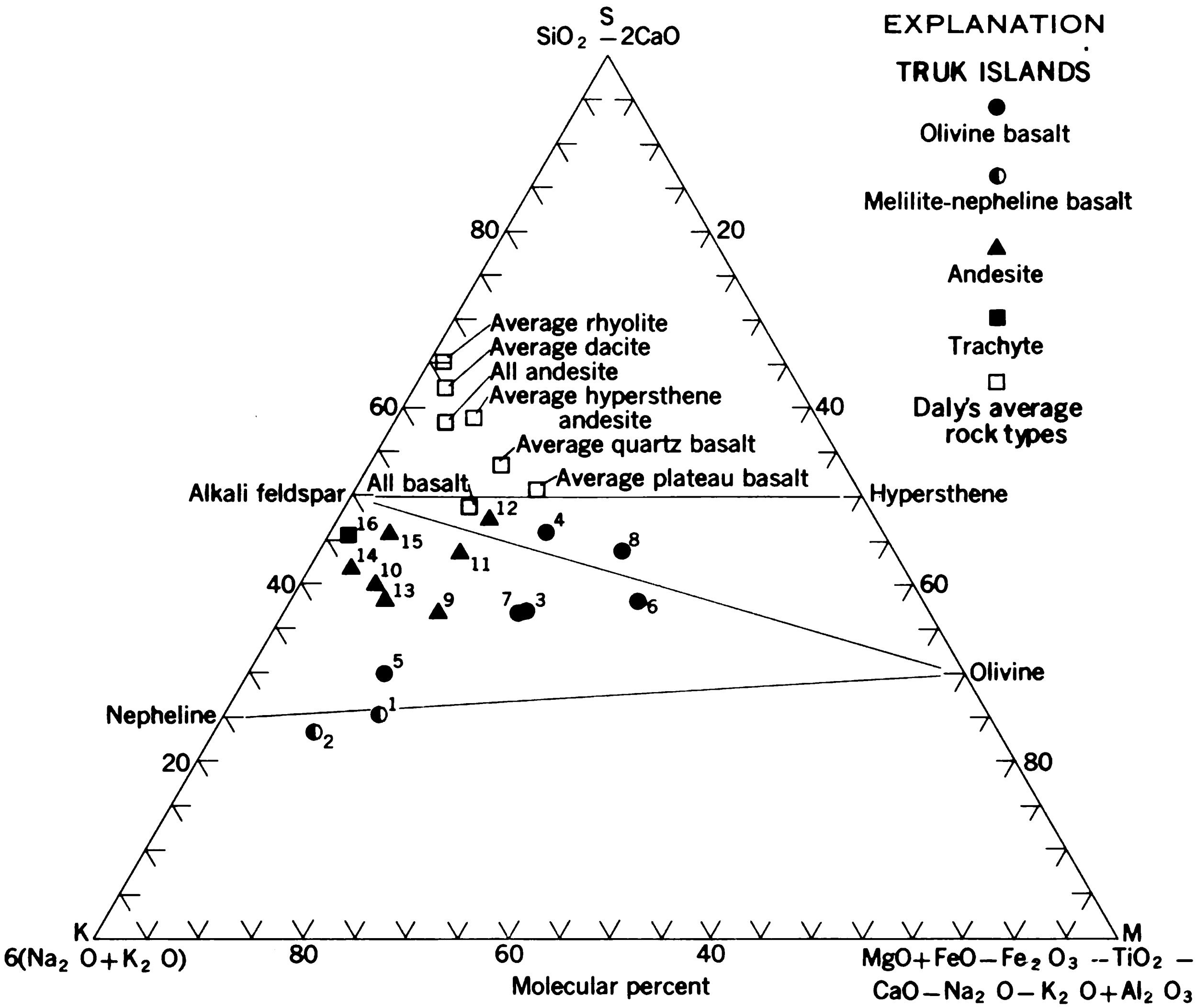

FIGURE 10.—*SKM* diagram of specimens of volcanic rocks of Truk Islands and Daly's average rock-types.

of Japan, and the Hawaiian Islands. All except the Hawaiian Islands lie west of the andesite line, in the zone of continental-type rocks of the Pacific margin. A close relation was pointed out between the islands of Saipan and Guam, and to the similarity in composition of their lavas with those of the Izu Peninsula region of Japan. The Truk Islands lie east of the andesite line and are closely related to the oceanic-type rocks of the Pacific basin. A comparison of the average composition of rocks of the Truk Islands with average compositions of the rocks of Guam (Stark, in prep., table 2) and Hawaii (Macdonald, 1949, table 5) is shown in figure 13. The analyses of two basalts from Ponape are also shown.

In the triangular *SKM* diagram (fig. 13), there is a close relation shown between the average andesite of the Truk Islands and the average andesine andesite of Hawaii. The average olivine basalt of Truk is slightly more mafic than the average olivine basalt of Hawaii, and both fall below the median alkali feldspar-hypersthene join. The melilite-nepheline basalts of Truk and Hawaii fall close to each other on the diagram, just below the nepheline-olivine join. The average andesite and average basalt of Guam fall well above the median alkali feldspar-hypersthene join or above the line of silica saturation. The trachyte of Truk and Hawaii fall close together just below this line and close to the alkali feldspar side of the *SKM* triangle.

Figures 14 and 15 are triangular diagrams or–ab–an and wo–en–fs, respectively, and show the relations of the normative feldspar and normative pyroxene of the rocks of Truk, Hawaii, Guam, and Ponape. In each diagram the similarity in composition between the lavas of Truk and Hawaii is indicated; however, certain significant differences occur. Macdonald (1949, p. 1545–1549) reports abundant pigeonite in the olivine

