LA HERRADURA ORE DEPOSIT: AN OROGENIC GOLD DEPOSIT IN NORTHEASTERN MEXICO.

by

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STATEMENT BY THE AUTHOR

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# TABLE OF CONTENTS

LIST OF FIGURES ........................................................................................................... 7

LIST OF TABLES .............................................................................................................. 9

ABSTRACT ....................................................................................................................... 10

CHAPTER 1: INTRODUCTION ......................................................................................... 12
  1.1. Introduction and Previous Work ............................................................................. 12
  1.2. Mining and Exploration History ............................................................................ 13

CHAPTER 2: GEOLOGIC AND TECTONIC SETTING ..................................................... 14
  2.1. Regional Geologic and Tectonic Setting ............................................................... 14
    2.1.1. Pre-Cretaceous Lithologies and Evolution: The Mojave-Sonora Megashear Hypothesis ........................................................................................................ 14
    2.1.2. Cretaceous and Younger Events and Lithologies .......................................... 18
  2.2. The Geologic Setting of La Herradura ................................................................. 20
    2.2.1. Precambrian Lithologies ................................................................................. 20
    2.2.2. Paleozoic Units ............................................................................................ 24
    2.2.3. Jurassic Units ................................................................................................ 24
    2.2.4. Tertiary and Quaternary Units ...................................................................... 25
  2.3. The Structural Setting of La Herradura ............................................................... 25
  2.4. Alteration of La Herradura ................................................................................... 26

CHAPTER 3: U-PB AND RE-OS GEOCHRONOLOGY FROM LA HERRADURA GOLD DEPOSITS ...................................................................................................................... 30
  3.1. Zircon U-Pb Dating ............................................................................................... 30
    3.1.1. Sampling ........................................................................................................ 30
    3.1.2. Preparation of Samples .................................................................................. 30
    3.1.3. Analytical procedures ..................................................................................... 30
3.1.4. Results.........................................................................................31

3.2. Pyrite-Magnetite Re-Os Dating.................................................................37
   3.2.1. Sampling.....................................................................................37
   3.2.2. Analytical Methods.....................................................................37
   3.2.3. Results........................................................................................38

CHAPTER 4: STABLE ISOTOPES..................................................................40
   4.1. Introduction.....................................................................................40
   4.2. Oxygen and Hydrogen Isotopes.........................................................41
      4.2.1. Sampling..................................................................................41
      4.2.2. Analytical Methods..................................................................45
      4.2.3. Analytical Results.....................................................................46
   4.3. Sulfur isotopes....................................................................................47
      4.3.1. Sampling..................................................................................47
      4.3.2. Analytical Methods..................................................................49
      4.3.3. Analytical Results.....................................................................50
      4.3.4 Sulfur Isotope Geothermometry....................................................50
   4.4. Carbon Isotopes..................................................................................53
      4.4.1. Sampling..................................................................................53
      4.4.2. Analytical Methods..................................................................54
      4.4.3. Analytical Results.....................................................................55
   4.5. Interpretation and Discussion...............................................................57

CHAPTER 5: FLUID INCLUSIONS.................................................................59
TABLE OF CONTENTS – continued

5.1. Introduction ........................................................................................................59
5.2. Sampling ...........................................................................................................60
5.3. Analytical Methods .........................................................................................61
5.4. Petrographic Characteristics of Fluid Inclusions .............................................65
5.5. Analytical Results ..........................................................................................67
5.6. Interpretation and Discussion .........................................................................72

CHAPTER 6: DISCUSSION AND CONCLUSIONS ............................................74
6.1. Discussion ......................................................................................................74
6.2. Conclusions ..................................................................................................85

REFERENCES
LIST OF FIGURES

Figure 2.1. Regional geological map of the La Herradura area..........................17
Figure 2.2. La Herradura geological map and cross section...............................22
Figure 2.3. Photomicrographs of representative rocks......................................23
Figure 2.4. Photograph of Victoria shear zone separating Proterozoic gneiss from Jurassic volcanic rocks.................................................................24
Figure 3.1. Map showing location of samples for U-Pb ages...............................35
Figure 3.2. U-Pb ages of zircons........................................................................36
Figure 3.3. Map of La Herradura mine showing locations of samples used for Re-Os isotopes.............................................................38
Figure 3.4. Re-Os isochron for La Herradura gold deposit.................................39
Figure 4.1. Map of La Herradura mine showing locations of samples used for oxygen and hydrogen isotopes.........................................................42
Figure 4.2. Photographs of representative samples used for stable isotope studies........44
Figure 4.3. Map of La Herradura mine, showing location of samples used for sulfur isotopes.................................................................48
Figure 4.4. Reflected light microphotography showing pyrite with strong fracturing and later gold mineralization in the fractures.................................49
Figure 4.5. Plot of $\delta^{34}S$ versus elevation....................................................53
Figure 4.6 Map of La Herradura mine, showing location of samples used for carbon isotopes.................................................................54
Figure 4.7 Plot of $\delta^{13}C$ versus 1/concentration of C........................................56
Figure 4.8. Plot of $\delta^D$ and $\delta^{18}O$ of water coexisting with gangue minerals at 330°C.................................................................58
Figure 5.1. Map of La Herradura mine showing locations of samples used for fluid inclusions petrography and microthermometry studies..........................62
Figure 5.2. Photograph showing shear veins and associated flat extension veins.....63
LIST OF FIGURES-Continued

Figure 5.3. Photograph showing extensional quartz vein with sericite-pyrite envelope ..............................................................................................................64

Figure 5.4. Photomicrograph of extensional quartz veins.................................................64

Figure 5.5. Microphotograph showing quartz vein with carbonate and Fe oxides as a selvage........................................................................................................65

Figure 5.6. Photomicrograph showing fluid inclusions......................................................66

Figure 5.7. Clathrate melting temperatures for all samples..............................................70

Figure 5.8. Temperatures of homogenization of CO₂ (to liquid) for all samples..............70

Figure 5.9. Final homogenization temperatures for all samples......................................71

Figure 5.10. Plot of maximum and minimum isochors for fluid inclusions......................71

Figure 6.1. Plot of δD Vs δ¹⁸O of water coexisting with gangue minerals.......................84
LIST OF TABLES

Table 2.1. Tectonic and deformation history from La Herradura gold deposit........28

Table 2.2. Paragenetic sequence of ore and gangue minerals from La Herradura gold deposit.................................................................29

Table 3.1. U-Pb ages from La Herradura gold deposit...............................35

Table 3.2. Results of analysis from Re-Os isotopes...................................39

Table 4.1. Description of samples for oxygen and hydrogen isotopes...........43

Table 4.2. Oxygen and Hydrogen isotope data........................................47

Table 4.3. S isotope database and calculated temperatures for deposition of sulfides.........................................................................52

Table 4.4. Carbon content and $\delta^{13}C$ values of fluid inclusion-bearing quartz and ankerite.................................................................54

Table 5.1. Description of samples used for fluid inclusion petrography-
Microthermometry................................................................................63

Table 5.2. Microthermometry summary...................................................72

Table 6.1. Oxygen and Hydrogen compositions of quartz-carbonate veins in the western North American Cordillera and counterparts worldwide.........83

Table 6.2. Comparison of fluid inclusions results from La Herradura with OGD located in the circum-Pacific rim of North and South America........84
ABSTRACT

The production plus reserves of La Herradura mine is 5.4 million ounces of gold. The deposit occurs within a northwest trending belt of metamorphic rocks of greenschist and amphibolite facies and granitoids of Proterozoic age. The ore deposit is hosted mainly in a 1.75 Ga quartz-feldspathic gneiss and strikes northwest. The orebody is bordered to the east by the Victoria shear zone, separating Jurassic clastic rocks and subvolcanic intrusions. To the west the orebody is bordered by the Ocotillo shear zone, separating Upper Paleozoic limestone.

Most of the gold mineralization occurs within thin quartz-sulfide veins that follow en echelon tensional swarms. These veins dip at low angles (<30°) and are interconnected to a thicker set of fault-fill higher-grade quartz vein systems. These higher-grade quartz vein systems are believed to be the feeders for gold mineralization in the deposit.

Oxygen isotopes of several quartz veins show consistent δ^{18}O values between +14 and +16 ‰. The calculated δ^{18}O values for the waters in equilibrium at 330°C for these quartz veins range between +8 to +10 ‰. These values are similar to those of metamorphic waters. Carbon isotopes from fluids inclusions-bearing quartz veins and ankerite, suggest that there are two sources of C, one associated with carbonate minerals, δ^{13}C=0 to -6 ‰, and other associated only with fluid inclusions, δ^{13}C < -18 ‰. The low δ^{13}C carbon source is organic C. Sulfur isotopes of sulfides in quartz-pyrite-galena-sphalerite veins from the Centauro pit have δ^{34}S values from +3.1 to +6.6 ‰. Geothermometry studies using pyrite-galena pairs show equilibrium temperatures on the order of 330 ±20°C for the quartz-sulfide-gold-bearing veins.
Fluid inclusion studies from these same veins show that the ore fluids are CO₂-rich, are low in salinity, have homogenization temperatures between 285-295 °C, and have calculated trapping pressures between 1.6 and 2.7 kbar. The fluids responsible for the mineralization are interpreted to have a metamorphic origin generated during the Laramide orogeny, as is consistent with isotopic measurements of Re-Os from pyrite and magnetite that accompany gold mineralization which is dated at 61.0 ± 2.1 Ma.

The style and geometry of mineralization, stable isotopes and fluid inclusions studies, show that La Herradura gold deposit can be classified as an orogenic gold-type deposit, the largest and the first orogenic gold-type deposit described in Mexico.
CHAPTER 1: INTRODUCTION

1.1. INTRODUCTION AND PREVIOUS WORK

The northwestern part of Mexico has similar geology to the southwestern United States. The most outstanding geologic feature in Mexico, however, is the postulated Mojave-Sonora megashear, which is suggested to have occurred during Late Jurassic time and separates two Proterozoic blocks of difference ages (Silver and Anderson, 1974). Nourse et al. (2005) alternatively suggest that the two different sequences of Proterozoic rocks in the region resulted from Paleoproterozoic accretion. Iriondo (2001) indicates that juxtaposition of Paleoproterozoic blocks in the Quitovac area represents a Laramide thrust zone and not the original Proterozoic suture between the basement rocks. Regardless of the precise reason, a large-scale structure occurs with the distribution of gold mineralization in northwest of Sonora. Consequently, previous works on gold occurrences in Sonora emphasize the relationship between gold mineralization and this large-scale structure (Silberman and Giles, 1988).

For La Herradura gold deposit, several internal studies from mining companies focus on the geology of the deposit in order to understand the genesis of the mineralization (Altamirano and Carreón, 1995; López and Díaz de León, 1995; Salvatierra and Novelo, 1995, de la Garza et al., 1998). Later, during the development of the mine, new studies are dealing with the structural framework and the geology of the deposit (Caddey, 2002; Caddey and García, 2003; de la Torre, 2004; Rhys, 2005; and Romero, 2005).

This study mainly addresses the geochemistry and origin of the ore fluids involved in the formation of the La Herradura gold deposit; in order to explain the geochemistry of the ore bearing fluids and to compare this deposit with orogenic gold deposits located in the Circum-Pacific rim, especially those in North America and Peru.
1.2. MINING AND EXPLORATION HISTORY

The La Herradura mine contains 5.4 million ounces of contained gold in production plus reserves. The deposit is owned by Minera Penmont, a Joint Venture between Peñoles and Newmont. As a result of an aggressive grassroots exploration program in northwestern Mexico that started in 1987, the first economic drill intersection in La Herradura came in 1991 (100m @ 0.85 g/t Au). Subsequent and continuous drilling campaigns resulted in the definition of an orebody containing 1.7 M oz by May 1998, when mine operations started. To date, 2 M oz of gold have been produced. Present reserves are 3.4 M oz of gold in ore with an average grade of 1 g/t, using a cut-off of 0.35 g/t Au. The mine produces 210,000 ounces of gold per year (Jose de la Torre, pers. commun., 2008).
CHAPTER 2: GEOLOGIC AND TECTONIC SETTING

2.1. Regional Geologic and Tectonic Setting

La Herradura mine is located in northwestern Sonora, Mexico. This deposit occurs within a northwest trending belt that consists of metamorphic rocks of greenschist and amphibolite facies and granitoids of Proterozoic age (Nourse et al., 2005). These rocks are intruded by a series of Triassic and Middle Jurassic granitoids and are overlain by younger sedimentary and volcanic rocks of Middle to Late Jurassic age (Figure 2.1). All these units are intruded by Late Cretaceous to early Tertiary granitoids related to the Laramide orogeny and are overlain by Miocene rhyolites, andesites, and basalts and Quaternary basalts. Basin and Range tectonics affect this area, as they do much of Sonora and adjacent Arizona. Basin and Range faulting occurred in the mid to late Tertiary. Faulting resulted in the formation of NW-trending linear ranges of crystalline rock, separated by deep basins filled with sand and gravel derived from the ranges (Figure 2.1). Correlation is difficult between ranges.

2.1.1. Pre-Cretaceous lithologies and evolution: The Sonora-Mojave megashear hypothesis

The interpretation of the pre-Cretaceous geological and tectonic evolution of the region is highly controversial and focuses on the possible existence of the Sonora-Mojave megashear, which is believed to have an important impact on the distribution of gold mineralization within the region. Many of the gold prospects in Sonora occur within or adjacent to the southwestern boundary of this regional fault that occurs
within Precambrian, Mesozoic, and Tertiary rocks (Silberman and Giles, 1988).

Evidence for the Mojave-Sonora megashear is based mainly on geochronological studies of the Proterozoic basement that indicate two different basement blocks with slightly different ages and interpreted differences in Paleozoic to Jurassic cover sequences across the proposed northwest-trending shear trace (Silver and Anderson, 1974). The trace of the shear zone is interpreted to pass north of the town of Caborca and to the west of Quitovac and Sonoyta before extending into southern California. The Caborca block to the west consists predominately of amphibolite and greenschist grade siliciclastic paragneiss and orthogneiss with ages ranging from 1.8 to 1.7 Ga. The North America block to the east consists predominately of orthogneiss and volcanosedimentary paragneiss with ages varying from 1.7 to 1.6 Ga old. La Herradura gold deposit is located in the Caborca block.

