SUPERGENE ENRICHMENT OF COPPER AT THE MIKE GOLD DEPOSIT, CARLIN TREND, NEVADA

A THESIS
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ABSTRACT

The Mike gold deposit, located about 15 kilometers northwest of the town of Carlin, Nevada, is one of many Carlin-type gold deposits along the Carlin Trend in northeastern Nevada. It is distinguished from other gold deposits of the Carlin Trend by highly anomalous concentrations of base metals. Many intervals of the oxidized portion of the deposit contain over 1 wt % copper, and the unoxidized portion of the Mike deposit contains abundant zinc, with lesser silver, lead, tungsten, molybdenum, and bismuth.

The purpose of this study was to document the distribution and mineralogy of copper in the deposit, in order to (1) determine the source of the copper; (2) understand the geological and geochemical controls governing the supergene copper enrichment process; and (3) determine the nature of the copper-arsenate mineral conichalcite, which is rare in most geological settings, but abundant in the Mike deposit.

Copper minerals are found predominantly in the oxidized zone of the deposit, and with the exception of very minor chalcocite which has replaced sulfides below the base of oxidation in a few areas of the deposit, there are no abundant sulfide copper minerals. Disseminated hematite with an aggregate botryoidal morphology similar to chalcocite marks the location of oxidized chalcocite blankets. These blankets are present throughout the deposit, including elevations far above the appearance of any copper oxide minerals. Conichalcite and chrysocolla are common at Mike, with chrysocolla typically forming lower and more central within the deposit than conichalcite. Malachite is rare, and is not observed in contact with conichalcite. Copper-bearing clays are observed in close proximity to some oxidized chalcocite blankets. Native copper and cuprite typically occur together in lower and peripheral parts of the oxidized zone, and are uncommon in areas containing an abundance of any other copper mineral. There is no apparent correlation between copper grade and the presence or absence of any particular copper oxide mineral.

Copper mass-balance calculations corroborate the observations supporting an exotic origin for the copper, transported in near-surface groundwater from an
undiscovered source northwest of the Mike deposit. There appear to be two stages of supergene copper enrichment: (1) Copper-bearing water migrated laterally under the paleohydraulic gradient and encountered pyrite-bearing wallrock at the Mike Deposit, where chalcocite replaced pyrite at the water table. (2) The water table later dropped to a lower elevation, exposing the chalcocite blankets to oxidation and leaching of copper, which reprecipitated locally as various copper oxide minerals.

The supergene copper enrichment at the Mike deposit was a consequence of copper-bearing waters encountering an area that had undergone prior Carlin-type hydrothermal activity. Previous decarbonatization, silicification, and hornfelsing of the wallrock removed any minerals which could serve as effective buffers to the copper-bearing, acidic waters. Abundant pyrite introduced to the rock with the emplacement of Carlin-type gold served to reduce the Eh and supply the sulfur to fix copper as chalcocite. Finally, the high arsenic content associated with Carlin-type pyrite helped retain the copper as conichalcite during oxidation of chalcocite and remobilization of copper during the second stage of supergene enrichment. Chrysocolla formed deeper in the supergene enrichment profile, after continued wall-rock interaction led to the hydrolysis of alkali feldspar and an increase in the silica concentration.

Conichalcite occupies a broad Eh-pH stability field at near-neutral conditions given sufficient arsenate, copper, and calcium in solution. This complete set of necessary conditions is absent in many environments, leading to the rarity of conichalcite. However, all these conditions for conichalcite precipitation are met in an oxidizing Carlin-type deposit, given a source of copper. Therefore, the recognition of even minor conichalcite in absence of other arsenic-bearing minerals in a Carlin-style mineral district may serve as a useful exploration tool.
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INTRODUCTION

Location and Significance of the Mike Deposit

The Mike gold deposit is located within the southern area of the Carlin Trend, Eureka County Nevada, about 15 kilometers northwest of the town of Carlin. The Carlin Trend is a 60 kilometer long, north-northwest trending linear array of Carlin-type gold deposits, sediment-hosted deposits of submicron size, disseminated gold (Teal and Jackson, 1997). The Mike deposit is unusual relative to other Carlin-type deposits of Nevada because of its high concentrations of Cu and Zn (both ranging as high as 10% each over 5 feet core assay intervals), with lesser of Ag, Pb, W, Mo, and Bi. Carlin-type deposits generally contain only traces of base metals (less than 0.01 %) (Arehart, 1996).

The Mike deposit is hosted in the Carlin tectonic window of the Maggie Creek district, about 4 kilometers northwest of the Gold Quarry Mine (Fig. 1). Published estimates for the Mike deposit metal inventory include 8.7 million ounces Au (average grade 0.021 oz/st), 1.02 billion pounds Cu (average grade 0.34 wt %), and 813 million pounds Zn (average grade 2.13 wt %). Cutoff grades established for this inventory calculation are 0.006 oz/st Au, 0.1% Cu, and 1.0 % Zn. The geometry of the ore body containing this inventory is based on a model using a $400/oz gold price and $0.8/pound copper price, and only includes the zinc reserves above cutoff grade which exist in the ore shape defined by the copper and gold values, rather than calculating a zinc price that could expand the volume of the deposit (Norby and Orobona, in press; J. Norby, pers. commun., 1999).

Early district exploration history

The area of the Maggie Creek district, and the Mike deposit in particular, did not represent an ore discovery in virgin territory. Earliest prospecting in the district began in the 1870s, resulting in the discovery of gold-silver-copper-lead ore (Roberts et al., 1967). Barite was mined from veins within the Good Hope jasperoid located along the fault of the same name northwest of Gold Quarry from 1906 to 1909 (Wilson, 1989). Doyle-Kunkel (1993) cites an internal Newmont report claiming that at the Copper King Mine, which is approximately 1 km southeast of the present Mike deposit, copper oxide
Figure 1. Location map

Location map for the Mike deposit, including outcrops of intrusions, major faults, and other gold deposits of the Carlin Trend. (map integrates elements from Norby and Orobona, in press; Ressel et al., 2000; Teal and Jackson, 1997; Newmont Mining Corporation TS Ranch Geology Map, scale 1:100,000)
minerals were discovered in the 1880s, and the mine produced 167 “railcar loads” (about 14,800 short tons) of copper oxide ore with average grade 3.4% Cu between 1952 and 1958. Chrysocolla, malachite, azurite, and cuprite are localized in the limestone beds of the Copper King Mine. No primary copper sulfides were reported (Erd et al., 1953).

Occidental Minerals explored this area of the Maggie Creek district for supergene copper during the early 1970s. Four core holes and two additional rotary chip holes were drilled southeast and within the presently recognized boundaries of the Mike deposit. Drill holes intersected chalcocite and oxidized chalcocite blankets, which they inferred to be the supergene enriched portion of a porphyry copper deposit, but they concluded that the size and grade of this copper prospect were too small to continue investment in the property. The majority of this drill core was not analyzed for gold (Occidental Minerals, 1974).

Newmont geologists explored the Carlin Window for gold, drilling approximately 130 exploration holes in the 1960s. This work led to the delineation of the Maggie Creek orebody, where mining began in 1979, overlying the present Gold Quarry pit. Gold Quarry was later discovered below 75 meters of unmineralized, basin-filling sediments east of Maggie Creek, and large-scale mining of Gold Quarry commenced in the mid 1980s (Rota, 1995). By the late 1980s, the Tusc orebody had been discovered along the Good Hope fault zone 2 km northwest of Gold Quarry, and early stage gold exploration work was underway at the Mike deposit (Fig. 1).

**Previous Work at the Mike Deposit**

The initial discovery of Carlin-type gold mineralization in the Mike deposit was made in 1989 by Newmont Gold Company. Subsequent characterization of the deposit was continued in the early 1990s, as documented by Branham and Arkell (1995). Teal and Branham (1997) present a deposit description summarizing the work done up to that time, and speculate that the base metals associated with the Mike system were introduced during a potassium feldspar flooding event, associated with a skarn proximal to an intrusion presumed to be below the Mike deposit.

The recognized boundaries and contained reserves of the Mike deposit were significantly expanded during a recent renewed interest, starting in 1998 and continuing
at present, with the drilling of several exploration holes deep into sulfide-bearing rock. Norby and Orobona (in press) present a thorough description of the Mike deposit, and synthesize much of the recent work, including brief characterization and grade distribution of several previously unexamined base metals. They propose a deep porphyry intrusion north or northwest of the Mike deposit as the source of all the base metals, and possibly even the source of gold in the entire Gold Quarry system. They also note the elevated zinc concentration in most drilled holes of the West Mike deposit present below the base of oxidized rock. Zn assays up to 1.65 weight % were reported from a very limited volume within the Tusc pit (Fig. 1), but no other rock of significant zinc concentration has been reported in the district (Wilson, 1989).

**Purpose of this study**

The anomalously high content of base metals found in the Mike deposit presents a number of challenges to our understanding of Carlin-type systems, which are generally considered to be gold-only deposits. The primary objective of this study is to document the mineralogy and spatial distribution of copper minerals and explain their patterns of distribution. Fieldwork was directed at collecting data and observations of the locations of the various copper minerals relative to (1) each other, (2) alteration associated with the core of the gold deposit, (3) the elevation of the boundary between the unweathered sulfide zone and the dominantly oxide zone, (4) hypogene quartz veins containing base-metal sulfides, and (5) any changes in bedrock lithology or alteration. This data set is then used to generate an interpretive model of the Mike deposit to understand the processes of supergene mobilization and concentration of copper relative to gold mineralization. Potential applications of this study lie in the use of this model as a predictive tool for delineation of further copper reserves at the Mike deposit, and in clarification of issues related to the benefaction of ores.
METHODS

Newmont granted full access to assay data from all Newmont holes drilled at the Mike deposit. In addition to over 75,000 gold assays, the database also contains over 14,000 base metal assays, and numerous multi-element ICP assays and sulfur and carbon whole rock assays, particularly from the most recent phase of exploration activity. Details of these assay data, grade information, and element correlations are summarized by Norby and Orobona (in press). The base-metal assay information in particular provided guidance for me in selecting drill core to log, and in modeling supergene remobilization of base metals.

Core logging

Due to the lack of surface exposures of the Mike deposit, this study is based on logging and sampling of drill core. Following Newmont convention, core assay and metal zone thicknesses are reported in inches and feet, while distances and stratigraphic thicknesses are reported in meters and kilometers, and my analytical work and sample descriptions were measured and are reported in metric μm, mm, and cm. I inspected split core with the associated geochemical assay information from many holes, and logged a total of 1845 meters of diamond drill core from seven holes (CV-16, REB-37, REB-68, REB-88, REB-89, REB-90, and REB-93), along with 60 meters of reverse circulation drilling rock chips from REB-90 (Fig 2). The split drill core from holes REB-37 and REB-68 consisted of rubble, and was logged at a scale of 1”=20’, while all of the other holes, due to better condition of the core, were logged at a scale of 1”=10’. The complete core of these seven holes were logged, except REB-93, and REB-90, in which the lowermost sulfide zones which lacked economic mineralization were not logged. Two of the deepest holes in the Mike deposit, REB-88 and REB-89, also drilled deep into sulfide zones below economic gold or base-metal mineralization, were logged in their entirety to provide detailed data on the unmineralized deeper parts of the system.

My logging efforts built upon lithologic, structural, and alteration logging completed by Newmont geologists by adding specific mineral identification, mineral assemblages and distribution, and macroscopic textural details (e.g. whether minerals were located along fractures, bedding horizons, or disseminated). Special attention was
Figure 2. Bedrock geology and drill-hole location

Interpreted bedrock geology of the Mike deposit (from Norby and Orobona, in press), with locations of drill holes logged and analyzed in detail for this study. Thick outline marks the spatial extent of the Mike deposit at bedrock surface as defined by gold mineralization at a 0.01 oz/st cutoff grade, and grade thickness equal to or greater than 5. Grade thickness is the sum, over the depth of the hole, of grade multiplied by feet thickness of each assayed interval (from Fig. 2, Norby and Orobona, in press).
given to copper oxide, iron oxide, and hypogene sulfide minerals, along with quartz veins and the minerals contained within them. Vein mineralogy on previous core logs was augmented by observations of thickness and abundance of veins, specific vein locations, and orientation (dip angle to core axis). I also measured and recorded permeability for hole REB-88, using a relative scale, to establish a representative example of the relationship between various alteration, lithology and mineralization types and rock permeability. Copper host minerals were identified by visual inspection and binocular microscope, also by plating copper onto steel nails with acid.

From the seven holes logged, I collected a total of 251 samples. 137 samples were selected for more detailed analysis at Stanford, using X-ray diffraction (XRD), transmitted and reflected light microscopy in polished thin sections, short-wave infrared spectrometry (using PIMA), scanning electron microscope (SEM) and electron microprobe analysis. Chemex analytical labs performed 32-element ICP analyses, a set of Sulfur assays, or particular element assays respectively, on selected samples. The oxide zone samples pictured in the images of Figures 7A, 8-11, and 14-20 are all from the Rodeo Creek Unit, while the sulfide zone vein samples in Figures 7B-E are from the Popovich Formation (both formations discussed in the lithologic overview subsection below). Oxide zone samples from the Roberts Mountains Formation of REB-37 were studied by XRD and microscope, but most were very friable, and none were used in the images of this thesis.

**XRD**

A Rigaku Geigerflex X-ray diffractometer was used to analyze 59 samples of hand-picked material from drill core, in order to identify unknown minerals and mineral mixtures. Powdered samples were run for a two-second scan time at each successive 0.05° increment, between X-ray incident angles of 5° to 65° relative to the plane of the sample surface. Clay samples produced unclear XRD spectra when analyzed using the method outlined above. My purpose in XRD analysis of these clay samples was primarily for clay mineral reconnaissance, rather than exact definition of all clay species in these mixed samples, so I used a simplified technique of clay particle XRD analysis. To orient the clay particles, I mixed a few drops of deionized water with the powdered clay sample on a glass slide, and allowed the mixture to dry for at least 24 hours.
Quartz was present in most of the samples, and was possibly introduced to some receptive fine-grained material during grinding with a chert mortar and pestle. The XRD spectrum for quartz was utilized as an internal standard to assist in the correct identification of the peaks of other minerals in the sample. This internal standard was particularly useful in the difficult cases of mineral mixtures in a sample, or in the identification of rare minerals.

**PIMA**

A short-wave infrared spectrometer was used to analyze of the mineral content of 25 samples. Due to the sensitivity of short-wave infrared spectrometry to cation-OH bonds, the PIMA proved useful in the verification of clay identity for some samples which were difficult to resolve with XRD. The PIMA (Portable Infrared Mineral Analyzer) shines an internal light source upon a sample surface, and measures reflected radiation between 1.3 and 2.5 μm wavelengths. This technique is particularly sensitive to cation-OH bonds, and certain radicals contained in mineral structures including CO₃, H₂O, and OH (Thompson et al., 1999). The reflectance spectrum of a sample is compared with a library of short-wave infrared spectra collected from known mineral samples and reference samples. A matching spectrum, or combination of mineral spectra identifies the mineral composition of the sample.

No special sample preparation was required for use of the PIMA. Samples were held flat against the PIMA detector window for approximately 30 seconds of collection time. In cases where flat surfaces were unavailable, external light, which would interfere with the PIMA measurement, was reduced with a piece of aluminum foil.

**Microscope and electron microprobe**

Sixty-five samples were selected to make polished thin-sections. These sections were reviewed with a Zeiss research-quality reflected and transmitted light microscope to observe larger-scale relations between minerals of particular assemblages, and to select the most appropriate samples for electron microprobe analysis.

Eight thin sections were analyzed using a JEOL JXA-733A electron microprobe. The electron microprobe is equipped with an energy-dispersive system (EDS), which was used to generate element maps. Five wavelength-dispersive spectrometers on the electron microprobe were used for quantitative elemental analysis at 15kV and 2 μm
beam size, yielding approximately 2-4 μm resolution. Quantitative elemental analysis is limited to an array of 10 elements maximum, or fewer depending on the specific elements selected for analysis. Fe, Pb, P, As, Zn, Cu, Ca, S, and Ni were analyzed in sulfide zone samples, whereas the same suite of nine elements except for the substitution of Si for Ni were analyzed in oxide zone samples.

SEM

Numerous rock fragments taken from 22 separate samples were analyzed with a JEOL JSM-5600LV Scanning Electron Microscope (SEM) equipped with an EDS detector. SEM was used to view textural relationships between minerals at the micron scale, on broken surfaces, and to generate element maps from two samples, one polished surface, and one broken, but relatively flat surface.
REGIONAL GEOLOGY

Tectonics and Magmatism

In the region of the Carlin Trend, the Roberts Mountains Thrust, a set of imbricated thrust sheets, separates overthrust siliciclastic Ordovician formations from underlying time correlative and younger carbonates. These thrust faults are related to the Late Devonian-Early Mississippian Antler Orogeny, the first period of compressional tectonics recognized in the region. After rifting sometime in the late Proterozoic, the western margin of the North American continent had been a passive margin, subject to widespread carbonate deposition and offshore sedimentation, producing the lithologic units currently observed at the Carlin Trend (Roberts et al., 1958).

The compressional, late Pennsylvanian Humboldt Orogeny affected the Carlin Trend area subsequent to the Antler Orogeny, but with somewhat less pronounced results. No major faults have been attributed directly to the late Pennsylvanian Humboldt Orogeny, which simply reactivated structures of the Antler orogeny (Snyder, 1991; Thorman et al., 1991). The existence of the Humboldt orogeny has been inferred from a preponderance of orogenic sedimentary rocks and prominently visible angular unconformities of Pennsylvanian through Permian age within the stratigraphic record of northeastern Nevada (Thorman et al., 1991). Local dome structures, believed by Roberts et al. (1971, p. 16) to have been initiated during late Pennsylvanian time, would therefore probably have been the result of the Humboldt Orogeny identified by Thorman et al. Later Mesozoic orogenies (discussed below) further accentuated these anticlinal structures, and subsequent denudation of the Roberts Mountains allochthon over the domes created windows exposing lower plate Silurian-Devonian silty carbonate sequences. The importance of these tectonic structures lies in the fact that Carlin Trend gold deposits are hosted on the margins of these dome-shaped structures, typically in the underlying silty carbonates (Teal and Jackson 1997).

East to southeast-vergent thrust faults, and other faults related to the Elko Orogeny of Jurassic age, are associated with folding and regional metamorphism of a large area of the Great Basin (Thorman et al., 1991). Granitoid plutons were intruded the
area at this time. K/Ar dates range between 168 Ma and 143 Ma, no error bounds reported (Roberts et al., 1971). Arehart et al. (1993) report preferred $^{40}\text{Ar}/^{39}\text{Ar}$ dates of 158.1±1.5 Ma, 157.2±1.7 Ma, and 157.5±1.7 Ma for hornblende and two biotite grains, respectively, from the altered monzonite Goldstrike stock (Fig. 1). Leonardson and Rahn (1996), and Volk et al. (1996) conclude that folds and faults in the area of the Goldstrike Stock also formed during this time, although these features have not been specifically linked by these authors to deformation associated with the Elko orogeny recognized by Thorman et al. (1991). The large offset, throughgoing, northwest striking faults (e.g. Post Fault, Good Hope Fault, Fig. 1), and the more common northeast striking faults exhibiting less offset, are both probably at least Jurassic in age. These northwest and northeast-striking fault sets are both occupied by Cretaceous-age andesitic dikes (Teal and Jackson, 1997).

From Cretaceous through Paleocene time, the compressional, east-southeast directed Sevier orogeny affected areas of Nevada and western Utah. In the region of the Carlin Trend, Sevier orogeny compression appears to have produced only broad folds and thrust faults of small displacement (Thorman et al., 1991). The Richmond quartz monzonite stock (Fig. 1) was emplaced at 106±2 Ma (K/Ar date from biotite) (Evans, 1980). A minority of researchers believe that extension in the Great Basin initiated during Cretaceous time between 90 Ma and 60 Ma, as a result of Sevier orogeny crustal overthickening (Livaccari 1991, Vandervoort and Schmitt 1990). The latter authors argue that scarpfront alluvial and landslide deposits late of Late Cretaceous age constitute evidence of this early horizontal extension.

There has been a considerable amount of research and speculation concerning the causes of North American Cordilleran compressional tectonism, the ensuing Great Basin extension, and the relationship between tectonism and ore deposits of the Great Basin (e.g. Coney 1987; Dumitru et al., 1991; Humphreys 1995; Livaccari 1991; Miller and Gans 1991; Opplinger et al., 1997; Seedorff 1991; Thorman et al., 1991; Vandervoort and Schmitt 1990). Late Mesozoic compressional tectonics are primarily attributed to the effects of shallow slab subduction of oceanic plates beneath the North American continent and linkage of the subducting plate with the base of the overriding North American continent. As an increasing amount of dense oceanic plate subducted, shallow
subduction is thought to have become unstable, and the slab either rolled back and fell away from the crust under its own weight, or more likely tectonically buckled (Humphreys 1995), perhaps induced by an ascending or overridden Yellowstone hotspot magma plume (Oppliger 1997). The proposed buckling, marking the end of shallow slab subduction, initiated about 50 Ma, was heralded by widespread and voluminous magmatism, which was time-transgressive from both the north and the southeast of the Great Basin, until these magmatic fronts met in southern Nevada about 20 Ma. This magmatic event was followed by southward migrating time-transgressive tectonic extension which commenced during late Eocene time; magmatism generally preceded the onset of significant Basin and Range extension (Gans et al., 1989).

Geochronological studies document the age of Cenozoic magmatism in the area of the Carlin Trend. Eocene volcanic centers have been described by Henry and Boden (1998), who determined that magmatism was particularly intense in the vicinity of the Carlin Trend from 40 to 39 Ma. Biotite grains from a granodiorite dike and a large quartz latite dike, both cropping out at Welches Canyon stock (Fig. 1) near the southern Carlin Trend, have been dated using the K/Ar method at 37±0.8 Ma, and at 36.6±0.7 Ma, respectively (Evans, 1980).

The current Yellowstone hotspot punctured the continental lithosphere in north-central Nevada about 17 Ma, resulting in widespread volcanism through Late Miocene time (McKee, 1996; Parsons et al., 1994). The basins in the area of the Carlin Trend contain basal volcanic rocks in some locations, and volcanic units are interlayered with clastic sedimentary rocks, limestone, and an upper unit of lacustrine clay. Regnier (1960) has paleontologically dated the sedimentary units as Late Oligocene to Present and concludes that an upper, thin welded tuff of this basin fill is of Pleistocene age.

Extensional deformation in northeastern Nevada is believed to have occurred during two distinct stages. The first stage began during late Eocene time (~35 Ma) and continued through early Miocene time (~20 Ma), and a later stage extended from Late Miocene (after ~10 Ma) through Holocene time, with a period of relative tectonic quiescence separating the two faulting stages (Thorman et al., 1991). The majority of extension in the northern Basin and Range area is thought to have occurred during the earlier stage, after the initial widespread Eocene volcanism (Seedorff 1991). However,
there appears to be only minor extension in the region of the Carlin Trend associated with Eocene magmatism, with little or no stratal rotation of local Paleozoic units (Henry and Boden, 1998). There also does not appear to be evidence for significant amounts of extension in this area during the later stage of extension, because most of the formations at the Carlin Trend are relatively flat-lying (Seedorff 1991).

In addition to Paleozoic thrust faults, there are three predominant later fault sets in the Carlin Trend (Figs. 1, 2). Continuous, northwest striking faults comprise the oldest set. These faults show multiple slip episodes, including oblique, reverse, and normal offset, indicating reactivation at different times. Younger, northeast striking normal offset faults crosscut the northwest faults. Both of these sets of faults are in turn offset by a localized third set of north-northeast striking normal faults, which also offset the contact between bedrock and overlying basin-filling, Neogene-aged Carlin Formation. The cumulative effects of the long sequence of several orogenic events acting over the region has produced the structurally complex collection of superimposed doming and folding, faulting, and minor tilting, observed in the geological record at the Carlin Trend today.

