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Geology and ore deposits of the Johnnie District, Nye County, Nevada

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A thesis submitted in partial fulfillment of the requirements for the degree of Master of Science in Geology

by

Stanley Wayne Ivosevic
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ABSTRACT

The Johnnie district, in the northwestern Spring Mountains, Nye County, Nevada, may have produced a little under 100,000 troy oz of gold, since the discovery of the district in 1890.

An approximately 13,000-ft-thick (4,000 m) section of east-dipping upper Precambrian through Middle Cambrian miogeosynclinal clastic and carbonate rocks is exposed in the district. The strata are, in order of decreasing age, the Johnnie Formation, Stirling Quartzite, Wood Canyon Formation, Zabriskie Quartzite, and Carrara and Bonanza King Formations. These are overlain by Cenozoic units which include an older unit and a younger unit of fanglomerate, the older containing a megabreccia deposit, and Quaternary alluvium.

The rocks were deformed by the Late Cretaceous Sevier orogeny and by subsequent tectonic events, which include Basin-and-Range faulting, of Miocene age. The oldest structures, formed in conjunction with Sevier tectonism, are disharmonic folds in the Johnnie Formation, which folds underwent rotation by later (Sevier orogeny) eastward tilting of the district during larger scale folding. Simultaneously with the later folding, competent rocks were translated across less competent ones along zones of tectonic readjustment; the most notable zone is at the top of the Johnnie Formation in the western part of the district.

High-angle fractures--in extension, conjugate, and pressure-release orientations--developed at about the same time and were the ancestors for most younger high-angle structures, including quartz veins.

Longitudinal faulting occurred at the end of the Sevier orogeny; and concurrently, the related Congress low-angle normal fault developed. The district was dropped down to the west along the Grapevine fault system during Basin-and-Range normal faulting at the west face of the Spring Mountains. Throughout the structural development of the district, displacement occurred along transverse faults in secondary readjustment to displacement along other features and also low-angle faults transposed younger rocks across older ones.

Some of the tectonism caused local carbonatization of the Bonanza King Formation.

The district was eroded in a series of stages during Basin-and-Range faulting. Finally, sometime between latest Miocene to middle Pleistocene times, pediments developed at the edges of the resultant basins, then were buried by bajadas of older fanglomerate. Later, parts of the pediments were exhumed by erosion.

High-angle and concordant quartz-bearing structures were emplaced during hydrothermal activity, probably between the Paleocene and early Miocene epochs High-angle quartz veins, the average of which strikes ENE and dips north, dominate and are the hosts for most economic, mesothermal mineralization in the district. Three ore mineralogic suites are present: gold-chalcopyrite-pyrite, chalcopyrite-galena, and galena (-calcite). Additionally, chalcopyrite occurs with specularite in stratabound quartz-poor lodes of apparent hydrothermal origin; these oxidize to low-grade malachite deposits. Placer gold deposits formed in the older fanglomerate during post Basin-and-Range faulting erosion of the district. The characteristic wall-rock alteration mineralogic suites in the hypogene deposits are sericite and pyrite in clastic rocks and also sericite, alone, in dolomite; the alteration minerals, chlorite, calcite, and specularite, occur locally. The ore mineralogic suites define a district-wide pattern of hypogene mineralogic zonation about gold centers at the Johnnie and Congress mines, the main producers in the district.

The fundamental control which admitted the hydrothermal fluids into the district is obscure. However, trains of quartz veins are concentrated within 2.5-mi-long (4 km), ENE-trending principal mineralized structures which lie astride an inferred 13-mi-long (21 km) N.-35°-E.-trending major longitudinal structure, which may be a manifestation of the fundamental control. The gold mineralization is localized in the Zabriskie Quartzite and the dolomitic rocks near the top of the Wood Canyon Formation, in part, by the retention of hydrothermal fluids beneath a blanket of shaly rocks at the base of the Carrara Formation. The base of the dolomitic rocks is also the base of the Cambrian section, which is recognized as a favorable stratigraphic site for ore deposition at other places in the southern Great Basin.

Although this report develops a theme in which the hydrothermal fluids were introduced from a remote source below the district, as the broad concepts of ore genesis change with time, it may become accepted that many of the constituents of the hydrothermal fluid were derived from local, syngenetic sources.

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INTRODUCTION

The Johnnie mining district, located in southern Nye County, Nevada (figs. 1 and 2), was an important local producer of native gold between approximately 1890 and 1915.

The present study was undertaken to determine the sequence of structural events in the district and the method of emplacement of the ore minerals and their quartz vein hosts. The detailed examination performed during this study accomplished these objectives and more fully delineated the regionality of the favorability of Lower Cambrian carbonate rocks as hosts for ore deposits. This study helps clarify the origin and morphology of disharmonic folding in the area, adds additional observations upon what had originally been called the "Johnnie thrust", and revises earlier interpretations concerning the trend of the concealed Grapevine fault. More observations are added to the literature of the enigmatic low-angle faults which transpose younger rocks across older ones in eastern Nevada and southern California.

Field work was conducted in the Johnnie district by the author intermittently during 1970-73 for a total of 18 months. The district was mapped at a scale of 1:12,000 with supplemental mapping of selected areas at larger scales. Most of the maps produced are presented herein. A structural analysis was undertaken, and mineralogic and chemical analytical studies were performed in the laboratory by the author and by Miles L. Silberman of the U. S. Geological Survey to amplify upon the field examination. Problems encountered during age dating of the ore deposits are being investigated by members of the U. S. Geological Survey, and their findings will be discussed in reports subsequent to

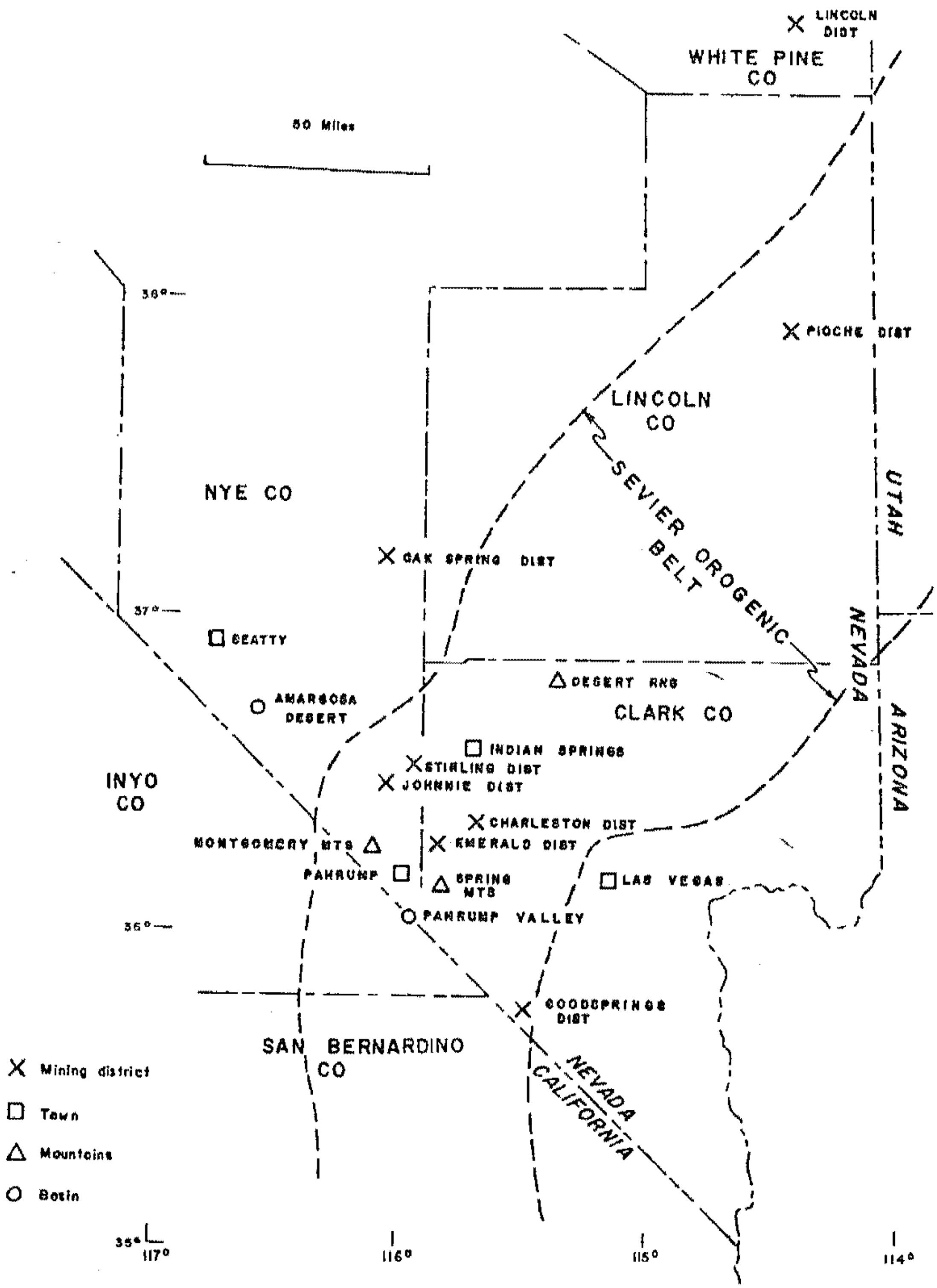

Figure 1. Index map of southern Nevada and adjacent areas showing locations of areas discussed in this report. Position of Sevier orogenic belt after Armstrong (1968) and Fleck (1970).

this one. Their findings will refine the estimates of age and temperature of ore deposition given herein but not otherwise necessitate modification of this report.

Where metric conversions are given in this report they are rounded to some extent from the English measurements.

Previous Work

The geology of the northwestern Spring Mountains, and of the included Johnnie district, was systematically examined and reported upon by Nolan (1924, summarized by Nolan, 1929). The only previous mention of the geology of the same general area may have been in a report, not available to me, by Gilbert (1875) to which Nolan (1924) alludes. The Johnnie district and vicinity are not mentioned in the pioneering geologic studies of the region (Wheeler, 1889; Spurr, 1903; Ball, 1906, 1907). Nolan (1924, 1929) recognized a structural feature in the Johnnie district which he named the "Johnnie thrust" and identified as a decollement surface. Although since then this has been reinterpreted as being a structural feature of lesser magnitude, the so-called "Johnnie thrust" persists in the literature as the type decollement thrust.

Hamil (1966) describes the geology of the Mt. Schader (15') quadrangle (fig. 2), in which the Johnnie district is included. Burchfiel (1961, 1964, 1965) and Livingston (1964) describe the geology of the adjacent Specter Range quadrangle (fig. 2); and Vincelette (1964), the Mt. Stirling quadrangle (fig. 2), which also is adjacent. The regional geology of Nye County, in which the district is situated, is discussed by Cornwall (1972); and Longwell and others (1965) describe the

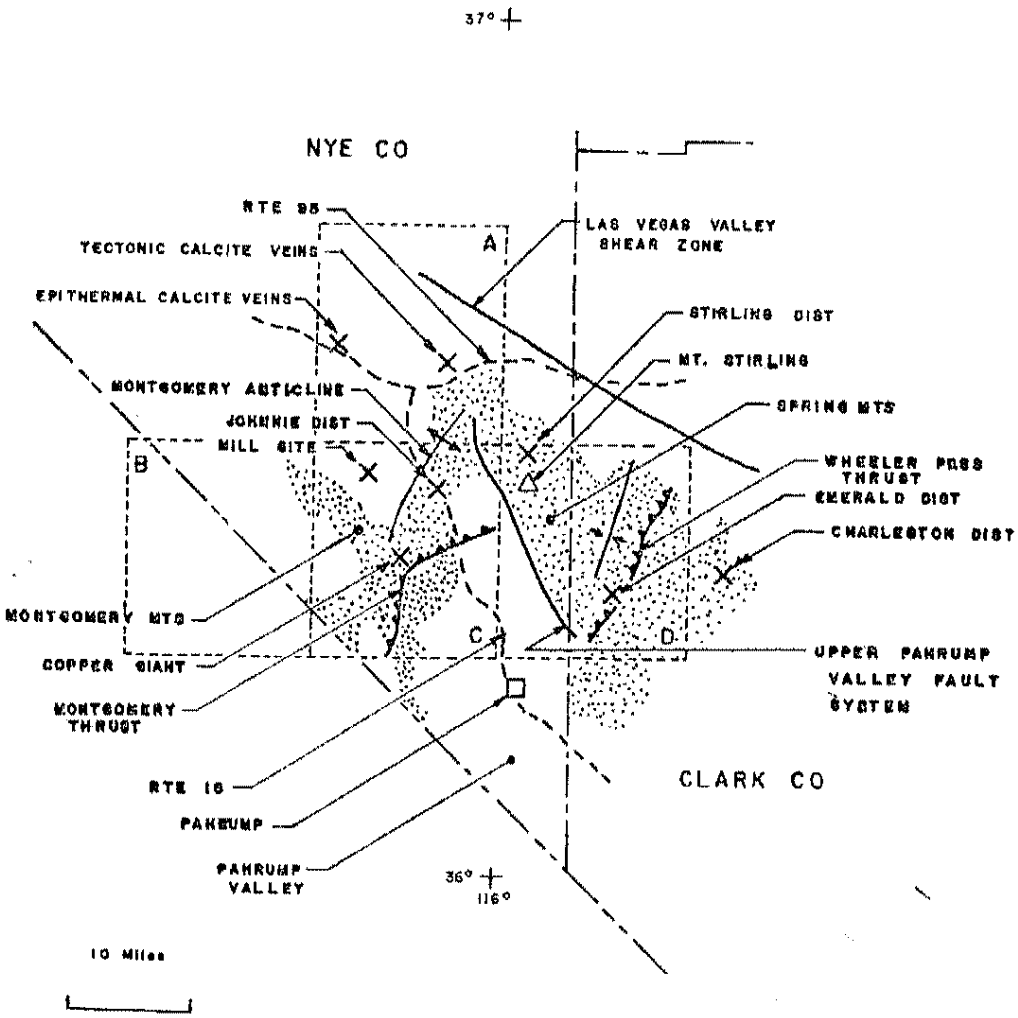

Figure 2. Index map of northwestern Spring Mountains and adjacent areas showing selected geologic features and locations of areas discussed in this report. Key to 15' quadrangles: A-Specter Range; B-Ash Meadows; C-Mt. Schader; D-Mt. Stirling. Outcrop areas of northwestern Spring Mountains, Montgomery Mountains, and nearby areas, only, are stippled.

regional geology of nearby Clark County, Nevada (fig. 1). The geology of the Spring Mountains is compiled in Burchfiel and others (1974). Ivosevic (1974) summarizes the genesis of some of the Mesozoic structures in the district.

As an adjunct to his general geologic report Nolan (1924) describes the mineral deposits of the Johnnie district, the nearby Copper Giant property, the Stirling district, and the Emerald district (figs. 1 and 2). Labbe (1921) reports on the placer deposits of the Johnnie district. Smith and Vanderburg (1932) and Vanderburg (1936) also report on the placers of the district generally by quoting Labbe (1921) but adding some new material.

The ore deposits and mining activity of the Johnnie district are mentioned in several tabulations of mining districts in the formal literature, which tabulations are based upon then recent work or upon previous work by the authors or by others. These are Hill (1912), Sanford and Stone (1914), Schrader and others (1917), Lincoln (1923), Nolan (1936), Kral (1951), and Cornwall (1972). Some less formal reports on the Johnnie district are listed in Appendix B.

Location, Accessibility, and Geography

The Johnnie mining district, as described herein, includes the traditional gold mining areas located along a northeast-trending diagonal through the approximately 40 sq mi (100 sq km) area roughly bounded by latitudes 36°23'30" and 36°29'30" N, longitudes 116°00'00" and 116°09'00" W. The area is centered over the mutual corner of Tps. 17 and 18 S., Rs. 52 and 53 E. in the southern corner of Nye County, Nevada. Not treated here is the Copper Giant property, to the southeast (fig. 2) in an area which is presently included in the Johnnie recording district. The Johnnie district is 55 mi (89 km) northwest of Las Vegas, Nevada (fig. 1), the nearest city, and 16 mi (26 km) north of Pahrump, Nevada (figs. 1 and 2), the nearest settlement.

State Route 16, a paved highway, passes through the center of the Johnnie district. Dirt roads lead into the main mining areas.

The district occupies low foothills flanked by alluviated basins in the northwestern Spring Mountains and the northeast spur of the unnamed range to the west, which is sometimes called the Montgomery Mountains (figs. 1 and 2). The total relief from the highlands at the east edge of the mapped area (elevation 6,000 ft or 1,830 m) to the west margin, along the Amargosa Desert, is approximately 3,300 ft (1,000 m). The average relief is 500 to 1,000 ft (150-300 m). (See pl. 1.)

The geography of the Johnnie district is portrayed on the Mt. Schader and Amargosa Flat (7.5') topographic quadrangle maps (U. S. Geological Survey, 1968).

Vegetation is sparse in the area. Annual precipitation is between 4 and 10 in. (10-25 cm). Annual temperatures range between 20° and 100° F. (Numerical data from Brown (1960).) Springs provide a small water supply.

GENERAL GEOLOGY

Regional Geology

The geologic evolution of the Johnnie district was directed by events within four successive, regional tectonic phenomena; namely, the Cordilleran miogeosyncline, the Sevier orogenic belt, the Las Vegas Valley shear zone, and the Basin-and-Range physiographic province. Igneous activity accompanied some of these events. The locations of some of the geologic features mentioned here are given in figures 1 and 2.

The Johnnie district lies approximately along the axis of the Cordilleran miogeosyncline, a north-trending, elongate area of late Precambrian to middle Mesozoic marine and later terrestrial sedimentation. The original miogeosynclinal section in the northern Spring Mountains was approximately 35,000 ft (10,700 m) thick and overlay a basement of earlier Precambrian metamorphic rocks. As many as four unconformities throughout the section elsewhere, beyond the Johnnie district, demark pulses of diastrophic activity within the miogeosyncline and adjacent areas (Hazzard, 1937; Vincelette, 1964; Longwell and others, 1965; Hamil, 1966; Fleck, 1967, 1970; Armstrong, 1968).

The sedimentary history of the region was completed by the deposition of two terrestrial clastic and lacustrine sedimentary suites. The first, a Late Cretaceous and early Tertiary post-Sevier orogeny suite, is absent in the Johnnie region. The second comprises Basin-and -Range basin fill of Neogene age and attains thicknesses in excess of 3,000 ft (900 m) in the basins surrounding the northern half of the Spring Mountains.

The Johnnie district is situated along the western margin of the Sevier orogenic belt (Harris, 1959; Armstrong, 1963; Fleck, 1970, fig. 7). The orogenic belt (fig. 1), which extends from southern California (Hewett, 1956; Burchfiel and Davis, 1974) to central Idaho (Armstrong, 1968), is an elongate zone of Cretaceous (Armstrong, 1968) folding and eastward directed thrust faulting which caused (Armstrong, 1963, 1968) the Spring Mountains and adjacent areas to become a region of broad longitudinal folds, paralleled by four major, laterally persistent thrust faults, the westernmost of which is comprised of two segments--the possibly once continuous Wheeler Pass and Montgomery thrust faults (fig. 2). Although some authors feel that all or part of folding followed thrust faulting, Fleck (1970) demonstrates that folding preceded and guided thrust faulting. He also shows that the latter is a Late Cretaceous (75-90 m.y.) event and that most of the deformation probably occurred during the Late Cretaceous Epoch.

A number of structural features which transpose younger rocks across older ones by any of several methods are exposed by deep erosion along the western margin of the Sevier orogenic belt (Armstrong, 1968).

The Las Vegas Valley shear zone (fig. 2), named by Longwell (1960), is a broad zone of right-lateral shearing and oroflexural bending which truncates the northeastern edge of the Spring Mountains. It is a part of a broader system of major right-lateral strike-separation structural features. Stratigraphic and structural features in northern Clark County are offset across the zone by approximately 45 mi (70 km). Lesser features with this approximate orientation throughout the region typically exhibit displacements of similar direction. Fleck (1967) and Longwell (1974) conclude that most movement along the Las Vegas Valley shear zone occurred during the middle of the Miocene Epoch (10-17 m.y., Fleck, 1967).

Northwest-trending Basin-and-Range normal faults, which control the trend of the western edge of the Spring Mountains and the locations of the adjacent basins of deposition of Basin-and-Range basin fill, are estimated, regionally, to be from middle to late Miocene (7-15 m.y., Fleck, 1967) in age, with activity continuing along some faults to present. Fleck (1967) also notes that Basin-and-Range faulting is, in part, coeval with movement along the Las Vegas Valley shear zone.

Although intrusive rocks are absent from the Johnnie district, an understanding of the regional igneous geology assists in reconstructing the genesis of the ore deposits there. The northern half of the Spring Mountains is devoid of igneous rocks, as is an approximately comparable area of the offset equivalents of the Spring Mountains north of the Las Vegas Valley shear zone. The nearest known intrusive rocks are in the Oak Spring and Goodsprings districts (fig. 1), approximately 50 mi (80 km) north and south, respectively, of the Johnnie district. Two separate episodes of volcanism and plutonism occurred in the region, one in the middle and late Mesozoic Era and one in the middle Tertiary Period.

The first episode, essentially along the Sevier orogenic belt in southern California and southern Nevada, preceded, accompanied, and followed folding and thrust faulting: Triassic and Cretaceous flows and tuffs are present (Hewett, 1956; Longwell and others, 1965; Adams and others, 1966; Fleck, 1967); and so are some acidic plugs--including those at Oak Spring and Goodsprings--dated between 82-110 m.y. (Adams and others, 1966; Fleck, 1967; Krueger and Schilling, 1971; Cornwall, 1972).

Abundant volcanic rocks, with related hypabyssal intrusives, were deposited in areas east and west of the Spring Mountains during the late Oligocene to middle Miocene epochs; waning volcanism continued into the Quaternary Period (Armstrong, 1963; Krueger and Schilling, 1971; Volborth, 1973).

The subdivision of the Cretaceous Period accepted by the Geological Society of London (1964) and the subdivision of the Cenozoic Era proposed by Berggren (1972) are used here in assigning relative ages to radiometric dates cited from the literature.

Stratigraphy

Introduction

A sequence of six sedimentary rock formations of late Precambrian and Early and Middle Cambrian age are exposed in the Johnnie district. A total of eight sedimentary rock units are recognized herein for general geologic mapping purposes. A ninth is recognized on intermediate scale maps. The contacts between all sedimentary rock formations are conformable and transitional through zones 50 to 100 ft (15-30 m) thick. Not all Cenozoic units are completely differentiated on geologic maps herein. The stratigraphy of the Johnnie district is summarized in table 1.

The stratigraphic section is nearly 14,000 ft (4,300 m) thick, of which the sedimentary formations comprise approximately 13,000 ft (4,000 m) and the Cenozoic deposits comprise the rest. These represent a lower miogeosynclinal assemblage of shallow water sediments, which change upward into finer through coarser grained clastic rocks which were derived (Stewart, 1970) from a cratonic source area to the

Table 1. Stratigraphic units present in the Johnnie district showing, in parentheses, subdivisions recognized by Stewart (1966, 1970).

Age	Name	Thickness (feet)	Character
Holocene	Alluvium	0-20 ± (0-6 m)	Colluvium, talus, stream bedloads
Late Pleistocene to Holocene	Younger fanglomerate	0-100 + (0-30 m)	Inactive alluvial fans of compact sand and gravel
Late Pliocene to Middle Pleistocene	Older fanglomerate	0-200 + (0-60 m)	Dissected, consolidated to caliche-cemented sandy gravel; includes a 50'-thick unit of volcanic sediments
	Megabreccia unit	300 ± (90 m)	Tabular body of slabs of upper unit of Wood Canyon Formation, quartz-veined Zabriskie Quartzite, and lower part of Carrara Formation
	Unconformity		
Middle Cambrian	Bonanza King Formation: Papoose Lake member	1,200 ± (365 m)	Massive dolomite with lamellar texture
Early and Middle Cambrian	Carrara Formation	700-1,300 (210-400 m)	Thin- to medium-bedded shaly rocks near base; thick-bedded limestone predominates near top
Early Cambrian	Zabriskie Quartzite	115-240 (35-73 m)	Medium-bedded to massive quartzite

Continued on next page.

Table 1. (continued)

Late Precambrian and Early Cambrian	Wood Canyon Formation	Dolomite bearing unit (upper member, in part)	175-310 (53-94 m)	Medium-bedded dolomite and thin-bedded quartzite and siltstone
		Lower unit (lower member; middle member; upper member, in part)	1,680 (510 m)	Medium-bedded quartzite and siltstone with some dolomite
Late Precambrian	Stirling Quartzite	Upper unit (C member; D member; E member)	2,160 (659 m)	Medium-bedded quartzite and dolomitic quartzite; shale at base; thick, massive quartzite near top
		Lower unit (A member; B member)	2,080 (634 m)	Medium-bedded to massive quartzite; thick, massive unit near base
	Johnnie Formation	Upper unit (middle unit, in part; Rainstorm Member)	850-1,300 (260-395 m)	Thin- to medium-bedded shale and quartzite with several layers of brown-weathering dolomite and dolomitic quartzite
		Lower unit (lower unit; middle unit, in part) - base not exposed	2,900-3,300 (800-1002 m)	Medium- to thick-bedded shale and quartzite; layer of gray dolomite 650' below top

east. This assemblage is transitional upward through a shale and carbonate sequence into massive marine carbonate rocks. The Cenozoic deposits are composed of colluvial and alluvial detritus and a deposit of megabreccia; most of these probably were deposited in ancestral desert basins.

Three intervals of the stratigraphic section were considered during this study to be of potential economic interest in the Johnnie district. These are, in ascending stratigraphic order:

(1) A quartzite horizon within the lower unit of the Stirling Quartzite which is amenable to the deposition of copper minerals. (See Malachite Deposits.)

(2) The Zabriskie Quartzite and the rocks of the Wood Canyon Formation immediately below, which, where in favorable structural configuration, localize important gold-quartz veins.

(3) The older fanglomerate, which buries some of the highest grade placer gold deposits in the district. (See Placer Gold Deposits.)

The Stirling Quartzite, Wood Canyon Formation, and Zabriskie Quartzite, which are present in the Johnnie district, correlate (Stewart, 1970) with the Prospect Mountain Quartzite, a late Precambrian to Early Cambrian formation which has received attention as a host for hydrothermal mineralization throughout eastern Nevada. The Carrara Formation of the Johnnie district correlates (Longwell and others, 1965) with the Pioche Shale, another important host rock for ore in the aforementioned region, as well as correlating with the overlying Lyndon Limestone and Chisholm Shale there.

The clastic rocks of the Johnnie district ubiquitously contain relicts of presumably syngenetic pyrite, which originally comprised less than 1 percent of the rock. The relicts usually consist of small (less than 1.0 mm), barren or iron oxide fringed or filled, cubic or irregularly shaped, equant cavities. In some places larger (1.0-5.0 mm) cubes of dense limonite remain.

Upper Precambrian Rocks

Johnnie Formation: The Johnnie Formation was named by Nolan (1924) after the locality of the type section at the head of Johnnie Wash (pl. 1) in the Johnnie district. The formation is generally composed of fine-grained quartzites and siltstones with some interlayered dolomitic rocks. Nolan (1924) and Hamil (1966) measured exposed thicknesses of 4,460 and 4,220 ft (1,359 and 1,286 m), respectively, at the structurally complex type locality. It is reasonable to suppose that the concealed lower portion of the formation includes a unit of carbonate rocks similar to that in the Desert Range (fig. 1) (Stewart, 1970), increasing the thickness of the Johnnie Formation in the district by a minimal estimated addition of 400 ft (122 m).

Stewart (1970) subdivides the formation into three units: a lower unit, 2,500 ft (760 m) thick; a middle unit, 800 ft (245 m); and the Rainstorm Member (Barnes and others, 1965), 900 ft (275 m). (Stewart's (1970) figures are here adjusted to conform with Hamil's (1966) descriptions and measurements.)

For mapping purposes within the Johnnie district, I divide the Johnnie Formation into an upper and a lower unit, with respective thicknesses of approximately 1,300 and 2,900 ft (400 and 880 m) and with their mutual contact lying at the base of the lowest dolomite bed within the middle unit defined by Stewart (1970).

The basal rocks of the lower unit mapped are predominantly green phyllonitic rocks formed by the epigenetic cleaving and chevron folding of shale and siltstone. These become brown upward and grade into a sequence of generally brown-weathering quartzites with some shaly interbeds, so that shale and quartzite, each, constitute approximately 50 percent of the bulk composition of the lower unit. A cherty dolomite layer lies approximately 650 ft (200 m) below the top of the unit. This layer, composed of thick to massive beds of gray- and sometimes brown-weathering, gray calcitic dolomite is approximately 75 ft (25 m) thick, contains 3-ft-thick (1 m) layers of black chert which weather brown in the lower half, and contains a few thin beds of shale.

The upper unit mapped here is composed of quartzite, siltstone, and mudstone with layers of usually rusty brown weathering, gray or brown dolomite to dolomitic quartzite up to 15 ft (5 m) thick, which are particularly abundant toward the base. The uppermost dolomite layer is the "Johnnie oolite" (Stewart, 1966) in the Rainstorm Member. This oolitic dolomite layer is tan or cream colored, weathering somewhat paler, thick bedded, and approximately 16 ft (5 m) thick. A generally light-green siltstone unit immediately underlying the oolite has conspicuous limonite filled and fringed, presumed pyritohedral, molds (up to 0.3 mm in diameter) after syngenetic pyrite. The Johnnie oolite is followed upward by thin beds of shale, quartzite, dolomitic quartzite, and quartzitic dolomite.

The base of the Johnnie Formation is not exposed in the Mt.

Schader or adjacent quadrangles. The formation either overlies a "basement" of older Precambrian metamorphosed rocks; or is separated from the basement by a pre-Johnnie Formation sequence of sedimentary rocks correlative with the younger Precambrian Pahrump Group or Noonday Dolomite of southeastern Inyo and north central San Bernardino Counties, California (fig. 1); or the Johnnie Formation is in thrust contact (fig. 3) with a lower plate of miogeosynclinal rocks which, in turn, overlie the older rocks.

Inspection of stratigraphic correlation diagrams in Stewart (1970) suggests that the Johnnie district lies along, or slightly east of the axis of a north-northeast-trending basin of Johnnie Formation sedimentation. The Johnnie Formation thins southward but may thicken northward toward the vicinity of the Desert Range (fig. 1). Relations are such (Wright and Troxel, 1967) that neither the Pahrump Group nor Noonday Dolomite can be projected confidently beneath the Johnnie district.

Stewart (1970) believes the Johnnie Formation thins to the northwest and southeast of the Johnnie district. I believe that the formation thins to nothing within 40 mi (65 km) to the southeast and that it rests directly upon the basement for at least the last several miles of this.

The Johnnie Formation is conformably overlain by the Stirling Quartzite throughout most of the Johnnie district except in areas where the contact has been subjected to epigenetic tectonism, which will be discussed in <u>Contact between the Johnnie Formation and Stirling Quartzite</u>. With the exception of minor local unconformities (Burchfiel, 1965; Stewart, 1970), this conformability is regional (Vincelette, 1964;

Burchfiel, 1965; Stewart, 1970; and personal reconnaissance).

The contact between the Johnnie Formation and the Stirling Quartzite is a conspicuously tectonic one in exposures along the east margin of the Amargosa Desert to a point, 3.5 mi (5.6 km) north of the jog in Route 16, in the NE 1/4 sec. 12, T. 17 S., R. 52 E. The stratigraphic distance from the Johnnie oolite to the base of the Stirling quartzite is approximately 550 ft (170 m) in this area as well as at the type section. However, with increasingly severe epigenetic tectonic thinning of the thin-bedded and laminated rocks at the top of the Johnnie Formation, this distance is progressively reduced to approximately 100 ft (30 m) in the area at the south margin of plate 1 for a total reduction in thickness of the upper part of the Rainbow Member of the Johnnie Formation of approximately 450 ft (135 m).

The Johnnie Formation is generally considered to be of Precambrian age, because it lies below the currently accepted boundary between the Precambrian and Cambrian periods.

Stirling Quartzite: The Precambrian Stirling Quartzite takes its name (Nolan, 1924) from the type section approximately 2 mi (3 km) east of the Johnnie district on the west slopes of Mt. Stirling (fig. 2).

The formation consists mostly of medium- to coarse-grained, frequently cross-bedded, thick beds of quartzite, and it contains a few thin-bedded or laminated shaly interbeds and a few layers of dolomitic rocks of thin to moderate thickness. The Stirling Quartzite forms bold ridges facing the Amargosa Desert and the north side and east end of Johnnie Wash. Hamil (1966) measured a total thickness of 4,241 ft

(1,293 m) for exposures in the Johnnie district overlooking Route 16 from the east (pl. 1).

Stewart (1966) divides the Stirling Quartzite into five members lettered, from the lowest, A through E. I use two major subdivisions for mapping in the Johnnie district: a lower unit, 2,080 ft (634 m) thick; and an upper unit, 2,161 ft (659 m) thick (thicknesses computed from Hamil (1966)).

In the lower unit mapped, the base of the Stirling Quartzite, which is the A member of Stewart (1966), is an approximately 300-ft-thick (90 m) section of pale colored to white, fine-grained, medium- to thick-bedded quartzite, the upper half of which essentially constitutes a single massive unit that forms a prominent white cliff along the base of Mt. Stirling and, locally, above the Amargosa Desert. The white quartzite is overlain by a sequence of pink to purple, coarse-grained to conglomeratic, cross-bedded, massive to medium-bedded quartzites which comprise the B member. Occasional thin layers of dark weathering shaly quartzite and quartzitic shale are present here.

The lower approximately 550 ft (170 m) of the upper unit mapped, the C member of Stewart (1966), is of interest, because parts of it are investigated herein as a potential source of iron ore. The unit is composed of green, tan, and maroon mudstone, siltstone, and quartzite, with the rocks near the base and the top of the unit being more quartzitic and with some medium beds of light-brown-weathering dolomite near the base. Hamil (1966) directs attention to the hematitic quality of this unit. My bulk sample of maroon shale contained 2.7 percent iron, as microscopic amorphous iron oxide. This is in keeping with the analyses published by Stewart (1970) but lower than the crustal average of 4.7 percent given for shale by Turekian and Wedepohl (1964).

The remainder of the upper unit of the Stirling Quartzite is predominantly quartzite but varies more lithologically than does the lower unit. The C member is followed upward by a sequence, the D member, of medium-bedded, light brown dolomites and gray dolomitic quartzites containing some shale interbeds which become more common near the top. This is overlain by an approximately 800-ft-thick (245 m) sub-unit of pink, medium-grained to conglomeratic, thick-bedded to massive quartzite which comprises the E member.

The contact with the overlying Wood Canyon Formation is conformable through a transitional zone in which the aforementioned quartzite becomes more thinly bedded and includes some shale, dolomite, and dolomitic quartzite. The top of the Stirling Quartzite is mapped at the top of the uppermost layer of white quartzite, which is approximately 5 ft (1.5 m) thick.

Like the Johnnie Formation, the Stirling Quartzite is placed in the younger Precambrian period, because it is beneath the currently accepted boundary with the Cambrian Period.

Upper Precambrian and Cambrian Rocks

Wood Canyon Formation: This formation is named (Nolan, 1924) for Wood Canyon, in which the type section is located, approximately 10 mi (16 km) southeast of the Johnnie district. The Wood Canyon Formation is composed of quartzite, siltstone, dolomite and intergradational varieties of rock. The formation underlies colluvium-covered slopes and low hills between the Stirling and Zabriskie quartzites, the latter of which overlies the Wood Canyon Formation.

Hamil (1966) measured a section of the Wood Canyon Formation having a total thickness of 1,992 ft (607 m) in an area 1 mi (1.6 km) northwest of the Congress mine (pl. 1). The upper approximately 300 ft (91 m) of this formation is tectonically thinned by about one-half in the Johnnie and Labbe mining areas. In addition, a minor amount of the base of the formation is thinned out along the contact with the Stirling Quartzite in the vicinity of the Lillyan claims in the NE 1/4 sec. 25, T. 17 S., R. 52 E., and SW 1/4 sec. 19, T. 17 S., R. 53 E. (pl. 2).

Stewart (1966) divides the Wood Canyon Formation into three members, which are of subequal thickness in the Johnnie district. The lower member, in the district, is an essentially siltstone unit containing a number of dark- to medium-brown weathering beds of dolomite of medium thickness. The middle member is composed predominantly of dark-red to purple, medium-grained to conglomeratic, medium-bedded quartzite. The upper member is divisible into two halves: a bottom half composed of thin-bedded quartzite and siltstone; and a top half of fossiliferous dolomite and dolomitic quartzite with subordinate quartzite overlain by a fine-grained clastic sequence which is transitional into the Zabriskie Quartzite. A portion of the upper member is more fully described in Appendix A. Parts of the upper member are mapped separately on larger scale maps herein.

The base of the Cambrian Period, regionally, is placed at the bottom of the dolomite-bearing top half of the upper member of the Wood Canyon Formation, because the dolomite contains the lowest appearance of identifiable Cambrian fossils. Stewart (1970) summarizes the paleontology of the dolomite and the bases for the above conclusion.

Therefore, current usage assigns an age of upper Precambrian, to the lower five-sixths of the Wood Canyon Formation and places the top one-sixth in the Early Cambrian Period.

Cambrian Rocks

Zabriskie Quartzite: The Zabriskie Quartzite (Hazzard, 1937; Wheeler, 1948; Barnes and others, 1965), of Early Cambrian age, is a unit of massive quartzite which forms low persistent cliffs, among other places, above the Johnnie, Overfield, and Labbe mines (pl. 2). (For the history of its recognition in the district see Nolan (1924), Hamil (1966), and Cornwall (1972).) The quartzite is 240 ft (73 m) thick as measured by Hamil (1966) in the vicinity of his measured section of the Wood Canyon Formation. Locally, the Zabriskie Quartzite is tectonically thinned, the minimum resultant thickness becoming 115 ft (35 m) on the west slope of Mineral Monument Hill, located between the Johnnie and Overfield mines in the SW 1/4 NW 1/4 sec. 20, T. 17 S., R. 53 E. (pl. 2).

The Zabriskie Quartzite is a pink- to purple-weathering, white to purple, fine- to coarse-grained, medium-bedded to massive quartzite containing numerous Scolithus tubes in the lower half. It is tectonically brecciated locally to form a pseudoconglomerate. A more detailed description of the Zabriskie Quartzite is given in Appendix A.

The Zabriskie Quartzite is conformably overlain by the Carrara Formation, in which, like the upper part of the Wood Canyon Formation, Early Cambrian fossils are commonly recognized. Thus, the Zabriskie Quartzite also is included in the Lower Cambrian Period.

Carrara Formation: The Lower through Middle Cambrian Carrara Formation (Cornwall and Kleinhampl, 1961) is lithologically transitional from the predominantly clastic rocks which underlie it in the Johnnie district into the overlying massive carbonate Bonanza King Formation. (The history of its recognition in the Johnnie district is given in Nolan (1924, 1929), Wheeler (1948), Hamil (1966), and Cornwall (1972).) The Carrara Formation forms concave slopes above the Zabriskie Quartzite but becomes steep and ledge forming beneath the Bonanza King Formation. The Carrara Formation is approximately 1,300 ft (400 m) thick in the district, thinning tectonically to as little as 700 ft (210 m) throughout the area generally east of the Johnnie and Overfield mines.