The Caborca block, in addition to containing the older Proterozoic basement, is also distinguished by the local presence of Neoproterozoic and Paleozoic strata, including carbonate sequences, that occur west of Caborca and near the La Herradura mine. Such strata are absent, or have non-correlative sequences northeast of the proposed trace of the megashear and east of the region extending from Sonoyta to southeast of Caborca. Pre-Jurassic lithologies are not exposed on the North America block in an area termed the Papago terrane (Haxel et al., 1984; Calmus and Sosson, 1995). Other sequences potentially unique to the Caborca block include Lower to Middle Jurassic clastic, volcanic, and plutonic rocks, which differ from the predominantly conglomeratic Upper Jurassic to lowermost Cretaceous sedimentary rocks, and rarity of older Jurassic units, northeast of the proposed Megashear trace (Anderson et al., 2005). Some of these Jurassic to Lower Cretaceous sequences are
proposed to have been deposited in pull-apart basins during active displacement along the Sonora-Mojave Megashear (Anderson and Nourse, 2005).

Based on the differences in Proterozoic and Phanerozoic sequences across the proposed trace of the megashear and the general age and lithological similarities of the rocks to the northeast of the megashear in California, several authors (e.g., Anderson and Silver, 2005; Stewart, 2005) propose approximately 800-1000 km of left-lateral displacement along the structure. The timing of the displacement suggested by Anderson and Silver (2005) is constrained to Late Jurassic by the youngest affected rocks (mid-Jurassic) and oldest unaffected and overlapping lithologies (Cretaceous). This time corresponds to the initial opening of the Gulf of Mexico. Alternative hypotheses which also invoke similar large scale displacements include the California-Coahuila transform, involving a major Late Paleozoic to Triassic left-lateral translation and juxtaposition of the Caborca block in a manner comparable to that of the Jurassic megashear hypothesis (Dickinson, 2000). An alternative to the megashear hypothesis is proposed by Nourse et al. (2005). They suggest that the two packages of Proterozoic rocks in the region are the result of Paleoproterozoic accretion.

Exposures of northwest trending mylonitic rocks 13 km southwest of the town of Sonoyta near the U.S. – Mexico border (Figure 2.1) have been interpreted by Campbell and Anderson (1998, 2003) to be the exposed fault rocks of the Sonora-Mojave megashear. Mylonites in this location affect rocks interpreted to be as young as Lower Jurassic age but are overprinted by Late Cretaceous granite, suggesting that the mylonites formed between these two times. The mylonites contain both sinistral kinematic indicators and locally shallow thrusts associated with down-dip
Figure 2.1. Regional Geological Map of the La Herradura area. Modified from Iriondo et al. (2005). Notice the location of Figure 2.2. MSM refers to the proposed locations of the Mojave-Sonora Megashear from Anderson and Silver (2005). MSM (alt) is the alternative position of this possible structure from Nourse et al. (2005).
plunging lineation indicating reverse shear, which collectively are interpreted as sinistral transpression by Campbell and Anderson (2003). However, it is also possible that these mylonites represent the effects of Laramide deformation in the area as documented by Iriondo et al. (2005) in Quitovac 40 km to the southeast, and thus postdate the proposed megashear. Elsewhere, the potential trace of the megashear is hidden under Tertiary and Quaternary rocks or is overprinted by Late Cretaceous plutons and the effects of Laramide deformation and mid-Tertiary Basin and Range extension. Further age dating, mapping, and structural studies are necessary to place the trace of the MSM into a more detailed context.

The presence of large-scale structures is a characteristic widely documented at OGD worldwide. For example, these regional structures are observed in the early-Eocene orogenic gold deposits in the Juneau gold belt, Alaska (Goldfarb, 1988), the late-Jurassic and Early Cretaceous Mother Lode in California (Böhlke and Kistler, 1986; Elder and Cashman, 1992), the Carboniferous orogenic gold deposits in Peru (Haeberlin, 2002), the Archean deposits in the Superior and Yilgarn cratons in Canada and Australia, respectively (Robert et al., 2005). Consequently, large-scale structure, related either to MSM or Paleoproterozoic accretion, is a positive factor in the exploration for OGD in northwestern Sonora, where several gold occurrences exist, and can be related at least in space to the same reservoir.

2.1.2. Cretaceous and younger events and lithologies

The Upper Cretaceous intrusions and all pre-Cretaceous units have been affected by regional Late Cretaceous to early Tertiary Laramide deformation, which is constrained in the Quitovac region, 40 km northeast of La Herradura, by the age of
the affected Cretaceous intrusions, $^{40}$Ar-$^{39}$Ar studies of metamorphic rocks in associated thrusts, and post-Laramide volcanosedimentary sequences range between 75 and 39 Ma (Iriondo et al., 2005). Laramide deformation in the area is associated with development of mainly northeast vergent folds and thrusts, inhomogeneous east-west to northwest trending foliation in lower greenschist grade rocks, and generally steeply plunging elongation lineation which formed in response to regional northeast-southwest to north-south directed shortening (e.g., Nourse, 2001; Iriondo et al., 2005).

Post-Laramide tectonic events in the region are dominated by extensional and transtensional faults that affect rocks as young as the Miocene calc-alkaline volcanic rocks of the El Pilar sequence, which occurs east of the La Herradura deposit (Iriondo et al., 2005). Mid-Tertiary, Basin and Range extension in the region is widespread. It is characterized by local development of metamorphic core complexes and shallow dipping foliations and by northeast- and southwest-dipping normal and detachment faults that locally remobilize Laramide thrust surfaces. These latter structures are probably largely responsible for the current topography, which is dominated by north to northwest trending outcrop ridges and intervening covered valley areas. In La Herradura district, these faults affect both Precambrian and Jurassic blocks and displace the orebody (See section in Figure 2.2). A second set of northeast trending dextral faults identified by Iriondo et al. (2005) in the Quitovac area overprint the northwest trending extensional structures. Late Miocene to Recent tectonics in the region is dominated by dextral activity of the northwest trending San Andreas and subsidiary fault systems associated with opening of the Gulf of California.

The youngest rocks in the area are basaltic lava flows associated with Cerro Pinacate, approximately 80 km north of La Herradura. A series of probable coeval
isolated minor eruptive centers, dikes, and erosional basalt flow remnants that occur along valleys in the region are potentially related to volcanism along dextral transform faults related to the San Andreas Fault system.

2.2. The Geologic Setting of La Herradura

La Herradura mine occurs within a northwest trending belt of Proterozoic rocks consisting of greenschist and amphibolite grade metamorphic rocks and granitoids. The deposit is hosted in biotite-quartz-feldspar and quartz-feldspathic gneisses that are bordered to the east by Jurassic clastic rocks and subvolcanic intrusions and to the west by upper Paleozoic limestone. Isolated outcrops of fresh andesite, trachyte, and basalt occur locally northeast of the mine.

2.2.1. Precambrian lithologies

The belt of Precambrian rocks that are the host rocks for La Herradura mine is part of the Caborca block as described by Silver and Anderson (1974). These rocks have also been described at Sierra Pinta (Araux-Sánchez, 2000) located 20 km to the north of La Herradura and to the southeast near the town of Caborca (Anderson and Silver, 2005; Nourse et al., 2005). This Proterozoic block comprises a northwest elongate, 2-km wide, Proterozoic structural block (de la Torre, 2004) that is bordered by two large shear zones. The eastern Victoria shear zone separates Precambrian rocks from Jurassic rocks (Figure 2.2 and 2.5), and the western Ocotillo shear zone separates Proterozoic rocks from Paleozoic rocks. The age for the gneiss ranges between 1.7 to 1.8 Ga (Silver and Anderson, 1974) and are coincident with new
isotopic U-Pb ages from zircons obtained in this study. These host rocks are composed principally of biotite-quartz-feldspar and biotite-poor quartz-feldspathic paragneiss. The protolith represents a metamorphosed clastic sequence of impure-sandstone-siltstone, and arkosic sandstone to quartzite. Pegmatite dikes and segregations dominated by K-feldspar-quartz also are common within this unit. Gradational contacts of the pegmatites with surroundings lithologies and common occurrences of coarse aggregates of anatectic K-feldspar + quartz + biotite segregations in the gneiss sequence suggest that the pegmatites formed by partial melting of the gneiss sequence at or near peak metamorphic conditions. Locally, coarser grained bodies of probable tonalite and granite composition are present to the northwest of Centauro Pit. The paragneisses are weakly foliated, internally massive to crudely banded biotite-poor quartz-feldspathic gneiss, forming several lenses up to 300 m thick. These rocks normally exhibit strong fracturing and are crosscut by numerous quartz veins. In thin section, the quartz feldspar gneiss is composed primarily of quartz and K-feldspar. Secondary minerals, in the core of the mineralized zone, are mainly albite and sericite replacing feldspars (Figure 2.3A), and relics of fresh and/or oxidized pyrite and hematite after pyrite. In the periphery of the mineralized zone, the plagioclase and biotite of the quartz-feldspar-biotite gneiss have been replaced partially by chlorite and epidote.
Figure 2.2. La Herradura geologic map and cross section. Modified from de la Torre (2004).
Figure 2.3. Photomicrographs. A is showing quartz-feldspar gneiss under microscope. Note the abundant sericite that is interpreted to have been derived mainly from metamorphic reactions and to a lesser extent by hydrothermal replacement of feldspar. Polarized light. B is showing andesite under microscope. Note strong chlorite-epidote replacing previous ferromagnesian minerals. Polarized light.
2.2.2. Paleozoic units

The oldest Phanerozoic lithologies in the area are Paleozoic limestone with subordinate quartzite and chert that lie to the west and northwest of La Herradura mine (Figure 2.2). Although these lithologies are frequently recrystallized to marble and highly strained, they still contain recognizable conodonts that range in age from Mississippian to Permian (Poole et al., 2004). This unit is separated from the mineralized Proterozoic gneiss block by the Ocotillo shear zone and does not host gold mineralization.

2.2.3. Jurassic units

To the east and northeast of the Proterozoic gneiss belt, and separated by the Victoria Shear zone, lies a Jurassic unit, composed of a diverse assemblage of clastic sedimentary rocks, rhyolitic sills and/or extrusive equivalents and andesitic
flows/dikes (Figure 2.2). These units are comparable to those units described previously in the Caborca block (Anderson and Silver, 2005). Under the microscope, the rhyolite exhibits a primary mineralogy dominated by quartz and feldspar. The secondary minerals are epidote, sphene, chlorite, sericite and calcite. This rock is weakly metamorphosed or contains only lower greenschist grade assemblages. The andesite shows a strong propylitic alteration where ferromagnesian minerals have been replaced by chlorite, epidote, and sphene (Figure 2.3B). In general, this rock is not deformed or metamorphosed. These Jurassic units are not affected by the gold mineralization in La Herradura mine.

2.2.4. Tertiary and Quaternary Units

Tertiary rocks are andesites and trachytes. These rocks outcrop to the northeast of La Herradura district and generally trend northwesterly. Younger basalts outcropping northeast of Centauro pit are associated with the Pinacate volcanic field, located 80 km north of the La Herradura. Outcrops of Quaternary basalts also have a general northwest trend.

2.3. The Structural Setting of La Herradura

As is published by de la Torre (2004) and Romero (2005), the orebody of La Herradura is hosted within shear zones in the more competent quartz-feldspar gneiss. Gold mineralization is bounded to the east and west by the Victoria and Ocotillo shear zones, respectively. Shear zones exhibit a northwest trend and are controlled by pre-mineral reverse faults that appear to have been reactivated during Laramide time. Most of the gold mineralization occurs within thin quartz-sulfide (leached in outcrop)
veins that follow en echelon tensional swarms. These veins are dip at low angles (<30°) and are interconnected to a thicker set of fault-fill higher-grade quartz vein systems. These higher-grade quartz vein systems are believed to be the feeders for gold mineralization in the deposit. At depth, these veins may be coalescing into one single large shear fault system that served as a channel for the ascending of ore fluids. In La Herradura mine, there is a strong correlation between the grade of gold in the thin extensional quartz veins and the proximity to fault-fill higher-grade quartz vein systems. Robert and Poulsen (2001) noted that extensional veins and vein arrays are generally of lower economic significance, but they can be important ore hosts in competent rocks or in areas adjacent to the fault-fill vein systems.

Based on structural mapping in the La Herradura mine area, it is possible to identify at least five tectonics events superimposed on all stratigraphic units outcropping in this area (de la Torre, 2004; Romero 2005, Table 2.1). These observations indicate that gold mineralization is associated with the third tectonic event, and they also tend to constrain the age of this mineralization to between 80 and 45 Ma.

2.4. Alteration of La Herradura

Reported alterations assemblages of this deposit (de la Torre, 2004; Romero, 2005) are quartz-sericite-albite in the core of the deposit and selectively follow the quartz-feldspar gneiss bands in the outer zones of the deposit. Iron-carbonates (ankerite-siderite) are widespread within the deposit, mainly restricted to haloes adjacent to quartz-sulfide veins within the core of the orebody. Iron carbonates also are found in the outer alteration aureoles of the deposits. Propylitic alteration is
located in the outermost portions of the deposit, and it occurs mainly in the biotite-bearing gneiss and in Jurassic rhyolitic and andesitic volcanic rocks.

The paragenesis for La Herradura gold deposit comprises seven different mineralogical associations; three of them are related to the gold mineralization (Table 2.2). Gold mineralization is mainly associated to quartz-galena-sphalerite and quartz-hematite veins (2nd and 3rd stages), however milky quartz veins are also carrying some gold (4th stage).
## Table 2.1. Tectonic and Deformation History from La Herradura Gold Deposit. Modified from de la Torre (2004) and Romero (2005).

