

**ECONOMIC GEOLOGY
RESEARCH UNIT**

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ARCHAEOAN LODE GOLD DEPOSITS OF CANADA

**Part II : Characteristics of the hydrothermal systems,
and models of origin**

R. KERRICH

— • INFORMATION CIRCULAR No.183

UNIVERSITY OF THE WITWATERSRAND
JOHANNESBURG

ARCHAean LODE GOLD DEPOSITS OF CANADA: PART II

CHARACTERISTICS OF THE HYDROTHERMAL SYSTEMS, AND MODELS OF ORIGIN

by

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ABSTRACT

Archaean lode gold deposits are characteristically enriched in a suite of rare elements including Au, Ag, B, W (+ Se, Te, Bi, Pd) which are concentrated by 10^3 to 10^4 times background, whereas the abundant and soluble base metals, Cu, Pb, Zn, are present at only 10^{-1} to 10^1 times background. This separation may be accounted for if mineralizing fluids are generated under crust-dominated conditions of low water/rock, and low Cl, such that the absolute abundance of rare elements is not constrained by solubility, but base metal solute concentration is limited by availability, of both H₂O and Cl. Sporadic development of Mo + graphite is a notable feature of some deposits.

The wide range of ambient Eh for gold deposition indicated by the coexistence of gold + hematite, to gold + pyrrhotite + graphite in different environments suggests that precipitation of gold is not simply mediated by redox and the availability of reduced S, as is the case for massive base metal sulphide deposits.

Gold-bearing vein systems formed during multiple episodes of hydrofracturing, from transient, pulsed hydrothermal discharge, under conditions where $P_{\text{fluid}} \geq \sigma_3 + T$, and P_{fluid} may have exceeded σ_1 , the maximum principal stress. Pulsed discharge-generated banded veins were probably induced by repeated breaching of a metamorphic fluid reservoir. The narrow recorded range in $\delta^{18}\text{O}$ of vein quartz at 12-15 per mil across banded veins, and of gold-bearing vein arrays within mines, independent of all-rock type, signifies an isotopically uniform fluid reservoir, a fluid-dominated system in the hydrofracture conduits, and a relatively small temperature span over which veins precipitated. Hydrothermal solutions transporting gold were isotopically characterized by $\delta^{18}\text{O}$ 6 to 9 per mil and δD -80 to -40 per mil, consistent with a metamorphic fluid reservoir. These fluids were at temperatures of 270 to 450°C, possessed low salinities of ≤ 2 wt. % NaCl equivalent, near neutral pH, Na/K ~ 0.1, and contained significant quantities of CO₂, along with the reduced gas species CH₄, CO, H₂, and ultrafine particulate hydrocarbons. The CO₂ may have been derived from mixtures of various possible carbon reservoirs, including marine and igneous carbonate, decarbonated marine carbonate or decarboxylation of organic hydrocarbons.

The fluid δD , $\delta^{18}\text{O}$, low salinity, high K/Na, variably elevated CO₂ and reduced gas species are all consistent with a metamorphic or magmatic fluid reservoir; whereas none of these geochemical features are typical of marine-water-derived ore-forming fluids. Hydrothermal fluids were more radiogenic ($^{87}\text{Sr}/^{86}\text{Sr} = 0.7010$ to 0.7025) and possessed higher μ than contemporaneous mantle, komatiites, or tholeiites, and thus carried a contribution from older sialic basement. Mineralized faults possess enrichments of LIL elements including K, Rb, Li, Cs, B, and CO₂, as well as rare elements such as Au, Ag, As, Sb, Se, Te, Bi, and W.

The ubiquitous gangue carbonate minerals formed via hydrolysis of Fe, Mg, Ca, Mn-silicates to Fe, Mg, Ca, Mn-carbonates, in which wall rocks donate the bivalent metal cations and CO₂ is indigenous to the hydrothermal fluids. Ultramafic rocks and oxide facies iron-formation are effective CO₂ sinks, given their inherently elevated complement of bivalent metal cations.

Wall rocks to veins are intensely reduced ($\text{Fe}^{2+}/\Sigma\text{Fe} \sim 0.9$) relative to background ($\text{Fe}^{2+}/\Sigma\text{Fe} \sim 0.7$), and this is also reflected in positive anomalies of Eu²⁺. The reductant involved was probably H₂, a product of the dissociation of H₂O under high T and P conditions in source rocks.

Of the various models proposed for greenstone-belt lode gold deposits, most are incommensurate with the geological and geochemical boundary conditions. Derivation of gold from the quartz-feldspar porphyry bodies spatially associated with many deposits is incompatible with known differences in isotopic, redox, REE, and alteration patterns of stocks and deposits; and given their small volume, with considerations of material balance for the gold source. Lateral secretion of gold and gold-associated chemical components into dilatant openings from enclosing wall rocks is not consistent with the known range of grain boundary diffusion distances over geological time spans. Lateral secretion does not account for the separation of noble from rare elements. Furthermore, this proposed mechanism is incompatible with the hydraulic, isotopic, redox, REE, and mineralogical disequilibrium between deposits and their peripheral host rocks.

There is no evidence that greenstone-belt rock types contain more abundant gold than their Proterozoic and Phanerozoic counterparts, except possibly for the volumetrically minor Algoman banded iron-formation. Indeed,

primary ultramafic rocks possess lower gold contents than mafic or felsic igneous rocks. Hence, there is no basis for proposed genetic links between various source rocks conjectured to have inherently elevated gold abundances and deposits.

Many authors have invoked seawater-hydrothermal fluids to account for the putative chemical sedimentary gold deposits; and some have proposed that the major auriferous quartz vein arrays are sites of seawater-hydrothermal fluid discharge, analogous to stockworks beneath the seawater-generated base metal massive sulphide deposits. However, no hard geological or geochemical evidence has been advanced to support seawater involvement in gold deposits. The seawater hypotheses do not account for the observed metal separation in gold deposits. Not only are the metal associations and alteration patterns of the greenstone-belt lode gold and base metal deposits markedly different, but the hydrothermal fluid δD , $\delta^{18}O$, redox, CO_2 , K/Na, and salinity of vein gold deposits are all commensurate with a metamorphic or magmatic fluid reservoir, whereas none of these geochemical features are typical of seawater.

Hence, the greenstone-belt gold deposits appear to form during the discharge of low-Cl, CO_2 -rich, aqueous fluids of metamorphic and/or magmatic origin, at depth in a compressive, crustal regime. This discharge represents the return flow of a cycle in which H_2O and CO_2 from the hydrosphere are initially transferred to rocks as hydrates, carbonates, and hydrocarbons, during cooling of volcanic flows at the ocean floor. These components are subsequently returned to the hydrosphere via outgassing of the volatile-laden volcanic sequence under conditions of burial accompanied by crustal heating. This hydration-dehydration cycle proceeds continuously during development of the greenstone belt, and is not considered to be primarily the product of a late putative orogeny. Later outgassing may occur during thrusting, and in the envelopes of rising batholiths, leading to additional concentrations of gold.

It is suggested that rift environments, with high heat fluxes, are appropriate for efficient gold concentration. Here, listric normal faults initially control extension along with ultramafic volcanism, provide pathways for mafic and felsic magmas, and are responsible for scarp-dominated sedimentary rocks. At later stages of greenstone belt development, reverse motion on the faults accommodates crustal compression, and structures act to focus hydrothermal discharge generating the gold deposits, as well as conduits for trondhjemite and alkalic magmas. The geometry of listric faults may be particularly efficient for tapping large volumes of crust.

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ARCHAEN LODE GOLD DEPOSITS OF CANADA:
CHARACTERISTICS OF THE HYDROTHERMAL SYSTEMS, AND MODELS OF ORIGIN

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CHARACTERISTICS OF THE HYDROTHERMAL SYSTEMS, AND MODELS OF ORIGIN

I. INTRODUCTION

This paper attempts to convey a summary of the geochemical characteristics of Archaean lode gold deposits of Canada. The survey is based on a synthesis of information presented in Part I (Kerrich, 1986), along with additional data pertaining to specific topics.

In the first section, a compilation of metal distributions, including rare earth elements, transition and base metals, is reported for a series of gold deposits selected to represent a wide range of geological settings. This is followed by a discussion of the geochemistry of alteration, with reference to deposits which encompass the variety of dominant alteration styles.

Specific questions considered in the context of the hydrothermal ore-forming system for Archaean lode deposits are:

1. source of volatile components such as H₂O, CO₂, H₂S;
2. the nature of rock reservoirs donating lithophile and other elements locally enriched in faults;
3. ambient temperature-pressure and hydraulic conditions;
4. the history of incremental displacement and fluid activity, and
5. the presence of highly sodic or alkaline magmas along some structural lineaments hosting gold deposits.

In the ensuing sections, the diverse lines of geochemical evidence reported are drawn together in an attempt to synthesize the data in a coherent manner. This is presented in a series of sections covering specific topics such as the properties of hydrothermal fluids involved in gold deposits, patterns of alteration, the gold-carbonate-potassium association, redox relations, aluminous rocks enveloping gold deposits, and a comparison of lode gold and base metal deposit characteristics, and of their hydrothermal reservoirs. This paper concludes with a critical discussion of the various models proposed for Archaean lode gold deposits.

II. METAL INVENTORY OF LODE GOLD DEPOSITS

Lode gold deposits are characterized by a distinctive metal inventory, or element association, raising the question as to why the rare element gold is preferentially enriched over the relatively abundant base and transition metals. Average abundances of Au, Ag, Pd, Fe₂O₃, MnO, S, CO₂, Be, B, As, Sb, V, Cr, Co, Ni, Cu, Zn, Mo, Sn, W, Hg, Pb, Cd, Ba, and Th in 30 separate ore bodies from 18 gold mines have been compiled by Kerrich and Fryer (1981), Kerrich and Hodder (1982), and Kerrich (1983). These results are graphically summarized in Fig. 1, where examples of auriferous veins from a wide range of geological environments encompassing ultramafic, mafic, and felsic volcanic rocks, along with detrital sediments are included. Metal distributions of deposits are portrayed in terms of enrichment factors, where enrichment is given by the average content of an element in a specified deposit, divided by its background abundance. A summary of the abundances of selected elements in various primary igneous rock types, and in Archaean greywackes, is reported in Table I, to facilitate comparisons of data from mines to background levels.

A divergence in the enrichment factors of two elements in a given deposit then indicates separation, or fractionation, of these elements during mineralization. For the case of lode gold deposits, all enrichments are calculated relative to background levels in the most abundant host rock type present. Where many different rock types are present enrichments have been computed with reference to element contents of primary basalts, because mafic volcanic rocks are volumetrically predominant as contiguous rocks to the deposits (Hutchinson, 1976; Hodgson, 1983).

For purposes of comparison, enrichments of Au, Ag, Cu, Zn, and Pb in base metal sulphide deposits are given in Figs. 2 and 3. These are expressed relative to abundances in basalts and also relative to felsic volcanic rocks, inasmuch as the latter can frequently constitute the immediate footwall. However, the patterns of element enrichments in lode gold and base metal sulphide deposits, respectively, are distinctive and fundamentally unchanged, irrespective of whether enrichments are referenced to an assumed mafic or felsic background rock (Figs. 1, 2).

A. Lode Gold Deposits

In gold-bearing veins Au, Ag, and Pd average 30, 8, and 0,16 ppm, respectively, compared to abundances of 2, 100, and 8 ppb determined for these elements in primary rock types remote from ore bodies and in mafic igneous rocks worldwide (Kerrich, 1983, Tables I and II; Tilling *et al.*, 1973; Kwong and Crocket, 1978; Frueh and Vincent, 1974; Parthe and Crocket, 1972). Concentration factors in the ore bodies are ~15,000 for Au, ~80 for Ag, and 20 for Pd relative to background abundances. Some of the mean Au enrichments given in Fig. 1 are greater than for the mines as a whole because predominantly high-grade ore was selected for determining metal distributions, with host rocks to the ore being collected and analyzed separately. Several elements are typically present in association with Au, such as W, As, Sb, and B, which have also undergone a high degree of concentration. Tungsten is a rare element with an average background abundance of 0,4 ppm in oceanic basalts (Helsen *et al.*, 1978), but is common as scheelite within gold-bearing veins, and has also been reported at anomalous concentrations in some chemical sedimentary rocks (Harmon *et al.*, 1975, 1978; Kerrich and Fryer, 1981, Table I). Tungsten is present at 5 to 50 ppm in most gold-bearing veins and sedimentary rocks, but lodes of scheelite are sporadically distributed in deposits, from some of which there has been past production, such as the Dome and Hollinger mines, Timmins District; thus estimates of bulk W contents of gold ore are difficult to make.