The lowest part of the formation is a zone of transition from the underlying Zabriskie Quartzite and consists of generally dark-weathering shaly rocks containing a few thin beds of quartzite and a single conspicuous bed of quartzite near the base. This grades upward into a section of light-colored shale and limy shale interbedded with limestone. The shales become overall less abundant upward and assume brilliant hues of green and maroon toward the top of the formation. Argillaceous limestones and limestones with shaly partings and shaly interbeds appear in the middle of the formation. These become increasingly more abundant and thicker and contain less clastic material toward the top. The uppermost limestones form thick gray and tan ledges, are fossiliferous, exhibit some current effects, and are calcitic to dolomitic. The limestones in the zone of transition into the conformably overlying Bonanza King Formation are thin- to medium-bedded and contain rusty weathering argillaceous partings. The lowest part of the formation is described in Appendix A.

Early Cambrian and Middle Cambrian fossils are commonly found, respectively, in the lower and upper parts of the Carrara Formation. Therefore, the Carrara Formation is considered to be of Early to Middle Cambrian age.

Bonanza King Formation: The lower, Papoose Lake, member (Barnes and Palmer, 1961) of the Middle Cambrian Bonanza King Formation (Hazzard and Mason, 1936) is composed of massive layers of marine dolomite and limestone exposed as bold, gray cliffs throughout the Johnnie district. An approximately 1,200-ft-thick (365 m) section of the Papoose Lake Member, which is probably a nearly complete section of the member, is present in the district. (The history of the recognition of the formation in the district is given in Nolan (1924) and Wheeler (1948).)

The Bonanza King Formation is composed of 20- to 100-ft-thick (6-30 m) layers of dark- to light-gray, thin- to thick-bedded dolomite and subordinate limestone which locally contains a few rust- or tan-colored shaly interbeds or argillaceous partings. Some thin, irregular, concordant patches of pale chert are interbedded with the carbonate rocks in sparse amounts. Generally, but not invariably, the more magnesium-rich layers of carbonate rocks are lighter colored than are the calcian ones.

The texture of the carbonate rocks is dominated by thin lamellae contained therein. The layers are composed of two or, sometimes, three sets of alternating 3- to 6-mm-thick, irregular, concordant lamellae ranging in color from black to white. Again, the more magnesian material is usually paler colored with the white lamellae consisting of crystalline dolomite.

The Bonanza King Formation conformably overlies the Carrara Formation. However, two possible unconformities are present due north of Mt. Montgomery.

The Papoose Lake Member of the Bonanza King Formation lies beneath what Barnes and Palmer (1961) consider, on the basis of fossil evidence, to be the boundary between the Middle and Upper Cambrian periods. Therefore, that part of the Bonanza King Formation present in the Johnnie district is considered to be of Middle Cambrian age.

Cenozoic Sedimentary Deposits

Older Fanglomerate: This name is assigned herein to a deposit of late Pliocene to early or middle Pleistocene fanglomerate, which includes some volcanic sediments and a body of megabreccia. It is one of the uppermost deposits in the local regime of basin fill.

The older fanglomerate crops out extensively in the washes at the margins of the Pahrump Valley and Amargosa Desert. Erosional remnants occur in the bedrock areas to elevations as great as 5,000 ft (1,525 m).

The older fanglomerate thickens from an erosional feather edge in the Johnnie district to a minimum thickness of 200 ft (60 m), the height of some banks in which it is exposed.

The fanglomerate is a consolidated to caliche-cemented, poorly- to well-sorted deposit of angular to subround, sand- to gravel-sized sediments. Parts of a possibly lensiatic body of volcanic sediments approximately 50 ft (15 m) thick (differentiated on pl. 1) crop out 2 mi (3 km) south of Grapevine Springs in secs. 32 and 33, T. 17 S., R. 53 E.

This unit is composed of cream to gray, gritty-surfaced, slightly indurated, fine sand- to sand-sized, festoon-cross-bedded, water-lain biotitic tuff. It is locally pebbly and contains sandy or pebbly clastic interbeds. The deposit is stratified and dips toward the present valleys at angles on the order of 5°.

Where the lower contact of the older fanglomerate is exposed in the Johnnie district, it unconformably overlies dissected surfaces of the bedrock or of the megabreccia deposit, from which it is separated by a thin paleosol. The dissected upper surface of the older fanglomerate is in unconformable contact with, and partially buried by, alluvial fans composed of younger fanglomerate.

The disposition of the older fanglomerate as thin veneers on hilltops and thick deposits in gullies suggests that it once buried the topography but was partially eroded away subsequently, effectively exhuming the topography.

By analogy with the ages determined for correlative deposits elsewhere in the region the older fanglomerate is of late Pliocene to early or middle Pleistocene age (Denny and Drewes, 1965; Haynes, 1965; Fleck, 1967; Tschanz and Pampeyan, 1970).

Megabreccia Unit of Older Fanglomerate: A megabreccia (Longwell, 1936) deposit of late Pliocene to middle Pleistocene age is present approximately 2 mi (3 km) southeast of the Labbe mine in an area southwest of the Grapevine fault in the vicinity of secs. 27, 28, 29, 32, and 33, T. 17 S., R. 53 E. (pl. 1). Parts of the deposit form steep banks and prominent, but low, dark-colored cliffs. The megabreccia deposit covers an elongate southwest-trending area of approximately 1 1/2 sq mi

(4 sq km) and is inferred to be a tabular mass approximately 300 ft (90 m) thick. The megabreccia is a landslide composed of subhorizontal slabs derived from the dolomite-bearing upper part of the upper member of the Wood Canyon Formation (Unit 2, Appendix A), the Zabriskie Quartzite, and the lower part of the Carrara Formation up to and including some green shales with dolomite interbeds (Unit 3, Appendix A).

Most of the rocks are strongly sheared along bedding, but the original layering can still be distinguished. The Zabriskie Quartzite is intensely brecciated and rehealed to form a durable pseudoconglomerate. Individual breccia fragments within the Zabriskie Quartzite contain segments of quartz veinlets and others are composed solely of fragments of vein quartz. Integral structures within the megabreccia deposit are broad, open folds, normal faults, and presumed reverse faults all of which trend northeast at approximately right angles to the Grapevine fault. Another set of faults is subparallel to the Grapevine fault.

The present lateral continuity of the megabreccia under the alluvial cover which surrounds its exposures, or its former continuity before post-landsliding erosion, cannot be demonstrated.

The megabreccia deposit may lie unconformably upon a post-late Miocene bedrock erosion surface (see <u>Geomorphology</u>), or it is enveloped within the older fanglomerate. There is a strong suggestion that it is in direct contact with the middle member of the Wood Canyon Formation immediately southwest of the Grapevine fault. The megabreccia deposit is overlain at the northern tip of its area of exposure, by large fragments of a silicified slump block of brecciated quartzite, belonging to the middle member of the Wood Canyon Formation and derived from the face of the west wall of the Grapevine fault. With the evidence available, the megabreccia deposit can only be dated approximately as being of late Pliocene to middle Pleistocene age.

The megabreccia deposit probably was derived from an area east of the Grapevine fault and perpendicular to it (fig. 3). The elongate nature of the deposit and its perpendicularity to the fault, and configurations and senses of displacement of some of the faults within it suggest transport across the Grapevine fault from an eastern source area. The quartz veined source area probably is now removed by erosion; but it is likely that this area was east of the Grapevine fault as some quartz veins in the megabreccia deposit contain chlorite, and this mineral is characteristic of quartz veins east of the fault (see <u>Hypogene Mineralogic Zonation</u>).

An alternate source of this megabreccia deposit is the area in the vicinity of Hill 4529 approximately 1 mi (1.5 km) northwest of the northernmost exposures of the deposit. This is a likely source; because a portion of the stratigraphic section which includes the portion of the section comprising the megabreccia deposit has been removed by a low-angle fault there, and because the profuse quartz veining present in the Zabriskie Quartzite within the megabreccia is characteristic of the nearest exposures of the Zabriskie Quartzite west of Hill 4529. The trends of the folds and configurations of some faults in the megabreccia deposit suggest movement parallel to the fault with concomitant slumping to the southeast.

Younger Fanglomerate: Late Pleistocene to Holocene alluvial fans composed of younger fanglomerate define most of the surfaces of the basins adjoining the Johnnie district. The alluvial fans thicken from zero to a minimum of 100 ft (30 m) in the district. They are composed of compact, locally caliche-cemented, fanglomeratic sediments which are generally finer, more rounded, and better sorted than those of the older fanglomerate deposits.

Fan building is inactive at present; the fans have developed stable, desert-varnished surfaces, and they have been broached by throughgoing streams. These are older features not presently undergoing the conventional channel filling and diversion processes of fan genesis.

The younger fanglomerate can be as old as latest Pleistocene, post-dating the lakes of regional distribution associated with late Pleistocene glaciation; and the fanglomerate is no younger than 2,000 years old, the minimum age of the desert varnish on rocks upon the stable surfaces of the fans (Hunt, 1961; Denny and Drewes, 1965).

Alluvium: This includes the colluvium, talus, and stream bedloads of most recent origin which are derived from the weathering and transport of the underlying bedrock and from the reworking of older Cenozoic deposits in the Johnnie district. This unit probably never exceeds 20 ft (6 m) in thickness.

Some of the talus and colluvium derived from bedrock and occurring at higher elevations probably is contemporaneous with the older fanglomerate but indistinguishable from more recent deposits.

Structure

Introduction

The structures present in the Johnnie district are the products of a series of structural events which began with the onset of the Cretaceous Sevier orogeny and continued into the Quaternary Period.

The initial tectonic activity, compression oriented across the subparallel axes of the miogeosyncline and the Sevier orogenic belt, caused disharmonic folds and coaxial, but larger, folds to develop. Several sets of high-angle fractures, including ones in extension, conjugate shear, and pressure-release orientations, developed also. The folding was accompanied by conspicuous interbed movement and high-angle reverse faulting.

Subsequent tectonic movements during at least two stages of normal faulting and one of strike-slip faulting, utilized the previously formed fractures and bedding planes. The most conspicuous later structures to develop are the longitudinal Congress and transverse Grapevine fault systems and structures ancillary to each. (The longitudinal faults parallel the strike of the miogeosyncline and orogenic belt and the transverse faults cross the strike at high angles.) Other transverse faults developed in secondary relationships to major faults.

Low-angle structures (grouped herein as low-angle faults), which transpose younger over older rocks, developed by different genetic processes throughout the structural history of the district. The origins of these low-angle structures are difficult to determine, because they are at once genetically unrelated but morphologically similar, and because of geometrical considerations in explaining the fates of the rocks which presumably were removed by faulting.

The emplacement of quartz veins, which apparently occurred between periods of normal faulting, will be discussed under Ore Deposits.

Several geologic features are named herein for convenience of reference in this report. These are the Grapevine fault system and segments thereof (the term "Grapevine fault" had been established through prior usage in the district), the Congress fault system, and the Labbe fault.

The fault terminology of Hill (1959), based upon the actual and relative directions of movement along faults, is used, where appropriate, in the following discussion.

Habit of Deformation

Most deformation within the Johnnie district was of an essentially brittle nature accomplished under conditions of moderate tectonic load. Stresses were relieved by movement along primary anisotropic features, particularly bedding planes, and along fractures, both during folding and during later structural events. The relative effects of the earlier and later tectonic events cannot always be identified.

Folding was accomplished by a flexural slip mechanism in which each successively younger layer of rock was translated updip (west) relative to the one below and in which internal rotation occurred in thicker layers of rock by a process of brecciation and rehealing. Locally prominent sites of relative translation at the bases of massive quartzite or carbonate units which overlie thinner bedded, hence, less competent, rocks are mapped as "zones of tectonic readjustment", after the terminology of Vincelette (1964).

The crustal shortening which folding represents is also seen in the pervasive development of small high-angle reverse faults throughout the district. The reverse faults take off from bedding planes, break upward through the section, and, locally, overturn to the east. The latter are probably early formed faults rotated beyond vertical during folding. The reverse faults on Mt. Schader appear to have originated from the zone of tectonic readjustment below (pls. 1 and 2), but this connection cannot be demonstrated at the surface.

Internal rotation within the four most massive stratigraphic units in the district--the basal A member and the higher E member of the Stirling Quartzite, the Zabriskie Quartzite, and the Bonanza King Formation--also occurred during folding and later deformation. The effects of rotation upon the quartzite units range from the development of an ordered breccia (Secor, 1962) through the development of a disordered breccia of disoriented fragments. Nolan (1924) describes petrographically the origin of the breccia at the base of the Stirling Quartzite and defines the ultimate product of the process as a "microbreccia". With increasing deformation, the bedding becomes obscured by a crude foliation which generally dips at angles lower than nearby bedding. The Bonanza King Formation exhibits comparable effects.

Thin-bedded quartzites, shaly quartzites, and shales develop phyllitic cleavages which are parallel to bedding but which in most fissile rocks are at angles to bedding. With some exceptions, medium-bedded rocks, principally quartzites, display little internal rotation aside from slippage along favorably oriented cross bedding planes.

Deformation in High-Angle Structures: The cataclastic products in fault and fracture zones in the Johnnie district vary with the thicknesses of the bedding and the compositions of the rocks involved in the individual zones. In faults in massive carbonate rocks, such as those of the Bonanza King Formation, rehealed breccia zones up to 3 ft (1 m) thick develop which occasionally are the sites of dolomitization or recrystallization (see Alteration in the Bonanza King Formation). Faulted material in thin- to medium-bedded clastic and carbonate rocks and the intermediate varieties are generally sheared subparallel to the fault by a combination of dragging and shearing along surfaces subparallel to the fault. Faulted medium- and thick-bedded quartzites form disordered breccias, sheets of which are usually dragged to make angles of approximately 45° with fault planes. The gouge and breccia formed thereby, as well as in similar rocks in zones of tectonic readjustment, is locally silicified (described in Nature of Quartz Filling). Angular breccias, which are seldom rehealed, form in thin- and medium-bedded quartzites which are only fractured but not faulted.

Bedding-Related Structures

Bedding: The sedimentary rocks of the Johnnie district characteristically dip moderately east because the district lies on the mutual flank of the Montgomery anticline which bounds the western edge of the district and the major syncline which, from regional map examination, I infer to underlie the western edge of the Spring Mountains to the east of the district (figs. 2, 3, and 8). Analysis of 237 selected measurements of bedding attitude indicates that the average bed in the district strikes N. 10° E. and dips 40° E.

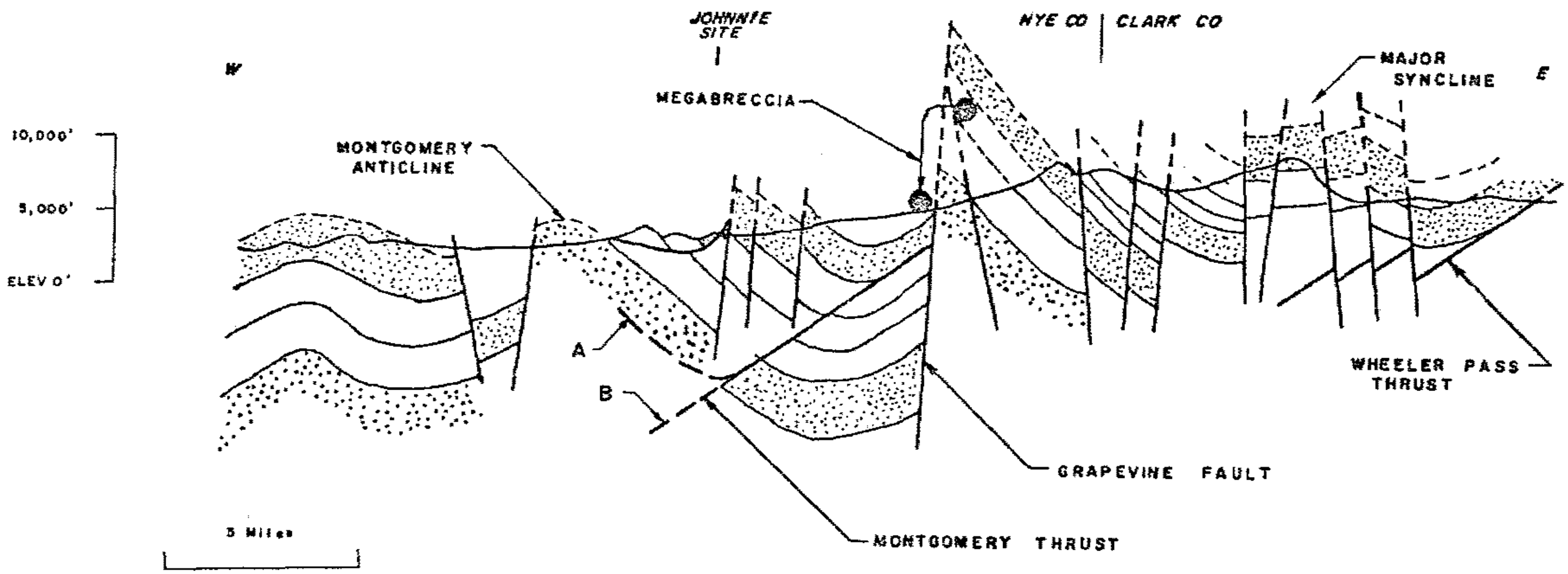

Figure 3. Geologic section - locally diagrammatic - through northwestern Spring Mountains along 18th parallel S., M. D. B. & M. showing large-scale folding, some possible thrust relations (A and B) beneath the Johnnie district, and probable origin of megabreccia deposit. Heavy stipple -- Johnnie Formation; light stipple -- Bonanza King Formation and younger Cambrian rocks. Compiled from Longwell and others (1965), Hamil (1966), Cornwall (1972), and mapping, this report.

There are local departures from this typically eastward dipping motif where the bedding swings into semi-parallelism with the northeast- to east-striking Montgomery thrust fault which lies approximately 3 mi (5 km) south of the area mapped on plate 1 and where major drag folding has occurred in the hanging walls of elements of the Grapevine fault system and of the Labbe fault (pl. 2). Also, some very low, but persistent, folds occur in the Stirling Quartzite at the east end of the district; and also the Johnnie Formation is disharmonically folded.

Deformation in the Johnnie Formation: Folding and ancillary features within the Johnnie Formation are local features related to the Sevier-orogeny shortening (Fleck, 1970) which is manifest as folding with accompanying thrust faulting elsewhere in the Sevier orogenic belt. The markedly contrasting styles of folding within the two areas separated by the Main and Northwest segments of the Grapevine fault system are summarized in the following table, and depicted in figures 4, 5, 6, and 7, and in cross section A-A' and C-C', plate 1:

	West of Grapevine	East of Grapevine
Geometry	Open, symmetrical	Tight, overturned assymmetrical to inclined isoclinal
Shape of axial surface	Planar	Non planar
Attitude of axial surface	N. 25° E, 83° E.	N. 8° E., 70° W.
Plunge of fold axis	19°, S. 24° W.	0°, N. 8° E.

The above parameters were derived from a structural analysis using 478 measurements of bedding (figs. 4 and 6) and 159 measurements of other features. The folds in both areas are disharmonic, as are the folds in the adjacent Mt. Stirling quadrangle (Vincelette, 1964). The

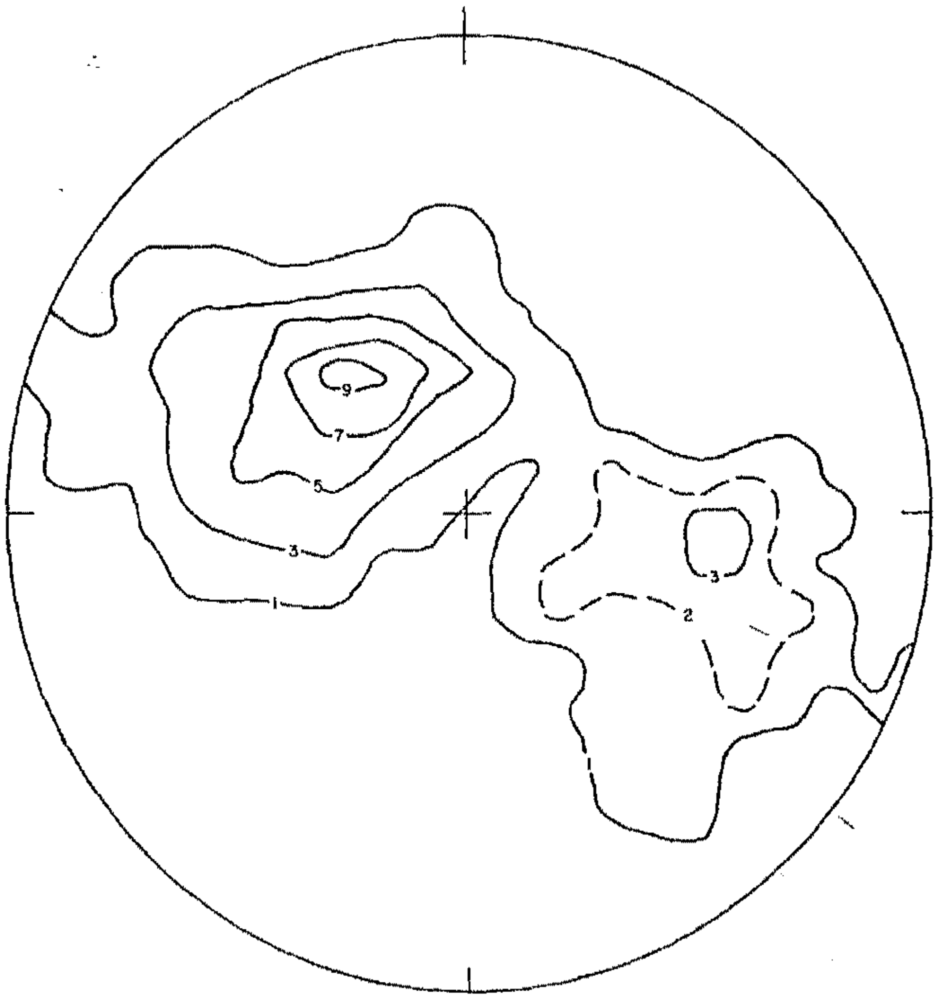

Figure 4. Contour diagram of lower hemisphere equal area (Schmidt) net plot of 299 poles to bedding in the Johnnie Formation in the area west of the Main and Northwest segments of the Grapevine fault system, showing fold geometry. Contoured at 1, 3, 5, 7, and 9 percent intervals with local supplemental contour at 2 percent.

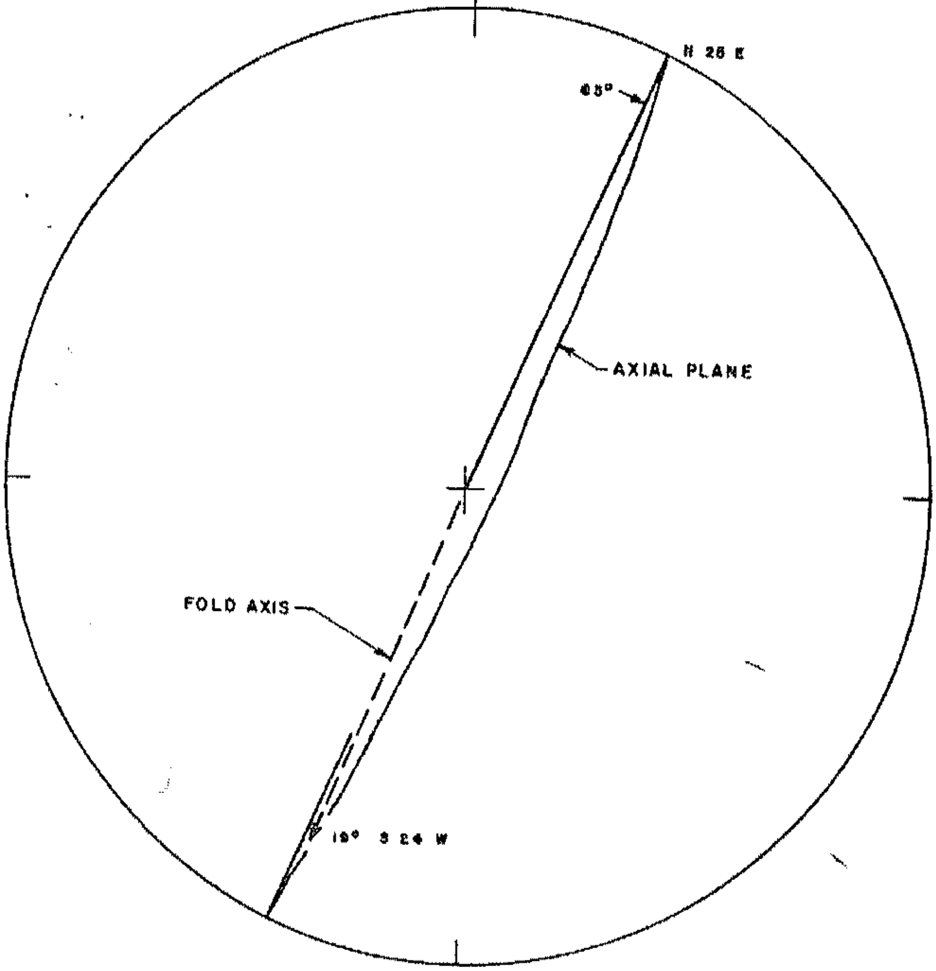

Figure 5. Lower hemisphere stereographic projection (Wulff net) of elements of folds in the Johnnie Formation in the area west of the Main and Northwest segments of the Grapevine fault system.

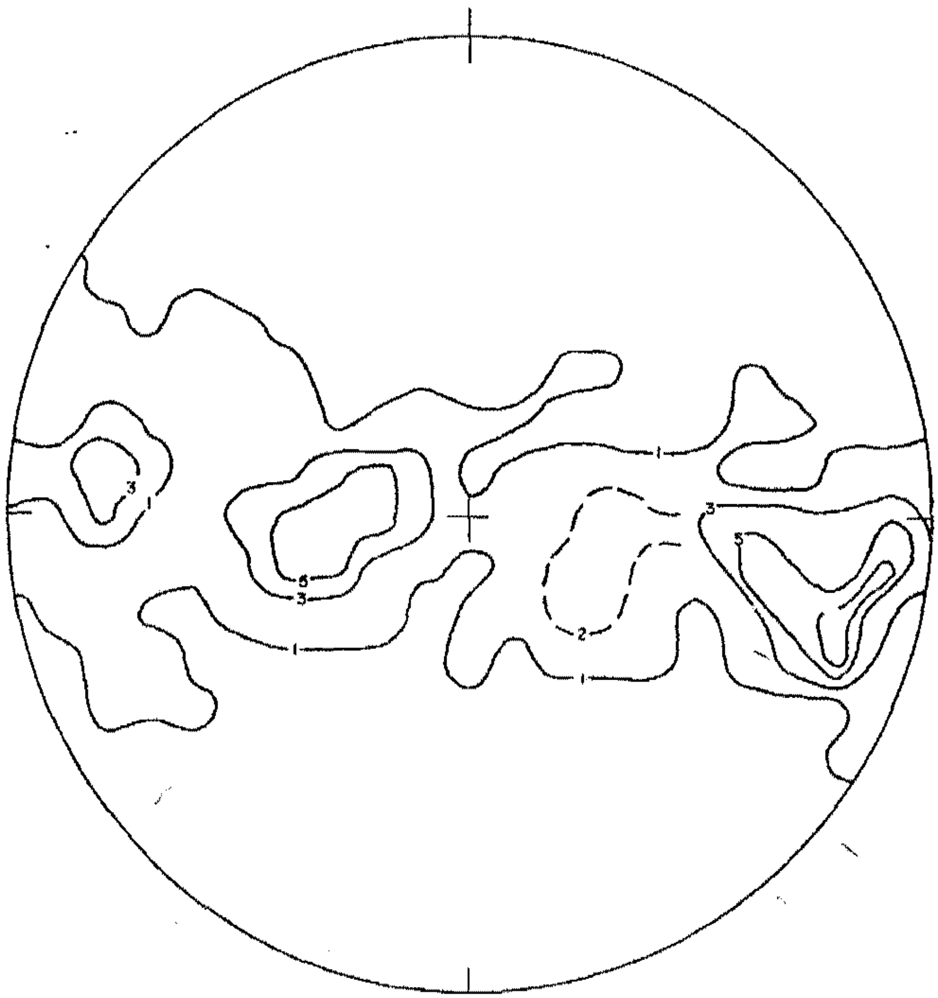

Figure 6. Contour diagram of lower hemisphere equal area (Schmidt) net plot of 179 poles to bedding in the Johnnie Formation in the area east of the Main and Northwest segments of the Grapevine fault system, showing fold geometry. Contoured at 1, 3, 5, and 7 percent intervals with local supplemental contour at 2 percent.

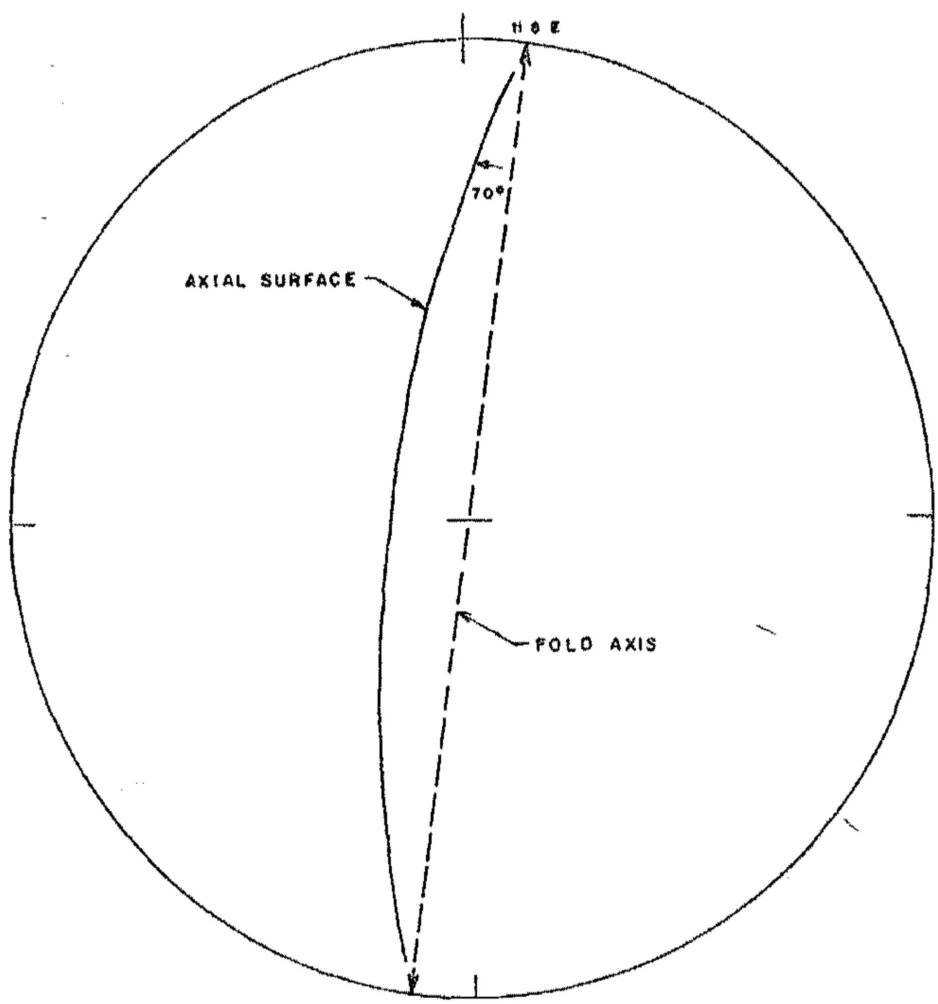

Figure 7. Lower hemisphere stereographic projection (Wulff net) of elements of folds in the Johnnie Formation in the area east of the Main and Northwest segments of the Grapevine fault system.

terminology summarized in Badgley (1965, p. 50-58) is used here and in the rest of this discussion on folding.

The open, symmetrical folds with subvertical axial planes in the western area are situated in an anticlinorial manner about the crest and the area immediately east of the crest of the Montgomery anticline, a feature named by Hamil (1966). The southward plunges of these lesser folds within the Johnnie district are explained by their being located south of the culmination of the Montgomery anticline.

The development of the folds higher in the Johnnie Formation in the area west of the Grapevine fault was accompanied by considerable eastward dipping low- to high-angle reverse faulting with subsidiary folding and step-bedding-plane thrusting and by east-trending transverse faulting. The interrelation between folds, reverse faults, and transverse faults in this western area is complex, but their development was essentially synchronous.

The westward dipping, non-plunging, tight, asymmetrical folds of the area to the east of the Main and Northwest segments of the Grapevine fault system may be further described as being of the non-planar cylindrical variety. In addition, they bear an incongruous relation to the enclosing major folds. The crests and troughs of some folds are recumbent. The fissile rocks at the base of the exposed part of the Formation at the head of Johnnie Wash are chevron folded. The pervasive shortening represented by reverse faulting in the western area is manifested here by moderate eastward dipping, concordant, isoclinal folding and local reverse faulting.

The difference in inclination of the axial surfaces of the folds in the Johnnie Formation in the areas on opposite sides of the Grapevine fault system is probably caused by rotation of the early formed vertical folds by subsequent folding of the entire miogeosynclinal stratigraphic section. This process is portrayed diagrammatically in figure 8. The overturned folds in the eastern area are on the mutual flank of the Montgomery anticline and complementary syncline to the east and are the rotated analogs of those vertical ones present along the crest of the Montgomery anticline; the average fold axes remain approximately perpendicular to the contact with the Stirling Quartzite. The somewhat incongruous nature of these folds indicates that they are not drag features formed during the broader scale folding.

Flexure on the mutual flank of the major folds resulted in the additional closing of, and distortion of, the axial planes of the disharmonic folds in the Johnnie Formation in the eastern area. Geologic maps of the Specter Range quadrangle (Livingston, 1964, pl. 1; Burchfiel, 1965, pl. 1) demonstrate that there is an even transition from one fold type to another across strike. The differences in trends of the fold axes on either side of the fault system derives from post-folding rotation during movement along the Grapevine fault system.

Alternative, less favorable, explanations for the disparity of fold morphology are: (1) the Grapevine fault system is an old system across which different styles of deformation occurred; or (2) that horizontal displacement along the fault has brought folds into juxtaposition which were formed originally at great distances from each other under different tectonic circumstances.

The style of deformation in the Johnnie Formation differs from

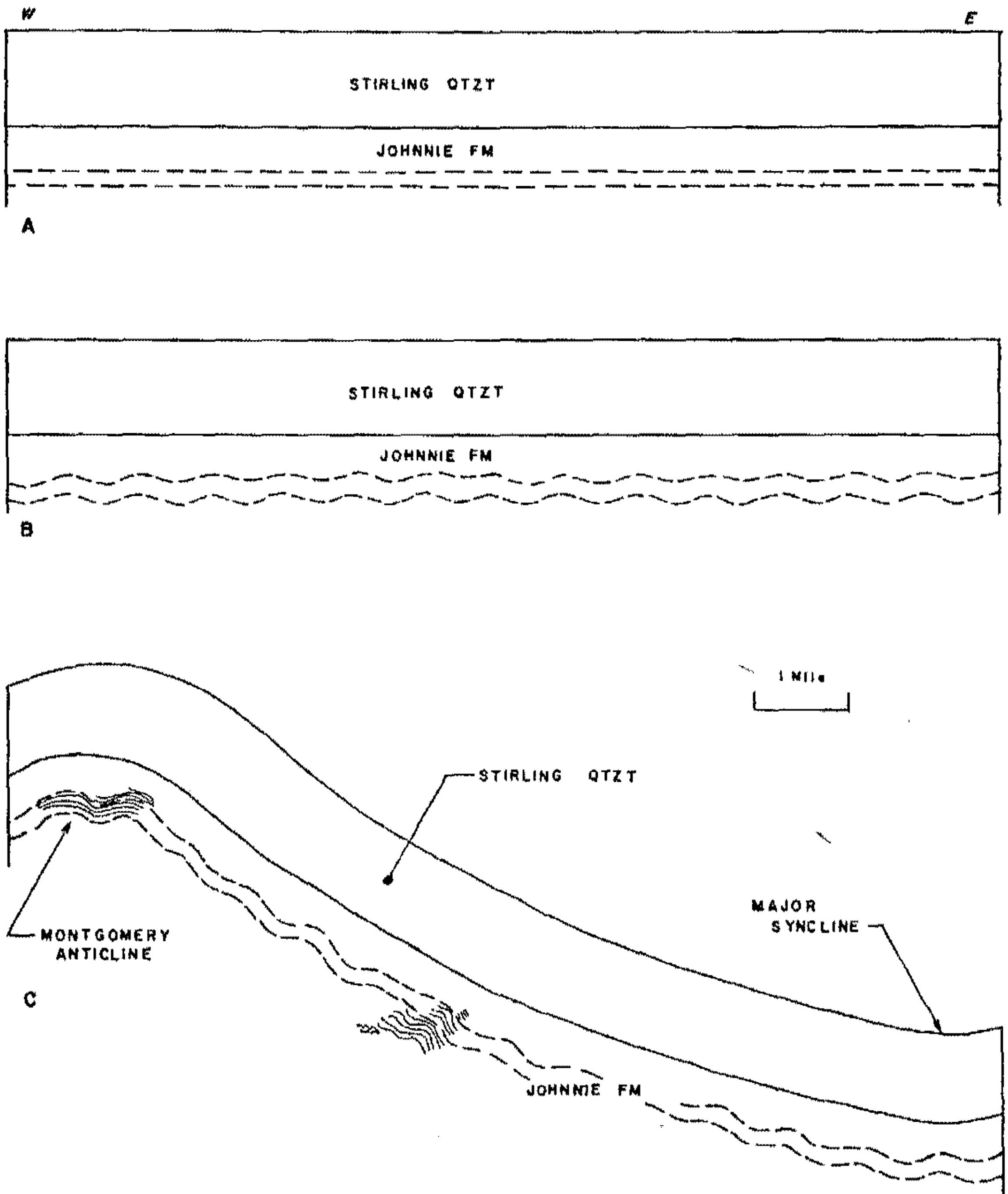

Figure 8. Diagram illustrating sequence of folding in the northwestern Spring Mountains and northern Montgomery Mountains. (A) Attitude of rocks before folding. (B) Disharmonic folding of the Johnnie Formation. (C) Folding of the entire stratigraphic section, showing disharmonic folding observed in the Johnnie district. For approximate locations of Montgomery anticline and the major syncline, see figure 2.

that in the overlying formations because of its gross relative incompetency: it is composed almost entirely of thin- to medium-bedded rocks and lacks any massive units. In addition, the massive basal member of the Stirling Quartzite confined local stresses to within the Johnnie Formation. Similar disharmonic relations are noted within other relatively incompetent stratigraphic units in the Spring Mountains by Hewett (1931, 1956), Vincelette (1964), Burchfiel and Davis (1971), Burchfiel and others (1974), and personal reconnaissance.