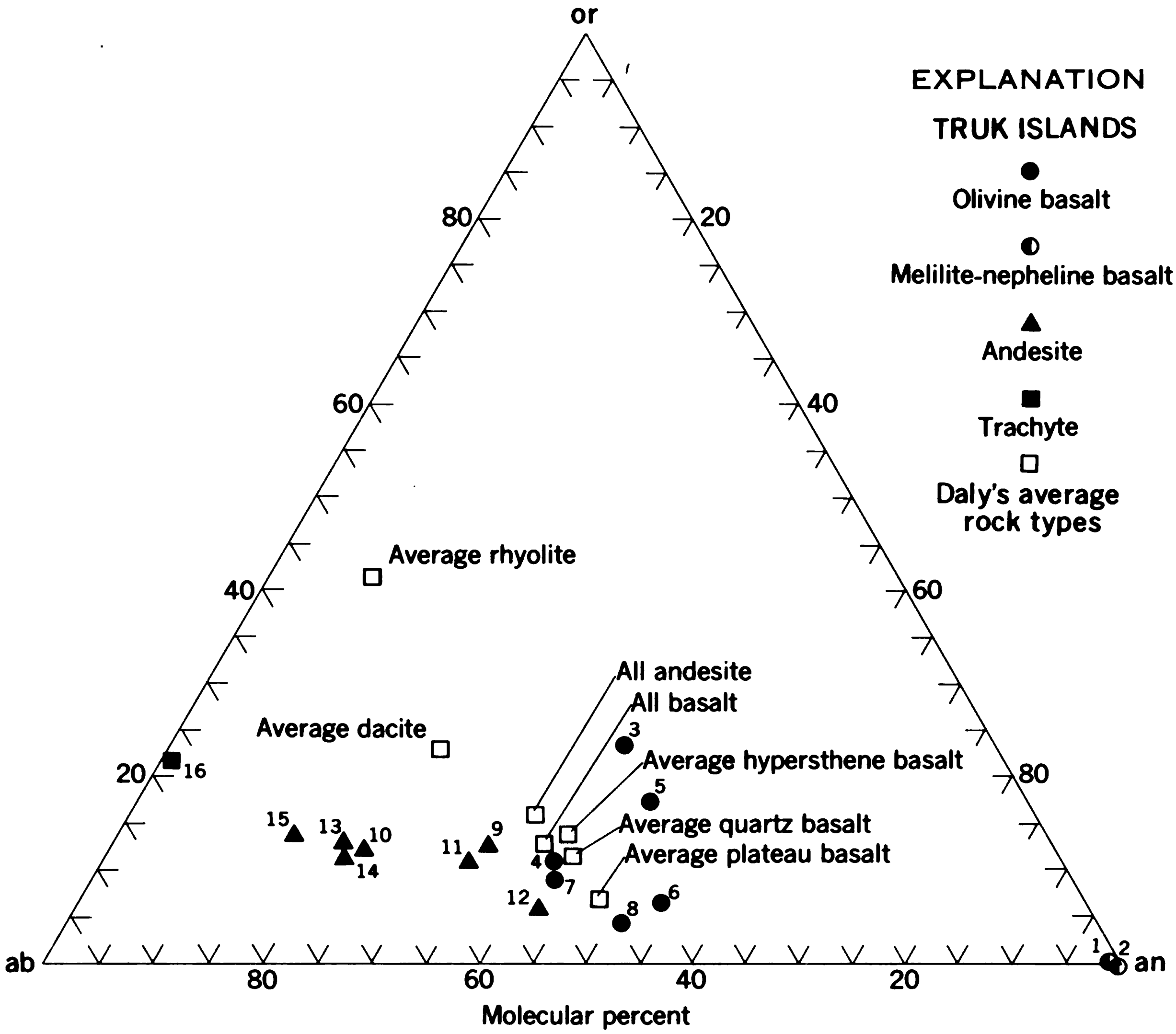

FIGURE 11.—Composition of normative feldspars of volcanic rocks of Truk Islands and Daly's average rock types.

basalt, basalt, and andesine andesite of Hawaii. Although some of the pyroxene in the Truk rocks have optic angles slightly lower than typical augites, in none of the measurements made is the optic angle small enough to identify pyroxene as pigeonite. Figure 16 is a triangular FeO–alkali–MgO diagram, with total iron plotted as FeO. The curve represented is similar to the curves of olivine basalt-trachyte series plotted elsewhere by Nockolds and Allen (1954, p. 246, 248). The nonporphytitic basalt lying closest to the MgO point represents the liquid closest in composition to the primary magma of the Truk Islands lavas. Chemically this basalt is essentially similar in all respects to the parental magmas of the olivine basalt-trachyte series as shown by Nockolds and Allen. (See table 3.)

CONCLUSIONS

EASTERN AND WESTERN ISLANDS

Various origins have been suggested to account for the parent magma of the olivine basalt-trachyte series. Some have advocated derivation from tholeiitic magma, a view recently reiterated by Powers (1955). Others (Kennedy, 1933) have postulated two basaltic layers in the crust, one of which yielded the alkalic olivine basalt magma series. Seismologists have recently shown that the Pacific Ocean is underlain by a basaltic layer only 3 to 5 kilometers thick (Raitt, 1956); this figure throws doubt on the feasibility of deriving oceanic volcanoes from a basaltic layer. Kuno and others (1957, p. 212–216) have summarized evidence that sug-