Arsenic is another rare element, which is typically present at the 100 to 10,000 ppm level in gold lodes, being contained within the ubiquitous arsenopyrite. In laminated pyritic rocks of the Agnico Eagle Mine, Joutel, in the silicate facies iron-formation of the East South C Zone of the Dickenson Mine, Red Lake District, and in quartz-carbonate veins of the Con Mine, Yellowknife, the content of As is locally 0,2 to 2 per cent (Boyle, 1961; Kerrich *et al.*, 1981; Kerrich, 1983, Table II), which therefore implies a concentration of 1000 to 10,000 times the background abundance of 2 ppm in basalts (Turekian and Wedepohl, 1961).

Boron is abundant in many gold-bearing veins and chemical sedimentary rocks, B being contained principally within tourmaline or axinite $(\text{Ca}, \text{Mn}, \text{Fe})_3\text{Al}_2\text{B}_3[\text{Si}_2\text{O}_5] (\text{OH})$. Notable examples include the Dome and Buffalo Ankerite mines of the Timmins District, where tourmaline is present both in chemical sedimentary rocks and massive quartz-carbonate veins. Tourmaline is abundant in quartz veins near the No. 6 zone at Mine Bousquet, at Mine Doyon - Cadillac District, at the Lamaque Mine - Val d'Or District, and at Passegem de Mariana in Brazil. With a background content of 5 ppm in basalts and 20 ppm in shales (Turekian and Wedepohl, 1961), and levels exceeding 1,5 per cent in some ore deposits (Fleischer and Routhier, 1973; Table I), the enrichment factor for B is 100 to 3,000 depending on the nature of the source region.

Tellurides, selenides, and bismuthides have a conspicuous association with lode gold deposits (Boyle, 1979); the Kirkland Lake gold district is among the prominent examples. Although the bulk Te, Se, and Bi abundances of Archaean lode gold deposits are not well known (Boyle, 1979, p. 151), the presence of Te-, Se- and Bi-bearing minerals signifies that these rare elements sporadically attain large degrees of enrichment in lode gold deposits, given their low background abundances of 0,01 to 0,001, 0,05 and 0,007 ppm, respectively (Table I; Turekian and Wedepohl, 1961; Rösler and Lange, 1972, p. 231).

In certain ore bodies, Cr abundance is in excess of 2000 ppm, as in the quartz Cr-muscovite veins hosted in carbonate-altered ultramafic rocks ('carbonate ore') at the Kerr Addison Mine, Larder Lake District, in the quartz-muscovite veins of the Dome Mine, and No. 12 'vein' of the Hollinger Mine open pit. In all of these examples the ore is enveloped by rocks of original ultramafic and basaltic composition. The mean Ni content in the East South C Zone sedimentary rocks, Dickenson Mine, which are hosted by basalts, is 900 ppm. Hence, elevated abundances of Cr and Ni in gold ore bodies typically appear to exist where ultramafic rocks are present, indicating that local rock composition mediates to a significant degree the abundance of these two transition metals. However, some hydrothermal transport of nickel and chromium may also occur along with the precious metals.

The average content of Pb, Zn, and Cu in gold-bearing veins and sedimentary rocks is only 0,1 to 5 times their abundances in contiguous mafic volcanic rocks and in primary basalts worldwide, signifying a low degree of enrichment, or in some instances lower than background abundances of the base metals (Fig. 1). Minor erratic enrichments of Pb occurs in tandem with gold at the Con Mine, Yellowknife, and in quartz-carbonate veins within 'porphyry' at the Dome Mine; and zinc is sporadically enriched along with gold in the East South C Zone, Dickenson Mine. There is as yet no clear pattern to these minor anomalies of base metals in lode gold deposits, but they are quantitatively insignificant in terms of the low overall base metal complement. Indeed, if base metal contents of lode gold deposits were significant, recovery of gold by conventional cyanidation would not be economically feasible, given the greater affinity of cyanide for Cu and Zn than for Au.

TABLE I

Abundances of Selected Metals in Primary Igneous Rocks and Archaean Greywackes

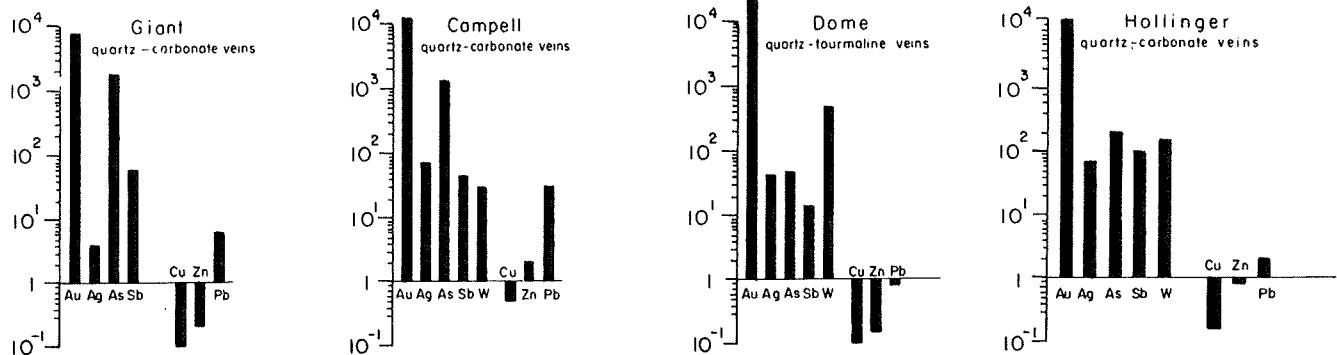
	Ultramafic	Mafic	Felsic	Archaean greywackes
Chromium	1600	170	22	74
Cobalt	150	48	7	25
Nickel	2000	130	15	48
Copper	10	87	30	61
Zinc	50	105	60	75
Lead	1	6	15	11
Gold	0,0008	0,0017	0,002	0,002
Silver	0,060	0,110	0,051	0,080
Palladium	0,009	0,016	0,002	
Platinum	0,011	0,009	0,008	
Arsenic		2		
Antimony		0,2		
Tungsten		0,4		
Selenium	0,05		0,05	
Tellurium			0,01 to 0,001	
Bismuth	0,001		0,01	

Abundances expressed in ppm.

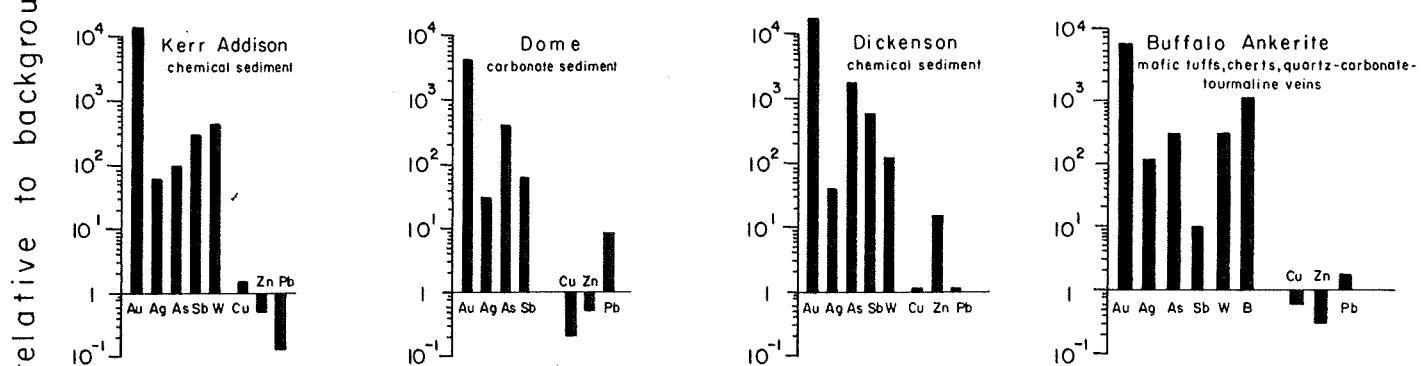
Data compiled from Turekian and Wedepohl (1981), Kwong and Crocket (1978), Parthé and Crocket (1972), and Röslér and Lange (1972). Data for Archaean greywackes represents the mean of 20 analyses of the Kewagama and Cadillac groups.

LODE GOLD DEPOSITS

vein systems in mafic to ultramafic environments



chemical sedimentary rocks in mafic to ultramafic environments



veins and chemical sediments in felsic intrusive-extrusive environments

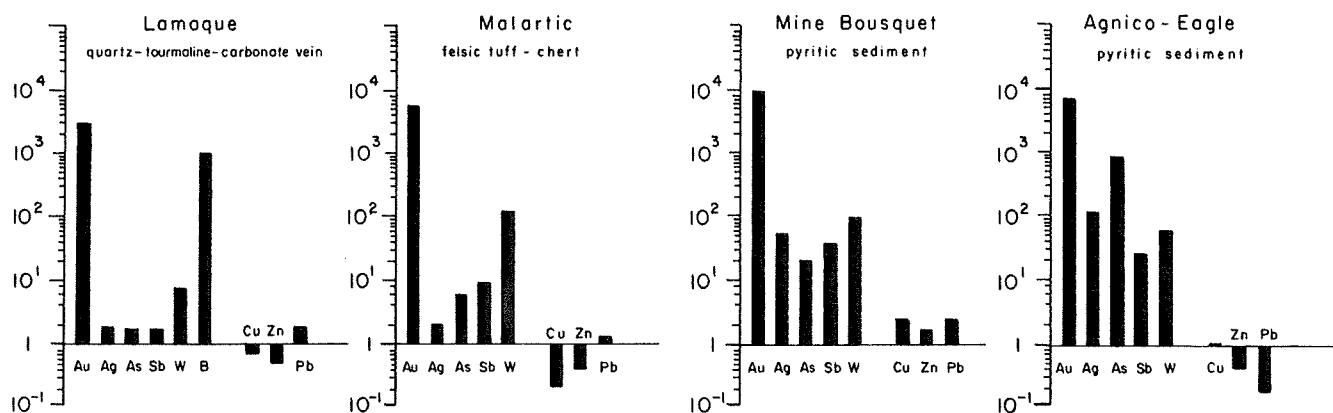


Figure 1 : Enrichments of selected elements in Archaean lode gold deposits, referenced to background abundances in basalts. Note logarithmic coordinate (modified from Kerrich, 1983).

Vanadium, Be, and Co are present at levels which broadly reflect the nature and extent of wall-rock fragments incorporated into veins, or of pyroclastic infall mixing with mineralized sedimentary rocks, inasmuch as they covary with essentially immobile elements such as Al, Sc, and Zr which are generally indigenous only to the host rocks (cf. Kerrich *et al.*, 1981). Cobalt, however, locally attains abundances in excess of 300 ppm in the Number 15 'syenite' ore body of the Kerr Addison Mine. Lode gold deposits generally have minor contents of Mo, Sn, Cd, U, and Th, approximately equal to or less than their average basaltic abundances of 1.5, 1.5, 0.3 and 4 ppm, respectively. However, molybdenum- and graphite-coated fractures are sporadically developed in some deposits: notable examples include those of the Kirkland Lake District, Taylor Township near Timmins, the Hemlo District, and Chadbourne Mine, Noranda.

Figure 1 illustrates the magnitude by which the rare elements and the abundant transition metals are separated. In the Con Mine, Yellowknife, the enrichment factors for gold and copper are 10,000 and 0.1, respectively, signifying separation of these two elements by a factor of 100,000. The essential features of this distribution are repeated in the other ore bodies, which encompass a wide variety of host rocks including chemical sedimentary rocks, and in diverse structural settings. There is thus an overall coherence to the patterns of relative enrichment of elements in lode gold deposits.

Within the overall characteristic enrichment pattern of gold and its associated elements (Ag, As, Sb, B, W, Se, Te, Bi) in Archaean lode deposits, there is significant variability in the distribution of the gold-associated elements that does not appear to be a simple function of the immediate host rock type. For example, arsenic is abundant in the pyritic gold deposit of Agnico Eagle, but not at the possibly comparable deposits of Thompson, Bousquet, or Dumagami, all three of which are hosted by felsic fragmental rocks. Gold-bearing quartz-carbonate veins at Yellowknife, located within a sequence of metabasalts, possess As contents in the 10,000 ppm range coupled with low B, yet at the Dome Mine similar quartz veins also transecting basalts have As contents of only 100 ppm, but abundant tourmaline.

