## ORE PETROLOGY AND

#### WALLROCK ALTERATION STUDIES

#### AT THE

## LAKE SHORE GOLD MINE, KIRKLAND LAKE, ONTARIO.

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# A THESIS SUBMITTED AS PARTIAL FULFILLMENT FOR THE HONOURS BACHELOR OF SCIENCE DEGREE.

FACULTY OF SCIENCE McMASTER UNIVERSITY HAMILTON, ONTARIO April, 1985.

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#### ACKNOWLEDGEMENTS

The production of this thesis required the help of many talented and generous people, to all of whom I an indebted. I wish to thank Tom Bryner, Abdul Kabir, Ota Mudroch, Jack Whorwood and Len Zwicker for their technical assistance, and John Cekr for his helpfulness during the past four years. Many thanks to Frank Ploeger, whose assistance during sample collecting Was oreatly appreciated. I am very grateful to Ian Williams and Steve Bonnyman for the use of their computing talents and facilities.

Special thanks are extended to Andy Fyon, for his extensive library, informative conversations and friendship throughout the past year.

A substantial debt is owed Jim Crocket, who had enough confidence in my ability to allow this thesis to be as comprehensive as it is in these days of financial restraint.

The preparation of this thesis is due to a large part to Noreen Evans whose patience and hard work has greatly improved the original manuscript. Finally, I would like to thank my family and friends for their continual support.

Funding for this work was provided by a National Research Grant awarded Dr. J.H. Crocket.

#### ABSTRACT

The Lake Shore Mine has accounted for over a quarter the gold produced in the Kirkland Lake of camp. Petrography suggests that gold is a later phase in the paragenesis of ore minerals. Sulphides precipitated first followed by tellurides and gold. Wallrock alteration adjacent to the quartz veins was studied by thin section geochemical analysis. Petrography enabled and an alteration facies scheme based on predominant alteration minerals to be established. The alteration assemblages suggest the fluid had a high K and  $CO_{2}$  content, low Na, and reduced S content. Geochemical analysis of these samples confirm petrographic observations and suggests major additions of KzO, SiOz, Rb, Sb and Au to the wallrock, and losses of NazO, Ba and Zr from the wallrock.

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#### INTRODUCTION

The Kirkland Lake camp is located 500 km north of Toronto in the Abitibi Greenstone Belt (fig.1). The first gold discovery in the camp was made in 1911 by W.M. Wright on the portion of the property now known as the Wright-Hargreaves mine (Burrows and Hopkins, 1923). Production fully underway around 1917, and since then was approximately 910,000 kg of gold have been produced from seven mines (Ploeger and Crocket, 1981). From west to east (fig. 2), these mines are the Macassa, Kirkland Lake 10m Te Gold, Teck Hughes, Lake Shore, Wright-Hargraves, Stannite, and Toburn. Only the Macassa Mine has been in continuous operation to present. The Lake Shore Mine began new operation in 1983.

#### Lake Shore Mine

The Lake Shore Gold Mine, from the beginning of operation in 1918, until the closing in 1965, produced 264,350 kg of gold and 60,800 kg of silver (Ploeger and Crocket, 1981). The mine was reopened in 1983 to remove the crown pillar section, which has proven reserves of approximately 82,000 tons, grading 16.8 gm/ton of gold (Northern Miner, 1985).



Geological Environment

The Kirkland Lake gold deposits are mainly hosted in an alkalic plutonic suite which intrudes sediments, volcanic flows, and pyroclastic rocks (fig.2).

The ore is localized predominantly in faults within the intrusives, which provided a structural control for the movement of auriferous hydrothermal fluids. This fault system is known as the Kirkland Lake Main Break (Burrows and Hopkins, 1923). Charlewood and Thomson (1948) described the ore controls as:

- 1) fractures in quartz veins,
- highly altered wallrock and in wallrock inclusions in quartz veins,
- 3) association with pyrite and
- 4) association with tellurides.

In this study, all types of gold mineralization were found.

Wallrock Alteration

Hydrothermal alteration of wallrock around vein and fracture systems is observable as gradational colouration changes from a purple-maroon to red-pink and then to a buff green-gray, with increasing alteration intensity. Colour variation is due to mineralogical and chemical changes in the rock. Figure 2. Geology and mine properties of the Kirkland Lake Camp. (Ploeger and Crocket, 1981).



This thesis presents studies of ore petrology and hydrothermal alteration, as well as geochemical analysis of samples taken at the Lake Shore Mine during the spring of 1984. An interpretation of the mineralogical and geochemical variations which accompanying gold mineralization in the Lake Shore Mine is also offered.

#### GENERAL GEOLOGY

The geology of the Kirkland Lake gold area has been described by Todd (1928), Thomson (1948), Jensen (1980), Ploeger, (1980), and Ploeger and Crocket, (1981). The gold mineralization is predominantly hosted in symmitic alkalic intrusives which cut a homoclinal south facing sequence of conglomerates, wackes, trachytic flows, and pyroclastic rocks, known as the Timiskaming Group (Hewitt, 1963). The intrusives have been dated at 2685  $\pm$  10 my. (Colvine et al., 1984).

The intrusive suite is composed of three phases: augite symmite, symmite and porpyritic symmite (Thomson, 1948). These phases may represent daughter products of one parent alkalic magma (Ploeger, 1980).

According to Colvine et al. (1984) the syenite intrusives serve as hosts to the gold mineralization because of the favourable mechanical properties of the rocks. Under high stress the sympites undergo brittle fracture, thus providing a conduit for the movement of Faulting ore-bearing hydrothermal fluids. as a structural control within the intrusives is apparent in that both the fault zones and the intrusive suite are concordant (see fig. 2).

Ore zones at the Lake Shore Mine are primarily hosted

in the symmite porphyry, especially in the eastern half, and at lower levels of the mine (Charlewood and Thomson. 1948). The displacement on the fault zones is not clearly understood due to the absence of well defined markers: however, Thomson (1948) suggested from available evidence that vertical movements of 400 to 490 meters are possible. This is an important factor, as it makes geochemical and mineralogical correlations from the footwall across the vein into the hanging wall of questionable significance, even though the rock type on either side of the fault is the same porphyritic syenite. Mineralogical and chemical are similar. but when dealing with changes 1 OW concentrations of trace elements, exact correlation across displaced fault zones is suspect because minor element content may be variable over small volumes of rock. For however, the sympite porphyry can major elements, be considered a homogeneous medium. Chemical enrichments and depletions across the displaced fault zone are considered to represent only the affects of hydrothermal alteration in the case of the major elements, and not to reflect structural displacements.

#### ECONOMIC GEOLOGY

Two vein systems trending approximately  $65^{\circ}$  at the surface cross the property. Thomson (1948) reported that the North (number 2) vein dips from  $75^{\circ}$  to  $85^{\circ}$  south and intersects the South (number 1) vein at depth. Samples from this study were taken from the North vein, across a section where symplet porphyry is the host intrusive.

The veins contain a variety of quartz, ranging in colour from black to smoky to milky white. The dark colouration is due to the presence of graphite and/or molybdenite, which line fractures and slips within the quartz. Altered blocks of syenite porphyry occur as inclusions in the quartz veins, and were probably emplaced by recurrent movement along the fault during quartz precipitation. Calcite, dolomite, and chlorite occur within the veins in varying abundance.

#### ORE MINERALS

The common ore minerals are native gold and the gold tellurides calaverite (AuTe2) and petzite ((Au,Ag)2Te). Native gold is found as small flakes, which generally range from 0.3 mm to a few  $\nearrow$  m in length. It often occurs intergrown with the gold tellurides, as well as with altaite (PbTe). These minerals are ubiquitous within Map 1

Map of a vertical section along the North Vein system showing "L4E" sample locations. Vertical Scale is equal to horizontal scale.



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1,

the veins. In hand sample, where molybdenite coats slip surfaces, a thin film of gold sometimes can be seen smeared over the molybdenite. The only visible gold noted was in a vein that assayed 450 gm/ton.

Pyrite is common in the vein material and wallrock adjacent to the vein is highly pyritiferous. Chalcopyrite also occurs in the vein, but is less common.

Magnetite and ilmenite are found predominantly in the wallrock and where adjacent vein material have been partly replaced by pyrite. The magnetite and ilmenite within the syenite porphyry are probably of magmatic origin. Hematite occurs as an alteration product in the wallrock. Hawley (1948) noted the occurrence of various other telluride minerals, including colorado\_ite (HgTe), melonite (NiTez), tetrahedrite (BizTezS), and joseite (BizTeS). Arsenopyrite, pyrrhotite, sphalerite, and galena were also noted. These minerals are not present in economic concentrations.

#### METHODOLOGY

#### Sampling Methods

Samples at the Lake Shore Mine were collected in May, 1984, with the help of Jim Crocket and mine geologist, Frank Ploeger. The two sets of samples used

for this study were obtained from an actively mined stope, and from core samples drilled roughly normal to the North vein system at about the 100 foot level in the mine. Figure 3 shows the sample locations. The stope samples are designated as the "L4E" series, and the core samples as the "L" series. The L4E sample map (map 1) shows a complex vein system in porphyritic syenite with slip surfaces and wallrock inclusions.

Sample L-65 was chosen as the least altered porphyry sample due to its dark maroon colour, easily identifiable porphyritic feldspars, and the general absence of fractures and chlorite slips. These observations were confirmed by thin section microscopy.

#### Analytical Methods

Major Element Analysis

powders Geochemical analysis was conducted on prepared by grinding rocks in a ceramic pulverizer and shatterbox to -200 mesh. Both series of samples were prepared for analysis in this way. Major elements were determined by X-ray fluorescence analysis, using fusion pellets on a Philips PW1450 automatic sequential spectrometer. Fusion pellets were prepared following the method of Norrish et al (1969). To get an estimation of the relative error produced by sample preparation and



## Figure 3.

**50** 100 m

analysis, duplicate samples were included for X-ray fluorescence work. A sample size of 19 with two duplicates is too small for rigorous statistical analysis, but table 1 provides the relative error on the oxides analysed for in this study.

Table 1.

Oxide	Detection limit (Fyon, 1980)	Precision
Si0z	0.1 wt%	0.3 wt%
Al 203	0.1 wt%	0.4 wt%
FeaOs	0.1 wt%	2.0 %
MgO	0.1 wt%	7.0 %
CaO	0.1 wt%	2.0 %
NazO	2.0 wt%	90.0% to 4.0 %
K <b>20</b>	0.1 wt%	2.0 %
Ti Oz	0.01 wt%	2.0 %
MnO	0.05 wt%	100.0 %
P20s	0.04 wt%	10.0 %

Trace Element Analysis

The following trace elements were determined by XRF analysis: Rb, Sr, Y, Zr, Nb, Ba, Ce, La, and Nd. Samples were analysed in powder pellet form prepared following the method of Marchand (1973). Duplicate samples were not included due to time constraints. Significant error is probable in the reported concentrations of Nb and Y since their abundance levels are near or below the detection limits which are approximately 3 ppm.