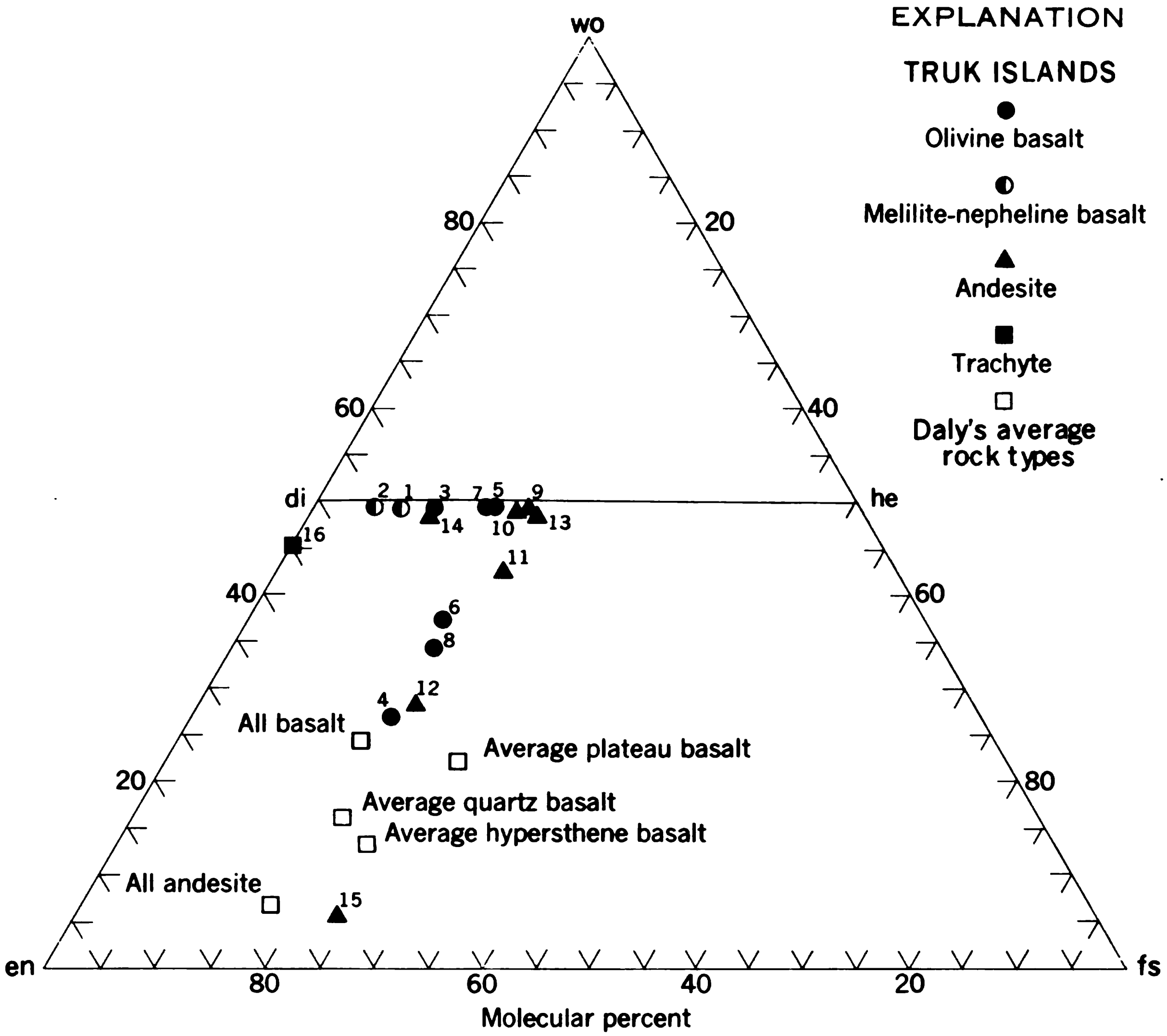

FIGURE 12.—Composition of normative pyroxenes of volcanic rocks of Truk Islands and Daly's average rock types.

TABLE 3.—*Chemical composition of nonporphyritic basalt of the Truk Islands, compared with basalts assumed by Nockolds and Allen (1954, p. 282) to represent primary magma of the olivine basalt-trachyte magma series*

	Truk Islands [1]	*Scottish Tertiary*	*Hawaiian*	*Polynesian*
SiO_2	47.2	46.1	46.7	46.0
TiO_2	1.7	2.5	3.0	2.9
Al_2O_3	14.2	15.1	15.3	16.9
Fe_2O_3	2.1	3.0	3.8	4.6
FeO	10.8	10.4	8.2	7.4
MnO	.2	.3	.1	.2
MgO	11.4	10.4	8.5	7.3
CaO	9.3	9.3	10.5	10.9
Na_2O	2.5	2.5	2.6	2.5
K_2O	.4	.3	1.0	1.0
P_2O_5	.2	.2	.3	.3

[1] Analysis by Lois Trumbull, calculated on a water-free basis.

gests the olivine basalt parent magma originated by fractional melting of peridotite at considerable depth in the earth's mantle. The present study offers little evidence if any on this controversial and fundamental problem.

CENTRAL ISLANDS

ANDESITE AND BASALT

The andesite and basalt lavas and dikes of the central islands are petrologically similar to those of other Truk Islands and indicate derivation from a similar (or the same) alkalic, undersaturated olivine basalt magma. Hornblende, however, is more common in andesite on Udot and Eot than elsewhere, a difference possibly reflecting a higher content of volatiles in the andesite erupted from the central conduit than in that erupted from lateral fissures. Some of the andesites and basalts are albitized, and a few contain quartz amygdules.

Most andesite and basalt blocks in the breccia resemble the lavas and dikes, but some are different. A small proportion of the blocks are more highly altered than the surface dikes and flows, and a few are coarser. The coarser rock crystallized below the surface, prob-