In summary, lode gold deposits have extreme enrichments of certain rare elements, including B, As, Sb, W, and Pd, with or without Te, Se, and Bi, along with low enrichments of the abundant transition and base metals, relative to their enclosing rocks. The salient question that must be resolved is what is the fractionation process whereby rare elements, and metals such as Cr, Ni, and W, conventionally considered to be immobile under conditions of low- to intermediate-grade metamorphism, are preferentially concentrated relative to the mobile and abundant base metals.

B. Base Metal Sulphide Deposits

Enrichment factors for Au, Ag, Cu, Zn, and Pb, where present, are depicted for 20 essentially Zn-Cu massive base metal sulphide deposits in Fig. 2. Most of the ore bodies are disposed within Archaean greenstone belts, but examples of Proterozoic age are also included. Patterns of relative enrichment are remarkably uniform; Au, Ag, Cu, and Zn are all concentrated by factors of 200 to 2000 time background, and there is thus no large separation of elements during concentration from background into the deposits. Data are for mines as a whole, notwithstanding the spatial separation of Cu and Zn within individual ore bodies. These patterns of relative enrichment are reproduced in some massive sulphide deposits of younger age, for instance those in the Buchans camp and in the Iberian pyrite belt (Strauss *et al.*, 1977; Franklin *et al.*, 1981; Fig. 3).

There is no comprehensive published information on the rare-element contents of more polymetallic massive sulphide deposits such as those of the Kuroko district, Japan. Mercer (1976) reports abundances of Mo, Sb, As, Bi and Cd for Kuroko deposits, which correspond to approximate enrichment factors of 200, 1500, 250, 10,000 and 3000 respectively.

C. Comparison of Lode Gold and Base Metal Deposits

Whereas there exists extreme separation of rare elements from the abundant and soluble base metals in lode gold deposits, massive base metal sulphide deposits do not possess any large separations of Au from Cu and Zn (Fig. 4). A continuum exists in gold content from Au-poor massive sulphide deposits such as at Flin Flon, Manitoba, through Au-enriched samples such as the Quemont and Horne mines of the Noranda district. However, the majority of Archaean base metal massive sulphide deposits have a significantly lower absolute Au abundance than the lode gold deposits. The salient feature is the overall paucity of base metal content in lode gold deposits.

BASE METAL DEPOSITS

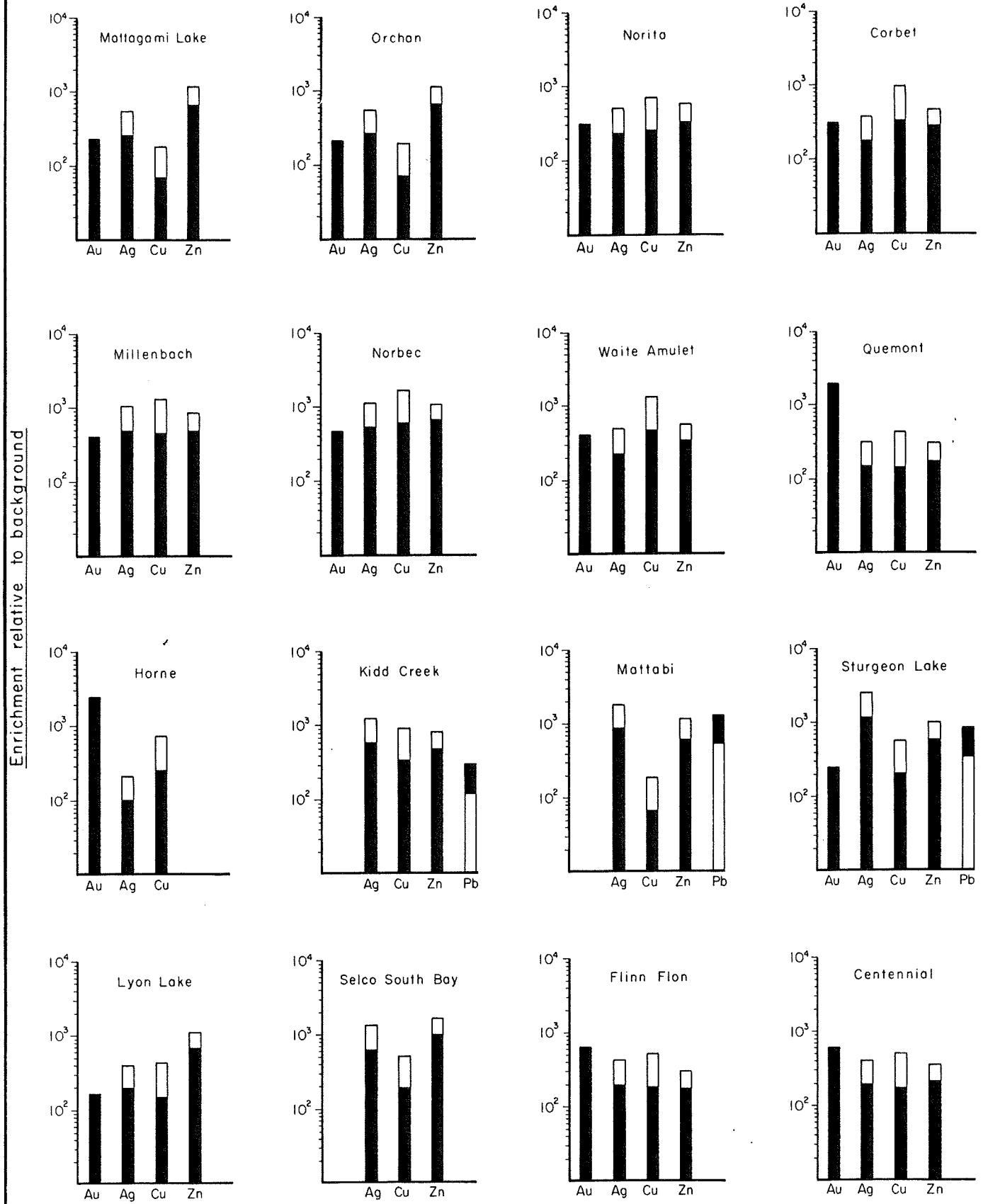


Figure 2 : Enrichments of Au, Ag, Cu, Zn, and Pb where recovered in some Precambrian base metal massive sulphide deposits. Solid bars represent enrichments relative to background abundance in unaltered mafic volcanic rocks; open bars signify enrichments relative to abundances in primary felsic igneous rocks. Note logarithmic co-ordinate. Based on data in Worobec, 1980, and additional sources (after Kerrich, 1983).

IBERIAN PYRITE BELT

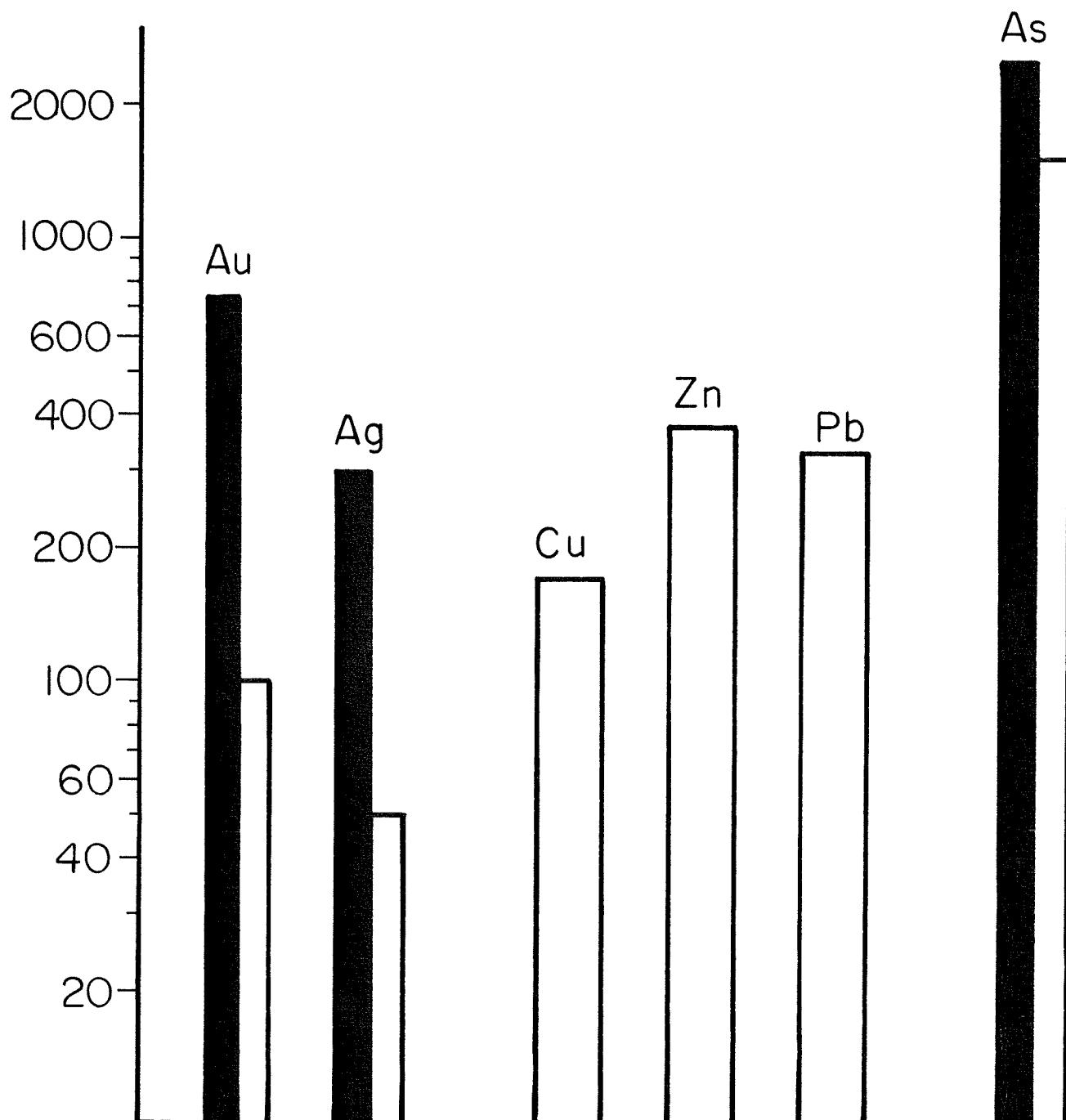


Figure 3 : Enrichment factors for specified metals in base metal massive sulphide deposits of the Iberian Pyrite Belt (data source, Strauss, 1974).

Further complexities in the patterns of metal separation are from the distribution of other elements, including cadmium. This rare element, having a background abundance of 0.3 ppm, and which displays geochemically coherent behaviour with zinc in polymetallic deposits, is not significantly enriched in Archaean lode gold deposits (Kerrich, 1983, Table II). Processes that have been advanced to account for the observed separation of rare elements from base metals in Archaean lode gold deposits are critically examined in the following section.

A key feature of the element distribution in Archaean lode gold deposits is the enrichment of a select group of lithophile elements (K, Rb, Ba, Cs) along with the rare elements such as those discussed above. Lithophile elements like Be, Sn, REE, U, and Th, with a general magmatic association, do not appear to be characteristically enriched by the ore-forming hydrothermal system. A second important observation to be emphasized is that the metal inventory is essentially independent of the composition of contiguous rocks hosting the deposits.

D. Separation of Base Metals From Rare Elements in Archaean Lode Gold Deposits

Suggested mechanisms for the fractionation of base metals from rare elements in Archaean lode gold deposits fall into four general categories:

- (1) those that appeal to special source rocks;
- (2) those involving selective precipitation of gold;
- (3) those that consider properties of the hydrothermal systems; and
- (4) special ligands.

1. Special Source Rocks

The question of the relationship of gold deposits to rocks of ultramafic composition is discussed below, and it is now clear that these 'primitive' magmas do not have intrinsically elevated gold contents, although they do possess slightly lower base metal abundances than mafic to felsic counterparts (Table I). Hodgson and MacGeehan (1982, p. 225) have suggested that the preferred disposition of gold deposits at the contact of major volcanic and sedimentary sequences may be somehow related to the presence of sedimentary rocks into which gold has been preconcentrated. This conjecture is difficult to reconcile with the observation that major classes of base metal ore deposits are hosted by sedimentary sequences (Gustafson and Williams, 1981; Bjorlykke and Sangster, 1981), and that typical greenstone-belt turbidites in proximity to and remote from gold deposits have neither inherently elevated gold abundances nor low base metal contents relative to average crustal background values (Table I).