Epithermal Instrumental Neutron Activation Analysis (EINAA) was used to determine the concentrations of Au, As, Sb, and W. The samples were sealed in poly vials and irradiated for 1 MW hour in the McMaster University pooltype reactor in a cadmium shielded position. The two sample packages each contained two rock powder standards, and were continuously rotated during the irradiation.

The samples were counted on a coaxial intrinsic germanium detector (APTEC Engineering Company, Toronto). The half-lives for each radionuclide are given below:

### Table 2.

Nuclide	Half-Life
1 <b></b> 7W	24.0 hours
<sup>7</sup> ≜As	23.6 hours
1 <b>≁</b> ⊜Au	64.8 hours
***Sb	66.0 hours

W and As were counted after 54 hours cooling because of their short half-lives. Due to the presence of high gold concentrations in the ore samples and As interference on the Sb peak, more reliable Au and Sb results were obtained on a second count, five days after irradiation. An example of elemental concentration calculations is given in Appendix 1.

#### Errors

Duplicate samples were run in order to determine the precision and relative error. The results of this analysis are presented below:

Table 3.

	Sb	As	Au	W
L4E-3a	40	70	10,044	168
L4E-3b	152	77	13,744	184
<del>x</del> =	96	73.5	11,874	176
	56	3.5	1,850	8
% error	5600 = 96	4.8% 58.3%	15.6%	4.5%
L-101a	38	6	27	3
L-1016	39	17	27	13
<del>x</del> =	38.5	11.5	27	8
20	0.5	5.5	ο	5
% error	50 = 38.5	48% 1.3%	0%	63%

Although rigorous statistical analysis cannot be carried out, comments on analytical error can be made. Firstly, sample L4E-3b contains an exceptionally high concentration of antimony compared to all other samples.

Table 4.	Major an	d trace e	lement an	alysis.		
5	L4E÷1A	L4E-2	L4E-3	L4E-4	L4E-5	L4E-6
SiO <sub>2</sub>	95.89	39.68	44.40	70.42	70.97	57.25
A1203	1.60	20.01	13.10	12.39	12.98	12.23
Fe <sub>2</sub> 03	0.58	14.63	11.39	2.52	2.74	9.88
Mg0	0.00	3.25	5.48	0.70	0.50	1.73
Ca0	0.22	1.23	7.38	1.55	1.03	2.56
Na <sub>2</sub> 0	0.47	0.16	0.41	0.44	0.35	0.30
к <sub>2</sub> 0	0.45	9.50	7.51	8.97	8.00	7.07
TiO <sub>2</sub>	0.03	1.34	0.91	0.21	0.25	0.82
MnO	0.00	0.01	0.13	0.02	0.01	0.05
P205	0.02	0.32	0.46	0.08	0.09	0.11
н <sub>2</sub> 0	0.27	8.38	0.00	0.19	1.37	3.91
co <sub>2</sub>	0.48	1.48	8.83	2.49	1.69	4.09
Ba	337.00	1437.00	1273.00	3037.00	2357.00	1223.00
Ce	5.00	143.00	70.00	48.00	71.00	57.00
Nd	2.00	74.00	52.00	36.00	54.00	48.00
La	6.00	76.00	27.00	18.00	32.00	50.00
Nb	19.00	2.00	2.00	14.00	14.00	4.00
Y	18.00	2.00	14.00	29.00	23.00	14.00
Zr	35.00	205.00	151.00	161.00	159.00	139.00
Rb	19.00	363.00	199.00	172.00	182.00	182.00
Sr	77.00	133.00	519.00	347.00	263.00	304.00
Sb	91.00	61.00	40.00	42.00	33.00	46.00
W	1.00	253.00	176.00	73.00	72.00	177.00
As	25.00	62.00	74.00	34.00	30.00	51.00
Au	792.20	7.67	11.89	8.04	6.13	40.60

(Majors expressed as weight percent, trace elements in ppm).

Table 4. con't.

	L4E-9B	L4E-10	L4E-13	L4E-14	L4E-16	L4E-17
SiO <sub>2</sub>	6 <b>4.8</b> 8	64.85	64.10	66.65	63.77	58.72
- A1 <sub>2</sub> 0 <sub>3</sub>	17,00	14.56	14.08	14.02	12.65	13.05
$Fe_2O_3$	3.31	2.89	2.97	3.78	2.76	3.87
MgO	1.23	1.78	2.31	1.63	2.84	3.26
Ca0	0.95	2.03	3.58	2.71	4.50	5.55
Na <sub>2</sub> 0	2.93	0.59	2.80	4.23	2.55	2.57
к <sub>2</sub> 0	6.00	7.95	5.59	2.90	3.69	3.74
TiO <sub>2</sub>	0.34	0.27	0.32	0.32	0.35	0.38
MnO	0.01	0.04	0.06	0.05	0.06	0.08
P <sub>2</sub> 0 <sub>5</sub>	0.21	0.17	0.19	0.17	0.29	0.31
н <sub>2</sub> 0	1.62	0.00	0.00	0.00	0.00	0.29
c0 <sub>2</sub>	1.51	2.86	3.99	3.38	6.54	8.17
Ba	981.00	1519.00	3 <b>58</b> 4.00	2059.00	888.00	1201.00
Ce	70.00	68.00	67.00	35.00	72.00	88.00
Nd	64.00	52.00	48.00	40.00	24.00	62.00
La	43.00	47.00	40.00	43.00	38.00	54.00
Nb	16.00	11.00	15.00	14.00	14.00	13.00
Y	30.00	23.00	29.00	28.00	32.00	31.00
Zr	216.00	171.00	196.00	210.00	217.00	249.00
Rb	167.00	190.00	118.00	68.00	127.00	114.00
Sr	261.00	259.00	285.00	559.00	604.00	920.00
Sb	18.00	20.00	15.00	57.00	89.00	77.00
W	28.00	52.00	31.00	41.00	22.00	26.00
As	5.00	21.00	5.00	7.00	23.00	12.00
Au	4.33	3.06	9.77	17.86	3.08	1.05

-Table4 . con't.

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	L-65	L-69	L-76	<b>L</b> -79	L-87	L-101	L-104
Si <b>0</b> 2	66,06	65.98	66.71	65.93	66.43	66.86	65.58
A1203	15.42	15.51	15.15	15.42	15.06	15.01	14.93
$Fe_20_3$	3.21	3.15	2.85	2.49	2.73	2.80	2.62
MgO	1.70	1.81	1.78	1.58	1.63	1.57	1.47
Ca0	2.87	2.55	2.33	2.72	2.62	2.55	3.17
Na <sub>2</sub> 0	4.53	4.53	4.69	4.20	3.91	4.46	4.61
K <sub>2</sub> 0	2.93	3.30	2.72	3.09	3.11	2.89	2.86
TiO2 .	0.32	0.31	0.30	0.31	0.31	0.31	0.31
MnO	0.03	0.02	0.04	0.04	0.04	0.04	0.05
P <sub>2</sub> 0 <sub>5</sub>	0.18	0.18	0.15	0.15	0.14	0.14	0.14
H <sub>2</sub> 0	0.53	0.62	0.09	0.22	0.27	0.46	0.02
co <sub>2</sub>	2.21	2.03	3.16	3.84	3.74	2.90	4.23
Ba	4117.00	2768.00	1808.00	1377.00	2446.00	1445.00	. 2870.00
Ce	61.00	50.00	58.00	67.00	47.00	50.00	50.00
Nd	26.00	32.00	14.00	32.00	28.00	20.00	34.00
La	52.00	29.00	40.00	40.00	27.00	31.00	34.00
Nb	13.00	12.00	17.00	13.00	13.00	11.00	15.00
Y	35.00	30.00	35.00	32.00	29.00	26.00	32.00
Zr	362.00	266.00	233.00	206.00	211.00	214.00	242.00
Rb	51.00	67.00	79.00	94.00	93.00	70.00	73.00
Sr	2008.00	1106.00	666.00	545.00	579.00	722.00	889.00
W	1.00	2.00	6.00	5.00	1.00	8,00	1.00
As	14.00	7.00	11.00	10.00	7.00	12.00	11.00
Au	•0 11	0.004	0.011	0.051	0.478	0.027	0.01

This is due to contamination or Sb heterogeneity in the sample. Secondly, very low concentrations (0 - 15 ppb) of W or As are liable to large errors up to 50 to 60% . Higher abundances are likely to have much lower errors, about 5%. Lastly, Au concentrations are quite reliable at both low ppb and high ppm levels.

#### Loss on Ignition

When weighed powdered samples are heated in an oven at 1040 C for an hour and then reweighed, there is usually a significant decrease in weight due to vaporization of HzO, CO<sub>2</sub> and SO<sub>2</sub> at high temperatures. Thus, "loss on ignition" (LOI) is equivalent to the volatile content of the sample. While HzO and CO<sub>2</sub> are abundant components of hydrothermal systems, SO<sub>2</sub> is generally not, and therefore the weight LOI is approximately equal to the HzO and CO<sub>2</sub> weight percent of the sample (ie. weight % HzO = LOI weight % CO<sub>2</sub>). These values are given in table 4.

## CO2 Analysis

CO2 analysis was performed on a LECO carbon detector and results are listed in table 4. Duplicate analysis showed errors of less than 30%. Weight percent H2O can be calculated by subtracting the CO2 content from the LOI value. In general, this gives adequate results for the

relative H2O content; however, samples with high sulfide content give erroneously high CO2 values since some of the sulfide is oxidized and counted as CO2. The CO2 values for pyrite-rich samples L4E-3, L4E-10, L4E-13 and L4E-14 are higher than the LOI values, due to sulfur oxidization.

#### Specific Gravity Determinations

In order to study the net gains and losses of chemical components with respect to alteration intensity, the composition-volume relationship derived by Gresens (1967) is used. Utilizing this procedure, the chemical changes in hydrothermally altered rocks can be determined (relative to the chemistry of a least altered rock), in terms of mass. This is done by finding the ratio of the density of an altered rock with respect to the least altered rock and substituting this value into the Gresens equation (Appendix 2).

Density determinations were carried out using a Berman balance as described by Hutchinson (1974). Rock chips weighing between 15 to 25 mg were weighed in air and then toluene to determine their density. The balance was calibrated using quartz and fluorite chips, which gave consistently precise values. Generally, three to four different chips from each sample were randomly chosen to provide an average density and to test reproduciblity. In all cases, reproduciblity was quite good. Results and precision for the analysis are listed in table 5.