The disharmonic folding of the Johnnie Formation can be explained on theoretical grounds by simplifying from Biot (1961). The relative incompetency and higher temperature, derived from greater depth of burial, of the Johnnie Formation rendered it more subject to flow than the overlying formations. This permitted the rocks of the Johnnie Formation to fold at the initially low compressive stress applied at the onset of orogeny. With increased stress, the weight and greater rigidity of the overlying formations, which had inhibited their folding earlier, were overcome; and folding proceeded in the post-Johnnie Formation rocks.

High-angle reverse faulting and isoclinal folding in the upper part of the Johnnie Formation probably resulted from the shearing effect of a force couple between the Stirling Quartzite and the underlying Johnnie Formation during the updip translation of the Stirling Quartzite along the contact with the Johnnie Formation during the large-scale folding of the region which followed disharmonic folding.

Folding is one of the oldest structural events recorded in the Johnnie district, having been coeval with some transverse faulting but later than some fracturing. The relatively tight folding of the Johnnie Formation was followed promptly by the regional-scale folding responsible for the Montgomery anticline and the major syncline to the east. The later folding caused shearing at the top of the Johnnie Formation, westward tilting of the earlier folds along the north and northeast edge of the district, and eastward tilting of the post-Johnnie Formation rocks of the district. Concomitant, or slightly later, thrust faulting caused the bending at the south end of the district.

Folding and related events, obviously cogenetic with the Sevier orogeny, must be of the same, Late Cretaceous, age.

<u>Contact between the Johnnie Formation and Stirling Quartzite</u>: The nature of the contact between the Johnnie Formation and the Stirling Quartzite has been the topic of considerable discussion in the literature. Nolan (1924, 1929) named the contact the "Johnnie thrust" and speculated that it was a major decollement surface from which the Wheeler Pass thrust fault (fig. 2), to the southeast, originated. Burchfiel (1961, 1965) and Burchfiel and Davis (1971) believe that a decollement surface complements the major thrust faults throughout the Spring Mountains; but, to the contrary, Fleck (1967, 1970) demonstrates that the major thrust faults in the Spring Mountains are of more local origin. Vincelette (1964) concludes that the Johnnie-Stirling contact in the Mt. Stirling quadrangle immediately east of the Johnnie district is of a normal sedimentary character and that, if any decollement surface is present, it is not exposed. Hamil (1966) reaches a similar conclusion for the contact within the Mt. Schader quadrangle.

From examination of cross sections (for example, fig. 3), the Montgomery thrust certainly must underlie the Johnnie district at depth, a concept also stated by Burchfiel (1965). But, whether it continues to dip steeply into its source area or flattens into a decollement surface remains problematical. (See Johnnie Formation.)

However, it remains that an anomalous structure is present at the contact of the Johnnie Formation and Stirling Quartzite. This is seen in: (1) strong brecciation of the Stirling Quartzite; (2) shearing and reverse faulting within the upper part of the Johnnie Formation; and (3) thinning by 450 ft (135 m) of the upper part of the Rainstorm Member of the Johnnie Formation along that portion of the contact with the Stirling Quartzite mapped on plate 1 as a zone of tectonic readjustment southward from just north of the mouth of Johnnie Wash.

This anomaly may be sedimentary, in part. Previous workers (Burchfiel 1964, 1965) note sedimentary irregularities near the top of the Johnnie Formation. Others (Vincelette, 1964; Stewart, 1970) observe variations in the distance from the Johnnie oolite to the top of the Johnnie Formation.

It is incorrect to describe the disturbance at this portion of the contact as a fault, because I observe no truncation of strata below the contact. Vincelette (1964) makes a similar observation in anomalous areas in the Mt. Stirling quadrangle. In addition, the divergence of the contact from the Johnnie oolite is very small. This is similar to an observation in the Mt. Stirling quadrangle by Vincelette (1964) who notes that this is a remarkable concordance for a decollement thrust fault.

Hamil (1966) describes a "downthrust" fault in the southern Mt. Schader quadrangle in which "thrust movement is distributed throughout a complex zone of shear" to bring and deposit younger rocks upon older with the stratigraphic sequence being maintained but the stratigraphic thickness being greatly reduced; such an account cannot be given here. Finally, whereas this disturbance is of Late Cretaceous age, most of the numerous structural features described which truly transpose younger across older rocks in the region are of young, say post-early Cenozoic, age.

There is a correlation between proximity to the crest of the Montgomery anticline, tectonic disturbance of the contact, thinning at the top of the Johnnie Formation, and the intensity of folding and related deformation. The minor isoclinal folding in the area east of the Main and North segments of the Grapevine fault system gives way to increasing amounts of reverse faulting through the area east of the Northwest segment and continues to increase into the area to the west which faces the Amargosa Desert. Then, the amount of reverse faulting, general shearing, and formational thinning increase southward along the tectonic contact under discussion. This suggests that the disturbance near the contact is related, not to thrusting, but to interbed shear caused by flexural slip during the large-scale folding of the region.

This is contrary to the classic example of flexural-slip folding in which slippage is confined to the flanks of folds and absent from their crests and troughs. However, Biot (1961) observes that, as folding proceeds, fracturing becomes concentrated at the crests and that this generally weakens the rocks there. The latter mechanism would have made the rocks near the top of the Johnnie Formation near the crest of the Montgomery anticline relatively more amenable to thinning by a process of distributive shear.

In summary, it seems that the anomolous appearance of the upper Johnnie Formation and its contact with the Stirling Quartzite derives from a folding related distributive shear near the crest of the Montgomery anticline during the Late Cretaceous Sevier orogeny, and it possibly derives partially from sedimentary thinning.

High-Angle and Related Structures

Introduction: High-angle structures present in the Johnnie district include faults, quartz veins localized in high-angle and ancillary structures (see Ore Deposits), fractures, and joints (see fig. 9). Fractures are here defined as structures, which do not host veins, along which no or minor displacement has occurred. Joints were not systematically studied during this project.

Many of the conclusions presented here are derived from a structural analysis using various measurements of 635 features. The analysis permitted average parameters and logical groupings of the features to be determined. The determination was to some extent arbitrary; and, although some parameters could be misplaced, the errors are of insufficient magnitude to materially affect the results of this study.

Antecedent Fractures: Four sets of early formed fractures played a continuing role in the deformational history of the district. Recognized on the basis of a consideration of all of the high-angle structures in the district, these are, in order of decreasing age, an

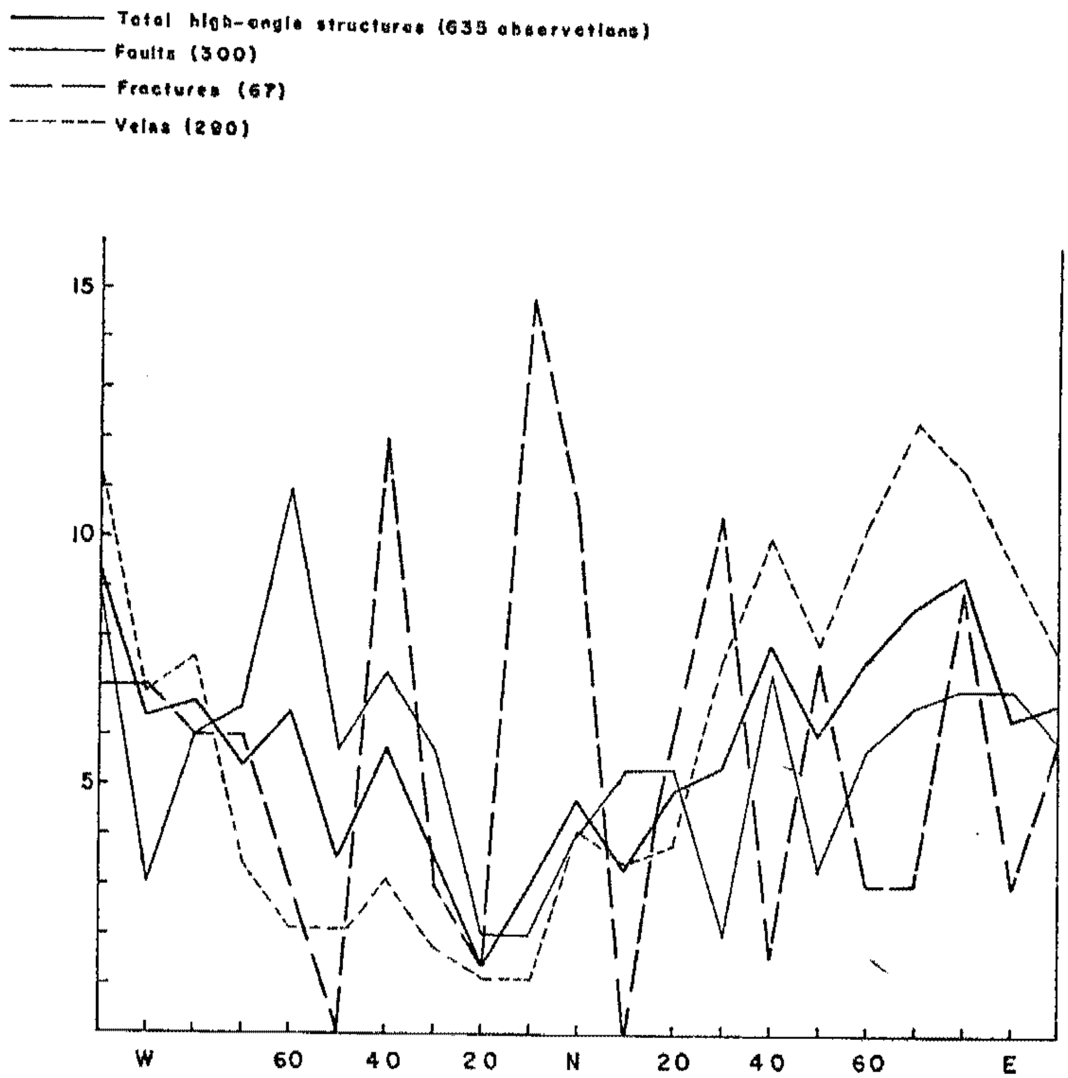

Figure 9. Strike frequency diagram of high-angle and related structures.

extension fracture set which strikes approximately east, two sets of conjugate fractures, one northwest and one ENE, and an approximately north-trending pressure-release fracture set. The characterisitics of these are illustrated in figure 10.

The extension fracture set probably developed at the onset of the Sevier orogeny and this fracturing was promptly followed by formation of the conjugate fracture set in response to east-oriented compression in the force field depicted in figure 11. These conclusions are suggested by the symmetrical disposition of the fracture sets about the strike and dip of bedding, which, in turn, reflects the orientation of the orogenic belt (fig. 1).

The conjugate fracture sets developed with a conjugate shear angle of 81° (fig. 11). This is a bit higher than predicted by shear failure theory, and it approaches the theoretical upper limit of 90°. This may result from the improper selection of values of the average conjugate fractures.

Folding ensued under continued application of similar stresses, rotating the structures, which previously were formed perpendicular to bedding, to their present attitudes as shown in figure 12. With the relaxation of compression, during which folding ceased, the force field became reversed and the now vertical pressure-release fractures developed.

It is possible to derive this simple picture of fracture genesis because of the homogeniety of this eastward dipping structural domain. Fractures dipping opposite to all of those discussed so far also formed; although less numerous than those shown in figure 10, in some local

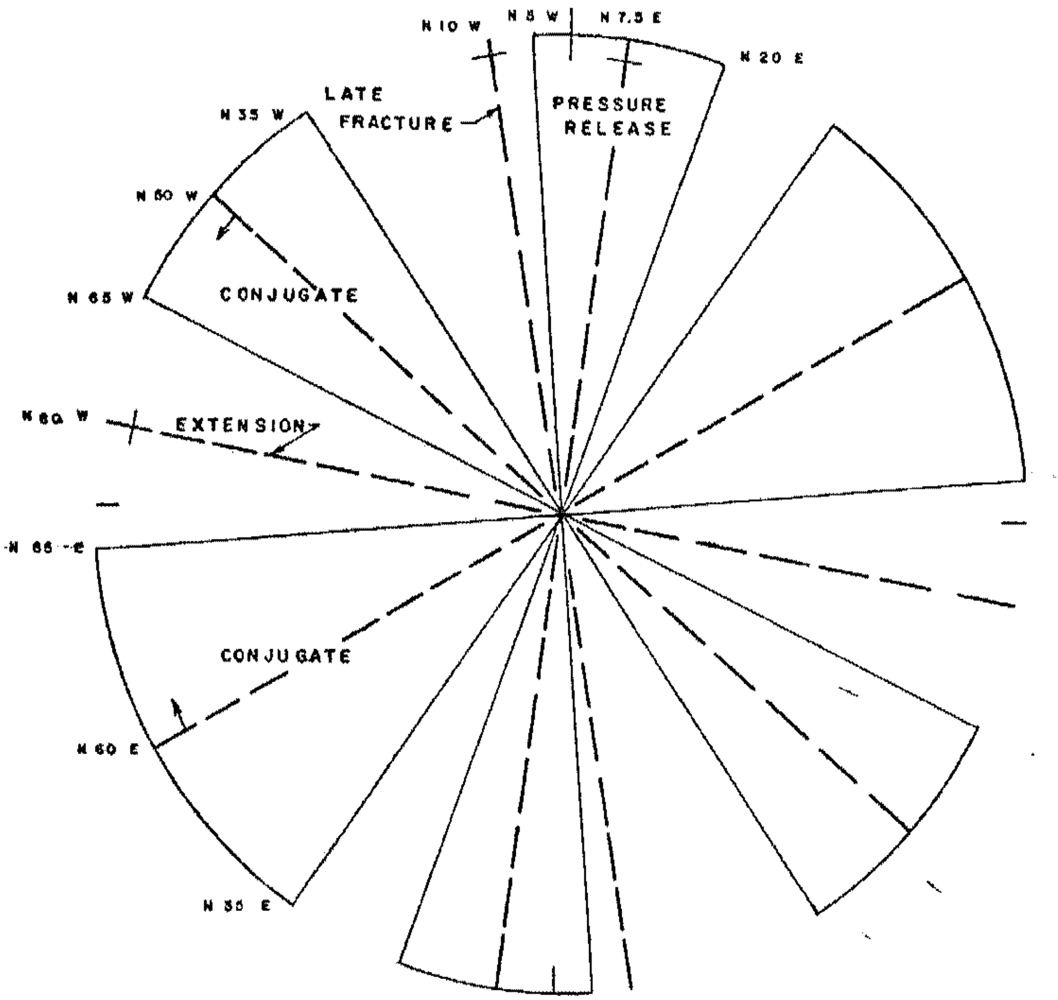

Figure 10. Diagram illustrating strike and direction of dip of average fractures (broken lines) and ranges of fracture sets. Ranges of extension and late fractures are narrow.

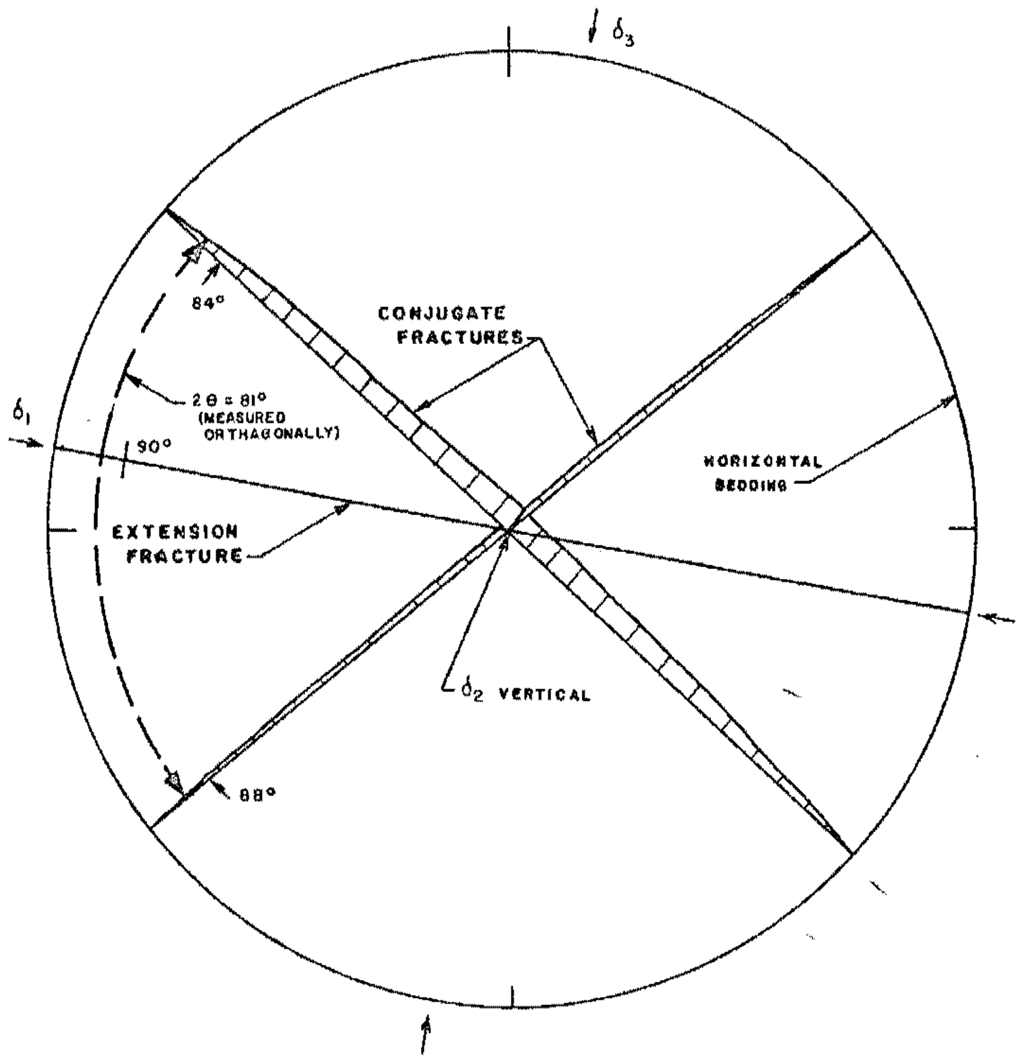

Figure 11. Lower hemisphere stereographic projection (Wulff net) of fractures and force field before folding. Obtained by rotating features by the amount required to restore the average bedding to horizontal.

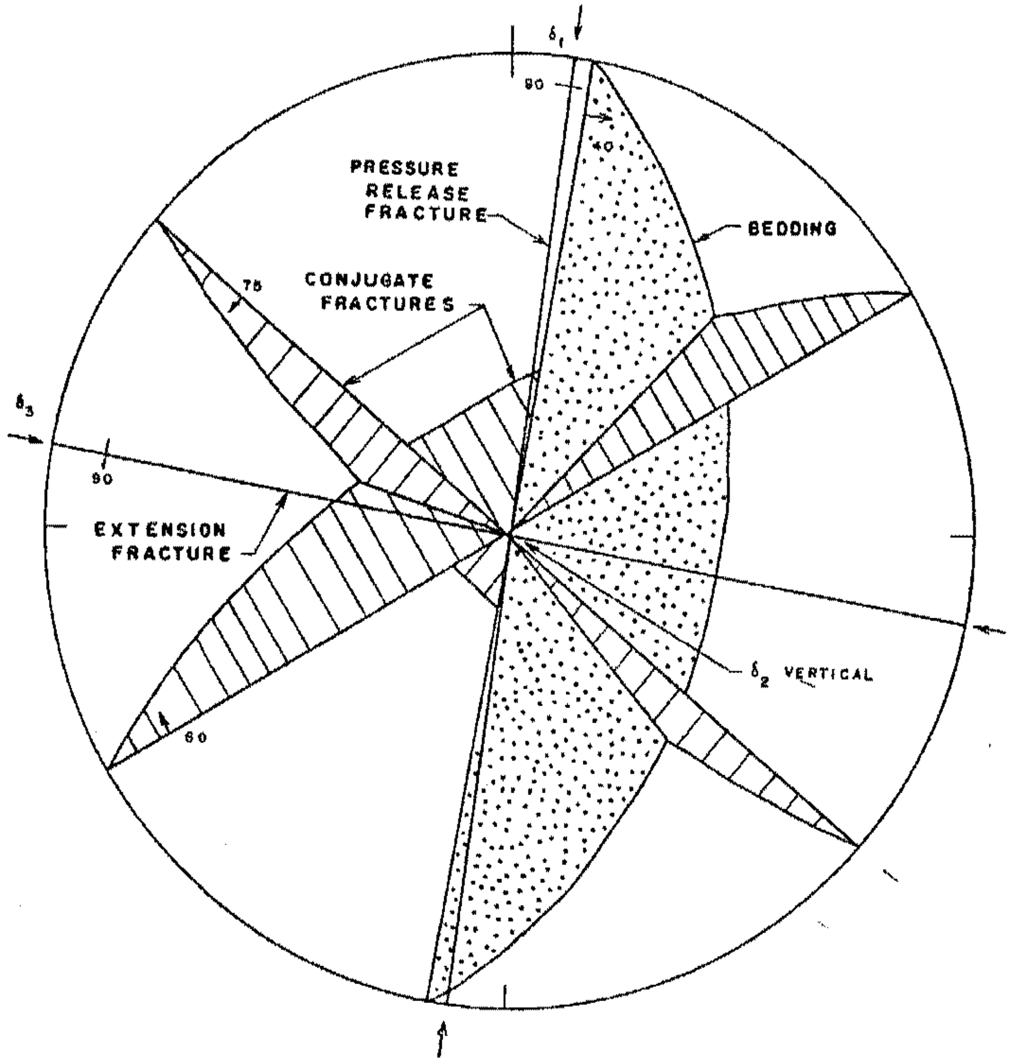

Figure 12. Lower hemisphere stereographic projection (Wulff net) of fractures and force field after folding (present attitudes of fractures and bedding).

cases they were the favored planes of structural development. In addition, a minor number of fractures developed between the ranges given for the various fracture sets.

Later Utilization of Antecedent Fractures: I believe that there was relatively little dislocation upon any of these fractures during the early stages of the structural evolution of the district. However, dislocation along the early fractures and bedding planes served to accommodate later displacement. All other features--faults, fractures, and quartz veins--can be related to these precursory fractures except for a single set which strikes N. 10° W. (see Fractures).

Examples of this are seen where there are gaps in the statistical distribution of faults and veins corresponding to areas intermediate between antecedent fracture sets (compare figs. 9, 14, and 21). In other words, few or no new fractures formed during faulting. Movement occurred along preexistent planes of anisotropy. Conversely, many transverse faults lie parallel to or merge with fractures along which no movement has occurred. Figure 13 shows a situation in which a portion, only, of at least one fracture was utilized in a displacement of the Zabriskie Quartzite.

Fractures: Fractures were not recorded systematically during this study. The 67 observations made in conjunction with the study of other structural features indicate that fractures are found in the same orientations as are faults and quartz veins.

The fracture set, probably late, which strikes N. 10° W. (fig. 10) has no earlier counterpart and bears no obvious relation to any faults

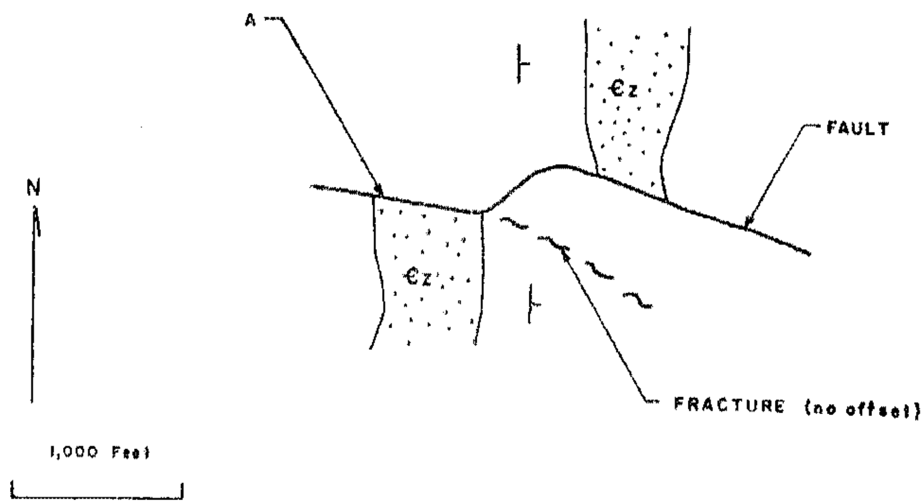

Figure 13. Interpretation of geology at the mutual corner of secs. 2, 3, 10, and 11, T. 18 S., R. 52 E. Fault segment "A" derived from movement along corresponding segment of preexistent fracture.

or veins identified in the district.

High-Angle Faults: High-angle faults fall into two broad sets in the Johnnie district, transverse and longitudinal, whose parameters are given in figure 14. They are related to the antecedent fracture sets as follows: transverse faults represent a portion of both conjugate fracture sets and the intermediate group of extension fractures; longitudinal faults include a portion of the ENE conjugate fracture set and adjacent pressure-release set.

The faults comprise four broad genetic groups (see fig. 14):

(1) Common transverse faults. This category of faults lies between the western limit and middle of the range of the transverse group and strikes somewhat perpendicularly to bedding. Displacement along these appears to have been largely strike slip, commonly as a secondary readjustment to displacement along other structural features. Examples are the transverse tear faults bounding the ends of the plate of rock displaced by the Congress low-angle normal fault (pl. 1), the transverse faults associated with the dextral bend in the vicinity of Mt. Schader (pl. 2), or the transverse faults associated with disharmonic folding discussed under Deformation in the Johnnie Formation.

(2) Transverse faults of the Grapevine trend. This includes the Main and Northwest segments of the Grapevine fault system, the Labbe and possibly some nearby faults to the west and east, and the largely concealed fault striking towards Route 16 from the Nomad No. 1 claim (pl. 3). This genetic group falls near the northern limit of the range of the transverse fault group,

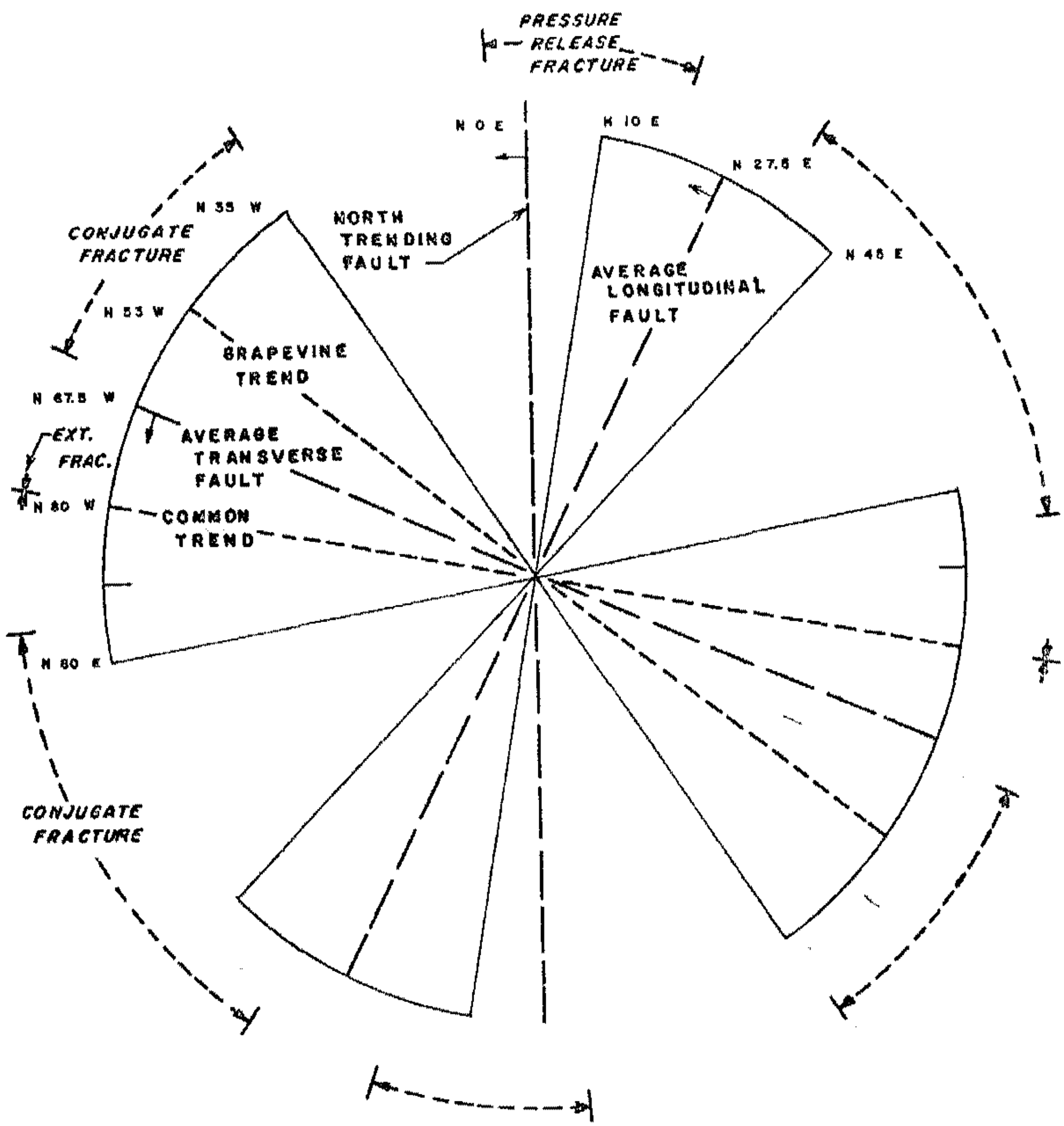

Figure 14. Diagram illustrating strike and direction of dip and ranges of genetic and statistical groups of faults and also illustrating their correspondence to fracture sets.

opposite from the majority of the remainder of the transverse faults. Displacement along them is largely dip slip, west side down, although the Grapevine fault system may be in part, an exception to this.

(3) Longitudinal faults. These occur in two places--along the Congress fault system, including its northern extension, and along the axial portions of some folds in the Johnnie Formation facing the Amargosa Desert. Movement along these appears to have been wholly dip slip, commonly with the west side down.

(4) North-trending faults. These faults comprise a statistically minor group in the Johnnie district. They occur east of the Main and Northwest segments of the Grapevine fault system where they exhibit apparent normal separation, usually with the west side down. They apparently belong to a group of north-trending faults which are of greater numerical importance (Burchfiel, 1965) in the adjacent southeast part of the Specter Range quadrangle.

If the two transverse fault groups are lumped together, then three regionally significant high-angle fault trends, which strike generally northwest, north, and northeast are recognized. These broadly correspond to the three groups, each, recognized by Burchfiel (1965) in the Specter Range quadrangle, and by Vincelette (1964) in the Mt. Stirling quadrangle.

The interrelationships among the groups of faults in the Johnnie district are unclear, except as noted that common transverse faults operate in conjunction with other faults. No simple stress field can be deduced which relates the pattern and senses of motion of the two main sets of faults in a conjugate manner. This is probably because more than one period of faulting, involving different stress fields, occurred. In addition, the antecedent fractures available for displacement to occur along may not have been in the classic orientation with the stress field or one set of faults may bear a lower order, not a conjugate, relation to the other.

Two major fault phenomena, the essentially longitudinal Congress fault system and its northward extension and the transverse Grapevine fault system, seem to be the products of separate events.

Congress Fault System and Related Structures: This is an arcuate system of high-angle faults of apparent dip slip displacement. It extends from the low hills in the alluviated area south of the Labbe mine to the area immediately west of the site of Johnnie where it diverges about a horst-like structure; thence, the Congress fault system proceeds southwest, and an apparent group of transverse faults trends southeastward toward Route 16 (pls. 1 and 3). This arcuate system may extend northward along the Labbe and adjacent faults up the south face of Mt. Schader. The system, as a whole, follows the longitudinal fault trend but is composed of individual longitudinal and transverse segments.

The Congress fault, itself (cross section D-D', pl. 1), is a low-angle normal fault analogous to those described by Longwell (1945) in the northward equivalents of the Spring Mountains in northern Clark County. It is a linear, steeply dipping structure northwest and southeast of the Congress mine but flattens with depth northwestward to emerge subhorizontally in places along the bluff overlooking the Amargosa Desert. A number of smaller, complementary, step like, convex features which also flatten at depth occur in the footwall of the fault southwest of the Congress mine.

The toe of the Congress fault apparently is deflected upward by the massive basal unit of the Stirling Quartzite which, in turn, is somewhat flattened there to produce the bend present in its contact with the underlying Johnnie Formation. The north and south ends of the toe emerge at the surface but the center of the toe dies out below the surface before it can emerge. The morphology of the flat portion of the fault displays a complex relation between transverse tear faults in the upper and lower plates and transverse faults which cut both plates.

As a low-angle normal fault, the Congress fault lies in an expectable position on the flank of the Montgomery anticline. It is in the correct position in the Sevier orogenic belt to be a feature which developed by the relaxation of east-oriented compression with the close of folding. Deformation in the toe of the fault resembles and seems to grade into that at the top of the Johnnie Formation discussed previously.

The implied kinship of deformation at the Johnnie-Stirling contact to the Congress fault indicates that the Congress fault is an older feature within the geochronologic framework of the Johnnie district. In addition: (1) elements of the Congress fault system host quartz veins; (2) other elements of the system localize dolomitization (see <u>Tectonic Alteration in the Bonanza King Formation</u>), which probably precedes quartz veining; and (3) minor folding and offsetting of beds caused by Congress-fault deformation give the impression of having taken place, in part, under conditions approaching ductile flow. This type of deformation probably could have occurred only under the tectonic load present earlier in the geologic history of the Johnnie district.

Considering the relationship between the Congress fault and the Montgomery anticline and considering its old age, the fault is taken to be a low-angle normal fault formed during relaxation of stress at the close of the Sevier orogeny. The origin and significance of the incompletely exposed remainder of the arcuate fault system is unclear, but it is probably related. A stress relaxation origin is also inferred for the normal, longitudinal faults near the crests of folds in the Johnnie Formation along the edge of the Amargosa Desert.

Grapevine and Related Fault Systems: The Grapevine fault system is part of a larger system of known and inferred high-angle faults which extends from 6 mi (10 km) north of the center of the Johnnie district (Livingston, 1964, pl. 1) to at least as far as the mouth of Wheeler Wash, 15 mi (24 km) south (Vincelette, 1964, pl. 1). The southern part of this system, which bounds the west face of the Spring Mountains, is named the Upper Pahrump Valley fault by Hamil (1966). (See fig. 2.)

Previous workers (Nolan, 1924; Hamil, 1966; Cornwall, 1972) portray the Grapevine fault as a single linear element crossing Johnnie Wash to connect what are shown as the Main and Northwest segments of the fault system on plate 1. Mapping conducted during this project shows that this actually is a composite feature in which early displacement along the path just described later was supplanted by displacement along the Main and North segments with accompanying abandonment and bending of the Northwest segment.

Examination of the minimum amounts of stratigraphic displacement or dip separation along the segments involved partially confirms this sequence of events. The aggregate displacement of 4,700 ft (1,433 m) along the North segment and the parallel fault immediately west added to the 2,500 ft (762 m) of displacement along the Northwest segment totals 7,200 ft (2,195 m). This is nearly identical to the 7,400 ft (2,256 m) measured on the Main segment northeast of Hill 3907 just south of the inferred junction between the Northwest segment with the Main and North segments.

It is difficult to determine precisely the origin of the Upper Pahrump Valley fault system because of its complex history, involved morphology, and the fact that it may offset the possibly once continuous Montgomery and Wheeler Pass thrust faults (fig. 2) by 8 or 9 mi (13-14 km), as estimated by Cornwall (1972). However, the Grapevine and associated faults probably should be considered Basin-and-Range frontal faults, because:

(1) They are in the correct orientation.

(2) Typical range front morphology is present.

(3) Considerable normal fault movement can be demonstrated. The synclinal drag north and east of the Johnnie mine is apparently a manifestation of such displacement since it is identical to the drag fold in the footwall of the Labbe fault. The latter is obviously a normal fault with nearly identical trend to but of lesser magnitude than the Grapevine fault.

(4) Displacement along the Upper Pahrump Valley fault system has continued into fairly recent times (Vincelette, 1964, pl. 1; Hamil, 1966; Cornwall, 1972, pl. 1; this examination) which is typical of Basin-and-Range faults.

However, earliest movement on the Grapevine fault system predated middle to late Miocene Basin-and-Range faulting, because: (1) it hosts quartz veins, which I believe also predate Basin-and-Range faulting; (2) its origin is complex; and (3) displacement of the Montgomery and Wheeler Pass thrust faults, if once continuous, by the Upper Pahrump Valley fault system could not have resulted simply from dip slip.

The fault system could be an old feature which predates the thrusting of the Sevier orogeny and across which deformation took different courses or it could be a tear fault in the overthrust plate which developed synchronously with thrusting as Hamil (1966) suggests. The objection to the first possibility is that there is no other instance of this type of deformation in the Spring Mountains or vicinity. The Upper Pahrump Valley fault system, including parts of the Grapevine fault system, could be a right-lateral-slip fault akin to the middle Miocene (Fleck, 1967) Las Vegas Valley shear zone (fig. 2) which does offset major thrust faults within the region. This explanation is not wholly acceptable because: (1) the Upper Pahrump Valley fault system is not in the same orientation; (2) it possesses different morphology; (3) elements of the Grapevine fault system, at least in part, predate quartz veining in the Johnnie district, which veining I believe predates activity of the Las Vegas Valley shear zone type; and (4) the Grapevine fault system exhibits apparent left-lateral and normal, as well as right-lateral drag patterns in the district.

Thus, the fault systems are old, but post folding and thrusting, features. They developed through a combination of actions, to include some right-lateral displacement, which possibly was sympathetic with displacement along the Las Vegas Valley shear zone; but they were profoundly affected by Basin-and-Range faulting.

Dextral Bend

A prominent zone of dextral bending of a part of the district about a subvertical axis deserves mention as a separate structure. Exposures of the Stirling Quartzite west and east of Mt. Schader are bent in the following manner:

(1) The medium-bedded quartzites of the B member of Stewart (1970) are offset by transverse faults of apparent right-lateral displacement which attenuate down section and do not penetrate the massive A member (pl. 1).

(2) The same transverse faults continue up section where they merge into a zone of bending within the more thinly bedded C and B members (pls. 1 and 2).

(3) The massive A member is bent and slides across the B member along a zone of tectonic readjustment (pls. 1 and 2).

The bend seems to be young, because so many lateral movements are involved and this implies a relationship to later strike separation tectonics. Bending must postdate quartz veining, in part, because some quartz veins appear to be localized in older structures reactivated by bending.

This dextral bend is probably a right-lateral drag feature caused by displacement along the Grapevine fault system, but how much displacement was required to produce this amount of bending is unknown.

Low-Angle Structures

This includes a group of similar appearing, near concordant, subhorizontal structural features of diverse origins which transpose younger rocks across older ones. Some of the problems in dealing with these are discussed in the section, Contact between the Johnnie Formation and Stirling Quartzite. Four types of low-angle structures are present in the Johnnie district:

(1) Essentially concordant zones of tectonic readjustment within which thin-bedded rocks beneath massive quartzite or carbonate units are thinned out with some shearing, but with no truncation of the stratigraphic section: these developed in response to updip translation of the section during folding and during dragging in the hanging walls of segments of the Grapevine fault system.