<table>
<thead>
<tr>
<th>Relative Age</th>
<th>Geological Age</th>
<th>Tectonic environment</th>
<th>Tectonic Major Event</th>
<th>Tectonic Secondary event</th>
<th>Strain type</th>
<th>$\sigma_1$ Orientation</th>
<th>Sense of Displacement</th>
</tr>
</thead>
<tbody>
<tr>
<td>Quaternary</td>
<td>Volcanic fissural rocks related to San Andrés Fault System (El Pinacate Complex). Rocks outcropping near the La Herradura mine.</td>
<td>D5</td>
<td>NW 56.47 SE</td>
<td>Right lateral</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Postmineral</td>
<td>Middle Miocene 20-12 Ma</td>
<td>Brittle Depth&lt; 10 km</td>
<td>D4</td>
<td>Left-lateral Reverse-oblique</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>Laramide</td>
<td>Pulses 5, 3e</td>
<td>NE 25.18 SW</td>
<td>Right lateral</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>Laramide</td>
<td>D3</td>
<td>NE 44.35 SW</td>
<td>Left-lateral Left-lateral Reverse-oblique</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Intracontinental</td>
<td>Late Cretaceous to Lower Tertiary 80-45 Ma</td>
<td>Brittle-Ductile Depth&lt; 10 km</td>
<td>D3a</td>
<td>NE 44.41 SW Inverted section and thrust faults</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>Laramide</td>
<td>Pulses 2, 3, 4</td>
<td>NE 80 SW</td>
<td>Left-lateral Left-lateral Reverse-oblique</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>D3b</td>
<td>Step 1 Quartz-goethite; barren to weak gold-bearing</td>
<td>D3c</td>
<td>Step 2 Quartz-hematite-gold ± wulfenite ± galena. Main gold mineralization</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>D3d</td>
<td>Step 3 Brown color quartz, calcite, ± goethite; barren to weak gold-bearing</td>
<td>D3e</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>Intense horizontal compression and vertical extension</td>
<td>5 compressive pulses</td>
<td>Intense horizontal compression and vertical extension</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>Mohave Sonora Megashear</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Premineral</td>
<td>Lower Jurassic 208-187 Ma</td>
<td>Brittle-Ductile Depth&lt; 13 km</td>
<td>D2</td>
<td>NE 51 SW Reverse</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>D1a</td>
<td>Extension</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>D1b</td>
<td>Intense Compression</td>
<td>NE SW</td>
<td>Left-lateral Left-lateral Reverse-oblique</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>D1a</td>
<td>Greenschist to amphibolite metamorphism grade on sedimentary rocks (1750 Ma). Sediments being deposited</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

Metasediments and metavolcanics, magmatic arc affinity. Greenschist metamorphic facies

Carboniferous Calcareous rocks: Limestones and dolomites with intense shearing, no foliation, no metamorphism. Corals and Brachiopods from Carboniferous

Younger events have rotated their original positions
Table 2.2. Paragenetic sequence of ore and gangue minerals from La Herradura gold deposit (modified from Altamirano and Carreón, 1995).

<table>
<thead>
<tr>
<th>Event</th>
<th>Mineralogical association</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>Quartz-pyrite</td>
</tr>
<tr>
<td>2</td>
<td>*Quartz-galena-sphalerite</td>
</tr>
<tr>
<td>3</td>
<td>*Quartz-hematite</td>
</tr>
<tr>
<td>4</td>
<td>*Quartz (milky)</td>
</tr>
<tr>
<td>5</td>
<td>Quartz-siderite-ankerite</td>
</tr>
<tr>
<td>6</td>
<td>Siderite-ankerite</td>
</tr>
<tr>
<td>7</td>
<td>Calcite</td>
</tr>
</tbody>
</table>

The precipitation of gold from hydrothermal ore-forming fluids, in order to form the orebody, is produced by changes in the fluid chemistry that remove gold from solution by destabilizing aqueous Au-S complexes. This destabilization could be accomplished by cooling, oxidation, reduction, decreasing of pH or lowering S of the fluids. This modification of fluids chemistry can be achieved by any of the following: (1) large-scale pressure and temperature gradients along the fluid plumbing system; (2) reaction of the fluid with the wall rocks surrounding the conduit; (3) transient pressure fluctuations inducing phase immiscibility in the fluid; (4) fluid mixing; or (5) chemisorption (Kerrich et al., 2000).
CHAPTER 3: U-PB ZIRCON GEOCHRONOLOGY FROM LA HERRADURA GOLD DEPOSIT.

3.1 Zircon U-Pb Dating

3.1.1 Sampling

Six samples from outcrops of different rock units were collected for dating by U-Pb isotopes (Figure 3.1, Table 3.1). The main purpose of this study was to obtain the age of the Precambrian rocks that host gold mineralization in the La Herradura gold deposit and of Jurassic rocks exposed to the east of the orebody. These ages are also crucial for an understanding of the tectonic history of the area.

3.1.2 Preparation of Samples

Zircon crystals were separated from the rocks by traditional crushing, followed by magnetic, and by gravimetric separation techniques. Heavy mineral concentrates of the <350 micron fraction were separated magnetically. Inclusion-free zircons from the non-magnetic fraction were then handpicked under a binocular microscope. At least fifty zircons from each sample were mounted in epoxy and polished for laser ablation analyses. Before the isotopic analysis, the zircons were studied by Scanning Electron Microscopy (SEM) and cathodoluminiscence in order to determine the presence of inherited cores or any other complexity.

3.1.3 Analytical Procedures

Zircon crystals were analyzed in polished section with a Micromass Isoprobe multi-collector Inductively Coupled Plasma Mass Spectrometer equipped with nine Faraday collectors, an axial Daly detector, and four ion-counting channels (Gehrels et al., 2008). The Isoprobe is equipped with an ArF Excimer laser, which has an emission wavelength of 193 nm. Analyses were conducted on 25- to 35-micron spots with output energy of ~80 mJ (at 23kV) and a repetition rate of 8 Hz. Each analysis consisted of one 20-second integration on peaks with no laser firing and twelve 1-second integrations on peaks with the laser firing. The Hg contribution to the $^{204}$Pb
mass position is accordingly removed by subtracting the on-peak background values. The depth of each ablation pit was ~15 microns. The total measurement time was ~90 s per analysis.

The collectors were configured for simultaneous measurement of $^{204}$Pb in an ion-counting channel while $^{206}$Pb, $^{207}$Pb, $^{208}$Pb, $^{232}$Th, and $^{238}$U are measured with Faraday detectors. All analyses were conducted in static mode. Inter-element fractionation was monitored by analyzing fragments of a large concordant zircon crystal from Sri Lanka with a known (ID-TIMS) age of 564 ± 4 Ma (2σ) (Gehrels et al., 2006). This reference zircon was analyzed once for every four unknown samples.

The reported ages for zircon grains are based on $^{206}$Pb/$^{238}$U ratios because errors of the $^{207}$Pb/$^{235}$U and $^{206}$Pb/$^{207}$Pb ratios are significantly higher. This is due primarily to the low intensity (commonly <1 mV) of the $^{207}$Pb signal from these young, low-U grains. The $^{206}$Pb/$^{238}$U ratios are corrected for common Pb by using the measured $^{206}$Pb/$^{204}$Pb, common Pb composition from Stacey and Kramers (1975), and respective uncertainties of 1.0 and 0.3 for $^{206}$Pb/$^{204}$Pb and $^{207}$Pb/$^{204}$Pb.

For each sample, the $^{206}$Pb/$^{238}$U ages are plotted with two-sigma error bars that reflect only the error from determining $^{206}$Pb/$^{238}$U and $^{206}$Pb/$^{204}$Pb. The weighted mean of each sample was calculated using the Isoplot program (Ludwig, 2003). For the age of each sample, additional uncertainty from the calibration correction, decay constant, common lead composition and variation in measured $^{206}$Pb/$^{238}$U and $^{206}$Pb/$^{207}$Pb of the standard are considered. These systematic errors (1.0-1.7 %) were added quadratically to the measurement error (~1.1-1.2%). The reported ages are based primarily on $^{206}$Pb/$^{238}$U ratios for <1000 Ma grains and $^{206}$Pb/$^{207}$Pb for >1000 Ma grains. All reported ages and weighted mean ages have uncertainties at the two-sigma level.

3.1.4 Results

Sample LH-1
Fifty-six zircon grains were measured from the quartz-feldspar gneiss hosting the orebody, sample LH-1. The sample was collected to west of the Victoria shear zone. Zircons analyzed are clear reddish in color and range from 80 to 300 μm in size. They are doubly-terminated prisms dominated by the [100] face (Pupin, 1980), with a 2-3:1 length to width ratio, which is typical morphologies of zircons in igneous rocks. Measurements were made at the center and tip of the zircons. Zircons from this sample have U concentrations that vary from 350 to 180 ppm and U/Th ratios of ~2-3, characteristic of igneous zircons (Rubatto et al., 2002). These zircons yielded a weighted average $^{206}\text{Pb}/^{238}\text{U}$ age of 1737 ± 19.5 Ma (n=56, MSWD =0.22, Figure 3.2). In the fifty-six grains analyzed, an older component was not detected.

Sample LH-2

Twenty-five zircon grains were measured from volcanic rhyolite, sample LH-2. The sample was collected to east of the Victoria shear zone. Zircons analyzed are clear, transparent to pinkish in color, and range from 120 to 300 μm in size. They are doubly-terminated prisms dominated by the [100] face (Pupin, 1980), with a 3-5:1 length to width ratio, which is typical morphologies of zircons in igneous rocks. Measurements were made at the center and tip of the zircons. Zircons from this sample have U concentrations that vary from ~600~400 ppm and U/Th ratios of ~1-2, characteristic of igneous zircons (Rubatto et al., 2002). These zircons yielded a weighted average $^{206}\text{Pb}/^{238}\text{U}$ age of 167.8 ± 3.5 Ma (n=22, MSWD =0.2; Figure 3.2). In the twenty-five grains analyzed, Mesozoic (186, 214 Ma), Paleozoic (372 Ma) and mid-Proterozoic (1.4 Ga) components were detected.
Sample LH-3

Twenty-eight zircon grains were measured from tonalite, sample LH-3. This rock outcrops northwest of Centauro pit. Zircons analyzed are reddish in color, which is the product of the radiation damage, and range from 90 to 250 μm in size. They are prisms dominated by the [100] face (Pupin, 1980), with a 2-4:1 length to width ratio, which is typical morphologies of zircons in igneous rocks. Measurements were made at the center and tip of the zircons. Zircons from this sample have U concentrations that vary from ~2800 to ~450 ppm and U/Th ratios of ~2-3, characteristic of igneous zircons (Rubatto et al., 2002). These zircons yielded a concordant age of 1651.9 ± 15.3 Ma (n=18; Figure 3.2). In the twenty-eight grains analyzed, an older component that is ~1716-1792 Ma in age was detected (n=8 cores).

Sample LH-4

Twenty-nine zircon grains were measured from granite, sample LH-4. This rock outcrops northwest of Centauro pit. Zircons analyzed are reddish in color, which is the product of the radiation damage, and range from 90 to 250 μm in size. They are prisms dominated by the [100] face (Pupin, 1980), with a 2-4:1 length to width ratio, which is typical morphologies of zircons in igneous rocks. Measurements were made at the center and tip of the zircons. Zircons from this sample have U concentrations that vary from ~3000 to ~800 ppm and U/Th ratios of ~2-4, characteristic of igneous zircons (Rubatto et al., 2002). These zircons yielded a concordant age of 1756.3 ± 33.4 Ma (n=18; Figure 3.2). In the twenty-nine grains analyzed, a ~1786-1855 Ma older component was detected (n=10 cores).
Sample LH-5

Twenty-two zircon grains were measured from quartz-feldspar gneiss, sample LH-5. This sample was collected from outcrops in Yaqui pit. Zircons analyzed are reddish in color, which is the product of the radiation damage, and range from 90 to 300 μm in size. They are prisms dominated by the [100] face (Pupin, 1980), with a 2-4:1 length to width ratio, which is typical morphologies of zircons in igneous rocks. Measurements were made at the center and tip of the zircons. Zircons from this sample have U concentrations that vary from ~850 to ~150 ppm and U/Th ratios of ~1 to 3, characteristic of igneous zircons (Rubatto et al., 2002). These zircons yielded a weighted average 206Pb/207Pb age of 1735 ± 21.8 Ma (n=22, MSWD =0.31, Figure 3.2). In the twenty-two grains analyzed, no older component was detected.

Sample LH6

Forty-two zircon grains were measured from metaandesite, sample LH-6. This rock outcrops east of the Victoria shear zone. Zircons analyzed are clear, transparent to pinkish in color, and range from 120 to 300 μm in size. They are doubly-terminated prisms dominated by the [100] face (Pupin, 1980), with a 3-5:1 length to width ratio, which is typical morphologies of zircons in igneous rocks. Measurements were made at the center and tip of the zircons. Zircons from this sample have U concentrations that vary from ~3000-~200 ppm and U/Th ratios of ~0.5 to ~3, characteristic of igneous zircons (Rubatto et al., 2002). This sample yielded a 206Pb/238U age, calculated using the TuffZirc Algorithm (Ludwig, K.J., 2003), of 159.0 +3.5 -3.7 Ma (n=23, Figure 3.2). In the forty-two grains analyzed, a ~1608-1785 Ma older component was detected (n=10 cores).
Figure 3.1. Map showing location of samples for U-Pb dating. Coordinate system is UTM, NAD27.

<table>
<thead>
<tr>
<th>Sample</th>
<th>east</th>
<th>north</th>
<th>masl</th>
<th>Rock type</th>
<th>age (Ma)</th>
</tr>
</thead>
<tbody>
<tr>
<td>LH-1</td>
<td>321,830</td>
<td>3,448,829</td>
<td>200</td>
<td>Quartz-Feldspar gneiss</td>
<td>1737 ± 19.5</td>
</tr>
<tr>
<td>LH-2</td>
<td>321,826</td>
<td>3,448,936</td>
<td>226</td>
<td>Metarhyolite</td>
<td>167.8±3.5</td>
</tr>
<tr>
<td>LH-3</td>
<td>320,493</td>
<td>3,448,943</td>
<td>231</td>
<td>Tonalite</td>
<td>1651.9 ± 15.3</td>
</tr>
<tr>
<td>LH-4</td>
<td>320,430</td>
<td>3,448,835</td>
<td>204</td>
<td>Granite</td>
<td>1756.3 ± 33.4</td>
</tr>
<tr>
<td>LH-5</td>
<td>322,661</td>
<td>3,447,941</td>
<td>98</td>
<td>Quartz-Feldspar gneiss</td>
<td>1735 ± 21.8</td>
</tr>
<tr>
<td>LH-6</td>
<td>322,302</td>
<td>3,448,844</td>
<td>230</td>
<td>Metaandesite</td>
<td>159 ±3.5, -3.7</td>
</tr>
</tbody>
</table>

Table 3.1. U-Pb ages from La Herradura Gold Deposit. Coordinates are UTM NAD27.
Figure 3.2. U-Pb ages from zircons.
3.2 PYRITE-MAGNETITE RE-OS DATING

3.2.1 Sampling

Five samples of quartz-pyrite and quartz-pyrite-magnetite veins from diamond drill cores were collected for dating by Re-Os isotopic analysis (Fig. 3.3). The main purpose of this study is to constrain the age of mineralization in the La Herradura gold deposit.