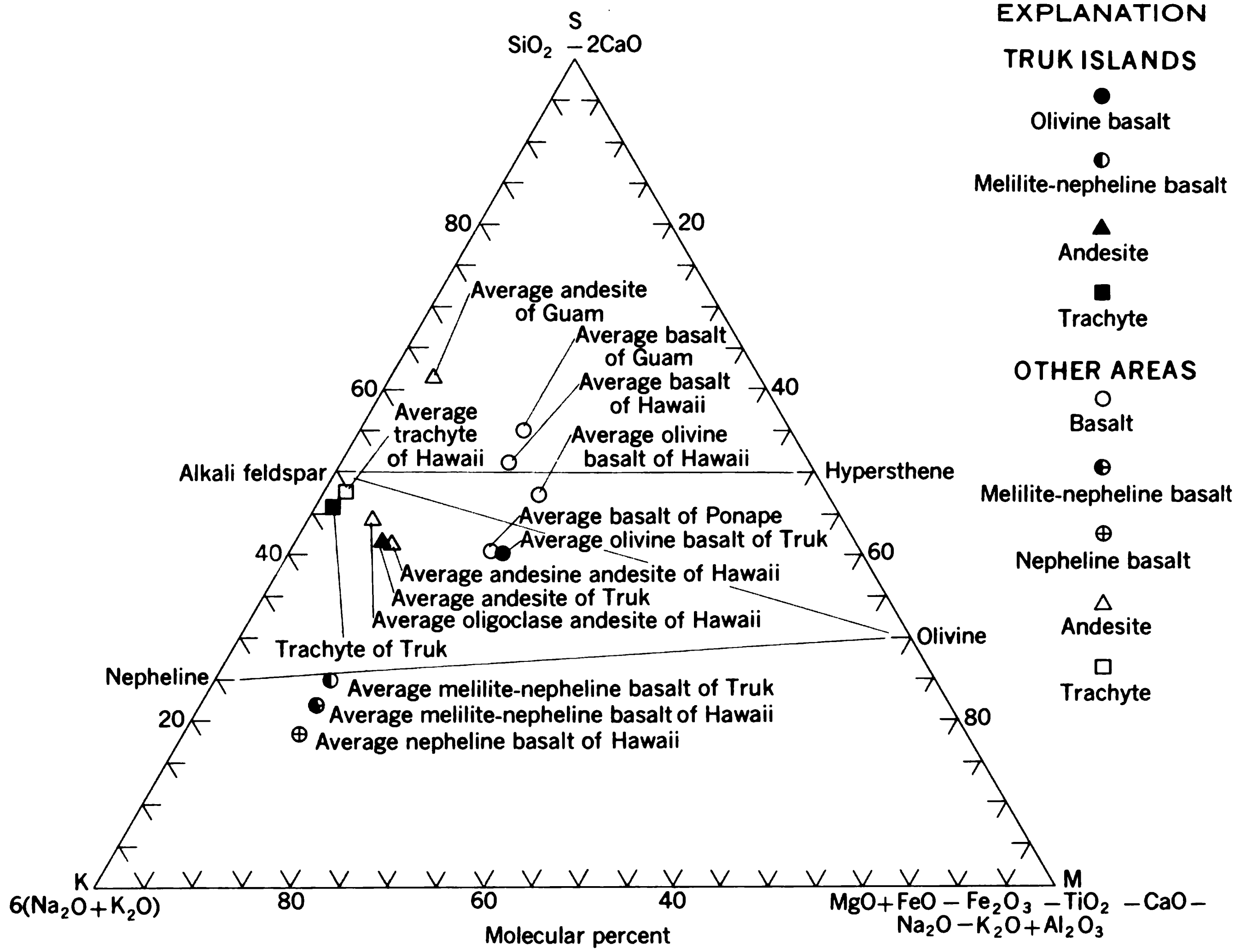

FIGURE 13.—SKM diagram of average volcanic rocks of Truk Islands and other areas.

ably in the central conduit. One block of coarse olivine basalt (or olivine dolerite) porphyry contains a higher proportion of olivine and plagioclase more calcic than any surface dikes or flows studied. This block may have been broken from a layer of settled crystals in the lower part of either the conduit or a magma chamber.

TRACHYTE AND QUARTZ TRACHYTE

Many of the trachyte lava flows, dikes and ejected blocks of the central islands contain quartz that is almost certainly primary, and thus they represent a magma oversaturated in silica. Texturally, these trachytes resemble the quartz-free trachytes of the eastern and western groups of the Truk Islands. The feldspar of the trachytes of the central islands is, however, a dusty complex of "low" albite and potassic feldspar (probably orthoclase), in contrast to the anorthoclase of trachytes on the other islands. Orthoclase mantles and replaces plagioclase in the Udot trachyte, but this fact does not prove that the orthoclase was postmagmatic. The normative plagioclase and content of normative orthoclase is about the same in the analyzed trachytes of Udot and Moen; hence, the "low" albite and potassic feldspar in the trachyte of Udot may have exsolved from anorthoclase under the influence of gases or solutions. The potassic feldspar may have been redistributed to mantle and replace plagioclase.

Origin of the quartz in these rocks is a petrologic problem, as all evidence points to an undersaturated alkalic olivine basalt primary magma for the Truk volcano. Crystallization of olivine and pyroxene in magma of this composition is eutectic (Hess and Poldervaart, 1951, p. 477) and lacks the reaction relation that enriches the residual magma in silica during fractional crystallization of tholeiitic basalt. This view has been recently reiterated by investigators of the

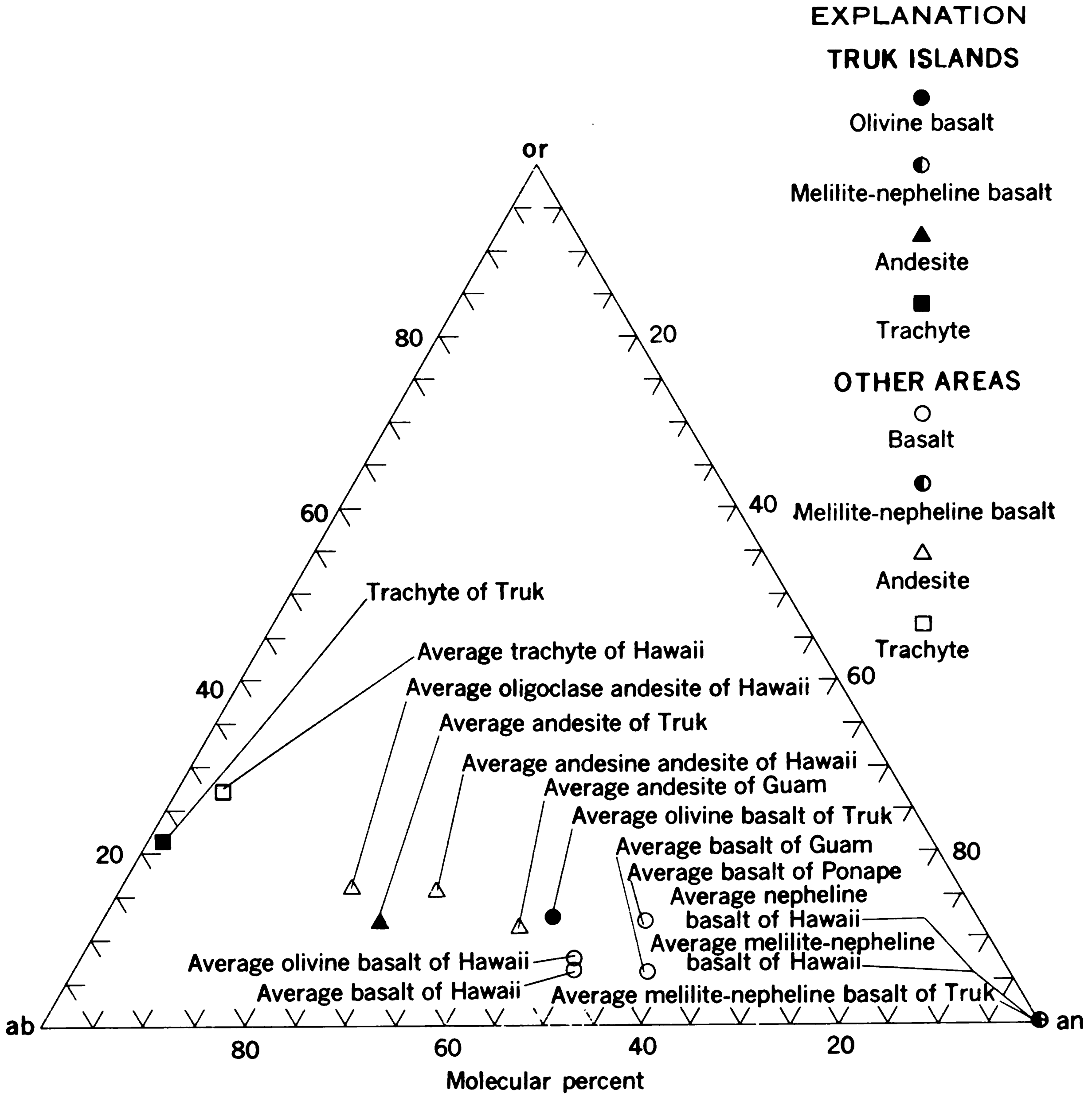

FIGURE 14.—Composition of average normative feldspar of volcanic rocks of Truk Islands and other areas.