(2) Similar appearing features, shown as low-angle faults on the accompanying geologic maps, transect the section at low angles: these occur immediately east and northeast of the Congress mine (pl. 3), immediately south of the Lilyan claims (pl. 2), and elsewhere. These evidently grade laterally into zones of tectonic readjustment; and the difference between this typical low-angle structure and zones of tectonic readjustment often being only one of degree. However, in discussing these two situations, I believe that I do not adequately account for missing or attenuated strata.

(3) The subhorizontal toe of the Congress fault is a low-angle fault. However, pervasive shear within and below it, as well as its near concordant attitude, cause it to be confused with the preceding two features. Some of the amount of the total deformation in each probably occurred at approximately the same time under similar conditions of tectonic load to heighten the similarity of appearance.

(4) The low-angle fault which extends from east of the Johnnie mine to west of the Grapevine Springs (cross section B-B', pl. 1) removes a wedge-shaped section of rocks from the dolomite-bearing upper part of the upper member of the Wood Canyon Formation (units 2 and 3, Appendix A) in the lower plate through the lower part of the Bonanza King Formation in the upper plate. It doubtlessly is another drag feature in the footwall of segments of the Grapevine fault system. Its relation to an overlying zone of tectonic readjustment at the base of the Bonanza King Formation suggests a genetic relation between the two types of deformation. However, the low-angle fault probably postdates quartz veining in the district, because it hosts none and because the quartz veined megabreccia to the southeast may be part of the wedge of rocks cut out by this fault after being veined (see <u>Megabreccia Unit of Older Fanglomerate</u>).

Tectonic Alteration in the Bonanza King Formation

Two unrelated types of apparently late tectonic carbonatization affect the Bonanza King Formation.

Elements of the Congress fault system, or transverse faults divergent therefrom, in the vicinity of the Congress mine are the sites of limited amounts (not portrayed on maps herein) of dolomitization or recrystallization of more limy parts of the Bonanza King Formation. Carbonatization is most common there along fault contacts with the Carrara Formation. Longwell and others (1965) note that such alteration commonly occurred sometime after thrust faulting and before Basin-and-Range normal faulting in the Spring Mountains and adjacent areas. Hewett (1931) makes the same observation in the southern Spring Mountains and places the time of dolomitization somewhat closer to the episode of thrust faulting and distinctly before mineralization. Therefore, by analogy with these authors' conclusions, dolomitization near the Congress mine is of comparable age.

A few thin veinlets of white, coarsely-cleavable calcite and others of similar dolomite occur locally in the Bonanza King Formation. These strongly resemble calcite veins and veinlets observed in the Bonanza King Formation 9 mi (14 km) north of the Johnnie district in the Specter Range quadrangle immediately north of Route 95 in sec. 12, T. 16 S., R. 52 E. (fig. 2). The veins in the Specter Range quadrangle demark faults and subsidiary features, and they probably originated from the local mobilization and recrystallization of calcite derived from carbonate gouge during faulting. Hence, the veinlets of calcite and of dolomite in the Johnnie district are probably the local products of tectonic activity, but no unique age can be assigned to them.

Geomorphology

Successively higher terracelike topographic features and accordant prominences throughout the Johnnie district and adjacent mountains are remnants of erosional surfaces which attest to the steplike degradation of the terrain during successive stages of Basin-and-Range faulting. The lowest of these features apparently are the exhumed remnants of several dissected, coeval pediments. Pediments are recognized also in nearby areas by Denny (1965, 1967) and Denny and Drewes (1965).

The pediments in the Johnnie district are located in the low hills facing the Amargosa Desert at elevations up to 3,000 ft (915 m); the higher of these hills probably are relict inselbergs. Other of these surfaces occur to elevations of about 3,700 ft (1,130 m) east of the Northwest segment of the Grapevine fault system; to 4,000 ft (1,220 m) east of the Main segment at the head of Johnnie Wash; and to about 3,900 ft (1,190 m) along the northwest and north edges of the Pahrump Valley. The pediments and older fanglomerate which filled channels in them and then buried them are deeply dissected, indicating later exhumation by erosion.

Deposits of older fanglomerate overlying the pediments are the remains of a bajada which ultimately buried the pediment. The accordant surfaces of older fanglomerate (not mapped during this project) in the area south of Grapevine Springs and west of the Main segment of the Grapevine fault may be relics of the surface of that feature protected by post depositional downfaulting from the subsequent vigorous erosion which exposed the other pediments in the district.

Being the youngest erosional surface of note, the pediments post-date the latest significant Basin-and-Range faulting, which was at least of an estimated late Miocene age, and are older than the earliest deposits of late Pliocene to middle Pleistocene older fanglomerate. So, pedimentation, later dissection, and burial can only be dated as having occurred between the latest Miocene to middle Pleistocene epochs.

Geologic History

The lower miogeosynclinal clastic and carbonate rocks which are exposed in the Johnnie district record only a portion of the total Precambrian and Paleozoic stratigraphic section which was deposited there. Earlier miogeosynclinal sedimentary rocks are not exposed in the district, and approximately 30,000 ft (9,000 m) of younger rocks have been eroded away.

Evidence of pre-Late Cretaceous deformation, recorded elsewhere in the region, is not present here; unless some of the earliest formed fracture sets date to earlier orogeny within, and parallel to the trend of, the Cordillera.

With the onset of the Late Cretaceous Sevier orogeny, early extension fractures, then later conjugate fractures, developed under a stress field in which the principal stress axis was oriented eastward, across the axis of the orogenic belt. Folding, with some transverse faulting, followed, under the influence of the same stress conditions, and rotated these fractures to their present positions. Folding began with small-scale, disharmonic folding within the Johnnie Formation and finished with larger scale regional folding, which rotated some of the earlier formed folds to their present inclined positions. Folding was accompanied by gliding and dragging along zones of tectonic readjustment, by westward oriented, longitudinal reverse faulting, and by shear along the contact between the Johnnie Formation and Stirling Quartzite. Folding was promptly followed by the translation of the Johnnie district eastward along the Montgomery thrust fault, which apparently underlies the district; but no obvious surface manifestation of thrusting is present beyond a deflection in bedding.

A pressure-release fracture set developed with the relaxation of stress at the close of folding and thrusting. All of the fractures hitherto named became the ancestors of all of the faults which were to form. An additional set, oriented N. 10° W., also developed, probably sometime later.

With the continued relaxation of horizontal compression, longitudinal faults--such as the Congress fault system--developed. Transverse faulting complemented movement along some of these. This faulting is probably of latest Cretaceous or early Tertiary age.

Local carbonatization of the Bonanza King Formation occurred at some time after this episode of faulting. Later, quartz veining, wallrock alteration, and mineralization occurred, probably at some time between the Paleocene to early Miocene epochs (see Ore Deposits).

Subsequently, the complex history of composite movement along various segments of the Grapevine fault system began. Although some earlier movement occurred, most activity along this system was of middle Miocene age. Displacement along the system caused secondary bending, utilizing additional transverse faults, of the central part of the district and also caused low-angle faulting. The history of movement on the Grapevine fault system included right- and left-lateral movement, possibly in conjunction with activity along the Las Vegas Valley shear zone, and culminated with substantial Basin-and-Range normal faulting. Some minor displacement along the system continued into Quaternary times. North-trending normal faults east of the Main and Northwest segments of the Grapevine fault system probably developed during the Basin-and-Range episode of activity. A potential deposit of megabreccia was prepared for eventual gravity transport at this same time.

Upland erosion and the deposition of a complex piedmont and lacustrine suite of basin fill accompanied and followed Basin-and-Range faulting. The only exposures of this in the Johnnie district are deposits of older fanglomerate which overlie the remnants of a presumed system of pediments, the youngest erosional surfaces present in the district. This late Pliocene to middle Pleistocene pediment system and some adjacent higher topography were dissected and overlain by the older fanglomerate during the late Pliocene to middle Pleistocene epochs. Some gold-bearing veins were exposed and eroded during this dissection and the gold redeposited in placers at the base of the fanglomerate (see Placer Gold Deposits).

The older topography was exhumed by erosion and the placer gold deposits exposed; then the exhumed topography and older fanglomerate were locally unconformably overlain by late Pleistocene to Holocene younger fanglomerate and alluvium. Later sedimentary activity has included some reworking of the older fanglomerate.

ORE DEPOSITS

Introduction

There are two major subdivisions of mesothermal ore deposits, of probable Paleocene to early Miocene age, in the Johnnie district: high-angle quartz veins containing various proportions of gold, chalcopyrite, galena, and pyrite; and, less importantly, concordant quartz-poor lodes of disseminated chalcopyrite with affiliated specularite and minor quartz. Gold-bearing quartz veins weather to produce placer gold deposits; chalcopyritic lodes oxidize to low-grade malachite deposits. Gold-bearing quartz veins and placer gold deposits account for the only significant production in the district.

The veins are epigenetic; the chalcopyritic lodes may be, in part, syngenetic deposits, but, if so, have been modified by epigenetic processes. Vein and chalcopyritic lode deposits and a few deposits with related morphologies are members of a spectrum of hydrothermal quartz-bearing structures in the district subdivided on the basis of configuration, attitude, and quartz content into: high-angle quartz veins; concordant quartz veins; concordant quartz stringer lodes; and concordant quartz-poor lodes. Quartz veins occupy preexistent fractures; the average high-angle vein strikes N. 70° E. and dips north.

The variation of the hypogene ore minerals from one quartz-bearing structure to the next defines a pattern of areal hypogene mineralogic zonation about two gold centers in the district. As the mineralogic proportions vary distally from these centers, zones of predominantly chalcopyrite mineralization are succeeded by galena zones.

Consequently, although intermediate mineralogic combinations are present, the main types of vein deposits distinguished are: gold-chalcopyrite-pyrite-quartz veins in the gold zones; chalcopyrite-galena-quartz veins and chalcopyritic lodes in the chalcopyrite zones; and galena- (calcite-) quartz veins in the galena zones.

The main wall-rock alteration products are widely distributed sericite and pyrite. Locally, specularite, chlorite, and calcite occur as alteration products or as cogenetic gangue minerals in uncommon host rocks whose original mineralogies furnished the reactants necessary to produce these three products. This permits the further subdivision of the hypogene mineralogic zones on the basis of gangue mineralogy. Thus, the chalcopyritic lodes, apparently inherent to the ferruginous parts of the Stirling Quartzite, include specularite; and the major galena-quartz veins, simultaneously falling in the galena zone and the dolomitic upper part of the Johnnie Formation, characteristically contain calcite.

The district may be fringed laterally and at depth by a zone of chlorite-bearing veins; and this and similar districts may have been overlain, before erosion, or now may be overlain by zones of epithermal calcite veins.

The main gold-chalcopyrite-pyrite-quartz veins and most other prominent veins in the district are localized in the Zabriskie Quartzite and underlying dolomitic rocks near the top of the Wood Canyon Formation. This was caused by a combination of mechanical and chemical qualities intrinsic to these rocks and by ponding of hydrothermal fluids beneath an impermeable blanket of sericitized shale gouge at the base of the Carrara Formation above the Zabriskie Quartzite. Significantly, correlative rocks, at the presently recognized base of the Cambrian section, host important ore deposits in other mining districts in the southern Great Basin.

Quartz veins and most other quartz-bearing structures in the Johnnie district are in trains within ENE-trending, north-dipping principal mineralized structures into which were funneled, through a series of other structures, hydrothermal fluids ascending from depth. These principal mineralized structures, four--or possibly five--in number, are aligned within a 15-mi-long (24 km) inferred major longitudinal structure, which trends N. 35° E. and also dips north. The major longitudinal structure, if real, is probably the surface expression of the fundamental control which guided heat and, maybe, metal-bearing hydrothermal fluids into the area to localize the Johnnie district.

The hydrothermal fluids may have originated from lower crust or upper mantle sources or were derived from interstitial connate brines in the miogeosynclinal rocks in or below the district. In either case, the metals in the fluids then could have been leached from the miogeosynclinal rocks transgressed by the fluids.

Wall-rock alteration, at temperatures below 200°C, added hydrogen, potassium, aluminum, iron, and sulfur to the rocks; and CO_2, sodium, magnesium, and calcium were released to the hydrothermal fluid. Quartz veining removed silica from the fluid; and metallization, at temperatures around 250°C, required the addition of gold, copper, lead, iron, and sulfur, from the hydrothermal fluid.

The term "fissure" is used hereforward as a convenience in discussing various aspects of quartz-bearing structures. I do not believe that open voids, as such, existed at any time. Widening of the fissures, by means such as those described in Vein Configuration and Localization of Ore Shoots, was accompanied simultaneously by filling by quartz.

History and Ownership

Activity began in the Johnnie district with the discovery of the Congress mine in 1890. The district received its present name around 1900-10; originally it had been called the Montgomery district.

The early owners and lessees of the Congress mine produced approximately 17,000 troy oz of gold by about 1900. Intermittent operations, largely by lessees, continued there until around the 1960's. Some of the early owners founded the Congress Mining Company, whose assets included the Congress mine, in 1905. The company was passed on to Leo I. Bartlett, the present owner.

The Johnnie mine was founded in 1894 by some of the early lessees of the Congress mine. Subsequently, ownership passed through the Johnnie Consolidated Gold Mining Company, O. T. Johnson, and the Congress Mining Company to Leo I. Bartlett. The mine's most significant production, approximately 45,000 oz of gold, ceased by 1915. The mine has been operated intermittently, mostly by lessees, since then, having been closed temporarily by Order L-208 in 1941. The Johnnie mine ultimately may have produced as much as 55,000 oz of gold.

Approximately 10,000 oz of gold were taken from the Overfield mine by the Crown Point-Globe Mining Company in 1908-9. Minor production has continued to now. Ownership of the mine passed to Charles E. Overfield, the present owner.

Charles H. Labbe acquired the Labbe mine from the Bullfrog-Johnnie Mining Company in 1910 and sold it to George E. Warner, the present owner, in 1964. There are no records of production from this gold property, but apparently production was small.

Placer gold production in the Johnnie district has been low. There was a small boom following the discovery of placer gold there in 1921 and additional production during the Depression.

Minor silver-lead production was taken from the district in the early 1900's.

There has been activity on other, smaller properties also. A more complete account of the history and ownership of the district and the sources of information given are presented in Appendix B.

There are 19 patented claims in the district and approximately 50 active unpatented claims, excluding a large block of Copper Giant claims. The claims comprising the properties discussed above are tabulated in Appendix C.

Production

Production records for the Johnnie district are incomplete, but it is likely that the district produced 80,000 to 90,000 troy oz of gold with a gross value of $1 1/2 to 2 million. This was accompanied by some silver and lead production. This estimate is in excess of the figures presented by previous writers (Nolan, 1924, 1936; Couch and Carpenter, 1943; Kral, 1951; Koschmann and Bergendahl, 1968), being approximately double the 40,000 oz of gold reported by Koschmann and Bergendahl (1968).

A recorded production of $600,399 from 28,721 oz of gold is shown in table 2a, which is adapted from figures given in Nolan (1936) and Couch and Carpenter (1943). Table 2b summarizes additional production, which figures I believe are plausible, given in various private reports. Combining the first two shows that probably $1,600,000 was produced from 80,916 oz of gold contained in 198,755 tons of ore.

Table 2c shows placer production estimated by Vanderburg (1936) along with my estimates of possible additional unreported lode and placer production which are based, in part, upon oral reports I have received. Adding this to the previous figures indicates that the district could have produced as much as 91,266 oz of gold with a gross value of $1,966,000.

The gold totals represent maximum ounces, recalculated from gross yields at the prevailing prices of bullion for the years under consideration.

Nolan (1936) states that the Johnnie district produced a reported 3,645 oz of silver and 6,017 lb (2,729 kg) of lead, in addition to 24,653 oz of gold, between 1908 and 1932. As gold mill recovery was by amalgamation, the galena from which the lead was recovered must have been mined from galena-calcite veins in the Johnnie Formation. This could have been derived from shipments totaling 9-10 tons of selected ore, at an assumed grade of 30-35 percent lead. Most of the reported silver could have been contained in the galena if the silver-lead ratio was 1 oz to 1 percent as reported in the Las Vegas Age (September 11,

Table 2a. Recorded Production in the Johnnie District

Year	Mine	Tons	Gold (oz)	Gross
1896-1897	Johnnie	4,397	3,800	$ 70,895
1908-1913	Johnnie	71,076	19,721	402,656
1914-1932	Unknown	17,769	4,930	100,664
1934-1935 and 1940	Unknown	1,513	465	16,274
Total		94,755	28,916	590,049

Table 2b. Reported Production Which is Not Recorded

Year	Mine	Tons	Gold (oz)	Gross
1890-1898	Congress	7,100	12,000	$ 250,000
1899	Congress	1,900	5,000	100,000
1908-1909	Overfield	Estimated 35,000	10,000	200,000
1910-1913	Johnnie	60,000	25,000	500,000
Total		104,000	52,000	1,050,000
Total Tables a & b		198,755	80,916	1,640,049

Table 2c. Estimated Additional Production Which is Not Reported

Year	Mine	Gold (oz)	Gross
1921-1935	Placer	Less than 1,000	Less than $ 20,000
1936-1974	Placer	100	3,500
1915-1970	Johnnie	7,150	250,000
1910-1974	Overfield	100	2,500
1891-1974	Rest of District	2,000	50,000
Total		10,350	326,000
Total Tables a, b, & c		91,266	$1,966,049

1926) for one of these properties. However, at a ratio of 1/4 oz silver to 1 percent of lead as concluded herein (see Galena-Calcite-Quartz Veins) there would have been excess silver which must have been alloyed with the gold recovered by amalgamation.

Mineralogy

Introduction

For classification purposes, there are three groups of hypogene minerals in the Johnnie district:

(1) Ore minerals: gold, chalcopyrite, galena, pyrite

(2) Gangue mineral: quartz

(3) Wall-rock alteration products: sericite, pyrite, specularite, chlorite, and calcite. Occasionally these occur within the quartz veins as gangue derived from the alteration of inclusions of wall rock or from the remobilization of wall-rock alteration products.

Supergene minerals include malachite, azurite, covellite, digenite, chalcocite, anglesite, cerrusite, limonite, quartz, calcite, manganese dioxide, and jasper (a mixture). Local concentrations of malachite have been prospected as a source of copper ore. Weathering releases free gold into placer deposits.

Most mineralized material exposed at the surface and in mine workings is leached, to some extent, by supergene processes. Therefore, the mineralogic descriptions which follow and the descriptions of hypogene ores in forthcoming sections largely are reconstructions based upon piecemeal information obtained from examples of each feature in the field and laboratory. Numerous specimens of gold from twelve locations within the district were examined.

Hypogene ore textures will be discussed in further detail in the individual descriptions of the types of ore deposits.

Hypogene Ore Minerals

Gold: Gold occurs in flattened to equidimensional grains from 0.01-1.0 mm in diameter and in single, convolute veinlets or branching and rejoining veinlets. Veinlets are tabular aggregates of gold up to 1 mm thick and 13 mm long, which are studded with irregular grains and spicules of gold. Labbe (1960) reports having seen aggregates of gold from one ore shoot from which "twenty dollar (gold) pieces" could have been cut. Grains of gold occur within and between the grains of the hypogene sulfide minerals present. Gold veinlets fill fractures which penetrate vein quartz and/or penetrate masses of hypogene sulfide minerals.

The surface of the gold is covered with closely, but irregularly spaced raised areas approximately 0.01 mm in diameter, which are more reflectant than the intervening depressions to give the surface of the gold a spangled appearance under low power (40X) magnification. These spangles may represent individual crystals.

Polished sections show that the gold is irregularly zoned. Equant patches of pinkish gold up to 0.15 mm in diameter, which are probably cupriferous, are discontinuously distributed around approximately one-third of the perimeters of most of the grains or aggregates observed. This is probably a hypogene growth zonation; observations of similar zonation of gold are summarized by Uytenbogaardt and Burke (1971).

Chalcopyrite: Individual grains, or tabular to equidimensional, walnut-sized aggregates of grains, of chalcopyrite are aligned along fractures in vein quartz. The individual grains have anhedral, equant forms and, on the average, are 0.5 mm in diameter.

Galena: Galena makes irregular, semi-continuous fracture fillings in quartz up to 5 cm in thickness and also occurs as single grains or small aggregates of grains adjacent to aggregates of chalcopyrite or nearby along the same host fractures. The individual grains of galena are anhedral and make irregular polygons up to 5 mm in width. They are strongly deformed; in polished section, cleavages are bent or, in extreme cases, kink banded.

Pyrite: Discontinuous groups of pyritohedrons up to 0.5 mm in diameter fill fractures, which are one crystal thick, in vein quartz or along the edges of aggregates of chalcopyrite. This apparently is new pyrite, introduced from without during metallization, and is not the remobilized and recrystallized derivative of the cubic, wall-rock alteration variety of pyrite.

Hypogene Gangue Minerals and Wall-Rock Alteration Products

Quartz: Quartz is the most abundant epigenetic mineral in the Johnnie district. It fills fissures in faults and fractures to give massive veins of milky quartz up to 5 ft (1.5 m) thick. To a lesser extent, quartz also cements or replaces breccia fragments or permeates gouge within those zones or permeates fractured country rock (see Nature of Quartz Filling).

Fissure filling vein quartz is composed of sutured, anhedral, elongate or equant grains up to 5 mm, respectively, in length or in diameter. Some grains contain nearly optically continuous internal structures, approximtely 10 times smaller, which are columnar and transverse to the elongate grains or which, in the case of equant grains, are equant. The elongate grains are comb structures, and the transverse internal structures probably are growth zones. The equant grains apparently are composites of several smaller grains which grew simultaneously from scattered nuclei.

The crystal faces of quartz are present locally and vugs are rare in the quartz veins. Where present, the vugs are partially lined with coarse, singly terminated quartz crystals which are continuous into the main mass of the quartz.

Sericite: Sericite is herein taken as the $2M_1$ variety of muscovite defined by Grim and others (1937). Two macroscopically different types are found in hydrothermally altered rocks in the Johnnie district. The first type occurs, with pyrite, in greenish rocks derived from shaly clastic rocks. It forms aggregates composed of flakes which range in size from less than 0.05 mm to, rarely, 25 mm. The other type is derived from dolomitic shales and shaly dolomites. Macroscopically, it is a compact, fine-grained, white material with a powdery surface; it coarsens locally to become very pale green with a slightly silken luster. Microscopically, it is a birefringent aggregate of intergrown bundles of flakes.

Sericite derived from dolomitic shale forms massive aggregates, where pure; but, ordinarily, sericite constitutes the matrix between detrital mineral grains or fills fractures through those grains or through the whole rock. Sericite replaces sedimentary feldspar, partially to completely, and slightly replaces detrital quartz. The latter is effected along growth zones and other internal crystal structures and along the sutures between detrital quartz grains and diagenetic secondary quartz overgrowths. Locally, sericite occurs in fine fractures through vein quartz, and very fine-grained silica locally accompanies sericite as a wall-rock alteration product.

Pyrite: Striated cubes of pyrite up to 15 mm across occur in the green sericitized rock derived from the hydrothermal alteration of shaly clastic rocks. Cubic pyrite also occurs as groups of single crystals and crystal aggregates dispersed throughout narrow tabular areas, which average 1 ft (0.3 m) on their long axes, within and subparallel to quartz veins. The latter are presumed to be relicts of completely altered wall rock; because a series of observations shows a succession from fresh wall rock inclusions concordant within veins through sericitized and pyritized inclusions replaced by increasing amounts of quartz until only cubic pyrite remains--unassociated with gold, chalcopyrite, galena, or pyritohedral pyrite.

Specularite: Specular hematite develops with a moderate amount of sericite as a hydrothermal alteration product of dark colored quartzites and shales. The specularite occurs within, and in quartz veins through, the dark host rocks. The coloration of the host rocks derives from small amounts of amorphous iron oxide in the matrices, and the specularite is taken to be a product of the local remobilization and recrystallization of this iron oxide.

Sheaf like, divergent, walnut-sized aggregates of euhedral plates of specularite are present in quartz veins. The individual plates are up to 20 mm in diameter and 0.05-0.3 mm thick; they thin toward the apexes of the bundles. The specularite in the clastic rocks occurs as sheets up to 1 mm in diameter in fractures through quartzite and in fractures through quartz grains therein to form masses of differently oriented groups of subparallel or divergent aggregates. Where less abundant, specularite makes reticulate structures in fractures or occurs as small individual crystals between and within quartz grains.

The specularite is green black; this coloration probably is, in part, a weathering effect. Locally, specularite merges into compact hematite.

Chlorite: Compact aggregates of very fine-grained chlorite occur sparsely in the Johnnie district within, and along the margins of, quartz veins. This chlorite is the hydrothermal alteration product of dolomitic clastic rocks and is the replacement product of specularite during continued hydrothermal alteration also.

Calcite: Calcite is the hydrothermal alteration product of dolomite country rock; it occurs in quartz veins which pass through or near horizons of dolomite or strongly dolomitic rocks. The calcite occurs in the veins as individual rhombs or as interconnected masses of coarsely cleavable aggregates up to 20 mm in diameter. These aggregates are composed of calcite grains 0.25-0.5 mm in diameter. Calcite also occurs along vein walls, where it becomes coarse and changes composition veinward from dolomite to calcite.

This calcite is light to dark brown in color due to small masses of amorphous iron oxide dispersed along its cleavage planes. The iron oxide constitutes as little as approximately 1 percent of the total weight of the calcite; and the higher the iron oxide content, the darker the color becomes. The iron could be exolved from hypogene ferroan calcite, or the iron oxide is an exotic hypogene or supergene deposit.

Hypogene Paragenesis

Three hypogene paragenetic events are evident in the Johnnie district: wall-rock alteration, quartz deposition, and metallization (fig. 15). There is some overlap between the first two events and within the third. Since alteration and quartz deposition are spatially separate they can be considered separately. In the absence of evidence to the contrary, this assumes that there was only one significant episode of hydrothermal activity in the district. The paragenetic sequence is evident from the textural relations discussed in Mineralogy and in Types of Ore Deposits.

Wall-rock alteration entailed the deposition of sericite with or without pyrite, the deposition of specularite (the hematite locally having been replaced by later chlorite), or the deposition of chlorite. Subsequently sericite replaced chlorite. Simultaneously with sericite deposition, calcite formed in dolomitic wall rocks.

Quartz veining began sometime during (locally simultaneously with) wall-rock alteration and persisted until approximately the end of alteration. For discussion purposes, quartz veining can be considered to have been effected in a single pulse.

Metallization involved the deposition of copper, lead, and iron

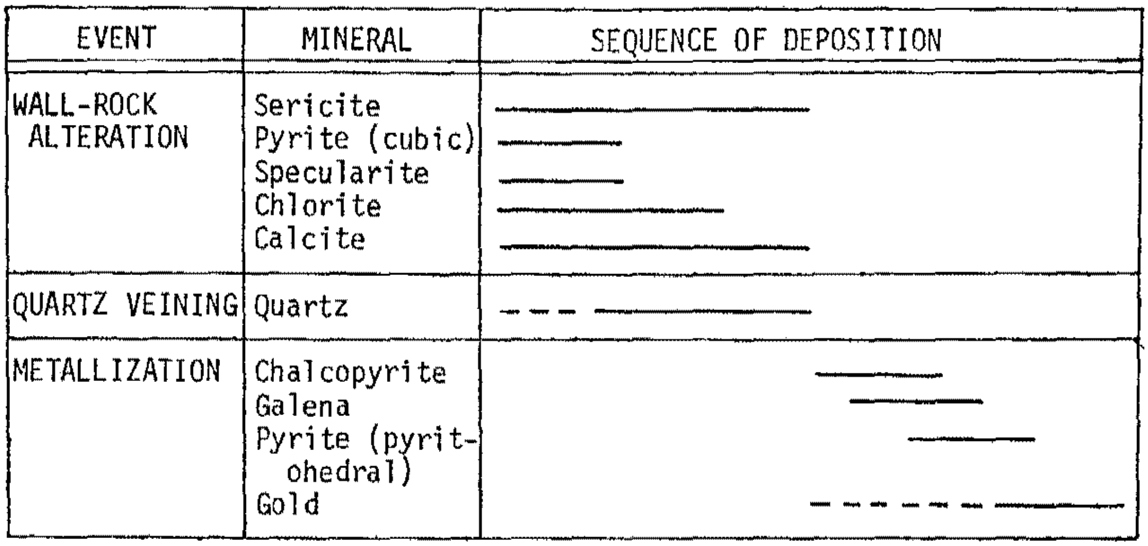

Figure 15. Paragenesis of hypogene ore deposits in the Johnnie district.

sulfides and the deposition of gold. A sequential order of deposition is apparent, although there are local reversals of the scheme. The general paragenetic sequence of the ore minerals is chalcopyrite, galena, pyrite, and gold. Some amounts of gold were introduced with the chalcopyrite and galena, but most of it was introduced with the pyrite or afterward.

Supergene Minerals and Paragenesis

Supergene processes began modifying the rocks and ore deposits of the Johnnie district during probable post-middle Miocene erosional exposure of the district. Locally, some of the processes are variously completed, arrested, or continuing. All aspects of supergene mineralogy, except for malachite (see Malachite Deposits), are discussed in this section. Many of these observations were used to infer hypogene ore, gangue, and wall-rock-alteration-product textures.

During oxidation, covellite, digenite, and chalcocite replace chalcopyrite. Limonite (the name here assigned to an amorphous, compact, brown secondary iron-oxide compound which probably approximates the composition of ferric oxide hydrate) commonly replaces the above supergene minerals and directly replaces pyrite and, to some extent, galena, as well as replacing specularite and calcite. Locally, galena is replaced by anglesite, cerussite, or covellite. The admixture of fine-grained supergene silica with limonite produces jasper. There is no evidence in the Johnnie district for supergene modification of gold beyond local limonite staining. Besides limonite, other transported supergene minerals are malachite, azurite, quartz, calcite, and manganese dioxide.

Supergene replacement proceeds inward through mineral grains within thin zones along the edges of grains and edges of throughgoing fractures. Replacement is guided by cleavage and other internal crystallographic planes, where present (such as in chalcopyrite and galena).

The edges of chalcopyrite grains are replaced by covellite along intersecting, internal regular lamellar structures. Then, the residual chalcopyrite which is bounded by replaced lamellae and also some of the covellite, itself, is replaced by later digenite. Chalcocite replaces both covellite and digenite and, in turn, is replaced by limonite which grades outward into copper pitch, a variety of jasper. Thus, weathered chalcopyrite grains are encircled by four concentric replacement zones --covellite-digenite, digenite, chalcocite, and limonite--each approximately 0.05 mm thick. This sequence of chalcopyrite replacement is characterized by an outward increase in the copper-sulfur ratio which may result from a depletion of sulfuric acid available from a dwindling supply of oxidizing pyrite.

In the presence of chalcopyrite, galena is replaced by a minor amount of covellite which is similarly succeeded outward by digenite, chalcocite, and, finally, limonite. Locally, galena develops a rind of anglesite. In a few cases in which galena is completely leached a few macroscopically fine crystals of cerussite remain in the mold.

As these supergene processes go to completion, the hypogene sulfide minerals, and specularite, and calcite are completely replaced by limonite. Fine-grained supergene silica invades the limonite to produce a reddish brown jasper. Microscopically, jasper consists either of a monotonous admixture of limonite in a fine siliceous webwork or of a banded sequence, sometimes exhibiting a colloform structure, of layers of differing limonite-silica proportions. Near-surface weathering leaches the limonite from the jasper, enriching it in silica and causing the jasper structure to collapse, leaving a barren cavity. Supergene boxworks of sulfide and other hypogene minerals or mixtures thereof develop locally and incidentally to the formation of jasper. The final oxidation product of chalcopyrite sometimes is a black, glassy, silica-rich jasper containing minor malachite and relatively rich in manganese and iron.

Indigenous limonite formed after pyrite does not alter to jasper. Malachite, produced from copper released by the limonite replacement of the secondary copper sulfide minerals, is deposited along fractures or enters jasper. Locally, minor amounts of azurite accompany malachite.

The various supergene replacement products, where below the surface, apparently contain some water which makes them soft. However, with the exception of malachite and azurite, they dehydrate and harden near the surface. Supergene quartz is common near the surface, cementing and rendering quite durable--to within 50 ft (15 m) of the surface --brecciated materials which are otherwise friable. Some supergene calcite is present in the same environment as compact fracture fillings or as crystalline fracture coatings.

Thin coatings of transported limonite are common throughout the entire supergene zone. It is locally mixed with the quartz named above, causing the quartz to become darkened. Approximately 2-mm-thick coatings of manganese dioxide are present on rocks in a few localities in the district.

Morphology of Quartz-Bearing Structures

Types

Hydrothermal activity in the Johnnie district produced four types of quartz-bearing structures as well as varieties gradational between different types of structures. Due to structural circumstances, particular quartz-bearing structures are characteristic in different areas of the district; their general distributions are given in figure 16. Although they are cogenetic, the types of quartz-bearing structures differ in orientation, configuration, and conditions of dilatancy during hydrothermal invasion:

(1) High-angle quartz veins. These are the fissure fillings in discordant high-angle fractures and faults. These widely distributed features are the most common, characteristic quartz-bearing structures in the district and are the hosts for all of the major ore deposits.

(2) Concordant quartz veins. These are veins localized along bedding planes, high-angle reverse faults, and shears in the axial areas of folds. They are most common in the Johnnie Formation, particularly in fissile rocks, and are generally small and unproductive.

(3) Concordant quartz stringer lodes. Stringer lodes are concordant bodies, up to 50 ft (15 m) thick, containing numerous veinlets of quartz localized along bedding planes, cross bedding planes, and discordant fractures. They are most commonly located in the B member of the Stirling Quartzite and in the Johnnie Formation. Stringer lodes are gradational locally into concordant

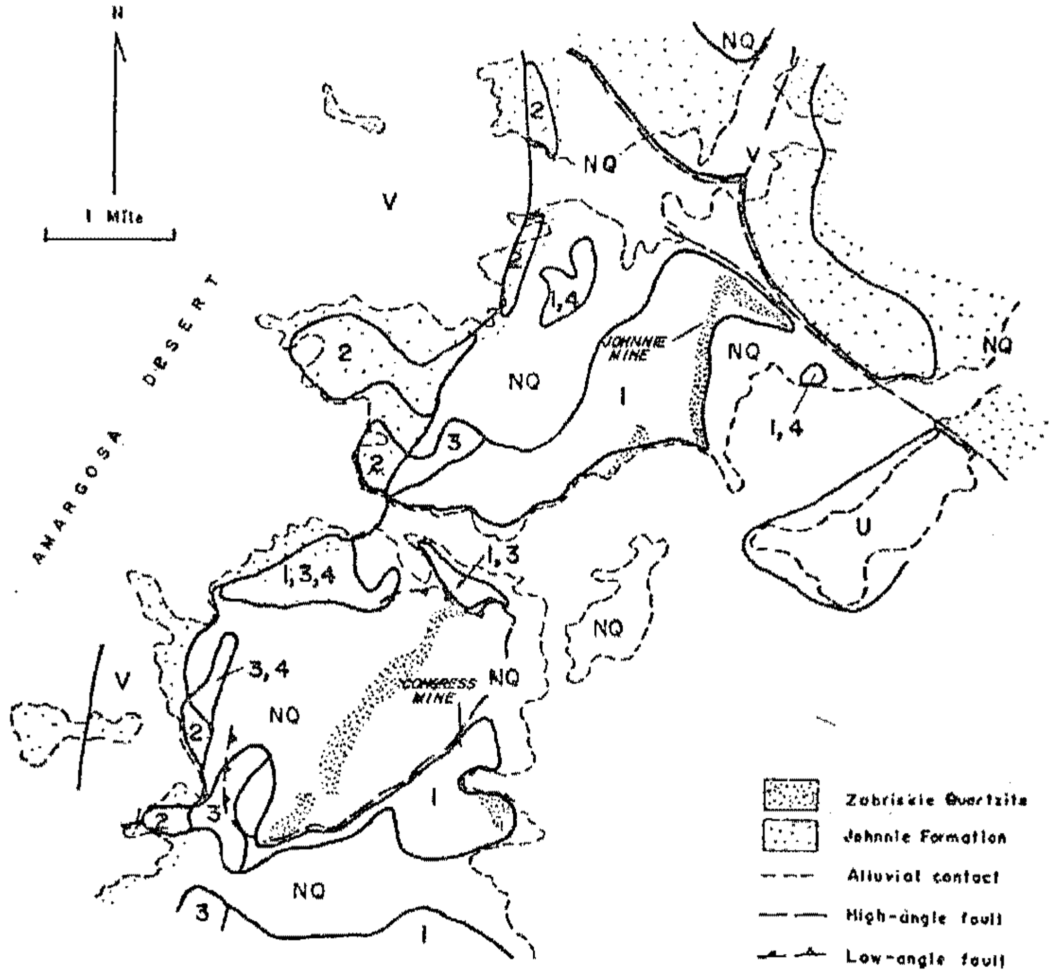

Figure 16. Map showing areas of hydrothermal activity and distribution of types of quartz-bearing structures in the Johnnie district.

1-4, outcrop areas of quartz-bearing structures: (1) high-angle quartz veins; (2) concordant quartz veins; (3) concordant quartz stringer lodes; (4) concordant quartz-poor lodes.

V, area of small outcrops containing numerous quartz veinlets.

U, allochtonous megabreccia deposit; type of quartz-bearing structure not classified.

NQ, areas essentially devoid of quartz-bearing structures.

quartz veins and they usually comprise the distal portions of high-angle quartz veins.

(4) Concordant quartz-poor lodes. These are weakly altered areas, within porous quartzite units, with some mineralization and, locally, some quartz. They are most common within the upper part of the B member of the Stirling Quartzite. These lodes are usually near or cut by thin high-angle or concordant quartz veins which were probably the hydrothermal feeders for the lodes.

Nature of Quartz Filling

The vein quartz which fills simple fissures, where not subjected to post ore fracturing (which is nearly ubiquitous in the district), is simply a massive aggregate of that mineral, as described in Mineralogy; but other circumstances modify the appearance of the quartz and of the entire mass of varied materials which constitute the veins.

A material here called quartz-impregnated breccia and gouge is present where quartz veins are in parts of gouge zones in predominantly clastic rocks. Quartz-impregnated breccia and gouge is a finely comminuted (fragments up to 5 mm) breccia of competent quartzite and dolomitic quartzite in a matrix of gouge derived from incompetent dolomite and shale. Grains of vein quartz up to 10 mm in diameter and larger pods of vein quartz are distributed throughout this. Although there is some rehealing of breccia and gouge fragments along points of contact, this material usually is quite friable.

Quartzite, itself, is replaced to become vein quartz by a process in which hydrothermal fluids invade the quartzite along microscopic fractures and intergranular boundaries, dissolving the shaly matrix minerals with or without the addition of quartz. Where quartz is not added to replace the matrix, the residual detrital quartz grains anneal along points of contact. These materials become part of the vein but retain replacement shadows--relict texture and coloration--of the original rock.