3.2.2 Analytical methods

The analytical procedure is the same as reported in Barra et al. (2003). Rock samples containing pyrite and magnetite were crushed, and mineral phases of interest were handpicked. About 0.8 g of sample was loaded in a Carius tube and dissolved in a 3:1 mixture of HNO₃ and HCl following the procedures described by Shirey and Walker (1995). Hydrogen peroxide was added to ensure complete oxidation of the sample and spike equilibration (Frei et al., 1998). The carius tube was heated to ~220 °C overnight and the solution was later treated in a two-stage distillation process for Os (Nägler and Frei, 1997). Re was extracted and purified through column chemistry (Shirey and Walker, 1995). Os was loaded on platinum filaments, whereas Re was loaded on nickel filaments and later analyzed by negative thermal ion mass spectrometry (NTIMS) (Creaser et al., 1991).

Reported uncertainties listed in Table 1 for both the $^{187}\text{Re}/^{188}\text{Os}$ and $^{187}\text{Os}/^{188}\text{Os}$ ratios were calculated using error propagation. The isochron age was calculated using Isoplot (Ludwig, 2001).
3.2.3 Results

The results for the analysis of three different pyrite samples and a magnetite sample (replicated) are shown in Table 3.2. The samples yielded a five-point isochron (Fig. 3.4) with an age of 61.0 ± 2.1 Ma and an initial $^{187}\text{Os}/^{188}\text{Os}$ ratio of 0.358 ± 0.009 (MSWD = 0.45). The concentrations of Re and Os range between 0.5-3.0 ppb and 12-41 ppt, respectively.
Table 3.2. Results of analysis from Re-Os isotopes.

<table>
<thead>
<tr>
<th>Sample</th>
<th>Mineral phase</th>
<th>$^{187}$Os/$^{188}$Os</th>
<th>Error</th>
<th>$^{187}$Re/$^{188}$Os</th>
<th>Error</th>
<th>Os (ppb)</th>
<th>Re (ppb)</th>
</tr>
</thead>
<tbody>
<tr>
<td>14-2</td>
<td>pyrite</td>
<td>0.7558</td>
<td>0.004</td>
<td>390.30</td>
<td>6.33</td>
<td>0.041</td>
<td>3.042</td>
</tr>
<tr>
<td>17-5</td>
<td>pyrite</td>
<td>0.7374</td>
<td>0.010</td>
<td>372.51</td>
<td>6.19</td>
<td>0.014</td>
<td>0.956</td>
</tr>
<tr>
<td>17-6</td>
<td>pyrite</td>
<td>0.6107</td>
<td>0.006</td>
<td>251.02</td>
<td>4.07</td>
<td>0.012</td>
<td>0.603</td>
</tr>
<tr>
<td>18-1-1</td>
<td>magnetite</td>
<td>0.5241</td>
<td>0.006</td>
<td>160.86</td>
<td>2.66</td>
<td>0.015</td>
<td>0.468</td>
</tr>
<tr>
<td>18-1-2</td>
<td>magnetite</td>
<td>0.5369</td>
<td>0.005</td>
<td>177.25</td>
<td>2.97</td>
<td>0.017</td>
<td>0.575</td>
</tr>
</tbody>
</table>

Figure 3.4. Re-Os isochron for La Herradura gold deposit.
CHAPTER 4: STABLE ISOTOPES

4.1 Introduction

Because water is the dominant constituent of ore-forming fluids, knowledge of the waters origin is fundamental to any theory of ore formation (Taylor, 1974). Natural waters of various origins exhibit systematic differences in their hydrogen and oxygen isotopic contents that can be deciphered by measuring stable isotopes ratios. Stable isotopic data can place a number of constraints on the source of the vein-forming fluids and associated gold mineralization and the conditions of vein formation (Ohmoto and Rye, 1979; O’Neil, 1986; Kerrich, 1987; Ohmoto and Goldhaber, 1997; Jia and Kerrich, 1999; Jia et al., 2000, 2001). For this reason, in the last four decades there have been a considerable number of studies that investigated the practical use of stable isotopes to explore for a range of hydrothermal mineral deposits. The studies have concentrated on the existence and extent of oxygen and hydrogen isotopic haloes within and beyond the limits of visible wallrock alteration, and the use of stable isotopes as vectors to ore zones by sampling limited exposures such as exploration drillholes (Bierlein et al., 2004).

Stable isotope studies in orogenic gold deposits (OGD) have been carried out extensively in the last twenty-five years for those deposits. This type of deposit is located mainly in the most gold-productive Precambrian shields in Australia and Canada, particularly the Late Archean greenstone belts and Paleoproterozoic fold belts, and in the late Neoproterozoic and younger Cordilleran-style orogens mainly along the margins of Gondwana, Laurentia and the more recent circum-Pacific (Cox et al., 1995; Goldfarb et al., 1988, 1991; Jia et al., 2001, 2003; Haeberlin, 2002; Bierlein et al., 2004; Shelton et al., 2004; Salier et al., 2005).
The purpose of this study is to obtain oxygen, hydrogen, and sulfur isotopic data from gold-bearing quartz-sulfide veins and their sericite alteration haloes in order to constrain the origin of the fluids. Coupled with geological observations, these measurements will lead to a better understanding of the orebody and can be compared with previous data published from known orogenic gold deposits from North America and South America.

4.2 Oxygen and Hydrogen Isotopes

4.2.1 Sampling

For oxygen isotopes, 16 samples of quartz and quartz-carbonate veins were collected from outcrops and diamond drill cores in and around the Centauro pit area (Figure 4-1). The sample set includes quartz veins from all previously distinguished events according to the paragenetic sequence of ore and gangue minerals described for La Herradura deposit (de la Torre, 2004). The set of samples is also trying to cover all types of quartz veins, according their differences in color, shape, thickness, mineralogical association, attitude, crosscutting relationships, and depth in the deposit (Table 4.1, Figure 2.2). All of these quartz vein samples are hosted in quartz-feldspar gneiss, and most of them have sericitic haloes (Figure 4.2). Some of the veins come from deep holes (e.g., 650 m) below and around the Centauro pit. These veins carry high values of gold that correlate closely with the presence of sulfide mineralization like pyrite, galena and sphalerite (Figure 4.2). Individual pieces of quartz veins were crushed gently and, from each sample a small fragment measuring between 100 to 200 microns size was mounted in epoxy along with oxygen isotope standards and polished to expose the sample interiors.
For hydrogen isotopes, 15 samples were collected from outcrops and cores from diamond drill cores in and around the Centauro Pit (Figure 4.1). Sericite was hand-picked under a microscope from intergrown vein quartz or from the innermost part of the sericitic alteration halo and was analyzed for its hydrogen isotope composition. The sericite is clearly hydrothermal in origin and is not remnant wall-rock phase; although in the samples coming from alteration haloes some isotopic heritage from the wall rock can be expected.

Figure 4.1. Map of La Herradura mine showing locations of samples used for oxygen and hydrogen isotopes. Coordinate system is UTM, NAD27.
Table 4.1. Description of samples for oxygen and hydrogen isotopes from La Herradura Gold Deposit.

<table>
<thead>
<tr>
<th>Sample</th>
<th>East</th>
<th>North</th>
<th>MASL</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>478-2</td>
<td>322,186</td>
<td>3,447,835</td>
<td>127</td>
<td>Quartz-pyrite-siderite-ankerite vein. 3 mm thickness. Host rock is Quartz Feldspar Gneiss.</td>
</tr>
<tr>
<td>478-4</td>
<td>322,186</td>
<td>3,447,835</td>
<td>127</td>
<td>Quartz vein 5 cm thickness, microbreccia, sericite haloes.</td>
</tr>
<tr>
<td>478-5</td>
<td>322,186</td>
<td>3,447,835</td>
<td>127</td>
<td>Quartz-sericite vein 3 cm thickness, siderite-ankerite at borders.</td>
</tr>
<tr>
<td>525-1</td>
<td>322,321</td>
<td>3,447,165</td>
<td>-65</td>
<td>Quartz-carbonate extensional vein 10 cm thickness, carbonates are in the borders. Host rock is Quartz Feldspar Gneiss. Sericite-pyrite alteration.</td>
</tr>
<tr>
<td>530-1</td>
<td>322,259</td>
<td>3,447,089</td>
<td>-58</td>
<td>Quartz-pyrite (now hematite) veins 2 cm thickness, Sericitic alteration halos.</td>
</tr>
<tr>
<td>530-2</td>
<td>322,259</td>
<td>3,447,089</td>
<td>-58</td>
<td>Quartz-pyrite (now hematite) veins 2 cm thickness, Sericitic alteration halos.</td>
</tr>
<tr>
<td>530-3</td>
<td>322,259</td>
<td>3,447,089</td>
<td>-58</td>
<td>Quartz crystals euhedral habit with hematite (on pyrite case) disseminated and on fractures. Sericitic alteration halos.</td>
</tr>
<tr>
<td>531</td>
<td>322,269</td>
<td>3,447,104</td>
<td>-60</td>
<td>Quartz-pyrite vein 1 cm thickness. Sericitic alteration halo</td>
</tr>
<tr>
<td>532-2</td>
<td>322,258</td>
<td>3,447,092</td>
<td>-35</td>
<td>Quartz-pyrite (now hematite) vein, 2 cm thickness, white color, low-dipping. This zone corresponds to high-grade mineral (2 g/t gold). Sample comes from a quartz crystal.</td>
</tr>
<tr>
<td>HGT-14, 487.30m</td>
<td>322,734</td>
<td>3,447,251</td>
<td>-317</td>
<td>Light gray quartz vein, 20 cm width on Quartz Feldspar Biotite Gneiss. Qz has disseminated pyrite clusters &lt;1 cm. Sericite alteration halo</td>
</tr>
<tr>
<td>HGT-14, 488.70m</td>
<td>322,734</td>
<td>3,447,251</td>
<td>-318</td>
<td>Quartz-pyrite-sphalerite vein, 8 mm thickness, sericitic alteration halo</td>
</tr>
<tr>
<td>HGT-14, 541.40m</td>
<td>322,724</td>
<td>3,447,243</td>
<td>-369</td>
<td>Quartz-pyrite-sericite 13 mm thickness. Host rock is Quartz Feldspar Gneiss. Sericite from the core of the vein.</td>
</tr>
<tr>
<td>HGT-14, 541.40m</td>
<td>322,724</td>
<td>3,447,243</td>
<td>-369</td>
<td>Quartz-pyrite-sericite 13 mm thickness. Host rock is Quartz Feldspar Gneiss. Sericite from alteration halo</td>
</tr>
<tr>
<td>HGT-14, 570.10m</td>
<td>322,718</td>
<td>3,447,238</td>
<td>-397</td>
<td>Quartz-pyrite-sphalerite vein, irregular form, 2 cm thickness, sericitic alteration halo.</td>
</tr>
<tr>
<td>HGT-14, 570.35m</td>
<td>322,718</td>
<td>3,447,238</td>
<td>-397</td>
<td>Quartz-pyrite-galen vein, 4 mm thickness. Sericite halo. Host rock is Quartz Feldspar gneiss</td>
</tr>
<tr>
<td>HGT-16 344.90m</td>
<td>321,576</td>
<td>3,447,823</td>
<td>-120</td>
<td>Light gray quartz veinlet, cut by siderite-ankerite. Drusic textures.</td>
</tr>
<tr>
<td>HGT-17 651.90m</td>
<td>322,481</td>
<td>3,447,297</td>
<td>-678</td>
<td>Quartz-Pyrite (5 mm) vein 10 mm thickness. Host rock is Quartz Feldspar Gneiss</td>
</tr>
<tr>
<td>LH-AE</td>
<td>322,263</td>
<td>3,447,325</td>
<td>118</td>
<td>Quartz milky color, 50 cm thickness.</td>
</tr>
</tbody>
</table>
Figure 4.2. Representative samples used for stable isotope studies. A. Sample 528. Quartz-iron carbonate-pyrite (now hematite) vein, sericite haloes with disseminated hematite after pyrite. B. Sample 530-2. Quartz-hematite after pyrite with sericite haloes. C. Sample LH-AE. Milky quartz, strong fracturing, collected from a quartz shear vein being the feeders for gold mineralization. D. HGT-14 541.40. Quartz-sericite vein with sericite haloes. This sample is coming from deep drill holes beneath the actual mining exploitation in Centauro pit. E. Sample HGT-13 643.30m. Quartz-pyrite-sphalerite-galena vein with sericitic halo. This sample is from deeper diamond drill holes in the Centauro pit. A 2-m width geochemical sample containing this vein has 6 g/t gold, indicating that there is a spatial relationship between gold and sulfides. F. Sample HGT-17 655.85m. Quartz-pyrite-sphalerite-galena vein from deeper diamond drill holes in the Centauro pit. Geochemical assay from a 2-m interval sample containing this vein assays 1.8 g/t gold. Ak: ankerite, Hm: hematite, Qz: quartz, Se: sericite, Sf: sulfide.
4.2.2 Analytical Methods

The oxygen isotopic compositions of quartz pieces were measured in situ within the quartz polished sections by using the UCLA CAMECA ims 1270 ion microprobe. Samples were coated with ~ 350 Å of gold. The ~0.1–0.4 nA primary Cs\(^+\) beam was shaped into a spot ~ 8–10 \(\mu\)m in diameter (usually in aperture illumination mode). Charge compensation was achieved using a normal-incidence electron flood gun (Slodzian, 1980). Low-energy (~0–30 eV) negative secondary ions were measured at high-mass resolving power (M/\(\Delta M\)) of about 6500. Measurements were made by magnetic peak switching through 25 cycles of counting \(^{16}\)O\(^-\) for 3s, and \(^{18}\)O\(^-\) for 5s. The secondary \(^{16}\)O\(^-\) current was measured in a Faraday cup equipped with a Keithley 642 electrometer, and \(^{18}\)O\(^-\) signals were pulse-counted on an electron multiplier with a deadtime of typical 20 ns. The analysis conditions typically yielded equivalent count rates of 20–40 million counts per second of \(^{16}\)O\(^-\). Measured secondary ion intensities were corrected for background (\(^{16}\)O\(^-\)), and for deadtime (\(^{18}\)O\(^-\)). With the count rates employed in these experiments, the total magnitude of these corrections is small (<2‰), and thus the additional uncertainties resulting from background and deadtime corrections are negligible (<0.1‰). Zircon was used as a standard (\(\delta^{18}\)O = 18.4 ‰) to correct for instrumental mass fractionation. Isotopic data are reported in standard \(\delta\) notation relative to Vienna SMOW.