Stratigraphy

The succession of stratigraphic units of the southern Carlin Trend are described in detail by Norby and Orobona (in press). Units are dated to geologic periods based on fossils (Evans 1980; Johnston 1995; Cole 1995). The lowermost unit is the Late-Middle Ordovician to Early Silurian, dolostone and dolomitic limestone of the Hanson Creek Formation. This unit is conformably overlain by the Early Silurian to Early Devonian Roberts Mountains Formation, a grey, planar-laminated, silty limestone. The Roberts Mountains Formation is in turn conformably overlain by the Popovich Formation, an Early to Late Devonian sequence of micrite, silty limestone, and fossiliferous debris-flow limestone. In most locations of the Carlin Trend, the Popovich Formation grades into the upper Middle to lower Late Devonian Rodeo Creek Unit, a thin-bedded, calcareous siltstone interbedded with siliceous mudstone.

The early Late to middle Late Devonian Marys Mountain sequence is recognized by Norby and Orobona (in press) to be the lowermost allochthonous unit of the Maggie
Creek district. This unit of silty limestone, siliceous mudstone, and limey mudstone is cut by numerous deformation structures, and is likely a transitional assemblage between the siliceous western assemblage and the underlying autochthonous carbonate platform, thrust some distance to its present location. Overlying the series of imbricated thrusts of the Roberts Mountains thrust fault is the Early to Late Ordovician Vinini Formation, an allochthonous chert, siliceous mudstone, and siltstone unit of the distal passive continental margin, which was later thrust eastward during the Antler Orogeny. Though chronologically younger than the Marys Mountain sequence, the Vinini Formation structurally overlies the Marys Mountain sequence. The erosional surface of the Paleozoic sedimentary units located in current downfaulted grabens was buried under Miocene siliciclastic, basin-filling Carlin Formation. The Roberts Mountains and Popovich Formations, and Rodeo Creek Unit, all host economically significant concentrations of gold in different areas of the Carlin Trend, and all three contain gold zones within the Mike deposit (Fig. 3). The Vinini Formation is known to contain smaller, fault-controlled gold deposits on the Carlin Trend (Teal and Jackson, 1997). The units that are present in the Mike deposit are described in more detail below.

**Carlin-type Mineralization**

Carlin-type gold deposits are both structurally and stratigraphically controlled, located along throughgoing northwest-striking faults with a history of reactiviation. The deposits are typically at intersections of these long faults with crosscutting northeast-striking normal faults. Hydrothermal fluids passed through these permeable fault conduits and infiltrated receptive stratigraphic units. A significant amount of arsenic, along with lesser antimony, mercury, and thallium, were introduced with gold. Gold is submicron in size, primarily concentrated in arenstian pyrite overgrowths on pyrite grains (Arehart, 1996). Common features of Carlin-type mineralization and alteration, and much of the extensive body of research on Carlin-type Deposits, have been summarized by Teal and Jackson (1997) and Arehart (1996).

The common forms of alteration which accompany gold are decarbonatization, silicification, "argillie" alteration, and sulfidation of host rocks to form pyrite and arsenical pyrite. Decarbonatization of host rocks is spatially the most pervasive form of
alteration, although the permeability and porosity of the stratigraphic host rocks exert some measure of control over the distance from the deposit core and extent to which the rock is decarbonatized. The replacement of host rocks by silica is similarly affected by rock type and hydraulic properties of the host. Intense silicification, especially along faults and fractures, formed jasperoids. "Argillic alteration" is a term assigned to the formation of fine-grained montmorillonite, kaolinite, or illite, and/or the concentration of these clays in the protolith by dissolution of carbonate minerals. Dependent on the relative amount of silicates present in the host rock, argillic alteration is more pronounced in the silty limestone horizons with a higher component of detrital feldspars and clays, and is very pronounced in many of the intrusive dikes. Detailed study of the interplay between these Carlin fluid alteration types and gold mineralization was undertaken by Bakken and Einaudi (1986), who discovered a positive correlation between gold grade and intermediate degrees of both silicification and decarbonatization in the Carlin mine.

**Dating of Carlin-type mineralization**

Studies addressing the age of gold mineralization along the Carlin Trend have produced a wide variety of results, leading to significant debate between researchers. Age estimates range from as old as 117 Ma (Arehart et al., 1993) to as young as 16 Ma (Radtke et al., 1980). The primary reason for disagreement lies in the lack of a definitive relationship between the limited datable minerals in Carlin-type deposits and the gold mineralization event, and in the case of Radtke et al. (1980) to assumptions that the age of the deposits were the same as local volcanic rocks, and must be young because the redox boundary, interpreted to be hypogene, was parallel to the present surface. Many researchers attempt to correlate some type of local magmatic activity with their age estimate for gold mineralization at the Carlin Trend. Examples include "...117 Ma age is consistent with one of the major magmatic pulses recognized in the northern Great Basin" (Arehart et al., 1993), the inferred Eocene position of the Yellowstone hotspot (Oppliger et al., 1997), Eocene magmatism (e.g. Henry and Boden, 1998), and late Tertiary magmatism (Radtke et al., 1980).

After a thorough examination of published radiometric dating of clearly mineralized and unmineralized features which cut through gold mineralized rock of Carlin-type deposits, Hofstra et al. (1999, p. 790) concluded that deposits of the northern
Great Basin formed between 42 and 30 Ma. The majority of recent researchers appear to favor such a mid-Tertiary age for Carlin Trend gold mineralization, based on textural relations and structural arguments. Hofstra et al. (1999) summarize stable isotope work, which in correlation with paleofloral and other methods of indirect dating, also provides evidence for a mid-Tertiary age of Carlin Trend deposits.

**Neogene paleoclimate**

There is limited information concerning paleoclimate in northeastern Nevada prior to mid-Miocene time. Based on their exhaustive summary of paleontological studies of the region, and of a steady decline in the value of $\delta D_{\text{H}_2\text{O}}$ in supergene alunite with decreasing age, which mirrors global ocean $\delta D_{\text{H}_2\text{O}}$ values for the same latitude and time period, Hofstra et al. (1999) argue that the climate of the northern Great Basin has become increasingly colder from late Cretaceous to its lowest point in mid-Oligocene time, about 30 Ma. There has been some variability since then in northeastern Nevada, but never a return to its warm climate of the Mesozoic.

Overall, the Late Eocene through Miocene time in northern Nevada appears to have been an arid environment, based on the lack of species diversity in paleoflora and fauna from that time (R. Madrid, pers. commun., 2000). However, supergene alunite, formed during Oligocene and Miocene time (Arhart et al., 1992), requires substantial meteoric water to form. Stronger evidence of punctuated periods of higher rainfall during this period lies in the development of some minor lateritic zones during the Miocene period. While the period as a whole may have been arid, there must have been some intervals of a wetter, more temperate climate (R. Madrid, pers. commun., 2000).
DEPOSIT GEOLOGY

Lithologic Overview

Past deposit descriptions of the Mike deposit, by Teal and Branham (1997) and by Norby and Orobona (in press), provide much of the material summarized below. The Mike deposit is located along the northwest-striking Good Hope Fault at its intersection with the northeast-striking Soap Creek Fault (Figs. 1, 2). Mineralized bedrock is buried under 120 to 240 meters of Carlin Formation (Fig. 3). The Good Hope Fault is a large, throughgoing fault with a history of reactivation, and appears to be the feeder structure for upwelling hydrothermal fluids which produced the gold mineralization. The fault currently exhibits a minimum reverse-sense offset of 750 meters, placing Silurian-Devonian Roberts Mountains Formation against Devonian Rodeo Creek Unit (Fig. 3) However, absolute vertical offset is difficult to establish because no upper or lower stratigraphic contact of Roberts Mountains Formation has been found in the hanging wall of the Good Hope Fault. This fault is a major feature of the deposit, and divides the Mike deposit into two parts: the hanging wall is referred to as Main Mike, and the footwall is referred to as West Mike (Fig. 3).

Bedding in the Mike deposit is nearly horizontal (Norby and Orobona, in press). Some dikes intrude fault planes and fractures. Hornfelsing at the Mike deposit (described below) obscures the subtle changes between the stratigraphic units, making interpretations of the contact between formations difficult to decipher. In addition to hornfelsing, the mineralized rock at Mike is predominantly decarbonatized, variably silicified, and exhibits argillic alteration in some places, particularly pronounced in the intrusive dikes.

Structure

Due to the deep Carlin Formation cover blanketing the Mike deposit, fault locations are estimated from stratigraphic offset observed between drill holes and from positions of faults and brecciated dikes in drill core. Gravity gradient data collected by Newmont, and current drainage topography developed on the surface, have also assisted in the interpretation of faults displaying post Carlin Formation movement. Regardless of
Figure 3. **Mike deposit cross-section**

Generalized SW-NE cross-section depicting lithology and metal-enriched zones of the Mike deposit (from Norby and Orobona, in press).
orientation, these faults all appear in drill core to focus slip on a few slip zones, each typically up to 5 meters thick (J. Norby and M. Orobona, writ. commun. 1998-1999). Drill-hole spacing is not sufficiently close to determine cross-cutting relationships between faults at Mike, so inferences were made from cross-cutting relationships in exposed faults of the Maggie Creek district (Norby and Orobona, in press).

The faults of the Mike deposit, as modeled by Norby and Orobona (in press), comprise three distinct fault sets (Fig. 2). Two continuous, northwest-striking faults are the oldest, with the northeast-dipping Good Hope Fault showing multiple slip episodes, including oblique, reverse, and normal offset, and the southwest-dipping Corridor Fault exhibiting normal offset. A set of younger, northeast-striking, normal offset faults crosscut the northwest faults. Both the northwest- and northeast-striking faults are in turn offset by a third set of north-northeast striking, apparently steeply dipping normal faults. The youngest set of faults exhibits compelling evidence of Neogene movement, offsetting the contact between bedrock and Carlin Formation. The northeast-striking faults display minor normal offset of this contact, while there is possible evidence of late normal sense Neogene reactivation on the oldest Good Hope Fault (Norby and Orobona, in press). The Roberts Mountains Thrust Fault is not observed in the Mike deposit.

Striking N40°-50°W and dipping 35°-45°NE, the Good Hope Fault zone is 15 to 45 meters wide, consisting of several shears through a brecciated zone with clay matrix. This fault appears to have served as a major conduit for gold-bearing fluids, and it contains both gold mineralization and also later copper mineralization. At the Mike deposit, the Good Hope Fault displays at least 750 meters of reverse sense stratigraphic offset, although the net offset cannot be ascertained due to the lack of a common stratigraphic marker observed in both hangingwall and footwall units. Drill data indicate a trough shaped depression of about 30-60 meters in the bedrock surface along strike on the hangingwall side of the Good Hope Fault, indicating possible later reactivation of the fault in a normal sense (Norby and Orobona, in press). The Good Hope Fault is inferred from its crosscutting relationships observed primarily in the Gold Quarry Mine to be offset by all other faults that intersect it at the Mike deposit.

The Good Hope Fault, particularly at the intersection with the northeast striking Soap Creek Fault, hosts a dense, silicified jasperoid which forms a bedrock high, and
which appears to extend down the plane of the Good Hope Fault zone (Norby and Orobona, in press). Where jasperoid is not found, dike intrusions along the Good Hope Fault are not uncommon. The fault zone and its infilling material serves as a gold ore host, and also appears to have been a conduit for transport of later supergene copper-bearing fluids. Rodeo Creek and Popovich Formation bedding planes in the footwall of the Good Hope Fault begin to dip monoclinally more steeply northeast with increasing proximity to the fault from their subhorizontal orientation in West Mike (Fig. 3).

The Corridor Fault, an apparent normal fault striking N40°-60°W and dipping approximately 60°SW, is defined by a few holes at the southwest edge of the Mike deposit that have intersected a 5-15 meter wide clay matrix breccia. This fault displays no geophysical gravity gradient signature or evidence of offset of the Carlin Formation-bedrock contact, but offsets bedrock stratigraphy by about 260 to 300 meters. The Corridor Fault is not known to be exposed anywhere in the district. The southwestern limit of thermal hornfelsing at the Mike deposit is coincident with the location of the Corridor Fault. Similar to the Good Hope Fault, the Corridor Fault appears to have been a primary channel for copper-bearing supergene fluids (Fig. 4).

A series of northeast-striking faults also dissect the Mike deposit, as recognized from stratigraphic offsets. The most pronounced of these features on a district scale is the Soap Creek Fault zone, the original target for Mike exploration activity by Newmont, but the group also includes the named Perseverence, Independence, Soap Creek Parallel, and Independence Parallel Fault zones (Fig. 2). This series of faults strike N40°-50°E and are estimated to dip 70°-80°NW, spaced apart by 120 to 300 meters, judging from drill hole and geophysical gravity gradient data (Norby and Orobona, in press). Apparent normal offset on each of these faults is less than 60 meters, while the offset of the bedrock surface under the Carlin Formation appears to be less, indicating that faults of this orientation may have formed before the deposition of the Carlin Formation, but were reactivated sometime after early Miocene. Where exposed elsewhere in the Maggie Creek district, faults of the same orientation consistently cross-cut the apparently older, northwest-striking faults. The Soap Creek Fault contains a dike which hosts gold
mineralization, and served as a conduit for supergene fluids (J. Norby, pers. commun. 1998).

North to north-northeast striking normal faults appear to dip 60°-80° west, with highly variable offset. Structures interpreted to belong to this group of nearly north-striking faults include the D-Day, Nebulous, Valley, and North-Pointing Dog Fault zones (Fig. 2). Interpretation of these faults is based on offsets necessary to account for elevations of features in cross-sections. The offsets of the bedrock surface appear to be equal to the offsets in the Paleozoic units, and the north-striking faults appear to locally offset the basal limit of oxidation as well as mineralized zones. These facts imply that the north-striking faults are late, post-dating the Miocene Carlin Formation (Norby and Orobona, in press).

**Stratigraphy**

**Formations and units**

Roberts Mountains Formation is a grey colored, planar laminated, silty limestone, with lamination more wavy near the contact with the overlying Devonian Popovich Formation. Minor siltstone occurs in Roberts Mountains Formation as thin, contorted layers of somewhat lighter color. The thickness of the Roberts Mountains Formation varies from about 360 to 450 meters thick in the area of the Carlin window (Rota, 1995). Roberts Mountains Formation has been encountered in the deepest drill holes at West Mike, where it exhibits very few signs of metamorphism.

Conformably overlying the Roberts Mountains Formation is the Devonian Popovich Formation. Lost to erosion at Main Mike, the Popovich Formation is a gold and zinc ore host at West Mike, where it has been approximated by Norby and Orobona (in press) to be about 360 to 400 meters thick, determined by a few deeper drill holes. This thickness at West Mike is approximately consistent with the thickness of the Popovich Formation throughout the Maggie Creek district. Three members of the Popovich Formation are recognized in this part of the Carlin Trend (Norby and Orobona, in press). The stratigraphically lowest and thickest of the three is a black, massive micrite, with subordinate layers of grey calcarenite and minor silty limestone. The middle member of Popovich Formation also contains thick sections of bioclastic debris.
interlayered with medium-bedded calcarenite and carbonaceous silty limestone. The upper member is dominated by dark, increasingly silty, medium- to thick-bedded carbonaceous silty limestone, with rare, light grey calcarenite. Much of this sedimentary detail is obscured by hornfelsing at West Mike.

The Popovich Formation along most of the Carlin Trend conformably grades into the overlying Rodeo Creek Unit. However, over much of the Maggie Creek district, including West Mike, clay-filled shears and clast-supported breccia are commonly coincident with this contact. The Rodeo Creek Unit itself is typically a grey colored, planar-laminated limey siltstone, with thin layers of interbedded siliceous mudstone in deeper sections of Rodeo Creek Unit. At West Mike, the uppermost section of the Rodeo Creek Unit has been lost to erosion. Elsewhere in the Maggie Creek district, the thickness of Rodeo Creek Unit as determined by drilling is approximately 320 meters (Norby and Orobona in press).

All pre-Miocene units at the Mike deposit were later eroded, and buried by the local basin-filling Carlin Formation. Regnier (1960) determined the age of the Carlin Formation to be Miocene to Early Pliocene, based on fossil evidence and correlative intervals with other Cenozoic basin-filling formations of northeastern Nevada. A tuff unit at the base of the Carlin Formation, sampled north of the Bootstrap Deposit at the northern end of the Carlin Trend, yielded $^{39}$Ar/$^{40}$Ar dates of 15.1-14.4 Ma from alkali feldspar grains, with reported error bounds no larger than ± 0.2 Ma per sample (Fleck et al., 1998). The Carlin Formation is a lacustrine and fluvial volcanioclastic siltstone and sandstone interlayered with conglomerates and gravels, with local volcanic tuff layers. Above the basal section of Carlin Formation lies 0-60 meters of lacustrine tuff, now altered to pale green, waxy clay, locally red-colored due to oxidization. In the southern portion of West Mike, the basal 100 meters of Carlin Formation appear to be an indurated tuff (Norby and Orobona, in press). The uppermost section of Carlin Formation consists of 75-135 meters of buff to white, reddish, and pale green siltstone and fine-grained sandstone, capped by 0-60 meters of pediment gravels (Teal and Branham 1997; Norby and Orobona in press).
Dikes

Clay-altered intrusive dikes fill many of the northeast-striking faults and fractures, and the northwest-striking Good Hope Fault. Norby and Orobona (in press) describe the dikes as "fine grained latite or microdiorite dikes, with finely porphyritic" texture, from a few inches to fifty feet thick. They range between intermediate and mafic composition, tending to be quartz poor. Minerals identified within the dikes include relict plagioclase, orthoclase, biotite, hornblende, and pyrite (Teal and Branham 1997). Where intruded into a gold-bearing stratigraphic horizon, the dikes also typically host gold, implying they are older than the introduction of gold. Similar to the sedimentary host units, that the dikes are also susceptible to reaction with hydrothermal fluids. A grain of K-feldspar from one of these dikes has been dated at 107±2 Ma (K/Ar), implying that the dike is at least Cretaceous in age (Branham 1994). (See discussion below.)

Alteration.

In addition to the typical forms of alteration associated with Carlin-type gold mineralization, the Mike deposit also contains areas of secondary K-feldspar (Teal and Branham, 1997), quartz veins with sodium-rich amphibole (Williams, 1994), and pervasive calc-silicate hornfelsing. Teal and Branham (1997) speculate that the area of the Mike deposit underwent a potassium metasomatic event, to explain the anomalous potassium enrichment in altered Roberts Mountains Formation from samples collected from Main Mike, compared with low potassium measurements (Bakken, 1990) which they cite from this unit in the Carlin pit. A sample of K-feldspar from this "potassium-flooded" section of the Roberts Mountains Formation has been K/Ar dated at 111±2 Ma, similar to a 107±2 Ma date from another K-spar grain at Main Mike, taken from an altered andesite dike (Branham, 1994). If the 107±2 Ma date does represent the age of the dike, this implies that the potassic metasomatic event may have been coincident with intrusion of the dikes.

Sedimentary units of the Mike deposit display variable affects of thermal alteration, dependent largely on protolith. Decarbonatized silty carbonate of the Roberts Mountains Formation at Main Mike is intercalated with lenses of white or maroon colored, decarbonatized, calc-silicate hornfels, and minor fine-grained marble (Teal and
Branham, 1997). Siliceous mudstone and limey siltstone of the Rodeo Creek Unit at West Mike has been altered primarily to a purple to tan colored, planar laminated, quartz hornfels, with minor thin intervals of calcsilicate hornfels. The underlying silty carbonate Popovich Formation of West Mike has been hornfelsed to a white to olive-brown colored calc-silicate hornfels and fine-grained marble, similar to the Roberts Mountains Formation at Main Mike (Norby and Orobona, in press). The marble appears to have formed preferentially from calcarenite protolith. Some degree of hornfels in at least a few stratigraphic horizons is observed throughout the Mike deposit, although hornfelsing decreases in intensity at the margins of the deposit, and with increasing depth below the lowermost gold zones in some locations where data from deep drilling exists (Fig. 3).

**Sulfide-bearing Veins**

Norby and Orobona (in press) describe two types of veins: base-metal sulfide veins, and sulfosalt bearing veins in the deepest portions of the deposit. Found in the hornfelsed and potassium metasomatized portion of the Mike deposit, base-metal sulfide veins are up to 5 cm wide, and contain quartz with pyrite and lesser base-metal sulfides. Dark-colored sphalerite is the most abundant of these base-metal sulfides, with subordinate chalcopyrite, galena, and rare occurrences of molybdenite, stibnite, and arsenopyrite. These veins are younger than dikes of West Mike that they cross-cut.

Sericite which appeared to Branham (1994) to be texturally affiliated with quartz base-metal sulfide veins cutting one such dike in Main Mike was dated by K/Ar methods at 71.6±1.8 Ma.

Shiny metallic minerals are observed to be present in quartz-carbonate veins of the hornfelsed portions of the Mike deposit. The identification of these minerals as sulfosalts is inferred from ICP analyses that show arsenic, bismuth, lead, and silver are higher within the vein-bearing core intervals than in core without veins (Norby and Orobona, in press). These authors also observed calcite veins deeper in the Mike deposit, and have logged quartz veins, and quartz-limonite (oxidized section) veins (J. Norby and M. Orobona, writ.commun., 1998-1999).
Metal Zones

Gold in the Mike deposit is located in the Good Hope and Soap Creek Fault zones, and in certain stratigraphic units in contact with these faults. There are two main, relatively flat, tabular gold zones (Fig. 3) that appear stratigraphically controlled in West Mike, (1) an upper, primarily oxidized gold zone (average gold grade 0.025 oz/st) in the Rodeo Creek Unit and in the upper section of the Popovich Formation, and (2) a higher grade, partially oxidized, lower gold zone (average gold grade 0.080 oz/st) in the hornfelsed middle section of the Popovich Formation. Gold mineralization discovered to date within Main Mike is concentrated in variably calc-silicated and marbleized, but dominantly decarbonatized, Roberts Mountains Formation. The gold zone at Main Mike (average gold grade 0.037 oz/st) is entirely oxidized.

The gold zone of the Tusc Deposit (Fig. 1) occurs in a setting that is very similar to that at Main Mike, in decarbonatized Roberts Mountains Formation in the hangingwall of the Good Hope Fault and within the fault zone itself at the same stratigraphic elevation. Norby and Orobona (in press) also relate the similarities between the West Mike and Gold Quarry gold mineralization, both occurring in the footwall of the Good Hope Fault and both hosted in the Rodeo Creek Unit into the upper section of the Popovich Formation. The only major difference between the gold zones of these deposits is that the Mike deposit exhibits hornfelsing due to thermal metamorphism, and significant additional copper content, which the other deposits of the Maggie Creek district do not exhibit.

Copper enriched zones are the result of secondary copper minerals, focused in two lobes along the two northwest striking faults, one lobe primarily in the hangingwall of the Good Hope Fault, and the other lobe primarily in the footwall of the Corridor Fault (Figs. 3, 4) (Norby and Orobona, in press). These lobes taper with increasing distance southeast, and their northwestern limit has not yet been defined by drilling. Copper assays on 5 foot intervals of drill core in the copper-enriched zones typically yield 0.2 to 0.6 wt % Cu, up to 3 wt % Cu in the highest grade zones.

Copper mineralization is well developed in both calc-silicate hornfels of the Roberts Mountains Formation and Popovich Formation, and in quartz hornfels of the Rodeo Creek Unit. Dikes within both of the copper mineralized lobes also host copper
Figure 4. Copper zones in plan view.