Geophysical Laboratory (Geophysical Laboratory, 1956, p. 169; Thornton and Tuttle, 1956). Despite this theoretical objection, Tilley (1950 p. 43) has maintained that small amounts of a siliceous alkalic residuum are produced by differentiation of undersaturated alkalic olivine basalt magma. In support of Tilley's argument, the average trachyte of the Hawaiian Islands contains 4 percent normative quartz (Macdonald, 1949, table 5, p. 1571), and one trachyte contains nearly 20 percent normative quartz. Tholeiitic basalt is now believed to form the bulk of the Hawaiian shield volcanoes, but the trachytes are more closely related in time, space, and chemical composition to undersaturated alkalic lavas than to the tholeiitic ones. Two other possible origins of the oversaturated magma on the Truk Islands will be considered as crystallization differention meets with a strong theoretical objection. The silica may have been concentrated in the magma by volatile transfer or thermal water (Fenner and Day, 1931). Secondary quartz in altered lavas and dikes was indeed transferred by gases or hydrothermal solutions, but this transfer does not prove volatile or hydrothermal concentration of the quartz. Conceivably the silica in rocks of the central islands was liberated

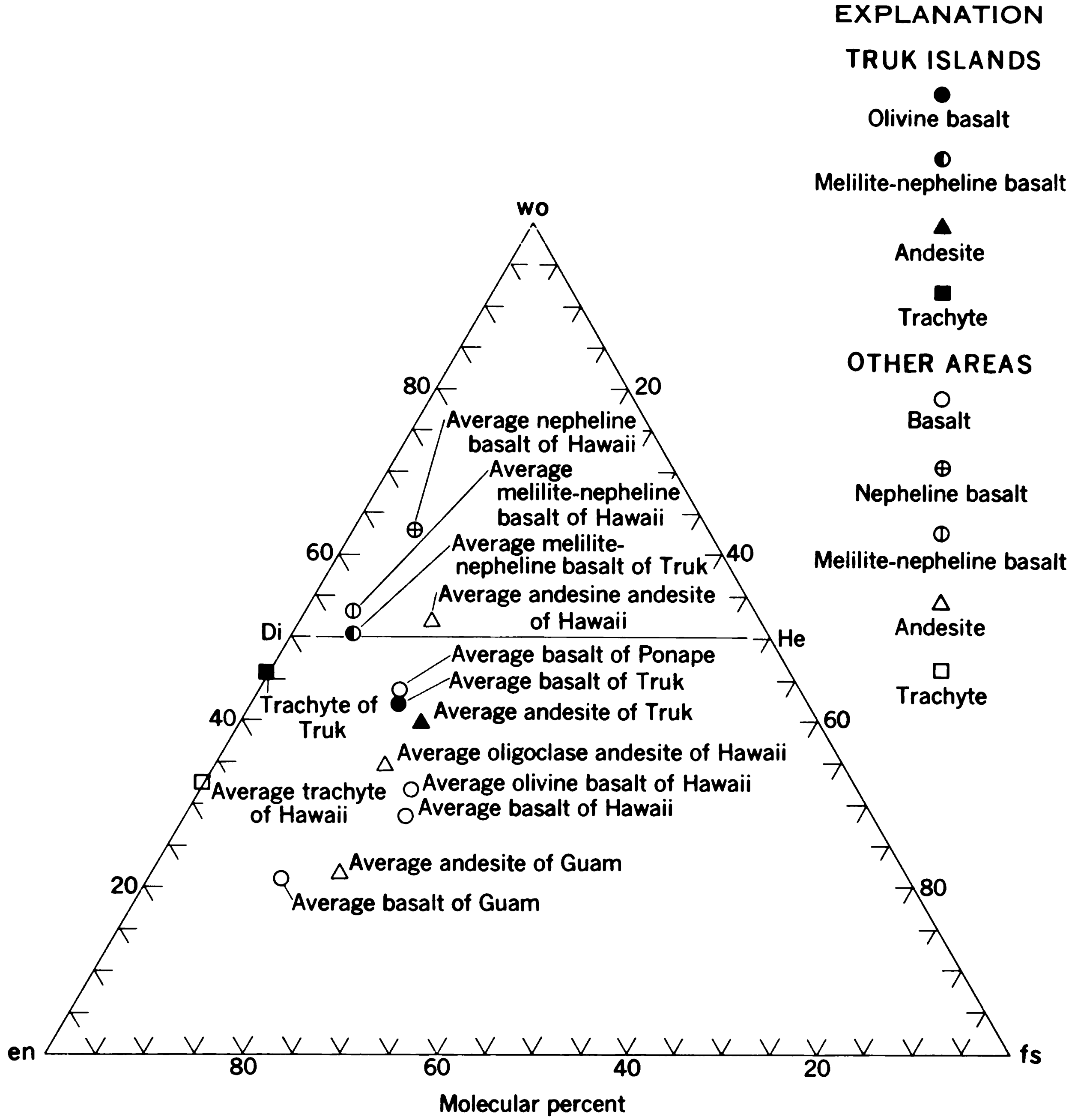

FIGURE 15.—Composition of average normative pyroxenes of volcanic rocks of Truk Islands and other areas.

by oxidation of ferrous ion (in silicates) and by the recombination of the ferrous ion into magnetite during thermal recrystallization of rocks adjacent to a magma chamber or the central conduit. This suggestion apparently receives some support from the abnormally low silica and high magnetite content of one block of recrystallized basalt that was chemically analyzed (specimen 17–Ud 141a). In conclusion, the writers believe that present data are inadequate to determine the process by which oversaturated alkalic magma of the central islands has been produced from the primary undersaturated parent magma.