The \(\delta\)D values for sericite were measured on a continuous-flow gas-ratio mass spectrometer (ThermoQuest Finnigan Delta PlusXL) in the Environment Isotope Laboratory at The University of Arizona. Samples were combusted at 1400°C using...
the Thermal Combustion Elemental Analyzer (TCEA) coupled to the mass spectrometer. Standardization is based on NIST SRM 8540. Precision ($1\sigma$) is better than ± 2.5 per mil on the basis of repeated internal standards. Isotopic data are reported in standard $\delta$D notation relative to Vienna SMOW.

4.2.3 Analytical Results

The range of $\delta^{18}$O values for vein quartz samples collected and analyzed during this study is 14.1 to 16.1 per mil, except for one sample whose value is 25.2 per mil. This erratic value is interpreted to come from silica formed by weathering of aluminosilicate minerals, or the sample was contaminated during its preparation. $\delta^{18}$O quartz values are very uniform in most of the samples (Table 4.2). This uniformity implies a very uniform temperature distribution and source of ore fluids, throughout the history of mineralization, regardless of their color, shape, thickness, attitude, mineralogical association, crosscutting relationships, and depth in the deposit. The quartz was in equilibrium with the ore fluids of between 8.2 and 10.2 per mil, using the quartz-water fractionation equation from Clayton et al. (1972) and assuming vein formation at 330°C, the temperature obtained from sulfur isotopes, although the actual temperature for ore deposition might be slightly higher or lower.

For sericite, the measured $\delta$D range is -109 to -57 per mil, but most values are between -76 and -61 per mil (Table 4.2). Extrapolation of water-muscovite fractionations factors (Suzuoki and Epstein, 1976) indicates about 41‰ fractionation at 330°C. This fractionation yields $\delta$D values for ore fluids between -65 and -15 per mil, with most values falling in a narrower interval of -34 to -19 per mil.
Table 4.2. Oxygen and Hydrogen Isotope data from La Herradura Gold Deposit.

<table>
<thead>
<tr>
<th>Sample</th>
<th>$\delta^{18}$O Quartz (‰)</th>
<th>$\delta$D Sericite (‰)</th>
<th>$\delta^{18}$O Ore fluids (‰)</th>
<th>$\delta$D Ore fluids (‰)</th>
</tr>
</thead>
<tbody>
<tr>
<td>478-2</td>
<td>15.9</td>
<td>1.0</td>
<td></td>
<td></td>
</tr>
<tr>
<td>478-4</td>
<td>15.7</td>
<td>-109</td>
<td>9.8</td>
<td>-67</td>
</tr>
<tr>
<td>478-5</td>
<td>25.2</td>
<td></td>
<td>19.3</td>
<td></td>
</tr>
<tr>
<td>525-1</td>
<td>15.3</td>
<td>-65</td>
<td>9.4</td>
<td>-23</td>
</tr>
<tr>
<td>525-2</td>
<td>14.2</td>
<td>-57</td>
<td>8.3</td>
<td>-15</td>
</tr>
<tr>
<td>528-1</td>
<td>14.1</td>
<td>-73</td>
<td>8.2</td>
<td>-31</td>
</tr>
<tr>
<td>528-3</td>
<td>14.5</td>
<td></td>
<td>8.6</td>
<td></td>
</tr>
<tr>
<td>530-1</td>
<td>14.4</td>
<td></td>
<td>8.5</td>
<td></td>
</tr>
<tr>
<td>530-2</td>
<td>14.6</td>
<td></td>
<td>8.7</td>
<td></td>
</tr>
<tr>
<td>530-3</td>
<td>15.4</td>
<td></td>
<td>9.5</td>
<td></td>
</tr>
<tr>
<td>531</td>
<td></td>
<td>-62</td>
<td></td>
<td>-20</td>
</tr>
<tr>
<td>532-2</td>
<td>14.9</td>
<td>-63</td>
<td>9</td>
<td>-21</td>
</tr>
<tr>
<td>533</td>
<td></td>
<td>-58</td>
<td></td>
<td>-16</td>
</tr>
<tr>
<td>HGT-14, 487.30m</td>
<td>14.3</td>
<td>-76</td>
<td>8.4</td>
<td>-34</td>
</tr>
<tr>
<td>HGT-14, 488.70m</td>
<td>16.1</td>
<td>-75</td>
<td>10.2</td>
<td>-33</td>
</tr>
<tr>
<td>HGT-14, 541.40m</td>
<td>-100</td>
<td></td>
<td></td>
<td>-58</td>
</tr>
<tr>
<td>HGT-14, 541.40m</td>
<td>-61</td>
<td></td>
<td></td>
<td>-19</td>
</tr>
<tr>
<td>HGT-14, 570.10m</td>
<td>-70</td>
<td></td>
<td></td>
<td>-28</td>
</tr>
<tr>
<td>HGT-14, 570.35m</td>
<td>-71</td>
<td></td>
<td></td>
<td>-29</td>
</tr>
<tr>
<td>HGT-16, 344.90m</td>
<td>15.5</td>
<td>-71</td>
<td>9.6</td>
<td>-29</td>
</tr>
<tr>
<td>HGT-17, 651.90m</td>
<td>14.5</td>
<td>-64</td>
<td>8.6</td>
<td>-22</td>
</tr>
<tr>
<td>LHAE</td>
<td>15.6</td>
<td></td>
<td>9.7</td>
<td></td>
</tr>
</tbody>
</table>

1. The $\delta^{18}$O values for ore-forming fluids were calculated from the quartz-water equilibrium equation of Clayton et al. (1972). 2. The $\delta$D for ore-forming fluids were calculated from the muscovite-water equilibrium fractionation factor of Suzuoki and Epstein (1976). Temperature of deposition was obtained from fluid inclusions and sulfur isotopes, $T = 330^\circ$C.

4.3 Sulfur Isotopes

4.3.1 Sampling

Twenty-six sulfur isotope measurements on pyrite, sphalerite, and galena were obtained from 12 samples of quartz-sulfide veins, collected from diamond drill cores deep below the Centauro pit (Figure 4.3). All quartz-sulfide veins have sericitic
alteration haloes and are hosted in quartz-feldspar gneiss. Individual grains of pyrite, galena, and sphalerite were hand-separated. The first suite of samples consisted of ten sulfide species that were used for determining the source of the sulfur in the deposit. Later, a second suite of samples was sent to the laboratory in order to use mineral pairs for the estimation of the temperature of deposition of the sulfides and gold mineralization (Figure 4.4) and compare those with the temperatures obtained from fluid inclusions.

Figure 4.3 Map of the La Herradura mine, showing location of samples used for Sulfur Isotopes. Coordinate system is UTM, NAD27.
49

Figure 4.4. Reflected light microphotograph, showing pyrite with strong fracturing and later gold mineralization in the fractures. The sample corresponds to DDH HGT-13, 367m. The Au assay in this area is 40 g/t gold.

4.3.2 Analytical Methods

Values of $\delta^{34}$S of different sulfides were measured on a continuous flow gas-ratio mass spectrometer (Finnigan Delta PlusXL). Samples were combusted at 1030 °C with V$_2$O$_5$ using an elemental analyzer (Costech) coupled to the mass spectrometer. Standardization is based on international standards OGS-1 and NBS123 and several other sulfide minerals that have been compared between laboratories. Calibration is linear in the range -10 to +30 per mil. Precision is estimated to be ± 0.15 ‰ or better (1σ), based on repeated internal standards. Isotopic data are reported in standard $\delta^{34}$S notation relative to Canyon Diablo Troilite (CDT).
4.3.3 Analytical Results

Sulfur isotopic data for 26 sulfide samples are summarized in Table 4.3. The $\delta^{34}S$ values of sulfides fall in a narrow range between 3.1 and 6.6 per mil. Individual species show a narrower range of $\delta^{34}S$, pyrite from 4.7 to 6.7 per mil, sphalerite from 4.8 to 5.6 per mil, and galena from 3.4 to 3.9 per mil. There is no systematic variation of isotope values with depth (Fig.4.5, Table 4.3). The $\delta^{34}S$ of pyrite $\geq$ $\delta^{34}S$ of sphalerite $> \delta^{34}S$ of galena, consistent with the order of equilibrium fractionation among these minerals (Bachinski, 1969).

4.3.4 Sulfur Isotope Geothermometry

The isotopic compositions obtained from coexisting sulfide species were used as a geothermometer, assuming that they have achieved isotopic equilibrium. Ohmoto and Lasaga (1982) demonstrated that at temperatures above 300°C, sulfides species can achieve isotopic equilibrium. The pyrite-galena pair, as the most sensitive geothermometer (Ohmoto and Rye, 1979), was chosen to estimate the temperature of formation of quartz-sulfide veins and deposition of related gold mineralization. Temperatures were calculated using equations in Campbell and Larson (1998). The calculated temperature for deposition of sulfides in samples HGT-13-401.20 and HGT-17-655.85 range between 319 to 343°C, with an average temperature of 330 $\pm$20°C (Table 4.3). For sample HGT-13-755.45, however, some disequilibrium is expected because the calculated average temperature is 407°C. Also, by using the pyrite-sphalerite mineral pair, the calculated temperature for sample HGT-13-643.30 the average temperature is 343°C.
Pyrite-galena values give the most precise temperature estimates. The temperatures obtained from samples HGT-13-401.20 and HGT-17-655.85 (328, 330, 342 and 319°C), whose average is 330 ±20°C, and is in agreement with temperatures obtained from fluid inclusions (275 to 285°C) in the same type of quartz veins.
Table 4.3. S isotope database and calculated temperatures for deposition of sulfides. Temperature deposition calculated from Campbell and Larson (1998).

<table>
<thead>
<tr>
<th>Sample</th>
<th>Vein morphology</th>
<th>mineral</th>
<th>$\delta^{34}\text{S} (\text{‰})$</th>
<th>$T , (\text{°C})$</th>
</tr>
</thead>
<tbody>
<tr>
<td>HGT-14,487.30m</td>
<td>vein</td>
<td>py</td>
<td>5.3</td>
<td></td>
</tr>
<tr>
<td>HGT-14,488.70m</td>
<td>vein</td>
<td>py</td>
<td>5.6</td>
<td></td>
</tr>
<tr>
<td>HGT-14,488.70m</td>
<td>vein</td>
<td>sp</td>
<td>5.2</td>
<td></td>
</tr>
<tr>
<td>HGT-14,514.40m</td>
<td>vein</td>
<td>py</td>
<td>4.7</td>
<td></td>
</tr>
<tr>
<td>HGT-14,570.10m</td>
<td>vein</td>
<td>sp</td>
<td>4.8</td>
<td></td>
</tr>
<tr>
<td>HGT-17,651.90m</td>
<td>vein</td>
<td>py</td>
<td>6.2</td>
<td></td>
</tr>
<tr>
<td>HGT-17,655.15m</td>
<td>vein</td>
<td>py</td>
<td>5.7</td>
<td></td>
</tr>
<tr>
<td>HGT-17,655.15m</td>
<td>vein</td>
<td>py</td>
<td>3.4</td>
<td></td>
</tr>
<tr>
<td>HGT-17,655.85m</td>
<td>vein</td>
<td>sp</td>
<td>5.3</td>
<td></td>
</tr>
<tr>
<td>HGT-13,367.20m</td>
<td>brecciated vein</td>
<td>Gn</td>
<td>3.6</td>
<td></td>
</tr>
<tr>
<td>HGT-13,401.20m</td>
<td>brecciated vein</td>
<td>Py</td>
<td>6.6</td>
<td>328</td>
</tr>
<tr>
<td>HGT-13,401.20m</td>
<td>brecciated vein</td>
<td>Gn</td>
<td>3.7</td>
<td>330</td>
</tr>
<tr>
<td>HGT-13,401.20m</td>
<td>brecciated vein</td>
<td>Py</td>
<td>6.6</td>
<td>330</td>
</tr>
<tr>
<td>HGT-13,643.30m</td>
<td>vein</td>
<td>Py</td>
<td>6.4</td>
<td>354</td>
</tr>
<tr>
<td>HGT-13,643.30m</td>
<td>vein</td>
<td>Sp</td>
<td>5.6</td>
<td></td>
</tr>
<tr>
<td>HGT-13,757.45m</td>
<td>vein</td>
<td>Py</td>
<td>6.1</td>
<td>417</td>
</tr>
<tr>
<td>HGT-13,757.45m</td>
<td>vein</td>
<td>Gn</td>
<td>3.9</td>
<td></td>
</tr>
<tr>
<td>HGT-17,655.85m</td>
<td>vein</td>
<td>Py</td>
<td>6.2</td>
<td>398</td>
</tr>
<tr>
<td>HGT-17,655.85m</td>
<td>vein</td>
<td>Gn</td>
<td>3.1</td>
<td></td>
</tr>
<tr>
<td>HGT-13,655.85m</td>
<td>vein</td>
<td>Py</td>
<td>6.0</td>
<td>319</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Gn</td>
<td>3.1</td>
<td></td>
</tr>
</tbody>
</table>
Figure 4.5. Plot of $\delta^{34}$S Vs elevation. Values for individual species are clustering. Arrow shows decreasing isotopic values from pyrite to sphalerite to galena. It implies that sulfide species may have achieved isotopic equilibrium.