2. Selective Precipitation of Gold

It is possible that the auriferous mineralizing fluids do contain significant concentrations of base metals, but that formation of base metal sulphides is inhibited by low availability of reduced S and/or the absence of appropriate redox gradients, whereas gold precipitation is not. Such specialized conditions are difficult to reconcile with the fact that there are gold-bearing vein deposits with high contents of sulphur and reduced carbon, along with examples having low contents of these reduced species, as well as prominent example where higher redox states are indicated by the presence of magnetite (cf. the Macassa Mine - Kirkland Lake). Locally, base metal sulphides are present, albeit in minor proportions; hence their formation is clearly not excluded by the ambient chemical environments of gold deposition, and an alternative explanation is required for their overall low content in the deposits.

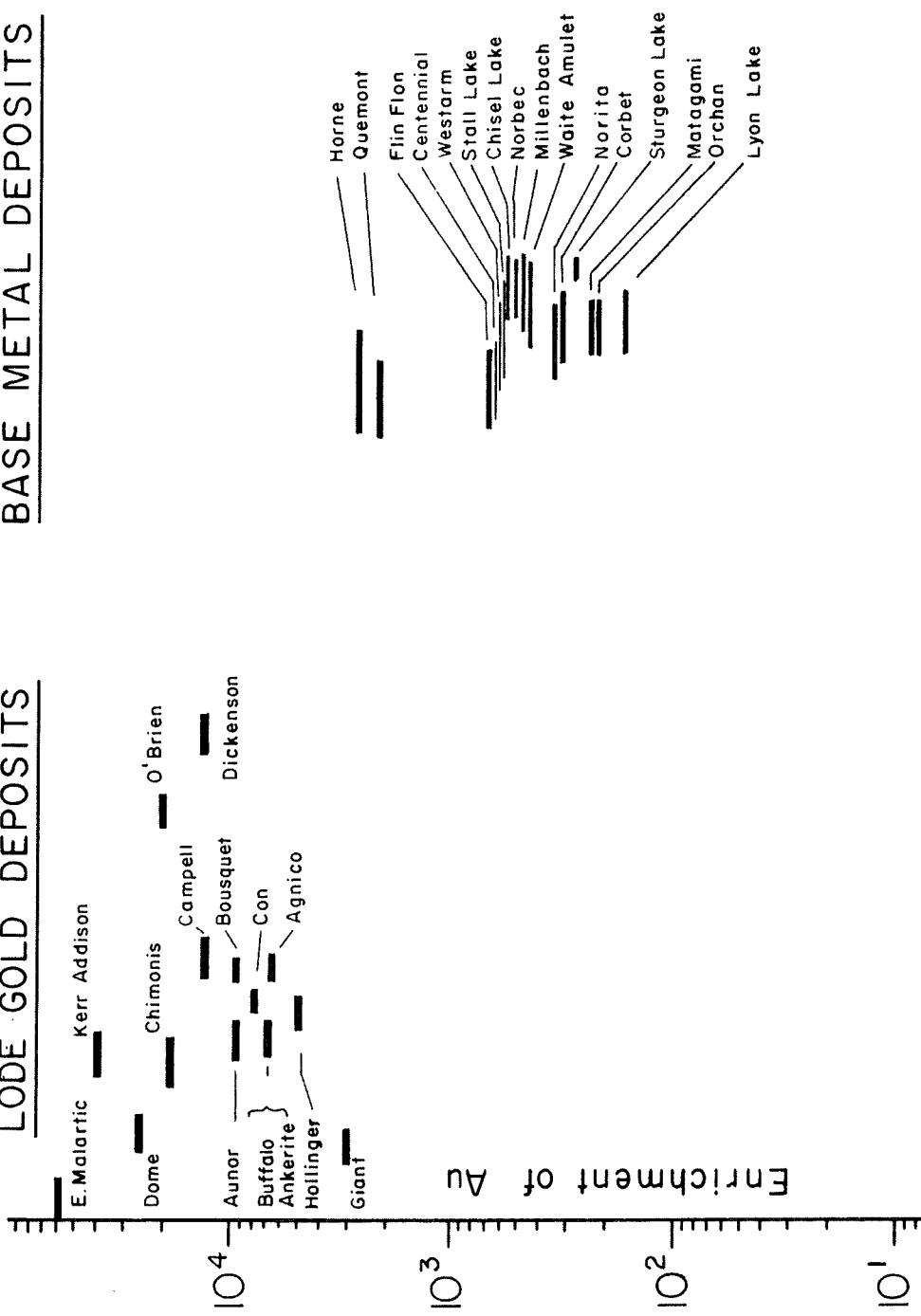
Some Tertiary epithermal precious metal deposits of the western USA typically contain distinct base metal and gold-rich domains in a given vein system, a notable example of which is the Sunnyside Mine, Colorado. In this instance, separation of gold and the base metals via differing precipitation mechanisms seems likely (cf. White, 1981; Barnes, 1979).

3. Special Properties of the Hydrothermal System plus Special Ligands

Separation mechanisms which appeal to properties of hydrothermal fluids have centred on the possible role of differences in the water/rock ratio of marine convective vs metamorphic-derived mineralizing fluid reservoirs (Kerrich, 1979; Kerrich and Fryer, 1979; Kerrich and Hodder, 1982). An essential feature of metamorphic fluids, which are generated under conditions of a low water-to-rock ratio, is the potential to separate rare from abundant elements. In order to illustrate this process, consider 100 grams of metabasalt with 2 ppb Au and 100 ppm Cu +

LODGE GOLD DEPOSITS

BASE METAL DEPOSITS



Enrichment of Cu plus enrichment of Zn

10⁻¹ | 10⁰ | 10¹ | 10² | 10³ | 10⁴

10⁻¹ | 10⁰ | 10¹ | 10² | 10³ | 10⁴

Figure 4 : Metal separation in lode gold vs. massive base metal sulphide deposits. Enrichment of Au versus the enrichment of Cu and the enrichment of Zn, for lode gold and base metal massive sulphide deposits. All enrichments calculated relative to background abundances in primary unaltered igneous rocks. The horizontal spread reflects enrichments computed utilizing background values for both mafic and felsic igneous rocks (cf. Fig. 2).

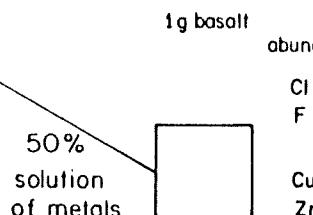
METAL SEPARATION IN DIFFERENT HYDROLOGICAL REGIMES

A SEAWATER CONVECTION

10 g water concentration in solution		
Cl	19,000 ppm	from sea
F	1.3	
Cu	4.3	
Zn	5.2	from rock
Pb	0.3	
Ag	0.005	
Au	0.0001	

WATER/ROCK \approx 10

Zn-Cu Deposits



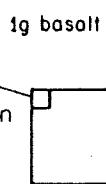
Σ HALOGENS \gg Σ METALS

B METAMORPHIC OUTGASSING

WATER/ROCK \approx 0.05

Au Deposits

0.05 g water concentration in solution		
50% solution of all species	selective fractionation of base metals out	
Cl	600 ppm	x 1 \longrightarrow 600
F	4000	x 1 \longrightarrow 4000
Cu	870	x 0.005 \longrightarrow 4.3
Zn	1050	x 0.005 \longrightarrow 5.2
Pb	60	x 0.005 \longrightarrow 0.3
Ag	1	x 1 \longrightarrow 1
Au	0.02	x 1 \longrightarrow 0.02



Σ CHLORINE $<$ Σ BASE METALS

Figure 5

: Schematic diagram illustrating the influence of water/rock ratio on separation of base metals and Au from source rock into solution. In A, water $>$ rock, and halogens \gg total dissolved metals. In B, rock $>>$ water, and halogen abundance is low. If the same proportion of base metals went into solution for B as in A (50%), base metals would exceed their solubility (870 ppm Cu, 1050 ppm Zn) and Cl would be deficient. If 0.005 of this amount actually partitioned into solution, the base metal solute concentrations would be restored to abundances as in A. No such solubility constraint is imposed on certain rare elements, hence uptake of these into solution is enhanced, whereas uptake of base metals is suppressed.

Pb + Zn. If 50 per cent of the metals are taken into solution as the rock evolves 5 weight per cent volatiles at the zeolite- greenschist or greenschist-amphibolite transition (water/rock ~ 0.05 , Fyfe *et al.*, 1978), then the solute concentration of Au will be 20 nano-grams ml $^{-1}$ (20×10^{-9} g per ml) and 1 mg ml $^{-1}$ base metals. Although the Au solute concentration is comparable to that reported for natural geothermal systems (Tilling *et al.*, 1973), the calculated figure for base metals is 100 times greater (Andrews and Fyfe, 1976) and implies that less than 1 per cent of the available base metals would be taken into solution. Therefore, in low water/rock systems the absolute abundance of rare elements in solution is not constrained by solubility, whereas base metal solute concentration is (Fig. 5).

This mechanism provides some degree of separation of the rare elements from abundant base metals in solution, but the metamorphic fluid would still be saturated with respect to Cu, Pb, and Zn, and it is necessary to postulate either a still lower intrinsic base metal content in the fluid, and/or a process for selectively precipitating rare elements.

Consider further the partitioning of elements from rocks into the fluid phase during metamorphic dehydration reactions. Following the reasoning above, a metabasalt with an average content of 60 ppm Cl and 400 ppm F (Turekian and Wedepohl, 1961) would provide 600 $\mu\text{g}/\text{ml}$ Cl and 4,000 $\mu\text{g}/\text{ml}$ F in solution, comparable to the estimated base metal solute concentrations. Inspection of alteration envelopes of auriferous veins (Bain, 1933; Kerrich and Fyfe, 1981) reveals that the most abundant cations added are typically K and Fe. These predominant metal cations in the hydrothermal fluid, in addition to the other cations present, such as Au, Ag, Cu, Pb, and Zn, will compete for the available halogens. If the rare elements compete more effectively for the limited halogens available for complexing, then this constraint may be a second factor suppressing base metal solute concentrations relative to the rare elements (Fig. 5).

Conversely, in submarine hydrothermal convective systems which generate massive base metal sulphide deposits, water-rock ratios may be of the order of 10 to 100 (Wolery and Sleep, 1976; Spooner *et al.*, 1977) and thus 200 to 2 000 times greater than for lode gold deposits. Cl is present at 19 000 ppm in seawater, or ~ 30 times the level calculated for metamorphic fluids. Here the solute abundances of base and precious metals are not constrained by solubility or availability of halogens for complexing, and probably broadly reflect background levels, with no large separations of elements in the hydrothermal solvent ore deposit. For example, in typical Archaean stratabound base metal massive sulphide deposits, such as Mattagami Lake Mines, and Millenbach or Corbet in the Noranda District, Zn and Cu are present in the low percentage range, silver at 10 to 50, and Au at 0.45 to 2 g metric ton $^{-1}$ (Worobec, 1980). This represents similar concentration factors for base metals (200 - 1 200), silver (200 - 1 100) relative to the background abundances (Fig. 2). Figures for the Zn, Cu, Ag, and Au contents of base metal massive sulphide ore deposits in the Iberian pyrite belt are in the same range (Fig. 3; Strauss *et al.*, 1977).

For lode gold deposits the ubiquitous presence of carbonate gangue requires hydrothermal solutions containing CO_2 , and this implies the possibility that gold may be transported in solution as a carbonyl (CO), carbonyl chloride, COS or related ligand (Kerrich and Fyfe, 1981; Fyfe and Kerrich, 1984). Several authors (cf. Bain, 1933) have observed the close linkage of gold content in veins to wall-rock type and intensity of wall-rock type and intensity of wall-rock alteration: this raises the possibility that gold precipitation may be mediated by reactions between fluid and wall rock. During carbonate-forming reactions, CO_2 is scavenged from solutions to form iron-magnesium-calcium carbonate minerals from Fe, Mg, Ca-silicate minerals in wall rocks, and one effect of these reactions would be to break down a proportion of carbonate complexes in solution, lowering P_{CO_2} , and hence forcing precipitation of the complexed species. It has also been suggested recently that in certain hydrothermal tungsten deposits of the greisen type, W may be complexed with CO^2- in solution (Higgins, 1980).