GABBRO BLOCKS

Origin of the gabbro blocks on Udot and Eot is of interest in view of recent conclusions that peridotite in some lavas of the Hawaiian Islands may represent fragments of a peridotite mantle within the earth (Ross and others, 1954). Geologic evidence suggests, how-

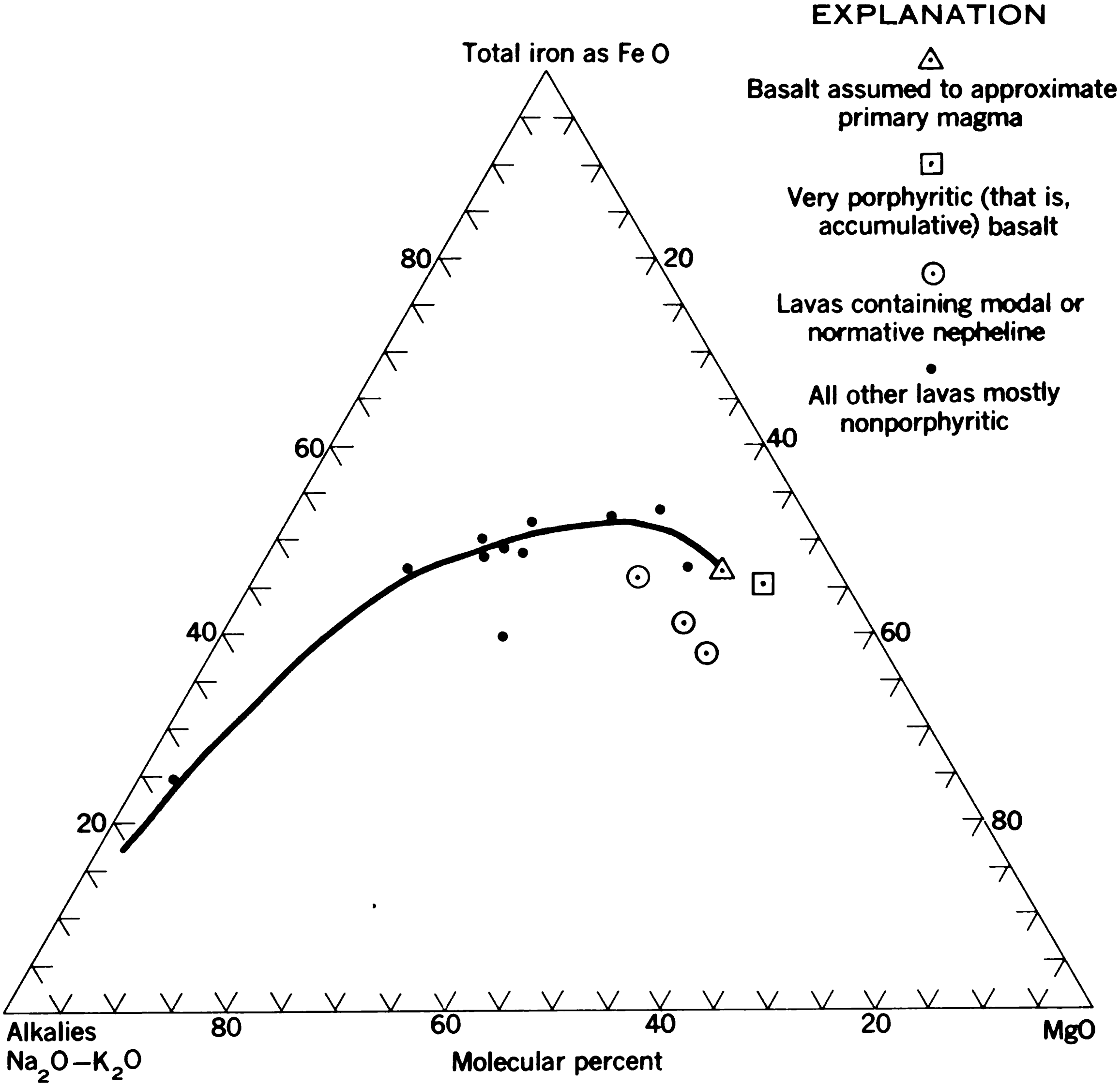

FIGURE 16.—Triangular FeO–alkali–MgO diagram of Truk Islands lavas. Total iron plotted as FeO.

ever, that the gabbro xenoliths of the Truk Islands were torn from an intrusion within the volcano itself.

The gabbro blocks contain nearly all the same minerals as do the basaltic lavas of Truk and indicat^ that the gabbro was olivine bearing and hence undersaturated in silica (as are the basalts). Some of the pyroxene in both basalt and gabbro is titaniferous, and the nontitaniferous pyroxene of both is augite of similar composition. These facts suggest the same or a similar primary magma for both basalt and gabbro. The color index (that is, proportion of mafic minerals) of the gabbro is, however, lower than that of basalts having the same plagioclase composition; this relation suggests some settling of mafic minerals from gabbro magma before intrusion. Chemical analyses of slightly altered gabbro are given in table 2.

Zoned plagioclase crystals indicate that cooling was too rapid for establishment of chemical equilibrium in the cooling magma; this rapid cooling favors interpretation as an intrusive stock rather than a crustal layer within the earth.

The fine- to coarse-grained texture and general lack of mineral orientation can most easily be explained by assuming the gabbro blocks to be fragments of an intrusive stock rather than a crustal layer. Linear orientation of plagioclase in a few blocks suggests magmatic flow rather than crystal settling.

Gabbro fragments occur only in the breccia; none were observed in the lavas. Intrusive relations of pyroclastic breccia to gabbro were observed in composite xenoliths; these relations suggest that the breccia was formed at or below the level of the gabbro. Pyroclastic breccia was more likely formed in the throat of the volcano than in a crustal layer 16,000 or more feet below the surface; hence, the gabbro is probably present in the volcano rather than in a crustal layer.

The hornfelsic, recrystallized gabbro and probably the gabbro in which the plagioclase is turbid were heated, possibly by magma rising in the central conduit penetrating the intrusive gabbro body. Some of the gabbro was actually remelted.

The epidote and actinolite of altered gabbro are probably hydrothermal in origin and were probably produced by aqueous solutions in and near the central conduit. Sodic plagioclase, potassic feldspar, and quartz were introduced in some blocks lacking alteration to epidote and actinolite, but may also be of hydrothermal or pneumatolytic origin.

RECRYSTALLIZED BASALT BLOCKS

The blocks of recrystallized basalt are probably fragments of lava flows that were adjacent to the central conduit and heated by rising magma. Inclusions of recrystallized basalt in the gabbro blocks are probably xenoliths incorporated by the gabbro magma during intrusion. A chemical analysis of recrystallized basalt is given in table 2, specimen 17 (Ud–141a). The chemical composition and normative minerals suggest that this rock was formerly olivine basalt. The silica percentage is considerably lower than that of any of the analyzed lavas, however, and suggests the block became desilicated (through oxidation of iron?) during recrystallization. Possibly release of silica during thermal recrystallization may have been a source of some secondary silica on the Truk Islands.