The product of replacement of dolomitic rocks by quartz adjacent to fissure fillings of vein quartz becomes part of the vein. This is very common in the Congress vein, particularly in the hanging wall where thicknesses up to 5 ft (1.5 m) are added (pl. 10); and it occurs locally in the veins at the Johnnie and Overfield mines.

Microscopically, replacement of dolomite occurs by invasion of vein quartz along fractures and between breccia fragments and involves embayment by the dissolution of dolomite along highly irregular boundaries. With advanced replacement dolomite appears to migrate toward the replacement front, evident because residual detrital quartz veins in quartzitic dolomite become more numerous toward the front. With continued replacement: these residual quartz grains anneal at points of contact in partially replaced rock; or the grains are engulfed by vein quartz and incorporated into its structure in completely replaced dolomitic rock.

The resultant vein quartz contains numerous inclusions of apparent iron oxide less than 2.5 microns in diameter and also contains larger inclusions of unreplaced dolomite. These inclusions impart the characteristic red color to the ore from the Congress mine.

Types of Ore Deposits

Three types of ore deposits are present in the Johnnie district. (Ore deposits are here taken as concentrations of minerals containing metals which are usually marketed, whether these metals are present in economic quantities or not.):

(1) Metal-bearing quartz veins. This category is divisible into three subtypes based on mineral assemblages:

(a) Gold-chalcopyrite-pyrite-quartz

(b) Chalcopyrite-galena-quartz

(c) Galena-calcite-quartz

The differences among the above three derive from changing relative proportions of minerals. Mineralogic types gradational between these characteristic types are present also.

(2) Malachite deposits: The hypogene precursors of these secondary copper deposits are variants of the metal-bearing quartz veins. The precursors are closely allied, spatially and perhaps genetically, with specularite-quartz veins, which are not considered as a separate type here because they are uneconomic in themselves.

(3) Placer gold deposits.

The following sections describe the individual types of ore deposits. Figure 17 shows the distribution of the ore minerals in the district, except placer gold. Placer gold deposits are shown in plates 2 and 3.

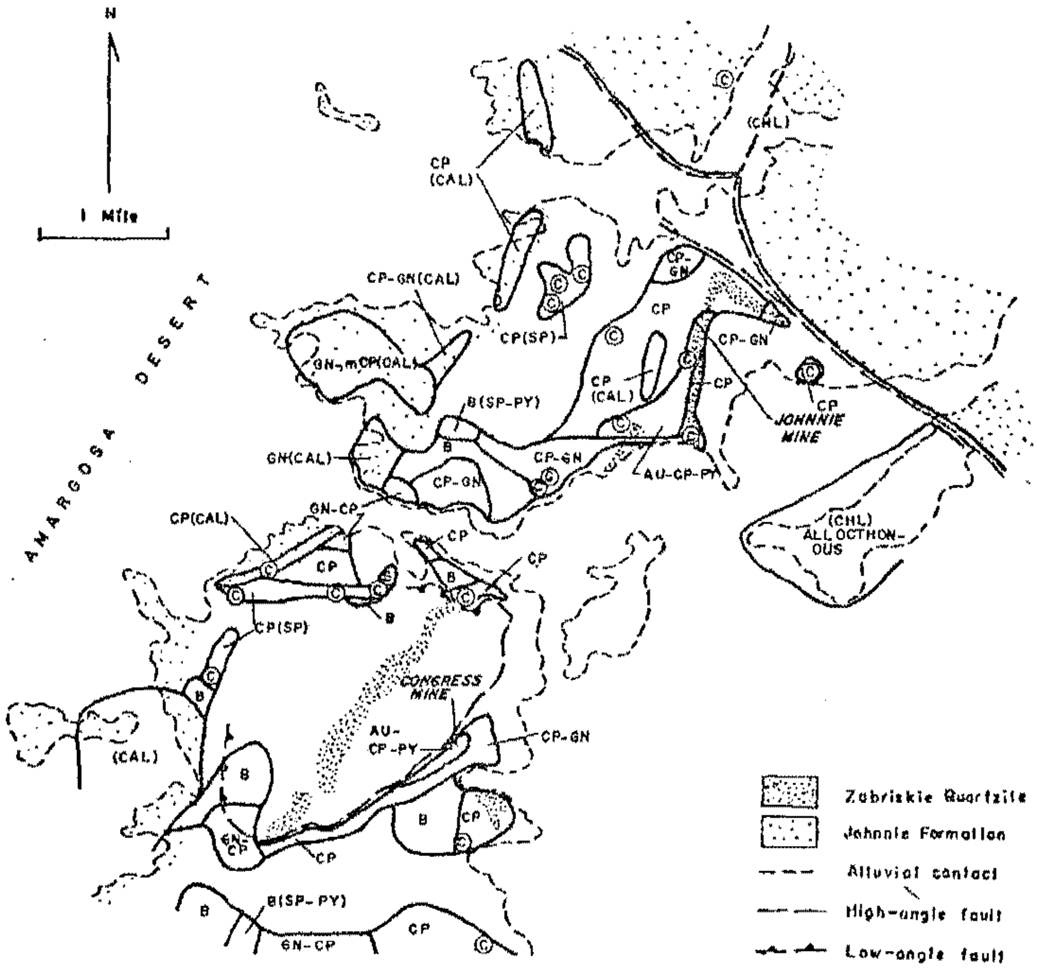

Figure 17. Map showing distribution of hydrothermal minerals in the Johnnie district, literally portraying areas in which ore minerals occur and locally showing general regions in which some gangue minerals occur (chlorite east of Grapevine fault; calcite in southwest part of district). Sub-zones with characteristic gangue minerals shown in parentheses.

Au-Cp-Py, gold-chalcopyrite-pyrite; Cp, chalcopyrite; Cp-Gn, chalcopyrite with subordinate galena; Gn-Cp, galena with subordinate chalcopyrite; Gn-mCp, galena with minor chalcopyrite; Gn, galena; Sp, specularite; Py, pyrite; Cal, calcite; Chl, chlorite; B, area of barren quartz veins.

(©), malachite deposit or group of closely spaced malachite deposits.

Based upon data from approximately 210 locations.

Gold-Chalcopyrite-Pyrite-Quartz Veins

Distribution: Gold-chalcopyrite-pyrite-quartz veins in the Johnnie district are localized along high-angle structures in the North and Congress mining areas. Known production in the North mining area has come from a triangular area the corners of which are at the vicinities of the Johnnie mine, Westend open cut, and the workings on the east end of the Doris A. L. claim (pl. 2). Economic production in the Congress mining area was probably restricted to the Congress mine.

During this study, I confirmed, by observation and assay, the presence of gold in the Johnnie and Crown Point mines (both veins, pls. 4 and 6); in the Boadway and Doris mines, the workings on the east end of the Doris A. L. claim, and the Westend open cut (pl. 2); and in the Congress mine and prospects 3,000 ft (900 m) southwest of the Congress mine at the end of the system of veins which includes the Congress vein (pl. 3).

Description: Grains and veinlets of gold, in some places quite densely distributed, occur in leached pockets, which contain varying amounts of sulfide minerals, in vein quartz which in most places has been shattered by postore movement along the host structures. Within 50 ft (15 m) of the surface (see Supergene Minerals and Paragenesis), the shattered vein quartz usually is cemented by supergene quartz which contains enough admixed iron oxide locally to impart a brownish or blackish color to the ore. Nolan (1924) and private reports indicate that the fineness of the gold ranged between 800 and 950.

Although most of the gold-chalcopyrite-pyrite-quartz veins are completely leached, reconstruction of a picture of the hypogene ore is possible from piecemeal observations. Varying amounts of native gold occurred in lenses within the fissure-filling portions of quartz veins along with chalcopyrite and pyrite, in approximately equal amounts, and with very minor quantities of galena. These lenses were equant tabular aggregates of grains 1 to 3 in. (2.5-7.5 cm) in diameter and about one-tenth as thick. The edges of these aggregates were fringed with individual grains or small aggregates of minerals. The lenses were zoned, with a sheet of chalcopyrite being sandwiched between two sheets of pyrite. Galena was dispersed at and beyond the distal portions of the pockets. Gold occurred as small grains within and between the sulfide mineral grains and in veinlets near or through the pockets and subparallel to them.

Groups of lenses were distributed along one or more sheetlike fractures subparallel to the veins or along the walls of veins. These were fractures which were opened by minor post-quartz-veining movement along the host structure, permitting the ingress of metal-rich hydrothermal fluids. These groups of lenses comprised tabular pockets of ore subparallel to the veins which usually were up to 3 ft (1 m) in diameter by 1 ft (0.3 m) wide, but locally were greater. The metallic minerals probably comprised up to 10 percent of the volumes of the pockets. Apparently there was a general increase in the amount of galena present toward and beyond the ends of the pockets as it is reported that the miners used this as a guide to prospecting.

The mineralized pockets were disposed in ore shoots, which are local swells of quartz which, depending on individual structural circumstances, rake along or down different veins (see <u>Vein Configuration and Localization of Ore Shoots</u>). The frequency with which pockets were distributed in the ore shoots is unknown and apparently their distribution was erratic.

<u>Tenor of Ore</u>: The grade of gold present in the mineralized pockets varied over a wide range. The highest grade ore was rich enough to constitute specimen grade or "jewelry rock"; but many gold-free pockets were present also. Charles H. Labbe (private report) observed a 750 lb (340 kg) batch of selected ore from the Johnnie mining area which contained 218 oz (580 oz/T) gold. The quartz vein between pockets and ore shoots was probably barren in most cases.

Private reports indicate that when the Johnnie mine was being systematically operated entire ore shoots were mined at a grade of approximately 0.5 oz/T or less.

Two composite samples of strongly leached ore from the Johnnie mine, selected during this study, contain the metal values given in table 3. Also given are the values for composite samples, which are similarly leached but contain some galena, collected from the Buldosa mine and from a prospect near the Doris mine (pl. 2). The latter two are near the fringe of what is here considered a gold-producing center of hypogene mineralogic zonation (fig. 19) which encompasses the Johnnie, Overfield, Broadway, and Doris mines.

Inspection of the ratios of the metals contained in the samples suggests that the silver is distributed between gold and galena and that the zinc is associated with galena. No silver or zinc minerals

Table 3. Analyses of ore samples from gold-producing zone in North mining area, showing estimated original volume percentages of metallic minerals contained therein*and showing informative metal ratios.

Sample	Johnnie mine** With visible gold	Johnnie mine** Without visible gold	Buldosa mine	Near Doris mine
Weight (lb)	0.62	0.69	3.97	1.76
Gold (oz/T)	1.13	0.95	0.25	0.15
Silver (oz/T)	0.89	0.60	0.53	0.57
Copper (%)	0.35	0.5	0.49	0.47
Lead (%)	0.4	0.36	7.02	3.34
Zinc (%)	0.02	0.02	0.14	0.80
Ag/(10Au+Pb)	0.08	0.06	0.06	0.12
Zn/Pb	0.05	0.06	0.02	0.24
Gold (volume %)	<0.001	<0.001	<0.001	<0.001
Chalcopyrite (vol. %)	15	10	2	1
Pyrite (volume %)	15	10	?	1
Galena (volume %)	<1	<1	2	1

*Estimated from examination of limonite etc. in leached specimens

**From pocket at surface near Johnnie shaft were detected in polished specimens, and presumably the silver and zinc are contained in solid solution in their host minerals.

<u>Wall-Rock Alteration</u>: Greenish sericite-pyrite hydrothermally altered selvages, usually up to 3 in. (7.5 cm) thick, are adjacent to gold-chalcopyrite-pyrite-quartz veins in clastic rocks. Selvages of the whitish variety of sericitized rock are present in adjacent hydrothermally altered dolomitic rocks; these are quite prominent as selvages up to 5 ft (1.5 m) thick locally adjacent to the Congress vein (pl. 10) and as thick bodies in fractured dolomitic shale, some of which contain small quartz veins, at the base of the Carrara Formation north and south of the Johnnie mine (for example, see pl. 4).

The layers of dolomitic rock in the dolomite-bearing upper part of the upper member of the Wood Canyon Formation (units 2 and 3, Appendix A) are silicified along elongate areas up to 25 ft (8 m) wide which are located in the vicinity of quartz veins in the North and Congress Mining areas. Silicified rock is shot through with quartz veinlets as little as 0.5 mm thick along which the adjacent dolomite is replaced to a minor extent by quartz in the manner described in <u>Nature of Quartz Filling</u>. Near the Congress mine, some of these areas of silicified dolomite are gradational into the zones of dolomite breccia replaced by quartz described in <u>Nature of Quartz Filling</u>. Dolomite breccia replaced by quartz differs from silicified dolomite in having more of the fractures (now filled with quartz) along which replacement occurred and in having had much more extensive replacement.

Iron oxide alteration of dolomite is present in places at the Congress mine and at one place near the Johnnie mine. This involves the concentration of iron oxide along dolomite grain boundaries and cleavages to as much as 10 percent of the volume of the rock. The resultant rock is pinkish in color.

The disposition of iron oxide alteration and, to some extent, silification appears to be, in part, stratigraphically controlled. The areal distribution of silicified areas and areas of iron oxide alteration are shown in plates 4 and 7.

Chalcopyrite-Galena-Quartz Veins

Chalcopyrite-galena-quartz veins occur in zones peripheral to areas in which gold-chalcopyrite-pyrite-quartz veins are concentrated. Chalcopyrite-galena-quartz veins are found throughout the entire stratigraphic section up to the base of the Carrara Formation but are most common in the Stirling Quartzite and Wood Canyon Formation. The veins usually are high angle.

Except for the absence of gold, the chalcopyrite-galena-quartz veins are mineralogically similar to the gold-chalcopyrite-pyrite-quartz veins; but galena, which is sparse in the gold-bearing veins, is relatively abundant in the chalcopyrite-galena-quartz veins and becomes the dominant mineral at progressively greater distances away from the gold-chalcopyrite-pyrite-quartz veins (see Hypogene Mineralogic Zonation). As with the gold-bearing veins, the sulfide minerals comprise less than 1 percent of the total volumes of the chalcopyrite-galena-quartz veins. Some pyrite, most of which probably originated as a wall-rock alteration product (see Mineralogy), occurs within the veins.

Calcite is present locally where the veins are hosted by dolomitic rocks. This makes calcite a characteristic gangue mineral in chalcopyrite-quartz and chalcopyrite-galena-quartz veins in the upper part of the Rainstorm Member of the Johnnie Formation in the areas north and south of the mouth of Johnnie Wash. Only rare quantities of calcite are present in chalcopyrite-galena-quartz veins in the dolomitic horizons in the Stirling Quartzite and Wood Canyon Formation.

The local controls on the ore minerals in the chalcopyrite-galena-quartz veins are the same as those for the gold-chalcopyrite-pyrite-quartz veins.

Similarly, the wall rocks adjacent to chalcopyrite-galena-quartz veins are sericitized and pyritized. This is usually manifested as bands of greenish rock approximately 10 mm thick where the wall rocks are of sufficiently high shale content for the alteration color to become evident.

Galena-Calcite-Quartz Veins

These veins of galena and calcite or of galena or of calcite, all in a quartz gangue, are modified varieties of the veins discussed so far. They are located throughout an area in the Johnnie Formation north of the jog in Route 16 in secs. 22, 23, 24, and 26, T. 17 S., R. 52 E. and a few are in the Johnnie Formation east of the Main and Northwest segments of the Grapevine fault system. A number of calcite-bearing veins are present in the area near the southwest corner of the area mapped on plate 1.

Although galena and calcite comprise less than 1 percent of the volumes of the veins, either or both, combined, are relatively more abundant than are the gold, chalcopyrite, and galena of the other veins. Chalcopyrite is usually inconspicuous where present at all. These veins contain negligible quantities of gold. The highest of four assays of selected ore grade material contained 0.015 oz/T of gold. Approximately 0.25 oz/T of silver is present to 1 percent of lead in this ore. This is similar to the 0.3 oz/T per 1 percent reported by Nolan (1924) for ore from the Coe property (see below). Apparently the silver is in argentiferous galena.

The galena is concentrated throughout areas subparallel to the veins, the locations of which areas are governed by premetallization fracturing of the quartz; but the calcite is less regularly distributed, because it crystallized before or during quartz veining.

Usually thin zones of sericitized and pyritized shaly rock are adjacent to the veins. Whitish sericitized rock is present in dolomitic shales, making selvages up to 2 ft (0.6 m) thick adjacent to some veins. These are some of the thicker selvages of sericitized rock in the district.

Although galena-calcite deposits are commonly in concordant quartz veins, the most strongly mineralized deposit of this type is a concordant quartz stringer lode localized in a drag fold adjacent to a transverse fault zone. This deposit is situated in the center of sec. 26 (pl. 1) and apparently is on the Albert Coe property to which Nolan (1924) refers.

The differences between these and the other metal-bearing vein deposits discussed so far are readily explainable. The high galena-to-chalcopyrite ratio of the galena-calcite-quartz veins results from their remote distances from centers of gold mineralization (see <u>Hypogene Mineralogic Zonation</u>). The circumstantial abundance of dolomitic country rocks causes calcite to be a common wall-rock alteration product here. Galena-calcite-quartz veins are localized in concordant structures, rather than in high-angle structures because the mechanical properties of the Johnnie Formation favored dilatancy in concordant structures.

Malachite Deposits

Description: Malachite coats surfaces of, and to some extent permeates, clastic rocks in the B member of the Stirling Quartzite and locally in other rocks, particularly in the Wood Canyon Formation, throughout the Johnnie district (pls. 2 and 3, fig. 17). The malachite is the supergene oxidation product of chalcopyrite disseminated in concordant quartz-poor lodes or in other quartz-bearing structures ancillary thereto. The lodes are on the order of 100 ft (30 m) thick and up to 300 ft (90 m) long in the outcrop and probably do not approach grades above 0.1 percent copper, even locally.

The prime characteristic of the malachite deposits is paucity of quartz veins--the parent copper sulfide mineral was truly disseminated throughout the nearby rock essentially without accompanying introduction of quartz; and the most important ones, by far, are those in the Stirling Quartzite. There are variations where: (1) chalcopyritic lodes are cut by or flanked by chalcopyrite-quartz veins; (2) minor chalcopyrite is disseminated in the walls of chalcopyrite-quartz veins; (3) malachite deposits occur in other stratigraphic intervals; and (4) the malachite in the deposits was transported moderate distances from its hypogene chalcopyrite source.

In the B member of the Stirling Quartzite, where this type of deposit is best developed, the chalcopyritic lodes occur in horizons of moderately-fractured, coarse- to medium-grained, light-colored quartzites. Grains and veinlets of chalcopyrite up to 3 mm thick occupy minor fractures between and through detrital quartz grains. (The moderate fracturing mentioned above is later.) Sericite and, locally, very fine-grained silica occur with some of the chalcopyrite. The sericite moderately to strongly replaces the edges of detrital grains, and also it occurs within the interiors of some grains. These quartzite horizons contain no to moderate amounts of quartz veinlets, and silty partings in the quartzite are altered to coarse sericite.

A few chalcopyrite-quartz veins occur in or nearby along strike with some of the chalcopyritic lodes in the B member of the Stirling Quartzite. Some specularite-quartz veins do the same and also occur elsewhere in the member. Neither type is rarely over 6 in. (15 cm) thick. Some of the specularite veins are quartz poor, in which case the specularite occurs in the hydrothermally altered walls of fractures in quartzite. The specularite-quartz veins are usually high-angle, but some of them and all of the chalcopyrite-quartz veins in this setting are localized in concordant quartz stringer lodes.

The malachite is deposited as macroscopically medium-grained crysstalline aggregates, along with some azurite and supergene quartz. These minerals are deposited on free surfaces in the chalcopyritic lodes. The surfaces are provided by bedding, cross bedding, joints, and late fractures and by sericite within early minor fractures and along detrital grain boundaries within the quartzite. The malachite deposits cited which occur in other parts of the stratigraphic section generally originate from leaching of chalcopyrite-quartz veins or of chalcopyritic lodes in coarse- to medium-grained, light-colored quartzite host rocks contiguous with quartz veins. This malachite is deposited on free surfaces within the quartzite host rocks or transported for distances up to 500 ft (150 m) to be deposited on medium- to fine-grained, dark-colored shaly quartzites and shales.

Origin: The chalcopyritic precursors of the malachite deposits developed in concordant quartz-poor lodes in porous host rocks in areas where quartz veins are thin and sparsely distributed and possibly were under compression at the time of chalcopyrite mineralization. This deprived the copper-bearing hydrothermal fluids of established conduits of migration and of dilatant fractures in quartz veins for the precipitation of chalcopyrite. The relatively porous coarser grained, light-colored (hence, pure) quartzites in both the Stirling Quartzite and the Wood Canyon Formation offered suitable alternative conduits for the passage of hydrothermal fluids, favoring formation of quartz-poor lodes rather than quartz veins. These quartzites are porous because they contain many interstitial spaces incompletely filled with diagenetic quartz overgrowths and lack closely packed shaly matrix minerals.

Therefore, quartz-poor chalcopyritic lodes developed in preference to the chalcopyrite-quartz veins which are the characteristic hosts for copper mineralization in the rest of the Johnnie district. The chalcopyritic lodes in the area south of Route 16 facing the Amargosa Desert are transitional northward along strike into the more typical quartz vein type of deposit.

The most prominent malachite deposits occur in beds within the B member of the Stirling Quartzite throughout an interval which also contains most of the specularite-quartz veins in the district. The deposition of the specularite, as well as sericite, as wall-rock alteration products were doubtlessly triggered by the passage of the hydrothermal fluid through local iron-rich, dark-colored shaly quartzite beds in the B member. Analyses show that these rocks normally contain an average of 3 percent iron, 100 times more than do the light-colored quartzites nearby.

Significant amounts of pyrite are lacking from the areas where malachite lodes occur; but, pyrite is present as a wall-rock alteration product in the specularite-bearing quartz stringer lodes and quartz veins in two areas where chalcopyrite is absent (fig. 17). (The distribution of this pyrite will be discussed further under <u>Specularite Alteration in the Stirling Quartzite</u>.) Consequently malachite was precipitated near its hypogene source; because, during supergene leaching, the absence of pyrite in the chalcopyritic lodes inhibited the production of excess sulfuric acid required to disperse the dissolved copper.

The role of sericite in malachite precipitation was not clarified during this study. The sericite-filled fractures in the host rocks of the malachite lodes in the Stirling Quartzite provided porous conduits for the migration of supergene solutions and may have provided free surfaces for the deposition of malachite. In addition, possible reaction of the copper laden supergene solutions with the surfaces of the sericite grains may have facilitated the nucleation of malachite crystals there in preference to the apparently less reactive surfaces of quartz grains. A reaction on the surfaces of minerals of sericitic affinity is suggested in the apparent preferential deposition of malachite on shaly rocks elsewhere in the stratigraphic section.

These deposits are intriguing, because three observations suggest that they are syngenetic, bedded copper deposits rather than being epigenetic as has been implied so far: (1) the deposits are concordant; (2) most are apparently confined to a single stratigraphic interval within the basal clastic series of a stratigraphic assemblage; and (3) they are widespread laterally throughout that interval. From personal observations and discussion, I know or infer that similar deposits are situated on the Copper Giant property (Appendix D; fig. 2), in the Stirling district (figs. 1 and 2), and maybe elsewhere in the region. At least some of the similar deposits are in the same stratigraphic interval.

Bedded copper sulfide deposits with affiliated iron oxides occur in Idaho, Montana, and British Columbia (Wedow, 1975) in the Belt-Purcell Supergroup (McGill, 1970) of younger Precambrian age (Obradovich and Peterman, 1968). The Belt-Purcell Supergroup is a lowest or, pre-, miogeosynclinal sequence of sedimentary rocks, which is stratigraphically equivalent to the pre-Johnnie Formation sedimentary rocks in southern Nevada and southern California (Crittenden and others, 1971).

The syngenetic appearance of the chalcopyritic lodes in the Johnnie district may be misleading. The stratigraphic restriction of these may have resulted from the presence of iron required for the hydrothermal formation of the specularite which is allied with the chalcopyrite and from the presence of the apparently unique local imbalance between fracture dilatancy and rock porosity, which imbalance was conducive to the formation of quartz-poor lodes. The diverse mineralogic and morphologic types of ore deposits in the Johnnie district are united by their mutual commonality of origin. To ascribe a syngenetic, hence uncommon, origin to a particular type of deposit is a departure from the rationale of unity, which departure further argues against there being syngenetic deposits here.

The possibility is broached in later sections that the ore deposits in the Johnnie district originated by a combination of syngenetic and epigenetic events. So, it may be that the chalcopyritic lodes are syngenetic deposits modified by later hydrothermal activity.

Placer Gold Deposits

Distribution: Eluvial and gulch placer gold deposits (terminology after Wells (1969)) occur on hillsides and gully bottoms in the vein gold producing areas of the Johnnie district. The locations of the most extensive placer workings in the district are shown on plates 2 and 3. Some of these, as well as others throughout the district, were probably very low grade but were worked because they presented large volumes of gravel to the miner.

The principal placer deposits (pls. 2 and 3) in the district follow. Gold was produced from parts, if not all, of each; and some reserves remain:

(1) The Hall placer, which extends northwest from the Johnnie mine through the Johnnie, Last Chance, and Butterfly nos. 1-3 claims.

(2) The Bartlett placer, extending north and south through the Globe claim. The term is used broadly here to include that portion of the deposit that extends through the Globe claim, which

is not a part of the present Bartlett property.

(3) The Labbe placer, which is in the same gulch as the Bartlett placer on the east end of the Doris A. L. claim.

(4) An unnamed placer on the Labbe mining property which extends north and south through the mutual end line of the Westend and Broadway claims.

(5) The Kusick placer, named here after its owner at the time of its operation in the 1930's, located in the gullies southeast of the Congress mine.

Description: Eluvial placer deposits are present on slopes downhill from outcropping gold-bearing quartz veins. The gold, derived from the weathering of these outcrops, is hackly to angular and locally contains bits of adherent quartz.

Gulch placer deposits, containing small, less angular gold nuggets derived from the reworking of previous eluvial placer deposits, begin at the bases of the slopes along which the eluvial deposits lie and extend for moderate distances down the gullies. Some gulch placers, such as that one north of the Broadway and Westend claims, are related to no presently outcropping quartz veins, the former presumably having been completely removed by erosion. The gold occurs in gravelly pay streaks on the bedrock surfaces of the gullies, where it is buried beneath late Pliocene to middle Pleistocene older fanglomerate, sometimes moderately caliche-cemented, as thick as 25 ft (8 m) (Labbe, 1921). Labbe (1921) indicates that more than one branching and rejoining pay streak may be present in an individual gully, and Labbe (1921) notes that, although the pay streaks are usually 1 ft (0.3 m) thick, one pay streak in the Bartlett placer was 8 ft (2.5 m) thick. A private report states that the placer gold was concentrated in pockets in natural traps on the bedrock along the pay streaks, and Labbe (1921) describes these traps in detail.

Tenor: Labbe (1921) and Vanderburg (1936) indicate that gold values in the pay streaks ranged up to approximately 1 oz/cu yd of gravel. Usually the pay streaks probably ran less than 0.1 oz/cu yd. Vanderburg (1936) and private reports indicate that the fineness of the placer gold was approximately 880.

Origin: During the late Pliocene to middle Pleistocene epochs, gold was concentrated in channels on the floors of gullies, which correspond to the present day ones, which were cut down through the post-latest Miocene pediment surfaces discussed under Geomorphology. The vigorous alluviation responsible for this downcutting provided a favorable environment for the accumulation of placer materials. The concentrations of gold were rapidly buried and preserved by deposits of older fanglomerate of late Pliocene to middle Pleistocene age. Subsequently, a major portion of the older fanglomerate was removed by erosion, reexposing the banks of the initial gullies with intact gold placer deposits still buried at their bottoms. The eluvial placers formed during erosion which followed and which continues into the present. The lode sources for the placer deposits which are not related to outcropping quartz veins were eroded away during pedimentation or during the ensuing erosion which dissected the pediments.

Hypogene Mineralogic Zonation

Introduction

This analysis of the distribution of ore and gangue minerals, summarized in figures 17, 18, and 19, in the Johnnie district is intended to clarify observations and problems recognized elsewhere throughout this study.

There are two centers of gold-chalcopyrite-pyrite mineralization at the surface in the district. Arranged symmetrically about these are zones of chalcopyrite-galena mineralization succeeded outward by zones of galena mineralization. The silver-lead ratio may increase outward, also (compare table 3 with Galena-Calcite-Quartz Veins). This general succession from copper-gold to lead (-silver) mineralization is observed in many zoned mining districts and is commonly referred to as thermal zonation. The controls on mineralogic zonation in the Johnnie district are unknown; but a local temperature control appears to have played a part in localizing gold-bearing ore shoots in the district (see Stratigraphic Localization of Gold Mineralization).

Before Basin-and-Range faulting and erosion, a zone of chlorite-bearing quartz veins may have flanked the district at depth and may possibly underlie it at present; and a zone of calcite veins may have overlain the district.

Literal Analysis

Figure 17 (p. 93) is a literal grouping of similar assemblages of ore minerals which is further subdivided on the basis of gangue mineralogy. It demonstrates the observed areal disposition of the minerals along gradients of changing mineralogy radiating from centers of hypogene mineralogic zonation in the area from the Johnnie to the Doris mines (pl. 2) and in the area of the Congress mine (pl. 3).

The distribution of the mineralogic zones (fig. 17) usually corresponds with the areas of hydrothermal activity in which quartz-bearing structures occur (fig. 16), because these are the areas containing the quartz-bearing structures which host later ore mineralization. Some areas of hydrothermal activity are shown in figure 17 as not being mineralized, because they fall beyond the limits of the areas entered by the hydrothermal fluids during metallization later.

Gangue Mineralogic Zones: Gangue mineralogic zones derive from the presence of unusual gangue minerals in quartz-bearing structures regardless whether ore minerals are present or absent. The unusual minerals result from the reaction of hydrothermal fluids with rocks whose compositions are out of the ordinary. Thus, calcite zones occur in and near horizons of rock containing numerous dolomite beds, particularly in the Johnnie Formation, and specularite usually occurs around iron-rich horizons in the lower part of the Stirling Quartzite (see Malachite Deposits). Chlorite is a characteristic gangue mineral in rocks east of the Main and Northwest segments of the Grapevine fault system.

Restricted Interpretation

Figure 18 is a restricted, somewhat conservative, expansion of the data for ore minerals summarized in figure 17. Here reasonable, but not uniquely possible, groupings of nearby zones are achieved by the projection of literal zone boundaries across barren or covered

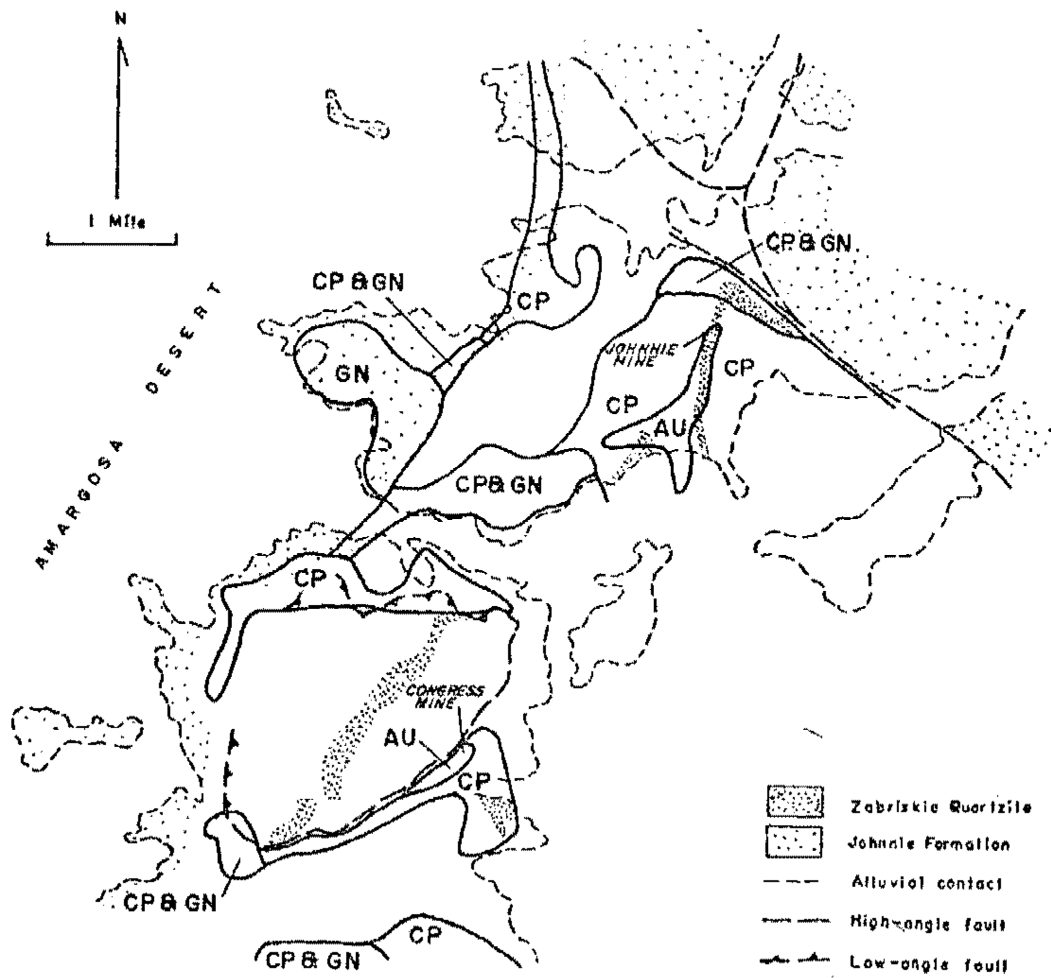

Figure 18. Map showing restricted interpretation of hypogene mineralogic zonation in the Johnnie district.

Au, gold zone; Cp, chalcopyrite zone; Cp and Gn, chalcopyrite-galena zone; Gn, galena zone.

areas. This presentation lends itself to approximately the same interpretation as that given for the Literal Analysis.

Broad Interpretation

Figure 19 is a very broad expansion of the data from figure 17, Literal Analysis, attained by liberally projecting known zones across unmineralized or covered areas.

The resultant pattern shows a symmetrical arrangement of the three main ore mineralogic zones about the two centers of hypogene mineralogic zonation. In a sense, this is how the situation might appear if the mineralization in underlying rocks could be revealed by removing all cover and overlying rocks which were not favorably situated for ore deposition. In another sense, this is how it would appear if mineralization had proceeded uniformly in three dimensions through all of the rocks in the district outward from the centers of hypogene mineralogic zonation.

Vertical Zonation

Chlorite is an associate of chalcopyrite and galena in quartz-bearing structures--with calcite, specularite, pyrite, and sericite--in the weakly mineralized Johnnie Formation east of the Main and Northwest segments of the Grapevine fault system; and chlorite apparently is common in quartz-bearing structures in the Stirling Quartzite, and possible overlying rocks, beyond to the east. So, a chlorite zone was present here structurally lower than the mineralogically zoned rocks now exposed west of the fault system at the time of hydrothermal activity, which preceded the Basin-and-Range normal faulting which

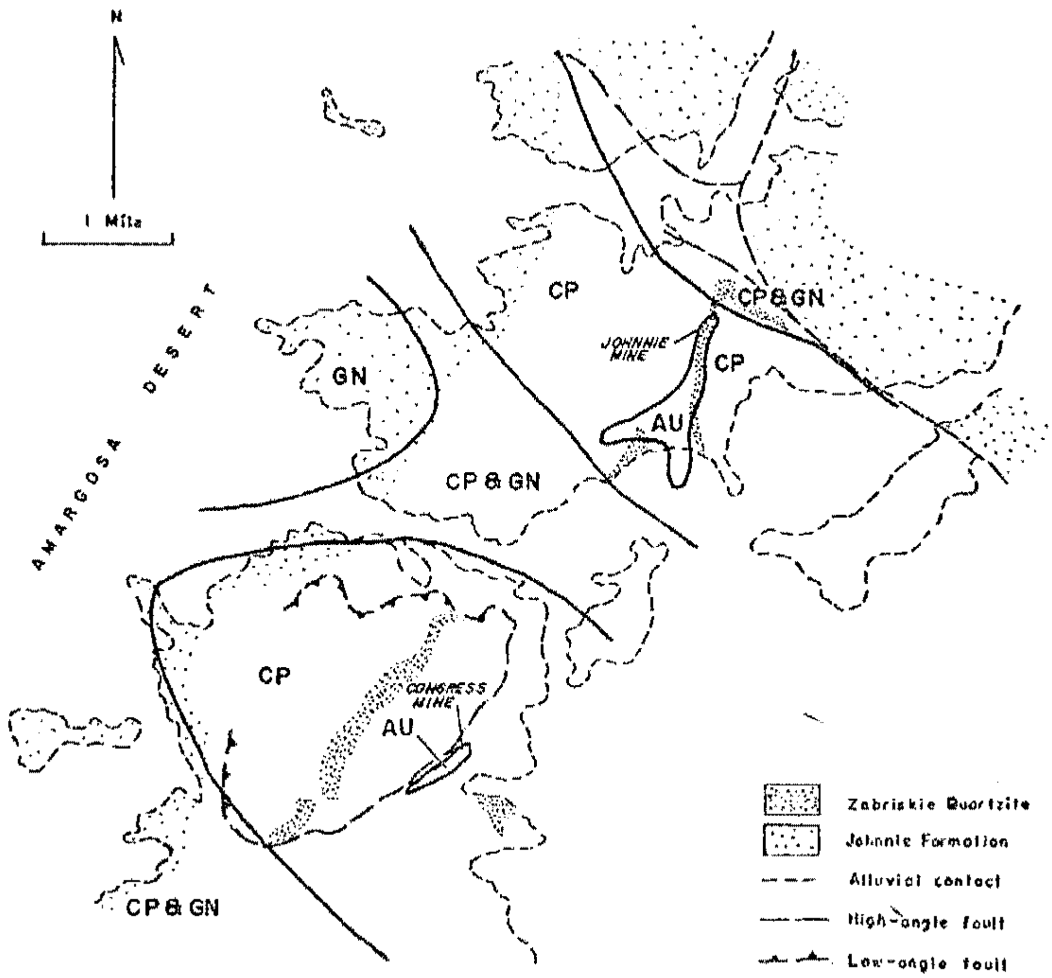

Figure 19. Map showing broad interpretation of hypogene mineralogic zonation in the Johnnie district.

Au, gold zone; Cp, chalcopyrite zone; Cp and Gn, chalcopyrite-galena zone; Gn, galena zone.

brought the two zoned areas into juxtaposition. Thus, the distribution of chlorite is related to a depth zonation as well as to the lateral zonation discussed up to now; this suggests that a chlorite zone may underly the district.

Some carbonate rocks in the portion of the stratigraphic section from the Bonanza King Formation up through the Silurian System throughout an area generally 10 mi (15 km) northwest of the Johnnie district (fig. 2) host epithermal calcite veins, of apparent postorogenic age, up to 8 ft (2.5 m) thick. Among other possible explanations, these could have been recrystallized from local rocks under the influence of waning, dilute hydrothermal systems. These veins, being the apparent products of silica-depleted systems under shallow depth conditions, may demonstrate phenomena which occurred in the rocks overlying the Johnnie district at the time of hydrothermal activity and before their erosional removal.