4.4 Carbon Isotopes

4.4.1 Sampling

Isotopic and carbon content measurements were carried out on ankerite crystals and on quartz veins, which contains CO$_2$-bearing fluid inclusions. Three ankerite samples were collected from outcrops in the Centauro pit and from the outer part of the deposit (Figure 4.6). Quartz veins were collected from outcrops within the Centauro pit and from diamond drill cores deep below the Centauro pit. All quartz-sulfide veins have sericitic alteration haloes and are hosted in quartz-feldspar gneiss. 100 mg of each sample was hand-separated and then finely crushed by using a mortar and pestle.
4.4.2. Analytical Methods

$\delta^{13}$C and carbon content were measured on a continuous continuous-flow gas-ratio mass spectrometer (Finnigan Delta PlusXL) coupled to an elemental analyzer Costech). Samples were combusted in the elemental analyzer. Standardization is based on acetonilide for elemental concentration, NBS-22 and USGS-24 for $\delta^{13}$C. Precision is better than $\pm 0.09 \%o$, based on repeated internal standards.
4.4.3 Analytical Results

Carbon content and carbon isotopic data for all samples are summarized in Table 4.4 and presented as a plot of $\delta^{13}C$ versus 1/concentration of C in figure 4.7. Carbon contents are based accurately on chromatogram peak areas, except for sample 14-488, whose carbon content is approximated from peak heights. The end members are fluid inclusion-bearing quartz and ankerite, which have approximately fixed compositions. Any mixing trend will appear as a straight line connecting the two end members. Some quartz yielded very little C concentrations, suggesting a strong influence of CO$_2$ from fluid inclusions. Some quartz yielded a relatively large amount of C (sample 17-651). Concentrations and isotopes are consistent with the presence of some carbonate mineral. One sample is intermediate, lying near the straight line between the two end members. These data suggest that there are two sources of C, one associated with carbonate minerals, $\delta^{13}C$=0 to -6‰, and one associated with fluid inclusions, $\delta^{13}C$ < -18‰. The low $\delta^{13}C$ carbon source is organic C.
Table 4.4. Carbon content and $\delta^{13}$C values of fluid inclusion-bearing quartz and ankerite.

<table>
<thead>
<tr>
<th>Sample</th>
<th>%C</th>
<th>% C corr</th>
<th>1 / %C</th>
<th>$\delta^{13}$C ‰</th>
</tr>
</thead>
<tbody>
<tr>
<td>528-FI</td>
<td>0.06</td>
<td>0.06</td>
<td>16.26</td>
<td>-8.1</td>
</tr>
<tr>
<td>530-FI</td>
<td>0.02</td>
<td>0.02</td>
<td>64.02</td>
<td>-17.8</td>
</tr>
<tr>
<td>13-643-FI</td>
<td>0.02</td>
<td>0.02</td>
<td>53.91</td>
<td>-18.4</td>
</tr>
<tr>
<td>17-651-FI</td>
<td>9.60</td>
<td>9.37</td>
<td>0.11</td>
<td>-5.7</td>
</tr>
<tr>
<td>14-488-FI</td>
<td></td>
<td>1.7</td>
<td>0.59</td>
<td>-6.4</td>
</tr>
<tr>
<td>488-A</td>
<td>12.0</td>
<td>0.08</td>
<td>0</td>
<td>-0.9</td>
</tr>
<tr>
<td>523-A</td>
<td>12.0</td>
<td>0.08</td>
<td>0</td>
<td>-1.6</td>
</tr>
<tr>
<td>528-A</td>
<td>12.0</td>
<td>0.08</td>
<td>0</td>
<td>-2.8</td>
</tr>
</tbody>
</table>

Figure 4.7. Plot of $\delta^{13}$C versus 1/ concentration of C.
4.5. Interpretation and Discussion

Together the calculated $\delta^{18}$O and $\delta$D values from La Herradura gold deposit plot in the field of metamorphic waters from Taylor (1974) (Figure 4.6). These results indicate a deep crustal source for the ore fluids, most likely of metamorphic origin, and are consistent with derivation of fluids from metamorphic dehydration reactions or by exchange of meteoric water with aluminosilicate-metamorphic rock. The narrow range of $\delta^{18}$O, 8.2 to 10.2 per mil for the ore fluids from La Herradura gold deposit, is typical for the estimated $\delta^{18}$O values for ore fluids from orogenic gold deposits of all ages from Neoarchean to Cenozoic, with values between 6 to 11 per mil for the Precambrian (McCuaig and Kerrich, 1998) and 7 to 13 per mil for the Phanerozoic (Bierlein and Crowe, 2000).

The $\delta^{34}$S values from sulfides fall in a narrow range between 3.1 and 6.6 per mil, consistent with a magmatic source for the sulfur. This magmatic source could be either a non-exposed igneous body or from igneous clasts in sedimentary protolith, that were devolatilized during metamorphic events. Carbon isotope data suggests sedimentary or metasedimentary origin of the carbon.

On the basis of stable isotopic data, there is no much evidence for variation in temperature and composition of ore fluids across the deposit or with the interpreted timing of vein formation. This indicates either that a very large volume of ore-forming fluid circulated at the same type of mineralization or that a large volume of rock was being subjected to a very uniform metamorphic facies. This scenario is similar to most of the Laverton gold province in Australia (Salier et al., 2005) and the OGD in Eastern Andean Cordillera of Peru (Haeberlin, 2002).
Figure 4.8. Plot of $\delta D$ vs $\delta^{18}O$ of water coexisting with gangue minerals at 330°C by using equation of Clayton et al. (1972).
CHAPTER 5. FLUID INCLUSION STUDY

5.1. Introduction

The main objectives of fluid inclusion studies on gold-rich and barren quartz veins in orogenic gold deposits are to estimate the chemical composition of the ore forming fluids, constrain the temperature and pressure at which these fluids were trapped, and define the P-T-V-X history of the vein-forming fluids. Most of the existing studies on fluid inclusions in orogenic gold deposits worldwide have documented aqueous and CO₂-rich fluid inclusions with estimated homogenization temperatures at 300 ± 50°C and 1-3 kbar (Groves et al., 1998). However, there are also many important lower and higher temperature deposits (200 to 700°C) formed over a continuum of depths that range between 2-20 km and at pressures varying from <1 to 6 ± 1 kbar (Barnicoat et al., 1991; Groves et al., 1998; Goldfarb et al., 2005).

The salinities of fluid inclusions generally vary between 3 to 12 equiv. wt percent NaCl, independently of the depth of formation of the deposit and the ages of mineralization. The low salinity has been interpreted to account for the low base-metal concentrations of gold deposits in metamorphic belts (Goldfarb et al., 2005), however, there are notable exceptions with higher salinities observed at certain deposits, such as Muruntau (Graupner et al., 2001) and in the Abitibi subprovince (Boullier et al., 1998), but these brine-bearing fluid inclusion generation postdates gold deposition.

The chemistry and origin of this type of ore fluid in northwest Sonora has not been studied in detail; however, some temperature and paleodepths have been obtained from several gold occurrences located in the trend of the Mojave-Sonora
megashear. Available studies include Albinson (1989), Silberman (1992), Perez et al. (1996), Lang (2003), and Noriega (2006), which all report low salinity fluids (all <10 wt% NaCl), a highly variable presence of CO₂, epithermal to mesothermal temperatures (~150 to ~340°C), and paleodepths between 1.5 to 7.0 km, mostly based on stratigraphic reconstruction and the chemical and physical properties of the ore fluid inclusion assemblage.

The objective of this study is to obtain information on temperature, pressure, and composition of the fluids that formed auriferous quartz veins and to compare these data with previous results published Phanerozoic orogenic gold deposits from North America and Peru. The latter is included because information has been published recently for these deposits.

5.2. Sampling

A total of 40 samples of quartz veins were collected from outcrops and diamond drill core in and around the Centauro pit area. Nineteen samples were prepared as removable doubly-polished fluid inclusion sections of 100 to 150 microns in thickness and, twelve of them were selected for petrographic descriptions of fluid inclusions and microthermometry studies in order to determine P-T-X characteristics of the fluids (Figure 5.1 and Table 5.1). Sampling was mainly focused on thin, discontinuous, extensional quartz veins with sericite envelopes, where the best identifiable fluid inclusions are found. These veins are spatially related to larger shear quartz veins, localized along major structures that control the distribution of ore-grade mineral (Figure 5.2 and 5.3). Most veins show development of elongated quartz
crystals growing at high angle to the wall of the vein, suggesting growth under tension (Figure 5.4). A few of them contain carbonate and iron oxides as a selvage on the quartz vein (Figure 5.5). All sampled veins were hosted by Proterozoic quartz-feldspar gneiss.

5.3. Analytical Method

Petrographic characteristics of fluid inclusions in 12 samples were examined in transmitted light using a Nikon binocular petrographic microscope. Seven samples were selected for thermometric analysis in order to determine P-T-X characteristics of the ore fluids. These studies were performed on 73 fluid inclusions hosted in quartz, using an SGE, Inc.-adapted USGS gas flow heating-freezing stage (Were et al., 1979). The SGE stage has a precision of $+0.1^\circ$C and $+1.4^\circ$C at 0.0°C and 374°C, respectively, and horizontal thermal gradients are below 0.2°C/cm. The stage was calibrated using SYN FLINC synthetic fluid inclusions (Sterner and Bodnar, 1984).
Figure 5.1. Map of La Herradura mine showing locations of samples used for fluid inclusions petrography and microthermometry studies. UTM, Datum NAD27.
Table 5.1. Samples used for fluid inclusion petrography-microthermometry, La Herradura.

<table>
<thead>
<tr>
<th>Sample</th>
<th>Area</th>
<th>East</th>
<th>North</th>
<th>MASL</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>LH-1</td>
<td>Megapit North</td>
<td>322107</td>
<td>3447864</td>
<td>154</td>
<td>Shear vein, white color (milky), 50 cm width, no Fe-oxides. Yellowish-white quartz vein 1-2 m thickness, sigmoidal form.</td>
</tr>
<tr>
<td>482-1</td>
<td>Centauro Pit</td>
<td>322376</td>
<td>3447114</td>
<td>16</td>
<td>The core is mainly quartz and scarce jarosite; at borders hematite in fractures, color is reddish black.</td>
</tr>
<tr>
<td>482-2</td>
<td>Centauro Pit</td>
<td>322376</td>
<td>3447114</td>
<td>16</td>
<td>Quartz vein, light gray color, cut by sericite-hematite veinlets. Fe-oxides disseminated and fractures, derived from pyrite.</td>
</tr>
<tr>
<td>525</td>
<td>Centauro Pit</td>
<td>322321</td>
<td>3447165</td>
<td>65</td>
<td>Quartz-carbonate extensional vein &lt; 1 cm thickness, carbonates are like selvages. Subhorizontal vein. Host is Quartz feldspar gneiss.</td>
</tr>
<tr>
<td>528</td>
<td>Centauro Pit</td>
<td>322276</td>
<td>3447114</td>
<td>60</td>
<td>Quartz-Fe-oxide (Py)-Fe-carbonate (ankerite). Sericite alteration halo.</td>
</tr>
<tr>
<td>530</td>
<td>Centauro Pit</td>
<td>322259</td>
<td>3447089</td>
<td>58</td>
<td>Quartz crystals euhedral habit with hematite (after pyrite) disseminated and on fractures. Sericitic alteration halos.</td>
</tr>
<tr>
<td>532</td>
<td>Centauro Pit</td>
<td>322258</td>
<td>3447092</td>
<td>35</td>
<td>Quartz-pyrite (now hematite) vein, 2 cm thickness, white color, low-dipping. This zone corresponds to high-grade mineral (2 g/t gold)</td>
</tr>
<tr>
<td>HGT-14</td>
<td>Centauro Pit East</td>
<td>322734</td>
<td>3447251</td>
<td>318</td>
<td>Quartz-pyrite-sphalerite vein, 8 mm thickness, sericitic haloes.</td>
</tr>
<tr>
<td>HGT-14</td>
<td>Centauro Pit East</td>
<td>322724</td>
<td>3447243</td>
<td>369</td>
<td>Quartz-pyrite vein, 13 mm thickness, sericite alteration haloes. Host rock is quartz-feldspar gneiss.</td>
</tr>
<tr>
<td>HGT-14</td>
<td>Centauro Pit East</td>
<td>322718</td>
<td>3447238</td>
<td>397</td>
<td>Quartz-pyrite-sphalerite vein, irregular form, 2 cm thickness, sericitic alteration halo.</td>
</tr>
<tr>
<td>HGT-16</td>
<td>North of Dunas Zone</td>
<td>321576</td>
<td>3447823</td>
<td>-119</td>
<td>Light gray quartz veinlet, cut by siderite-ankerite. Drusic textures.</td>
</tr>
</tbody>
</table>

Coordinate system is UTM NAD27

Figure 5.2. Photograph showing shear veins and associated flat extension veins (from Rhys, 2005).
Figure 5.3. Photograph showing extensional quartz vein with sericite-pyrite envelope. Centauro pit, bench -72.

Figure 5.4. Photomicrograph of extensional quartz veins. Note fibrous quartz crystals growing at high angles to the walls of the veins, which indicate growth under tension. El Centauro pit; full section view in crossed polars (from Lang, 2003).
5.4. Petrographic Characteristics of Fluid Inclusions.

Most of the studied samples contain quartz veins with deformational textures, commonly with crystals of quartz growing 120° to the hanging wall, suggesting that the quartz vein grew under tension (Lang, 2003). The fluid inclusions assemblages selected for microthermometry are located near the center of the veins, where the euhedral quartz crystals are less deformed and preserve the larger and best fluid inclusions to be studied. The shape of the fluid inclusions studied vary from irregular, elongate, ovoid and circular, with size ranging from 4 to 20 microns, averaging 8 microns in size. Most of the fluid inclusions studied were interpreted as primary according criteria of Shepherd et al. (1985). In the same way, there is a large group of secondary planes (fractures?), containing small and secondary fluid inclusions with sizes less than 1 micron. Only one type of fluid inclusion was observed (Figure. 5.6)
containing three phases (CO$_2$-rich vapor, CO$_2$-rich liquid and aqueous liquid), arranged concentrically. The CO$_2$ phase covers from 8 to 50 % of the inclusion area, but the vast majority fall in a narrower range of 15 to 30 %. No daughter minerals were seen.

Figure 5.6. Photomicrograph showing fluid inclusions. The inclusion contain an aqueous fluid, CO$_2$ liquid and CO$_2$ vapor. Sample HGT-14-488.70m. Plane polarized light; inclusion is 20 /_m long.
5.5. Analytical Results

Microthermometry data were collected from 54 fluid inclusions scattered in 7 samples, and is summarized in the Table 5.2. Freezing and heating analyses were focused on large groups of inclusions with constant liquid/CO$_2$ ratios, which were considered representative of most inclusions in a given sample. The temperatures obtained for each fluid inclusion were run at least ten times in order to assure the quality of the data obtained. Freezing runs were carried out before heating runs, because larger inclusions, independent of shape, were destroyed by decrepitation during heating runs. The principal observations are summarized below.