Table 5. Specific grav	ity determinations.	
Calibration:		
Fluorite 3 3 3 X=3 known =3	L4E-6 2.65 2.80 $\overline{X} = \frac{2.80}{2.75}$ $\overline{C}_{c} = 2.78$	L-69 2.67 2.58 2.64 $\overline{X} = \frac{2.67}{2.64}$ $\mathbf{e}_{c}^{=} 2.67$
$\boldsymbol{\rho}_{c} = \boldsymbol{X}$	L4E-9B 2.63 $\overline{X} = \frac{2.64}{2.64}$ $\overline{P}_{c} = 2.67$	L-76 2.61 $\vec{X} = \frac{2.61}{2.61}$ $\boldsymbol{Q}_{c} = 2.64$
L4E-1A: 2.66 2.64 $\overline{X} = \frac{2.66}{2.65}$ $\mathbf{r}_{c} = 2.68$	L4E-10 2.67 2.69 $\overline{X} = \frac{2.70}{2.69}$ $e_{c} = 2.72$	L-79 2.67 2.66 2.63 2.64 2.63
L4E-2 2.93 2.85 $\vec{X} = \frac{2.94}{2.91}$ $\rho_{c} = 2.94$	L4E-14 2.68 $\overline{X} = \frac{2.67}{2.68}$ $\mathbf{e}_{c} = 2.71$	$\overline{X} = \frac{2.65}{2.68}$ L-87 2.67 2.60 2.62
L4E-3 2.90 $\overline{X} = \frac{2.90}{2.90}$ $e_{c} = 2.93$	L4E-16 2.67 2.64 $\bar{x} = \frac{2.64}{2.65}$ $e_{c} = 2.68$	$\chi = \frac{2.62}{2.63}$ $\chi = 2.66$ $\chi = 2.66$
L4E-4 2.62 2.64 $\overline{X} = \frac{2.64}{2.63}$ $\mathcal{C}_{c} = 2.66$	L4E-17 2.66 2.67 $\overline{X} = \frac{2.69}{2.67}$ e = 2.70	$\frac{2.58}{X} = \frac{2.59}{2.59}$ $\mathbf{e}_{c} = 2.62$ L-104 2.62
L4E-5 2.59 2.63 2.65 $\overline{X} = \frac{2.85}{2.68}$ $\overline{x} = 2.71$	L-65 2.61 2.59 $\bar{X} = \frac{2.59}{2.60}$ $\boldsymbol{c} = 2.63$	$ \begin{array}{r} 2.68 \\ 2.61 \\ 2.68 \\ \underline{2.62} \\ \underline{2.62} \\ \underline{2.64} \\ \mathbf{e}_{c} = 2.67 \end{array} $

#### ORE PETROLOGY

The principal ore minerals of the Kirkland Lake camp ore are native gold and gold tellurides. The mineralogy of the ores has been discussed by Todd (1928) and Hawley (1948). With the help of X-ray analysis, Hawley (1948) was able to identify the rarer tellurides in the camp to account for a total of 21 minerals. These minerals are listed below in table 6.

#### Table 6.

Туре	Mode of occurrence	Mineral	Remarks In graphic intergrowths: primary constituents of syenites and por- phyry.	
MAGMATIC		Magnetite-ilmenite		
Hydrothermal	Elements	Gold Graphite	Alloyed with minor silver. Minor	
	Oxides	Specular hematite. Red dusty hematite. Magnetite	Rare.	
	Tellurides	Altaite, PbTe. Calaverite, AuTe <sub>2</sub> . Coloradoite, HgTe. Melonite, NiTe <sub>2</sub> Petzite, (Ag, Au) <sub>2</sub> Te. Tetradymite, Bi <sub>2</sub> Te <sub>2</sub> S. Joseite, Bi <sub>3</sub> TeS	Rare. Ag: Au=3:1. Locality unknown. Locality unknown.	
	Sulphides	Pyrite. Arsenopyrite. Chalcopyrite. Pyrrhotite. Molybdenite. Sphalerite. Galena.	Very rare. Very rare. Very rare. Very rare.	

(After Hawley, 1948)
In this thesis, at the 100 foot level of the Lake Shore mine, the following minerals were identified in 16 polished sections from the high-grade "L4E" vein :

> Mineral Occurrence native gold COMMON altaite (PbTe) COMMON calaverite (AuTez) minor coloradoite (MgTe) rare chalcopyrite COMMON pyrite major sphalerite rare molybdenite rare magnetite primary constituents in ilmenite porphyry syenite graphite COMMON

## Gold

Native gold occurs throughout the sections as minute grains, generally in microfractures and microscopic shear zones, and is of extremely fine grain size (0.02-0.09 mm). Gold within quartz gangue is rarely present as "free" gold, but is commonly intergrown with one or more tellurides. The telluride with which gold is most often associated is altaite. The complex intergrowths formed by these minerals can be described as follows:

- Small, rounded blebs, entirely within altaite or vice versa (plate 1).
- 2) Gold and altaite with interpenetrating amoeboid

boundaries (plate 1 and 2).

3) Course dendrites of gold within altaite, perhaps

filling and replacing altaite cleavage or grain boundaries (plate 3).

Due to the complexity of the intergrowths, it is difficult to ascertain paragenesis betweeen these two minerals. In the majority of cases, it appears that altaite replaces gold, but this relationship is not clear. Gold and altaite together, however, definitely replace pyrite, chalcopyrite, and sphalerite, and may replace coloradoite (plate 1).

Hawley (1948) mentions the presence of very fine gold as inclusions in pyrite. This type of gold occurrence was not observed in these sections; however, gold and altaite sometimes occur as a coating on grain boundaries of euhedral pyrite. Gold appears to replace the pyrite in some sections.

## Altaite

Altaite is the most abundant telluride in the

Plate 1: Intergrowths of altaite (A) and gold (G) replacing coloradoite (Co). Assemblage is 0.17 mm long.(PP). (700x).

Plate 2: Gold (G) and altaite (A) with interpenetrating amoeboid boundaries. Chalcopyrite (Ch) is a tan brown colour. Note the presence of calaverite (Ca) within altaite. (PP). (140x).





Kirkland Lake camp, (Hawley, 1948), and its presence is ubiquitous in all the sections investigated in this work. It was identified by its high reflectivity, bright bluishwhite colouration, and by its common occurrence with gold. Altaite appears to be the youngest mineral in these samples, replacing virtually all other metallic minerals. This texture is not unlike that reported for epithermal vein telluride ores from Colorado (Kelly and Goddard, 1969). Altaite occurs often as a thin coating around calaverite, both minerals having the same habit (plate 4).

Minute inclusions of altaite within euhedral pyrite grains (plate 5), and interstitial altaite particles between subhedral to euhedral pyrite grain boundaries (plate 5) are also noted. This texture suggests that pyrite recrystallization has occurred, enveloping the primary altaite, and incorporating the latter into the pyrite.

#### Calaverite

Todd (1928) reported that 17% of the gold ore mined in the Kirkland Lake camp comes from calaverite (AuTe<sub>2</sub>). It is distinguishible from altaite by its creamy pink colouration. When seen isolated, calaverite can be distinguished by its weak anisotropism which contrasts with the isotropic nature of the other minerals in this study. When calaverite is present, it is usually Plate 3: Gold (G) replacing altaite (A) along grain boundaries. Pyrite (P) is being replaced by altaite.(PP). (280x).

Plate 4: Rim of bluish-white altaite (A) enveloping pink calaverite (Ca). (PP). (110x).





Plate 5: Subhedral pyrite (P) with inclusions of altaite (A) and chalcopyrite (Ch), being plated by gold (G), altaite, and calaverite (Ca). (PP). (110x).

Plate 6: Chalcopyrite (Ch) being replaced by calaverite (Ca)
and altaite (A). Note the planar boundary between
chalcopyrite and calaverite, suggesting co-precipitation.
The small gray inclusion is coloradoite. (PP). (128x).





associated with altaite or altaite and gold. It is usually enveloped by the altaite (plate 4). Planar boundaries sometimes exist between chalcopyrite and calaverite (plate 6), suggesting coprecipitation; however, in other cases, scalloped embayments in chalcopyrite suggests replacement by calaverite. Calaverite is known to have a small amount of silver substituted for gold (Ramdohr, 1980); however, analysis by Todd (1928) shows that the calaverite present in the camp is relatively pure AuTe2. This suggests that during the deposition of the gold telluride, the vein solutions were free of silver.

#### Coloradoite

The presence of coloradoite (HgTe) in L4E vein samples is relatively rare, however its occurrence at the Lake Shore mine has been noted by Todd (1928) to be quite spectacular. It is identified in polished section by its dark gray to purple colouration and isotropic nature. Ramdohr (1980) suggests that coloradoite may be mistaken for petzite ((Au,Ag)=Te), if it were not that the latter is weakly anisotropic. Petzite was not seen in this study, which agrees with the work of Hawley (1948), where only one example of petzite was noted in the 60 sections studied.

The coloradoite was always found within altaite, altaite and gold (plate 1), or altaite and calaverite (plate 7). It appears to replace calaverite, and in turn to be replaced by gold and altaite. It is difficult to ascertain the paragenesis of coloradoite, as only a half dozen grains were observed, but its position in this paragenetic scheme is in agreement with that of Hawley (1948).

## Chalcopyrite

Chalcopyrite is observed in most slides as a minor constituent, although it may make up 20 to 30% of the total metallic mineral population. It occurs within microfractures in all of the sections and also as small blebs replacing gangue minerals. This association was thought by Todd (1928) to represent chalcopyrite precipitated either before or during ore deposition, and it is the only type identified in this study. The other association of chalcopyrite noted by Todd (1928) is its occurrence with galena, barite and calcite in post-ore faults.

Chalcopyrite appears to be a pre-telluride mineral, since it is veined and replaced by both the telluride assemblage and gold. This texture is observed in specimens which occur with massive gangue, and do not appear to be fracture controlled. Hawley (1948)

interprets the occurrence of chalcopyrite, gold and tellurides, when they occur together within fractures, as contemporaneous deposition. Another interpretation of these relationships is that post-ore faulting has redistributed these ductile minerals giving mechanically produced intergrowths, rather than intergrowths due to replacement by an aqueous medium.

Association of chalcopyrite and sphalerite is observed in a few samples, with sphalerite showing embayments, inclusions and veinlets of chalcopyrite. When pyrite and chalcopyrite occur together, the latter clearly replaces the former.

#### Sphalerite

The occurrence of sphalerite in the Kirkland Lake camp, according to Todd (1928) and Hawley (1948) is very rare. The former author noted only one grain of sphalerite embedded in pink calcite gangue unassociated with any other metallic mineral. Thus its position in the paragenetic scheme could not accurately be established.

Intergrowths of sphalerite with chalcopyrite, altaite, calaverite, coloradoite and gold were observed (plate 8) as well as intergrowths with chalcopyrite only. The sphalerite occurs only as blebs in the massive gangue, and never in microfractures. In each of the observed Plate 7: Coloradoite (Co) replacing calaverite (Ca), and enveloped by altaite (A). (PP). (510x).

Plate 8: Complex assemblage of altaite (A), calaverite (Ca), chalcopyrite (Ch), sphalerite (Sp) and gold (G). Sphalerite is veined by chalcopyrite and altaite. Note the inclusions of coloradoite along altaite-calaverite boundaries. (PP). (104x).





mineral intergrowths, the sphalerite is always veined or replaced by other minerals. Thus it appears that sphalerite is a relatively early mineral in the ore paragenesis at the Lake Shore Mine and probably in the whole Kirkland Lake camp. It is indeed remarkable that this relatively small-scale study (in comparison to the work of Todd (1928) and Hawley (1950)) uncovered a significant amount of sphalerite. It can be concluded from microscopic observation, that this ore zone at the Lake Shore Mine must be zinc-rich in comparison to the other veins studied in the camp.