Structural Aspects of Hypogene Ore Deposits

Introduction

This section of the report discusses, in order of decreasing magnitude, the structural features responsible for localizing the hydrothermal ore deposits at and within the Johnnie district. In general outline, these are, successively: fundamental features responsible for the localization of the district; an inferred major longitudinal structure within the district in which are located the principal structures in which most quartz veins are located; the quartz veins; and finally gold-bearing ore shoots in them. Some stratigraphic controls on gold deposition which interacted with the structural controls are included at the end of the discussion.

Localization of District

Any discussion of the localization of the district must speculate upon, if not explain, the source, magmatic or otherwise, of heat and metal-bearing fluid in the hydrothermal system(s) at the Johnnie and Stirling districts and Copper Giant property (see Characteristics of Principal Structures). It must also explain the fundamental structure which guided hydrothermal activity and which controlled the observed N. 35° E. alignment of mineralized areas (fig. 20), and it must explain the dilatant conditions in the fundamental structure. A magmatic source is unlikely because of the apparent absence of any igneous activity in or near the district (see Regional Geology).

The localizing feature could have been a lineament in the basement of Precambrian metamorphic rocks which permitted the flow of heat, magma, or magmatic products from the lower crust or upper mantle and also controlled the alignment of the mineralized areas.

Likewise, if a portion of the floor of the Pacific Ocean had been subducted beneath the district (Menard, 1955), a deep fracture in the oceanic crust would have given access to upper mantle sources of heat and/or magma and influenced the observed alignment of the mineralized areas. If a portion of a mid-oceanic ridge had been subducted, its medial rift would have provided the requisite deep fracture and the overlying rocks might have arched up, becoming dilatant. Lipman and others (1971) and Lowell (1974a and 1974b) demonstrate the possibility, but not the presence, of paleosubduction beneath the central Cordillera.

Premiogeosynclinal topography could have influenced the localization of the district. An elongate basin would explain the alignment of the mineralized areas, or have permitted a high flow of heat from subcrustal sources, or have permitted the spontaneous initiation of hydrothermal activity by the locally high hydrostatic pressures and high heat from the geothermal gradient in these thick, basinal sediments (see Summary of Ore Genesis). Although the basal miogeosynclinal sediments may thicken beneath the district (see Johnnie Formation), a basin cannot be inferred with certainty.

An elongate ridge in the basement may explain the alignment of mineralized areas, and/or have provided a barrier which diverted upward any hydrothermal fluids migrating at depth (Park and McDiarmid, 1970, p. 65), and--possibly--localized high heat flow. Simplifying, heat flow (in part, a function of thermal conductivity and geothermal gradient (Jaeger (1965)) above the surface of a basement ridge would be greater than above the deeper surface of an adjacent depression, because the thermal conductivity of basement metamorphic rocks is expected to be higher than that of the overlying sedimentary rocks.

A hypothesized major longitudinal structure, which trends N. 35° E. through the Johnnie district, could be a surface manifestation of the fundamental structure speculated upon here. The major longitudinal structure could be a feature propagated upward from--or developed above--a basement lineament, a tensional feature developed above a mid-oceanic ridge or a basement ridge, or a supracrustal feature which permitted the ingress of hydrothermal fluids generated within the lower part of a coaxial basin or migrating upward from another deeper source.

Major Longitudinal Structure and Localization of Principal Structures

Characteristics of Principal Structures: The term "principal structure" is used here to refer to the loci of hydrothermal activity within the mineralized areas in the region of the Johnnie district. The principal structures are trains of the biggest and most numerous quartz veins in their respective localities; some of them are the loci of gold mineralization; and all important production in the region has been taken from these structures.

The five principal structures recognized are 2 to 3 mi (3-5 km) long, strike--on the average--ENE, and generally dip north. Their strikes change clockwise from NNE to ESE; the average is here considered to be ENE, similar to the N. 70° E. average trend of the high-angle quartz veins in the Johnnie district (see Statistical Analysis of Quartz Veins and fig. 21). The principal structures probably dip north, because most of the lesser structures in them, including most of the prominent veins, dip north.

The Johnnie and Congress structures have been documented well by this study. The presences of the others are inferred with varying amounts of certainty. The principal structures are (fig. 20):

(1) Stirling structure. This 2-mi-long (3 km), NNE-trending structure near the Stirling mine, in sec. 7, T. 17 S., R. 54 E., localizes a number of approximately east-trending gold-quartz veins in the approximately 10 sq mi (25 sq km) mineralized area of the Stirling district (fig. 2).

(2) Johnnie structure. This trends SSE from the Oversight claim to the Crown Point claim; curves west to the West end claim

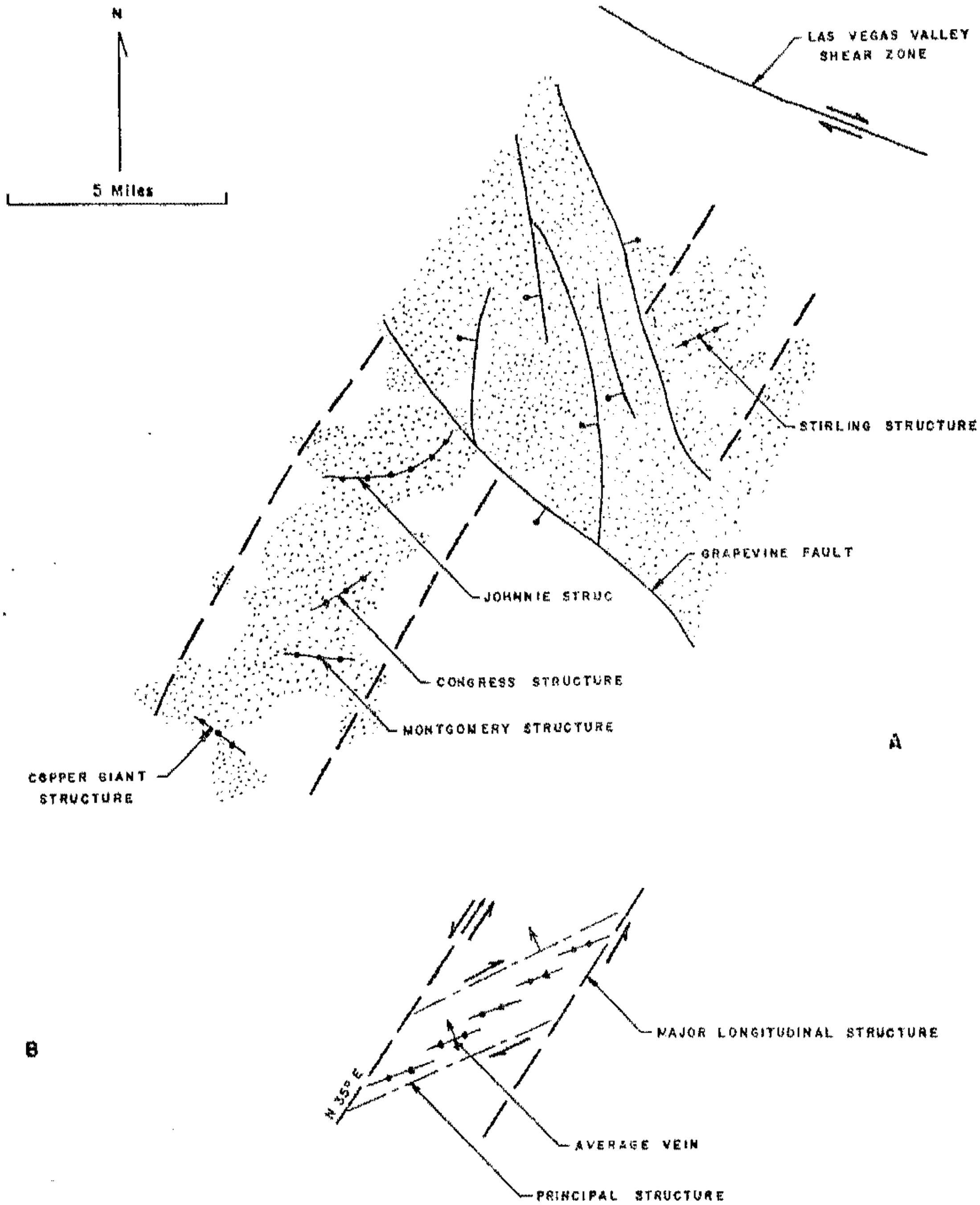

Figure 20. (A) Map of principal structures, showing inferred offset of major longitudinal structure (broken line). Faults from Cornwall (1972, pl. 1); ball on downthrown side. (B) Diagram illustrating relations among mineral localizing structures, showing inferred directions of strike separation.

Outcrop areas in (A) stippled.

(pl. 2); appears to continue west through the Lilyan group of claims (pl. 2); and gradually curves WNW to the vicinity of the Coe property (pl. 1). Its total length is approximately 3.5 mi (5.5 km).

Interestingly, the east end of the Johnnie structure curves in a sense opposite to the dextral bend in that part of the district (see Dextral Bend), but the significance of this is not evident.

(3) Congress structure. Trends northeast through the Congress mine for about 2 mi (3 km).

(4) Montgomery structure. Trends WNW across the south end of the area mapped in plate 1 for at least 2.5 mi (4 km).

(5) Copper Giant structure. Reconnaissance and map interpretation suggest that the locus of quartz veining on the Copper Giant property (see Appendix D and fig. 2) is along a WNW structure 1 mi (1.5 km) long.

Alignment of Principal Structures and Character of Major Longitudinal Structure: The areas in which the Johnnie, Congress, Montgomery, and Copper Giant structures are located lie along a line which trends N. 35° E. (fig. 20). This alignment suggests, inconclusively, that the principal structures are controlled by a N. 35° E. major longitudinal structure. The locality of the Stirling structure is offset approximately 4 mi (6.5 km) off the northeast extension of this line (fig. 20). This may reflect either the original distribution of the structures or displacement of the major longitudinal structure along the Grapevine fault system and parallel faults mapped in the Spring

Mountains (Vincelette, 1964, pl. 1; Burchfiel and others, 1974, fig. 3) which lie between the Stirling district and the rest of the principal structures.

The major longitudinal structure has no other evident surface manifestation but parallels regional structure and the trend of the longitudinal fault set defined in High-Angle Faults (fig. 14). The major longitudinal structure is probably a supracrustal feature; and, considering that the Johnnie district probably is in a plate of overthrust rocks, the major longitudinal structure may terminate at depth against the thrust surface.

Origin of Longitudinal and Principal Structures: Examination of vein morphology suggests that their host fractures became dilatant by shear, rather than tensional, movements along them at the time of quartz veining. Therefore, given a shear origin, the *en echelon* right vein pattern and acute intersection of the veins with the trends of the principal structures (fig. 20) can be explained by right-lateral strike slip along the principal structures. The northward dip of the veins and *en echelon* pattern suggest that there was a northward dip slip component of movement along the principal structures as well.

The occurrence of four, or five, of these north-dipping, ENE-trending, right-lateral shear structures along a N. 35° E. major longitudinal structure the strike of which diverges 30° from the average for the principal structures, in turn, suggests that the principal structures originated from left-lateral movement, with a component of north-side-down normal movement, along the apparent major longitudinal structure which enclosed them. If movement along the major longitudinal structure were right-lateral, the included principal structures would display tensional, not shear origins. Considering only the plan relations: the inferred directions of separation along, and the average angle between, the longitudinal and principal structures suggests the two could be in conjugate shear relation (fig. 20).

The major longitudinal structure, if real, could be genetically related to either the longitudinal faults in the district or to the Las Vegas Valley shear zone. The implication that the major longitudinal structure is genetically related to the longitudinal fault set, the west sides of the average members of which are displaced downward, is reinforced by the fact that inferred normal separations to the north along the principal structures could be the consequences of similar separation on the major longitudinal structure. This implies, perhaps correctly, a tensional aspect to the origin of the major longitudinal structure not ascribed to it in the preceding paragraph. The lack of other surface manifestations of the major longitudinal structure suggests that movement along it was small, and this may be additional evidence for its having had a tensional origin. The major longitudinal structure could bear either a conjugate or second order relation to the Las Vegas Valley shear zone; however, the geometric and temporal relations between the shear zone and the major longitudinal structure are unclear.

Areas of Hydrothermal Activity and Hypogene Plumbing

Areas throughout which hydrothermal activity is pronounced in the Johnnie district are shown in figure 16. These are areas containing significant numbers of the quartz veins and related structures discussed in Morphology of Quartz-Bearing Structures. Minor numbers of quartz veinlets occur beyond the limits of the areas indicated. Hydrothermal activity was distributed in all or parts of each of the following geologic features:

(1) The Johnnie, Congress, and Montgomery structures.

(2) The Zabriskie Quartzite and rocks immediately below.

(3) The contact of the Johnnie Formation and Stirling Quartzite and the sole of the Congress low-angle normal fault.

(4) Moderate to profuse, usually sub-map-scale quartz veinlets occur in all exposures of the Johnnie Formation in the Johnnie district. They are localized along bedding-related structures, particularly in fissile rocks, and in high-angle joints and small fractures.

When considered in three dimensions, this distribution was a consequence of the hypogene plumbing.

Widespread veinlets beyond the limits of the areas of hydrothermal activity, particularly in the Johnnie Formation, appear to represent the general path of hydrothermal fluids ascending along a broad front through the major longitudinal structure or through or above the fundamental structure which localized the district.

Most of these ascending fluids were collected and canalized along the Johnnie-Stirling contact, the sole of the Congress low-angle normal fault, and--probably--along the elements of the Grapevine fault system in existence at the time. Some of the fluids migrating upward to the west along the Johnnie-Stirling contact probably were diverted in the Congress low-angle fault along their line of intersection. A similar process of collection and canalization may have occurred at depth along the Montgomery thrust, but the thrust and structures ancillary to it do not appear to have exerted any control upon exposed ore deposits.

The concentration of hydrothermal activity along the principal structures results from their having penetrated, and consequently having been fed hydrothermally by the Johnnie-Stirling contact and the sole of the Congress low-angle normal fault.

Profuse hydrothermal activity in the Zabriskie Quartzite and the rocks immediately below resulted from flooding by hydrothermal fluids where the principal structures intersected the rocks named (see Stratigraphic Localization of Gold Mineralization).

Statistical Analysis of Quartz Veins

Inspection of figures 9, 10, and 21 shows that high-angle quartz veins occur in any orientation for which pre-existent structures are present in the Johnnie district. However, a minor group of veins, striking parallel to the northwest-trending conjugate fracture set, dips oppositely. The most common high-angle veins (fig. 21) presumably were derived from parts of an antecedent conjugate fracture set and the adjacent extension fracture set (fig. 10).

The most common high-angle veins trend through approximately 70° of arc, with the average being N. 70° E., and dip north. North dips are progressively steeper in the more easterly trending veins.

The average attitude of the concordant quartz veins and quartz stringer lodes reflect the average of bedding attitude in the district (fig. 21).

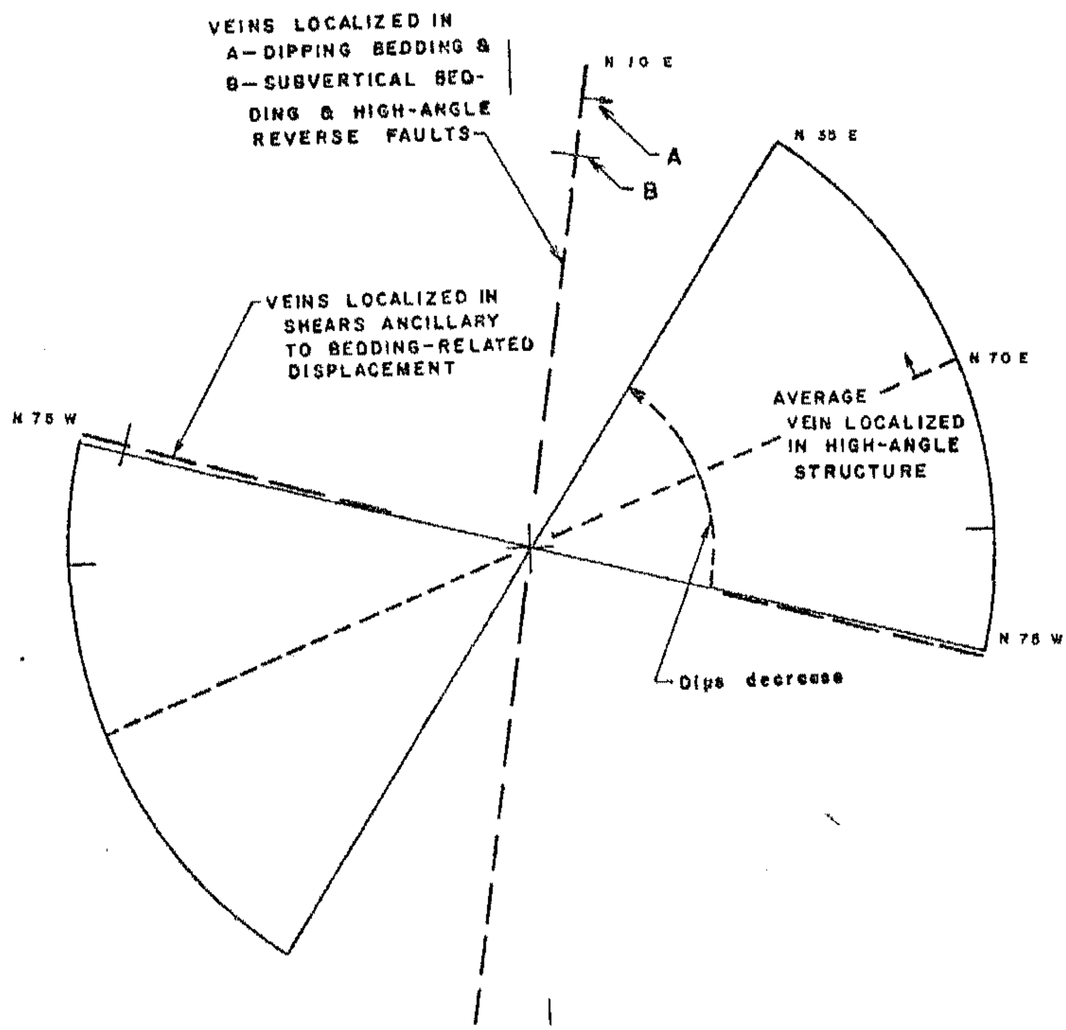

Figure 21. Diagram illustrating strike and direction of dip and ranges of important groups of veins localized in high-angle and related structures and veins localized in bedding-related structures. Refer to figure 10 for correspondence of veins localized in high-angle and related structures with antecedent fractures.

Appendix E amplifies upon these conclusions and presents the balance of the statistical analysis conducted during this study. The balance does not furnish conclusions essential to understanding the genesis of the ore deposits in the district.

Localization of Quartz-Bearing and Related Structures

Basically, at the time of hydrothermal activity, the hydrothermal fluids and resultant quartz-bearing structures occupied whichever fractures and related host structures happened to be dilatant at the time. This had two consequences. First, hydrothermal fluids ascending from structurally low areas were funneled against porous features such as dipping contacts and fault planes, then diverted into the ENE-trending principal structures where the most abundant high-angle quartz veins were localized (see Areas of Hydrothermal Activity and Hypogene Plumbing). Other quartz veins, some quite prominent, such as the one on the Nomad No. 1 claim (pl. 3), developed in areas outside of the principal structures but conformed to the same pattern of orientation. Second, although a variety of pre-existent potential host structures were available, only ones in the structurally favored positions necessary for them to become dilatant at the time of hydrothermal activity actually received quartz veins (see Statistical Analysis of Quartz Veins). (Conversely, since not all favorably oriented fractures were structurally active at the time of quartz veining, some are devoid of veins also.)

The orientation of the stress field in existence at the time of dilatancy cannot be deduced, because: (1) conjugate and lower order relations within and/or among groups of quartz-bearing structures, principal mineralized structures, and the major longitudinal structure are not certain; (2) the quartz-bearing structures apparently occupied available pre-existent fractures which were favorably, but not necessarily exactly, located with respect to their ideal positions in the stress system in effect; and (3) by the advent of quartz veining, the miogeosynclinal rocks of the district had been rendered quite anisotropic by folding and high- and low-angle faulting so that any large-scale (regional- or even district-scale) stress pattern was subject to local reorientation (McKinstry, 1955, p. 184).

Vein Configuration and Localization of Ore Shoots

High-angle veins are pinching and swelling, tabular bodies of quartz bent in a curviplanar manner; the veins are elliptical to equidimensional in longitudinal view and are up to 750 ft (230 m) long. The swells constitute raking ore shoots up to 20 ft (6 m) thick, 75 ft (25 m) wide, and 200 ft (60 m) long (for example, the ore shoot followed by the Johnnie shaft to the Third level, pls. 5 and 6); the ore shoots are thickest toward the centers of the veins and thinner outward (fig. 22). The veins terminate distally in horsetail and *en echelon* structures which generally curve to follow bedding and which are filled with gouge or sparse amounts of quartz (figs. 22 and 23). Pinches within the veins are likewise gougy or contain sparse amounts of quartz or quartz-impregnated breccia and gouge (fig. 24 and pl. 6). Gouge zones at the ends of veins and along pinches result from the prolonged milling action there of the walls of the host fractures, which walls were not removed from contact during filling as were the walls of ore shoots (fig. 25).

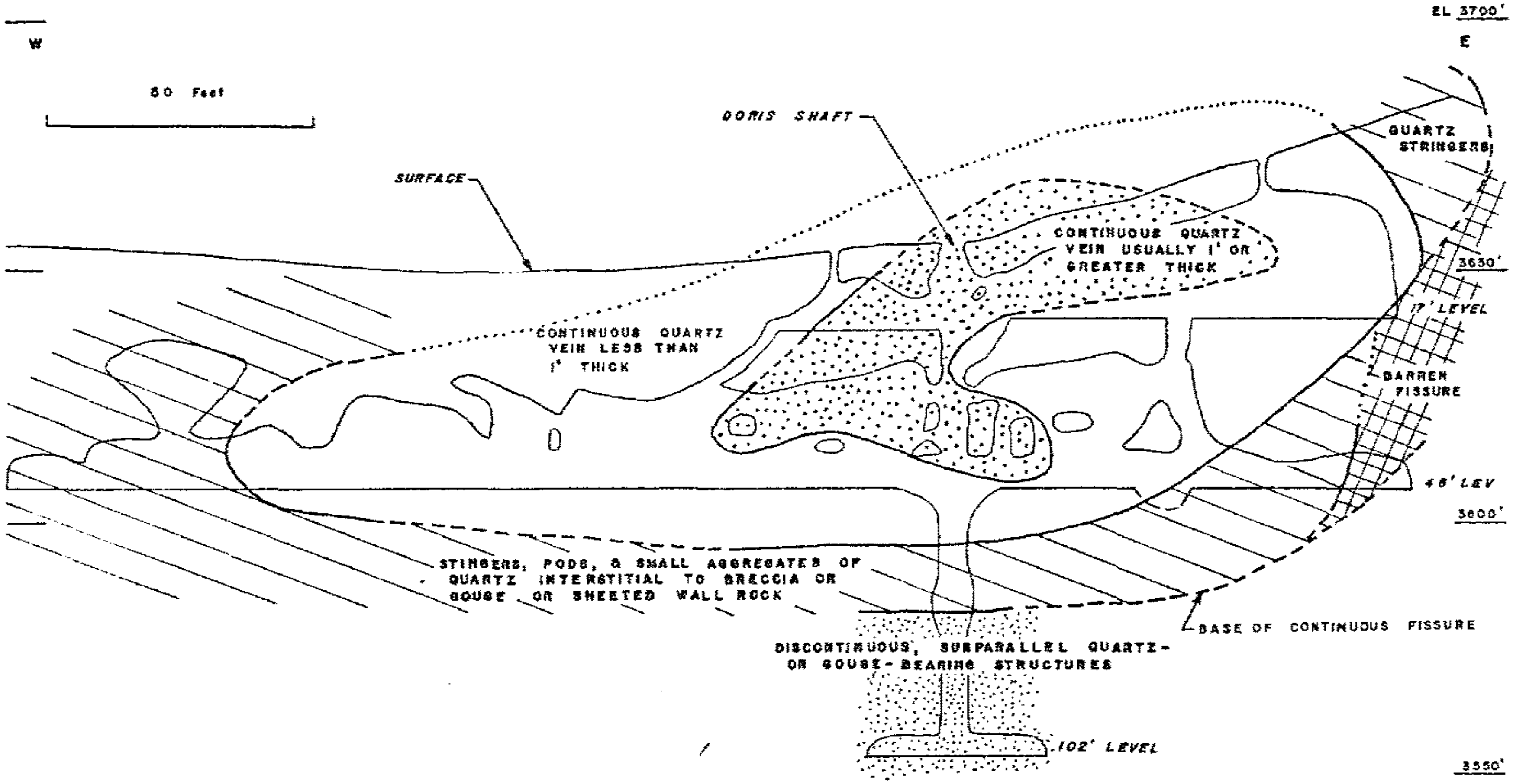

Figure 22. Illustration of general structure of a simple fissure: Composite longitudinal projection of workings and vein in the Doris mine in a plane approximately N. 72° W. Looks north. Generalized from mapping by the author at a scale of 1" = 20'. For location of Doris mine, see plate 2.

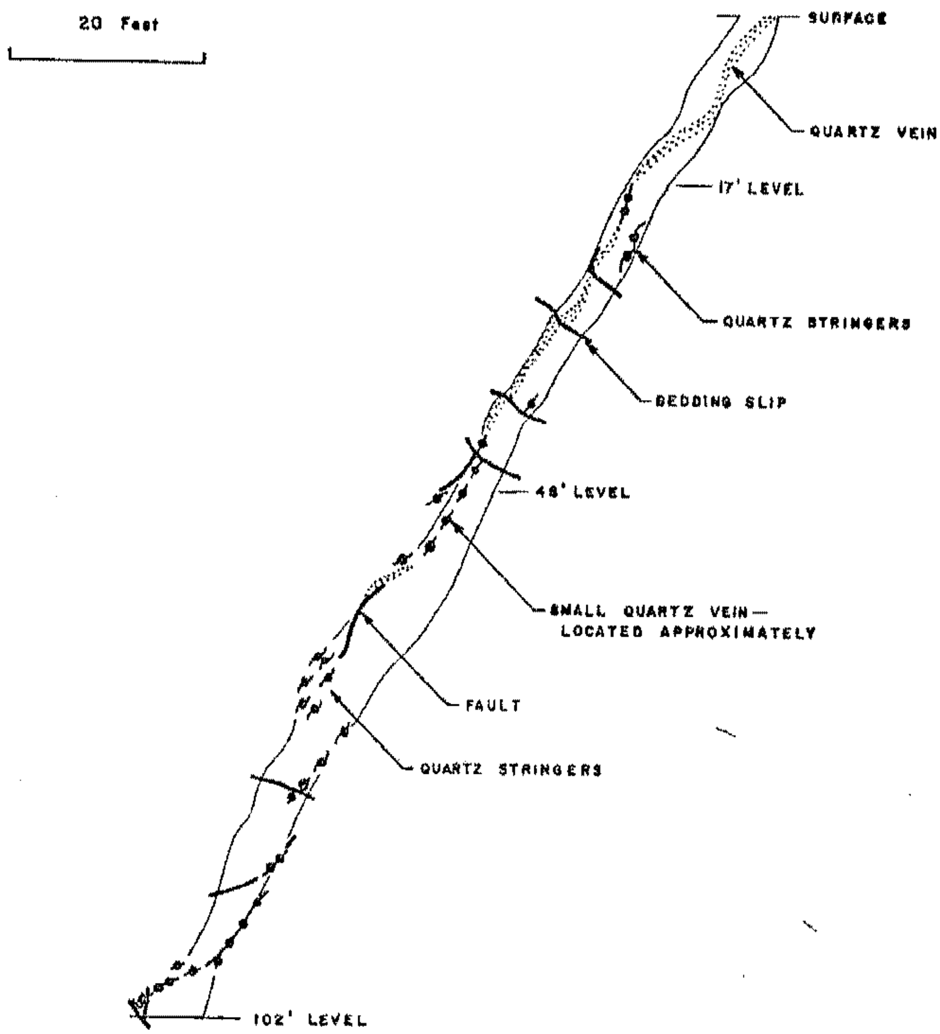

Figure 23. Doris mine. Section through shaft on line N. 18° W. Looks east. Compass-tape survey by S. W. Ivosevic, August 27, 1972.

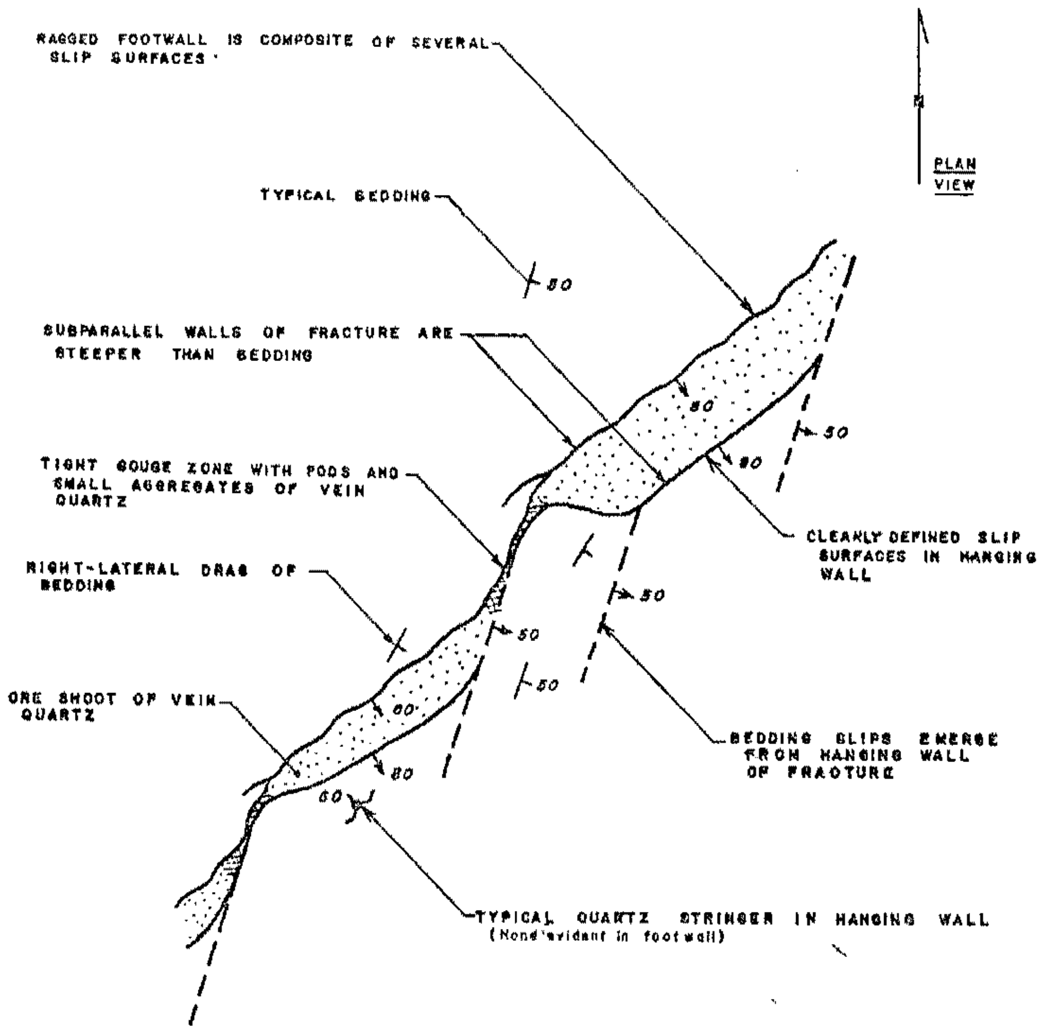

Figure 24. Diagram illustrating typical vein configuration along stoped area of Johnnie vein at surface southwest of Johnnie shaft. Generalized from mapping by the author at a scale of 1" = 10'. For location of Johnnie vein, see plate 4.

Scale of this figure is 1" equals approximately 20'.

Notes:

1. Stopes (ore shoots) become longer and wider toward the northeast.

2. Vein walls become less clearly defined toward the southwest (may be a consequence of decreasing wall-rock competence).

3. Viewing this diagram from the west side depicts the typical inferred vertical section and suggests normal separation.

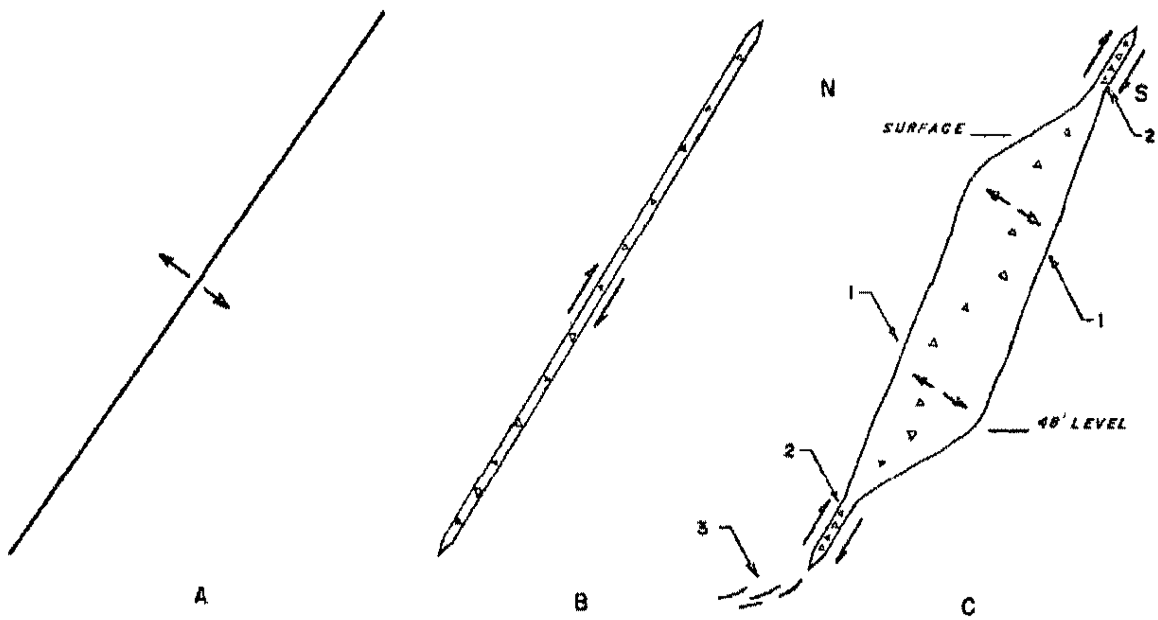

Figure 25. Serial diagrams illustrating, in vertical section, development of a simple fissure and its distal termination. (A) Fracture initially develops. (B) Inception of movement along fracture with minor displacement, causing brecciation along entire fracture. (C) With continued movement, center of fracture, in an aggregate sense, widens, (1), which separates opposite walls and removes them from furthur milling action. Distal parts of original fracture, (2), remain in contact and brecciation continues there. Beyond ends of fissure, dislocation is distributed throughout discontinuous, subparallel structures, (3). Fiducials in (C) show applicability of this situation to the Doris mine (compare with fig. 23). Scale across fissure is greatly exaggerated.

Although ore shoots are usually clean fissures filled with quartz, the dilatancy of the veins resulted from shear; that is, movement subparallel to the walls of the veins as opposed to tensional separation normal to them. The direction of shear locally varied from the general condition described in Origin of Longitudinal and Principal Structures in accord with local circumstances. The fissures opened, during quartz veining, in the classical manner near deflections in the host fractures during continued shear along the fractures. These deflections in veins in the Johnnie district frequently resulted from the distortion of the fractures by bedding slip before quartz veining as illustrated in figure 26. Figure 24 illustrates the relation between bedding slips and ore shoots.

These bedding slips which subsequently localized deflections are frequently in argillaceous rocks and localized ore shoots in the manner discussed under Stratigraphic Localization of Gold Mineralization, Zabriskie Quartzite. The bedding slips formed gougy bodies, which later were sericitized, in pinches between ore shoots and along one or both walls of many ore shoots in the district. These impervious, sericitized gougy bodies trapped hydrothermal fluids within ore shoots in veins throughout the district.

The thickest and most persistent ore shoots, which are in the largest veins, usually are the ones in veins in the Zabriskie Quartzite and dolomitic rocks of the Wood Canyon Formation immediately below. The ore shoots in the Congress mine, which are in veins in these rocks, are additionally enlarged by chambering, the collapse of the hanging wall into the fissure to form a breccia zone on that wall of the

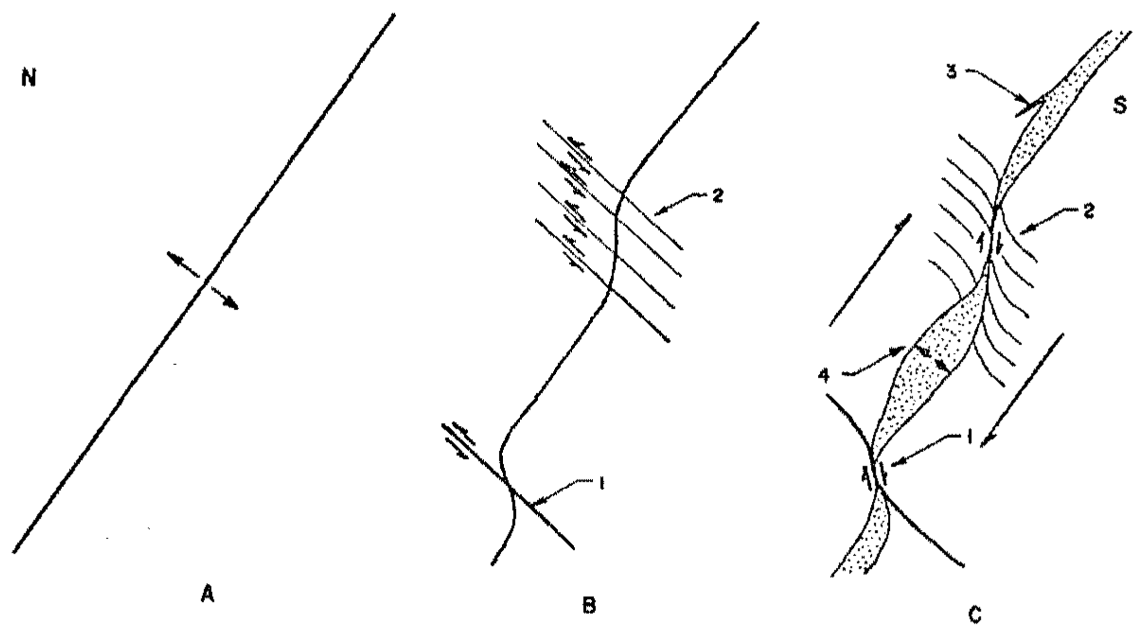

Figure 26. Serial diagrams illustrating, in vertical section, development of horizontal ore shoots in a simple fissure: Generalized vertical section through a portion of the vein in the Doris mine in a plane approximately N. 0° E. Looks east. Generalized from mapping by the author at a scale of 1" = 20'. For location of Doris mine, see plate 2. (A) Fracture initially develops. (B) Fracture offset by (1) bedding slip or (2) cumulative updip translation of bedding. (C) Reverse movement occurs along fracture, as shown by (1) drag pattern in bedding slip, (2) drag pattern in thin-bedded rocks, and (3) attitude of feather veinlet. Subvertical portions at (1) and (2) of fracture impinge, causing north-dipping portions, such as (4), of fracture to widen.