DIKES OF ANDESITE AND BASALT IN GABBRO BLOCKS

These dikes indicate that magma similar to that of the dikes and flows at the surface penetrated fractures in the gabbro at depth. A chemical analysis of a partly recrystallized andesine basalt dike contained in a gabbro block is given in table 2 (specimen 19, Ud–139c). The dike is similar in most respects to an andesine basalt flow of Moen, table 2 (specimen 12, Mo–159). As previously discussed, gabbro was melted adjacent to some of the dikes; the melting suggests intrusion of the dike rock at high temperature. Many of these dikes are recrystallized, indicating later heating of the dike rock. Biotite was formed in some dikes during thermal recrystallization.

MONZONITE AND QUARTZ MONZONITE

The monzonite and quartz monzonite veins and blocks crystallized from a magma more calcic than but otherwise similar to that of the quartz trachytes. Partly assimilated labradorite xenocrysts are present in several monzonite specimens; hence, lime has probably been added beyond that present in the melt before intrusion. In other words, a quartz trachyte (that is, quartz syenite) magma may have been modified to a quartz monzonite magma by the assimilation of labradorite. In support of this suggestion, labradorite crystals of the gabbro are albitized where protruding into the monzonite veins, and recrystallized labradorite bordering the monzonite veins grades out into the coarser vein oligoclase. Pyroxene of the monzonite veins is similar in composition to that of the adjacent gabbro and was probably incorporated from the gabbro.

The oligoclase in the monzonite is extensively mantled and veined by albite (An_{0-5}), which was probably introduced by late magmatic or deuteric solutions. Some irregular areas of orthoclase resemble exsolution perthite, but other orthoclase appears to be a massive replacement of the primary plagioclase. The replacing orthoclase may have been a secondary introduction (as in the altered gabbro), but it may be that the orthoclase mobilized and concentrated after exsolution by "stewing" in the late magmatic or deuteric solutions which deposited the albite.

The marginal selvages of recrystallized gabbro adjacent to the monzonite veins indicate that the monzonite magma was emplaced at a temperature sufficiently high to melt the gabbro. Under anhydrous conditions a temperature of about 1,200° or 1,300° C would be required to melt gabbro, but the temperature may have been as low as 700° or 800° C under a high H_2O pressure (Yoder and Tilley, 1956, p. 170, fig. 6). Melted augite of the gabbro was incorporated in the monzonite veins to form the dominant ferromagnesian mineral. Apatite and magnetite of the veins may have been incorporated from the melted gabbro, but quartz and zircon probably crystallized from the melt as they do not occur in the gabbro. The fringe of biotite enclosing magnetite in the contact zone of the monzonite and gabbro resulted from deuteric reaction of the magnetite with the hydrous monzonite magma or related deuteric solutions or gases. Lenticular quartz crystals between the lamellae in biotite crystals of one vein suggest hat the albitic deuteric solutions contained silica as well.

Siliceous veins such as granophyre are commonly associated with sills of tholeiitic basalt and gabbro, from which they can be shown to have originated by fractional crystallization. Consequently, the possibility that the quartz monzonite may be the residuum produced by fractional crystallization of the gabbro stock(?) within the Truk volcano requires consideration. This possibility seems unlikely, as one of the monzonite veins contained in a composite basalt and gabbro xenolith cuts both fine-grained basalt and gabbro. The basalt is a dike chilled against the gabbro during intrusion at depth and is andesine basalt similar in composition to lavas exposed at the surface; hence, the basalt was emplaced after the gabbro had cooled sufficiently to chill intruded basalt. The interval of cooling and of basalt intrusion between the gabbro and the monzonite suggests that the latter are unrelated in origin.

The quartz monzonite veins and blocks of the Truk Islands shed no new light on the origin of oversaturated magma outside the andesite line. This quartz monzonite is, however, similar in some respects to the alkalic granite blocks of Ascension Island, which Tilley (1950, p. 43) believes a differentiation product of undersaturated alkalic olivine basalt magma. As already pointed out, the quartz monzonite blocks of Truk may represent contaminated quartz alkalic syenite magma (perhaps more accurately an alkalic granite magma). The blocks of Ascension Island are characterized by alkalic pyroxenes or amphiboles, and some of the quartz monzonite in Truk contains amphibole. Pyroxene in blocks of the latter is not alkali rich, probably because it was incorporated from gabbro.

HYDROTHERMAL ALTERATION

The secondary minerals in dikes, flows, and blocks include albite, potassic feldspar (probably orthoclase), chlorite, calcite, quartz, sphene, and less commonly epidote, actinolite, sericite, hematite, and pyrite. This overall assemblage suggests a hydrothermal or pneumatolytic origin. The hydrothermal minerals are restricted to the islands near the former central crater. Altered blocks in otherwise unaltered breccia prove that the alteration occurred during the active magmatic history of the volcano. The sodium and potassium are easily explained as residual materials concentrated in the magma chamber during crystallization. Origin of the silica is a problem that is presently unsolved.

SUMMARY OF CONCLUSIONS

The lavas of the Truk Islands are, excepting the quartz trachyte, an undersaturated assemblage ranging from nepheline-melilite basalt to trachyte; this assemblage resembles lavas of the declining phase in the Hawaiian Islands. Most of the lavas of Truk can be explained by fractional crystallization of undersaturated olivine basalt parent magma. This study supplies no new evidence on the origin of the olivine basalt magma.

Gabbro xenoliths in pyroclastic breccia of the central islands were probably torn from an intrusive body in the volcano. The gabbro is undersaturated and may have been derived from the same source as the Truk Islands lavas.

Quartz trachyte lavas, dikes, and ejected blocks were formed by the crystallization of magma oversaturated in silica. This silicic magma was probably produced from the same undersaturated magma that gave rise to the undersaturated lavas on Truk. The process by which this silica-rich magma was produced from an undersaturated magma is not known.

Monzonite and quartz monzonite xenoliths and dikes contained in gabbro xenoliths crystallized from a quartz syenite magma which had assimilated pyroxene and labradorite from adjacent gabbro in the volcano. The quartz syenite magma is probably the same magma that erupted to form the quartz trachyte just discussed.