Plan view extent of copper zones, based on 0.1 wt % Cu cutoff grade, and grade thickness interval equal to or greater than 25 (after Norby and Orobona, 2000). Note the localization of copper along the two northeast-trending faults of the Mike deposit.
minerals. Copper minerals occur along fractures, and as replacements, the latter particularly in the less hornfelsed, more clay-altered areas. Copper zones are typically interlayered with, but do not overlap, oxide gold zones. Newmont reports observations of chrysocolla, copper-bearing montmorillonite, malachite, azurite, turquoise, libethenite, conichalcite, tenorite, cuprite, native copper, copper-bearing limonites, chalcocite, and secondary digenite and covellite (Norby and Orobona, in press). However, only some of these minerals have been positively identified by XRD. Chalcopyrite and bornite are also found in veins deeper in the Mike deposit.

Speculation on Magmatic Sources of Base Metals

Teal and Branham (1997) interpret a magnetic anomaly, approximately 1 kilometer in diameter, lying immediately below the Mike deposit as an apophysis of a larger Mary's Mountain intrusion. They speculate that this was the heat source that drove a thermal metamorphic event causing the hornfelsing and the introduction of copper, bismuth, molybdenum, tungsten, iron, zinc, silver, tellurium, and possibly gold. The authors assume the age of this metasomatic event is Cretaceous. Chalcopyrite and molybdenite-bearing garnet-pyroxene skarn in silty limestone of the Roberts Mountains Formation was encountered in one deep drill hole northwest of Main Mike. Teal and Branham (1997) believe that middle Tertiary supergene processes leached copper vertically down from remnants of quartz base-metal veins observed above the present location of secondary copper minerals.

Upon correcting the Mike deposit magnetic data for the effect of Earth's magnetic poles, a deep igneous pluton was modeled below the northern margin of the Mike deposit (Fig. 5) (Wright, 1998; J. Wright, pers. commun., 1999). The deeper source appears to be a 1.8 km deep, vertically oriented cylinder, which has an airborne magnetic signature nearly identical to that of the roughly 37 Ma old Welches Canyon stock located 3.7 km west-southwest of the Mike deposit (Norby and Orobona, in press; J. Wright, pers. commun., 1999). Norby and Orobona (in press) indicate that the Welches Canyon stock does exhibit limited hornfels development on its margins, but there is no recognized mineralization associated with the intrusion.
Figure 5. Magnetic data vicinity of Mike deposit.

Modelled deep magnetic signature of inferred buried plutons in the vicinity of the Mike deposit.
There is another airborne magnetic anomaly in the area of the Mike deposit (Fig. 5), a protuberance on the magnetic expression of the Richmond stock which extends to within 1.8 km northwest of the Mike deposit (Norby and Orobona, in press; J. Wright, pers. commun., 1999). The Richmond stock, dated at 106±2 Ma, as discussed above, exhibits hornfelsing and base metal mineralization at its periphery (Norby and Orobona, in press). Norby and Orobona argue that the morphology of the hornfels envelope at Mike, opening in a northwestern direction, implies a northern heat source. They invoke a model of a molybdenum-rich intrusion at one of these north-northeastern locations to explain the abundance of tungsten, zinc, silver, and molybdenum in addition to the large amount of copper proven to be in the vicinity of the Mike deposit. Assuming that a porphyry system was responsible for the production of the quartz base-metal sulfide veins, Norby and Orobona (in press) argue that the dominance of sphalerite in veins at the Mike deposit implies their location within the system is distal to the porphyry intrusion.

Norby and Orobona (in press) note a paucity of chalcopyrite within these quartz base-metal veins at the Mike deposit, believing that there would not have been enough copper in the oxidized part of the deposit to account for the amount of copper presently observed at Mike. This observation of rare chalcopyrite in veins, in addition to the shape of the copper lobes open northwest and tapering southeast, has led Norby and Orobona (in press) to conclude the secondary copper is of exotic origin, most likely from a northwestern location, with a flow direction structurally influenced by the Good Hope and Corridor Faults (Fig. 4).

Teal and Branham (1997) describe late supergene alunite cross-cutting secondary copper minerals, placing a minimum relative age on the supergene copper mobilization event. Branham (1994) reports a K/Ar date of 19.7±0.5 Ma from a sample of this alunite. This date for the supergene alunite is significantly younger than three dates of 28.0±0.7 Ma, 27.7±0.7 Ma, and 27.4±0.7 Ma, determined by K/Ar dating of three supergene alunite grains from nearby Gold Quarry (Heitt, 1992).
DISTRIBUTION AND MINERALOGY OF SUPERGENE COPPER

Background Environment and REDOX Boundary

The primary focus of the fieldwork was to determine the identity and spatial distribution of copper minerals, veins, and associated minerals. Some revisions to the Mike deposit stratigraphy emerged from this core logging study, and these revisions are described below. A detailed description of the oxidized zone and its boundary with the unoxidized portion of the deposit also follows in this section.

Redox

The "redox" boundary indicated on Figure 6 and later figures marks the base of oxidation in the weathering profile, the contact between oxide and sulfide portions of the Mike deposit, and is commonly observed as a sharp contact less than 1 cm thick. On the scale of the deposit, the redox is roughly planar, although oxidizing fluids locally permeated fault zones, oxidizing rock to an elevation deeper than the surrounding redox level (e.g. REB-37, Fig. 6). Above this contact, the core is yellow-tan to red and purple color, with the hue dependent on the abundance and particular mixture of iron oxides and the amount of clay that has formed locally. Below the redox, the core is colored pale grey to beige where there are few sulfide minerals, and dark grey to black in regions of abundant sulfide minerals or organic carbon. Rare lenses of unoxidized rock containing relict sulfide minerals are observed above the redox. The redox boundary is used as the spatial reference point for subsequent Eh-pH diagrams illustrating the geochemical conditions governing the formation of supergene copper minerals at the Mike deposit.

Lithology

An extensive geologic logging and reinterpretation program at Gold Quarry during 1999 (in which the author participated) established the gradational contact between the Rodeo Creek and Popovich Formations in the Maggie Creek subdistrict. The contact is placed below the last appearance of clustered, 0.5 cm thick, 2-5 cm diameter, bedding parallel, white lenses that are believed to be phosphate nodules (J. Norby, pers. commun., 1999). This interval in the stratigraphic column is coincident
with the appearance of a pale green, waxy material in many of the Gold Quarry drill holes.

Laminations and subtle lithologic changes in the sedimentary rocks near the Rodeo Creek-Popovich Formation contact at Mike are obscured by hornfelsing, making it difficult to identify the contact. Based on the appearance of the phosphate nodules, and occurrences of the same pale green waxy mineral, the location of the contact is placed about 100 meters stratigraphically below the location shown in the older Newmont geologic logs of West Mike, and as much as 130 meters lower in one hole (REB-89). There is also a marked decrease in the abundance of quartz veins on passing from the Rodeo Creek Unit into the Popovich Formation (See Fig. 6 for the interpreted geology in section from the holes logged for this study.) This new position for the contact increases the thickness of the Rodeo Creek Unit, and decreases the thickness of the Popovich Formation relative to Norby and Orobona (in press) estimates of 320 meters and about 380 meters, respectively. After rectification of the Rodeo Creek Unit-Popovich Formation contact in West Mike, the gold zones of Gold Quarry and West Mike (the link between them made by Norby and Orobona (in press) and described above) are in approximately the same stratigraphic horizon. Including the reinterpreted section, the Rodeo Creek Unit is the host to the copper-enriched zones of West Mike, and to some of the gold zones as well (Figs. 3, 22 and 23, discussed later).

**Lithologic control of the redox boundary**

Lithology may exercise some measure of control over the location of the redox boundary in West Mike. The depth of the redox appears very close to the Rodeo Creek-Popovich Formation contact in some of the drill holes of West Mike, including REB-88 (where the redox and lithologic contact are both faulted) and REB-93 along the fence diagram of Figure 6. The siliceous Rodeo Creek Unit appears to fracture more pervasively than the Popovich Formation, which would enhance Rodeo Creek permeability, permitting greater circulation of groundwater. Outside of intervals of minor faults, the Popovich Formation (all in sulfide zone) is typically also less clay-altered, and less porous than the Rodeo Creek Unit in the oxidized zone above the redox. As a less permeable unit, groundwater may be more likely to pond above the Popovich
Figure 6. Geologic section of Mike deposit.

SW-NE fence diagram of Mike deposit interpretive geology. See Figure 2 for the locations of the drill holes shown here.
dikes, or sills intruded along Drc-Dp contact (?)
dikes, faulted dike margin

KEY

Tc  Carlin Formation
Tint  intrusive dike
Drc  Rodeo Creek Formation
Dp  Popovich Formation
SDrm  Roberts Mtns Formation
Formation, maintaining an elevated water table locally protecting saturated Popovich Formation from penetration by vadose zone oxygen.

The Rodeo Creek-Popovich contact in REB-90 is marked by islands of Popovich Formation interfingered with clay altered, felsic dikes or sills intruded along the contact. The redox boundary in REB-90 lies below the lithologic contact, at the base of the lowest sill, the bottom of a zone of increased fracture abundance, and thus increased permeability, which is spatially associated with the sills.

The lithologic contact dips northeast approaching the Good Hope Fault Zone, as noted by Norby and Orobona (in press), whereas the redox is located at more shallow depths to the northeast (Fig. 6). However, it should be noted that even though the redox is located far above the lithologic contact observed in REB-89, the thickness of oxidized bedrock in hole REB-89 overlying the redox is greater than in any of the other holes logged. Therefore, oxidizing ground waters apparently penetrated deeper into the Rodeo Creek Unit than into any other local formations.

**Veins and Base-metal Sulfides**

Classification of hydrothermal veins yields six recognized types, listed in Table 1 in order of decreasing abundance. The most common are barren quartz veins (Fig. 7A), observed primarily in the Rodeo Creek Unit. Carbonate veins, composed of dolomite ±quartz ±calcite ±ankerite siderite (Fig. 7B) occur in lower parts of the Mike deposit. A third vein type, pyrite-quartz veins, are less common in the sulfide zone, and are marked by analogue hematite + quartz veins in the oxide zone (Fig. 7A). Zones of silica-encapsulated orpiment veins (Fig. 7C) are observed within some sulfide gold zones. Uncommon veins include pyrite-base-metal veins (Fig. 7D), and sphalerite-dominant veins (Fig. 7E). Attributes and distributions of these vein types through the Mike deposit are summarized in Table 1.

Over the course of logging 680 meters of sulfide zone drill core, I encountered at most 20 veins of either the pyrite-base-metal, or sphalerite-dominant vein type. The abundance of these veins increases with depth. No obvious oxidized analogues of either of these two least common type of veins were observed in the oxidized portion of the Mike deposit. However, it would be surprising to find remnants of either of these base-
<table>
<thead>
<tr>
<th>vein type</th>
<th>minerals</th>
<th>size</th>
<th>abundance</th>
<th>location and concentration</th>
</tr>
</thead>
<tbody>
<tr>
<td>barren quartz</td>
<td>quartz</td>
<td>0.1 to 1 mm, rarely over 2 mm</td>
<td>common</td>
<td>most common in hornfelsed Rodeo Creek Unit as abundant small veinlets pervading the core, in other units as 2-4 feet wide concentrations of veins separated by 30-40 feet intervals of very few Barren quartz veins</td>
</tr>
<tr>
<td>carbonate</td>
<td>dolomite + calcite ± quartz ± ankerite-siderite</td>
<td>typically 0.1 to 1 mm, but up to 1 cm</td>
<td>common</td>
<td>observed at the periphery and outside the gold mineralization at the Mike deposit; quartz content decreases and calcite content increases with distance from gold mineralized areas</td>
</tr>
<tr>
<td>pyrite-quartz</td>
<td>pyrite + quartz</td>
<td>0.1 to 1 mm, up to 2 mm</td>
<td>less common</td>
<td>throughout the study area, no recognizable spatial patterns</td>
</tr>
<tr>
<td>orpiment</td>
<td>silica-encapsulated orpiment</td>
<td>2 mm to 6 cm veins in clustered in intervals up to 1.5 m</td>
<td>less common</td>
<td>veins occur intermittently over intervals up to 75 feet thick, associated with gold zones in the unoxidized portion of the deposit</td>
</tr>
<tr>
<td>pyrite-base-metal</td>
<td>pyrite + quartz + sphalerite + chalcopyrite ± galena ± arsenopyrite</td>
<td>typically less than 2 cm, up to 6 cm, with a 0.5-1 cm selvage of sericite and clay</td>
<td>rare</td>
<td>observed stratigraphically lower elevations of the deposit, and apparently restricted to hornfelsed intervals or interfingered unhornfelsed rock; no remnants of any former pyrite-base-metal veins identified in the oxide zones of the deposit; vein abundance increases northward and with depth same pattern of distribution as pyrite-base-metal veins</td>
</tr>
<tr>
<td>sphalerite-dominant</td>
<td>sphalerite ± quartz ± pyrite</td>
<td>0.5 cm, with no additional selvage</td>
<td>rare</td>
<td></td>
</tr>
</tbody>
</table>
Figure 7. Vein photographs.

(A) Numerous thin, barne quartz veins oriented from upper left to lower right. Probable pyrite-quartz veins, now oxidized, oriented sub-horizontally in photo. Sample REB-90-1172. (B) An example of a carbonate vein, sample REB-89-2204. (C) A silica-encapsulated orpiment vein, from sulfide gold zone, REB-88-1765. (D) Pyrite-base-metal vein, containing pyrite + quartz + sphalerite, with other minor sulfide minerals, REB-90-1629.5. (E) A sphalerite-dominant vein, containing dark brown, iron-rich sphalerite ± pyrite ± quartz, sample REB-89-2059.5. Note the visible selvage flanking the pyrite-base-metal vein (Fig. 7D), which is absent from the sphalerite-dominant vein (Fig. 7E).
metal sulfide vein types in oxidized core which is less competent, and commonly broken into rubble, particularly given the low concentration of these veins throughout the sulfide zone intervals. Deep in the sulfide zone of the Mike deposit, presence of a pyrite-base-metal or sphalerite-dominant base-metal bearing veins results in a concentration of base metals of approximately 3000-5000 ppm combined Zn, Pb, and Cu, with Zn commonly at least twice as high as Pb, and with minor Cu. This high base metal concentration in a five foot assay interval hosting a base metal vein contrasts with background concentrations of approximately 300-500 ppm combined base metals.

Due to the limited continuous three dimensional exposure afforded by drill core and the low abundance of pyrite-base-metal and sphalerite-dominant veins, I encountered no intersections of these veins with each other, to enable a determination of relative age relations. However, I did observe one very thin carbonate veinlet clearly crosscutting a pyrite-base-metal vein in thin section. Pyrite-base-metal veins and sphalerite-dominant veins both crosscut intrusive dikes, and are therefore younger than the dikes.

Barren quartz veinlets are a milky grey, fine-grained quartz, and have an irregular, mildly sinuous morphology (Fig. 7A). Carbonate veins in some cases contain very fine-grained euhedral crystals, particularly in the wider veins (Fig. 7B). Carbonate veins are abundant at the periphery and below gold mineralization at the Mike deposit, while essentially no carbonate veins are observed in the core of the gold mineralized areas. The first carbonate veins encountered on approaching the central gold core of the deposit appear to contain considerable dolomite as determined from their reaction to HCl acid, along with some quartz and ankerite as determined by XRD analysis. The quartz content of carbonate veins decreases and calcite content increases with depth, with the deepest carbonate veins observed well below the lowermost gold zone being composed of near pure calcite. The barren quartz and carbonate veinlets appear to decrease in abundance with lateral distance from the core of the Mike deposit, and there is a dearth of either barren quartz or carbonate veins in hole CV-16, located far northwest of West Mike (Fig. 2).

The abundance of barren quartz veinlets varies somewhat dependent on host lithology. Hornfelsed Rodeo Creek Unit observed at West Mike typically contains about 3% quartz veinlets described above, while the quartz veinlets in Main Mike Roberts
Mountains Formation and in lower West Mike Popovich Formation (Popovich Formation still within the area most affected by hydrothermal fluids) are typically 1 meter intervals of abundant veinlets separated by 10-12 meter intervals with very few veinlets. The average quartz contained in these formations is about 0.5%. The increased abundance of quartz veinlets within the Rodeo Creek compared to the other lithologies may be the result of the greater abundance of fractures in the siliceous Rodeo Creek Unit.

The less common pyrite-quartz veinlets (and their oxidized counterparts) also contain very fine-grained quartz, and have a similar thickness and mildly sinuous morphology as the barren quartz veins; however, the pyrite-quartz veinlets appear to be less common than the barren quartz variety, and occur throughout the study area, with no observed pattern of abundance dependent on lithology. Pyrite-quartz veins do not display any recognizable spatial relation to any of the other vein types. In addition to pyrite-quartz veins, dark-colored, fine-grained pyrite is observed in some cases filling fractures.

Veins of silica-encapsulated orpiment, are observed in gold zones of the sulfide portions of the deposit. These anastomosing orange and milky white vein clusters dominate local zones 1-2 meter thick (Fig. 7C) and tend to occur intermittently over intervals of up to 25 meters in drill core. Orpiment veins occur only in the general vicinity of gold zones, although there is no direct one-to-one correlation between orpiment vein abundance and gold concentration, or even a correlation between arsenic and gold concentrations in the assay data.

Pyrite-base-metal veins have a planar attitude with mostly transparent, commonly euhedral quartz, and contain the following sulfide minerals in order of decreasing abundance after quartz: pyrite, Fe-rich sphalerite (averaging 13 mole % FeS, as determined by electron microprobe analysis), chalcopyrite, galena, and rare stibnite, arsenopyrite, marcasite, and molybdenite. Veins are typically less than 2 cm thick, displaying a 0.5-1 cm selvage of sericite and clay. Pyrite-base-metal veins up to 6 cm wide have been observed, although in the few observed cases of veins this wide, there appears to be no significant additional volume of sulfides over the typical 1-2 cm counterpart pyrite-base-metal veins, with the extra volume of the thick veins filled by quartz.
Rare sphalerite-dominant veins, also have a very planar morphology on the scale of core samples. The mineral composition of these veins varies over lateral distance, although on average they are predominantly dark brown, Fe-rich (8-18 mol % FeS) sphalerite, with subordinate pyrite (spl:py ratio of 2:1 or greater), and quartz in well-defined crystals, which fill between 0 to 60% of the vein. One sphalerite-dominant vein in sample REB-89-2059.5 varies from a zone of nearly pure blackjack sphalerite through a 1 cm wide pyrite zone into nearly pure quartz fill over a distance of less than 3 cm. Minor constituents of sphalerite-dominant veins include marcasite, galena, rare chalcopyrite, and rare late calcite. The sphalerite-dominant veins are consistently about 0.5 cm thick, and differ from the pyrite-base-metal veins in that no apparent selvages flank sphalerite-dominant veins, in either hand sample or thin section (Figs. 7D, 7E).

The sphalerite-dominant veins and pyrite-base-metal veins (both types now collectively referred to as "base-metal sulfide veins") are observed in stratigraphically lower portions of the deposit, in intruded dikes and in hornfelsed sections of the deposit. These veins are observed in some intervals that are not hornfelsed, but only where this rock is interlayered with hornfelsed intervals; base-metal sulfide veins appear to be restricted to areas of the deposit which have been subjected to some measure of hornfelsing. These veins are not present in the fringes of the hornfelsing (e.g. hole REB-90, Fig 2) or at depths approaching the limit of hornfels in other holes (e.g. REB-89, REB-93), even below hornfelsed stratigraphic units which contained concentrated base-metal sulfide veins (Fig. 3). Within rock that is moderately hornfelsed, the abundance of base-metal sulfide veins appears to increase to the north (e.g. REB-94) and with depth. There is no evidence in the oxidized portion of the deposit of former base-metal sulfide veins, although they were undoubtedly present. No significant intercepts had been drilled below the oxidized portion of Main Mike at the time of this study, so the presence of base-metal sulfide veins in Main Mike hornfelsed units cannot be addressed.

In a few short intervals below the redox, most notably in hole REB-93, there are clusters of intermingled large sphalerite and pyrite grains, each grain measuring a relatively uniform 0.5 cm wide. These clusters do not show any relation to veins on the scale of the core samples. The sphalerite in the clusters is the same dark brown shade as the sphalerite in the sphalerite-dominant veins, and the proportion of sphalerite to pyrite
in the clusters is also consistent with these sphalerite-dominant veins. A thin calcite veinlet cross-cuts sphalerite in one of these sphalerite grain clusters in REB-93.

Absent Secondary Minerals

Eight separate samples of a material believed during core logging to be jarosite (KFe$_3$(SO$_4$)$_2$(OH)$_6$) were prepared for XRD analysis. The samples were collected from five holes (CV-16, REB-37, REB-68, REB-88, and REB-89). XRD indicated no jarosite was contained within any of these samples, which are instead a mixture of goethite with quartz and kaolinite. One sample may contain alunite. A powdery, olive-green mineral observed in REB-90 was discovered by XRD to also be primarily goethite, in combination with quartz, kaolinite, and additional minor hematite. A dark grey metallic mineral observed during core logging was confirmed by XRD to be hematite.

Suspected scorodite was observed during core logging, and similar to jarosite, the geochemical implications of its formation at the Mike deposit would be significant. The two most plausible scorodite samples and two additional samples of potential As-oxide minerals, were analyzed with XRD. The four samples were determined to be clays, in some cases with a little quartz, but none of the samples appeared to contain any arsenic-bearing mineral phases. No scorodite or any other arsenic-bearing oxide mineral with the exception of conichalcite were identified in the Mike deposit by any analytical methods. A small white lens of suspected phosphate material, collected from the base of the Rodeo Creek Unit near the Popovich Formation lithologic contact in hole REB-90 was confirmed by XRD to be crandallite, CaAl$_3$(PO$_4$)$_2$(OH)$_5$·H$_2$O.

Copper Minerals of the Mike Deposit

Background copper concentration within the oxidized host rock units of the Mike deposit is typically 100-200 ppm Cu, although the copper concentration in core where secondary copper minerals have formed has been measured in excess of 1 wt % Cu. The several copper minerals identified during the course of this study are summarized below (Table 2) and their occurrences described in detail in the subsections that follow. Figure 21 displays the generalized distribution of copper minerals observed between the core logs of five holes (REB-90, REB-88, REB-93, REB-89, and REB-37) which form a
Table 2. Summary of copper minerals of the Mike deposit

<table>
<thead>
<tr>
<th>Mineral</th>
<th>Formula</th>
<th>Basis for identification</th>
</tr>
</thead>
<tbody>
<tr>
<td>Conichalcite</td>
<td>CaCu(AsO₄)(OH)</td>
<td>visual identification, XRD, electron microprobe, SEM/EDS analysis</td>
</tr>
<tr>
<td>Chrysocolla</td>
<td>CuSiO₃·nH₂O</td>
<td>visual identification, SEM/EDS analysis, PIMA, XRD</td>
</tr>
<tr>
<td>Cuprite</td>
<td>Cu₂O</td>
<td>XRD, electron microprobe, SEM/EDS analysis</td>
</tr>
<tr>
<td>Native Copper</td>
<td>Cu</td>
<td>visual identification, XRD, electron microprobe</td>
</tr>
<tr>
<td>Copper-bearing clays</td>
<td>kaolinite, montmorillonite, halloysite</td>
<td>ICP analysis, PIMA, assay correlation, visual identification</td>
</tr>
<tr>
<td>Chalcocite</td>
<td>Cu₂S</td>
<td>assay correlation, visual identification</td>
</tr>
<tr>
<td>Malachite</td>
<td>Cu₂(OH)₂(CO₃)</td>
<td>visual ID, acid tests, electron microprobe</td>
</tr>
<tr>
<td>Chalcopyrite</td>
<td>CuFeS₂</td>
<td>visual and microscope identification, electron microprobe</td>
</tr>
</tbody>
</table>
section across the deposit (Fig. 2). This figure and the mineral distribution patterns which have emerged will be discussed later, after the descriptions of individual minerals.