# Pyrite

Pyrite is the most abundant sulfide reported in the Kirkland Lake ores (Hawley, 1950). It is concentrated along fracture zones and seams in the vein material and country rock.

Where the porphyry is extremely sheared and altered within the ore zone, pyrite may account for up to 50% of the rock. This type of pyrite is usually subhedral to anhedral and shows intense cataclastic deformation. Primary magmatic titaniferous magnetite shows extensive replacement by pyrite with the exsolved ilmenite lamellae displaying partial replacement by sulfide.

Within the vein quartz pyrite is concentrated along small scale shear zones and is associated with

molybdenite, chlorite and graphitic material. This pyrite is generally subhedral and of very fine grain size (< 0.3 mm). Locally, it may show rounded and corroded edges within altaite, indicating replacement by the latter.

Euhedral pyrite is most often found within massive quartz gangue but in lower abundances than the shear zone related pyrite. This euhedral pyrite is generally less than 0.5 mm in diameter, and frequently contains inclusions of altaite, gangue minerals and chalcopyrite (plate 5). Hawley (1950), also reports inclusions of exceptionally fine grained gold ranging in size from 0.5 to 10 microns. These inclusions can be explained in two ways:

- Replacement textures along minor aberrations in the pyrite structure in the third dimension.
- 2. As impurities incorporated into the pyrite lattice during crystal growth.

Some sections show clear replacement of pyrite by chalcopyrite , gold and altaite (plate 9), and thus provide evidence for the first explanation. No inclusions of gold within pyrite were observed in this study, although gold and altaite often occur as films plating the euhedral pyrite (Plate 5). From this evidence, pyrite is thought to be an early mineral in the paragenesis of the

Plate 9: Pyrite (P) showing subrounded edges due to replacement by altaite (A) and gold (G). (PP). (160x).

Plate 10: Molybdenite (gray) near microshear zone with anhedral pyrite (P). (PP). (180x).



Kirkland Lake ores; however, pyrite crystallization at some later stage is also supported by inclusions of altaite and chalcopyrite.

#### Molybdenite

Molybdenite (associated with chlorite and graphite), occurs as a bluish-black coating on major slip surfaces. It usually shows slickenslides, sometimes with an extremely thin film of gold smeared out along the molybdenite surface. In other examples, pulverized pyrite grains are strewn out along the striae.

Polished sections with molybdenite (plate 10) show very minor amounts (< 0.1 mm wide) along micro-shears that include cataclastic pyrite. Analysis by Todd (1928) of the molybdenite film shows 90% sericite and chlorite, 6.5% molybdenite and 3.5% pyrite.

## Graphite

The presence of carbonaceous material in the form of graphite is documented in many gold deposits (Wilson , 1985: Boyle, 1979). At the Lake Shore Mine, ( Lovell et al., (1984); this study) and throughout the Kirkland Lake camp (Kerrich and Watson, 1984; Hawley, 1950) the presence of graphite within quartz veins produces a black to gray colouration in some of the vein material. Other milkywhite guartz veins have ribbon-like patterns of graphitic material. perhaps reflecting periods of cvclical deposition of graphite plus quartz and quartz. Another example of this texture is observed in fractured wallrock, with graphite in quartz lining one side of the vein filling, and pure quartz making up the remainder of the vein material.

## Paragenesis

When discussing the paragenesis of an ore deposit which has been spatially oriented by fault controlled processes, and subsequently deformed by syn-and postdepositional movement, one must look to the textural which evidence is least affected by mechanical remobilization of metallic minerals. The paragenetic scheme developed in this study is derived from textural relationships between minerals occurring within massive quartz gangue which shows little evidence of shearing. It must be stated, however, that the ductile nature of gold, molybdenite, chalcopyrite, and all the tellurides could conceivably produce intergrowths caused by physical processes, which do not reflect their true depositional relationships.

Taking the above into consideration, the paragenesis of the ore minerals within the Lake Shore Mine is;

QUARTZ

GRAPHITE

PYRITE------

SPHALERITE

CHALCOPYRITE

CALAVERITE

COLORADOITE

⊢ GOLD----

This paragenetic scheme is in close agreement with that of Hawley (1950), with the exception of sphalerite. The position of sphalerite was changed because in the previous studies, sphalerite was not observed in conjunction with any other metallic mineral, as is seen here.

# HYDROTHERMAL ALTERATION

## Introduction

Wallrock alteration by hydrothermal fluids has been studied in many ore environments from porphyry copper deposits (Beancand Titley, 1981; Meyer and Hemley, 1967), and massive sulfide deposits (Riverin and Hodgson, 1980: Franklin et al., 1975), to Archean lode gold deposits (Robert and Brown, 1984; Kerrich and Watson, 1984). The process of hydrothermal alteration is one in which a mineral assemblage undergoes chemical changes in order to reequilibrate with a fluid-rich phase, which is usually at a different temperature than the host rock. This type of alteration involves the introduction of an aqueous phase elevated temperatures thus the at and term "hydrothermal".

Hydrothermal alteration, as it applies to lode gold deposits within various lithologies of Archean age, is discussed by Colvine et al. (1984), Kerrich et al. (1982) and Boyle (1979). Table 7 (Colvine et el., 1984), shows typical alteration assemblages for lithologies of intermediate to felsic composition, including the symmite intrusives in the Kirkland Lake camp. Investigations on the processes and patterns of hydrothermal alteration within the camp at this point in time are limited to

# <u>Table 7</u>.

ALTERATION ASSEMBLAGES IN INTERMEDIATE TO FELSIC LITHOLOGIES (selected examples)								
CAMP /DEPOSIT AND LOCATION	ROCK TYPE	DIAGNOSTIC Background	ALTERATION PHASES Ore Zone	REFERENCES				
		**************	***************************************	***********************************				
TIMMINS Ontario, Canada	Q¥Р	albite quartz white mica chlorite calcite anhydrite hematite pyrite	white mica dolomite pyrite quartz albite	Davies and Luhta, 1978 Gorman et al., 1981 Karvinen, 1981 Bain, 1933				
RED LAKE (Howey-Hasaga) Ontario, Canada	ÚŁ6	albite quartz white mica carbonate	white mica albite quartz	Horwood, 1945				
CROW RIVER (Pickle Crow)	QFP	albite quartz white mica chlorite	quartz albite calcite pyrrhotite pyrite	Рүе, 1976				
GERALDTON Ontario, Canada	58.b	aibite quartz white mica chlorite catbonate	white mica carbonale chlorite quartz pyrite	Horwood and Pye, 1955				
KIRKLAND LAKE Ontario, Canada	Syenite	chlorite dolomite calcite white mica	white mich chlorite pyrite chematite	Thomson, 1948 Ploeger, 1980 Watson and Kerrich, 1983				
CANADIAN ARROW (Abitibi belt) Ontario, Conada	Monzonite	chlorite calcite pyrite white mica hematite	pyrite hematite	Cherry, 1983				
RED LAKE Ontario, Canada	Granodiorite	chlorite epidote calcite	white mica quartz calcite chlorite hematite	Horwood, 1945 Ferguson and Watt, 1965 Lakind, 1984				

Notes: (1) All the above examples occur within host rocks of greenschist grade, (2) Only those miniral phases considered to be diagnostic have usen listed. (3) Descriptions of mesoveric and solicite are disted as "white minia". (4) Quartz listed above occurs as a component of rock replacement. All of the above examples contain quartz verus in ore zones. (5) QPF = intermediate to felsic quartz and/or feldspar porpnyry bodies.

,

(Colvine et al., 1984).

Kerrich et al. (1984) at the Macassa Mine, Thomson et al. (1948) throughout the Kirkland Lake camp, and this study at the Lake Shore Mine. Igneous and Alteration Petrology

Petrographic descriptions were made of mineralogical trends from samples taken within the L4E ore zone and from the L-series core drilled approximately perpendicular to the north vein system (see figure 3). From these samples, 22 thin sections were investigated, and the mineral trends discerned were used to establish alteration facies assemblages (fig. 4). The discussion of the mineralogical facies is presented in order from the least altered "parent" syenite porphyry, to its most altered product within the L4E vein. All the samples show some degree of alteration. but the least altered rock in this study is sample (L-65) which is farthest away from the vein system. As the vein system is approached, the alteration intensity increases.

#### Hand Sample Scale

At a hand specimen scale, the least altered porphyry contains abundant feldspar grains. These range up to 10 mm in diameter but average around 2 mm, and are set in a fine grained quartz-feldspar matrix. The white porphyritic feldspar is easily distinguished from the maroon-gray matrix material, as are the subrounded inclusions of mafic country rock (<3 cm diameter) noted in some samples.

As the vein system is approached, the colour of the syenite changes gradually to a reddish-purple, then to a reddish-green, and finally to a buff-green. The rocks showing the highest degree of alteration are gray-green and highly deformed.

The red colouration is caused by the presence of hematite produced by the breakdown of biotite to chlorite in the presence of a hydrothermal fluid, which liberates and oxidizes some of the iron from the biotite (Smith, 1974), producing extremely fine-grained hematite. The green colouration is probably due to the preponderance of white mica in the rock. Samples that are at this stage of alteration lose their porphyritic texture, as the colour of the phenocrysts becomes indistinguishible from that of the groundmass. The intensity of fractures and chloriteslip features increases towards the ore zone, and reaches a maximum next to the vein. Away from the ore zone, these fractures contain little quartz-carbonate material.

Microscopic Scale

Least Altered Rock (L-65)

Before chosing a least altered rock for mass balance calculations, the samples were studied in thin section for

textures and mineral assemblages that would most closely resemble those of a fresh syenite. Sample L-65 was chosen as the least altered rock because:

- It is the only rock to have magmatic biotite as a constituent, although the biotite is partly altered to chlorite.
- The degree of seritization, carbonitization, and microfracturing is minimal compared with the rest of the suite.

The prominent mineral constituents of this sample are sodic plagioclase, orthoclase, and biotite-chlorite as porphyritic minerals, with quartz, feldspar, and carbonate as matrix components. Carlsbad twinning is common in the orthoclase, while albite, pericline and penetration twinning is ubiquitous throughout the plagioclase. The feldspars sometimes show zoning.

Accessory minerals are zircon, rutile, titanomagnetite, hematite, and sericite. Most of the zircons present show well defined crystal habit and some display zoning (plate 11C), The maximum size of the zircons is 0.5 mm. Rutile occurs as acicular reticulated inclusions within the chlorite-altered biotite (plate 12A ). This three-dimensional array of rutile is referred to as sagenitic texture. Iron opaques, presumed to be titanomagnetite, are seen throughout the slide as elongated inclusions in altered biotite (plate 12B), and in the groundmass as blade-like crystals associated with chlorite. Hematite occurs near the biotite/chlorite assemblages as minute red grains. Sericite (fine-grained white mica) is ubiquitous in all the feldspars as a secondary phase.

# ALTERATION FACIES

In order to simplify the mineralogical trends in this hydrothermal environment, an alteration facies sequence, based on the predominant mineral assemblages, is presented in figure 4. Much of the terminology used in discussing hydrothermal alteration assemblages stems from the early work in porphyry copper environments (Sales and Meyer, 1948). This nomenclature is also applicable to hydrothermal alteration in lode gold deposits, but in the interest of simplification, the alteration facies will be categorized according to the characteristic mineral assemblage of that facies.