The higher ambient temperatures of hydrothermal gold systems compared to submarine geothermal convection, together with the possibility of ligands such as CO in the former, may promote aqueous transport of some relatively insoluble metals. Thus, the extent of enrichment of W, Cr, Ni, and Pd documented above for lode gold deposits is much greater than is typical for massive base metal sulphide ore bodies.

Several additional factors, including local perturbations of the chemical or physical environment, undoubtedly govern to some extent the separation of elements in ore deposits. Such factors may have controlled the recorded variations in abundance of B, As, Cr, Ni, and Hg between individual ore bodies as well as the sporadic occurrence of Mo plus graphite. Exceptions may represent anomalous concentrations in the source regions, mixing of different hydrothermal reservoirs, variations in the relative proportions of complexing agents, availability of reduced S,

or many other possible factors. There is, however, an over-all coherence to the patterns of enrichments in lode gold deposits, on the one hand, and of seafloor base metal massive sulphide deposits on the other hand (Fig. 4).

The observed distributions of elements could be accounted for in terms of two different hydrological regimes: a lower-temperature seawater system with a high water-rock ratio and abundant halogens, in which base metals, silver, and gold are concentrated to similar extents; and conversely, a high-temperature metamorphic system where the combination of a low water-rock ratio with a limited supply of halogens results in selective partitioning of certain rare elements into the fluid phase relative to abundant and mobile base metals. This interpretation is consistent with isotopic data for the temperature and fluid sources of the two systems and with geological relationships of the deposit types.

Irrespective of the metal complexing agents, a low water/rock in the source region coupled with low chloride would serve to suppress uptake of abundant base metals into solution, resulting in a preferential concentration of certain rare elements into the hydrothermal reservoir.

4. Special Ligands in Near-Neutral, Low-S Fluids

Lydon (1980) has proposed that in solutions of neutral to slightly alkaline pH, base metal and gold-thiocomplexes are the dominant solute species. If the total sulphur content of the solution is low ($< 10,000$ ppm), then according to Lydon (1980) the extent of base metal thiocomplex formation is insignificant, whereas that of gold is still appreciable (e.g. base metals $\gg 1$ ppm, Au 1-2 ppm in solutions of $C_1 = 3,0$ m, $pH = 6,5$ $\log [S]$ reduced/S oxidized] = 10,0, and at $200^\circ C$; Lydon, 1980, Fig. 21). Phillips and Groves (1983) suggested that $HAu(HS)_2$ complexes are the most likely candidates for hydrothermal transport of gold, and further that such ligands would selectively fractionate gold relative to base metals in solution.

Any such ligand-specific mechanisms for gold deposits must not only account for the separation of gold and base metals, but must also explain cosolution of the gold-associated elements such as Ag, As, Sb, B, W, Se, Te, and Bi. Furthermore, the specified gold-transporting complex must be stable in neutral to slightly acidic fluids, as signified by the muscovite-dominated alteration assemblages (cf. Boyle, 1961; Kerrich and Fye, 1981; Rose and Burt, 1979).

It should be noted that under intermediate to high temperature-pressure conditions the concentrations of sulphur-bearing species in equilibrium with Fe^{2+} -bearing minerals is very low (cf. Carmichael *et al.*, 1974, p. 306; Kerrich *et al.*, 1977). However, given the low estimated gold solute concentrations (~ 20 ppb) in hydrothermal fluids, along with the absolute introduction of sulphur into vein deposits, molecules incorporating sulphur may provide a significant gold transporting ligand.

III. SOURCE OF VOLATILE COMPONENTS IN FAULT ZONES

The source, temperature, and magnitude of aqueous fluid reservoirs participating in the gold deposits may be evaluated from oxygen and hydrogen isotope data. Five distinct terrestrial fluid reservoirs have been identified, based on their oxygen and hydrogen isotope compositional ranges. These are marine, meteoric, magmatic, and metamorphic fluids, together with formation brines that generally contain a large component of 'evolved' meteoric recharge waters (Taylor, 1974; Fig. 6). Oxygen isotope studies also permit estimates of ambient mineral and fluid temperatures at equilibrium, given that the fractionation of oxygen isotopes between coexisting silicates or metal oxides is a function of temperature alone (Javoy, 1977; Friedman and O'Neill, 1977). Problems of identifying fluid flow through rocks are particularly amenable to evaluation by oxygen isotope analyses, since isotopic exchange between rocks and an aqueous reservoir is reflected in shifts of $\delta^{18}O$ rock or mineral from an initial value by a magnitude that depends on the $\delta^{18}O$ fluid, ambient temperature, and the fluid to rock ratio (Taylor, 1974, 1979).

A compilation of oxygen isotope data for hydrothermal silicates from mineralized domains is given in Part I, and calculated fluid isotopic compositions are summarized in Fig. 6.

A significant feature of the data is that the oxygen isotope composition of quartz, muscovite, and chlorite from individual auriferous vein systems lies within narrow intervals (Table I, Part I), signifying uniform hydrothermal conditions. For example at the Dome Mine, all of the vein quartz from five different ore bodies has a restricted range of $\delta^{18}O$ values at 14 to 15,2 per mil: Δ (quartz-muscovite) and (quartz-chlorite) remain approximately constant at 3,5 to 4,0 per mil and 5,5 to 6,1 per mil, respectively (Table I, Part I). The mineral

pair fractionations correspond to isotopic temperatures of 270°C to 450°C and a calculated $\delta^{18}\text{O}$ fluid of 6 to 8 per mil. All of the individual vein systems are broadly similar in terms of $\delta^{18}\text{O}$ (quartz), fluid temperature, and fluid isotopic composition, implying some common characteristics in the ore-forming process. Fluid pressures are estimated at 2-3 kb, equivalent to crustal depths of 8-12 km.

Considering the data for auriferous veins from twelve different mines collectively, the generally estimated temperature interval for gold precipitation is 270° - 450°C, with $\delta^{18}\text{O}$ (fluid) ranging from 6 to 9 per mil, and δD (fluid) from -80 to -40 per mil. The calculated or measured fluid oxygen and hydrogen isotope values are consistent with the range of most fluids implicated in metamorphism or derived from magmas (Taylor, 1974), but are isotopically heavy with respect to $\delta^{18}\text{O}$ and light for δD compared to seawater that has undergone either low- or high-temperature exchange with the oceanic crust. In most vein systems solutions are slightly enriched in $\delta^{18}\text{O}$ relative to primary magmatic fluids, but a magmatic component cannot be ruled out. The inferred metamorphic fluids, with a possible magmatic component, are considered to be generated during progressive accumulation and burial of the volcanic sequences, and hence were broadly contemporaneous with volcanic-plutonic activity.

The observed narrow range of $\delta^{18}\text{O}$ -values at 11 to 15 per mil for quartz in auriferous veins is independent of unaltered wall-rock $\delta^{18}\text{O}$, signifying a fluid-dominated system. A corollary to the above results is that lateral diffusion of silica, gold, etc. (cf. Boyle, 1961, 1979) from wall rock into veins, or remobilization of gold, etc., from nearby sediments, i.e. a rock-dominated system, is incommensurate with the isotopic relations.

In the Timmins, Matheson, Kirkland Lake, and Val d'Or regions of the fault systems, sulphates occur close to or within the structures. The sulphates, typically anhydrite or selenite, coexist with albite, Mg-chlorite, calcite, and hematite. On the basis of stable isotope data and mineralogical considerations these domains have been interpreted as sites for incursion of Archaean marine water into submarine volcanic rocks (Kerrich and Hodder, 1982; Kerrich and Fyfe, 1981; Kerrich and Watson, 1984): ambient temperatures were 150 to 200°C, and $\delta^{18}\text{O}$ fluid = + 1.5 per mil (Fig. 6).

A third fluid regime is evident from stable isotope data in plutonic rocks of the Kirkland Lake and Noranda districts. In these districts quartz-magnetite and quartz-chlorite veins were precipitated at temperatures 120-250°C from low- $\delta^{18}\text{O}$ meteoric water (Fig. 6).

Thus cooling of the volcanic-plutonic complexes was initially dominated by ocean water, in the submarine environment, with a transition to meteoric water cooling as the Archaean greenstone terrane emerged above sea level. This scheme is corroborated by the presence of late fluvial clastic sediments.

A. Implications of the Isotopic Data

Two important constraints on the nature of Archaean gold-forming hydrothermal systems are provided by the isotopic data. First, the disparity in $\delta^{18}\text{O}$ between veins and unmineralized wall rock contraindicates suggestions that the isotopic data record an overprint of the late putative metamorphism, and for this reason yield a 'metamorphic' signature. Rather, field and petrographic relations demonstrate that most Archaean gold deposits postdate any early marine water induced "spilitic", or regional metamorphism of the volcanic sequences, and this observation has been known from the time of the classic work of Knopf (1929) and Bain (1933). In addition, the preservation of a distinct meteoric water isotopic signature, as well as that of marine water-related-alteration and the main gold-forming hydrothermal event (Fig. 6), rules out any pervasive isotopic overprinting.

Second, the restricted range of calculated temperature, $\delta^{18}\text{O}$ and $\delta\text{D}_{\text{H}_2\text{O}}$, implies a uniform hydrothermal reservoir or reservoirs, isotopically buffered by a large source volume. This reservoir cannot have mixed with distinct fluid sources, and was thus hydrologically isolated from the hydrosphere, as independently deduced from the high fluid pressures implied by hydrofracturing (Part I). Such relations are in contrast to the isotopic characteristics of base metal massive sulphide or copper porphyry deposits where the incursion of an external marine or meteoric water source to the main ore reservoir, induces large dispersions in the isotopic composition of minerals (cf. Taylor, 1974, 1979).

B. Data from Fluid Inclusions

The dissolved solid content of ore-forming fluids may be estimated by measuring the melting point of aqueous liquid inclusions that have been frozen, and such 'salinities' are generally expressed as wt % NaCl equivalent

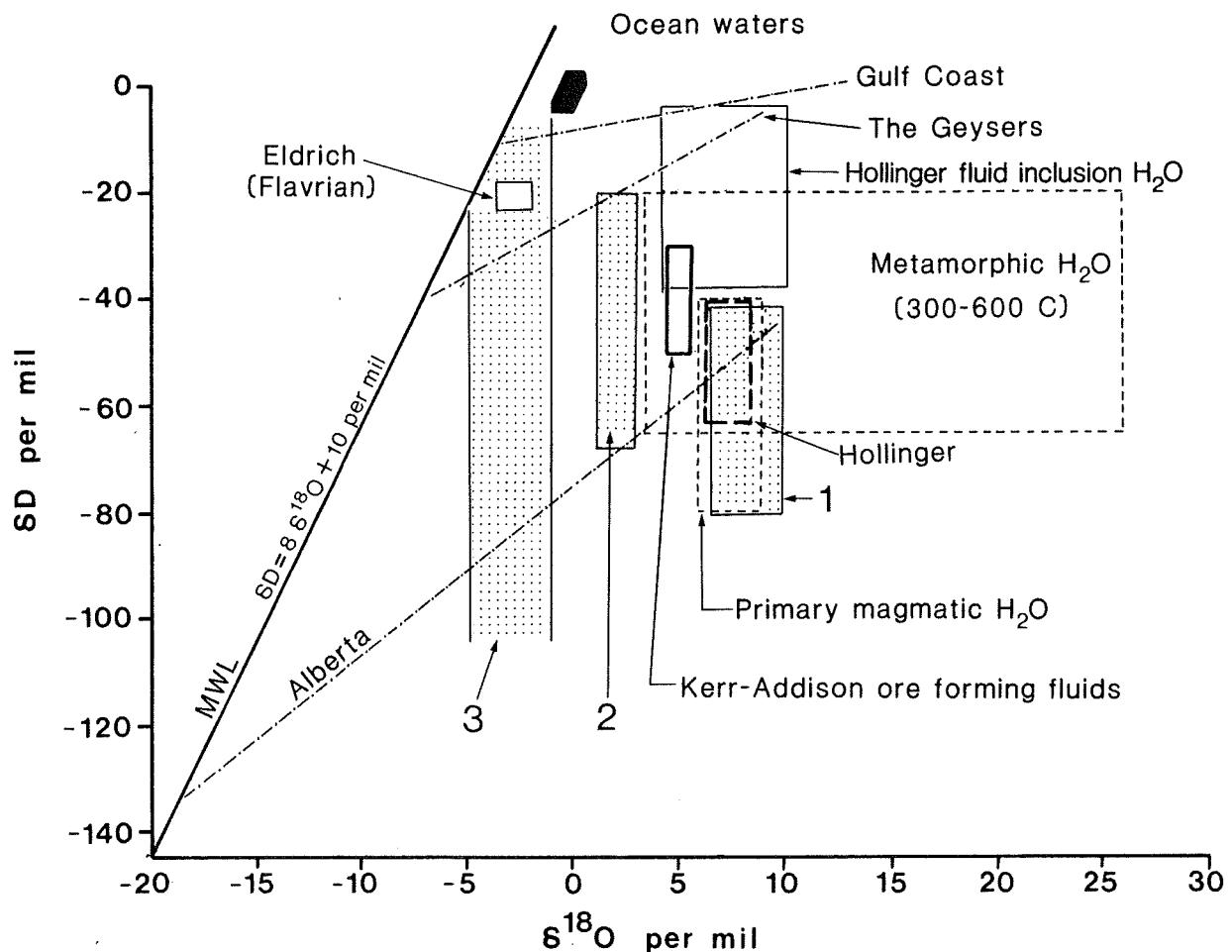


Figure 6 : Isotopic composition of natural terrestrial fluid reservoirs in δD , $\delta^{18}\text{O}$ co-ordinate space modified from Taylor (1974). Superimposed are calculated fields of 3 vein stages at Macassa, 1 ore solutions, 2 fluids associated with selenite deposition in faults, 3 late stage quartz-magnetite stringers (after Kerrich and Watson, 1984), Hollinger (this paper), Kerr Addison (Kishida, 1984), and Eldrich. Data for H_2O decrepitated from quartz veins of the Hollinger Mine is from Fyon et al., (1983).