This figure illustrates ore shoots of all sizes, so that the distance between (1) and (2) can be between 5' to 40'.

fissure; this is a common situation, which is summarized by Lindgren (1933, p. 163).

Structural Localization of Gold Mineralization

Not only are the largest veins in the Johnnie district localized in the Zabriskie Quartzite and rocks immediately below, but also--with exceptions--the gold-producing veins in the Johnnie district are those in the Johnnie and Congress principal mineralized structures at their intersections with these rocks at the heads of gradients of hypogene mineralogic zonation along these structures (see Hypogene Mineralogic Zonation). Conversely, prominent veins, not in centers of hypogene mineralogic zonation were localized by the same controlling features and only contain chalcopyrite and/or galena mineralization.

Stratigraphic Localization of Gold Mineralization

The localization of the richest gold ore in the largest ore shoots in the Zabriskie Quartzite and rocks immediately below resulted from the imposition of stratigraphic controls on the prevailing structural controls. The stratigraphic controls were the interrelated results of the mechanical and chemical behaviors of the rocks of the upper part of the Wood Canyon Formation, the Zabriskie Quartzite, and the lowest rocks of the Carrara Formation. (The detailed lithologies of these rocks are given in Appendix A, and the lithologies of the three formations are summarized in table 1.) These same rocks are important hosts for ore deposits elsewhere in and near southern and eastern Nevada.

The effects of the stratigraphic controls were fourfold: (1) the competent rocks held permeable fissures open; (2) widely spaced argillaceous interbeds in the Zabriskie Quartzite, which guided bedding slips, favored the development of large ore shoots; (3) the reactive dolomitic rocks in the Wood Canyon Formation caused a disequilibrium in the composition of the hydrothermal fluid which probably persisted to the top of the Zabriskie Quartzite; and (4) impermeable blankets, particularly at the base of the Carrara Formation, promoted ponding of hydrothermal fluids in the underlying rocks.

Ponding of hydrothermal fluids increased circulation of ore fluids, promoted the deposition of the ore minerals, and--most importantly--permitted the attainment of the higher temperatures apparently necessary for gold deposition in the veins.

Wood Canyon Formation: The rocks of the dolomite-bearing upper part of the upper member of the Wood Canyon Formation (units 2 and 3, Appendix A) are dolomite with some thin interbeds and interlayers of argillaceous and clastic rocks. These rocks all offered permeable conduits and/or reactive surfaces to the entering hydrothermal fluids.

The relatively high competency of the rocks permitted them to generate permeable fissures during fracturing and faulting. Also, when the rocks were comminuted during the shearing which opened the fissures, they formed porous breccias, rather than impermeable gouge. These breccias presented large surface areas for reaction with the hydrothermal fluids, multiplying the innate reactivity and solubility of the dolomite.

The reactive dolomite wall rocks and breccia were conducive to the precipitation of the ore minerals because "the presence of strongly reactive wall rock, out of equilibrium with passing solutions (sic) would produce a rapid change in the chemical character of the solutions"

(Lovering, 1942, p. 5). The ready solution and removal of the dolomite breccia, the process of which is described in Nature of Quartz Filling, created more room in which quartz could be deposited.

Zabriskie Quartzite: This formation of quartzite contains some widely spaced argillaceous interbeds and partings, but is practically monolithic in its upper half (unit 3, Appendix A). During quartz veining, this competent quartzite held open fractures; but when it did brecciate, the breccia was porous and the permeability of the structure was not lessened. The wide stratigraphic spacing between argillaceous interbeds favored the development of large ore shoots by the mechanism described in figure 26.

The argillaceous interbeds controlled the locations of the bedding slips, the distances between which, in turn, controlled the sizes of the ore shoots (fig. 24). The wide spacing between the interbeds in the lower half of the formation caused the ore shoots there to be generally larger than those in the upper part of the underlying Wood Canyon Formation with its more closely spaced argillaceous interbeds. The absence of argillaceous interbeds and physically discrete bedding planes in the upper half of the Zabriskie Quartzite accounts for the very large ore shoots there.

The presence of large ore shoots apparently promoted the precipitation of ore minerals; it is noted (McKinstry, 1955, p. 190) that in many veins the widest ore shoots are also the highest grade. McKinstry (1955) hypothesizes that this is the result of the great volume of circulation there and the presence of a great volume of breccia which presents a large surface area for chemical reactions and cooling.

Carrara Formation: The basal rocks of this formation (units 1 and 2, Appendix A), which overly the Zabriskie Quartzite, are relatively incompetent argillaceous rocks whose initial impermeability was increased by their conversion to gouge during preore bedding slip and by later pre-quartz sericitization (pl. 4). This formed a blanket to the tops of vein systems which caused ponding of hydrothermal fluids below the blankets. Locally, other blankets were formed by zones of tectonic readjustment within the Zabriskie Quartzite in which the quartzite had been reduced to an impermeable powdery gouge.

The ponding by the Carrara Formation in principal mineralized structures is manifested: by the profusion of quartz veins, veinlets, and stringer lodes in the Zabriskie Quartzite and by the silicification of dolomite, below (see Gold-Chalcopyrite-Pyrite-Quartz Veins, Wall-Rock Alteration); by the generally large sizes of the ore shoots beneath the Carrara Formation; and by the local upward flaring of shoots terminated against blankets. The efficiency of the barring of hydrothermal fluids by the basal rocks of the Carrara Formation is evidenced by the passage of quartz veins in the Zabriskie Quartzite into thin, barren structures in the Carrara Formation and by the nearly complete lack of veins in post-Zabriskie Quartzite rocks in the district and surrounding areas. The veins which do occur in the latter rocks in the district are thin (less than 1 in. (2.5 cm) thick) and unmineralized.

Localizing Effects: The action of the four effects of stratigraphic control cited at the opening of this section and of additional unrecognized controls and effects were interrelated and, in an aggregate sense, they influenced ore deposition positively. The effects of hydrothermal ponding can be isolated and these appear to have played a dominant part in the net influence upon gold deposition. The information available also permits the consideration of surface-volume relations and possible pressure effects in the gold-bearing veins.

The trapping of ore fluids within ore shoots by ponding caused them to become thick; therefore, having greater volumes than thin ore shoots, thick shoots developed more sheetlike, post-quartz, pre-metallization fractures than thin shoots; this provided more conduits for the circulation of later, metallizing hydrothermal fluids and provided more sites for the deposition of ore minerals. Ponding also caused the retention of the hydrothermal fluids within the ore shoots for longer periods of time than in unblanketed veins; this permitted more of the chemical reactions involved to go to completion and reduced the loss of heat from the bigger veins, permitting them to maintain higher temperatures than the other veins.

Toulmin and Clark (1967) demonstrate that the cooling of hydrothermal fluid at a point in a prospective ore deposit by heat exchange with the wall rock is minimized where the values of the following are large: width of fissure, porosity of fissure, time of reaction, and velocity of flow; and where the distance from the source of the hydrothermal fluids is small. The fissures hosting ore shoots in and immediately below the Zabriskie Quartzite were the largest in the district and certainly were porous; the time of reaction was prolonged in this setting due to ponding. However, ponding beneath blankets reduced the velocity of flow, mitigating against low heat loss; but ponding may have had an offsetting effect by permitting the large ore shoots to build up in the first place. The distance from the source was probably great enough to be considered constant throughout the district.

High surface-volume ratios promote the reaction between hydrothermal fluids and wall rocks and breccia derived from wall rock, and this, in turn, promotes ore deposition. The ratios were low in the important veins in the Johnnie district because: the host fissures opened without much brecciation; and because the areas of the walls of the veins having been constant, the ratios were inherently lower in the wider, more important veins. However, since the wide veins are productive, the low surface-volume ratios do not seem to have had an important effect. Inasmuch as a low surface-volume ratio also retards cooling (McKinstry, 1955, p. 190), the reduced ratios observed in the district may have aided the retention of heat within the veins and, thus, countered any adverse effect of the low ratio upon the rate of chemical reaction.

Reductions in pressure are generally thought to influence the deposition of ore minerals from hydrothermal fluids. However, pressure changes do not seem to have affected ore deposition here; because nothing in the texture of the vein quartz suggest departures from the prevailing pressure.

Disseminated Gold Possibilities: The hypothesis was tested, with negative results, that hydrothermal ponding in the Johnnie district resulted in enrichment of the wall rock in gold in areas of gold mineralization to produce low grade, bulk mineable ore situations. Thirty-one systematically collected bulk samples of representative materials, including quartz stockworks, from the surface near the Johnnie mine contained no gold or silver upon fire assay.

The results of this test, in conjunction with other observations, shows that epigenetic mineralization in the quartz vein environment in the Johnnie district is restricted to ore shoots in veins.

Regional Implications: The lowest carbonate rocks in the Cambrian sections of the Johnnie district and other districts in the southern Great Basin appear to be exceptionally favorable hosts for epigenetic mineralization. The dolomitic rocks of the upper part of the Wood Canyon Formation, the base of which part (base of Unit 2, Appendix A) demarks the base of the Cambrian System of rocks in the Johnnie district, are (along with the contiguous Zabriskie Quartzite above) the most favorable hosts in the vicinity of the district. In other districts, the lowest Cambrian carbonate rocks, which happen to be the oldest carbonate rocks exposed in their respective stratigraphic sections, influence ore deposition prominently. For example: the comparable rocks in the Pioche district, Nevada (fig. 1), are those of the Combined Metals Member of the Pioche Shale; although mineral deposits occur in rocks stratigraphically below and above the Member, the striking localization of replacement deposits of base metals in the Member at its intersection with high-angle structures is frequently cited as a premier example of stratigraphic control of ore deposition (for example, Westgate and Knopf, 1932; Gemmill, 1968; Tschanz and Pampeyan, 1970). A system of quartz veins containing significant tungsten, beryllium, and fluorine mineralization are localized at the intersections of high-angle structures with the Wheeler bed of the

Pioche Shale in the Lincoln district, White Pine County, Nevada (fig. 1) (Stager, 1960).

It has been speculated that the favorability of these lowest Cambrian carbonate host rocks derives from their being the first rocks in their respective stratigraphic sections to contain fossiliferous material, and, hence, organic carbon which may influence the precipitation of metals from hydrothermal solution. It has also been speculated that the rocks are good hosts simply because they are the lowest carbonate rocks in their stratigraphic sections and, hence, the first reactive rocks encountered by ascending hydrothermal fluids.

The correlation between carbon and gold deposition is well documented, and organic carbon liberated from dissolved dolomite could have had an extremely active influence on the deposition of gold in veins in the upper part of the Wood Canyon Formation in the Johnnie district. Although these rocks, exactly above the base of the Cambrian section, are the stratigraphically lowest prominently fossiliferous ones in the district, Stewart (1970) cites a number of examples of the presence of fossil material in the rocks stratigraphically below the Wood Canyon Formation in the region; thus negating the uniqueness of this influence to the Wood Canyon Formation alone.

Carbonate rocks occur throughout the Johnnie Formation, Stirling Quartzite, and lower units of the Wood Canyon Formation; so the carbonate rocks in the upper unit of the Wood Canyon Formation are not the oldest carbonate rocks exposed in the Johnnie district, negating the second speculation above. Further, Woodward (1972) notes that it is frequently only assumed that the Cambrian carbonate rocks in the other districts in the region are the lowest in their respective stratigraphic sections and Hewitt (1968) and Woodward (1972) speculate that unexposed older rocks could host significant hydrothermal mineralization.

Although this discussion suggests that the upper unit of the Wood Canyon Formation possesses no intrinsic qualities to make it stand apart as a host for hydrothermal mineralization, the general relations observed in the Johnnie and other districts reinforce, in my opinion, the conclusion that these lowest Cambrian carbonate rocks do, indeed, have a unique influence upon ore deposition.

Hypogene Geochemistry

Geochemistry of Wall-Rock Alteration

Inspection of and calculations from chemical and mineralogic data on rocks in the region presented by Stewart (1970) gives insight into some of the wall-rock reactions which occurred and therefrom gives insight into the character of the hydrothermal fluids in the Johnnie district. The compilation of selected data from Stewart (1970) gives an approximation of the average compositions of the rocks in the district. Only end member shale, quartzite, and dolomite, are considered.

Alteration produced the wall-rock alteration minerals or mineral assemblages: sericite-pyrite-silica of clastic rocks; chlorite, locally, in argillaceous rocks; sericite in dolomite; calcite in dolomite; and specularite-pyrite in some quartzites. Wall-rock alteration appears to have been the chemical modification of the original minerals in place, although some alteration products probably were remobilized from nearby and some additions from the hydrothermal fluid were required.

<u>Sericite-Pyrite-Silica</u> <u>Alteration</u>: The production of this suite of alteration products primarily involved the isochemical recrystallization of sedimentary muscovite and the sericitization of feldspar and chlorite. The reactants comprised approximately half of the total volume of the shale and the matrix of the quartzite. Half of the reactants in the shale were sedimentary muscovite; and most of the reactants in the quartzite were feldspars. Chlorite was important only locally.

Sericitization of potassium feldspar and plagioclase is a hydrolytic base leaching reaction which consumes hydrogen ions from the hydrothermal fluid to produce sericite and silica (Meyer and Hemley, 1967, p. 206-207). Calculations show that enough potassium was released from the potassium feldspar present to sericitize the average amount of plagioclase present in the affected rocks in the district; and the sericitization of the plagioclase released sodium and calcium into the hydrothermal fluid. The sericitization of chlorite (Meyer and Hemley, 1967, p. 207) proceeded similarly; addition of hydrogen, potassium, and aluminum from the hydrothermal fluid was required, and iron and magnesium were released from the altered minerals. When potassium feldspar, plagioclase, and chlorite in the molar ratio 3:1:3 which was present are sericitized, 0.5 moles of potassium and 1.0 mole of hydrogen are consumed and 0.2 moles, each, of sodium and calcium and 1.0 mole, each, of iron and magnesium are released. This assumes plagioclase of probable andesine composition and chlorite containing subequal amounts of iron and magnesium. The behavior of aluminum, which is added to wall rocks during the sericitization of chlorite, is ignored.

Similarly, sericitization of the primary chlorite in shale of the composition present could have released enough iron to convert 1.0 volume percent of the rock to pyrite, where sulfur was available; the iron given in the analyses of unaltered shale could have produced 4.5 volume percent of pyrite. Since pyrite commonly constitutes up to 10 volume percent of sericitized shale and since chlorite is not universally present, an outside source of iron was required. The situation with quartzite is comparable.

Chloritic Alteration: Secondary chlorite is in the rocks east of the Main and Northwest segments of the Grapevine fault system (see Hypogene Mineralogic Zonation), where the alteration minerals hematite, sericite, and calcite are present also. Chloritization is the product of the reaction of feldspar and probably muscovite with a hydroxyl-bearing hydrothermal fluid with the release of the hydrogen ion into the fluid (Meyer and Hemley, 1967, p. 207). The iron incorporated into chlorite could have been derived from the hematite which chlorite is seen to have replaced, and the magnesium could have been from the calcitization of dolomite (see Calcitization of Dolomite). The hematite probably was remobilized primary iron oxide.

Chlorite, where present, appears to have been the initial product of alteration of the wall rock by a hydrothermal fluid which was alkaline initially but whose pH eventually dropped into the acid range. With lowered pH, the primary minerals in the rocks and the earlier chlorite were sericitized.

Sericitization of Dolomite: There is no clear reason why dolomite, an agent expected to neutralize an acidic hydrothermal fluid, was pervasively sericitized and without the formation of pyrite, a mineral present in the sericitic alteration assemblage of shale. Acid in hydrothermal fluids dissolved dolomite to produce the HCO_3^- ion at the expense of the hydrogen ion to raise the pH into the alkaline field. Barnes and Czamanske (1967, p. 349) show that at 250°C, under geologically reasonable conditions, the pH in this system would then lie between 8 and 11. Sericite could not have been precipitated (Hemley and others, 1961; Hemley and Jones, 1964) if this pH prevailed to the below 200°C inferred to have prevailed during wall-rock alteration in the Johnnie district (see <u>Depth-Temperature Environment of Hydrothermal Activity</u>).

In some structural situations in the Johnnie district, dolomite can be construed to have been sericitized at the low pressure (exit) end of a throttle. Under such circumstances the pressure of a hydrothermal fluid released through a throttle drops, permitting vast expansion (adiabatic boiling) with an attendant temperature drop (Barton and Toulmin, 1961; Toulmin and Clark, 1967, p. 446-451). As the temperature falls, the stability field of sericite expands into higher pH ranges (Hemley and others, 1961). There is no evidence for such drastic pressure changes, but if this questionable construction had been the case in the Johnnie district, then the rise in pH also explains the inhibition of pyrite deposition; but this does not explain the lack of any iron oxide mineral (Barnes and Czamanske, 1967, p. 351).

The exit of the hydrothermal fluid through the throttle would have had additional possible effects. The temperature drop favored the solubility of dolomite (Holland, 1967, p. 404). Boiling acid volatiles escaped from the solution, raising its pH (Hemley and Jones, 1964, p. 565). The drop in total pressure increased the partial pressure of CO_2, increasing the solubility of dolomite and raising the pH.

At any rate, the sericitization of dolomite required the addition of all of the constituents of sericite from the hydrothermal fluid and the release of considerable HCO_3^-, calcium, and magnesium.

Calcitization of Dolomite: This involved the selective removal of magnesium carbonate from dolomite or replacement of magnesium by added calcium and involved the recrystallization of the resultant product as calcite. Calcitization in the Johnnie district could have proceeded as a combination of the incongruent solution of dolomite and dedolomitization, adding magnesium to the hydrothermal fluid and depleting it of calcium.

Incongruent solution (Krauskopf, 1967, p. 87) results from the greater solubility of magnesium carbonate than calcium carbonate, particularly at lower temperatures. Where dissolution is incomplete, magnesium is added to the hydrothermal fluid, lowering the calcium-magnesium ratio.

Dedolomitization (Holland, 1967, p. 413) is the replacement of magnesium in dolomite by calcium. As with incongruent solution, the reaction is favored at lower temperatures (Holland, 1967, p. 412) and the reaction also is favored where the calcium-magnesium ratio in the hydrothermal fluid is high initially.

Specularite Alteration in the Stirling Quartzite: Specularite was deposited where abundant iron was available for reaction with the hydrothermal fluids ascending into the B member of the Stirling Quartzite (see Malachite Deposits, Origin). The oxygen fugacity of the fluid was increased in this stratigraphic interval containing abundant apparently syngenetic iron oxide, an oxidizing agent. This should have favored the recrystallization of iron oxide rather than the precipitation of iron sulfide (Barnes and Czamanske, 1967, p. 351).

In two areas (see Hypogene Mineralogic Zonation) cubic pyrite was precipitated with the specularite. This pyrite probably was a wall-rock alteration product; but, alternatively, this pyrite may have been deposited as an ore mineral in place of chalcopyrite in hypogene ore mineralogic zones distally from the ones characterized by chalcopyrite.

Seemingly, the pH of the hydrothermal fluid should have been raised during its passage through the upper unit of the Johnnie Formation, below, through neutralization by reaction with the numerous dolomitic horizons near the top of the Johnnie Formation. However, conditions in the B member of the Stirling Quartzite must have remained acidic because sericite developed there in conjunction with the specularite and because the precipitation of the specularite-pyrite assemblage present would have been favored in an acidic environment (Barnes and Czamanske, 1967, p. 351).

Composition of Hydrothermal Fluid

In a net sense, hydrothermal alteration in the Johnnie district consumed hydrogen, potassium, aluminum, iron, and sulfur from the hydrothermal fluid and added CO_2, sodium, magnesium, and calcium to the fluid. The consumption of calcium during calcitization of dolomite was a local event, offset by additions of calcium from sericitization reactions. Quartz veining, which followed wall-rock alteration, removed vast quantities of silica from the fluid.

During metallization, gold, copper, lead, iron, and sulfur were provided from the hydrothermal fluid--assuming a classic magmatic-hydrothermal origin for the ore deposits in the Johnnie district. Alternatively, some or all of the metals were leached from surrounding rocks.

Gold may have been present as low grade placer deposits in some of the quartzite units; however, my sampling found none in the Zabriskie Quartzite in an unmineralized part of the district. As suggested in Malachite Deposits, Origin, syngenetic copper sulfide minerals may be present in certain stratigraphic intervals. The presence of conspicuous galena deposits in the upper unit of the Johnnie Formation (see Galena-Calcite-Quartz Veins) suggests that syngenetic galena was there, but their presence actually is a consequence of mineralogic zonation. Although iron is, in part, locally derived (see Malachite Deposits, Origin and Sericite-Pyrite-Silica Alteration) an outside source was required for much of the iron deposited in pyrite and chalcopyrite as hydrothermal alteration products and vein minerals.

The original sulfur content of the rocks in the district is uncertain, but the widespread occurrence of inconspicuous molds of pyrite suggests that syngenetic pyrite could have been sufficiently abundant to have provided all of the sulfur (and, therefore, iron) required for hydrothermal alteration and metallization. However, it is plausible that the sulfur was derived from the progressive oxidation of some sulfide species, such as H_2S, in the hydrothermal fluid. The latter explanation also provides a convenient source for the hydrogen ion required for sericitization during the wall-rock alteration stage of hydrothermal activity.

Nothing in this study precludes the possibility of the volumetrically great aqueous phase, itself, of the hydrothermal fluid having originated from interstitial, connate brines in the miogeosynclinal rocks in or below the district. Carrying this hypothesis one step further, the silica, deposited as quartz veins, in the hydrothermal fluid may have originated by lateral secretion from rocks at the same source as the connate hydrothermal fluid or from other rocks transgressed by hydrothermal fluids from any (for example, magmatic or connate) source.

One objection to this hypothesis is that, by the time of hydrothermal activity, the rocks of the district may have been so tectonically mature that they were too thoroughly dewatered to have provided the requisite amount of connate water.

Geochemical Leakage Dispersion Patterns

Some of the products of wall-rock alteration--CO_2, sodium, calcium, and magnesium--freed into the hydrothermal fluid may have produced geochemical leakage dispersion patterns in post-Zabriskie Quartzite rocks above the Johnnie district and above other mineralized areas; so may have surplus reactants in the spent fluids. These leakage patterns may be manifested as: metallic minerals in veins, wall rocks, or country rocks; potassium and aluminum enrichment, as sericite, in carbonate wall rocks or country rocks; variations in potassium, sodium, and calcium ratios in feldspar in clastic wall rocks or country rocks; or variations in calcium-magnesium ratios in carbonate wall rocks or country rocks or in carbonate minerals in veins. The calcite veins northwest of the Johnnie district (fig. 2) may be a manifestation of geochemical leakage (see Vertical Zonation).

Depth-Temperature Environment of Hydrothermal Activity

This section reasons that hydrothermal activity occurred at estimated temperatures of 150°C, later having risen to 250°C, and pressures of approximately 1,000 atm. The depth of burial corresponding to this is 15,000 ft (4,600 m).

For a working figure, at a depth of burial of 30,000 ft (9,000 m), the inferred amount of cover over the Johnnie district before erosion (see Regional Geology), the prevailing temperature along an average geothermal gradient of 25°C/km was approximately 255°C and the pressure with an assumed specific weight of overburden of 144 lb per cu ft (sp gr 2.3) was approximately 2,000 atm. However, two factors modify these calculated temperatures and pressures: (1) an unknown amount of erosion had occurred prior to hydrothermal activity, reducing the figure for pressure below 2,000 atm; and (2) the failure of the potassium-argon clock in sedimentary muscovite to reset during sericitization suggests temperatures less than 200°C prevailed (Miles L. Silberman, pers. comm., 1974) during the wall-rock alteration stage of hydrothermal activity (see Age of Hydrothermal Ore Deposits). Nonetheless, the temperatures probably rose to within the mesothermal range (see below) by the time of metallization; the temperature could have exceeded that along the average geothermal gradient if it is correctly assumed that heat was introduced from the lower crust or upper mantle (see Localization of District).

The texture of the quartz veins and, to some extent, the hydrothermal minerals present suggest deposition above 200°C under mesothermal conditions (200°-350°C, 400-1,600 atm, Ridge, 1968, p. 1817); so I assume a temperature around 250°C. The texture of the quartz is uniform throughout the entire vertical interval exposed in the district; this interval is approximately 1,500 ft (460 m) in the area west of the Main and Northwest segments of the Grapevine fault system and may have been in excess of 10,000 ft (3,000 m) across the fault system before Basin-and-Range faulting. The quartz exhibits no epithermal features, such as symmetrical, crustiform banding; the vug structures which are present (see Quartz) can be expected in mesothermal ore; and there is none of the pervasive alteration or mineralization in the walls of the veins characteristic of epithermal mineralization.

Age of Hydrothermal Ore Deposits

It is likely that hydrothermal activity in the Johnnie district occurred as a single episode confined to within a reasonably limited time interval. I conclude in this section that the episode and hydrothermal deposits probably are of early or middle Tertiary (Paleocene to early Miocene) age.

A single episode of activity is suggested by the similarity of appearance of the vein quartz throughout the district, the uniformity of texture--including lack of banding--of the veins, and the absence of cross cutting of quartz veins of apparently different ages.

The quartz veins can be no older than latest (75 m.y., Fleck, 1967) Sevier orogeny age (Late Cretaceous; see Regional Geology), because some veins are hosted by latest or post-Sevier orogeny structures such as the Congress low-angle normal fault (see Congress Fault System and Related Structures). Other quartz veins, possibly coeval with those in the Johnnie district, in the Spring Mountains, such as those in the Emerald district (figs. 1 and 2), are localized in Sevier orogeny thrust faults and structures ancillary thereto, also.

Parts of the Grapevine fault system guided migrating hydrothermal fluids. For example, the Grapevine fault hosts large horses of quartz-replaced quartzite breccia and gouge of the Wood Canyon Formation near Grapevine Springs and at the north tip of the megabreccia deposit. The fault also is the locus of some pervasive, jasperoidal silification in dolomite of the Johnnie Formation near the latter site. The Grapevine fault system (see Grapevine and Related Fault Systems) is an old feature, and hydrothermal activity preceded the major displacement on the fault system which prepared the topographic situation which caused the post-quartz transport of the late Pliocene to middle Pleistocene megabreccia deposit to its present position (see Megabreccia Unit of Older Fanglomerate and Geomorphology). Thus, the quartz veins in the district can be no younger than late Miocene (7-15 m.y., Fleck, 1967) (see Regional Geology), the age of major Basin-and-Range faulting. This span can be extended to 17 m.y. (Fleck, 1967) if the major movement began concurrently with movement along the Las Vegas Valley shear zone (see Regional Geology).

An age nearer to 15-17 m.y. rather than to 7 m.y. is indicated by:

(1) the observation that quartz veining preceded significant erosion of the rocks above the district, because the texture of the quartz suggests that the veins were emplaced at a moderate (not at a near surface) depth (see Depth-Temperature Environment of Hydrothermal Activity); and (2) Basin-and-Range faulting seems to have put the chlorite alteration zone into anomalous juxtaposition with rocks across the Grapevine fault system (see Hypogene Mineralogic Zonation).

There is no good correlation between the genesis of the hydrothermal ore deposits in the Johnnie district and either of the two episodes of igneous activity in the region (see Regional Geology). It is unlikely that ore deposition accompanied the late Mesozoic (pre 82 m.y.) synorogenic episode of intrusion along the orogenic belt; because the deposits in the Johnnie district are postorogenic, and because the deposits of the Johnnie district do not have any evident magmatic affiliations (see Localization of District). Also, it is unlikely that ore deposition in the Johnnie district accompanied the mid-Tertiary intrusive activity, because the latter was restricted to the areas beyond the flanks of the site of the orogenic belt, within which the Johnnie district is situated.

Potassium-argon age dating of hydrothermal sericite in the Johnnie district was attempted but was inconclusive because the temperatures of ore deposition were apparently too low (less than 200°C) to reset the K/Ar radiometric clock (Miles L. Silberman, pers. comm., 1974). Younger Precambrian and Pennsylvanian ages were obtained, and these are incompatible with the structural evidence. This is under investigation by the U. S. Geological Survey. The discrepancy will be explained and a revised age given in later publications.

The dispositions of the average quartz veins, principal structures, and the major longitudinal structure can be compared, inconclusively, with a simple model inspired by Rehrig (1971) and Rehrig and Heidrick (1972) in an effort to determine the age of mineralization. The model involves the localization of mineralization in transverse structures dilatant during orogenic (Late Cretaceous) transverse compression or localization in longitudinal structures dilatant during later (Tertiary) conditions of tension across the site of the orogenic belt. The quartz veins and principal mineralized structures in the district are oriented along a trend which bisects the transverse and longitudinal trends and, so, the veins and structures cannot be compared with the model. The inferred major longitudinal structure is oriented favorably with the Tertiary tensional trend but not enough is known about the exact orientation of the structure or its genesis to permit making any comparisons safely.

In conclusion, the deposits in the Johnnie district could be from Late Cretaceous to early Miocene (75-15 m.y.) in age. I favor a geologically young age, because post-ore offsetting of the veins usually is not excessive and because their host fractures became dilatant by very subtle movements along the principal mineralized structures and major longitudinal structures, which do not appear to be related to the Sevier orogeny, which left a very prominent structural signature. Therefore an early (say, Paleocene) or middle Tertiary age is assigned here to the hydrothermal ore deposits in the district.

Summary of Ore Genesis

Hydrothermal activity and attendant ore deposition occurred in the Johnnie district during early to middle Tertiary times after the structural preparation of the area by the Late Cretaceous Sevier orogeny and subsequent tectonism of lesser magnitude. The subsequent tectonism included inception of movement on the ancestral Grapevine fault system. After hydrothermal activity, the district was disturbed by Basin-and-Range normal faulting which offset some veins and thoroughly fractured the quartz in all of them. Although some erosion probably preceded hydrothermal activity, the remainder, most after Basin-and-Range faulting, exposed the district at its present level in a series of geomorphologic events culminating in the deposition of gold placers.

The sources of the heat, water, and contained elements in the hydrothermal system were not established by this study; but they probably came from some combination of remote and local sources.

The hydrothermal fluid may have been derived from a remote source in the upper mantle or lower crust or from a pluton (considered remote for discussion purposes) emplaced at depth below the district. Alternatively, the source could have been local: connate water expelled from the miogeosynclinal rocks in or below the district during or after diagenesis; groundwater transported downward along the ancestral Grapevine fault system; or water in wet surficial or near surface materials overridden by the upper plate of the Montgomery thrust.

The heat, only, of the hydrothermal system may have been derived from remote sources or the heat was generated locally by the geothermal gradient. In either case, hydrothermal convection of local fluids, if such was their source, would have resulted.

The hydrothermal fluid, with its metals and silica, could have originated from a remote source or, the metals and silica originated by the leaching action of a hydrothermal fluid, from any source, migrating through the miogeosynclinal rocks in or below the district. If the fluid was a connate brine it would have provided the halogens considered necessary to enter into the complexes required to transport some of the metals, or the halogens came from evaporite units, since removed by leaching, in the miogeosynclinal stratigraphic section.

The hydrothermal fluid probably ascended along a gravity-controlled pressure gradient within a tabular, vertical dilatant zone now manifest as the inferred major longitudinal structure and/or the principal mineralized structures. The fluid either entered this zone by migrating directly from a remote upper mantle, lower crust, or plutonic source or by migrating essentially laterally from a local source. Fluid from a remote source may have migrated directly along the major longitudinal structure, if it is the upward propagation of a basement lineament; or the fluids from a remote source migrated from depth along a poorly defined front until they were collected and canalized along the Montgomery thrust or Grapevine fault system to ultimately enter the major longitudinal structure or principal mineralized structures.

Whether the major longitudinal structure exists or not, once the hydrothermal fluid entered the mineralized area it was canalized at the base of the Stirling Quartzite, by the sole of the Congress low-angle normal fault, and possibly canalized by the Grapevine fault system and Montgomery thrust. From those sites, the fluid was diverted into the principal mineralized structures, which were the loci of fractures and faults made dilatant, by shear, under the influence of the local stress system in effect at the time. Finally in the fractures and faults, the hydrothermal fluids caused--sequentially--wall-rock alteration overlapped by quartz veining, and--after minor fracturing of the quartz--metallization.

A Fanciful Hypothesis for Ore Genesis

I am fascinated by the possibility that the hydrothermal ore deposits in the Johnnie district originated from syngenetic materials in the vicinity.

Concepts of ore genesis change with time. The once popular belief that all of the components of ore fluids originate from magmatic sources has been modified greatly, in the last decade, by studies demonstrating that the aqueous phase of the fluids can arise from a variety of nonmagmatic sources (for example, Gross, 1975; Guy, 1975; Skall, 1975). Another trend, which has roots in the same studies, points to examples (such as Carpenter and others, 1974; Gross, 1975) of the enrichment of metals in potentially hydrothermal subsurface waters in which the metals have been concentrated by solution of trace elements from country rock. As this trend takes conceptual form in terms of ore genesis, it will not be unreasonable to interpret the origins of deposits, such as those in the Johnnie district where there is no obvious magmatic source, in terms of lateral secretion.

Among other phenomena, the apparent affinity of some metals for certain stratigraphic horizons in the Johnnie district may be more than a series of coincidences; the affinities are galena in the upper unit of the Johnnie Formation, chalcopyrite in the B member of the Stirling Quartzite, gold in the Zabriskie Quartzite, and the affinity of concordant quartz stringer lodes for particular stratigraphic horizons. The lack of an obvious magmatic source for the hydrothermal system and lack of obvious relations between the ore deposits and deeper structures cause one to seek alternate interpretations for their origins.

Perhaps unimaginatively, this report explains these features on structural and lithologic grounds in light of currently prevailing concepts of ore genesis. However, if contradicting details of current concepts of ore genesis--which concepts are, in part, the outgrowth of contemporary geologic prejudices--are ignored, a case could be made for a syngenetic method of origin modified by lateral secretion processes.

This report demonstrates (see Composition of Hydrothermal Fluid and Summary of Ore Genesis) that, although not necessarily true, most of the constituents of the hydrothermal fluid could have been derived from local sources. Perhaps the profusion of quartz veinlets in the Johnnie Formation does not represent the paths of hydrothermal fluids ascending across a broad front, from below, before canalization into overlying structures to form quartz veins; perhaps the profuse veinlets are a system of collectors of hydrothermal fluids migrating laterally from adjacent rocks.

Then all that is required for a lateral secretion origin is a lateral temperature and pressure differential to initiate hydrothermal activity and concentrate it along the controlling structures. The requisite heat could have been generated by the geothermal gradient, and the necessary low pressure conditions would have prevailed along the controlling structures at the time of the dilatancy which permitted the ingress of hydrothermal fluids.

The following speculation, one of a number possible, is appealing because it unites a lot of observations. If, conveniently, the Johnnie district is above a narrow, longitudinal basin or basinal sequence of younger Precambrian sedimentary rocks below the base of the Paleozoic miogeosynclinal section, as suggested earlier in this report, several aspects of the district can be accounted for: (1) the ore deposits would lie within an elongate, northerly trending area; (2) ignoring the ramifications of thrust faulting for simplicity, the geothermal gradient might have been higher than in adjacent rocks, localizing hydrothermal convection and negating the requirement for a magmatic source; (3) high lithostatic pressures (Secor, 1962, 1965) would have localized hydraulic fracturing and the simultaneous collection of connate fluids within these fractures, thus making it, at least in part, a case of lateral secretion.

Additional isotope studies are necessary to: determine the source of the aqueous phase and other components of the hydrothermal fluid in the Johnnie district; and to define the temperature and age of ore deposition, which incidentally permits the deduction of depth and pressure. These studies, along with ones meant to prove or disprove the regionality--suggested here--of possible syngenetic copper deposits, would contribute greatly to assessing the validity of this fanciful hypothesis in the district and might give new direction to mineral exploration in the region.

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APPENDIX A

Partial section of the upper member of the Wood Canyon Formation, Zabriskie Quartzite, and lower part of the Carrara Formation in S 1/2 sec. 35, T. 17 S., R. 52 E. and NW 1/4 sec. 2, T. 18 S., R. 52 E.

(Composite section for Johnnie district from examination by S. W. Ivosevic Sept. 13, 1972 with observations interjected from elsewhere in district and utilizing some thicknesses (asterisked) for the Wood Canyon Formation and Zabriskie Quartzite from measured section by Hamil (1966) in the vicinity of the coordinates given above.)

Top of section measured of interest for influencing gold mineralization.

CARRARA FORMATION, incomplete:

		Feet
Unit 3	Dolomite, brown and gray, thin to medium bedded (Kelley, 1956), fossiliferous, locally crudely oolitic. Contains a few interbeds of argillaceous dolomite. Not measured. (Correlates with Hamil (1966) unit 6.)	
Unit 2	Shale, green, irregularly laminated to papery, with brown calcareous shale interbeds. (Correlates with Hamil (1966) unit 5.)	145 (44 m)
Unit 1	Carrara-Zabriskie contact zone. Shale, laminated, and thin bedded shaly quartzite. Most rocks maroon, but some, notably a thin papery shale bed near the top, are green. Contains a 5-10 ft (1.5-3 m) quartzite bed, similar to Zabriskie Quartzite, below, within 20 ft (6 m) of the base. Tectonically thins to as low as approximately 100 ft (30 m). (Correlates with Hamil (1966) units 1-4).	170 (52 m)
	Total thickness measured Carrara Formation	315 (96 m)

ZABRISKIE QUARTZITE:

Top placed at top of uppermost ledge forming thick quartzite bed.

		Feet
Unit 3	Quartzite, white to purple, weathers pink to purple, fine to medium grained, locally coarse grained, medium bedded to massive. Internally, the beds are finely laminated and frequently cross bedded. Deforms as a unit.	ca 120 (35 m)
Unit 2	Sandstone or shattered impure quartzite, white, medium grained. A single bed of said material infrequently crops out through the talus from the overlying unit; its ubiquity throughout the district cannot be ascertained. If universally present, it may serve as the plane of anisotropy, recognized in areas of marked deformation as a zone of tectonic readjustment, along which the upper unit of the formation glides over the lower. This may correlate with a bed of porcellaneous white quartzite occasionally observed in approximately this same stratigraphic position elsewhere in the district.	ca 2 (0.6 m)
Unit 1	Quartzite; same as unit 3 but parts along bedding during deformation. Becomes more thinly bedded toward base. *Scolithus* tubes abundant.	ca 120 (35 m)
	Total thickness Zabriskie Quartzite	240* (73 m)

Tectonically thins to as low as 115 ft (35 m).

(Correlates with Hamil (1966) unit 12, Wood Canyon Formation, Zabriskie Quartzite Member.)