Many blocks in breccias of the central islands were hydrothermally altered before they were erupted, and all rocks of the central islands have been subsequently altered to a lesser extent. Calcic plagioclase is extensively albitized, and primary anorthoclase in the trachytes is now represented by intergrowths of "low" albite and orthoclase. Orthoclase extensively mantles the albite of some trachyte and suggests redistribution of alkalies by hydrothermal solutions or volatiles. Minerals deposited by hydrothermal solutions include albite, potassic feldspar (probably orthoclase), chlorite, calcite, quartz, and pyrite. Epidote, actinolite, sphene, sericite, and hematite have been formed by alteration of the primary igneous minerals. The soda and potash of the hydrothermal solutions are readily explained as residual materials concentrated in the magma chamber during crystallization. Origin of the silica is an unsolved problem.

REFERENCES CITED

Betz, Frederick, Jr., and Hess, H. H., 1942, The floor of the North Pacific Ocean: Am. Geophys. Union Trans., 1942, pt. 2, p. 348–349.

Bridge, Josiah, 1948, A restudy of reported occurrences of schist on Truk, Eastern Caroline Islands: Pacific Sci., v. 2, no. 3, p. 216–222.

Chayes, Felix, 1952, Notes on the staining of potash feldspar with sodium cobaltinitrite in thin section: Am. Mineralogist, v. 37, p. 337–340.

Darwin, Charles, 1909, The voyage of the *Beagle:* republication of original as v. 29 in the "Harvard Classics," New York, P. F. Collier and Son Co., 547 p.

Fenner, C. N., and Day, A. L., 1931, Borehole investigations in the geyser basin of Yellowstone National Park [abs.]: Washington Acad. Sci. Jour., v. 21, no. 20, p. 488–489.

Geophysical Laboratory, 1956, Primitive magmas: Carnegie Inst. Washington Year Book 55, p. 161–171.

Hess, H. H., 1948, Major structural features of the western North Pacific, an interpretation of the H. O. 5485, Bathymetric Chart, Korea to New Guinea: Geol. Soc. America Bull., v. 59, p. 417–426.

——— 1949, Chemical composition and optical properties of common clino-pyroxenes, Part I: Am. Mineralogist, v. 3–4, nos. 9–10, p. 621–666.

Hess, H. H., and Poldervaart, Arie, 1951, Pyroxenes in the crystallization of basaltic magma: Jour. Geology, v. 59, p. 472–489.

Hobbs, W. H., 1944, Mountain growth, a study of the southwestern Pacific region: Am. Philos. Soc. Proc., v. 88, p. 221–268.

Irving, E. M., 1952, Geological history and petroleum possibilities of the Philippines: Am. Assoc. Petroleum Geologists Bull., v. 36, p. 437–476.

Kennedy, W. Q., 1933, Trends of differentiation in basaltic magmas: Am. Jour. Sci., v. 25, p. 239–256.

Kramer, A. F., 1908, Aus der Schutzgebitfender Südsee, Studienerise nach den Zentral-und Westkarolinen: Deutschen Schutzgebieten Mitt., v. 2, no. 3, p. 169–189.

Kuno, Hisashi, 1950, Petrology of Hakone volcano and adjacent areas of Japan: Geol. Soc. America Bull., v. 61, p. 957–1020.

Kuno, Hisashi, Yamasaki, Kazuo, Iida, Chūzō, and Nagashima, Kōzō, 1957, Differentiation of Hawaiian magmas: Japanese Jour. Geol. and Geog., v. 28, p. 179–218.

Ladd, H. S., Ingerson, Earl, Townsend, R. C., Russell, Martin and Stephenson, H. H., 1953, Drilling on Eniwetok Atoll Marshall Islands: Am. Assoc. Petroleum Geologists Bull., v 37, p. 2257–2280.

Macdonald, G. A., 1949, Hawaiian petrographic province: Bull Geol. Soc. America, v. 60, p. 1541–1596.

MacGregor, A. G., 1931, Clouded feldspar and thermal metamorphism: Mineralog. Mag., v. 22, p. 524–538.

Nockolds, S. R., and Allen, R., 1954, The geochemistry of some igneous rock series, pt. 2: Geochim. et Cosmochim. Acta, v. 5, no. 6, p. 245–285.

Powers, H. A., 1955, Composition and origin of basaltic magma of the Hawaiian Islands: Geochim. et Cosmochim. Acta, v 7, p. 77–107.

Raitt, R. W., 1956, Seismic refraction studies of the Pacific Ocean Basin: Geol. Soc. America Bull., v. 67, p. 1623–1640

Ross, C. S., and others, 1954, Origin of dunites and olivine-rich inclusions in basaltic rocks: Am. Mineralogist, v. 39, p 693–737.

Schmidt, R. G., 1957, Geology of Saipan, Mariana Islands, Part 2, Petrology of the volcanic rocks: U.S. Geol. Survey Prof Paper 280–B, 174 p.

Stark, J. T., Paseur, J. E., Hay, R. L., May, H. G., and Patterson, E. D., 1958, Military geology of Truk Islands, Caroline Islands: U.S. Army Chief of Engineers, Intelligence Div. Office Eng., Headquarters, U.S. Army Pacific, 207 p., maps.

Stearns, H. T., 1946, Geology of Hawaiian Islands: Hawaii Dept Pub. Lands, Div. Hydrography Bull. 8, 100 p.

Stearns, H. T., and Vaksvik, R. V., 1935, Geology and ground water resources of Oahu, Hawaii: Hawaii Dept. Pub. Lands Div. Hydrography Bull. 1, 478 p.

Tayama, Risaburo, 1952, Coral reefs in South Seas: Suirobu Hokoku (Hydro. Office Bull.), v. 11, no. 941, 292 p.

Thornton, C. P., and Tuttle, O. F., 1956, Applications of the differentiation index to petrologic problems [abs.]: Geol Soc. America Bull., v. 67, p. 1738–1739.

Tilley, C. E., 1950, Some aspects of magmatic evolution: Geol Soc. London Quart. Jour. 106, pt. 1, p. 37–61.

Winchell, Horace, 1947, Honolulu series, Oahu, Hawaii: Geol Soc. America Bull., v. 58, p. 1–48.

Yoder, H. S., and Tilley, C. E., 1956, Natural tholeiitic basalt water system: Carnegie Inst. Wash. Year Book 55, p 169–171.

INDEX

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7 ~~6~~ MAPS IN POCKET 7/02 4/04

5 ~~6~~ MISSING

7/91 DATE (MO/YR) 7/02

Missing 2-409
4-408