(Roedder, 1979). CO_2 also acts to depress the melting point of frozen aqueous inclusions, and hence the true fluid salinity will, in general, be less than the measured value for CO_2 -bearing inclusions, which are ubiquitous in Archaean vein gold deposits. In $\text{H}_2\text{O}-\text{CO}_2$ inclusions, a clathrate phase, $\text{CO}_2 \cdot 5.75\text{H}_2\text{O}$, forms during cooling runs, which, when pure, melts at about +10°C (Collins, 1979). Solutes present in hydrothermal fluids also act to depress the clathrate melting temperature, providing an independent means of determining hydrothermal fluid salinities for the case of CO_2 -rich fluids. However, methane and solutes exert opposed effects on the clathrate melting behaviour: hence, melting-point data from liquid inclusions are not easy to interpret without knowledge of the gas species present.

In a review of fluid inclusion data on gold deposits, Roedder (1982, 1984c) concluded that the hydrothermal fluids are generally characterized by low salinities and high CO_2 contents. Inclusions in Archaean lode gold deposits also indicate rather dilute solutions of 0.5 to 2.6 wt % NaCl equivalent, well below the value of 3.6% for modern seawater (Guha *et al.*, 1982, Mine Doyon; Kerrich, 1983; see also Graves and Zentilli, 1982, and Roedder, 1972 for veins in metamorphic rocks).

In a detailed fluid inclusion study of the Hollinger Mine, Timmins District, Smith and Kesler (1985) have shown that fluids associated with gold mineralization were $\text{CO}_2-\text{H}_2\text{O}$ mixtures, where $\text{CO}_2 \sim 12$ mole %. Zoning peripheral to the domains of economic mineralization is characterized by diminishing grades, in conjunction with lower proportions of CO_2 in the fluids (1-2 mole %) and ankerite in alteration envelopes.

It is important to note that when marine water is trapped in pore spaces during the accumulation of rock sequences, the pore fluid will initially have a salinity close to that of seawater (~3.5% salinity, ~19,000 ppm Cl). Pore fluids will tend to increase in salinity during progressive burial at low temperatures, as some of the water is consumed by participating in hydration reactions of anhydrous minerals in the rock, leaving the remaining pore fluid as a relatively more concentrated salt solution.

Once such saline fluids have been expelled during burial and compaction, any further evolution of fluids from the rocks will proceed by dehydration reactions at higher temperatures. Such metamorphic fluids, in contrast to seawater or expelled pore water, will tend to have a relatively low chlorine content, because of the low chlorine abundance of rocks at ~60 ppm (Turekian and Wedepohl, 1961) coupled with the stability of chlorine in amphibole. Evidence for some saline fluids is indicated by the presence of halite crystals in liquid inclusions from late-stage, post-Au quartz veins at the Con and O'Brien mines (Boyle, 1961; Krupka *et al.*, 1977, respectively). These veins with saline fluid inclusions could arise from incursion of lower-pressure hypersaline formation brines into a fracture array induced by discharge of a high-pressure, low-salinity, auriferous metamorphic fluid reservoir, after decay in the hydraulic pressure of the latter. Alternatively, such saline fluids could represent condensates of a supercritical fluid, into which solutes are preferentially partitioned, leaving a vapour phase possessing low solute concentrations.

In addition to ubiquitous CO_2 , methane is a well-known product of fluid inclusion rupture in the Con Mine, Yellowknife, and other major Archaean deposits. Smith and Kesler (1985) have identified CH_4 in fluid inclusions in vein quartz at the Hollinger Mine where such veins intersect graphitic and/or carbonaceous units. Guha *et al.* (1982) have identified several generations of liquid inclusions at Mine Doyon (Silverstack), Quebec, of generally low salinity, containing 16 to 95% CO_2 and methane. Immiscibility of CO_2 , identified from highly variable $\text{CO}_2/\text{H}_2\text{O}$ ratios, appears to have accompanied gold precipitation in some examples (Kerrich, 1983; Colvine *et al.*, 1984; see also Higgins and Kerrich, 1982).

Modern seawater has Na/K of about 28. Analysis of liquid inclusions from vein stockworks beneath base metal massive sulphide deposits, which are generally thought to have formed from modified seawater, reveals that $\text{Na} \gg \text{K}$ in accord with the proposed dominant seawater origin of mineralizing fluids (Franklin *et al.*, 1981). Data for Archaean gold-bearing vein deposits are available only for the O'Brien Mine, Quebec, where Krupka *et al.* (1977) report $\text{Na}/\text{K} \sim 1$ for liquid inclusions in quartz associated with the main stage of mineralization. A more plausible value is ~ 10, based on independent considerations (see below). In general, data from liquid inclusions indicate that high-temperature fluids have much smaller Na/K than seawater (Roedder, 1972).

The greater proportion of K to Na, and of CO_2 and H_2O in high-temperature fluid sources compared to seawater, could account for the general potassio- CO_2 addition to rocks enveloping gold veins. This is in contradistinction to the dominant Fe, Mg, Si, S fixation in footwall rocks to base metal massive sulphide deposits.

A summary of fluid inclusion data for Archaean lode gold deposits is given by Colvine *et al.* (1984, Table 6). In general, the fluids were characterized by dense CO₂ (0.8 g/cm³), filling temperatures of 200-400°C, and variable CO₂/H₂O mixtures. According to Colvine *et al.* (1984) the variable CO₂/H₂O ratios may be accounted for by several mechanisms including mixing of CO₂ and H₂O-rich fluids, or immiscible separation of CO₂ from a CO₂-H₂O mixture.

IV. MINERALOGICAL CONSTRAINTS ON FLUID CHEMISTRY

Mineral assemblages of metasomatized fault zones vary considerably, but in general the characteristic species are quartz plus carbonate, along with subordinate muscovite, albite, chlorite, talc, tourmaline, scheelite, and sulphides or arsenides. The relative proportions of minerals and their composition depend critically on the lithology faults intersect. For instance where structures traverse granitic rocks the dominant carbonate is calcite, whereas in tholeiitic basalts and ultramafic rocks ferroan dolomite and magnesite predominate, respectively. Similarly, hydrothermal chlorite and muscovite may contain several per cent Cr in faults hosted by ultramafic flows, whereas their counterparts in structures propagating through felsic rocks possess low Cr abundances. A summary of diagnostic alteration mineral assemblages in lode gold deposits of Canada is reported by Colvine *et al.* (1984, Tables I, II, IV), classified according to area, lithology, and hydrothermal-metamorphic grade.

In general, hydrothermal fluids venting through fault zones donated volatiles and potassium, with wall rocks contributing bivalent metal cations. Thus hydrothermal carbonates reflect the carbonation of Fe, Mg, Ca-silicate precursors such as chlorite + epidote to Fe, Mg, Ca carbonates; hydrothermal muscovite or K-feldspar represent the hydrolysis of albite + chlorite with aqueous K addition; and pyrite, the transfer of Fe from a silicate or oxide reservoir to sulphide minerals accompanying the consumption of aqueous sulphur species.

Progressive mineralogical reactions of these types, involved in hydrothermal alteration of Fe-tholeiites and ultramafic rocks along the Kirkland Lake fault zone, have been studied in detail by Kishida (1984). Such reactions can be evaluated in terms of T, X_{CO_2} , $a(K^+/H^+)$ and $a(Na^+/H^+)$, according to equilibria given by Montoya and Hemley (1975). Two reactions are considered here, namely chlorite + calcite + CO₂ + ankerite + muscovite and quartz; and chlorite + muscovite + albite (Fig. 7a,b). Reaction 1 depends on small changes of X_{CO_2} and $a(K^+/H^+)$, whereas reaction 2 is mediated by $a(K^+/H^+)$ but is independent of X_{CO_2} . Assuming a temperature of 300°C coupled with isothermal conditions an $a(Na^+/K^+)$ of $\geq 10:1$ can be deduced for carbonate + albite + muscovite alteration (cf. Montoya and Hemley, 1975).

From the above consideration it can be deduced that fluids responsible for the specified hydrothermal mineral assemblages along the Kirkland Lake fault were characterized by $X_{CO_2} \sim 0.1$, a neutral to slightly acidic pH, and aqueous Na⁺/K⁺ $\sim 10:1$.

For gold deposits within the 2.8 \pm 0.1 Ga granitoid-greenstone terranes of Australia, Phillips and Groves (1983) have determined that the mineralizing solutions were H₂O + CO₂ mixtures, of slightly alkaline pH, low Eh and salinity, at $\sim 400^\circ C$.

In summary, fluids implicated in Archaean lode gold deposits possess low salinity, high CO₂ contents, reduced gas species such as methane, relatively high K/Na ~ 0.1 , and in some instances exhibit evidence of transient CO₂ immiscibility accompanying precipitation of gold.

V. PATTERNS OF OXIDATION-REDUCTION REACTIONS

Reduction of iron in rocks adjacent to auriferous veins was noted by Bain (1933). Inasmuch as rocks are generally resistant to changes of oxidation state during metamorphism, the observed dominance of Fe²⁺ requires the introduction of large volumes of reductant in the auriferous hydrothermal solutions, implying high fluid fluxes through fracture conduits. It is probable that the reductant involved is H₂, which is a product of the dissociation of water in equilibrium with the QFM buffer system (cf. basalts) at elevated temperatures (Eugster, 1959; Eugster and Wones, 1962; Eugster and Skippen, 1967).