WOOD CANYON FORMATION, UPPER MEMBER, incomplete:

Unit 3	Zabriskie-Wood Canyon contact zone. Quartzite, pale, and green to light brown siltstone in subequal amounts along with intergradational rock types; thin to medium bedded. Contains a few thin beds of dolomite and quartzitic dolomite throughout including a horizon through which dark brown weathering dolomite and dolomitic quartzite beds are common 20 ft (6 m) above the base of the unit. A 5 ft (1.5 m) bed of irregularly laminated green shale immediately underlies the Zabriskie Quartzite. *Scolithus* tubes common in the clastic rocks and abundant in the upper shale bed.	130* (40 m)

WOOD CANYON FORMATION, UPPER MEMBER, continued:		Feet
	Tectonically thins to as low as 45 ft (14 m).	
	(Correlates with Hamil (1966) unit 11b.)	
Unit 2c	Dolomite (75%) to quartzite (25%). Dolomite, gray, weathers rust brown, fine to medium crystalline. Quartzite, gray to brown, fine grained. Unit medium to thick bedded, becoming more thinly bedded and quartzitic upward, this being especially evident in the topmost 20 ft (6 m). Internally, 1-3 ft (0.3-1 m) thick, cross bedded layers of dolomite or quartzitic dolomite are succeeded upward by increasingly thinner, more quartzitic cross bedded layers of dolomitic quartzite to quartzite which locally exhibit fossil debris on weathered surfaces.	70-80 (21-24 m)
Unit 2b	Quartzite, gray, fine grained, medium bedded, finely laminated internally.	10-20 (3-6 m)
Unit 2a	Similar to unit 2c except that dolomite is oolitic and unit becomes more quartzitic downward, this being especially marked in the lowermost 20 ft (6 m). Lowermost beds are very dark weathering, locally developing a conspicuous coat of desert varnish.	90-100 (27-30)
	Total thickness unit 2	178* (54 m)
	Tectonically thins to as low as 130 ft (40 m).	
	(Correlates with Hamil (1966) unit 11a.)	
	Total thickness units 2 and 3	308* (94 m)
	Tectonically thins to as low as 175 ft (53 m) combined.	
Unit 1b	Siltstone to shaly quartzite, green, laminated to thin bedded. Trail or drag mark like features and mud cracks on parting surfaces. *Scolithus* tubes common.	140 (43 m)
Unit 1a	Quartzite, gray to pale green, fine grained, medium bedded with laminated internal structure. Local small mineral molds contain indigenous and/or fringing limonite (Blanchard, 1968), probably after pyrite.	100 (30 m)
	Total thickness unit 1	240 (73 m)
	(Correlates with Hamil (1966) unit 10.)	

WOOD CANYON FORMATION, UPPER MEMBER, continued:

	Feet
Total thickness measured Wood Canyon Formation	548 (167 m)

Section of Wood Canyon Formation continues downward for a total of 740 ft* (226 m) to the base of the upper member (Stewart, 1966) of the Wood Canyon Formation (correlates with Hamil (1966) units 10 and 11).

TOTAL THICKNESS	1,103 (336 m)

Base of section measured of interest for influencing gold mineralization.

APPENDIX B

History and Ownership of the Johnnie District

The following discussion of the history of the Johnnie district is derived from numerous private and published reports and some oral statements. Mining claim records in the Nye County Courthouse and U. S. Bureau of Land Management plats of patent surveys also form part of the basis for this discussion.

Inasmuch as the individual conclusions presented here are compilations of details, frequently conflicting, from more than one source, the citation of individual references for each conclusion is not feasible. The published references are:

> The Arrowhead (July, 1907), Los Angeles Mining Review (March 27, 1909), Labbe (1921, 1935, 1960), Las Vegas Age (April 9, 1912, September 11, 1926, March 12, 1927, December 31, 1927), Lincoln (1923), Nolan (1924, 1936), Smith and Vanderburg (1932), Vanderburg (1936), Mott (1937, 1940), Couch and Carpenter (1943), Kral (1951), Koschmann and Bergendahl (1968), and Paher (1970).

The ultimate source of many of the historical facts and production figures cited by the various authors is Charles H. Labbe, a mine owner and operator in the district from 1910 to 1964. In general, Labbe's work is accurate but should be accepted with reservation.

The following sections discuss the history of the most important mining properties in the district. The claims comprising these properties and the names of their present owners are tablulated in Appendix C.

Other properties have been operated in the district on exploratory or small-scale production bases. Approximately fifty unpatented mining claims were active in the district at the time of this study, excluding those of the Copper Giant property which embraces a rather large number of additional claims.

Discovery of District

Activity in the Johnnie district was initiated in 1890 by a small group of men from Indian Springs (fig. 1), 25 mi (40 km) to the northeast in Clark County, Nevada. They were in search of the Breyfogle discovery, one of the strikes which figures in the lore of lost mines in western America (for example, see Kral, 1951). The group included the Montgomery brothers, who had been mining in the Stirling district (figs. 1 and 2), approximately 10 mi (16 km) northeast of the Johnnie district. They were led by George "Monty" Montgomery and guided by Indian Johnnie, a local rogue who was acquainted with the countryside and who knew the location of the outcropping gold ore which they were to discover.

The group discovered the Chispa mine (Chispa: literal Spanish for "spark" or colloquialism for "ore with visible gold") which was to be renamed the Congress mine. This native gold in boldly outcropping quartz veins was known to Indians and prospectors, alike, previously; and Labbe (private report) notes that the district could have been worked in the 1860's had it not been so inaccessible.

This was originally named the Montgomery district, but gradually began to be called the Johnnie district around 1900 to 1910.

Congress Mine

M. B. Bartlett joined George Montgomery by 1891 and located the Chispa, California, and Freeland claims near the Congress mine. These were renamed the Congress, Phoenix, and Nevada claims, respectively, by 1900 and augmented with the Gold Dollar, Gold Eagle, and Gold King claims in 1900 to round out the Congress group of claims which, although unpatented, persists to the present time.

By about 1899, the owners had removed approximately 12,000 oz of gold from the older Congress shafts (pl. 8). In about 1899 a Utah group leased the Congress mine and extracted around 5,000 oz of gold from the ore shoot in the Mormon shaft. Parts of the surface installation were destroyed by fire and dynamite after about ten months of operation when either Bartlett refused to sell the mine to the operating group or when he cancelled their lease for non payment of royalties.

Sometime prior to 1905, Harry Ramsey, who by then was an associate of Bartlett, evicted a group of claim jumpers led by Phil Foote in an episode involving the use of firearms and the demise of Foote and a Mr. Gilespie (sic) (Labbe, 1960).

In 1905, Bartlett and Ramsey founded the Congress Mining Company and included the Congress group of claims therein. Subsequently Bartlett obtained ownership of the company, which--along with its assets--was passed through his sons Henry J. and George to his grandson Leo I. Bartlett, who now owns the company.

The Congress mine apparently has been worked intermittently by lessees since. The new Congress shaft evidently was sunk by 1939 and the Air shaft and Winze level developed since then (pls. 8 and 9). A new headframe was built between 1940 and 1950 (Kral, 1951) but this was destroyed around 1969.

In the first year of operation, the ore from near surface workings at the Chispa mine was hauled to the Horseshutem Springs (pl. 1) for treatment in a Huntington mill and in a Kendall one-stamp mill (Labbe, 1960). A ten-stamp mill was then erected at the mine, using water piped from the springs (Labbe, 1960), but this mill was destroyed in the 1899 explosion and fire. At least some of the ore from the Congress mine must have been treated at the Minnie Mae mill site (pl. 1), near the town of Johnnie in the NE 1/4 sec. 1, T. 18 S., R. 52 E. In later years, additional ore must have gone to a small mill site (fig. 2) in the Amargosa Desert in the S 1/2 NE 1/4 sec. 24, T. 17 S., R. 51 E., where an arrastra and the foundation of an Ellis mill remain.

Johnnie Mine

The Johnnie mine was located in 1894 by apparently the same Utah group (Los Angeles Mining Review, March 27, 1909) which leased the Congress mine in approximately 1899. They apparently worked the surface stopes southwest of the Johnnie shaft (pls. 4 and 5) possibly down to the present Second level, hauled selected ore directly to Salt Lake City, Utah (Los Angeles Mining Review, March 27, 1909), and milled other ore on the site with the Huntington mill (Labbe, 1960) which previously had been set up at the Horshutem Springs.

The Utah group's involvement with the Johnnie mine apparently was supplanted by that of the Johnnie Consolidated Gold Mining Company, probably at the time in 1903 that Carl Anderson located the Johnnie

Consolidated (pl. 2) and Tiger Consolidated (fig. 27) groups of claims. The Johnnie Consolidated Gold Mining Company patented both groups in 1905, developed the Johnnie shaft to the Seventh level (pl. 5), and installed a Nissen ten-stamp mill and 85 tpd amalgamation plant nearby in 1907 (The Arrowhead, July, 1907) or 1908 to replace the older Huntington mill and companion Chilean mill (Labbe, 1960) which they had been using earlier. During patent proceedings, the company also took title to the Minnie Mae mill site and to the April Fool mill site (pl. 1), near Horseshutem Springs in secs. 27 and 34, T. 17 S., R. 53 E.

The Johnnie Consolidated Gold Mining Company changed ownership at least once (The Arrowhead, July, 1907) and was taken over by the Johnnie Mining and Milling Company in 1909 (Los Angeles Mining Review, March 27, 1909). They operated the mine until 1915 when, the mine's most significant output having been achieved, they apparently sold out to O. T. Johnson. By then the mine had produced as much as a reported $3 million.

By some time between 1910 and 1925, the owners of the Johnnie Consolidated group of claims located the adjacent Battery, Buldosa, Flagstaff, Omaha, Oversight, Queen, and Teddy's Terrors nos. 2 and 3 claims. The Protection and Butterfly nos. 1, 2, and 3 claims were added later. None of the latter two groups were ever patented. (See pl. 2.)

Johnson apparently operated the Johnnie mine until 1923, extending the shaft to the 11th level and connecting the Second level with the Overfield mine and to some workings on the Johnnie property which were near the Overfield mine. Johnson apparently leased the mine to various operators who performed intermittent pocket mining until 1941 when about simultaneously Johnson died and War Production Board Order L-208, which terminated much U. S. gold mining, was issued (Labbe, 1960; Paher, 1970).

Apparently at this time, the Congress Mining Company acquired the Johnnie and Tiger consolidated groups of claims and the adjacent claims and appurtenant mill sites from the Johnson estate. The property has been leased to various small operators to present. Leo I. Bartlett is the current owner of the properties.

Overfield Mine

The Crown Point and Globe claims (pl. 2) were located by W. W. Booth and F. C. MacNeil (Labbe, 1960) in 1892 or 1893. The property was acquired by Ed "Happy Hunch" Overfield (the nickname was earned previously in the Goldfield district, Esmeralda County, Nevada), who organized the Crown Point-Globe mine under the Crown Point-Globe Mining Company. These were changed subsequently to the Overfield mine and Overfield Mining Company, respectively. The Overfield Mine passed on to a son of the founder, Charles E. Overfield, who is the present owner.

These two claims were patented as the Crown Point Consolidated Mine in 1911 along with the Mono claim (fig. 27), which was located near the Grapevine Springs in 1906.

Ed Overfield encountered high grade ore on the Globe claim in 1908 after fruitless exploration, apparently on the Globe claim, in that same year (Labbe, 1960). Production from here ceased in 1909 when the boundary of the property was reached and a lease could not be obtained from the Johnnie Consolidated Gold Mining Company to extend the workings (Labbe, 1960); but, during this time $200,000 worth of ore was mined (Labbe, private report, ca. 1960). Although substantially all of the ultimate development of this mine was accomplished during this period (Kral, 1951), a small amount of production has been obtained intermittently from the property up to the present.

The Kendall one-stamp mill was moved (Labbe, 1960) from Horshutem Springs to the Crown Point-Globe mine in 1909 (Kral, 1951) and remained in use there at least until 1919 (Labbe, 1960). Subsequently, this was replaced with an Ellis mill (Kral, 1951) which remains on the property at present.

Labbe Mine

The early history of the Labbe Mining property (pl. 2) is unclear. The original claims, Bluebell, Golden Eagle, Primero, Howitzer, and Annam, apparently were located in 1902 and 1903. These were sold by the Bullfrog-Johnnie Mining Company to Charles H. Labbe in 1910. Labbe amended and renamed the claims, respectively, the Broadway, Doris A. L., Westend, Hillside, and Digmore in 1926 and patented all but the Digmore in 1928. Subsequently, he added the Marge and Reef claims to the group. George E. Warner acquired the property from Labbe in 1964 and added the David and Dyke claims to the group. (See pl. 2.)

There is no record of the history of the development of the Labbe mining property. The 1928 plat of the patent survey indicates that the Broadway, Doris, and Bluebell mines were in existence by that time. The open cut on the Westend claim is a relatively recent development.

Sometime between 1923 (Lincoln, 1923) and 1940 (Mott, 1940) a ten-stamp mill, which is operable presently, was installed on the property. Lincoln (1923) mentions a two-stamp mill installed and operated by a

Eureka-Johnnie concern in 1917-1919, and this may be another mill which is present also on the Labbe property. It is not certain with which property the latter mill was associated at the time.

Placer Mining

Placer mining for gold yielded a relatively minor amount of the total production from the Johnnie district (see Production).

Although Vanderburg (1936) reports that Mormon operators discovered placer gold at an early date, the date of discovery is generally taken as being in the early part of 1921 (Las Vegas Age, April 9, 1921; Labbe, 1921). The first discoveries apparently were made by Robert Wedekind and Frank Buol (sic) near the Congress mine (Las Vegas Age, April 9, 1921) and by Walter Dryer near the Johnnie mine (Vanderburg, 1936).

Vanderburg (1936) notes that a short boom followed this discovery. Intermittent placer activity is reported in the district through 1951 (Nolan, 1924; Las Vegas Age, September 11, 1926, December 31, 1927; Vanderburg, 1936; Kral, 1951; Paher, 1970). The Matt Kusick operation (reported, in part, by Vanderburg (1936) and Kral (1951)) is noteworthy in that it apparently supported a small population of indigent lessees during the Depression. In general, most of the earlier placer mining took place in the vicinity of the Congress mine; and the later, near the Johnnie mine.

The lack of water near the placer workings prompted the development of an air jig or dry washer to extract placer gold (Labbe, 1921), which jig was somewhat unusual for its times and which permitted 75-90 percent recovery of gold values. Some of the miners transported gravel to nearby springs to be washed in rocker boxes (Smith and Vanderburg, 1932). Where the gravels exceeded 8 ft (24 m) in depth, mining was by shaft and drift methods (Nolan, 1924). Later operators used heavier equipment and piped water in from the springs (Kral, 1951).

The mill tailings below the Congress and Johnnie mines, which could be considered a placer resource, were partially washed away during a storm in July, 1956.

Silver-Lead Properties

Silver and lead production from galena-calcite-quartz veins (see Galena-Calcite-Quartz Veins) in the Johnnie Formation along the west margin of the district, most apparently during the early decades of this century, is reported by Lincoln (1923), Nolan (1924), The *Las Vegas Age* (September 11, 1926), and Kral (1951). The ore was upgraded either by hand sorting (Nolan, 1924) or by milling (*Las Vegas Age*, September 11, 1926).

Culture

Paher (1970) chronicles the close association between the volume of mining activity and history of population in the Johnnie district. He notes that post offices with varying life spans were opened there in 1891, 1905, and in the late 1930's.

The Johnnie townsite was inhabited from 1905 until the 1930's, reaching its heyday in 1907 (Labbe, 1960). The populace then shifted to a camp at the Johnnie mine, which camp was inhabited until 1957.

APPENDIX C

Significant Properties in the Johnnie District

(Listed for being patented, appurtenant to patented claims, or having yielded significant production)

Congress mine: Congress Mining Company (Leo I. Bartlett, Secretary-Treasurer), owner.

Congress group of unpatented claims: Congress, Gold Dollar, Gold Eagle, Gold King, Nevada, Phoenix

Johnnie Mine property: Leo I. Bartlett, owner.

Patented claims

Johnnie Consolidated group (pl. 2): April Fool, First Chance, Fraction, Fraction No. 2, Johnnie, Last Chance, Los Angeles, Minnie Mae, Teddys, Teddys Terrors

Tiger Consolidated group (fig. 27): Tiger, Chas. Swab

Unpatented claims (pl. 2): Battery, Buldosa, Butterfly nos. 1, 2, and 3, Oversight, Protection, Queen, Teddys Terrors nos. 2 and 3

Patented mill sites (pl. 1): April Fool, Minnie Mae

Labbe Mine Property: George E. Warner, owner.

Patented claims (pl. 2): Broadway, Doris A. L., Hillside, Westend

Unpatented claims (pl. 2): David, Digmore, Dyke, Marge, Reef

This property includes patented ground on springs east of Horshutem Springs (pl. 1)

Overfield Mine property: Charles E. Overfield, owner.

Patented claims

Crown Point Consolidated mine (pl. 2): Crown Point, Globe

Additional claim: Mono (fig. 27)

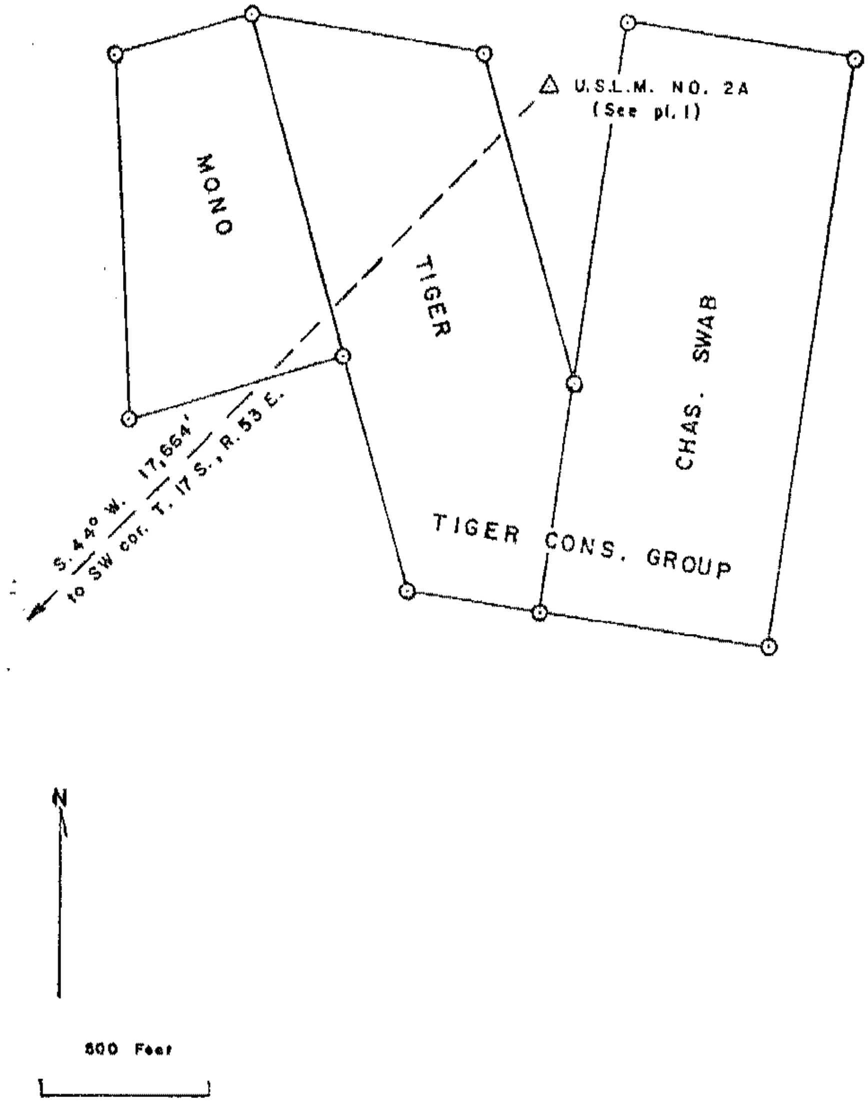


Figure 27. Plat of patented claims in the vicinity of Grapevine Springs, SW1/4 sec. 21, T. 17 S., R. 53 E. Redrawn from U. S. Bureau of Land Management plats of mineral surveys 2182-A and 3800. For approximate location of U. S. L. M. No. 2A, see plate 1.

APPENDIX D

Geology of the Copper Giant Property

Supergene malachite deposits derived from apparent concordant quartz-poor lodes of chalcopyrite which are associated with specularite veins are present at the Copper Giant property in secs. 20 and 29, T. 18 S., R. 52 E. (fig. 2).

The property is situated approximately 1 mi (1.5 km) north of the concealed trace of the Montgomery thrust fault in northeastward dipping rocks of the lower Stirling Quartzite. Approximately east-trending specularite-quartz veins up to 3 ft (1 m) thick are opened up by old shafts there. The specularite occurs in masses up to fist sized of large flakes which weather a patent leather black.

Malachite is deposited on white quartzite which overlies a dark shaly unit. This is exposed in several cuts and recognized in float which occur together over an area of somewhat restricted size.

APPENDIX E

Statistical Analysis of Quartz Veins

There are two broad morphologic groups of quartz veins in the Johnnie district: (1) high-angle quartz veins and veins ancillary thereto; and (2) concordant quartz veins and concordant quartz stringer lodes (see Morphology of Quartz-Bearing Structures).

Inspection of figures 9, 10, 21, and 29 shows that high-angle quartz veins can occur in any orientation for which pre-existent fractures and faults are present. However, a minor group of veins, which strikes parallel to the northwest-striking conjugate fracture set and transverse fault group, dips oppositely.

Veins are most common throughout an angular range (fig. 21) bridging the transverse and longitudinal fault sets and involving about half of each (fig. 10). This area, spanning approximately 70° of arc, relates to, and is presumably derived from, one-half of the northeast-trending conjugate fracture set and the adjacent extension fracture group. All dips in this area are north and are progressively steeper in the more easterly trending veins.

Figure 28 (summarized in fig. 21) shows the attitudes of veins localized in bedding-related structures in the district. An east-trending train of maxima reflect the disposition of veins along bedding which are bent from the average 40° easterly dip along the flanks of folds. The subvertical maxima at the east and west positions of figure 28 reflect veins localized in high-angle reverse faults as well as those localized along vertical bedding. The west-trending group of

Figure 28. Contour diagram of lower hemisphere equal area (Schmidt) net plot of 159 poles to veins localized in bedding-related structures. Contoured at 1, 3, 5, 7, and 9 percent intervals with local supplemental contours at 2 percent.

vertical veins are in shears which are usually localized along the axial areas of folds, which shears developed in an ancillary fashion to dislocation along the axial areas.

The veins localized in bedding-related structures cannot be related to those in high-angle structures in a simple conjugate or lower order manner. This is because the truly dominant bedding-related feature, which includes discordant fractures within quartz stringer lodes, was not isolated during this study. Assuming a single episode of quartz-veining, bedding-related veins occupy dilatant zones which developed under the same stress field in effect at the same time the rest of the veins were forming.

Stereographic Analysis of Veins Localized in High-Angle and Related Structures

The dispositions of the poles to quartz veins localized in high-angle and related structures plotted in figure 29 can be related to four small circles and to four inclined and three vertical great circles. This indicates that the poles are dispersed along the portion of the surface of a double cone the axis of which is horizontal and strikes north-northwest. The mutual apex of the double cone is pierced by a vertical line and the figure is bisected by a vertical plane (fig. 30).

The maxima of the north-dipping, statistically most prominent veins probably lie along small circle 1 (SC-1) in the southeast quadrant of figure 29; because the maxima are related somewhat less clearly along SC-2, the alternative possibility. The maxima of the subordinate, southward dipping group, in the northwest quadrant, define SC-3, which

Figure 29. Contour diagram of lower hemisphere equal area (Schmidt) net plot of 240 veins localized in high-angle and related structures. Contoured at 1 percent intervals to 5 percent. Explanation: crosses-maxima; SC-small circles; GC-great circles; B-locus of bisectors of GC-5, GC-6, and GC-7. The traces of GC-2 and GC-6 are similar and are not shown separately.

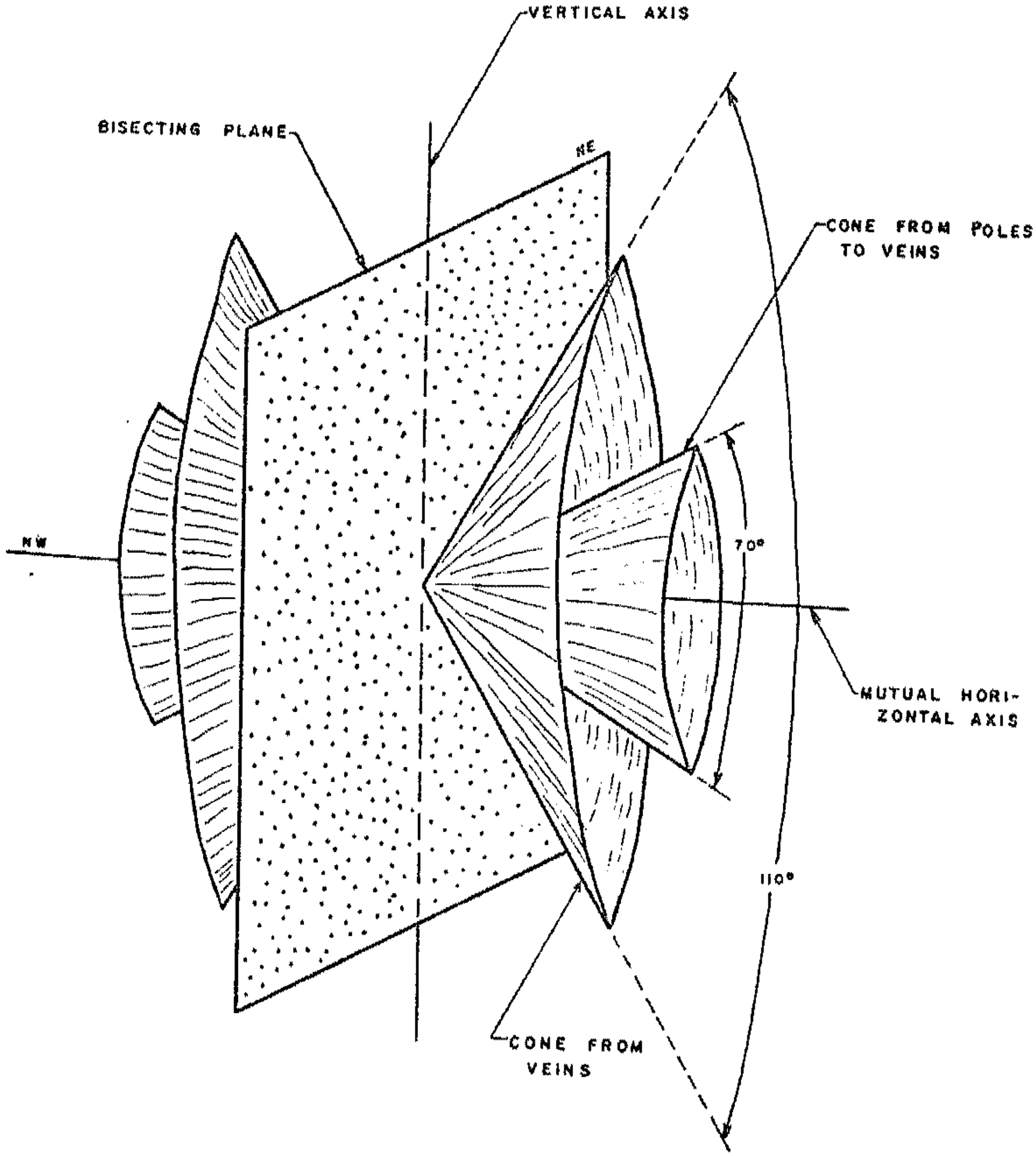


Figure 30. Diagram summarizing statistical elements of high-angle veins.

is approximately 180° away from its mirror image, SC-1.

These small circles are related by three inclined great circles--GC-1, GC-2, and GC-3--which approximately follow lines of maxima dispersed partially across the hemisphere. These great circles are related statistically in two ways. First, the points of intersection of the three great circles with the elongate minimum zone lie along another great circle, GC-4, the strike of which is approximately parallel to that of the average high-angle vein in the district (fig. 21) but which great circle is inclined in the opposite direction. The strike of GC-4 also is approximately parallel to those of SC-1 and SC-3. Second, the strike of another small circle, SC-4, which is defined by an elongate minimum zone, is approximately parallel to the strike of SC-2.

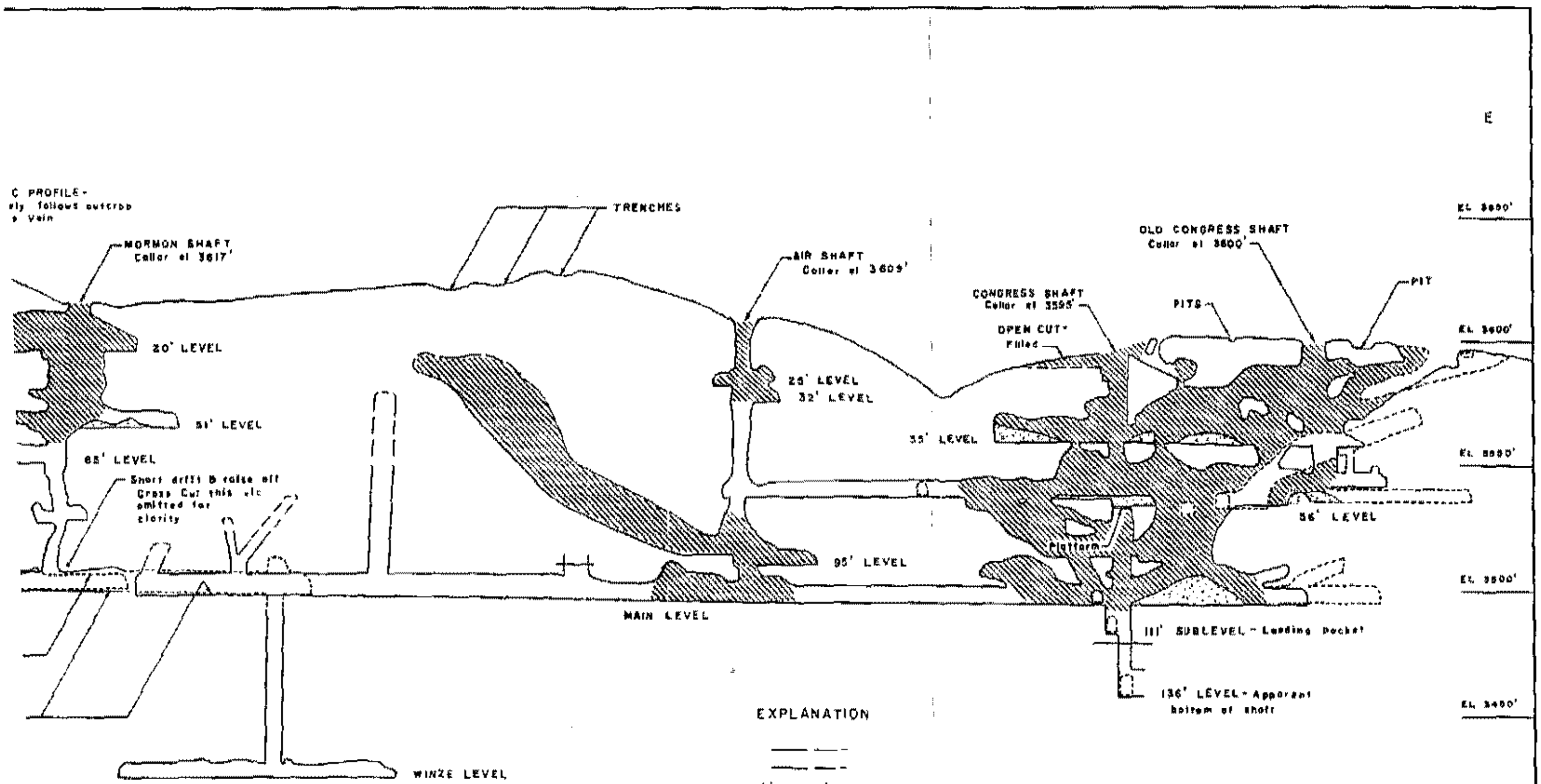
The apical angle of the cone along which SC-4 lies is so obtuse that the figure nearly becomes a plane, and the attitude of this plane is similar to that represented by GC-4. SC-4 does not lie along the same conical surface as does SC-3.

It is seen from mapping that the major veins within any one area of the district are parallel. (For example, see the map of veins between the Johnnie mine and the Westend open cut, pl. 2). Mapping also shows that subordinate, oppositely dipping veins do occur in and near the walls of major veins (for example, mapping of stopes at surface south of the Johnnie shaft, pl. 6, and fig. 24). However, the actual relation between any one major vein and its statistical analog in the opposite quadrant is unclear; although, within any one area, it is intuitively evident that oppositely dipping veins should occur together which are statistically related along either the inclined great circles, GC-1, GC-2, or GC-3, or along the vertical great circles, GC-5, GC-6, or GC-7. Studies of equal area net plots of poles to veins within selected individual areas do not give enough data to clarify this intuitive observation.

Maxima in opposite quadrants are related by vertical great circles, GC-5, GC-6, and GC-7, the mutual intersection of which at the origin of the hemisphere is a vertical line, the bisectors, B in figure 29, of opposite pairs of maxima are the poles of similar subvertical lines. In general terms, all four small circles are approximately coaxial; their common horizontal axis parallels the strikes of GC-1, GC-2, and GC-3 and is approximately perpendicular to GC-4. The vertical line, or axis, which intersects this horizontal axis at the origin is suggested by: the subvertical dip azimuth of GC-4; by the subvertical dip azimuth SC-4 would have if it were generated by a plane; by the mutual intersection of GC-5, GC-6, and GC-7 at the origin; and by the near verticality of the bisectors of the maxima of the latter.

SC-1 and SC-3 define cones having apical angles of 72° and 70°, respectively. From manipulation of the stereographic net, it is seen that the poles by which these cones are recognized are generated from planes (veins) tangential to a coaxial double cone having apical angles of approximately 110°.

This interpretation of the statistical analysis may be more simplified than the actual situation. An orthagonal relation about a level line is assumed for the two axes recognized; although of minor significance, this is contrary to any suggestion that the vertical axis designated herein is actually an inclined one. A third axis, mutually perpendicular to the two, cannot be inferred, because there is no evidence to support the existence of such tetragonal symmetry. Note that a third axis, mutually perpendicular to SC-2 and SC-4, may exist and that these three statistical elements may operate independently of the other elements; possibly they represent activity during a separate structural event. This assumes that SC-2 actually exists and is not the product of my overinterpretation.



EXPLANATION

Mine workings

Inaccessible shaft or raise

Cross cut projected from drift away from observer, toward observer

Cross cut projected through drift

Filled workings

Stope

SCALE 50 0 50 FEET

Vertical longitudinal projection to a plane N. 76° E. Looks north.

For location of section see Composite Plan Projection of Workings

JOHNNIE DISTRICT, NYE CO., NEV.

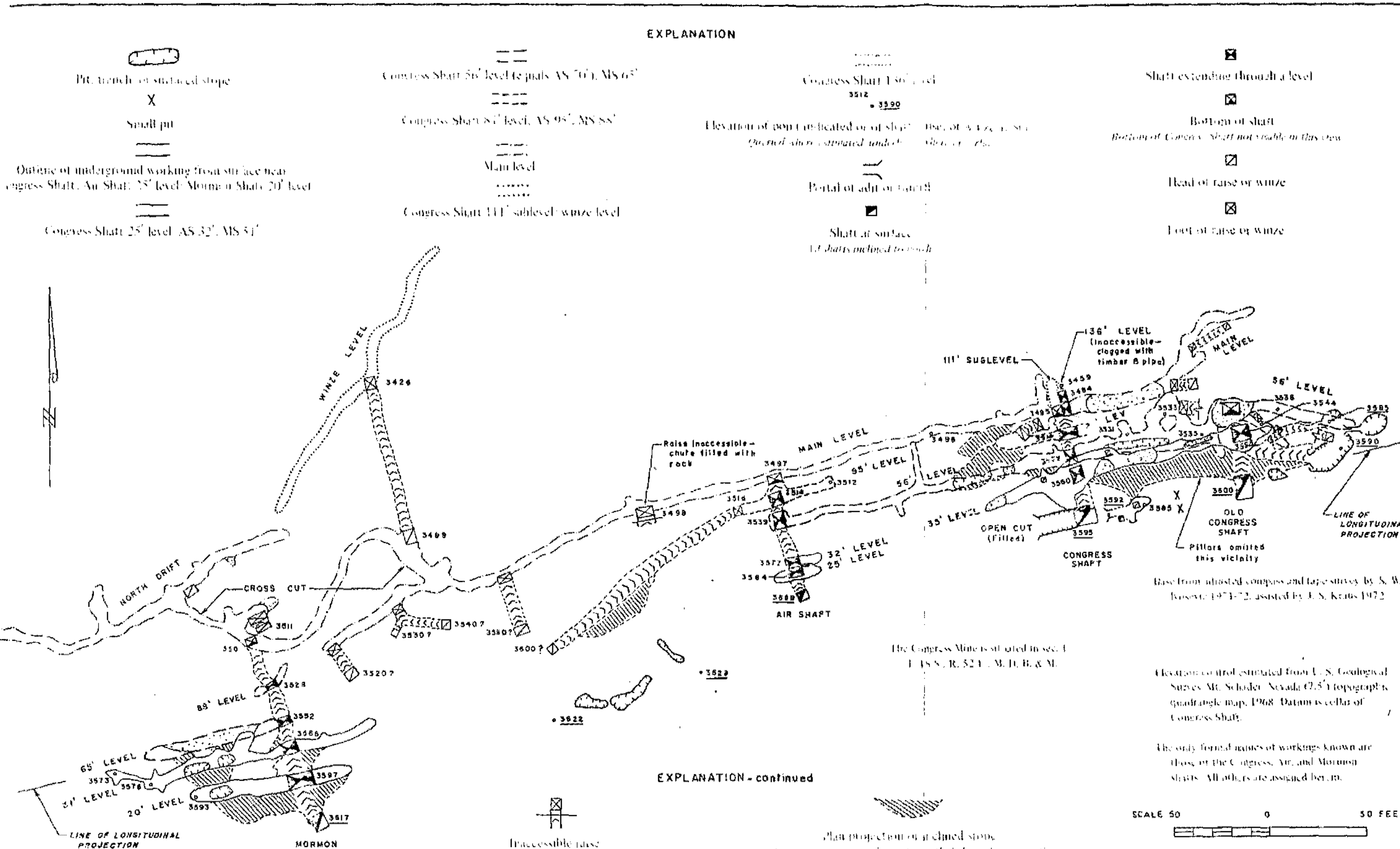
CONGRESS MINE

COMPOSITE LONGITUDINAL PROJECTION OF WORKINGS

Scale 1"=50'

By S. W. Ivosevic 1973

PLATE 9



JOHNNIE DISTRICT, NYE CO., NEV
CONGRESS MINE
COMPOSITE PLAN PROJECTION OF WORKINGS

Scale 1"=50'

By S. W. Ivosevic 1973

PLATE 1