**Chalcocite, and oxidized chalcocite blankets**

Chalcocite is known to form a sub-horizontal "blanket" morphology below the base of copper leached during the oxidation of primary copper sulfide minerals. Chalcocite precipitates at an elevation in the weathering profile where reducing conditions are low enough that it becomes stable, typically thought to be at the top of the water table and below. Chalcocite replaces pre-existing sulfide grains at these elevations below the water table (Spencer, 1913; Ransome, 1910). Chalcocite replacement of pyrite at the Mike deposit (pictured in Fig. 8A at left, and Fig. 8B at right) is visible in REB-88 in a nearly 12 meter thick, unoxidized, relic chalcocite blanket above the redox.

Chalcocite also is present in some drill holes immediately below the redox boundary, although this core is often rubble, and black colored due to carbon in the rock, which obscures recognition of chalcocite during core logging. The reason that chalcocite is suspected to be present at this location is due to the Newmont chemical assay data from the redox boundary intervals. Chalcocite is the only copper mineral normally associated with supergene enrichment profiles which typically yields a near complete cyanide-leach assay for copper, but also a very poor (about 5% or less) acid-leach assay for copper, both relative to the total copper assay, thus making the chemical assay signature of chalcocite very distinctive. Redox level chalcocite blankets in the Mike deposit are typically marked by copper grades in excess of 0.3 wt % Cu for only a 5 feet thick assay interval, although in some cases, such as in hole REB-91 (Fig. 24), a 10 foot thick interval grading over 1 wt % Cu is present.

Chalcocite forms botryoidal grains replacing pyrite. On the scale of hand samples, these botryoidal grains typically congregate either as aggregate chalcocite which clusters into groups of 2-4 mm diameter spheres which commonly merge into each other (Fig. 8A), or as more dendritic accumulations up to about 1 cm across (Fig. 8B). Oxidized hematite analogues of the original chalcocite replacement textures display the same size and shape characteristics of their sulfide precursor clusters, as clearly illustrated in the photographs of Figure 8. For this reason, the approximately 25 oxidized chalcocite blankets noted in the core holes logged are relatively easy to identify from the
Figure 8. Chalcocite and oxidized chalcocite.

A comparison of chalcocite textures with oxidized analogues of hematite in the same morphology. Samples collected from REB-88. (A) Rounded clusters of chalcocite from 1247' below the surface, and hematite from the overlying oxidized portion of the chalcocite blanket at 1210' deep. (B) Dendritic, botryoidal chalcocite blossoms at center bottom of sample at right, from 1248', and left sample exhibiting hematite of same morphology from 1166'.
striking red color of hematite replacement of the exact chalcocite morphological textures. This particular hematite morphology indicates the former presence of pyrite within the chalcocite blanket, prior to chalcocite replacement rims forming on those pyrite grains, whether or not the pyrite grains were completely replaced. The dearth of drill holes over the area of the Mike deposit precludes a reliable correlation between holes of individual, distinctly separate, now oxidized chalcocite blankets.

Most of the oxidized chalcocite blankets of the Mike deposit, with the exception of the very thick sequence in REB-88, are from 10 cm to 1.5 meters thick. In most cases, the chalcocite blankets at the redox boundary also appear to be about this same thickness. There is a constant character of the rock in which both chalcocite blankets and their oxidized analogues are observed: in all cases the rock has been completely decarbonatized, but shows no silicification, a common style of alteration in Carlin-type deposits. Most of the chalcocite blankets are also in rock which has been logged on a relative alteration scale by Newmont geologists as exhibiting the maximum level of alteration to clay. Although the present alteration in the rock could be due in large measure to the secondary processes of copper enrichment, rather than the Carlin-type gold introduction event as logged by Newmont geologists (a topic which will be addressed in the discussion section), it is interesting to note this striking consistency of alteration type in rock where chalcocite blankets occur.

Intervals of oxidized chalcocite blankets contain significantly less copper than their precursor chalcocite blankets. With one very notable exception of the 21 meter thick horizon of oxidized chalcocite observed in REB-88 immediately above the remnant chalcocite blanket in that hole, copper concentrations in the oxidized chalcocite blankets do not exceed 0.2 wt % Cu, and are typically only slightly higher than local background copper concentrations. The underlying reasons for the exceptionally high copper grade associated with the particular oxidized chalcocite blanket from REB-88 will be addressed below in the discussion section.

The degree of chalcocite replacement of sulfide minerals, as documented by the concentration and distinctive form of hematite textures, appears to vary between different oxidized chalcocite blankets. There does not appear to be any correlation between the
extent of original chalcocite replacement to any other alteration type or lithologic features of the core, or to the presence or absence of any particular surrounding minerals.

There is great variability in the thickness, degree of original chalcocite replacement, elevations, and number of distinct oxidized (or relic) chalcocite blankets observed in each logged drill hole. No strong inference can be drawn about the lateral extent of these blankets from the hole where they were observed due to the dearth of core holes drilled in the deposit, as discussed earlier. Chalcocite blankets, and their oxidized analogues, may also be laterally discontinuous across the deposit, occurring specifically only where geochemical conditions were suitable (as outlined above) and pre-existing sulfide minerals present, to precipitate the chalcocite. For example, there are abundant and obvious oxidized chalcocite blankets in REB-88, whereas there are only a few minor ones in REB-93 and REB-89, and only weak, incompletely oxidized chalcocite blankets remaining in REB-90 (Fig. 21). This abundance of original chalcocite replacement in REB-88 is not associated with the Valley Fault, a very late feature, as discussed in the geological history section, which apparently cross-cuts the redox boundary.

There is also an issue of incomplete data concerning the distribution of chalcocite blankets throughout the deposit. The type of drilling will exert significant bias over the locations where chalcocite, and particularly oxidized chalcocite blankets, will be recognized and logged. Observation of solid drill core makes positive identification of oxidized chalcocite from the hematite morphology possible, whereas observation of drill chips, the product of reverse circulation drilling, obscure recognition of oxidized chalcocite blankets. Each small sample (on the order of 50-75 grams of material) of chips collected and archived during reverse circulation drilling are taken from an entire five-foot sample interval, a very small and scattered amount of material from a significantly larger original volume, meaning that any horizontal layers of an oxidized chalcocite blanket will be diluted, destroyed, or missed altogether. Individual drill chips may consist of hematite, which is very common in the oxidized portion of the Mike deposit, but rock chips will not be large enough to clearly discern characteristic hematite blossoms remaining after the oxidization of a former chalcocite blankets.

Copper assay data can assist recognition of chalcocite blankets, as described above, but there is no way to recognize oxidized chalcocite textures without visual
recognition. Note that most reverse circulation, chip-drilled sections of drill holes are from higher elevations of the deposit, less likely locations for a geochemically reducing environment to preserve chalcocite replacement, and no remnant chalcocite blankets were identified in chip drilled intervals from copper assay data.

After drilling through the thick cover of Carlin Formation, many of the drill holes were continued as chip-drilled holes deeper into bedrock before switching to core drilling for the remainder of the hole. As a result, many of the holes which were logged in detail for this study did not have a complete section of core. (See the "top of core drilling" boundary marked on all the holes of the Figure 21 section through the deposit.) Core drilling in REB-90 was not initiated until deep within the copper zones of the hole, and core drilling in REB-89 only began at the very top of the copper zone. An oxidized chalcocite blanket is present at the very top of the core from REB-89, and another oxidized chalcocite blanket, exhibiting thorough replacement of the original chalcocite is now observed near the top of core available from REB-88. The two older holes of the Mike deposit that were logged in detail, REB-37 and REB-68, were core drilled for their entire length through bedrock, and both of those core holes display numerous oxidized chalcocite blankets, including some near the top of bedrock. These observations lead me to believe that there are additional oxidized chalcocite blankets in the deposit through the chip-drilled, uppermost oxidized portion of the deposit, even though none could be recognized and noted on Figure 21.

**Conichalcite**

Conichalcite, CaCu(AsO$_4$)(OH), is perhaps the single most common copper oxide mineral in the Mike deposit. Although conichalcite appears to account for much of the copper contained in the oxidized zone of the deposit, there does not appear to be any correlation between the presence of conichalcite and higher or lower copper grade. Conichalcite occurs alone or is commonly observed with chrysocolla and with copper-bearing clays. Conichalcite is also common within the base or immediately below oxidized chalcocite blankets. Most descriptions of localities containing conichalcite describe an association with chrysocolla (e.g. Bogoch et al., 1994; Crowley, 1975; Gillson, 1926), and one very detailed deposit description, from the Dome Rock Mine in Australia, describes the association of chrysocolla and copper-bearing clays with
conichalcite (Ryall and Sengit, 1976). There is no mention of the presence of oxidized chalcocite blanket textures at any of these conichalcite localities.

Conichalcite at the Mike deposit occurs in variable shades of olive to dark green, as illustrated by the examples in Figure 9. On hand sample scale, conichalcite forms botryoidal clusters of spheres up to 0.5 mm diameter (Fig. 9A). In locations where conichalcite is less abundant, spheres are more perfectly spherical (Fig. 15A), while more concentrated occurrences of conichalcite form imperfect spheres in botryoidal clusters due to impingement (Fig. 10).

Conichalcite spheres are not comprised of concentric growth bands, as if they were formed from a colloidal suspension. The conichalcite spheroids are actually composed of a collection of radiating acicular crystals, each crystal less than 10 microns by 10 microns in width, but over several tens of microns in length (Figs. 10, 15). The individual bladed crystals are rhombohedral and appear to decrease in width towards the centers of the conichalcite clusters (illustrated in Fig. 10B). There appears to be a minimum diameter of about 100 μm for a concentration of individual conichalcite blades to aggregate into spheroidal shapes (e.g. Figs. 10A and 15A). Any smaller clustering of the acicular conichalcite crystals forms an irregular collection of the individual crystals (e.g. Figs. 11, 15B). Although isolated, single, bladed crystals of conichalcite have been reported (Bogoch et al., 1994), I did not observe such crystals in samples from Mike. Descriptive studies of conichalcite at Dome Rock Mine, Australia (Ryall and Sengit, 1976), and at the Zinc Hill Mine, California (Crowley, 1975), document only botryoidal accumulations of spheres of acicular crystals of conichalcite; however, the California locality does not appear to have been studied using an SEM. SEM analysis from the Mike deposit samples also identified tiny, 5 μm rosettes of likely arseniosiderite (Ca$_3$Fe$_4$(AsO$_4$)$_4$(OH)$_6$·3H$_2$O) attached to the outer ends of conichalcite crystals.

Though it appears to be texturally homogenous in hand samples, conichalcite is chemically variable. The BSE (backscattered electron) image in Figure 11 illustrates one example of this variability. The bright yellow to medium orange colored, irregular grain at the center of the image is conichalcite, but the brightest areas (higher average atomic number) contain up to 13 weight percent lead, while the darker areas (lower average atomic number) contain 1.5 wt % Pb. Table 3 presents quantitative electron microprobe
Figure 9. Conichalcite photographs.

Botryoidal conichalcite samples, all confirmed by XRD and/or SEM analysis, exhibiting variable color and grain size. (A) Dark green crusts on conichalcite from REB-88-1154. (B) Bright green botryoidal spheres of conichalcite, precipitated on fracture with sky blue chrysocolla, from 24 feet deeper in hole, from REB-88-1178. (C) Fine-grained, army green, fracture-coating of conichalcite intermixed with minor quartz, sample from REB-89-1134.5.
Figure 10. SEM images of conichalcite texture.

(A) Botryoidal spheres of conichalcite grown on a fracture surface from REB-88-1178, the same sample photographed in Fig. 9B. (B) Magnified view of a cut surface through intergrown conichalcite spheres showing the individual acicular crystals of which the spheroidal growths are composed. Sample from REB-88-1154, the same sample photographed in Fig. 9A.
Figure 11. BSE image depicting heterogeneity of conichalcite

Backscattered electron image of conichalcite from electron microprobe study of sample REB-89-1054. Brightness correlates with average atomic number, darker areas representing relative lower atomic number for the material in that area. Note the heterogeneity of the conichalcite pictured here. Bright areas were measured by microprobe analyses (Table 3) to have incorporated up to 13-17% Pb substitution for calcium in the conichalcite structure. Black background material is composed of silicate minerals.
Table 3. Selected microprobe analyses of conichalcite from REB-89-1054.

<table>
<thead>
<tr>
<th></th>
<th>Average-higher at. #</th>
<th>Average-lower at. #</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>areas of higher average atomic number</td>
<td>areas of lower average atomic number</td>
</tr>
<tr>
<td></td>
<td>Analysis</td>
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</tr>
<tr>
<td>FeO</td>
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</tr>
<tr>
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<td>36.63</td>
</tr>
<tr>
<td>ZnO</td>
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<td>&lt;d.l.</td>
</tr>
<tr>
<td>SiO2</td>
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<td>0.35</td>
</tr>
<tr>
<td>Totals</td>
<td>95.95</td>
<td>98.24</td>
</tr>
</tbody>
</table>

Formula coefficients balanced on 5 Oxygen atoms and 1 H atom (for adelite group arsenates)

<table>
<thead>
<tr>
<th></th>
<th>Average-higher at. #</th>
<th>Average-lower at. #</th>
</tr>
</thead>
<tbody>
<tr>
<td>Fe</td>
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<td>0.073</td>
</tr>
<tr>
<td>Pb</td>
<td>0.137</td>
<td>0.152</td>
</tr>
<tr>
<td>P</td>
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<td>0.052</td>
</tr>
<tr>
<td>As</td>
<td>0.920</td>
<td>0.918</td>
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<tr>
<td>Zn</td>
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<td>NA</td>
</tr>
<tr>
<td>Cu</td>
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<td>0.962</td>
</tr>
<tr>
<td>Ca</td>
<td>0.871</td>
<td>0.852</td>
</tr>
<tr>
<td>Si</td>
<td>0.016</td>
<td>0.017</td>
</tr>
<tr>
<td>O</td>
<td>5.000</td>
<td>5.000</td>
</tr>
<tr>
<td>H</td>
<td>1.000</td>
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</tbody>
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Detection limit for Zn = 0.05 wt %
NA = not applicable
analyses of conichalcite, separated into areas of higher and lower average atomic number within grains. Some analyses for zinc were below the electron microprobe detection limit of roughly 0.05 wt % for 20% error in the measured elemental amounts (B. Jones, pers. commun., 2001). Formula units are calculated on the basis of five oxygen atoms and one hydrogen atom (Table 3).

Considerable atomic substitution is possible within the arsénate group minerals, as noted by a number of researchers. Jambor et al. (1980) list members of the adelite mineral group which are believed to be isostructural, including conichalcite, duftite (PbCu(AsO_4)(OH)), adelite (CaMg(AsO_4)(OH)), austinite (CaZn(AsO_4)(OH)), and gabrielsonite (PbFe(AsO_4)(OH)), all of which are variants on the conichalcite mineral structure with substitution of Zn, Fe, and/or Mg for Cu, and Pb for Ca. Their study of arsénate solid solution provides a basis from which to understand the variability between microprobe analyses of conichalcite (Table 3). Plotting the calculated formula concentration of atoms illuminates some patterns of elemental variation between separate analyses (Figs. 12 and 13).

A plot of calcium versus lead concentration (Fig. 12), demonstrates a strong inverse correlation, implying that the lead present occupies the calcium sites of conichalcite, thus forming a partial solid solution of conichalcite with duftite. All of the Pb analyses are above the electron microprobe detection limit for Pb. The partial totals of the normalized formula concentration of Ca and Pb sum to nearly one, as would be expected if Pb, and no other metals, occupy Ca sites.

Figure 13 illustrates the increase of iron concentration with increasing lead concentration in the conichalcite, with the lead-rich areas of conichalcite containing between one and two percent FeO, with over 2% FeO in one analysis (Table 3). This correlation in Fe implies that there may be a very minor component of gabrielsonite, or even some yukonite, a rare Ca-Fe arsénate (Pieczka et al., 1998), in the lead-rich areas of the arsénate. It is interesting to note that although the Mike deposit contains much more zinc than lead, very little zinc was detected (Table 3). Zinc as austinite is present at numerous other localities.

The formula atoms of Cu, Zn, and Fe, potential occupants of the Cu site in conichalcite, fail to sum to one. The incomplete totals from the electron microprobe
Figure 12. Ca-Pb formula concentration in conichalcite.

Plot of Ca vs. Pb formula concentration in electron microprobe quantitative analyses of conichalcite from REB-89-1054. The plot illustrates an inverse correlation between calcium and lead in conichalcite from the Mike deposit. The line marks the position of an ideal one-to-one substitution of lead atoms into the calcium atom site in conichalcite mineral structure. All of the analyses were measured above the 0.05 wt % detection limit for Pb, which would be 0.0006 formula atoms for Pb in conichalcite.
Ca vs. Pb formula concentration

![Graph showing the relationship between Ca and Pb formula atoms concentration. The graph includes a line of best fit and data points.](image-url)
Figure 13. Pb-Fe formula concentration in conichalcite.

Plot of Pb vs. Fe atomic concentration in electron microprobe quantitative analyses of conichalcite from REB-89-1054. The plot illustrates increasing iron concentration with an increasing lead concentration. All analyses were above detection limits for Fe and Pb, which in conichalcite structure translate into 0.002 for Fe, and 0.0006 for Pb.
Pb vs. Fe formula concentration

![Graph showing the concentration of Pb vs. Fe in formula atoms](Image)

- X-axis: formula atoms of Fe
- Y-axis: formula atoms of Pb

The graph indicates a positive correlation between the concentration of Pb and Fe in the formula.
analyses in Table 3 may be due to magnesium, which could not be quantitatively analyzed given the array of elements selected for analysis. Mg could potentially occupy the Cu site of conichalcite, and may account for material missing from the analyses. Likewise, formula atoms of arsenic do not total one, implying complete arsenate in the conichalcite analyses from the Mike deposit. Ryall and Segnit (1976) note that vanadium, phosphorus, and also silicon and aluminum have replaced arsenic in conichalcite from the Dome Rock Mine, and a similar substitution of these components for arsenate may be occurring in the Mike deposit conichalcite. However, overall totals from microprobe analyses are quite good, indicating relatively pure conichalcite, other than formation of limited solid solution with duftite as noted above.

**Chrysocolla**

Abundant in the Mike deposit, chrysocolla (CuSiO₃·nH₂O, with possible Al₂O₃ incorporated) varies between deep-blue to pale-blue in color, though the color is usually homogenous within individual samples (Figs. 14A, B). Chrysocolla was recognized during logging by its blue color, correlation with copper assays, and from the observation that chrysocolla sticks to the tongue. No greenish color was observed in chrysocolla samples from the Mike deposit, although pale green and blue-green varieties of chrysocolla are noted at other localities (e.g. Ryall and Segnit, 1976; Sun, 1963). Experiments by Newberg (1967) imply that more greenish color may reflect the extent of dehydration of the chrysocolla.

Due to the substantial color variability of material believed to be chrysocolla, including some which resembled turquoise (Fig. 14A), samples were analyzed using EDS to confirm their identity as chrysocolla. These qualitative analyses indicated that elements of higher atomic number consisted primarily of copper and silicon, with minor aluminum (EDS has a detection limit for Al of approximately 2-3 %). XRD analysis of chrysocolla consistently provided poorly defined spectra, as expected for amorphous materials. PIMA results proved to be more conclusive than XRD. The samples analyzed with PIMA were compared to one sample which is known to be chrysocolla from its visual appearance, SEM textures and EDS results. The PIMA spectra, including that from the known sample, were convincingly similar to each other, and also a reasonably close match to the USGS library PIMA reference spectrum for chrysocolla.
Figure 14. Chrysocolla photographs

Variably colored chrysocolla: (A) Deep blue variety from REB-89-1253.5. (B) More commonly observed, light colored sky-blue chrysocolla, a thick accumulation on a fracture in sample REB-88-1192. (C) Conichalcite and chrysocolla precipitated on the same fracture, but formed in mutually exclusive areas, from REB-93-1461.5.
Chrysocolla was commonly observed alone (Figs. 14A, B), but also in contact with copper-bearing clays, and in some cases with other minerals such as malachite. In one sample (Fig. 14C), a large chrysocolla crystal was found, indicating the presence of some researchers. Further studies are needed to clarify the role of chrysocolla in this context.
No particular textures such as the formation of growth bands were observed on the scale of hand samples in chrysocolla from the Mike deposit. SEM imaging of chrysocolla revealed some desiccation cracks, and also apparent layers within the chrysocolla which appeared to have contracted to differing degrees, implying that these differentially contracted layers formed successively as growth bands of slightly different chemical composition or degree of hydration. This observation of microscale colliform structure in chrysocolla coincides with the observations of chrysocolla at Chuquicamata by Newberg (1967) and is compatible with the suggestion that chrysocolla forms from a hydrogel (as reviewed by Sun, 1963). These desiccation features also imply that the material was a hydrate, and thus provide textural evidence to support the argument that this blue material from the Mike deposit is chrysocolla.

Chrysocolla was commonly observed alone (Figs. 14A, B), but also in contact with copper-bearing clays, and in some cases proximal to malachite, though there were too few samples of malachite to be able to evaluate this mineral assemblage. Conichalcite and chrysocolla were commonly observed together on fractures (Fig. 15), but in some places they are separated into mutually exclusive areas (Fig. 14C). Chrysocolla was not observed in contact with or near chalcocite or native copper, but was observed near cuprite in drill core from hole REB-88.

**Timing relations between conichalcite and chrysocolla**

SEM analysis of samples containing both conichalcite and chrysocolla together clarified some of the timing relationships between these two coexisting minerals (Fig. 15). In none of the samples containing both minerals was conichalcite observed attached to wallrock, whereas chrysocolla commonly is, implying that conichalcite is younger than some chrysocolla. In one sample (Fig. 15B), chrysocolla is attached to a conichalcite crystal, indicating that chrysocolla continued to form after conichalcite. Some researchers provide details of the order of formation between coexisting chrysocolla and conichalcite at other localities. Both at Dome Rock (Ryall and Segnit, 1976) and at Zinc Hill Mine (Crowley, 1975), conichalcite coats chrysocolla, and is interpreted to be younger. Bogoch et al. (1994) also report that conichalcite is younger than chrysocolla at a prospect in southern Israel, although there is also an SEM image in their paper (p. 317)
Figure 15. SEM images of conichalcite with chrysocolla

SEM images from a fracture containing both conichalcite and chrysocolla, from sample REB-88-1178, the same sample location also pictured in Figures 9B and 10A. (A) A sphere of conichalcite crystals, surrounded by a matrix of chrysocolla. (B) Smaller cluster of radiating acicular conichalcite crystals, also surrounded by chrysocolla. Note the attachment of chrysocolla to a conichalcite crystal at the front upper-left portion of the cluster, indication that this chrysocolla is younger than the conichalcite.
showing some chrysocolla after conichalcite, similar to the evidence in Figure 15 from the Mike deposit.

**Copper-bearing clays**

Some clays observed in the Mike deposit core are pale green to pale blue in color, relative to the typical white clays more commonly observed in the core (Fig. 16A). Suspected copper-bearing clays, logged from visual appearance and correlation with assay data, were confirmed by ICP assays. These copper-bearing clays typically only occur over 15 cm to 1.2 m of core length, but in one location from REB-88, near the base of a 21 meter thick sequence of oxidized chalcocite, two copper-bearing clay lenses in close proximity to each other comprise 5 m of a 6 meter interval of core (Fig. 21). In some locations, concentrations of colored clays are accompanied by conichalcite, chrysocolla, and hematite, possibly with minor cuprite, but in most occurrences, the suspected copper-bearing clay contains no evident copper oxide or other minerals. These colored clays are among the most permeable material in the drill core. Many of these intervals of clay concentrations with high copper assays over 1 wt % Cu also contain anomalously high zinc assays between 0.1 and 0.5 wt % Zn.