In this study there are four distinct alteration zones:

1) magmatic biotite

2) chlorite

3) phlogopite-penninite-hematite

Distance from Ore Zone(meters)	• <b>8</b> 0	•7.0	+ 5.0	•4.0	+1.5	
Sample Locations	L65	L69	L76	L79	L 87	
biotite chlorite		<b></b>				
rutile hematite						NE .
phlogopite						RE ZO
pyrite penninite						-
secondary orthoclase	•			1 1 60		-
FACIES	magmatic biotite	chlorite		phlogopit( penninite	hematite  secondary or thoclas	

Figure 4. Alteration Assemblages and Facies.

4) sericite-orthoclase-pyrite.

The intensity of hydrothermal alteration increases from zone one through four, and the relative distances of the assemblages from the ore zone can be seen in figure 4. The spatial extent and relative proportion of the alteration products are shown in figure 5.

The micas are perhaps the most useful mineral group on which to study the effects of hydrothermal alteration. They are common rock-forming minerals, have a relatively open framework structure, enabling them to undergo fluidmineral reactions, and have a broad chemical composition that is susceptible to exchange with fluids of varying composition and temperature. The geochemistry of biotite has been studied in porphyry copper environments by Jacobs and Parry (1976;1979) and Larier et al. (1978), and in other hydrothermal environments by Munoz (1984), Ferry (1979) and Parry and Downey (1982).

The susceptibility of the mica group to hydrothermal alteration is displayed in the alteration facies listed earlier. Figure 4 shows the trend from magmatic biotite; to chlorite; to a secondary biotite with a phlogopitic nature, and finally to a sericite as alteration intensity increases. The changes in these alteration facies may provide information on the nature of the aqueous medium responsible for the hydrothermal alteration adjacent to

Distance from Ore Zone(meters)	•8.0	•7.0		•5.0	+4.0		+1,5		
Sample Locations	L65	L69		L76	L79		L87		
microfracturing carbonatization sericitization hematization pyritization silicification albitization sagenite									ORE ZONE
FACIES	magmatic biotite	       	chlorite		phlogopite-	penninite - hematite		orthoclase- pyrite	

Figure 5. Semi-quantitative Distribution of Alteration Products.

the ore zone. If these fluids are linked to gold mineralization in any way, mineralogical and geochemical changes in the altered zone may be useful as an exploration tool in hydrothermal precious metal environments.

## MAGMATIC BIOTITE FACIES

This facies is characterized by the presence of biotite with igneous character. Magmatic biotite an is distinctive in its physical nature from all other micas studied here. It is pleochroic from brown to colourless, rutile-free, and generally in the process of being altered to chlorite. Chloritization of biotite is common (Speer, 1984) and generally occurs in the presence of an aqueous phase. Plate 11A and 11B show magmatic biotite (high birefringence) almost completely replaced by chlorite (very low birefringence). The process of chloritization appears to take place along the OO1 planes, with the exterior of the biotite phenocryst being replaced first. Basal sections show complete chloritization.

Partly chloritized biotite has inclusions of threedimensional acicular rutile (sagenite), within the chlorite layers. The development of sagenite (Plate 12A ) reflects the inability of chlorite to incorporate titanium

# Plate 11 MAGMATIC BIOTITE FACIES

Plate 11A

Chlorite (low birefringence) replacing magmatic biotite (high birefringence) along OO1 plane. Plagioclase is partially sericitized. (63 x). (XP).

Plate 11B

Chlorite (C) replacing magmatic biotite (B). (125x). (XP).

Plate 11C

Euhedral zircon with zoning. Note the deterioration on the right side of the crystal suggesting alteration. Zircon is 0.5 mm long. (PP).



into its structural lattice (Speer, 1984). The process of alteration of titaniferous biotite, in the presence of a fluid phase, to chlorite and rutile can be depicted as follows:

Equation 1. 3KaMgeTiaSisAlaOao + 16H+ + 12HaO

4

 $\frac{Mg_{10}A1_{4}Si_{6}O_{20}(OH)_{16} + 6TiO_{2} + 6K^{+} + 2A1^{3+} + 2Mg^{3+}}{+ 9H_{4}SiO_{4}}$ 

Other possible reaction mechanisms for the chloritization of biotite can be found in Ferry (1979) and Parry et al. (1982). Microscopic evidence suggests that no change in volume accompanies this chloritization process, although small inclusions of quartz and/or potassium feldspar are noted in some chlorite grains, probably as daughter products of the chloritization process.

The transition from biotite to chlorite, in the presence of water, causes the oxidation of some ferrous iron, which gives rise to hematite formation, as follows:

Equation 2.  $2Fe^+ + 3H_{20} \rightleftharpoons Fe_{203} + H_{2} + 4H^+$ 

Although hematite is not abundant in this facies, it is

noted in greater proportions around biotite which is partly altered to chlorite.

## CHLORITE FACIES

This facies represents the total replacement of magmatic biotite by chlorite. Hematite is more abundant than in the previous facies, but is still a relatively minor phase. Elongate inclusions of titano-magnetite and sagenite occur along and parallel to the OO1 plane (plate 12B).

Calcium and magnesium carbonates commonly occur within and around chlorite. Siderite is present around grains of titano-magnetite. These carbonates are secondary phases, produced by the reaction of a  $CO_2 - H_2O$ fluid phase with Mg<sup>2+</sup> from the chloritization process,  $Ca^{2+}$  from the sericitization of plagioclase, and Fe<sup>2+</sup> from biotite or titano-magnetite.

Plagioclase and orthoclase both show increased sericitization relative to the magmatic biotite facies. Some plagioclase grains have altered to sericite + carbonate. Other feldspars, however; have clear optically oriented rims around sericitized interiors. This texture can be explained in two ways:

1) Preferential sericitization of the interior of the
#### Plate 12 CHLORITE FACIES

Plate 12A Basal section of chlorite with inclusions of acicular rutile (sagenite). (125x). (PP).

Plate 12B

Hydrothermal chlorite with inclusions of titanomagnetite parallel to the cleavage direction. (63 x). (PP).

Plate 12C

Sericitized feldspar core with clear rims of optically oriented albite. (63x). (XP).



feldspar phenocryst.

2) Growth of secondary feldspar around partially

sericitizated feldspar grains (plate 12C). The later mechanism (albitization) is a common texture in porpyhry copper and epithermal environments (Boyle, 1979).

The best evidence for albitization over preferential sericitization occurs when a microfracture cuts a plagioclase grain. The fracture is filled with guartzcarbonate material. The plagioclase is moderately sericitized away from the microfracture, but close to it, the plagioclase is clear, and in perfect twin continuity with the rest of the feldspar phenocryst. This indicates that hydrothermal fluids passed through the microfracture, carrying Na+, H+, H4SiO4 and Al3+ which reacted with the anorthite component of the plagioclase to produce Ca<sup>2+</sup>, which in turn reacted with the CO<sub>2</sub> in the fluid to produce calcite.

## PHLOGOPITE-PENNINITE-HEMATITE FACIES

This facies is characterized by the alteration of chlorite to phlogopite, the development of penninite and the widespread and dense distribution of hematite throughout the samples.

The development of phlogopite in hydrothermally altered environments has been noted by Lanier et al.

# Plate 13 PHLOGOPITE - PENNINITE - HEMATITE FACIES

Plate 13A

Phlogopite with attenuated sagenite (compare to plate 2). Black granules in groundmass are hematite. (125 x). (PP).

Plate 13B Same as A under crossed polars. (125 x). (XP).

Plate 13C

Porphyritic plagioclase showing calcite replacement along microfractures. (63 x). (XP).



(1978), Jacobs and Parry (1976;1979). This secondary biotite is physically and geochemically distinct from the primary magmatic biotite (Speer, 1984, this study). In this facies it is slightly pleochroic from yellow to greenish-blue, and has a lower refractive index than the magmatic biotite, while still retaining the "bird's-eye" extinction characteristic of the biotite group. To establish geochemical differences, an electron microprobe study would need to be conducted.

The development of phlogopite from chlorite occurs in much the same way as chlorite from magmatic biotite. Phlogopite replaces chlorite along the OO1 planes, the exterior of the chlorite being replaced first. The reaction of chlorite alteration to phlogopite can be written using the magnesium end members (Parry and Downey, 1982):

Equation 3. MgzAlzSisOie(OH) = + 1.4K<sup>+</sup> + 2H<sup>+</sup> + 1.2HzSiO4

1.4KMgæAlSiæOı₀(OH)₂ + 0.8Mg<sup>æ+</sup> + 0.6Al<sup>∞+</sup> + 1.2H<sub>æ</sub>O

5

This reaction requires the addition of  $K^+$ ,  $H^+$  and water.

The hydrothermal fluid responsible for this alteration must, therefore, have a higher  $K^+/H^+$  ratio than the fluid causing the alteration of magmatic biotite.

Sagnetic texture from the chlorite is preserved in the phlogopite but the rutile asters (Deer, Howie and Zussman, 1966) are considerably attenuated (plate 13A and 13B).

Penninite ((Mg11Al)(Si7Al)Ozo(OH)14), distinguished by its anomalous "berlin-blue" birefringence, occurs as a major constituent of slip features that occur throughout this facies, and as inclusions within the feldspars. The conversion of plagioclase to chlorite by the introduction of Mg<sup>2+</sup>, H<sup>+</sup> and water is documented by Smith (1974) as follows:

Equation 4. 2CaAl 2Si 20e + 5Mg2+ + 2H20 + 4H+

 $Mg = A1Si = A10 = (0H) = + 2Ca^{2+} + 2A1^{3+} + Si^{4+}$ 

The Mg<sup>2+</sup> ions needed for this reaction may have been supplied by the phlogopitization of chlorite (equation 3). The Ca<sup>2+</sup> produced in equation 4 explains the abundance of calcite within the plagioclase while the Al<sup>3+</sup> and Si<sup>4+</sup> components are utilized in the production of sericite and

kaolinite, which are major constituents of this facies.

The process of hematization reaches a maximum within this facies and produces a red colouration in all the samples. Not only does hematite occur in the groundmass near mafic minerals, but throughout the section, and even within the feldspars where a red clouding is seen. The hematite within the feldspars generally consists of granules less than 0.01 mm in diameter, identified by their very high relief and ruby red colouration. The process of hematization via hydrothermal fluids, and the nature of red clouded feldspars is fully discussed by Boone (1969).

The advanced development of hematite throughout this facies suggests that the process of oxidizing ferrous iron (equation 2) to ferric iron has increased towards the ore zone. The oxidation potential of the hydrothermal fluids has increased from the magmatic biotite to the phlogopite- penninite-hematite facies.

### SERICITE-ORTHOCLASE-PYRITE FACIES

This mineral assemblage occurs adjacent to the ore zone and reflects the highest intensity of hydrothermal alteration. This is supported by the highly fractured nature of the rock (plate 14A and 14C), the appearance of

# Plate 14 SECONDARY ORTHOCLASE - PYRITE FACIES

Plate 14A

Highly fractured, sericitized and carbonatized section, within one meter of the ore vein. Pyrite (P) with inclusion of white mica. (63 x). (XP).