The extreme reduction observed in wall rocks implies extensive interaction with fluids originating under high-temperature reducing conditions. Note that for aqueous fluids at high temperatures, the equilibrium $P_{H_2} \gg P_{O_2}$ on account of reaction of the type Fe²⁺-silicate + H₂O + Fe₃O₄ + silicate + H₂. Hydrogen fixation in wall rocks takes place in response to the difference in equilibrium H₂ pressure at decreasing temperature-pressure states, as

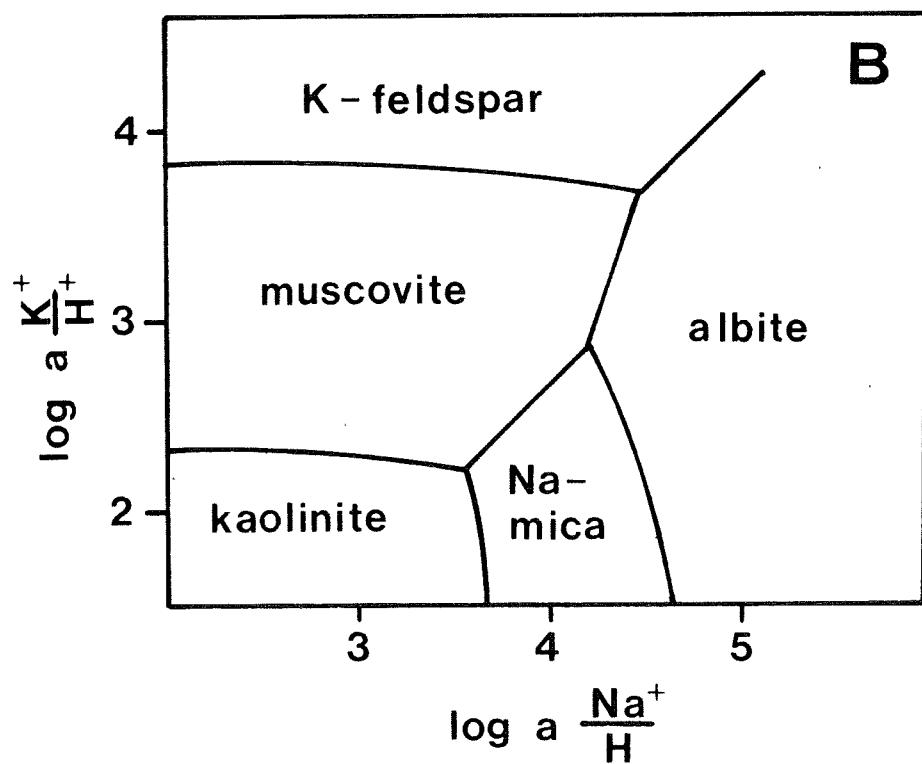
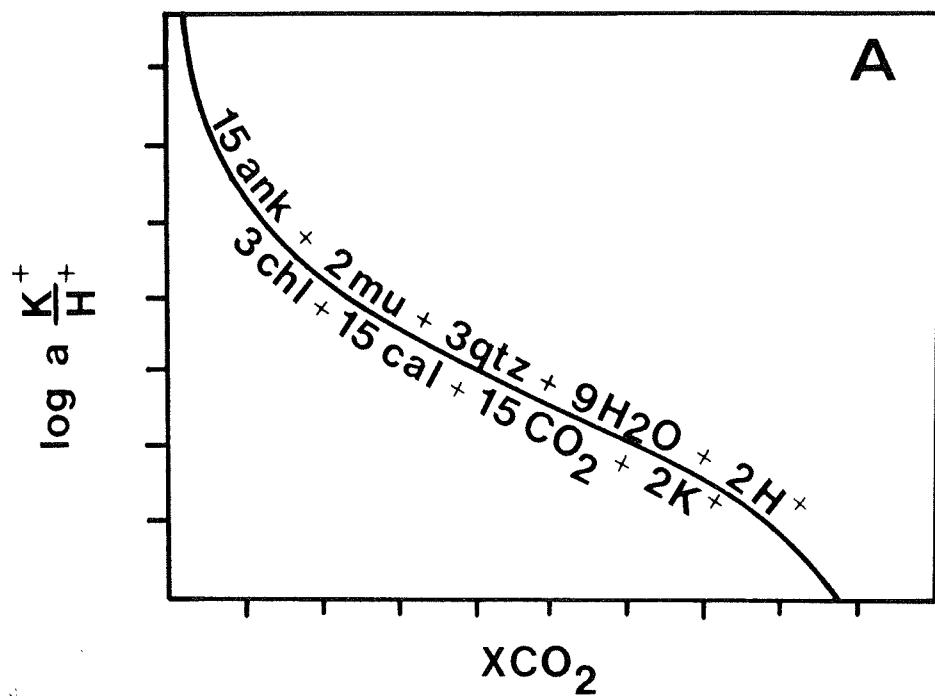


Figure 7A,B: Equilibria relations for specified reactions (after Montoya and Hemley, 1979).

the mineralizing solutions ascend through the crust. Estimates of minimum fluid/rock ratios obtained from shifts in the oxidation state are >3:1 (see Kerrich *et al.*, 1977). REE distributions in veins and wall rocks to veins exhibit marked positive Eu²⁺ anomalies, which, along with the presence of methane and hydrocarbons, provide independent evidence for chemical exchange with fluids of low redox potential (see Kerrich, 1983, Table 17; Kerrich and Fryer, 1979).

Locally, where vein systems intersect carbonaceous sediments, such as in stope 15-26-16 of the Schumacher Mine at the Pearl Lake porphyry, or in Carlin-type deposits (Radke and Scheiner, 1970), reduced carbon may act as an additional reductant or precipitant, as well as generating methane.

In contrast to the pattern of reduction accompanying Au-mineralization described above, some sediments and their enveloping rocks are significantly oxidized. For instance, the felsic relative to background (Kerrich, 1983), and has high contents of sodium, thus exhibiting chemical features typical of low-temperature, ocean-floor spilites (Fyfe and Lonsdale, 1980; Munha *et al.*, 1980). Patterns of oxidation with concomitant albitization have also been recorded from portions of the Lamaque granodiorite, Pearl Lake porphyry and Preston porphyry, where hematite + carbonate + anhydrite are locally abundant. As discussed in a following section, the presence of both reductive-K₂O and oxidative Na₂O patterns of alteration in the seafloor environment around lode deposits is probably due to the interplay of two distinct fluid reservoirs: an early, shallow circulating, relatively oxidized marine water system driven by the thermal anomaly of subvolcanic stocks and their extrusive counterparts, in conjunction with a later reduced auriferous hydrothermal reservoir.

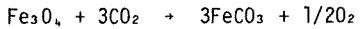
Perhaps one of the clearest examples of the interaction of oxidative and reductive fluid reservoirs in the near-marine environment is present at the Chadbourn Mine, Noranda, Quebec. Here, gold mineralization is confined to a subconically shaped domain of intense hydrofracturing, within Archaean andesites of the Blake River Group (~2,7, Ga). Calculations of chemical mass balance in which the coherent behaviour of Al, Sc and Zr are used to define volume change reveal two domains of alteration. Firstly, there is an outer zone, 250m wide, peripheral to the mineralized breccia, characterized by additions of Na₂O and H₂O with concomitant loss of K₂O, MgO, MnO, SiO₂ and some transition metals. Secondly, in the Au-breccia, massive fixation of CO₂, H₂O and K₂O is accompanied by depletion of MgO, CaO and Na₂O. The outer zone is oxidized ($Fe^{2+}/\Sigma Fe = 0,64 \pm 0,23 \text{ } \sigma$) and the inner zone reduced ($Fe^{2+}/\Sigma Fe = 0,9$) relative to background ($Fe^{2+}/\Sigma Fe = 0,7$). The presence of hydrofractures in all orientations signifies $P_{\text{fluid}} \geq P_{\text{load}}$ during mineralization. In the outer alteration zone, the $\delta^{18}\text{O}$ of quartz (8,9 to 12 per mil), albite (6 to 8,6 per mil) and chlorite (0,4 to 1,7 per mil) corresponds to a T_{fluid} of 220-280°C, and $\delta^{18}\text{O}_{\text{H}_2\text{O}} = 0 \pm 1$ per mil. Within the domain of gold mineralization, the $\delta^{18}\text{O}$ of quartz (12 to 13,5 per mil), chlorite (6,1 to 6,9 per mil) and magnetite (-0,8 to +1,9 per mil) defines a T_{fluid} of 300-380°C and $\delta^{18}\text{O} = 7 \pm 1$ per mil (Kerrich and Walker, 1981).

The data are interpreted in terms of forceful discharge of a low-redox, auriferous metamorphic fluid reservoir at lithostatic pressure, in which K > Na such that muscovite is stabilized over albite, and in which elevated CO₂ contributes to the hydrolysis of Fe, Mg, Ca-silicates to Fe, Mg, Ca-carbonates. The thermal anomaly arising from heat transport upward in the metamorphic plume forced local, shallow convection of oxidized marine water under hydrostatic pressure, in which Na > K such that albite was stabilized. Pressure differentials between the two reservoirs appear to have limited mixing.

Lister (1980) has addressed the question of the growth and decay of hydrothermal cells in the seafloor environment, with incursion of cool, oxidized water into the vent region during the waning stages of discharge.

Several important exceptions, to the generally reduced character of Archaean lode gold deposits are known. At the Macassa Mine - Kirkland Lake, and Canadian Arrow despost - Matheson, trivalent iron is present in magnetite, signifying relatively more oxidizing conditions during precipitation. Lambert *et al.* (1983) and Schwarz and Rees (1984) have noted that where more oxidizing conditions of mineralization occurred, the $\delta^{34}\text{S}$ of pyrite is depleted relative to pyrite in the majority of other deposits. For instance at the Golden Mile deposit, Australia, $\delta^{34}\text{S}$ pyrite = -1 to -7, compared to +1 to +4 per mil for populations of pyrite from most other deposits, and this reflects the redox-dependent fractionation of S-isotopes during sulphide precipitation (Ohmoto and Rye, 1979).

According to Lambert *et al.* (1983) oxidation of fluids may be in response to reactions such as:



and



The $\delta^{18}\text{O}$ of vein quartz at Macassa and Canadian Arrow is within the range of lode gold deposits ensemble, implying that in terms of oxygen isotopic composition and temperature, hydrothermal fluids at these two deposits did not differ significantly from gold deposits in general, although the inferred redox state was higher.

VI. FLUID VOLUMES - SOLUTE CONCENTRATIONS

Total production of gold from the Porcupine District up to 1976 was $1,7 \times 10^9\text{g}$ (Pyke, 1976). If it is assumed that this mass of gold was derived by 50% efficient leaching of metabasalts with an average Au abundance of 2 ppb (Tilling *et al.*, 1973; Kwong and Crocket, 1978), then the source volume would be of the order of 600km^3 (1km^3 of rock contains $5,6 \times 10^6\text{g}$ Au at 2 ppb). If hydrothermal fluids are derived from the freezing of magmas, then similar volumes of magmatic rock would be required. Based on the premise that Si, Au, Ag, Pd, W, Ni, and Cr together with other elements present in veins, were leached and transported during metamorphic outgassing within the greenschist facies, or at the greenschist-amphibolite transition, when approximately 5 wt % structural water is released from a hydrated rock of basaltic composition, then the solvent volume would be of the order of 90km^3 .

An independent estimate of the hydrothermal fluid volume may be derived as follows using gold-bearing quartz veins, which are essentially pure quartz, as examples. In cooling from 550°C to 300°C , 1kg of quartz-saturated water will precipitate 2,6g of quartz, assuming a thermal gradient of $100^\circ\text{C km}^{-1}$, and an hydraulic gradient equal to the lithostatic gradient (Holland, 1968). Gold is present in the veins at an average level of 10 ppm, and therefore if Au and quartz were coprecipitated from a solution cooling along a geothermobar the 1kg of solvent would precipitate $2,6 \times 10^{-5}\text{g}$ Au and the average decrease in concentration of gold in the hydrothermal solution would be 25 ng ml^{-1} . The total carrier fluid volume for $1,7 \times 10^9\text{g}$ Au is then $65 \times 10^{12}\text{kg}$, or 65km^3 , based on the generalized assumption that gold is dominantly precipitated over the stated temperature interval. Estimates of fluid volume are slightly increased if a lower thermal gradient is assumed or if CO_2 consumption from the hydrothermal fluid acts to decrease the rate of change of silica solubility during cooling. Further, consider 100g of metabasalt containing Au at 2 ppb. If 50% of the gold is taken into solution as the metabasalt evolves 5 wt % water at the greenschist-amphibolite transition, then the solute concentration will be 20 ng ml^{-1} Au (20 ppb), which is comparable to the figure estimated above and the value of 50 ng ml^{-1} reported by Tilling *et al.* (1973) for gold in natural thermal waters. These calculations are intended to demonstrate the magnitude of source and fluid volumes, and provide estimates of Au solute concentrations. Independent estimates for gold transport in solution are given by Helgeson and Garrels (1968) and Fyfe and Henley (1973).

Seward (1979) reports deposition of ore-grade gold material, along with Ag, Sb, Hg, and Tl, from dilute fluids with ~ 1000 ppm chloride, 0,04 ppb Au, and 0,7 ppb Ag, in the Wairiki geothermal system, New Zealand. For fluids carrying 0,04 ppb Au, 800km^3 of hydrothermal solution would contain a million ounces of gold. Although these solute concentrations for gold in a modern system are about one hundred times lower than the maximum values calculated above for the higher-temperature Archaean vein deposits, and the values determined by Tilling *et al.* (1973), they serve to illustrate that high metal concentrations in hydrothermal fluids are not necessary to form ore-grade deposits.

VII. GOLD CARBONATE-POTASSIUM ASSOCIATION

A ubiquitous feature of gold-bearing vein deposits in Archaean greenstone belts is the presence of abundant carbonate minerals. Quartz and carbonates are the dominant gangue in hydrothermal veins, and are typically present in the extensive domains of alteration which envelop veins.