Most of the copper-bearing clays occur proximal to oxidized chalcocite blankets (Fig. 21). Assays from 5 foot intervals in such areas consistently show high copper concentration, typically about 1 wt % Cu. Whether or not they are also associated with oxidized chalcocite blankets, intervals of copper-bearing clays containing lower copper concentration (less than 0.6 wt % Cu) commonly contain more copper than adjacent assay intervals.

Many samples of clay from intervals of anomalously high copper assays, some of which also yielded high zinc assays, were analyzed by XRD. Results indicate that most of these samples are composed of kaolinite, some with a mixture with quartz, goethite, hematite and other unidentified mineral species. One of the samples from hole REB-90, collected from a five foot assay interval yielding approximately 0.5 wt % of both copper and zinc, was identified by XRD to be primarily montmorillonite, likely mixed with other clays. The PIMA was then used to corroborate XRD results and to analyze additional samples. The PIMA measurements led to limited success in samples with low clay content. In one sample from an assay interval containing >1.8 wt % Cu and >1.5 wt %
Figure 16. Copper-bearing clay, and malachite, photographs

(A) Example of pale greenish-blue copper-bearing clay from REB-88-1154.5, with metal concentrations given in the text. More strongly green-colored areas in the clay-filled breccia matrix pictured here contain small grains of conichalcite. Note the thin ribbon of chrysocolla filling a fracture through the clay along the left edge of the breccia matrix. Other areas of concentrated blue within this image are artifacts of the scanning and processing. (B) Emerald-green malachite with a concentric growth habit on a fracture in sample REB-90-1435.
Zn in REB-90, PIMA analysis suggested the presence of sauconite, a Zn-montmorillonite.

Many of the indefinite XRD analyses of suspected copper-bearing clays from assay intervals of hole REB-88 which contained high copper concentrations (but no anomalous zinc concentrations) were identified as primarily halloysite by the PIMA. Copper-bearing halloysite clay has been described from the Dome Rock deposit in South Australia, associated with copper oxide minerals (Ryall and Segnit, 1976). These authors also cite a Russian study of naturally occurring copper-bearing clays from deposits in the USSR, which were determined to be montmorillonite with some halloysite.

Three hand-picked clay samples were sent to Chemex Labs for ICP analysis to determine their chemical composition. One light blue colored clay sample from REB-90 of primarily montmorillonite, with PIMA analysis indicating possible minor kaolinite, contains 1.14 wt % Cu, in addition to 1.13 wt % Zn, and slightly over 1% Ca, but no Pb or other unexpected metal cations. A second sample of very pale blue clay from REB-90, identified as predominantly kaolinite by both XRD and PIMA, contains 0.44 wt % Cu, with lesser zinc (0.12 wt %) and also some magnesium (0.32 wt %).

PIMA analysis of a third, pale green sample of clay from REB-88 (the example shown in Figure 16A) suggested halloysite as the primary clay mineral, although the sample also contains some very fine-grained conichalcite, and potential minor cuprite. ICP analysis of material that was hand-drilled from this sample indicated 1.08 wt % As, in addition to 2.13 wt % Cu. The sample contained only minor Zn and Pb, 0.04 wt % and 0.05 wt %, respectively. This arsenic is accounted for by the included conichalcite, given that As substitution into clays is not a widely documented phenomenon. Subtracting the stoichiometric equivalent amount of copper which would be in this conichalcite results in 1.21 wt % Cu contained in the colored clay. Under 20x magnification, no cuprite or chrysocolla was visible within the clay sampled with the dental drill.

**Malachite**

Malachite is scarce at the Mike deposit, found only in a few locations, typically at the top and fringes of the oxide copper zone. The nine malachite zones identified during core logging are each less than a half meter thick, and most are only 10 cm thick. Malachite forms concentric bands of emerald green color on fracture surfaces (Fig. 16B),
and can be distinguished from conichalcite by its vigorous fizz on contact with HCl. Four of the malachite localities in the Mike deposit occur with copper-bearing clays, and the other five malachite localities are near copper-bearing clay. Malachite has also been observed in contact with cuprite, and associated with chrysocolla, but was not in contact with conichalcite. These malachite zones commonly occur at shallow depths in the oxide copper zone, within conichalcite mineral zones, but are observed in thin pods everywhere flanked by chrysocolla or copper-bearing clay which form a transition zone between the malachite and conichalcite. (Malachite zones are too small in size to appear in Figure 21 given the scale of the section.)

Native copper

Native copper is observed as shiny, dendritic copper leaf along fractures, particularly at margins and deeper within the copper zone (Fig. 21, images in Figs. 17B and 17C). Abundant native copper leaf was found in REB-90, CV-16 (Fig. 2), and REB-40, located 130 meters north of REB-37 on Figure 2. Cuprite is associated with most native copper, commonly in greater abundance than native copper, where the two minerals are observed together. At lower elevations of the oxide zone in REB-90, and throughout the copper-bearing interval of hole CV-16, native copper predominates over cuprite (Fig. 21). Native copper is only observed in wallrock that contains little or no clay alteration.

Though it is typically visible as dendritic leaf along fractures, some of the native copper (and all of it in hole CV-16) is too fine-grained to be recognized with a 20X hand lens. Material from a few intervals containing no visible copper minerals, but copper grades greater than 0.25 wt % Cu, was confirmed to contain native copper by XRD. Native copper produces a clear XRD spectral pattern consisting of only two distinct peaks, making it easy to identify even in a sample with a very low concentration of native copper mixed among other minerals. All of the samples that contain very fine-grained native copper are powdery and have a greenish-blue tarnish, with quartz, clay, and trace K-feldspar comprising the bulk of the sample. Small amounts of native copper are also present as islands on the order of a few microns in diameter within larger clusters of cuprite observed with the electron microprobe.
Figure 17. Native copper, and cuprite, photographs

(A) Abundant red cuprite precipitated on fracture in REB-90-1161. (B) Thin cuprite filling micro-fractures, surrounding black chalcocite clusters at top of sample. Note thick leaf of native copper along right side of this sample (black area adjacent to the copper leaf is a shadow). Sample from REB-88-1225.5. (C) Dendritic native copper precipitated on open fracture from REB-90-1435.5. The yellow-brown material is goethite.
**Cuprite**

Although not observed in great abundance in core, cuprite has been identified in several different textural forms during petrographic study. It is present in contact with or near all other copper minerals in the oxidized copper zone of the deposit. Transmitted light microscope study revealed cuprite up to 80 μm across filling spaces in malachite-filled fractures from hole REB-90. Electron microprobe analysis confirmed the identity of these minerals. However, the bulk of the cuprite is associated with native copper or hematite. In hand sample, aggregates of crimson red cuprite crystals, as confirmed by XRD, and SEM/EDS analysis, are observed on fractures (Figs. 17A, 18) in hole REB-90 (Fig. 21). Isolated aggregates of cuprite, smaller than the accumulation shown in Figure 17A, are most commonly observed with native copper, with cuprite being more common than native copper at the higher elevations of the hole, as shown on Figure 21. Cuprite is also observed filling micro-fractures, surrounded by chalcocite and adjacent to native copper at the upper part of a relic chalcocite blanket (Fig. 17B). In a supergene enrichment profile actively forming at the Pleysumi prospect in Papua New Guinea, Titley (1978, p. 781) reports a texture similar to the sample in Figure 17B: "In the upper part of the old blanket, chalcocite is being converted to native copper..... It appears that native copper is subsequently converted to cuprite."

Cuprite has also been detected with hematite in some areas of the Mike deposit, commonly within conichalcite and chrysocolla dominant zones, although the cuprite has not been observed in contact with conichalcite or chrysocolla. Some intervals of core assaying up to 0.6 wt % Cu, contain no obvious copper minerals, only abundant hematite, and it was estimated that red-colored cuprite could occur among the hematite. Several such samples collected from all the holes except REB-93 and REB-68 were analyzed by XRD, and four of them showed evidence of cuprite, and in one sample native copper, in a mixture of hematite and quartz, with minor clay. There may be very limited cuprite in some other samples, too little to be detected by XRD.

Electron microprobe analyses of a sample from such a copper-bearing zone with only abundant hematite visible in hand sample (hole REB-90) revealed that many porous cuprite clusters, as large as 100 μm across, are enclosed in abundant hematite (Fig. 19).
Figure 18. SEM image of cuprite crystals

Crystals of cuprite from sample REB-90-1161, the sample photographed in Fig. 17A.
Figure 19. BSE image of cuprite with hematite

Backscattered electron image of cuprite (bright area) surrounded by hematite (orange color), from electron microprobe study of sample REB-90-1157.5. No direct contact intermixing between cuprite and iron oxide minerals was observed in any samples, even on the micron-scale, as pictured here.
Cuprite is generally near hematite, but not in contact with hematite or with any other iron-oxide minerals.

**Copper minerals not observed at the Mike deposit**

Additional copper minerals have been listed in Newmont company documents as being present in the Mike deposit, including tenorite, libethenite\((\text{Cu}_2\text{(PO}_4\text{)(OH)})\), azurite\((\text{Cu}_3\text{(CO}_3\text{)}_2\text{(OH)}_2\text{)}\), and turquoise. However, none of these minerals were identified during the present study. A few samples of material believed to be turquoise \((\text{CuAl}_6\text{(PO}_4\text{)}_4\text{(OH)}_8\cdot\text{4H}_2\text{O})\) were found to be a darker-blue variety of chrysocolla (Fig. 14A) based on PIMA analysis.

Numerous occurrences of black, fine-grained surface coatings of manganese-oxide were observed during core logging. Contrary to company reports suggesting relatively common occurrence of tenorite \((\text{CuO})\) in the Mike deposit, no tenorite was identified in samples which resembled tenorite. Careful assay correlation and copper plating tests performed during core logging provided no evidence of tenorite, nor did any follow-up XRD and electron microprobe analyses of collected samples. These minerals may simply be very rare in the Mike deposit and not encountered in any of the areas studied during this project.

**Association of Base-metals with Iron Oxyhydroxides**

Some assayed intervals of the oxidized portion of the Mike deposit contain concentrations of well over 0.1 wt % copper and 0.1 wt % zinc (with some intervals containing over 1% of either of these metals) with only brown to rust-red colored, powdery iron-oxide material visible with 20x hand lens. No zinc oxide minerals were identified, in spite of rigorous examination of assayed intervals containing one percent zinc; this examination included liberal sprinkling on the core with "zinc zap", a solution specifically formulated to color any zinc oxide minerals bright orange upon contact. Newmont conducted extensive metallurgical assay work on the base-metal rich iron oxides, and determined that the zinc present in these core intervals was only retrievable from the iron oxides by total acid digestion, which completely dissolves all material including iron oxide minerals. This particular zinc was not retrievable by any less extreme leach methods which are typically effective for zinc minerals (S. Acar, pers.)
commun., 2000). These tests imply that the zinc (along with other leach-tested base metals in these iron oxides) is not contained as very fine-grained oxide minerals mixed with iron-oxides, but instead is somehow more intimately associated with the iron oxides.

One two-foot assay interval of core with 8900 ppm Zn contained a distinct powdery orange ribbon of material amidst weathered, pale red, hematite-coated small pieces of core. A sample of the orange material from a 2 cm coherent piece of core was scraped off and sent to Chemex Labs for ICP analysis. The bulk sample was also sent for ICP analysis to determine if the zinc in this interval was segregated into the orange material. The orange material contained 8550 ppm Zn, whereas the bulk sample after removal of the orange material contained 9190 ppm Zn. The two ICP analysis results were extremely similar in composition, with the bulk rock containing about 1% more Al and K than the orange material, but all other measured components of the ICP analysis, including Cu, As, Ba, Ni, and Pb, were nearly equal. This result implies that the base metals are present in a relatively homogenous distribution in iron oxide minerals coating the rock and within the rock matrix.

Electron microprobe studies of two samples provided further insight into the association of base metals with iron oxides. Figure 20 illustrates the textures of the iron oxides coating the surfaces of other minerals, although the iron oxides appear to preferentially form rims on potassium feldspar grains. Selected quantitative electron microprobe analyses from samples REB-89-1054 and REB-90-1175.5 both demonstrate the presence of significant As, Zn, and Cu contained in the iron oxide rims, particularly in the sample from REB-89 (Tables 4 and 5). Numerous quantitative point analyses, with the beam focused to a few microns in diameter, failed to identify discrete zinc or copper oxide minerals, lending further support to the evidence presented above that the base metals are somehow associated with the iron oxide minerals, rather than contained in very fine-grained base metal oxide minerals admixed with iron oxides.

The electron microprobe analyses from REB-89-1054 (Fig. 20) were calculated as an FeO(OH) (iron oxyhydroxide) chemical composition (Table 4). The fact that this calculation brought the totals close to 100% supports the idea that these iron oxide rims are goethite. Totals for the analyses of the iron oxide rims from sample REB-90-1157.5 were all well over 100% when calculated as goethite, so they were calculated as hematite,
Figure 20. BSE images of metal-enriched iron oxyhydroxide coatings

Two backscattered electron images from electron microprobe study of sample REB-89-1054. Bright areas indicate higher average atomic number, with the whitish material in these images all iron oxyhydroxides. The brightest orange grains which are completely surrounded by iron oxyhydroxide material in (A), and also encased by the abundant iron oxyhydroxides in the center of the image in (B), are feldspar. The darker orange areas of the microprobe images are quartz and other silicate grains, and the black areas are clay.
<table>
<thead>
<tr>
<th>Analysis</th>
<th>1</th>
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<td>0.36</td>
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<td>99.12</td>
<td>99.26</td>
<td>99.41</td>
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1.00 Hydrogen atoms, and stoichiometric oxygen has been added to analyses to calculate these totals as FeO(OH) composition.
Table 5. Electron microprobe analyses of hematite coatings from REB-90-1157.5

<table>
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<tr>
<th>Analysis</th>
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<th>2</th>
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<th>4</th>
<th>5</th>
<th>6</th>
<th>7</th>
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<td>64.15</td>
<td>64.39</td>
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<tr>
<td>PbO</td>
<td>&lt;d.l.</td>
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<td>&lt;d.l.</td>
<td>&lt;d.l.</td>
<td>&lt;d.l.</td>
<td>&lt;d.l.</td>
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<td>&lt;d.l.</td>
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<tr>
<td>ZnO</td>
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<td>0.33</td>
<td>0.36</td>
<td>0.53</td>
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<td>0.50</td>
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<td>0.07</td>
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<tr>
<td>SiO2</td>
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<td>96.21</td>
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<td>96.06</td>
</tr>
</tbody>
</table>

All analyses were corrected for oxide forms.

detection limit = 0.05 wt % Pb
which produced totals near 100% (Table 5). These electron microprobe totals for the hematite were all a bit low, which may likely be due to partial hydration of the hematite (B. Jones, pers. commun., 2001), or to the presence of magnesium or other metals which are not able to be measured with this particular array of elements. Alpers and Brimhall (1989) review literature discussing the incorporation of Al into hematite and goethite, and also the preference of Mg\(^{2+}\) for hematite over goethite.

The concentration of base metals analyzed in the hematite of the REB-90-1157.5 sample was considerably lower than the concentration of those base metals in the goethite of REB-89-1054, but this decrease is especially pronounced for copper. Some zinc also appears to be intimately associated with these cuprite clusters, with electron microprobe measurements revealing that the interiors of some of the larger porous cuprite clusters contain up to 3 wt % Zn.

These amounts of other metals associated with the iron oxides of the Mike deposit are not surprising, given the amounts of metal concentrated in iron oxide minerals measured by researchers at other localities. Iron oxides in an oxidized deposit in southern Utah are reported to contain as much as several percent of accumulated Si, Ca, Al, K, As, Pb, Cu, and Zn, as determined from SEM study (Bernstein, 1986). Thornber and Wildman (1984) conducted a series of experiments to simulate the production of gossan minerals during sulfide deposit oxidation, and they repeatedly measured base-metals bound to the iron oxides precipitated in their experiments. With total metal concentration of 0.5 M and proportions of base metal to iron of 1:100 in that solution, about 1 wt % Cu or Zn, or 0.9 wt % Pb, were contained in the precipitated iron oxides. Higher base metal to iron ratios yielded higher corresponding base metal concentrations in the iron oxides precipitated. Their timed experiments with initially equal amounts of iron and copper in solution produce cuprite first, followed by base-metal bearing iron oxide, similar to the observations of cuprite and hematite in sample REB-90-1157.5 from the Mike deposit (Fig. 19). Concerning arsenic in the iron oxide and hydroxide oxidation rims in the Mike deposit, Bowell (1994) measured similar concentrations of As in iron oxides, with up to 3.5 wt % As in iron oxide and oxyhydroxide soils from the oxidized portion of an arsenic-bearing gold deposit in Ghana.
Patterns of Copper Mineral Distribution

Figure 21 displays the pattern of copper mineral distribution based on detailed core logging across the central area of the Mike deposit. Variability northwest and southeast of the plane of this SW-NE section, from the limited evidence available to address that question, will be discussed below. These mineral zones depict which particular copper oxide mineral predominates over others in terms of relative abundance, with no implication regarding absolute abundance, the concentration of copper, or the exclusion of other copper oxide minerals.

Copper grade distribution in the deposit (Fig. 22) illustrates the obvious fact that most significant areas of copper concentration occur in the oxidized portion of the deposit, above the redox elevation. There is a good correlation between locations of significant copper grades and locations where copper minerals are observed, although some zones of lower, yet significant copper concentration (500 to 1000 ppm Cu) have no visible copper mineral hosts. Comparison of Figures 21 and 22 shows no large-scale correlation of high copper concentration with zones containing any particular dominant copper oxide mineral.

Aside from limited suspected chalcocite enrichment below the redox boundary in a few drill holes (e.g. REB-93 and REB-37 on Fig. 21, and REB-91 south of this section), only two other deeper thin zones (ten feet thick assay interval) of sulfide zone core containing over 0.2 wt % Cu were encountered during the course of core logging (Fig. 22). Neither of these anomalous sulfide copper zones contained base-metal sulfide veins, although both zones contained stratigraphically conformable sulfides of very fine grain-size. Electron microprobe study of a sample containing stratiform sulfide grains from hole REB-88 revealed that chalcopyrite is present within composite, 50-80 μm diameter, sulfide grains containing pyrite, sphalerite, and lesser chalcopyrite.

A few broad patterns concerning the distribution of copper minerals become apparent from the compilation of Figure 21. In each of the logged holes where any chrysocolla was observed, conichalcite was typically observed at least 40 feet above the first abundant concentration of chrysocolla. Chrysocolla dominant zones appear to be located deeper and more centrally located within the Mike deposit relative to conichalcite.
Figure 21. Copper mineral distribution

SW to NE fence diagram of copper mineral and oxidized chalcocite distribution in the Mike deposit. The locations of these roughly coplanar, vertical drill holes are shown in plan view on Figure 2. Mineral fields illustrate dominant copper minerals observed. Rare malachite zones are too thin to include as separate fields, present up to 0.5 m thick zones primarily at the outer fringes of the deposit. (Note that the horizontal length of the chalcocite blankets as shown here correlates to the degree of the original chalcocite replacement, and is not related to any actual spatial length. Due to the scale of the figure, the lines also do not express the actual thickness of the chalcocite or oxidized chalcocite blankets.)
KEY

- chrysocolla
- cuprite > native Cu
- conichalcite
- native Cu >> cuprite
- copper impregnated clays
- hematite after chalccite blanket oxidation
- relict chalccite blanket
- incompletely oxidized chalccite blanket

Length of line marking chalccite locations indicates relative thickness and completeness of original sulfide replacement by chalccite. Length of copper impregnated clay fields follows a similar convention for intensity of blue-green color to clay, indicating probable copper content. Boxes in red or grey mark locations of chalccite blanket textures much thicker than line width.
Figure 22. Copper grade distribution

SW to NE fence diagram depicting copper grade distribution. Shaded area outlines the areas which have been assayed, while no assay data exists for white areas. Movement on the Valley Fault post-dates copper emplacement, and there appears to be minor smearing of copper-bearing rock in the fault zone where it intersect REB-88.
KEY

- > 1% Cu
- 0.5-1.0 % Cu
- 0.25-0.5 % Cu
- 0.1-0.25 % Cu
- 0.05-0.1 % Cu
- < 0.05 % Cu
dominant zones (Fig. 21). However, conichalcite, or chrysocolla (if deep enough below the first occurrence of macroscopic conichalcite) is commonly observed within zones dominated by the other of those two minerals.

Minor native copper is common in cuprite dominant zones, and limited cuprite is not unusual in native copper dominated zones (Fig. 21). Cuprite and native copper dominant zones, particularly where encountered at the southwestern fringe of the deposit (REB-90), appear interfingered with conichalcite and chrysocolla dominant zones (Fig. 21). These separate zones (either cuprite plus native copper, or conichalcite or chrysocolla) are typically at least 5 meters thick where logged in REB-90, and within either of these two mineral zone types, typically only very rare occurrences of the minerals from the alternate mineral zone type are present. This pattern of mineral intervals is encountered on a scale of the deposit, unlike the local scale example of malachite in small pods less than a half meter thick, as described above.

Zones of significant gold concentration are interpreted to be subparallel to stratigraphic bedding planes (Figs. 3, 23). The intervals of high gold concentration typically do not coincide with intervals of high copper concentration, and there is no recognized correlation with the presence or absence of any particular minerals.

Conichalcite occurs only over an area approximately limited to the vicinity of gold mineralization (Fig. 24). The conichalcite boundary in Figure 24 marks the extent of significant conichalcite in the deposit, although there are rare conichalcite occurrences outside of this boundary. Only two documented observations of conichalcite occurrences in core from the two holes logged in detail for this study which lay outside of this boundary. CV-16 contained a few grains of conichalcite, visible only with a 20-power hand lens, in the vicinity of the uppermost gold zone (only 10 feet, at 0.02 oz/st Au), and a second location of a few conichalcite grains below the deepest oxidized layer of once pervasive chalcocite replacement in REB-68. This conichalcite boundary of Figure 24 is also an approximate limit for chrysocolla.

Southeast of the plane of the fence diagram in Figure 21, copper concentrations decrease significantly (Fig. 4). Very few copper minerals were observed in REB-68, the exceptions being a possible weakly concentrated copper-bearing clay, associated with a near completely oxidized layer of former chalcocite, and the conichalcite grains.
Figure 23. Gold grade distribution

SW to NE fence diagram depicting gold grade distribution. Shaded area outlines the areas which have been assayed, while no assay data exists for white areas. Offset of gold grade zones is due to movement on the post-mineralization Valley Fault.
Figure 24. **Conichalcite distribution in plan view**

Vertical projection of the extent of conichalcite observed in the Mike deposit. The drill holes investigated to assist in the placement of the conichalcite boundary (as shown in the figure) include more than the seven which were logged in great detail. The approximate outline of Carlin-type gold is also shown for comparison.
mentioned above. There were numerous oxidized chalcocite blankets in REB-68, commonly in and near the intervals of highest gold grade (although relatively low gold concentration compared to gold intervals in other holes). In spite of their frequency in the core of REB-68, the original chalcocite replacement of pyrite in these blankets appears to be far less complete than in the blankets observed in REB-88. REB-68 was not drilled deep enough to reach the redox and determine if there was chalcocite enrichment there.

An isolated concentration of copper oxide minerals is located at the Copper King mine 1 km southeast of the conichalcite limit of the Mike deposit (Fig. 24). The Copper King contained chrysocolla, malachite, azurite, and cuprite, hosted in limestone beds in the footwall of the Good Hope Fault footwall (Erd et al., 1953). This abandoned mine is now buried from operations at the Tusc Pit (Fig. 1).