- The melting temperature of the CO$_2$ phase for all fluid inclusions was between -56.6 and -56.9°C (Table 5.2). A pure CO$_2$ solid phase melts (triple point) melts at -56.6°C. The slightly more negative values indicates the presence of additional, but in this case very minor, gaseous components like CH$_4$, N$_2$, and others gases in the CO$_2$ bubble (Shepherd et al., 1985).

- Clathrate melting temperatures were measured on 45 fluid inclusions, with temperatures ranging from 6.5 to 8.5°C in nearly all cases, but mostly around 7.5 to 8.5°C (Fig. 5.7). Based on these temperatures, the salinities of fluid inclusions were estimated using the FLINCOR program (Brown, 1989), with values ranging from 3 to 6.6 equiv. wt % NaCl, although the vast majority fall in a range of 3.5 to 4.8 wt % NaCl (Table 5.2). These estimates agree with others estimated by using the equation of Bozzo et al. (1973), valid for clathrate melting temperatures between 10 and about 0°C, with salinities ranging from 0 to 16 equiv. wt % NaCl.
• Homogenization temperatures of the CO₂ vapor phase into the CO₂ liquid phase were measured in 54 fluid inclusions. The temperatures obtained range from 24.1 to 30.2 °C, with the majority occurring between 27 and 29 °C (Figure 5.8).

• The final homogenization temperature of the CO₂ bubble into the aqueous liquid phase was obtained in 39 fluid inclusions. The measured temperatures range consistently between 245 and 295°C, but the majority are between 275 and 285°C (Figure 5.9). Many of the larger inclusions, independent of shape, but with higher volumes of CO₂, decrepitated at temperatures around 240 and 270°C due to high internal pressure.

• Calculations of fluid compositions using the FLINCOR program (Brown, 1989), indicate fluids containing between 2.2 and 7.4 mol % CO₂.

• Pressures (Table 5.2) were estimated for each sample using the melting temperature of clathrate, the homogenization temperature of the CO₂ vapor to liquid, the estimated volume fraction of CO₂, and the final homogenization temperature obtained, using the methods outlined in Brown and Lamb (1989), Bowers and Helgeson (1983), and the FLINCOR Program (Brown, 1989). Isochores were plotted for each sample from its final homogenization temperature to 334°C (Fig. 5.10), and the equations of the “tendency line” were obtained from each one. The algebraic “average isochore” was obtained,
plotted and, then the “average tendency line” was obtained. The intersection of this isochore (average) with the average temperature from all the samples fixes the minimal P-T trapping (average) for fluid inclusions. The upper intersection of the “average isochore” with an independent geothermometer (Roedder and Bodnar, 1980), in this case 330 ±20°C for deposition of sulfides obtained from sulfur isotopes, fixes the maximal P-T trapping for fluid inclusions. Based in the calculations outlined above and using the equation from Bowers and Helgeson (1983), the trapping pressure of the fluid inclusions lies between 1.6 and 2.7 kbar (Figure 5.10). This corresponds to paleodepths of between 4.3 and 7.3 km, considering an average density of 2.7 g/cm³ for the quartz feldspathic gneiss that hosts the mineralization in the La Herradura mine. The density for the host rock was obtained from thousands of measurements carried out by the Peñoles exploration team. The lower values for temperature and trapping pressure for sample HGT-16-344.90m compared to all other samples might reflect a decrease in the pressure regime related to episodic extension and opening of fractures during entrapment of these fluid inclusions.
Figure 5.7. Clathrate melting temperatures for all samples. The salinity of fluid inclusions can be estimated from these measurements (Bozzo et al., 1973).

Figure 5.8. Temperatures of homogenization for CO$_2$ (to liquid) for all samples.
Figure 5.9. Final homogenization temperature for all samples. The results span a remarkably narrow range in temperature, even though the samples come from across the entire deposit as well as from depth beneath Centauro pit, with most results between 265 and 285°C.

Figure 5.10 Plot of maximum and minimum isochores for fluid inclusions. Intersection of homogenization temperature with the isochores establishes trapping pressure for fluid inclusions.
Table 5.2. Microthermometry summary.

<table>
<thead>
<tr>
<th>Sample</th>
<th>Petrography</th>
<th>Microthermometry of fluid inclusions *</th>
<th>Flincor program (Brown, 1989)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td>Volume % CO₂ melting T</td>
<td>Clathrate melting T</td>
</tr>
<tr>
<td>482-1</td>
<td>7x8</td>
<td>20</td>
<td>-56.6</td>
</tr>
<tr>
<td>482-2</td>
<td>10x5</td>
<td>20</td>
<td>-56.6</td>
</tr>
<tr>
<td>528_1</td>
<td>5x8</td>
<td>25</td>
<td>-56.7</td>
</tr>
<tr>
<td>530_1</td>
<td>10x8</td>
<td>30</td>
<td>-56.6</td>
</tr>
<tr>
<td>HGT-14-488.7</td>
<td>6x5</td>
<td>20</td>
<td>-56.6</td>
</tr>
<tr>
<td>HGT-14-570.1</td>
<td>11x20</td>
<td>15</td>
<td>-56.6</td>
</tr>
<tr>
<td>HGT-16-344.9</td>
<td>10x4</td>
<td>50</td>
<td>-56.6</td>
</tr>
</tbody>
</table>

* At least 10 measurements for each fluid inclusion. T is Celsius.

5.6. Interpretation and Discussion

The results obtained in this study are very similar to those obtained from Lang (2003). Petrographic studies indicate that the ore fluids were entrapped at the same time during the syn-deformational precipitation of the auriferous quartz veins, because higher values of gold correlates closely with extensional quartz-sulfide veins (Figure 4.4). There is no significant variation of temperature or fluid composition in the deposit. This is in agreement with previous information obtained from stable isotopes. The uniformity in the characteristics of ore fluids implies that these fluids remained uniform and homogeneous over much of the vertical extent of the hydrothermal conduit.

On the basis of the depth of formation, the ore fluids, where fluid inclusions were collected, lie on the transition between epizonal and mesozonal classes (Groves et. al, 1998), corresponding to the greenschist metamorphic facies. Summarizing, ore fluids have the following characteristics:

- Mostly liquid-rich fluids.
• H₂O-CO₂ composition (2.3 to 6.8 mol % CO₂).
• A nearly pure CO₂ phase
• Low salinities, with values between 3.5 and 4.8 equiv. wt % NaCl.
• Final homogenization temperatures of the CO₂ bubble into the aqueous liquid phase range consistently between 245 to 295°C, but the majority are between 275 and 285°C (Figure 5-11).
• Trapping pressures between 1.6 and 2.7 kbar. This is equivalent to 4.3 to 7.3 km under a lithostatic pressure regime, lying in the transition between epizonal to mesozonal classes, in the greenschist metamorphic facies.
CHAPTER 6: DISCUSSION AND CONCLUSIONS

6.1. DISCUSSION

Mojave-Sonora Megashear

La Herradura mine occurs within a northwest trending belt that consists of metamorphic rocks of greenschist and amphibolite facies and granitoids of Proterozoic age (Nourse et al., 2005). The most outstanding tectonic feature in northwestern Mexico is the controversial Mojave-Sonora megashear (MSM) that separates two different basement blocks. The Caborca block lies to the west and has basement ages ranging from 1.8 to 1.7 Ga. The North American block to the east has basement ages ranging between 1.7 and 1.6 Ga. The La Herradura gold deposit is located within the older basement, i.e., in the Caborca block, as is confirmed with the new ages obtained by using U-Pb isotopes in zircons from within intrusive rocks and the gneiss belt hosting the orebody. These new ages agree with those previously published for similar rocks from the Caborca block (Anderson and Silver, 2005). Metavolcanic rocks outcropping to the north of La Herradura deposit yield ages of $167.8 \pm 3.5$ and $159 + 3.5, -3.7$ Ma for rhyolite and andesite, respectively. These data are also in agreement with the characteristic Lower to Middle Jurassic ages in the Caborca block. Moreover, on the basis of biostratigraphic studies, González et al., (2005) argued that the Late Permian-Early Jurassic El Antimonio Group, which is part of the Caborca Block, was originally positioned in southern Nevada and southeastern California, and later translated to its present location by MSM. However, Boucot et al. (2008), from paleontological studies indicated a continuous Devonian continental shelf along the entire southern margin of Laurentia that
contradicts the early Late-Jurassic Mojave-Sonora megashear as a viable hypothesis for large-magnitude offset of Proterozoic through Middle Jurassic rocks from California to Sonora.

As an alternative hypothesis to the MSM, Nourse et al. (2005) suggest that the two sequences of Proterozoic rocks in the region are the result of Paleoproterozoic accretion. No ultramafic rocks (which might expected along a suture) of this age have been reported in northwestern Sonora. However, little of the trace of the putative suture would be exposed, because of the development of later Tertiary alluvial basins.

A different alternative for the large-scale structure could be a combination of the paleoproterozoic accretion proposed by Nourse et al., (2005), and its subsequent reactivation as a result of younger tectonic events. Reactivation of this structure might have occurred during the emplacement of granitoid intrusions related to the Permo-Triassic magmatic arc of Mexico (stretching from Chihuahua to Chiapas) described by Torres et al., (1999). These authors report ages from 287 to 232 Ma, for the magmatic arc. A 233 Ma monzodiorite intrusion outcropping to the southwest of Sonoyta (Campbell and Anderson, 2003) should be part of this magmatic arc, bridging the gap between the Permo-Triassic granitoid belt of México and the Late Permian-Triassic (243 to 208 Ma) plutonic rocks of southern California (Barth et al., 1997). In addition, González et al. (2005) reported 270-240 Ma zircon grains within the El Antimonio Group, an Upper Permian-Triassic and Lower Jurassic sedimentary succession in northwestern Sonora. It is also possible this magmatic arc supplied part or all of those Permian volcanic detritus reported for the Upper Triassic Chinle Formation of the Colorado Plateau (Stewart et al., 1986).
Younger reactivations of the mega-structure would have occurred during subsequent tectonic events, like the Jurassic magmatic arc, the Late Cretaceous-Early Tertiary Laramide orogeny and, the Mid-Tertiary Basin and Range extension. However, the limited number of known Permo-Triassic intrusions in the area leaves room for doubts about these interpretations. Intrusions of this age are not reported for the North America plate. They could be an exclusive characteristic of the Caborca block; only detailed mapping and geochronology will cast further light on the discussion.

Association of gold and the Mojave-Sonora megashear

Regardless of the origin and the age, a large-scale structure is believed to play an important role in the distribution of gold deposits, displaying a northwestern belt from southern California to northwestern central Sonora. The belt includes the Picacho mining district, La Choya, Quitovac, La Herradura, Tajitos, Cerro Colorado, El Chanate, San Francisco and Lluvia de Oro, among other important gold prospects (Figure 2 of Silberman and Giles, 1988). Most of these gold deposits occurs within or adjacent to a regional structure, hosted by Precambrian, Mesozoic, and Tertiary rocks (Fig. 2 from Silberman and Giles, 1988). La Herradura is an outstanding example of this class of deposits hosted in Precambrian rocks, as well as La Choya for Mesozoic rocks, and Cerro Colorado for Tertiary rocks.

There are several geochemical analyses published from the gold occurrences located in northwest Sonora and Southern California (Silberman and Giles, 1988; Durgin and Teran, 1996, and Losh et al., 2005). These data show that the structurally controlled Au deposits considered by Silberman and Giles (1988) have Au/Ag ratios
of 2/1 to 6/1. In Arizona, the Au/Ag ratios range between 1/44.5 to 1/88.9 (Titley, 1997). The contrasting Au/Ag ratios suggest different metallogenic provinces in terms of underlying basement sources of precious metals. The higher ratios are found on both sides of the MSM.

Origin of the fluids

The ore fluids leading that deposited the La Herradura orebody were most likely generated following the prograde metamorphism of the continental plate during the Laramide orogeny. This timing is consistent with the age obtained for mineralization from Re-Os isotopes, which is 61.0 ± 2.1 Ma. The change in plate motions, passing from a compressional to a transpressional regime, could have reactivated older healed large-scale structures, which permitted the ore fluids migrate upwards, until they reached the more favorable ductile-brittle transition, which permitted the precipitation of gold in second-order structures. This structural relationship has been observed in most OGD (Goldfarb et al., 1988, 1991, 2005; Rushton et al., 1993; Haeberlin 2002; Haeberlin et al., 2003; 2004; Sillitoe and Thompson, 2006). In the La Herradura deposit, first-order structures could correspond to the trace of the postulated MSM or to the boundary of accretion between paleoterranes proposed by Nourse et al. (2005). Second-order structures may correspond to thrust faults developed during Sevier and/or Laramide orogenies.

The gold mineralization in La Herradura gold deposit came after peak metamorphic conditions. In the Centauro pit, some mineralization is hosted in metamorphic pegmatites, which have a retrograde greenschist overprint. The petrographic studies of fluid inclusions indicate that the ore fluids were entrapped at
the same time as the syn-deformational precipitation of the auriferous quartz veins, because of the correlation of distribution of gold with quartz-sulfide veins (Figure 4.4).

The calculated $\delta^{18}O$ and $\delta^D$ values from the La Herradura ore fluids at 330 $^\circ$C range mainly in a narrow interval between 8.2 to 10.2 per mil, and -34 to -19 per mil, respectively. These values indicate a deep crustal source for the ore fluids, most likely of metamorphic origin, and are consistent with derivation of fluids from metamorphic dehydration reactions or by exchange of meteoric water with aluminosilicate metamorphic rocks. The uniformity of $\delta^{18}O$ values also implies a very uniform source of fluids, normally being driven at orogen scale, and related to a single distal source, as has been documented for OGD hosted in the Superior and Yilgarn cratons (Robert et al., 2005), and the OGD in Eastern Andean Cordillera of Peru (Haeberlin, 2002).

The high radiogenic initial ratios of $^{187}\text{Os}/^{188}\text{Os}$ (0.3578) rules out mantle and oceanic sources, at least for the gold-associated pyrite. Therefore, source must be continental crust. In the same way, carbon in fluid inclusions is of organic origin implying a sedimentary or metasedimentary origin for the carbon. This fact discards an igneous source for the carbon.