Plate 14B

Metasomatic pyrite (P) with inclusions of white mica near quartz-carbonate vein. Quartz (Q) fills a pressure shadow adjacent to pyrite. (63 x). (XP).

Plate 14C

Secondary "clear" alkali feldspar (AF) internally replacing sericitized plagioclase (P1) along fracture. Matrix is highly fractured, carbonatized and sericitized. ( 63x). (XP).



metasomatic pyrite with quartz pressure shadows (plate 14A and 14B), and the abundance of quartz-carbonate veinlets (plate 14A and 14B).

The plagioclase phenocrysts within this facies are almost totally replaced by sericite. In some sections, deformation has been so intense that no phenocrysts remain. Secondary orthoclase is present in the less deformed samples as subhedral to euhedral phenocrysts with inclusions of sericite. It is not clear whether this secondary orthoclase grew around and included existing sericite, or if a later hydrothermal event caused the sericitization of the metasomatic orthoclase. What is suggested by the presence of this K-feldspar is that the K+/H+ activity of the hydrothermal fluid was high enough to stabilize the formation of KAlSisOs in the presence of sericite (see figure 6).

Pyritization of the rock in this facies is developed at the expense of hematite. Samples with pyrite contain little hematite and rocks within this facies are generally a green to buff-green colour. This suggests that the hydrothermal fluids had a reduced sulfur component. The presence of pyrite, however; is limited to a short distance from the vein material (generally < 1m), and is not a major constituent (usually < 20% of the rock alteration assemblage). This indicates that although the

fluids may have carried reduced sulfur, this element was a relatively minor component. If not, more wallrock pyrite would be present. Metasomatic pyrite with inclusions of sericite are frequently seen in this facies (plate 14A and 14B). Pressure shadows of quartz often occur along the edge of the pyrite.

Phlogopite is present in this assemblage, but only as a minor constituent, and often in a highly sericitized form. There are virtually no inclusions of sagenite remaining in the phlogopite.

Penninite is present with sericite along the many slip surfaces throughout this facies.

#### DISCUSSION

It is hoped that by studying the change in mineral assemblages with distance from the ore zone one can interpret the physical and chemical nature of the ore fluid. Information accumulated in this work suggests that the hydrothermal fluids were continuously reacting and evolving as they penetrated the country rock. An important question to be considered is whether the fluid that caused the hydrothermal alteration of the country rock was the same as that which precipitated the gold. hydrothermal alteration of the country rock is The

probably associated with fluids that precipitated the bulk of the quartz vein material. From the ore petrology, gold is clearly one of the later phases introduced into the system. and is definitely later than the quartz. Rather than speculate about the existence of two separate precipitating silica fluids. one and the other precipitating gold, it is perhaps more logical to envision one fluid which evolves and cools through time to produce the above mineral associations. The higher temperature fluids may have been responsible for wallrock alteration with gold being deposited as this fluid cooled. The cooler fluids that precipitated gold could not generate any further wallrock alteration.

Characteristics of the Hydrothermal Fluid from Mineral Assemblages

Alteration mineral assemblages provide information about the hydrothermal fluid responsible for the alteration. The presence of sphene and epidote as alteration products of biotite and plagioclase has been noted in hydrothermal systems by Ferry (1979), Kerrich and Fyfe (1981), Robert and Brown (1984), and Parry and Downey (1982). The absence of epidote and sphene in this study indicates that the fluid has a significant COz Kerrich and Fyfe (1981) suggest the following content. carbonation reactions for the breakdown of the calcium silicates:

Equation 5. CaTiSiOs + COz = CaCOs + TiOz + SiOz sphene

Equation 6. 3(Mg,Fe) + Al + SizO10(DH) + 6CazAl = SizO12(DH) aluminous chlorite epidote

+  $SiO_2$  + 24CO<sub>2</sub> + 10K+

**ک** 

10KAlaSiaOio(OH) = + 12Ca(Mg,Fe)(COs) = + 10H+

The absence of sphene and epidote in this study suggests that the above reactions have gone to completion. Both reactions occur in the presence of  $CO_2$  but no quantitative measurement of the Xco<sub>2</sub> of the fluid can be derived from this work.

Fluid inclusion studies from the McIntyre-Hollinger Gold Deposit in Timmins, Ontario (Smith et al., 1984), have an Xco2 between 3 to 24 mole percent. Kerrich and Fyfe (1981) suggest CO2/H2O ratios between 0.2 to 0.5 in hydrothermal lode deposits at the Con Mine, N.W.T., and implicate that Au, carried as a carbonyl (CO) or carbonate (COs) complex, may precipitate out of solution during fluid-wallrock interaction, causing carbonatization of the wallrock. Since carbonate alteration adjacent to the ore zone is noted in this study, it is possible that this same process occurred at the Lake Shore Mine. To further investigate this carbonate-gold relationship, a fluid inclusion study of the quartz vein and alteration minerals in the wallrock is necessary.

Figure 6A and B (Parry and Downey, 1982), show equilibrium stability relations for the  $K_{20}$ -MgO-Al\_2O<sub>3</sub>-H<sub>2</sub>O system at 100 bars and 200°C. Although oxygen isotope studies from Macassa (Kerrich and Watson, 1984), suggest fluid temperature between 380 °C to 490 °C, the relationships shown on figure 6 will still be applicable.

Figure 6 . Calculated stability relations of chlorite, phlogopite, muscovite and K-feldspar at 200<sup>°</sup>C and 100 bars for volume conservative reactions (A) abd Al-conservative reactions (B). (Parry and Downey, 1982).



Guantitative changes may occur, but theoretically, mineral relationships with respect to K<sup>+</sup> and Mg<sup>2+</sup> will remain the same under different temperature conditions. Referring to these diagrams, and recalling the alteration assemblages, one can derive information about the ak/am<sup>+</sup> and am<sup>2</sup>/am<sup>+</sup> of the hydrothermal fluid with increasing alteration intensity and proximity to the ore zone. For example, in the phlogopitic facies, chlorite is altered to phlogopite (increasing ak/am<sup>+</sup>), whereas in the facies closest to the ore zone, phlopopite alters to sericite and K-feldspar precipitates (increasing ak/am<sup>+</sup> and decreasing af<sup>2</sup>/am<sup>+</sup>). The ak/am<sup>+</sup> is lowest farthest away from the ore zone, where magmatic biotite alters to chlorite.

From the alteration facies, it is possible to deduce the following about the character of the hydrothermal fluid as it permeated the wallrock:

- 1) decreasing ak/am\*
- 2) increasing  $a_{H_{a}}^{2}/a_{H_{a}}^{2}$
- 3) reducing and relatively S-poor with first wallrock

reaction, then more oxidizing with further permeation.

4) increasing and/ant

5) highly CO2 charged.

#### GEOCHEMISTRY

Lithogeochemistry of the Least Altered Syenite Porphyry

The elemental geochemistry of the least altered sample (L-65) is given in table 4. Barium contents of 4000 ppm suggest substantial substitution of Ba for K in the K-feldspars. This corresponds to a 14% BaAlzSi $_2O_{e}$  molecule in the orthoclase, which classifies the feldspars as hyalophanitic (Deer, Howie and Zussman, 1966). Some Ba will substitute for K in the magmatic biotite, but since most of the biotite has undergone partial alteration to chlorite, the preferred structural site for Ba is in the K-feldspar.

Similarily, Sr concentrations of 2000 ppm suggest a significant degree of Sr substitution in the Ca lattice site in the anorthite molecule of the plagioclase.

The presence of large euhedral zircons in sample L-65 explains the relatively high zirconium content (370 ppm).

These three trace elements are instrumental to a discussion of geochemical trends with hydrothermal alteration, and also provide information as to the chemical nature of the hydrothermal fluid.

Geochemistry of Hydrothermal Alteration Introduction

Variations in geochemical trends with alteration intensity were studied using the "L"-series core samples and "L4E" vein samples. Absolute gains and losses of elements were calculated using variations of Gresens (1967) mass balance equations. Constant volume was assumed for the L-series samples and confirmed by petrology. Volume factors for the L4E-series were calculated using isochemical parameters where constant volume conditions did not apply.

Geochemical trends in the L-series samples compare well with the L4E-series, and therefore conclusions concerning the chemical nature of the fluid and its interaction with the wallrock can be made.

"L"-Series - Geochemical Trends Versus Distance from Ore Zone

The L-series samples were used to examine the variations in geochemistry, due to hydrothermal alteration, versus the distance from the ore zone. Sample L-65, (least altered rock) was taken 8 meters from the ore system. The chemical parameters and specific gravity of this sample were taken to be representative of the parent

rock in Gresens (1967) mass balance equation.

Although some of the samples showed microfracturing, volume changes, such as those due to quartz-carbonate veinlets, were generally absent. Therefore, constant volume was assumed when calculating absolute gains and losses of elements. The ratio of the volume of parent and product rocks (volume factor) is equal to one in calculations of this nature (Appendix 2).

Figures 7A to X show gains or losses in weight percent, versus distance from the ore zone. Alteration intensity increases as the ore zone is approached, and therefore the gains or losses of the chemical constituents of the wallrock can be attributed to hydrothermal alteration. The following trends are seen in figure 7A to X.

Figure 7A) Increase of SiO2 as vein system is approached.

- B) Fluctuating trends of Al<sub>2</sub>O<sub>3</sub> with distance.
- C) Loss of Fe<sub>2</sub>O<sub>3</sub> from 8 meters to 4 meters from the ore zone. Increase in Fe<sub>2</sub>O<sub>3</sub> from 4 meters to zero meters from the ore zone.
- D) Gain of MgO away from the ore zone, and decrease as ore zone is approached.
- E) Loss of CaO on approach to ore zone.
- F) Gain of Na<sub>2</sub>O from 5 meters to 7 meters, and

Figure 7.



Distance from Ore Zone (m)







Distance from Ore Zone (m)

then significant loss as ore zone is approached.

- G) General increase in K<sub>2</sub>O as ore zone is approached.
- H) Minor losses of TiO2 in altered rock.
- I) Minor gains of MnO as ore zone is approached.
- J) Significant loss of P2Os with increasing alteration intensity.
- K) Major losses of Ba with increasing alteration intensity.
- L) Fluctuating variations in Ce with distance.
- M) Fluctuating variations in Nd with distance.
- N) General loss of La in altered rocks.
- 0) No significant loss or gain of Nb.
- P) Minor loss of Y with increasing proximity to ore zone.
- Q) Major loss of Zr with increasing alteration intensity.
- R) Moderate gain of Rb with increasing alteration intensity.
- S) Major loss of Sr with greater proximity to ore zone.
- T) Fluctuating variations in W with distance.
- U) Moderate gain of Sb with increased alteration intensity.



Distance from Ore Zone (m)

Figure 7.





Figure 7.

















Figure 7.





Figure 7.

-9



Distance from Ore Zone (m)

-3

-1

1

-5





- V) General loss of As in altered rocks.
- W) Significant gain of Au with increasing alteration intensity.
- X) Significant gain of volatiles with alteration.