Northwest of the fence diagram section in Figure 21, hole CV-16 exhibits a less hornfelsed section of Rodeo Creek Unit than present in rock central to the West Mike deposit (Fig. 2). CV-16 contains only one interval of significant gold concentration, but in spite of its lower gold concentration, the hole contains significant copper concentrations, comparable to the levels measured in REB-90 (Fig. 22). However, native copper is the predominant copper mineral in CV-16, and other copper minerals observed elsewhere in the deposit are rare or absent. Oxidized remnants of only two weakly developed chalcocite replacement blankets were observed in hole CV-16, in spite of the fact that this core is markedly more competent than the core from the other two older Mike Deposit holes logged in detail for this study, REB-37 and REB-68, allowing for greater confidence identifying textures such as oxidized chalcocite in CV-16. Aside from the minimal hornfelsing, hole CV-16 also exhibits much less extensive Carlin-style alteration than any of the other mineralized holes, perhaps as expected, given that hole CV-16 is fairly distal to the core of the Mike deposit.

Patterns of Iron Oxide Distribution

In spite of their varying appearance, the numerous iron oxides observed in the core are all hematite and goethite. Reddish-purple and golden yellow to dark brown stains pervading the core (e.g. Fig. 17B), along with uncommon red or dark brown spots
up to 1.5 mm across, are inferred to be indigenous hematite and goethite, derived from the immediate vicinity, if not \textit{in situ}. Botyroidal hematite 2-4 mm in diameter is observed in clumps marking oxidized chalcocite blankets, as discussed above (Fig. 8). Other hematite and goethite occupy fractures (e.g. Fig. 17C), and in some core permeate fracture walls for a short distance (e.g. Fig. 7A). Hematite samples analyzed with XRD typically contain minor quartz, whereas goethite samples all contain quartz and varying amounts of clays which probably are responsible for the variable color between medium brown and golden yellow resembling jarosite. No jarosite was found, despite efforts to identify it.

Although the colors and textures of the iron oxide minerals commonly vary over short distances, some very general distribution patterns were noted. Goethite, both exotic and indigenous, is the predominant iron oxide mineral in intervals of core containing sparse copper minerals, although some local zones of hematite are observed interfingering with goethite in some locations. However, hematite dominates intervals containing abundant copper-oxide minerals, with rare goethite present. The specific sections of core and actual fractures hosting the secondary copper minerals typically contain no iron oxide minerals with the copper-bearing minerals (Figs. 9, 14, 16, and cuprite in 17A and 17B), or rare iron oxides (Fig. 17C). Although the abundance of iron oxide minerals decreases with depth, among the iron oxides present, most holes exhibit a goethite-dominant interval over at least 1.5 meters immediately above the redox, and over 21 meters in hole CV-16. The only hematite present in the oxidized interval immediately overlying the redox is that associated with oxidized chalcocite, if present.
DISCUSSION

Eh-pH Environments of Supergene Copper Minerals

Except for copper-bearing sulfides, copper-bearing clays and remnant chalcocite, copper minerals occur dominantly along fractures that cut all other features in the core, implying that they are secondary minerals deposited by percolating groundwater. Larger scale patterns delineated from field observations (Fig. 21) show a clear association of the copper minerals other than chalcopyrite (including chalcocite at the redox boundary in some holes) with oxidative weathering. The particular copper minerals encountered are those commonly observed in oxidized zones of base metal deposits (e.g. Chávez, 2000; Magalhaes et al., 1988). Knowing that the minerals are secondary, calculations were undertaken to model Eh-pH conditions and their spatial variability during supergene copper enrichment, and to test for thermodynamic equilibrium between the copper minerals observed.

A simplified Eh-pH diagram for the Cu-O-S-H system at 25°C is presented in Figure 25, calculated for species concentrations intended to be representative of the oxidizing environment at the Mike deposit. With the addition of carbonate to this system, malachite may also form, as illustrated in Figure 26. These figures were generated with Geochemist's Workbench 3.0 computer software, using an internally consistent thermodynamic database compiled by the geochemical modeling group at Lawrence Livermore National Laboratories (Bethke, 1998). Establishment of the theory involved and details of the construction of Eh-pH diagrams are thoroughly described by Garrels and Christ (1965). Fields within the diagram illustrate which copper mineral, aqueous species, or ion is stable. Minerals will form at the Eh and pH value corresponding to their stability fields if the activity of aqueous complexes and bare metal ions are equal to the inset value shown (Fig. 26). Other species or copper minerals may exist at that Eh and pH, but they would be metastable relative to the minerals and species shown. The dashed lines within the diagram separate areas where different sulfur species are dominant, thus determining which sulfur species will be involved in any chemical reactions with other aqueous species or minerals in that Eh-pH area.
Figure 25. Eh-pH diagram for Cu-O-S-H system at 25°C, 1 atm

Eh-pH stability diagram for 25°C, 1 atm, with the activities of copper and sulfate as shown. Stability fields for copper minerals are marked by solid lines, while thin, dashed lines separate areas of dominance for sulfur species.
\[ a_{\text{Cu}^{2+}} = 2.6 \cdot 10^{-5} \]
\[ a_{\text{SO}_4^{2-}} = 5.7 \cdot 10^{-4} \]

25°C, 1 atm

Figure 26: Eh-pH diagram for Cu-O-S-H-C system at 25°C, 1 atm.
Figure 26. Eh-pH diagram for Cu-O-S-H-C system at 25°C, 1 atm

Eh-pH stability diagram for 25°C, 1 atm, with the activities of copper, sulfate, and carbonate as shown. Stability fields for copper minerals are marked by solid lines, while thin, dashed lines separate areas of dominance for sulfur species.
$a_{\text{Cu}^{2+}} = 2.6 \cdot 10^{-5}$

$a_{\text{SO}_4^{2-}} = 5.7 \cdot 10^{-4}$

$a_{\text{HCO}_3^-} = 2.3 \cdot 10^{-3}$
Assumptions

The concentrations of components in solution that were chosen to calculate the Eh-pH diagrams shown in Figures 25 through 27 are based on analyses of water samples from natural environments that best approximate those thought to be present at the Mike deposit (Table 6). Concentrations of aqueous sulfur as sulfate ion and of carbonate as bicarbonate ion reflect their predominance at the Eh and pH conditions of the sampled waters. The concentrations assumed for copper and sulfate are based on a water sample collected from an active supergene enrichment profile over the Pleysumi copper prospect in Papua New Guinea (Titley, 1978). This is a tropical climate with high rainfall; thus these concentrations may be lower than the actual value during the semi-arid environment of enrichment at the Mike deposit, but they provide a reasonable first approximation. (The uncertainty in assumptions and the resulting affects on the appearance of the Eh-pH diagram are discussed below.) Sulfate concentration is significantly higher (about 1500 mg/L SO$_4^{2-}$) in lake waters filling the Carlin-type Getchell Mine pit (Tempel et al., 2000), but this value is higher than I would expect in groundwater during oxidation of the Mike deposit. An open pit exposes a large volume of rock containing sulfide minerals to oxygenated surface water, which would increase the amount of dissolved sulfate over that present in groundwater in a subsurface environment. Therefore, I believe that the sulfate concentration measured in groundwater from an active copper enrichment zone, despite a tropical climate, is a better estimate. Values of dissolved carbonate concentration in natural environments vary greatly. The assumed concentration shown in Table 6 is an average of three similar analyses of HCO$_3^-$ in river water from three rivers in arid and semi-arid climates (Livingston, 1963, reported in Krauskopf and Bird, 1995).

Activities of components in the Cu-O-S-C-H system

Concentrations reported from analyses of water samples (Table 6) were converted to activities at 25°C with the aqueous speciation modeling software program EQ3nr version 8.0 (Wolery, 1992) using the slop98 data base from Shock and coworkers (1998). As discussed below, the groundwater responsible for supergene oxidation and enrichment of the Mike deposit is believed to have been mildly acidic. In addition to excess H$^+$ relative to OH$^-$ concentration, the solution most likely contained Na$^+$, Ca$^{++}$, and other
Table 6. Estimated chemical activity values for copper Eh-pH diagrams

<table>
<thead>
<tr>
<th>component</th>
<th>concentration</th>
<th>source environment of this concentration</th>
<th>resulting chemical activity</th>
<th>cited source</th>
</tr>
</thead>
<tbody>
<tr>
<td>copper</td>
<td>2.1 ppm</td>
<td>water sample from enrichment zone of an actively forming copper supergene profile of a copper prospect</td>
<td>$2.6 \cdot 10^{-5}$</td>
<td>Titley, 1978</td>
</tr>
<tr>
<td>sulfur (SO$_4^{2-}$)</td>
<td>68 ppm</td>
<td>same water sample</td>
<td>$5.6 \cdot 10^{-4}$</td>
<td>Titley, 1978</td>
</tr>
<tr>
<td>carbonate (HCO$_3^-$)</td>
<td>147 ppm</td>
<td>average of dissolved carbonate concentration in three river water samples from arid and semi-arid environments</td>
<td>$2.4 \cdot 10^{-3}$</td>
<td>Livingston, 1963</td>
</tr>
<tr>
<td>arsenic</td>
<td>0.57 ppm</td>
<td>water sample from pit lake filling old mine of an arsenic-rich, Carlin-type gold deposit, Nevada</td>
<td>$7.0 \cdot 10^{-6}$ for H$_2$AsO$_4^-$</td>
<td>Tempel et al., 2000</td>
</tr>
<tr>
<td>calcium</td>
<td>70 ppm</td>
<td>average of ten samples of Ca$^{2+}$ from pit lakes filling old mines of Carlin-type deposits in Nevada, value agrees well with river water analyses from same arid and semi-arid rivers above</td>
<td>$1.4 \cdot 10^{-3}$</td>
<td>Price et al., 1995</td>
</tr>
</tbody>
</table>
cations, which must be assumed to have been present in order to maintain the requirement of electrical charge balance in aqueous solutions. Though it can be assumed that there must have been calcium present in the groundwater due to the appearance of supergene conichalcite, it is impossible to determine the identities of all other cations or to accurately constrain any of the concentrations in the groundwater during oxidation of the Mike deposit. The EQ3nr runs were balanced on Na\(^+\) ions for simplicity, and the resulting concentrations of Na\(^+\) calculated by EQ3nr to maintain the required electrical neutrality of the solution were within the range of Na\(^+\) in semi-arid and arid environment river water analyses (Livingston, 1963, reported in Krauskopf and Bird, 1995) for all modeled pH values, implying that this simplification did not lead to unreasonable modeled results. Separate EQ3nr models were run for pH values of 3.5, 4.5, 5.5, and 6.5, and all of these models resulted in nearly equal activities for carbonate, sulfate and copper in the Cu-O-S-C-H system of Figure 26.

**Variations in Eh-pH diagrams due to uncertainty of assumptions**

The general form of the Eh-pH diagram for the system Cu-S-O-C-H at 25°C does not change significantly in response to changes in the activities of any of the dissolved species. With increasing copper activity, the malachite, native copper and cuprite stability fields expand into regions of higher Eh and lower pH, meaning that within a system of higher copper concentration and no other chemical changes, these copper minerals are stable under more acidic and oxidizing conditions. However, even with an order of magnitude increase or decrease in copper activity, the phase boundaries shift only slightly. For example, the vertical malachite-Cu\(^{++}\) boundary of Figure 26 is the boundary most affected by changes in the copper activity, but with an order of magnitude change in copper activity, the boundary shifts only two-thirds of one pH unit. The cuprite-Cu\(^{++}\) and native Cu-Cu\(^{++}\) boundaries are even less affected by changes in copper activity. Therefore, any inaccuracies regarding the original copper concentration estimate would need to be large (two orders of magnitude or greater) to significantly affect the size of stability fields and their location within the Eh-pH model.

Likewise, changes in the chemical activity of sulfur do not produce a significant shift in any of the stability boundaries. The Cu\(^{++}\)-chalcocite and native Cu-chalcocite boundaries shift almost imperceptibly with one order of magnitude change in sulfate
activity. The covellite stability field occupies a greater or lesser proportion of the chalcocite stability field with increasing or decreasing sulfur concentration, respectively. Predominance boundaries for aqueous sulfur species, located at Eh-pH values where two bordering species have equal activities, are not affected by changes in sulfur concentration. The stability field of native sulfur decreases and eventually disappears with decreasing sulfur concentration. If the activity of sulfur becomes large enough, copper sulfate minerals may form in place of malachite, although this concentration of sulfur would be realistic only under hyperarid conditions. The malachite stability field shown in Figure 26 shrinks in response to a decrease in carbonate concentration, but by a smaller rate than it decreases in response to decreasing copper concentration, as discussed above. With the estimated concentrations of copper and sulfate as listed in Table 6, malachite would not be stable at any Eh-pH conditions if the activity of bicarbonate ion is less than $6.0 \times 10^{-4}$, which corresponds to a concentration of about 39 ppm $\text{HCO}_3^-$. For reference, concentration of $\text{HCO}_3^-$ is in a water in equilibrium with the atmosphere is $10^{-5}$.

**Limitations of Eh-pH diagrams**

Eh-pH diagrams such as those presented in Figures 25 and 26 represent equilibrium conditions between minerals and an aqueous solution. Eh-pH diagrams do not take into account the kinetics of chemical reactions, which determine how long a system may require to reach complete chemical equilibrium. Determination of whether or not chemical equilibrium was attained in an ancient water-rock system, particularly when attempting to assess this condition only from the existence of remnant minerals, is a difficult task. Observations from supergene enrichment profiles and experimental data imply that the weathering environment in oxidizing ore deposits is not well-represented by equilibrium reactions as depicted in Eh-pH diagrams. For example, abiotic reduction of aqueous sulfate to sulfide is kinetically inhibited at low temperatures (25°C), as shown by Ohmoto and Lasaga (1982), who calculated from experiments that it would require a minimum time of 9 billion years for 90% equilibrium to be attained between sulfate and sulfide in a 0.01 molal total sulfur system at a pH range of 4 to 7.

From his electrochemical experimental studies, Sato (1992) has modified Eh-pH stability diagrams for environments of supergene enrichment, generating what he names
"persistency-field" Eh-pH diagrams. He notes that, during oxidative weathering of ore bodies, some supergene mineral assemblages are metastable. For example, under oxidizing conditions, chalcocite persists within the stability fields of both native copper and cuprite, in spite of the fact that cuprite is not stable with chalcocite. Sato's chalcocite persistency-field diagram is consistent with field observations at the Mike deposit, where chalcocite is found in contact with cuprite (e.g., Fig. 17B).

Eh-pH diagrams may not represent a perfect model for supergene enrichment profiles, but as long as their limitations are stated and understood, stability diagrams are useful as a basis for discussion of geochemical environment and processes. At the very least, they provide a map that allows consideration of directions of change, and metastability can be accounted for in a qualitative way.

**Addition of conichalcite to the system**

The presence of conichalcite among the secondary copper minerals requires addition of both arsenate and calcium to the Cu-O-S-C-H system modeled in Figure 26. Arsenic concentration is assumed to be 0.57 mg/L, based on the average of measured concentration in two samples (0.55 and 0.59 mg/L) collected at different depths of the more oxygenated epilimnion of the pit lake filling the North Pit, Getchell mine, Nevada (Tempel et al., 2000). The pit lake had been full for approximately 8.5 years prior to collection of the water samples, allowing time for the arsenic in the system to reach chemical equilibrium. The Getchell Mine is a Carlin-type gold deposit containing abundant arsenic sulfide minerals and a very high arsenic-content (Cline, 2001).

Analyses of ten water samples from two pit lakes in abandoned Carlin-type deposits (Price et al., 1995) yield an average concentration of 70 ppm Ca++. This average agrees well with Ca++ concentrations of 94 ppm, 23 ppm, and 17.4 ppm, reported for the three, earlier cited, rivers in arid and semi-arid climates (Krauskopf and Bird, 1995). The addition of aqueous arsenic and calcium species increases the ionic strength of the modeled groundwater, thereby slightly decreasing the activities of all species in solution relative to activities used in diagrams for the system Cu-O-S-C-H system described above (Figs. 25 and 26), although the assumed concentrations of all these components have not been changed.
The Eh-pH diagram of Figure 27 shows the speciation of arsenic in the system As-O-S-H, illustrating that at the range of pH and oxidizing conditions believed to represent the groundwater during supergene enrichment of the Mike deposit, as discussed below, H$_2$AsO$_4^-$ is the dominant species of aqueous arsenic. Using this arsenic species, the boundary between conichalcite and aqueous Cu$^{++}$ is described by the reaction:

$$\text{CaCu(AsO}_4\text{(OH)} + 3\text{H}^+ = \text{Ca}^{++} + \text{Cu}^{++} + \text{H}_2\text{AsO}_4^- + \text{H}_2\text{O}$$

Modeling the ionic strength and activities of all the species in the groundwater, the result for H$_2$AsO$_4^-$ activity is approximately $7.0 \cdot 10^{-6}$. Seventy ppm Ca$^{++}$ in the fluid translates into an activity of about $1.4 \cdot 10^{-3}$ for calcium. An equilibrium constant, log $K = 1.29$, has been determined for the above reaction through experiments conducted by Magalhaes et al. (1988). Equilibrium constants for other chemical reactions involving conichalcite can be calculated from the standard free energies of formation for conichalcite, which was determined by Magalhaes et al. (1988), and all other minerals and species involved in the reaction, according to the method outlined by Garrels and Christ (1965). Standard free energy of formation values for all other minerals and aqueous species involved in the reactions were taken from the slope98 thermodynamic database compiled by Shock and coworkers (1998). Redox reactions and equilibrium constants calculated are listed in Table 7. Activity of the arsenic species involved depends on the relative ionic charge of the predominant species at any particular value of Eh and pH, with species of higher ionic charge having lower calculated activity than the concentration of arsenic in system.

The equilibrium stability field for conichalcite in relation to other copper oxide minerals is illustrated in the Eh-pH diagram of Figure 28. Conichalcite is stable at oxidizing conditions above a pH of 4.7 (Fig. 28), in contrast with malachite and tenorite, which are stable above a pH of 6.0 (Figs. 25 and 26). Given the assumed concentrations of copper, arsenic, calcium, and carbonate in the system, conichalcite is a stable mineral over a large range of Eh and pH conditions. Decreases in the aqueous concentration of arsenic and/or calcium will cause the conichalcite stability field to shrink from its size in Figure 28. Calculating the redox reactions for the boundary between malachite and conichalcite, it becomes apparent that with the assumed concentrations of species in this groundwater system, malachite is not stable relative to conichalcite anywhere within or near the pH range for malachite as delineated in Figure 26. Note that because three H$^+$
Figure 27. Arsenic speciation: Eh-pH diagram for As-S-O-H system

Eh-pH diagram showing areas of dominance for arsenate and arsenite ions, and stability fields for arsenic sulfide minerals at 25°C, 1 atm, with the activities of arsenic and sulfate as shown. Thin, dashed lines separate areas of dominance for sulfur species.
$a_{\text{As}} = 7.6 \cdot 10^{-6}$

$\text{AsO}_4^{3-} = 5.7 \cdot 10^{-4}$

25°C, 1 atm
Table 7. Redox chemical equations and calculated equilibrium constants which define conichalcite stability

<table>
<thead>
<tr>
<th>Redox chemical reactions</th>
<th>log K</th>
</tr>
</thead>
<tbody>
<tr>
<td>1) CaCu(AsO₄)(OH) + 3H⁺ = Ca⁺²⁺ + Cu⁺²⁺ + H₂AsO₄⁻ + H₂O</td>
<td>+1.29</td>
</tr>
<tr>
<td>2) 2CaCu(AsO₄)(OH) + 6e⁻ + 10H⁺ = Cu₂O + 2Ca⁺²⁺ + 2HAsO₂(aq) + 5H₂O</td>
<td>+52.75</td>
</tr>
<tr>
<td>3) CaCu(AsO₄)(OH) + 4e⁻ + 6H⁺ = Cu⁺⁰ + Ca⁺²⁺ + HAsO₂(aq) + 3H₂O</td>
<td>+34.18</td>
</tr>
<tr>
<td>4) CaCu(AsO₄)(OH) + 2e⁻ + 2H⁺ = Cu⁺⁰ + Ca⁺²⁺ + HAsO₄⁻² + H₂O</td>
<td>+5.85</td>
</tr>
<tr>
<td>5) CaCu(AsO₄)(OH) + 2e⁻ = Cu⁺⁰ + Ca⁺²⁺ + AsO₄⁻³ + OH⁻</td>
<td>-19.83</td>
</tr>
<tr>
<td>6) 2CaCu(AsO₄)(OH) + 2e⁻ = Cu₂O + 2Ca⁺²⁺ + 2AsO₄⁻³ + H₂O</td>
<td>-27.31</td>
</tr>
<tr>
<td>7) CaCu(AsO₄)(OH) + OH⁻ = CuO + Ca⁺¹² + AsO₄⁻³ + H₂O</td>
<td>-11.00</td>
</tr>
<tr>
<td>8) 2CaCu(AsO₄)(OH) + HCO₃⁻ + 3H⁺ = Cu₂(OH)₂CO₃ + 2Ca⁺²⁺ + 2H₂AsO₄⁻</td>
<td>-3.87</td>
</tr>
</tbody>
</table>
Figure 28. Concichalcite stability field in Eh-pH space

Eh-pH stability diagram for 25°C, 1 atm, at the activities of all components as shown, showing the broad stability field for concichalcite at these conditions, relative to other copper minerals (e.g., Fig. 26). Stability fields for copper minerals are marked by solid lines, while thin, dashed lines separate areas of dominance for sulfur species, and dotted lines for arsenic species.
arsenic speciation

sulfur speciation

\[ a_{\text{As}} = 7.6 \cdot 10^{-6} \]

\[ a_{\text{Ca}^{2+}} = 1.3 \cdot 10^{-3} \]

\[ a_{\text{Cu}^{2+}} = 2.4 \cdot 10^{-5} \]

\[ a_{\text{Cu}^{2+}} = 4.4 \cdot 10^{-4} \]
ions are involved in the reaction forming conichalcite from water containing Cu$^{++}$ ions (reaction 1 on Table 7), the location of this boundary is not strongly affected by variations in the assumed concentration of individual species. A change of three orders of magnitude in the same direction of each of arsenic, calcium and copper species would be required to shift the boundary one pH unit.

**Addition of chrysocolla to the system**

The Mike deposit contains abundant chrysocolla, which requires the addition of silica to the modeled aqueous system. Because it is not a mineral in a strict sense, having an amorphous structure and variable chemical composition, it is more difficult to model than the other copper minerals described above. However, the results of observational and experimental studies do allow an approximation for chrysocolla formation. Experiments by Newberg (1967) produced chrysocolla by a stepwise process, where copper hydroxide was stabilized by monomeric silica, with metastable precipitates converting to a more stable final product of chrysocolla, for which he determined a solubility constant. These experiments, combined with analysis of previous studies, led Newberg (1967) to conclude that chrysocolla forms at (a) low copper concentrations, (b) low concentrations of anions with which copper could precipitate as insoluble mineral salts, and (c) neutral to basic conditions. Williamson (1994) observed an aluminum-rich chrysocolla coagulating in a series of settling ponds into which drainage water flows from a supergene copper deposit. This groundwater pH was 5.7 to 7.0, and Williamson states that the primary transport mechanism of chrysocolla in the groundwater appears to be by colloidal silica. Williamson estimated a free energy of formation that is in close agreement with Newberg's (1967).