The uniformity of $\delta^{34}S$ (3.1 and 6.6 per mil) suggest no partial reduction of sulfate. These values indicate that the sulfur source was isotopically uniform and derived directly from the mantle, from magmas, or indirectly from dissolution and/or desulfidation of primary sulfide minerals of average crustal sulfur isotopic composition (Hagemann and Cassidy; 2000). For La Herradura, the source of sulfur is
either a non-exposed igneous body or igneous clasts in sedimentary protolith that were devolatilized during metamorphic events.

Drawing all isotopic data from La Herradura deposit, the fluids probably originated in Au-enriched metasediments, containing organic C. The gold, sulfides, and other metals, are removed from a previously gold-enriched rock.

Pressure-temperature (P-T) gradients of regional metamorphism drive gold mobilization and desulfidation reactions from amphibolite or higher grade rocks, into lower-grade rocks. Under amphibolites facies, pyrite breaks down to pyrrhotite ± Fe-oxides ± Fe-silicates at amphibolite facies temperatures (Thompson, 1972; Connolly and Cesare, 1993), which produces HS\(^-\) ligands for selective complexing of gold. The fluids transport gold as Au-S complexes, either Au (HS)\(_2\)\(^-\) or HAu(HS)\(_2\). Base metal transport is restricted to the low Cl contents, because brines in paleobasins did not survive to greenschist or higher metamorphic facies.

Causes of ore deposition

Several generations of fluids were generated during episodic fracturing during Laramide time, probably as a response to the fluctuations in lithostatic-hydrostatic pressure. Fluids migrate through large-scale structures to shallower levels, until they reach the brittle-ductile transition, where permeable rocks improve movement of fluids, thereby pressure and temperature decrease and some gold precipitates, as is the case for the gold in extensional veins. However, the precipitation of gold is mainly driven by changes in the pH of the ore fluids, as a result of chemical reactions between ore fluids and wall rock, as is evidenced by the alteration of feldspar to sericite. Also, the high Fe/Fe + Mg ratios of wall rock promote the deposition of gold.
In La Herradura, the deposition of gold is favored for the higher Fe/Fe + Mg ratios in the granitoids within the Caborca block, compared to the granitoids within the North American block. Iriondo (2001) published Fe/Fe + Mg ratios ranging from 0.85 to 0.89 for granitoids within the Caborca block, and 0.72 to 0.82 for those located within the North America blocks. The precipitation of early pyrite, galena and sphalerite reduce the sulfur activity of the ore-forming fluids in the early mineralization stages. This change promotes gold deposition by destabilization the gold bisulfide complex (Henley and Brown, 1985). Later, as the hydrothermal system evolved, precipitation of gold continues in the shift of the fluids, from a slightly acid-near neutral to a slightly higher pH, into a more oxidized ore-fluid that favored the precipitation of gold accompanied by hematite. Subsequent stages in the evolution of fluids precipitated iron and magnesium carbonates in the periphery of the deposit.

Based on fluid inclusion and stable isotopic studies, there is no evidence of variations in temperature and composition, except for C isotopes, for the ore fluids across the deposit or with interpreted timing of vein formation. The isotopic uniformity of quartz veins implies that isotopic values of ore-forming fluids remained uniform and homogeneous over much of the vertical extent of the hydrothermal conduit.

Comparison of La Herradura ore fluids versus OGD in North America and Peru

The geochemistry of the ore-forming fluids from La Herradura compares well with those from OGD described in North America and Peru. The calculated $\delta^{18}$O and $\delta^D$ for the ore-forming fluids, combined with the temperature of formation of veins
and deposition of gold mineralization, overlap with the isotopic data field from other orogenic gold deposits from North America and Peru (Figure 6.1). In the Juneau gold belt, Goldfarb et al. (1991) estimated $\delta^{18}O$ and $\delta D$ values for the ore fluids between 7.2 to 12.8 and -25 to -15 per mil, respectively, for a vein hosted in late Paleozoic (?) through Cretaceous metasedimentary (± metavolcanic) sequences. In the Carson and Counterville deposits from the Mother Lode, California, Jia et al. (2003) reported $\delta^{18}O$ and $\delta D$ values between 9 to 12 and -20 to -15 per mil, respectively, for the ore fluids involved in the formation of quartz vein hosted in Paleozoic-Mesozoic metasedimentary, volcanic, and ultramafic rocks. Also Jia et al. (2003) published $\delta^{18}O$ and $\delta D$ calculated values between 7 to 8 and -130 to -120 per mil, respectively, for the quartz veins hosted in mid-Paleozoic to early Mesozoic metasedimentary and igneous rocks in the Sheba district, in the Klondike, Yukon Territory. The low $\delta D_{H2O}$ values reflect secondary exchange between recrystallizing mica and meteoric waters.

Leitch et al. (1991) reported calculated $\delta^{18}O$ between 10 to 14 per mil for the Bralorne and Pioneer mesothermal veins in Bridge River, southern British Columbia, and Maheux (1989) published $\delta D$ calculated values between -38 ±18 per mil for antimony-associated gold mineralization in the Bridge River district, British Columbia, Canada. These veins are hosted in Permian (?) to Jurassic metavolcanics and metasedimentary rocks. Haeberlin (2002) obtained $\delta^{18}O$ and $\delta D$ calculated values between 8.2 to 10.2 and -130 to -120 per mil, respectively, for the ore fluids involved in the Carboniferous orogenic gold deposits in the Pataz gold province, Peru. These deposits are hosted in Paleozoic intrusions and in volcanic and sedimentary rocks.
From fluid inclusion study, the more common characteristics are the chemistry of the fluids (Table 6.2), these characteristics include low salinities, the presence of CO$_2$, which in individual deposits can be variable (Goldfarb et al., 1998), dependant in the depth and pressure of trapping for fluid inclusions. Another very important shared feature is the temperature of homogenization for fluid inclusions, mostly on the order of 200 to 350°C, which means that the ore fluids were trapped in the greenschist metamorphic grade host rocks, with pressures ranging from 1.2 to 2.3 kbar. In the Pataz gold province in Peru, however, the estimated pressure in the early stages was 4.5 kbar, and in the later evolution of fluids, these were moving to smaller pressures in response to a tectonic uplift (Haeberlin, 2002).

Based in the tectonic and structural setting, the geometry and style of mineralization, the chemistry and the origin of the ore fluids involved in La Herradura gold deposit, it is conclusive that this deposit corresponds to an orogenic gold deposit type, the first OGD described in the northwestern Mexico.
Table 6.1 Oxygen and Hydrogen Compositions of Quartz and Quartz-Carbonate veins in the Western North American Cordillera and Counterparts Worldwide (modified from Jia et al., 2003).

<table>
<thead>
<tr>
<th>Gold Province/District</th>
<th>Vein location</th>
<th>$\delta^{18}$O quartz (‰)</th>
<th>$\delta^D$ sericite (%)</th>
<th>T (°C)</th>
<th>$\delta^{18}$O ore fluids (%)$^1$</th>
<th>$\delta^D$ ore fluids (%)$^2$</th>
<th>References</th>
</tr>
</thead>
<tbody>
<tr>
<td>Juneau Gold Belt</td>
<td>Juneau</td>
<td>15.2 to 20.8</td>
<td>-75 to -53</td>
<td>300</td>
<td>7.2 to 12.8</td>
<td>-35 to -15</td>
<td>1</td>
</tr>
<tr>
<td>Mother Gold, California</td>
<td>Carson and Coulterville</td>
<td>16.5 to 18.6</td>
<td>9 to 12</td>
<td>300</td>
<td>6 to 11</td>
<td>-20 to -15</td>
<td>2</td>
</tr>
<tr>
<td>Oro Rico and McAlpine</td>
<td></td>
<td>15.0 to 17.5</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Klondike, Yukon Territory</td>
<td>Sheba</td>
<td>14.6 to 15.1</td>
<td>7 to 8</td>
<td></td>
<td></td>
<td>-130 to -120</td>
<td>1</td>
</tr>
<tr>
<td></td>
<td></td>
<td>14.7 to 15.1</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>3</td>
</tr>
<tr>
<td>Bridge River, southern BC</td>
<td>Pioneer</td>
<td>17.8 to 18.1</td>
<td>11 to 12</td>
<td></td>
<td></td>
<td>-15</td>
<td>1</td>
</tr>
<tr>
<td>Tasman orogenic belt</td>
<td>Central and North Deborah</td>
<td>14.4 to 17.2</td>
<td></td>
<td>350</td>
<td>8 to 11</td>
<td>-37 to -17</td>
<td>7</td>
</tr>
<tr>
<td>Bendigo, Australia</td>
<td></td>
<td>14.4 to 17.2</td>
<td>8 to 11</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Birimian greenstone belt</td>
<td>Ashanti gold field</td>
<td>12.8 to 15.6</td>
<td>9 to 12</td>
<td>400</td>
<td></td>
<td>-53 to -37</td>
<td>8</td>
</tr>
<tr>
<td>Ashanti, west Africa</td>
<td></td>
<td>12.8 to 15.6</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Superior Province of Canada</td>
<td>Hollinger-McIntyre, Dome</td>
<td>12.5 to 15.0</td>
<td></td>
<td>220 ~ 450</td>
<td>6 to 11</td>
<td>-80 to -20</td>
<td>9</td>
</tr>
<tr>
<td>Abitibi belt</td>
<td></td>
<td>12.5 to 15.0</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Yilgarn Craton, W. Australia</td>
<td>Kalgoorlie and Norseman</td>
<td>11.4 to 13.4</td>
<td></td>
<td>220 ~ 500</td>
<td>5 to 9</td>
<td>-40 to -10</td>
<td>10</td>
</tr>
<tr>
<td>Norseman-Wiluna belt</td>
<td></td>
<td>11.4 to 13.4</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Eastern Andean Cordillera, Perú</td>
<td>Pataz-Marañon Valley Au belt</td>
<td>10.9 to 14.2</td>
<td>-60 to -39</td>
<td>330</td>
<td>6.2 to 8.6</td>
<td>-19 to -40</td>
<td>11</td>
</tr>
<tr>
<td>La Herradura, Mexico</td>
<td>Centauro Pit</td>
<td>14.1 to 16.1</td>
<td>-109 to -57</td>
<td>330</td>
<td>8.2 to 10.2</td>
<td>-19 to -34</td>
<td>12</td>
</tr>
</tbody>
</table>

$^1$The $\delta^{18}$O values for vein-forming fluids were calculated from quartz-water equilibrium equation of Clayton et al. (1972)

$^2$The $\delta^D$ for vein-forming fluids were calculated from muscovite-water equilibrium fractionation factor of Suzuki and Epstein (1976)

Figure 6.1. Plot of $\delta^1$D Vs $\delta^{18}$O of water coexisting with gangue minerals. Data from selected orogenic gold deposits from North America and Perú, compared to La Herradura gold deposit.

Table 6.2. Comparison of fluid inclusions results from La Herradura with orogenic gold deposits located in the circum-Pacific of North and South America.

<table>
<thead>
<tr>
<th>Deposit</th>
<th>Temp °C</th>
<th>Salinity wt% NaCl</th>
<th>CO2 mol%</th>
<th>Pressure (Kbar)</th>
<th>Source</th>
</tr>
</thead>
<tbody>
<tr>
<td>Juneau Gold Belt, SE Alaska</td>
<td>200-325</td>
<td>&lt;5 wt%</td>
<td>&gt;50%</td>
<td>1.5</td>
<td>Goldfarb et al.(1988)</td>
</tr>
<tr>
<td>Mother Gold, California</td>
<td>250-325</td>
<td>2</td>
<td>&lt;10</td>
<td>&lt;2 Kb</td>
<td>Weir and Kerrick (1987)</td>
</tr>
<tr>
<td>Klondike in Yukon, Canada (Sheba)</td>
<td>323 ± 18</td>
<td>4.3 ± 1.5</td>
<td>5.8-7.9</td>
<td>.3 - .9 to 2.3 ± .7</td>
<td>Rushton et al. (1993)</td>
</tr>
<tr>
<td>Bridge River, British Columbia, Canada</td>
<td>230-350</td>
<td>&lt;5</td>
<td>5-15</td>
<td>1.75</td>
<td>Leitch et al. (1991) and Maheux (1989)</td>
</tr>
<tr>
<td>Pataz-Marañon, Peru</td>
<td>140 - 330</td>
<td>1 to 8</td>
<td>low</td>
<td>2.1-4.5</td>
<td>Haeberlin (2002)</td>
</tr>
<tr>
<td>La Herradura</td>
<td>245-295</td>
<td>3.5-4.8</td>
<td>2.3-6.8</td>
<td>1.6-2.7</td>
<td>This study</td>
</tr>
</tbody>
</table>
6.2. CONCLUSIONS

The La Herradura mine is located in northwestern Sonora Mexico. It occurs within a northwest trending belt of metamorphic rocks of greenschist and amphibolite facies, and granitoids, of Proterozoic age. The deposit is hosted preferentially in the more competent quartz-feldspar gneiss, 1.75 Ga age. It is limited to the east and west for Victoria and Ocotillo shear zones respectively; these shears exhibit a northwest trend, that are separating the Proterozoic structural block to the east from Jurassic clastic rocks and subvolcanic intrusions, and to the west by Upper Paleozoic limestone. Most of the gold mineralization occurs in thin quartz-sulfide (leached in outcrop) veins following en echelon tensional swarms. These veins are low-dipping (<30°) and are interconnected to a thicker set of fault-fill higher-grade quartz vein systems, that are believed to be the feeders for gold mineralization in the entire deposit.

Stable isotope data indicate that ore fluids have a metamorphic origin, probably derived from a deep crustal source as a result of regional metamorphic dehydration reactions of the continental plate, following a prograde metamorphism during Laramide Orogeny. The calculated temperature deposition of sulfides accompanying gold mineralization is 330 ±20°C. Fluid inclusion studies in quartz-sulfides-gold bearing veins show that the ore fluids are CO₂-rich, low in salinity, and have homogenization temperatures between 285-295 °C, and a calculated trapping pressure between 1.6 and 2.7 Kbar.
The age obtained from Re-Os isotopes on pyrite and magnetite-gold associated is
61.0 ± 2.1 Ma. This is the first age obtained on mineralization for this type of deposits in
Mexico.

The style and geometry of mineralization, stable isotopes and fluid inclusions
studies, indicate that La Herradura gold deposit can be classified as an orogenic gold-type
deposit, the first OGD described in northwest Mexico.
REFERENCES