Elemental Gains and Losses - Discussion

It can be concluded fom figure 7A to X that hydrothermal fluids introduced SiOz, KzO, MnO, Rb, Sb, Au and volatiles (COz and HzO) to the system.

The addition of SiOz as an H4SiO4 component of the hydrothermal fluid is expected since appreciable volumes of quartz fills the veins. The addition of K<sub>2</sub>O indicates that the fluid had a high K<sup>+</sup> content. which is substantiated by the petrography. The addition of MnO is relatively minor and may reflect substitution of Mn in phlogopite. Gains of Rb suggest that the fluid had a high Rb<sup>+</sup> concentration, which correlates well with high K<sup>+</sup> The net gain of Sb and Au indicates that the contents. hydrothermal solutions carried substantial amounts of each element, although Sb has permeated the country rock to a greater extent than Au. The increase in volatile content is generally less than 2 weight percent as the ore zone is approached. At a molecular scale this increase becomes substantial due to the light atomic weight of  $H_{2}O$  and  $CO_{2}$ 

relative to the other oxides. This hydration of the wallrock confirms that an aqueous phase was responsible for the alteration.

FeaOs, CaO, NazO, TiOz, PzOs, Ba, La, Zr, Sr and As show net losses due to alteration. The loss of iron from the wallrock is compatible with the destruction of the mafic phases biotite and chlorite. Some of the iron is taken up in hematite (or oxidizes to form hematite). Excess ferrous iron may be transported into the phlogopite-penninite-hematite facies, where it becomes oxidized, or is precipitated in pyrite by the sulfur phase of the fluid. The loss of CaO, La, and Sr are probably related to the breakdown of plagioclase, releasing  $Ca^{2+}$ ,  $La^{3+}$  and  $Sr^{2+}$  into the fluid. Some of these would be taken up in carbonation reactions, but it appears that most of the La<sup>3+</sup>,  $Sr^{2+}$ , and some of the Ca<sup>2+</sup> ions were leached from the wallrock. Sodium is gained from the ore zone into the chlorite facies which is probably related to the albitization process, however; substantial losses of NagO occur as alteration intensity increases. The activity of Na\* in the hydrothermal fluid therefore, increases with distance from the ore zone, which is substantiated by petrographic evidence. The losses recorded for TiOz suggest that the use of TiOz as an immobile element (isochemical mass balance equations) may
not always be valid when dealing with hydrothermal alteration. The losses of TiO2 here are, however, relatively insignificant. P=Os is definitely lost to the hydrothermal fluids due to its consistent and marked decrease with increasing proximity to the ore zone. Ba shows substantial losses with increasing alteration This reflects the loss of Ba+ intensity. during the hydration of K-feldspar and magmatic biotite. It does not appear that Ba is taken into phlogopite. The phenomenon of Ba-poor hydrothermal biotite has been noted in porphyry copper deposits (Jacobs and Parry, 1979). Arsenic is removed from the wallrock during alteration, implying that the fluid was As-poor. As arsenopyrite is an extremely rare mineral in the Kirkland Lake ores (Hawley, 1948) the petrography supports the As depletion. Zr loss with increasing alteration intensity is particularily interesting since Zr (like TiO2) is often taken to be an immobile element when computing isochemical mass balance equations. In this study, zirconium is particularily Zircons are known to be resistant to normal mobile. chemical attack (Deer, Howie and Zussman, 1966), but may be altered in the presence of an alkali-rich fluid phase (R.P. Taylor, pers. comm.) at elevated temperatures. Petrographic evidence indicative of zircon instability

was found in this study: large euhedral zircons which have undergone some alteration (plate 11C), are seen in the magmatic biotite facies. The zircons present in the zones of higher alteration intensity are all of smaller size and are characteristically subhedral to anhedral in shape. Because the altered rocks show evidence of sodium and potassium metasomatism, the presence of alkalis in the hydrothermal fluid may have caused the destabilization of zircon, and subsequently the loss of zirconium to the fluid.

"L4E"-Series - Geochemical Trends Versus Volume Factor

The L4E samples were taken from within the ore zone, and therefore are the most highly altered rocks of the study. Because deformation within the ore zone is intense. with ubiguitous guartz-carbonate veinlets, the To assumption of constant volume is no longer valid. utilize mass balance equations in this instance, a volume factor must be calculated using an isochemical parameter. The latter is derived from the least "mobile" elements in the system (ie. those least affected by hydrothermal alteration). Elements often assumed to be immobile are; AlaOs, Sc, Zr (Kerrich and Fyfe, 1981), combinations of Al**20**3, SiO2, TiO2 (Davies and Luhta, 1978), and Al2O3/TiO2 (Gresens, 1967).

In this study there are no totally immobile elements, however; to distinguish general trends of net gains and losses within the L4E samples, to a first approximation, a weighted mean volume factor was calculated using SiOz, Al**20s.** TiO2 and Nb (table 8). These components are known from the L-series plots (figure 7A to X) to be the least mobile parameters of the system. Figure 7A shows that SiOz increases with alteration intensity and therefore is mobile. This is expected since the vein material is quartz, emplaced by silica remobilization. SiOz was utilized in the volume factor calculations since the ratio of weight % SiOz gained / original weight % SiOz is small compared to similar gain or loss ratios of AlaOs, TiOz and Thus SiOz tends to buffer the mean volume factor in Nb. SiO<sub>2</sub> does not significantly effect the the calculations. weighted volume factor nor does it alter the mean arrangement of sample alteration intensity.

Generally, the larger the volume factor, the higher the alteration intensity. This is substantiated by petrography. L4E-9B contains relatively unaltered phlogopite, while L4E-6, L4E-2 and L4E-3 are highly sheared and pyritiferous. The calculated volume factors were used to compute percent gains or losses of components in the L4E-series rocks, relative to the least altered

	L4E-9B	L4E-14	L4E-13	L4E-17	L4E5	L4E-10	LAE-16	L4E-4	L4E6	L4E-2	L4E3
SiO <sub>2</sub>	1.00	0.96	1.04	1.10	0.90	0.96	1.02	0.93	1.09	1.49	1.34
A1203	0.89	1.05	1.11	1.15	1.15	1.02	1.16	1.23	1.19	0.69	1.06
TiO2	0.91	1.07	1.00	0.83	1.22	1.15	1.17	1.55	0.37	0.21	0.31
Nb	0.80	0.90	0.88	0.97	0.90	1.14	0.91	0.92	3.08	5.78	5.84
$F_v^w$	0.90	1.01	1.03	1.04	1.06	1.06	1.08	1.17	1.38	1.77	1.92

Table 8. Calculated Volume Factors.

 $F_v^w$  = Weighted mean volume factor.

syenite porphyry (L-65). Results are given in table 9.

Elemental Gains and Losses - Discussion

Percent losses or gains are plotted against volume factor (alteration intensity) in figure 8A to W. From these plots, the wallrock displays unequivocal net gains of SiO<sub>2</sub>, K<sub>2</sub>O, Ce, Mg, Rb, Sb, W, As, and Au. The gains in SiO<sub>2</sub>, K<sub>2</sub>O, Rb, Sb, and Au were discussed in the last section.

Ce, Mq, W, and As are added to the rocks within the ore zone. The gain of Ce in hydrothermal gold systems has been noted by Ludden et al. (1984). Magnesium is added to the majority of the samples with major gains in samples L4E-3 and L4E-2, which are closest to the ore zone, and most highly deformed. These enrichments of magnesium do not conflict with mineral stabilities (figure 6) since the  $a_{H_{\Phi}^{2}/a^{2}H^{\dagger}}$  remains low adjacent to the fluid conduit where The L4E the H<sup>+</sup> ion concentration should be the highest. samples all show extreme enrichments in ω. The hydrothermal fluids must have introduced W into the wallrock, however; the L-series samples suggest that W interacted quickly with the wallrock, and therefore did not accompany the fluids that altered the L-series rocks. Scheelite is a common mineral within the ore zones at

Table 9. Calculated Percent Loss or Gain of Components.

Element	L4E	-9B	L4E-14		L4E-13		
	wt. %	% loss or gain	wt. %	% loss or gain	wt. %	% loss or gain	
SiO <sub>2</sub>	-6.78	-10.26	+3.30	+5.00	-1.20	-0.79	
A1203	+0.11	+ 0.71	-0.64	-4.15	-1.08	-7.00	
Fe <sub>2</sub> 03	-0.19	-5.92	+0.72	+22.43	-0.19	-5.92	
Mgo	-0,58	-34.12	0.00	0.00	+0.65	+38.24	
Ca0	-2.00	-69.69	-0.05	-1.74	+0.78	+27.18	
Na20	-1.85	-40.84	-0.13	-2.87	-1.68	-37.07	
к <sub>2</sub> 0	+2.55	+87.03	+0.09	+3.07	+2.76	+94.20	
TiO <sub>2</sub>	-0.01	-3.13	-0.02	-6.25	+0.01	+3.13	
MnO	-0.02	-66.67	+0.02	+66.67	+0.03	+100.00	
P205	+0.01	+5.56	0.00	0.00	+0.01	+5.56	
Ba	-3221.00	-78.00	-1974.00	-48.00	-468.00	-11.00	
Ce	+3.00	+5.00	-25.00	-41.00	+7.00	+11.00	
Nd	+33.00	+127.00	+16.00	+62.00	+23.00	+88.00	
La	-13.00	-25.00	-7.00	-13.00	-11.00	-21,00	
Nb	+2.00	+15.00	+2.00	+15.00	+2.00	+15.00	
Y	-8.00	-23.00	-6.00	-17.00	-5,00	-14.00	
Zr	-165,00	-46.00	-143.00	-40.00	-162.00	-45.00	
Rb	+102.00	+200.00	+20.00	+39.00	+69.00	+169.00	
Sr	-1770.00	-88.00	-1426.00	-71.00	-1412.00	-70.00	
Sb	-7.00	-30.00	+36.00	+157.00	-8.00	-35.00	
W	+25.00	+2500.00	+42.00	+4200.00	+31.00	+3100.00	
As	-9.00	-64.00	-7.00	-50.00	-9.00	-64.00	
Au	+3.95	+35909.0	+18.58	168909.0	+9.94	90364,00	

# Table 9, con't

Element	L4	E-17	L4	E-5	L4E-10	
	wt. 🕇	% loss or gain	wt. %	% loss or gain	wt. %	% loss or gain
SiO <sub>2</sub>	-3.37	-5.10	+11.46	+17,35	+7.32	+10.94
A12 <sup>0</sup> 3	-1.49	-9.66	-1.24	-8.04	+0.54	+3.50
$Fe_2O_3$	+0.92	+28.66	-0.22	-6.85	-0.04	-1.25
MgO	+1.78	+104.71	-1.15	-67.65	+0.25	+14.71
Ca0	+3.06	+106.62	-1.74	-60.63	-0.64	+22.30
Na20	-1.79	-39.51	-4.15	-99.56	-3.88	-85.65
к <sub>2</sub> 0	+1.06	+36.18	+5.81	+198.29	+5.79	+197.61
TiO <sub>2</sub>	+0.09	+28.13	-0.05	-15.63	-0.02	-6.25
MnO	+0.06	+200.00	-0.02	-66.67	+0.01	+33.33
P <sub>2</sub> 0 <sub>5</sub>	+0.15	+83.33	-0.08	-44.44	+0.01	+5.56
Ba	-2835.00	-69.00	-1543.00	-38.00	-2452.00	-588.00
Ce	+33.00	+54.00	+17.00	+28.00	+14.00	+23.00
Nd	+40.00	+154.00	+33.00	+127.00	+31.00	+119.00
La	+6.00	+12.00	-17.00	-33.00	0.00	0.00
NЪ	+1.00	+8.00	+2.00	-15.00	-1.00	-8.00
Y	-2.00	-6.00	-10.00	-29.00	-10.00	-29.00
Zr	-96.00	-27.00	-188.00	-52.00	-175.00	-48.00
Rъ	+71.00	+139.00	+148.00	+290.00	+157.00	+308.00
Sr	-1026.00	-51.00	-1721.00	-86.00	-1724.00	-86.00
Sb	+59.00	+257.00	+13.00	+57.00	-1.00	-4.00
W	+27.00	+2700.00	+78.00	+7800.00	+56.00	+5600.00
As	-1.00	-7.00	+19.00	+136.00	+9.00	+64.00
Au	+1.11	+10C91.00	+6.68	+60727.00	+3.34	30364.00

<u>Table 9.</u> con't.