Yates et al. (1998) conducted laboratory experiments aimed at understanding the mechanism of chrysocolla formation. Using solutions with high concentrations of silica (though maintained below amorphous silica saturation), these authors studied early stages of chrysocolla formation at varying pH and copper concentrations. They determined that copper in solution under mildly acidic conditions was attracted to the surface charge of aqueous silica and induced coagulation of that aqueous silica at concentrations as low as 15 ppm Cu (at which point the uptake of aqueous copper at their experimental conditions was near 100%). At a pH higher than 7, Cu(OH)$_2$ becomes the increasingly dominant
copper species, and this species is not involved in $H^+$ cation exchange with aqueous silica (Yates et al., 1998). The results of this study imply that chrysocolla will form easily from aqueous copper ion in mildly acidic to near neutral oxidized groundwater, provided that there is sufficient silica in solution, and no abundance of anions with which the aqueous copper may react to form other minerals. This study supports Newberg's (1967) idea of chrysocolla being able to form under conditions of low copper concentration, but Yates et al. (1998) have established experimentally that chrysocolla appears to form over a more narrow pH range, primarily from 5.5 to 7.5 pH in solution.

Newberg (1967) determined solubility constants from his experiments on relatively pure chrysocolla. However, these solubility values are of questionable utility, given the amorphous structure and highly variable chemical composition and degree of hydration in natural chrysocolla. Using Newberg's (1967) thermodynamic data, with the assumed concentrations of other components of the modeled groundwater system for the Mike deposit, I calculated reactions for the stability of chrysocolla, and its reaction with malachite, tenorite and conichalcite. Every one of these calculated stability reactions overestimates the stability for chrysocolla by at least several orders of magnitude above reasonable estimates relative to the body of geological observations regarding chrysocolla.

While chrysocolla defies any efforts to generate rigorous stability field boundaries in Eh-pH space, some generalizations can be made concerning its stability relative to other minerals. For example, chrysocolla is not stable in acidic groundwater, as predicted from field relations (Newberg, 1967) and expressed by the reaction:

$$\text{CuSiO}_3 \cdot \text{H}_2\text{O} + 2\text{H}^+ = \text{SiO}_2(\text{aq}) + \text{Cu}^{+2} + 2\text{H}_2\text{O}$$

Chrysocolla also appears to be favored over tenorite in relatively silica-rich solutions (neglecting kinetic effects and upper pH limitations on chrysocolla formation), based on the experiments of Yates et al. (1998):

$$\text{CuSiO}_3 \cdot \text{H}_2\text{O} = \text{CuO + SiO}_2(\text{aq}) + \text{H}_2\text{O}$$

At concentrations of carbonate and silica that are appropriate for the stability of both malachite and chrysocolla, a boundary at a particular pH value would separate malachite and chrysocolla (with Malachite stable at lower pH than chrysocolla):
\[2\text{CuSiO}_3\cdot\text{H}_2\text{O} + \text{HCO}_3^- + \text{H}^+ = \text{Cu}_2(\text{OH})_2(\text{CO}_3) + 2\text{SiO}_2(\text{aq}) + 2\text{H}_2\text{O}\]

Also, given appropriate concentrations of arsenate, calcium, and silica for both conichalcite and chrysocolla to be stable, a boundary at a particular pH value would separate chrysocolla from conichalcite (with conichalcite stable at a more basic pH than chrysocolla):

\[
\text{CaCu(AsO}_4)(\text{OH}) + \text{SiO}_2(\text{aq}) + \text{H}_2\text{O} + \text{H}^+ = \text{CuSiO}_3\cdot\text{H}_2\text{O} + \text{Ca}^{++} + \text{H}_2\text{AsO}_4^- 
\]

The relationships between all of these minerals together will be complicated by the problems of dealing with numerous components simultaneously, and the fact that the thermodynamic properties of chrysocolla cannot be strictly established, but these relations again form a basis to evaluate the interplay of the copper oxide minerals found in the Mike deposit.
INTERPRETATION

Processes of Supergene Enrichment

**General relationships**

The Eh-pH diagram of Figure 29 is a summarization of the geochemical conditions during the supergene copper enrichment of the Mike deposit. Conichalcite, chrysocolla, malachite, and tenorite are grouped together in one idealized stability field because they all form under similar Eh-pH conditions, and the conditions under which any of these minerals forms are governed by factors other than simply Eh and pH, as discussed below. Because the stability field combining these four copper oxide minerals is idealized, and phase boundaries between this field and other species are not defined thermodynamically, those boundaries are dashed in the figures. Predominance fields for aqueous sulfur species also are shown by dashed lines, and the thick lines denote the stability fields of Fe-O-S minerals and of Fe ions in solution. The boundary between hematite and pyrite is insensitive to changes of less than a few orders of magnitude in iron activity, so any minor fluctuations from the assumed value of $10^{-3}$ for iron activity will not affect its location. The hematite-pyrite boundary is chosen as a proxy for the redox boundary in the Mike deposit, and this agrees well with the known compositions of natural waters, as shown by the gray ellipse marked "groundwater" (Garrels and Christ; 1965). Also shown are waters labeled: "oxidizing sulfides", based on Eh-pH of mine waters; "surface waters" representing rain and stream conditions; and "organic-rich water", representing relatively acidic and reduced waters, based on Eh-pH values of bog waters (Garrels and Christ; 1965).

The likelihood that pyrite persists as a metastable phase in the hematite field, and thus can act as a sink for copper under acidic conditions, is indicated on Figure 29 by the label "metastable pyrite". The likelihood of metastable pyrite, based on the experimental work of Sato (1992), explains the association of chalcocite with pyrite rather than with hematite in chalcocite blankets formed at low pH in the absence of covellite (tip of arrow "1", Fig. 29).
Figure 29. Idealized Eh-pH diagram depicting supergene copper enrichment

Specific features of the diagram, and possible fluid pathways (marked by arrows in the figure) are described in detail in the text.
Paths taken by acidic and oxidized copper-bearing waters generated at the surface through weathering of copper-bearing pyritic rocks are highly variable. The paths need to be considered in both time and space, and depend on many factors, including the wallrock mineralogy of the weathering outcrop, and the permeability and mineralogy of the surrounding wallrock if transport of copper is involved. Representative paths through Eh-pH space are shown on Figure 29 by numbered arrows, and are described in the sections that follow.

**Supergene chalcocite**

Arrow "1" illustrates the path toward decreasing Eh that forms supergene chalcocite and, in rare cases, native copper. Minor buffering of the fluid towards more neutral pH occurs with time spent in contact with wallrock, with the path more nearly vertical (constant pH) in less reactive wallrock. This path, if representing downward percolation from the vadose to the saturated zone, generates an "in situ" chalcocite blanket. Such in-situ chalcocite blankets, underlying the leached rocks that served as the source for copper, are represented by the majority of the great supergene copper orebodies of the southwestern United States and the Andes of Chile (Titley and Marozas, 1995). Given appropriate lithologies and groundwater flow, path "1" also could involve lateral transport away from the copper source, generating "exotic" chalcocite blankets (Munchmeyer, 1996) in areas where the copper in migrating solutions encounters a combination of reduced groundwater and reducing wall rocks containing pyrite, to become chemically trapped though chalcocite precipitation.

The key requirement for path "1", in addition to sufficient copper and a source of reduced sulfur, is that the pH remains low, meaning that rocks along the flow path and at the water table are incapable of neutralizing the acidic, copper-bearing fluid. Such rocks normally consist of quartz-rich rocks with a significant content of micas and clays, and lack carbonate or feldspars. The silicified and pyritic core of the Mike deposit represents this environment very well. Given a sufficiently high copper concentration in solution, chalcocite will precipitate in pores and fractures if aqueous sulfate is reduced to H₂S at the water table. Alternatively, in the absence of sufficient transported aqueous sulfur, chalcocite precipitates as the solution encounters metastable pyrite, forming rims on pyrite, or, if the process goes to completion, totally replacing pyrite. The presence or
absence of relict pyrite in chalcocite blankets will determine the processes and products if the blanket is exposed to oxidizing conditions after it is formed (discussed below).

**Supergene copper-oxides**

In contrast with supergene chalcocite, the formation of supergene copper-oxide deposits requires a path toward neutral pH, while maintaining high oxidation states (arrow "2", Fig. 29). Conichalcite, chrysocolla, malachite, and/or tenorite, form at the high oxidation states typical of the vadose zone. Arrow "3" shows a possible continuation of path "2" that could result if surface waters at neutral pH were reduced as they encounter the water table. A fluid represented by paths "2" and "3" would generate the commonly observed supergene oxide profile involving tenorite – cuprite – native copper, and could precipitate chalcocite if they contained sufficient copper on reaching the water table (tip of arrow "3", Fig. 29).

**Copper Mass Balance Calculation and Interpretation**

Abundant copper minerals in the Mike deposit are all low temperature, secondary minerals, including limited relic chalcocite blankets, and more common copper oxides hosted along fractures in the oxide portion of the deposit. This supergene copper enriched zone of the Mike deposit contains over one billion pounds of copper of an average concentration of 0.34 wt % Cu at a cutoff grade of 0.1% (Norby and Orobona, in press). However, in spite of this large amount of enriched copper, the source of copper in the chalcocite blankets and copper oxides above the redox zone is unclear. Two possible explanations exist: either (1) the copper source was bedrock overlying the present Mike deposit, and the copper was leached vertically downward in percolating groundwater during oxidation of the source, or (2) the copper source was "exotic", located laterally up the hydraulic gradient from the Mike deposit.

Regarding a local source above the present deposit, the only identified copper minerals below the redox boundary are chalcopyrite, as described above, and rare bornite, as identified in Newmont mineralogical studies. Both are contained in hypogene base metal veins, with minor additional chalcopyrite contained within composite sulfide grains among stratigraphically conformable sulfide grains. In the oxide zone, no
evidence was observed that would indicate a greater abundance of Cu-Fe-sulfides were once present above the redox than are currently observed below the redox.

A copper mass balance calculation employing the basic method outlined by Brimhall et al. (1985) was undertaken to provide a semi-quantitative assessment of a vertical leaching model for supergene copper enrichment. This model determines the thickness of rock above the current top of oxidized bedrock that would be needed to provide sufficient copper to account for copper now present in chalcocite and copper oxides. Two pieces of information are required: (1) estimates of copper concentration in the copper-enrichment zones and in the unenriched sulfide rock prior to oxidation, and (2) bulk-rock densities corresponding to these environments. Copper concentration and bulk density measurements (performed by Dawson Metallurgical Laboratories for Newmont) are available for the entire section, both oxide and sulfide, for seven deep drill holes that penetrated the sulfide zone.

The mass-balance calculation includes the following assumptions: (1) copper concentration in the sulfide zone below the redox is equivalent to the former copper concentration in the now-oxidized rock prior to weathering and this copper grade continued unchanged above the present surface up to the requisite thickness; (2) present bulk density of leached rock below the present bedrock surface is equivalent to the bulk density of rock now eroded; (3) 100% efficiency in both leaching and re-deposition of copper; (4) none of the copper-enriched rock was lost to erosion prior to burial by the Carlin Formation, which halted the enrichment process and preserved the enriched copper zone. This latter assumption may not be true in at least two of the seven holes. These three assumptions combined establish a minimum thickness of leached cap required because the efficiencies of leaching and deposition are likely to be significantly less than 100%.

A relative mass of copper normalized by vertical distance (copper concentration multiplied by bulk density, divided by the number of feet in that interval) was computed for each assay interval (typically about 5 feet) of core from the seven holes employed for this study. Table 8 contains a summary of the results of copper concentration per meter in each of the supergene enriched oxide and the hypogene sulfide zones. These were compared to determine what multiple of rock thickness would need to be leached per unit
Table 8. Summary of data from copper mass-balance calculations

<table>
<thead>
<tr>
<th>hole and zone</th>
<th>thickness (m)</th>
<th>copper conc. per meter (%Cu g cm⁻³/m)</th>
<th>eroded cap required (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>REB-87 oxide</td>
<td>148.5</td>
<td>3114</td>
<td>242</td>
</tr>
<tr>
<td>REB-87 sulfide</td>
<td>361.2</td>
<td>1909</td>
<td></td>
</tr>
<tr>
<td>REB-88 oxide</td>
<td>213.5</td>
<td>24304</td>
<td>2170</td>
</tr>
<tr>
<td>REB-88 sulfide</td>
<td>172.5</td>
<td>2391</td>
<td></td>
</tr>
<tr>
<td>REB-89 oxide</td>
<td>292.6</td>
<td>2358</td>
<td>476</td>
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<tr>
<td>REB-89 sulfide</td>
<td>328.5</td>
<td>1449</td>
<td></td>
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<tr>
<td>REB-90 oxide</td>
<td>273.4</td>
<td>20447</td>
<td>6764</td>
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<td>277.1</td>
<td>826</td>
<td></td>
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<td>REB-91 oxide</td>
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<td>REB-91 sulfide</td>
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<td>REB-93 sulfide</td>
<td>216.6</td>
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<tr>
<td>average oxide</td>
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<td>9759</td>
<td>1729</td>
</tr>
<tr>
<td>average sulfide</td>
<td>269.3</td>
<td>1725</td>
<td></td>
</tr>
</tbody>
</table>
thickness of the current enriched zone to account for the mass of copper contained. Using the total data averaged from the intervals of all seven holes, the averaged relative mass copper per meter for the sulfide zone and oxide zone was compared. Subtracting the average sulfide zone value from the oxide value, to account for the amount of copper that must be assumed to have been present before copper enrichment, results in an average copper concentration per meter in the oxide zone of 4.9 times the average copper concentration per meter in the sulfide zone (Table 8). Given an average thickness of the copper enrichment zone of 229 m, a minimum thickness of source rock above the current erosional surface is 1124 m.

This value of 1.1 km of rock can be compared to studies of erosion rates and duration of supergene enrichment in the Chilean porphyry deposits (Sillitoe and McKee, 1996; Alpers and Brimhall, 1988), Arizona (Graybeal, 1982) and Butte, Montana (Brimhall et al., 1985). Primarily due to higher-grade protore in these deposits than at Mike, estimates for minimum eroded thickness of leached capping of at least 140 meters at both Butte (Brimhall et al., 1985) and at La Escondida (Alpers and Brimhall, 1988) are considerably less than the thickness required for the Mike deposit.

Alpers and Brimhall (1988) have determined an approximate erosion rate of 42 m/m.y. during supergene copper enrichment at La Escondida by using their estimate of a 3.3 m.y duration for enrichment (based on K/Ar dates of supergene alunite) and dividing that time into their 140 meter erosion estimate. This erosion rate is less than a value of 92 m/m.y. for a transition between the Colorado Plateau and Basin and Range provinces in Arizona, arrived at by dating mesa-capping basalts at different elevations (Shafiquallah et al., 1975), cited by Graybeal (1982) when he calculated 2 million years as the length of time required for a cycle of supergene enrichment at El Tiro in Arizona. Sillitoe and McKee (1996) employed a similar approach using K/Ar dating of supergene alunite in Chilean porphyry Cu deposits to estimate the duration of at least intermittent supergene activity at individual supergene enrichment profiles at between 0.4 and 6.2 m.y. Even accepting the highest of the erosion rates for semi-arid climates of approximately 92.5 m/m.y., the minimum thickness of rock overlying the Mike deposit would require over 12 million years to erode. Note that this time estimate uses the highest erosion rate, highest efficiencies of leaching and fixation, and neglects the fact that some of the copper
enriched zone likely is missing. These factors could lead to a cumulative underestimate of the minimum time for erosion by at least 5 times.

Based on the fact that the zone of supergene enrichment contains so much more copper than is present in the unoxidized sulfide zone, and that this necessary leached cap is not present, it appears that an unrealistic thickness of overlying rock must have been leached and later eroded to account for the supergene copper observed at the Mike deposit. Therefore, the vertical leaching hypothesis is not considered applicable to the Mike deposit, and the copper currently present in the oxide zone appears to be exotic. The pattern of copper concentration open and increasing to the northwest (Fig. 4) led Norby and Orobona (in press) to suspect a source to the northwest of Mike. Copper mineral and textural evidence from this study strongly support this hypothesis.
DEPOSIT EVOLUTION

Overview

Figure 30 illustrates the stages of development of the Mike deposit. The cross section represents the same view as shown in previous sections (Figs. 6, 21-23), and the complementary longitudinal section for each frame is oriented perpendicular to the cross section. At some time prior to oxidative weathering and the development of supergene copper enrichment, and likely before Carlin-type gold deposit formation, as discussed below, a hypogene copper source was emplaced somewhere northwest of the Mike deposit (Fig. 30B). The copper source may have been related to a Mesozoic stock, which also caused the generation of hornfels in the vicinity of the Mike deposit. Nothing more can be said about this speculative copper source, other than it having necessarily contained a larger amount of copper than the current amount of copper known to be contained in the area of the Mike deposit. Only very scattered and inconsequential base-metal sulfide plus quartz veins were present at the site that later became the Mike deposit (Fig. 30A).

The Carlin-type gold deposit at Mike formed at the intersection of the Good Hope Fault and Soap Creek Fault zone. Hydrothermal activity associated with the introduction of gold also introduced abundant arsenic-rich pyrite and silicified and decarbonatized the sedimentary rocks (Fig. 30C-D). The net result of both Mesozoic hornfelsing and Tertiary silicification and decarbonatization was removal of reactive carbonate minerals from a central corridor that is oriented NW-SE and passes through the center of the Mike deposit (shown in Figure 30, frames C and D).

From the patterns of copper minerals and hematite observed in the deposit, and from the estimates of Eh-pH conditions for the formation of different minerals, the evolution of supergene copper at the Mike deposit required two stages, as discussed below.
Figure 30. Time frame sequence of the evolution of the Mike deposit

Each pair of frames (showing perpendicular views through the area of the Mike deposit, with the approximate intersection point marked at the top of the frame) depict a stage in the development of the Mike deposit supergene copper enrichment. These stages, and their relation to supergene copper enrichment, are described in thorough detail in the text.
cross sections (160% vert exaggeration)

A  hypogène Cu-sulfides

C  Mike Au deposit

E  supogne Cu Stage 1

G  supogne Cu Stage 2

longitudinal sections (250% vert exaggeration)

B  hypogène Cu-sulfides

D  Mike Au deposit

F  supogne Cu Stage 1

H  supogne Cu Stage 2
Stage-1 Supergene Copper

The sequence of events postulated here starts with erosion and weathering extending down to the level of the hypogene copper source (Fig. 30F). Copper is leached by acidic water (unfilled oval, Fig. 31), and this acidic, copper-bearing water flows down the hydraulic gradient to the southeast (Fig. 30F). Copper remains in solution as Cu$^{+2}$ ions until either the water is reduced at relatively acid pH on reaching the water table (path "1a", Fig. 31), or it remains oxidized and within the vadose zone where it encounters minerals capable of buffering it to near-neutral pH (path "1b", Fig. 31). In the former case, reducing conditions imposed by the water table do not effectively reduce aqueous sulfate to aqueous sulfide because of kinetic factors (Ohmoto and Lasaga, 1982). Therefore, chalcocite may not form until the migrating water encounters pyrite already present in the wallrock. Pyrite provides a source of reduced sulfur as it is replaced by chalcocite. Although chalcocite in chalcocite blankets may not in every case replace hypogene sulfides, chalcocite formation is not known to have been documented in any location where there were no precursor sulfide minerals.

 Fluids following an Eh-pH path represented by arrow "1a" may precipitate native copper under reducing conditions in the absence of metastable pyrite, a scenario represented by the right-hand portion of the gray oval at the tip of the arrow (Fig. 31). Hole CV-16, northwest of the heart of the Mike deposit (Fig. 2), contains high concentrations of copper, although very limited gold concentration, implying that it is at the fringe of the Mike gold deposit. The copper mineralization in CV-16 consists of extremely fine-grained native copper, exclusive of other copper minerals. Oxidized remnants of weakly developed chalcocite replacement blankets are present in only two locations within CV-16. These observations support the idea that native copper precipitated directly by the mechanism proposed above, rather than having formed as an oxidation product of chalcocite (cf., Titley, 1978). This native copper indicates that copper-bearing fluids passed CV-16 along their travel southeastward toward the Mike deposit.

Waters migrating toward the Mike deposit (Fig. 30C-D) are passing through rock that contains essentially no carbonate, little potassium feldspar, and essentially no
Figure 31. Eh-pH diagram representing Stage-1 supergene copper enrichment

Eh-pH diagram depicts the fluid pathways producing the Stage-1 supergene copper enrichment: chalcocite formation in the Mike deposit, and copper oxide precipitation at the Copper King deposit. Details of the diagram are described in the text.
plagioclase feldspar, meaning that the rocks along the fluid path (Fig. 30F) were not capable of buffering a large acidic fluid envelope to a near-neutral pH where copper oxide minerals are stable (Fig. 31). Upon reaching the Mike gold deposit, copper-bearing waters passing through pyritic rocks below the water table were reduced and precipitated chalcocite (path "la", Fig. 31). The replacement of pyrite by chalcocite proceeds by the following redox reaction:

$$\text{FeS}_2 + \text{Cu}^{+2} + 4\text{H}_2\text{O} = \text{Cu}_2\text{S} + \text{SO}_4^{-2} + \text{Fe}^{+2} + 6\text{e}^- + 8\text{H}^+$$

This process of chalcocite blanket formation locally generates copious amounts of acid, which would likely convert any remaining feldspar in the wallrock to clay, explaining the observation by Newmont geologists that all intervals where chalcocite and oxidized chalcocite blankets occur exhibit maximum clay alteration. This clay production is likely not due to Carlin-type gold mineralization, but rather chalcocite formation during Stage-1 supergene copper enrichment.

Exotic copper oxide minerals have been documented up to 8 km away from the inferred source in some of the larger exotic copper ore bodies in Chile (Munchmeyer, 1996). Exotic chalcocite is only noted in two exotic deposits, both associated with the El Salvador porphyry complex, where it forms in the "proximal zone" of the exotic ore body, 0-2 km from the center of hypogene porphyry copper mineralization (Munchmeyer, 1996). However, these distances are only representative, while the actual copper transport distances, and suite of particular copper minerals formed depends entirely on the geologic details of the system where the exotic mineralization forms. The Damiana and Quebrada M exotic copper deposits at El Salvador formed at the site where the copper-bearing fluids first encountered pyrite-bearing wallrock. There is no reason to believe that exotic chalcocite could not form further away from the copper source, as long as the copper-bearing fluids do not encounter pyrite-bearing wallrock, or minerals which could buffer the acidic fluids and form copper oxide minerals. The Mike gold deposit was likely the first rock containing abundant pyrite that the acidic, copper-bearing fluids encountered during their flow from the northwest, thus explaining the location where abundant chalcocite formed (Fig. 30E-F). Although numerous examples of oxidized chalcocite blankets are observed in both REB-88 and REB-68, the examples
in REB-88 display a more pervasive degree of chalcocite replacement of pyrite than the examples in REB-68, further southeast (Fig. 22).

Southeast of the Mike deposit, down the hydraulic gradient from the inferred copper source, is the Copper King deposit, consisting of chrysocolla, malachite, azurite, and cuprite, localized in limestone units of an interbedded chert and shale sequence with minor limestone (Erd et al., 1953). This deposit formed at the outer edge of the envelope of decalcified and hornfelsed rock (Fig. 30F), in fresh carbonate rocks that effectively served to buffer the pH to neutral values where chrysocolla and copper carbonate minerals become stable (arrow "1b", Fig. 31). Interaction between acidic fluids and wallrock provided sufficient carbonate for malachite and azurite to form. Reduction of the same water as it descended toward the water table could have led to precipitation of cuprite if sufficient copper remained in solution (arrow "1c", Fig. 31).

**Stage-2 Supergene Copper**

At some point during or after Stage-1 supergene enrichment, the elevation of the water table decreased a significant distance. As a result, chalcocite blankets formed during Stage-1 supergene enrichment were now above the water table and began to oxidize, which leached the copper they contained into water percolating through the oxidation profile (Fig. 30G). This Eh-pH environment is shown by the dotted pattern in the diagram of Figure 32, at the tip of the arrow marked "2a". Conichalcite, chrysocolla, and malachite most likely precipitated directly from Cu^{++} bearing fluids, represented by the path marked "2b" in Figure 32, given that no replacement of cuprite by these minerals was observed. Minor cuprite observed with chrysocolla, malachite and conichalcite, may indicate locations where the supply of silica, carbonate, or arsenate and calcium was depleted, leaving excess copper in solution to form cuprite. Note that the conichalcite stability field (Fig. 28) covers much of the former cuprite field in addition to the entire tenorite and malachite fields of Figure 26, implying that a decrease in either arsenate or calcium activity would cause the conichalcite stability field to shrink in Eh-pH space, and the fluid would then likely precipitate cuprite.

Conichalcite and chrysocolla were the most likely minerals to form as a result of fluid path "2b". The high arsenic content of the Carlin-type pyrite released sufficient
Figure 32. Eh-pH representing Stage-2 supergene copper enrichment

Eh-pH diagram depicts the fluid pathways producing the Stage-2 supergene copper enrichment: chalcocite blanket oxidation, and local reprecipitation of that leached copper, primarily as conichalcite and chrysocolla. Details of the diagram are described in the text.
Stage-2

H₂O

Fe³⁺

HS⁻

SO₄²⁻

Cu²⁺

O₂

H₂S

chalocite

covellite

Mike

cuprite

conichalcite, chrysocolla

groundwater

hematite stable

pyrite stable

magnetite

Fe(OH)₃

a_{Cu^{2+}} = 2.6 \cdot 10^{-5}

a_{SO₄^{2-}} = 5.7 \cdot 10^{-4}

25°C, 1 atm
arsenate upon oxidation to make conichalcite thermodynamically stable. Feldspar destruction resulting from water-wallrock interaction not only contributes additional silica to solution, but also buffers the supergene fluids to a pH where chrysocolla becomes stable. The more arid climate inferred during Stage-2 enrichment would involve a lower input of water percolating through the oxidized zone, which decreases water-wallrock ratio, thus increasing the buffering capacity of the wallrock. The same wallrock would now be able to dominate the fluid pH to a greater extent than it did during Stage-1 enrichment, when the water-wallrock ratio was higher, and the same amount of potassium feldspar in the wallrock had less effect on the fluid. The smaller volume of fluid would effectively concentrate arsenic and aqueous silica released by water-wallrock interaction. The observation noted above that the first appearance of significant chrysocolla consistently forms at lower elevations than the first appearance of abundant conichalcite (Fig. 21) may be related to the fact that the copper-bearing fluid descending from the oxidizing chalcocite blanket would then have more time in contact with wallrock, to react with the wallrock and increase the aqueous silica concentration.

The pH of the fluids in the oxidizing chalcocite blanket would have varied between locations, primarily dependent upon the amount of relic pyrite in the chalcocite blanket, which would generate acid upon oxidation. Oxidation of chalcocite alone will not generate sufficient acid to mobilize the copper, as will the oxidation of a mixture of abundant relic pyrite with chalcocite (Tittley and Marozas, 1995). Other minerals which typically form during supergene enrichment, but are not observed anywhere in the Mike deposit, serve as further indicators of Eh-pH conditions during Stage-2 supergene enrichment. Jarosite is a mineral commonly formed during supergene copper enrichment (Tittley and Marozas, 1995), which has been shown to be stable at moderate to high Eh, and acidic environment of pH less than 3 (Brown, 1971). However, in spite of efforts described above, no jarosite was identified during this study, implying that the pH was greater than 3 during Stage-2 enrichment. From their study of iron oxide minerals in the supergene enrichment profile of La Escondida, Alpers and Brimhall (1989) note that jarosite is the dominant iron oxide mineral precipitating under strongly acidic conditions. Hematite predominates in the least acidic oxidative weathering environment, while goethite forms at pH conditions intermediate to those favoring jarosite or goethite. As
described above in the observations section, hematite and goethite are both abundant in the oxide zone of the Mike deposit, in approximately equal amounts, except for the copper zones, where hematite is dominant and very little goethite has formed. These observations of iron oxide minerals formed during Stage-2 enrichment help to constrain the pH to the zone with the dotted pattern on Figure 32, marking the Eh-pH conditions in an oxidizing chalcocite blanket.

The distance that copper migrates from its oxidized source to the location where it is redeposited is dependent on a combination of fluid acidity and the buffering capacity of the wallrock (Chávez, 2000). Thorough replacement of pyrite by chalcocite during Stage-1 supergene enrichment will produce a chalcocite blanket which will generate less acid upon oxidation during Stage-2, and the resulting copper-bearing fluid requires a smaller pH change for copper oxide minerals to become stable. As a result, this fluid would be more quickly buffered by wallrock to the pH required for mineral stability, limiting the mobility of the copper leached from that particular chalcocite blanket. There are no apparent large variations in the buffering capacity of the oxide zone rock between the holes logged during this study, implying that the relic pyrite content of chalcocite blankets may be the factor having the greatest impact on copper mobility during Stage-2 oxidation of chalcocite blankets. The initial pH of the oxidizing fluid, represented by a higher or lower pH location within the dotted oval on Figure 32, would therefore be the determining factor for the distance the remobilized copper could travel from its oxidizing chalcocite blanket source before being fixed as a copper oxide mineral, analogous to changes in the length of fluid path "2b".

For example, REB-88 contains significant copper concentration in an interval also occupied by oxidized chalcocite blankets, as described above. These oxidized chalcocite blankets from REB-88 show pervasive replacement of pyrite. There likely remained so little pyrite after Stage-1 chalcocite replacement, that Stage-2 oxidation of the chalcocite blanket generated very little acid. The copper remobilized during Stage-2 enrichment was therefore not able to travel more than tens of centimeters before being fixed in clays and local copper oxide minerals within and proximal to the oxidized chalcocite.

Meanwhile, REB-37 does not contain any significant copper concentration in the upper intervals of oxidized chalcocite blankets (Figs. 21, 22). The highest oxidized
chalocite blankets in REB-37, however, are near an interval of much more concentrated gold (Fig. 23) and the abundant arsenic-rich pyrite which accompanies that Carlin-type gold mineralization. The more abundant pyrite in REB-37 (relative to REB-88) was not completely replaced by chalcocite, and as a result, more pyrite was present among the chalcocite during oxidation. Stage-2 oxidation then generated more acid, and the copper in REB-37 during Stage-2 enrichment was more mobile, travelling further below the oxidized chalcocite blankets before being fixed, primarily as conichalcite deeper below the high grade gold zones in REB-37 (Figs. 21, 23). In locations where the initial pH in oxidizing chalcocite blankets was sufficiently acidic, the fluid could descend to the redox level before precipitating copper minerals, and replace sulfides below the water table with redeposited chalcocite.

Except for the case of the native copper in CV-16, as described above, cuprite and native copper in the Mike deposit typically occur together, as illustrated by the thick cuprite-native copper dominant zones of REB-90 (Fig. 21). As described above, very few other copper minerals are present in these cuprite-native copper dominant zones, which would imply that they likely formed by fluid path "2c" in Figure 32. Time may have played a role in the development of the co-existing cuprite and native copper, which typically occurs lower in the deposit. Copper may have first precipitated under moderately reducing Eh as native copper, but with the water table elevation lowering with time, there would have been a lower degree of pore saturation, and thus more available free oxygen in groundwater percolating through that location. Cuprite may then have formed later by partial oxidation of the native copper. Cuprite is also observed on fractures after copper or possibly directly replacing chalcocite (e.g. Fig. 17B). The cuprite and chalcocite stability fields share no boundaries in Eh-pH space, implying that this scenario of cuprite replacing chalcocite in situ is a metastable case, which is certainly possible as indicated by the experiments of Sato (1992), discussed above.

Stage-2 supergene copper enrichment may have continued until the burial of the bedrock surface by the deposition of the Carlin Formation (Fig. 30H). Burial by the Carlin Formation, and possible additional downfaulting of the Mike deposit by Basin and Range tectonism, caused the water table elevation to rise relative to the stratigraphic location of the Mike deposit. This increase of the water table elevation halted the Stage-2
supergene enrichment, and preserved the Mike deposit as it exists today. Normal faulting along northeastern striking faults down-dropped the area northwest of the Mike deposit prior to and following burial by the Carlin Formation. This tectonism dropped the inferred copper source further below its original elevation (Fig. 30H).

**Interactions between green copper oxide minerals**

The option of whether conichalcite, chrysocolla, malachite, or tenorite will precipitate in their shared stability field at the upper left of the Eh-pH diagram in Figure 32 is not due strictly to Eh and pH. The stability of these minerals relative to each other is primarily dependent upon the competitive relationship between activities of carbonate, arsenate and calcium, and silica, interacting with the Cu$^{+2}$ ion in solution under oxidized conditions. Conichalcite and chrysocolla possibly co-precipitate, as discussed below. In cases where the individual activities are high enough that more than one of these minerals are stable, the preferential formation of one over the others may be due to kinetic effects. For example, it is possible that intermediate copper-arsenate complexes temporarily bind the copper, which later precipitates as conichalcite, similar to the coagulation of aqueous silica by Cu$^{+2}$ ions into a gel as an intermediate stage in the formation of chrysocolla, as described by Yates et al. (1998). However, no thermodynamic data was found in slop98 (Shock and coworkers) to critically evaluate the possibility of copper-arsenate complex formation.

Conichalcite is considered a relatively rare mineral; the Mike deposit may be one of the world's largest concentrations of conichalcite. Its great abundance at Mike, together with stability relations as derived from experimentally determined solubility for conichalcite (Magalhaes et. al, 1988), implies that it is a stable mineral in over a large region of Eh-pH space (Fig. 28). The probable reason that conichalcite is not commonly observed in abundance is not that it is thermodynamically unstable, but rather that the complete set of precursor conditions necessary for conichalcite precipitation are not commonly met. Precipitation requires a minimum concentration of arsenate, calcium, and copper, and an oxidizing environment of mildly acidic to near neutral pH. Conichalcite is noted as a very minor product in the weathering profiles of some ore deposits, although the other localities where a significant amount of conichalcite is noted underwent supergene enrichment in arid climates (Bogoch, 1994; Ryall and Segnit,
Northeastern Nevada also appears to have become increasingly colder during the time of supergene enrichment of the Mike deposit (Hofstra et al., 1999), and likely more arid, supported by the evidence of a significantly lowered water table elevation during Stage-2 enrichment.

In locations at Mike where chrysocolla and conichalcite occur together, it appears that some chrysocolla formed first, lining the fracture walls as noted in the observations section (Fig. 15B). Additional chrysocolla may have filled the void space remaining in mineralized fractures after conichalcite precipitation (Fig. 15A), as implied by the chrysocolla attaching to a conichalcite blade in Figure 15B. It is unclear from the SEM/EDS studies if chrysocolla formed continuously before, during, and after the conichalcite, although no chrysocolla was incorporated into conichalcite in any of the samples analyzed. This relationship between the two minerals could perhaps be explained by the fact that chrysocolla first forms a gel before solidifying (Yates et al., 1998). Copper-silica gel must have begun to coagulate early, coating the fracture walls and partially filling the fracture, before the arsenate and calcium activities in solution became high enough to precipitate conichalcite. Conichalcite likely precipitated in its characteristic spherical forms before the chrysocolla gel solidified, whether or not the gel continued to accumulate additional copper-coagulated silica during and following conichalcite precipitation. As the solid chrysocolla took shape, it may have formed rare attachments to conichalcite crystals, similar to the example in Figure 15B. In most locations, the chrysocolla gel, which was likely displaced by conichalcite crystal growth, appears to have solidified in place. Solidification of chrysocolla from a gel may cause it to shrink, generating the micron-scale void space between conichalcite spheres and surrounding chrysocolla commonly observed (Fig. 15A).

In some places, chrysocolla and conichalcite are observed together on fractures (Fig. 14C), but are separated into mutually exclusive zones. This relationship may have been an example of the two copper minerals forming simultaneously, but the silica, or arsenate and calcium activities were locally higher at one site over the other, where precipitation of one mineral initiated and simply continued at each location. Perhaps it is also an example of either the conichalcite or the chrysocolla replacing the other, by the chemical reaction:
CaCu(AsO$_4$)(OH) + SiO$_2$(aq) + H$_2$O + H$^+$ = CuSiO$_3$·H$_2$O + Ca$^{++}$ + H$_2$AsO$_4^-$

A third possibility is that a chrysocolla gel formed, locally impeded fluid flow through that fracture, and then conichalcite precipitated from later fluid which was diverted around the chrysocolla through the unblocked part of the fracture.

Malachite is uncommon in the Mike deposit, encountered in scattered locations at the fringes of the oxide copper zone. Formation of malachite likely was restricted to chance encounters of the mildly acidic, Stage-2 groundwater with minor local carbonate remaining after the decarbonatization alteration associated with Carlin-type gold emplacement. The rarity of malachite implies that the carbonate activity value was low in the groundwater during supergene enrichment. Stability calculations using the assumed concentrations of the species in the system demonstrate that conichalcite is stable over malachite in all regions of Eh-pH space where malachite would precipitate. In spite of being present in the same general locations of the deposit, malachite is not observed in contact with conichalcite. The chemical reaction between conichalcite and malachite (reaction 8 in Table 7) may lead to direct replacement, whereas chrysocolla may not react directly with the other green copper oxides, because it forms an intermediate gel phase before solidification. No tenorite has been observed in the Mike deposit. It may not have precipitated simply because there was enough silica, arsenate and calcium, and/or carbonate in solution, to stabilize other copper oxide minerals over tenorite.

**Adsorption of base metals**

Base metals are closely associated with iron oxides, as shown by the electron microprobe analyses presented above (Figs. 19 and 20, Tables 4 and 5). Here I am concerned especially with measured concentration of base metals in core intervals where no base metal minerals are observed. The electron microprobe and other analytical techniques employed during this study do not yield the information necessary to choose between different mechanism of incorporation of these base metals into or onto the surfaces of the iron oxides. Auxiliary evidence points to the likelihood of adsorption of base metals to the iron oxides. Inner-sphere complexes, which bind metals tightly to the sorbent surface, unlike the electrostatic adsorption of outer-sphere complexes, are the sorption mechanism favored by both transition metals and arsenate (Krauskopf and Bird,
Newmont metallurgical leach tests from these intervals of abundant iron oxides lacking obvious base-metal minerals indicated extremely poor metal recovery, until a total acid digestion method was used, at which time metal recoveries were near complete.

Amorphous species, such as ferrihydrites and other secondary hydrated iron and manganese oxide minerals, are known to have a greater number of potential sorbing sites than well-structured minerals (Smith, 1999). The base metal and arsenic concentrations measured in Mike deposit iron oxyhydroxides, which could be amorphous or a combination of FeO(OH) structures (Table 4), are significantly higher than the concentrations in hematite (Table 5). Arsenic is highly concentrated in the samples, consistent with measured arsenate adsorption of over 95% total arsenic in the pH range 4-6 by ferrihydrite (data from Pierce and Moore, 1982, as recalculated by Davis and Kent, 1990). Given that this is comparable to the estimated pH range during supergene enrichment, and that the oxidizing environment implies that the arsenic in the Mike system was arsenate (Fig. 27), this high arsenic concentration in iron hydroxides at Mike is consistent with adsorption.

**Incompatibility between copper and iron oxide minerals**

Even using the electron microprobe to resolve fine-scale textural relations, hematite and cuprite were nowhere observed in contact with each other (e.g., Fig. 19), implying that the two minerals are not in chemical equilibrium. This observation is consistent with experiments of Yund and Kullerud (1964) in the system Cu-Fe-O. Their work concludes that cuprite and hematite would not coexist at any temperature, but each should coexist with an intermediate phase, the mineral delafossite, CuFeO$_2$. Neither delafossite, nor any other copper-iron oxide phases, were identified in samples from the Mike deposit. With the activity of iron and copper, and with oxide minerals of both metals precipitating in the system, the lack of delafossite would imply that the low temperature environment of oxidative weathering was not at complete chemical equilibrium.
SUMMARY AND CONCLUSIONS

Summary of age constraints

Many of the critical crosscutting relationships that could be used to evaluate the relative age of various geological features at the Mike deposit have not been seen in drill core. However, the few radiogenic dates and the crosscutting relationships that have been observed allow a first approximation of the timing of different events.

Contact metamorphism, quartz veins, and base-metals

Barren quartz veins are ubiquitous in the hornfels zone of the Mike deposit and are rare outside that zone. Quartz veins may be coeval with thermal metamorphism, or may be younger and related to the greater fracture permeability of hornfelsed rocks contrasted with unaltered Rodeo Creek, Popovich and Roberts Mountains lithologies. No absolute radiometric dates are tied to the hornfelsing event and Barren quartz veins, although both are likely the result of emplacement of granitoids along the central Carlin Trend, which occurred at 106 and 37 Ma (Evans, 1980).

Although the base-metal sulfide veins at Mike do not appear to be the source of the supergene copper there, it is likely that introduction of base-metal sulfides at Mike coincided in time with emplacement of the metal source for the secondary copper minerals. Barring the possibility of multiple episodes of base-metal sulfide veins, these veins formed at some point in time after the intrusion of the dikes at Mike, which the veins crosscut. Potassium feldspar from an altered dike at Mike has been dated by K/Ar methods at 107±2 Ma implying that the dike is at least that age if the potassium feldspar date represents a potassic alteration event (Teal and Branham, 1997). This Cretaceous date is much earlier than the inferred Eocene age of the gold mineralization in the Carlin Trend. Although the veins crosscut the dikes, they may be broadly coeval with dike emplacement. The veins are limited to the thoroughly hornfelsed parts of the deposit, and therefore the veins and the dikes may be products of the same thermal event. The potential source of fluids related to these veins is unknown, although the Richmond stock has been proposed as a potential source, given the demonstrated affiliation of this stock
with sphalerite in surface exposures (P. Mallette, pers. commun., 1999). The Richmond stock has been dated at 106±2 Ma by K/Ar dating methods on biotite (Evans, 1980).

A sericite grain from a selvage on a base-metal sulfide vein in the Mike deposit yielded a K/Ar date of 71.6±1.8 Ma (Branham, 1994), which would provide a younger bound on the age of the pyrite-base-metal or sphalerite-dominant veins. However, this sericite sample comes from an assay interval of very high gold values, indicating the possibility that Carlin-type hydrothermal fluids altered or partially reset the grain. It is thus unclear whether this date has any relevance to the emplacement of the original copper into the local area, although if this date is accurate for that event, it would rule out the Richmond stock and related magmatism as a potential source of the exotic copper. However, no known local igneous activity has been dated at this age to corroborate this radiometric date for sphalerite, casting some doubt on its validity.

It may be possible to determine a date for the sphalerite in the sphalerite-dominant (or even pyrite-base-metal) veins by the Rb-Sr method for sphalerite devised by Nakai et al. (1990), although results may be of limited value because of the large uncertainties involved with this dating technique (Pettke and Diamond, 1996).

Calcite and silica-orpiment veins

The spatial correlation of the carbonate veins with the periphery of the gold-bearing portion of the deposit implies that these veins are the products of Carlin-style hydrothermal activity. Because calcite veinlets are observed cutting sphalerite-dominant veins, the paragenetic sequence establishes that the gold mineralizing event occurred after the emplacement of hypogene base metal sulfides. Silica-orpiment veins, which are thought by the majority of Carlin researchers to represent a late stage of Carlin-style mineralization, also are interpreted as younger than these base metal sulfide veins.

Supergene copper

Copper enrichment at the Mike deposit clearly post-dates gold mineralization, given the fact that the supergene copper can be related to a redox profile parallel to the 15 Ma Carlin Formation unconformity. Copper remobilization from a distal source depended on near-surface and surface oxidation, which leached copper and put it into solution. The transport by groundwater and sites of eventual fixation of copper as secondary minerals were affected by the elevation (and changes in the elevation) of the
water table. Although there may be minor disagreement among Carlin researchers regarding the depth of formation of the gold deposits, they certainly formed at much greater depth than the near-surface oxidation that affected the area prior to emplacement of the Carlin Formation at 15 Ma. Supergene alunite from the Mike deposit, interpreted to be supergene, yielded a K/Ar date of 19.7 Ma (Teal and Branham, 1997). Whether or not the supergene enrichment process was brought to an end by a change to arid conditions, burial of the area by the Carlin Formation certainly halted the supergene process and preserved the profile currently observed in drill core. A minimum age for supergene enrichment can be placed at 15.1-14.7 Ma, based on $^{39}\text{Ar}/^{40}\text{Ar}$ dates of the Carlin Formation in the Carlin Trend (Fleck et al., 1998).

**Conclusions Regarding Copper Enrichment**

The focus of this study has been to document the mineralogy of copper-bearing phases and interpret the geochemical mechanisms responsible for creating a supergene copper deposit among the gold deposits of the Carlin Trend. A secondary objective was to identify the source of the copper at Mike. The main conclusions are summarized below.

1) The source of the copper was exotic, and most likely was located northwest of the Mike deposit. The style of hypogene copper mineralization is unknown, but by analogy with base-metal sulfide veins present in the Mike area and with other known base-metal occurrences in the Carlin Trend, the source most likely consisted of disseminated and vein pyrite, chalcopyrite, and sphalerite related to the emplacement of a Mesozoic stock.

2) Enrichment is interpreted to have taken place in two broad stages (Figs. 31 and 32), although each "stage" may have involved repeated events. During Stage-1, chalcocite was formed in reducing, pyritic zones of the Mike gold deposit at the water table. Oxidized and acidic near-surface waters transported some of the copper beyond Mike, eventually reaching the carbonate fringe, where the waters were neutralized at
high Eh and precipitated chrysocolla, malachite, and azurite to form the exotic copper oxide deposit at Copper King. No prior Carlin-style hydrothermal activity had occurred there, and conichalcite is unknown at this locality.

3) A second stage of supergene enrichment was focused on the Mike deposit, where chalcocite blankets oxidized as the water table fell. The remobilized copper was fixed nearby as copper oxide minerals, primarily conichalcite and chrysocolla.

4) Conichalcite is thought to be a rare mineral, although on the basis of thermodynamic calculations, its stability field is large relative to both pH and Eh. The abundance of conichalcite at the Mike deposit supports this prediction. The rarity of conichalcite in other settings where copper enrichment has taken place is due to the requirement of elevated arsenic concentrations in near-neutral pH waters which also contain sufficient copper. These rare conditions were met at Mike because of the presence of chalcocite blankets oxidizing in rocks containing arsenic-rich pyrite.

5) The descending, mildly-acidic, copper-bearing fluids derived from oxidation of former chalcocite blankets precipitated first conichalcite, then chrysocolla deeper in the profile. This pattern of distribution is thought to be dependent on the degree of water-rock interaction, with chrysocolla forming only after hydrolysis reactions with alkali feldspar had resulted in an elevated silica concentration in solution.

6) Supergene chalcocite enrichment occurred at the Mike deposit because of the prior formation of the Carlin-type gold deposit at that location. Carlin-style hydrothermal activity led to decarbonatization and removal of the most effective mechanism to buffer the pH of the migrating copper-bearing fluids. At the same time, introduction of pyrite provided the reducing agent and sulfur source to fix copper as chalcocite. Upon oxidation, the high arsenic concentration of the Carlin-type pyrite provided the arsenate necessary to stabilize conichalcite, thus more effectively fixing copper during Stage 2 supergene enrichment, when copper was remobilized through oxidation of the earlier formed chalcocite blankets.
7) Conichalcite could serve as a pathfinder for "invisible gold" deposits of the Carlin-type. The recognition of conichalcite in a prospect, in the absence of other arsenic-bearing phases, provides a useful exploration tool, a natural analogue of dusting for the fingerprints left by Carlin-type, gold-bearing hydrothermal fluids.
REFERENCES


Occidental Minerals, 1974, Status Report, Copper King Project (308), Eureka County, Nevada, internal company report, 18 p.