Element	L4E-	16	L4E-	4	L4E-	-6
	wt. %	% loss or gain	wt. %	% loss or gain	wt. %	% loss or gain
SiO2	+4.12	+6.24	+17.27	+26.14	+17.45	+26.42
A12 <sup>0</sup> 3	-1.50	-9.73	-0.76	-4.93	+2.42	+15.69
Fe <sub>2</sub> 03	-0.17	-5.30	-0.23	-7.17	+11.20	+348.91
Mg	+1.43	+84.12	-0.87	-51.18	+0.82	+48.24
Ca0	+2.08	+72.47	-1.04	-36.24	+0.86	+29.97
NaO2	-1.72	-37.97	-4.01	-88.52	-4.09	-90.29
к <sub>2</sub> 0	+1.13	+38.57	+7.68	+262.12	+7.38	+251.88
TiO <sub>2</sub>	+0.07	+21.88	-0.07	-21.88	+0.88	+275.00
MnO	+0.04	+133.33	-0.01	-33.33	+0.04	+133.33
P <sub>2</sub> 0 <sub>5</sub>	+0.14	+77.78	-0.09	-50.00	-0.02	-11.11
Ba	-3140.00	-76.00	-481.00	-12.00	-2333.00	-57.00
Се	+18.00	+30.00	-4.00	-7.00	+22.00	+36.00
Nd	0.00	0.00	+17.00	+65.00	+46.00	+169.00
La	-10.00	-19.00	-31.00	-60.00	+21.00	+40.00
NЪ	+2.00	+15.00	+4.00	+31.00	-7'.00	-54.00
Y	0.00	0.00	-1.00	-3.00	-15.00	-43.00
Zr	-123.00	-34.00	-171.00	+47.00	-15900	+44.00
Rb	+89.00	+175.00	+153.00	+300.00	+214.00	+420.00
Sr	-1343.00	-67.00	-1597.00	-80.00	-1565.00	-78.00
Sb	+75.00	+326.00	+27.00	+117.00	+44.00	+191.00
W	+23.00	+2300.00	+85.00	+8500.00	+257.00	25700.00
As	+11.00	+79.00	+26.00	+186.00	+60.00	+429.00
Au	+3.38	30727.00	+9.50	86364.00	+59.21	+538273.0

Table 9. con't.

Element	L4E	-2	L4E	-3
	wt. %	% loss or gain	wt. %	% loss or gain
SiO <sub>2</sub>	+12.45	+18.85	+28.91	+43.76
A12 <sup>0</sup> 3	+24.17	+156.74	+12.60	+81.71
$Fe_2O_3$	+25.74	+801.87	+21.15	+658.88
MgO	+4.73	+278.24	+10.02	+589.41
Ca0	-0.44	-15.33	+12.92	+450.17
Na <sub>2</sub> 0	-4.21	-92.94	-3.65	-80.57
к <sub>2</sub> 0	+15.87	+541.64	+13.13	+448.12
TiO <sub>2</sub>	+2.33	+728.13	+1.63	+509.38
MnO	-0.01	-33.33	+0.25	+833.33
P <sub>2</sub> 0 <sub>5</sub>	+0.45	+250.00	+0.80	+444.44
Ba	-1274.00	-31.00	-1394.00	-34.00
Ce	+222.00	+364.00	+89.00	+146.00
Nd	+120.00	+462.00	+85.00	+327.00
La	+98.00	+188.00	+6.00	+12.00
Nb	-9.00	-69.00	-9.00	-69.00
Ŷ	-31.00	-89.00	-5.00	-14.00
Zr	+44.00	+12.00	-39.00	-11.00
Rb	+667.00	+1308.00	+375.00	+735.00
Sr	-1745.00	-87.00	-898.00	-45.00
Sb	+98.00	+426.00	+63.00	+274.00
W	+500.00	+50000.00	+375.00	+37500.00
As	+109.00	+779.00	+144.00	+1029.00
Au	+15.17	137909.00	+25,42	231091.00



Percent Loss or Gain of Al2O3 vs. Volume Factor



Percent Gain or Lass of Al203







Volume Factor













Percent Loss or Gain of La vs. Volume Factor

N.







1.08

1.17 1.38 1.77

1.92

1.06

-60

-70

-80

-90

0.9

1.01

1.03 1.04















Macassa (Kerrich and Watson, 1984) where massive additions of tungsten are also noted. Arsenic shows depletions in the less altered samples, and enrichments in the highly altered samples. This suggests a "dumping" of As into the samples which underwent the highest degree of fluid-rock interaction. The more evolved fluid became undersaturated in As with increased permeation of the country rock, and thereby leached the L-series samples.

NagO, Ba, Zr, Sr, and Y were lost from the wallrock during hydrothermal alteration. The loss of the first four components has been discussed in the L-series section. the loss of Y within the ore zone is depicted in figure 8P. This suggests that Y was undersaturated in the fluid relative to the wallrock, and that diffusion of Y into the hydrothermal fluids occurred during alteration.

# K - Metasomatism

The predominant alteration process noted in this study is K-metasomatism. The massive introduction of K into the wallrock is shown in figure 86. In general, increasing alteration intensity correlates with increasing K addition. This gain of K is instrumental in allowing the precipitation of K-feldspar in the secondary orthoclasepyrite alteration facies.

Figure 9 shows the antipathetic nature of potassium

Na20 (wt.%) vs. K20 (wt. %)



Rb (ppm) vs. K20 (wt. %)



and sodium in the porphyritic sympite. Those rocks with the highest K20 and lowest Na20 content have undergone the highest degree of hydrothermal alteration.

Figure 10 shows the covariance of K and Rb, indicating that Rb has the same chemical nature as K in hydrothermal systems.

## Summary

The following geochemical trends are noted with increasing intensity of hydrothermal alteration:

1) Addition of KaO, SiOa, Rb, Sb, and Au.

2) Depletion of NazO, Ba, Sr, and Zr.

These variations are consistent with petrographic observations.

#### CONCLUSIONS

Petrologic and **geochemical** evidence provide information on ore deposition and hydrothermal alteration at the Lake Shore Mine. Ore petrography established a paragenetic scheme that fundamentally agrees with previous work of Todd. (1928) and Hawley. (1948). Petrography of the altered wallrock revealed mineralogical variations with increasing alteration intensity, and allowed alteration facies to be recoonized. Chemical gains and losses due to hydrothermal alteration substantiate the observed petrography.

Ore petrography concluded that the sulfide minerals were early phases in the paragenesis of the Kirkaland Lake ores. The paragenetic position of sphalerite could be deduced (c.f. Todd, 1928; Hawley, 1948) since the L4E samples contained numerous intergrowths of sphalerite and ore minerals. Telluride minerals were precipitated at a later stage than the sulfides. Gold emplacement occurred toward the end of the paragenetic scheme, possibly contemporaneous with altaite. There is evidence that pyrite precipitation occurred throughout the paragenesis.

Thin section petrography established mineral assemblages which show progressive alteration intensity toward the ore zone. The mica group minerals can be used as markers to interpret four alteration facies. With increasing alteration intensity, magmatic biotite is altered to chlorite, chlorite to phlogopite, and finally to sericite. The facies with the highest degree of alteration contained orthoclase which is interpreted to have precipitated from a fluid with a high and and This fluid evolved as it permeated, and interacted with the wallrock to produce the observed facies of alteration minerals. The development of carbonate minerals and the absence of epidote and sphene suggest that this fluid was highly CO2 charged.

Geochemical evidence leads to the conclusion that the process of hydrothermal alteration added  $K_{20}$ , SiO<sub>2</sub>, Rb. Sb, and Au to the wallrock, and removed NazO, Ba and Zr from the wallrock. No "immobile" elements were identified in this study, but certain "least mobile" elements (AlaOs, TiOz, SiOz, and Nb) allowed semi-quantitative analysis of elemental gains or losses. Isochemical and isovolumetric balance relationships are in agreement. The mass antipathetic relationship between NazO and KzO and the sympathetic relationship between Rb and KaO is due to pervasive K-metasomatism accompanying the hydrothermal alteration.

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## **APPENDIX 1**

Element concentrations were calculated as follows: Au concentration in sample 8;

Ts/2 = 64.8 hours = ln2/Ts/2 = 0.01069672 Sample weight = 0.4967 grams Standard concentration = 92.11 ppb. t = \Delta t = count time of standard-count time of sample = 11.75 - 9.4 N = measured activity of sample N= = corrected activity of sample Decay Equation:

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N = N_{\bullet} e^{-\lambda_{\bullet}}
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therefore

 $N_{e} = N/e^{-\lambda_{e}}$ 

= 28,661/e-0.01044472 × 2.38

= 29391

IF standard has concentration of 92.11 ppb., and an activity of 1886.5

THEN, sample 8 has a raw concentration of

 $Au = 92.11 \text{ ppb } \times 29391/1886.5$ 

= 1435 ppb.

To obtain the actual concentration in one grain of sample, divide the raw concentration by the sample weight:

1435 ppb / 0.4967 grams = 2889 ppb.

APPENDIX 2

Gresens (1967) Equation:

$$\Delta X_n = a \left[ \left\{ K_v \chi_n^{\beta} \left( \rho^{\beta} / \rho^{\gamma} \right) \right\} - \chi_n^{\alpha} \right]$$

where

 $\Delta X_n$  = chemical transfer in grams of component n from rock x to rock  $\beta$ 

a = grams of parent rock

Kv = ratio of the volume of rock to rock  $\chi_n^{\alpha,\beta}$  = weight fractions of component n in rocks  $\propto$  and  $\beta$  $\rho^{\tilde{\alpha}}$  and  $\rho^{\beta}$  = densities of  $\propto$  and  $\beta$ , respectively.