The nature of wall-rock alteration in proximity to Au-bearing veins has been comprehensively documented by Bain (1933) for studies of lode deposits in Ontario. Bain identified carbonate, chlorite, and sericite as the predominant alteration minerals. Extensive development of carbonate mineral in the Bendigo and Mother Lode systems was reported by Lindgren (1905) and Knopf (1929), respectively. The subject of carbonate alteration occurring in conjunction with gold vein deposits has been discussed by several workers, including Boyle (1961, 1976, 1979), Meyer and Hemley (1967), Rose and Burt (1979), Kerrich and Fyfe (1981), Kerrich (1983), and Colvine *et al.* (1984). In general, there is a consensus that reactions between CO_2 -rich hydrothermal fluids and Ca-, Fe-, Mg-, Mn-bearing silicates in the wall rocks are implicated in the formation of the carboante minerals.

Au-bearing deposits hosted in altered chemical sedimentary rocks, like their vein counterparts, typically contain a high proportion of carbonate minerals. At the Agnico Eagle Mine, Joutel, Quebec, gold is present in banded pyrite-siderite "iron formation" (Barnett and Hutchinson, 1976; Barnett *et al.*, 1982). In the Timmins District, one of the common ore types, the so-called 'ankerite veins', present in several mines including Dome, Paymaster, Aunor, Delnite, McIntyre, and Coniarum, is composed of a laminated ferroan-dolomite and chert sediment (Fryer and Hutchinson, 1976; Karvinen, 1982; Fryer *et al.*, 1979). Such auriferous carbonate sediments are typically bounded by flows or volcaniclastic rocks possessing pervasive carbonate alteration with attendant quartz-carbonate veining. This gold-carbonate association has attracted intense interest from the point of view of timing in the volcanic-sedimentary sequences, significance to gold precipitation, and potential as an exploration criterion.

It is generally accepted that metals and gangue constituents present in veins have been transported into the fracture arrays by hydrothermal solutions ascending through structurally controlled conduits in the crust, with precipitation occurring in response to cooling of the solvent along a geothermobar (Helgeson and Garrels, 1968; Fyfe and Henley, 1973; Holland, 1967). This premise is entirely consistent with predictions from solubility data, inasmuch as quartz, Au, Ag, and W (a metal association characteristic of these deposits) all have a positive temperature coefficient of solubility (Fyfe and Henley, 1973; Seward, 1976, Foster, 1977; Holland and Malinin, 1979). Other parameters, such as pH and wall-rock reactions, may also influence the solubility of some aqueous species (Meyer and Hamley, 1967).

The solubility relations of carbonate minerals are more complex than for quartz (Holland and Malinin, 1979). In general, calcite solubility decreases as a function of increasing temperature (T) at constant P_{CO_2} , or diminishes with decreasing P_{CO_2} , at constant T . The negative temperature coefficient of solubility dominates the positive effect of pressure, such that carbonates become less soluble with respect to increasing temperature, and carbonate precipitation is not anticipated from cooling fluids under typical crustal P-T cooling paths. Specialized conditions such as immiscible separation of CO_2 from an aqueous fluid are required for carbonate deposition from cooling solutions. Thus, the abundance of carbonate and its coprecipitation with quartz in Au-bearing vein deposits requires explanation.

In contrast, the distribution of quartz and carbonate minerals in convective hydrothermal systems is as predicted for simple temperature-solubility relations. Carbonate veins are present in the peripheral recharge areas where surface fluids heat during penetration into the crust, whereas quartz is the dominant gangue mineral in vein stockworks which represent the loci of hydrothermal discharge of fluids cooling during return flow to the surface (Andrews and Fyfe, 1976; Spooner *et al.*, 1977; Fyfe and Lonsdale, 1981; Meyer and Hemley, 1967). Examples of such hydrothermal systems include seawater circulation through hot basalts at mid-ocean ridges and convection of meteoric fluids in proximity of Cu-porphyry deposits (Lowell and Guilbert, 1970; Sheppard *et al.*, 1971; Wolery and Sleep, 1976; Fyfe and Lonsdale, 1981).

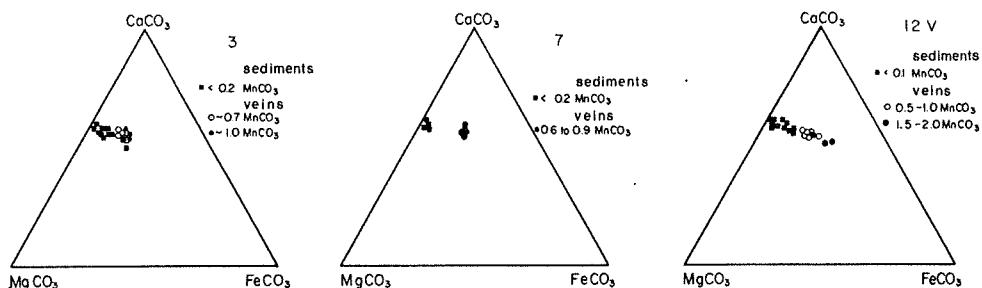
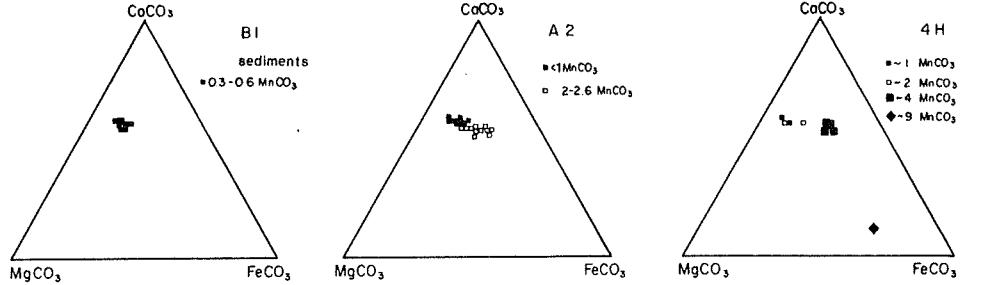
A similar distinction exists between gold deposits hosted by chemical sediments on the one hand, which typically contain abundant carbonate, such as at Timmins and Red Lake, Ontario, and Lapa Seca, Brazil; and, on the other hand, base-metal-bearing sediments in which the carbonate content is not generally significant (Franklin *et al.*, 1981).

Several explanations have been advanced to account for the precipitation of carbonate minerals from cooling fluids, as discussed by Holland (1967): additional processes are discussed in the section on the source of CO_2 . An important factor is believed to be boiling of solutions, which will take place during fluid ascent at the point when the confining pressure becomes less than the vapour pressure of the solution. This condition will be met in proximity to the terrestrial surface (<2km), and boiling of hydrothermal solutions has been identified during stages of calcite deposition at the shallow-level Providencia Mine, Mexico (Sawkins, 1964).

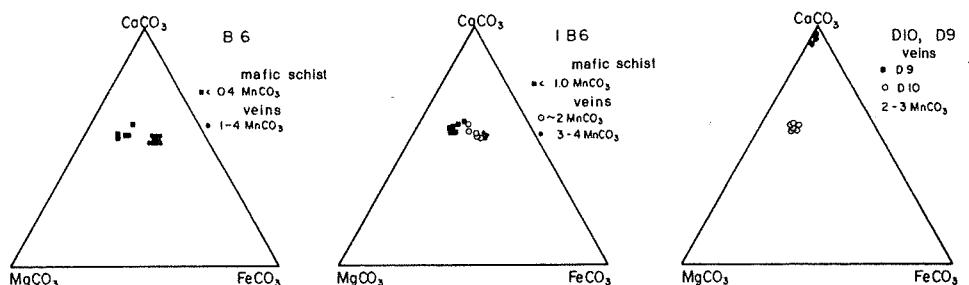
There is no direct evidence available on the ambient hydraulic pressure or crustal depth at which gold-bearing veins formed. However, based on the premise that fluids were derived by outgassing reactions under greenschist-facies conditions, or at the greenschist-amphibolite transition, and assuming a geothermal gradient of $50^\circ C km^{-1}$ with $P_{fluid} > P_{lithostatic}$, fluid pressures and depths of 8km, 2 kb (200 MPa) to 10 km, 2.5 kb (250 MPa) are indicated. These conditions are well above the vapour curve for H_2O-CO_2 solutions of 6 wt % equivalent NaCl (Takenouchi and Kennedy, 1965). Furthermore, inspection of primary fluid inclusions in vein quartz from the Con Mine, Yellowknife, has not revealed evidence for boiling of the hydrothermal fluids (Kerrich and Fyfe, 1981).

DOME MINE CARBONATE COMPOSITIONS

carbonates in sediments and veins



carbonates in mafic schists and veins



veins in Timiskaming

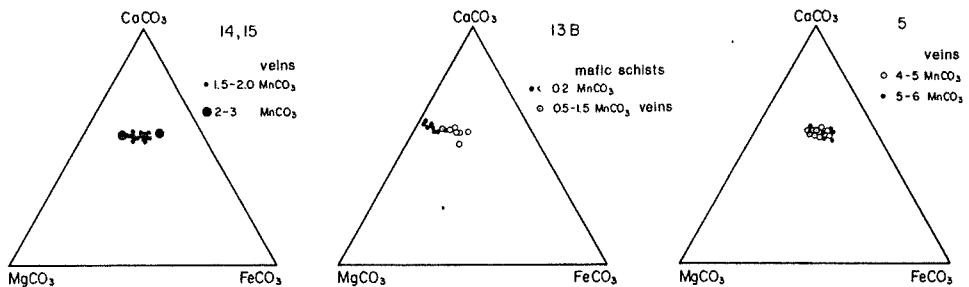
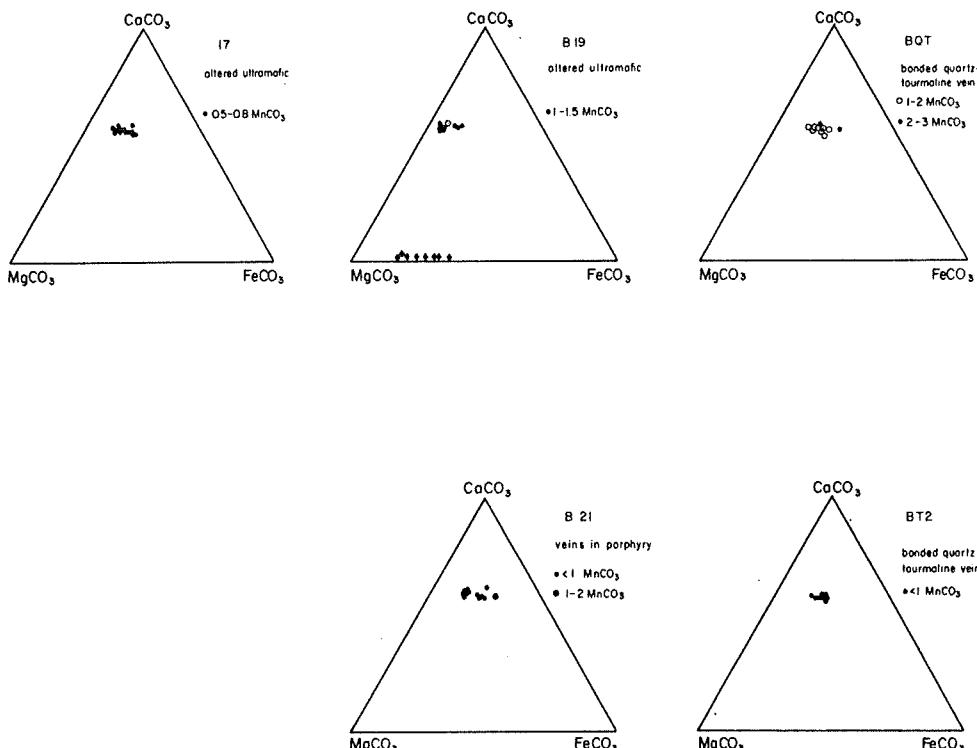


Figure 8A : Compositions of primary sediments and secondary vein carbonate minerals in various rock types at the Dome Mine, plotted in Ca, Mg, Fe-co-ordinate space. Manganese contents are indicated by different symbols. Note vein carbonates are systematically more Fe- and Mn-rich than neighbouring sedimentary counterparts.

DOME MINE CARBONATE COMPOSITIONS

veins hosted by altered ultramafic rocks



R. L. BARNETT, analyst

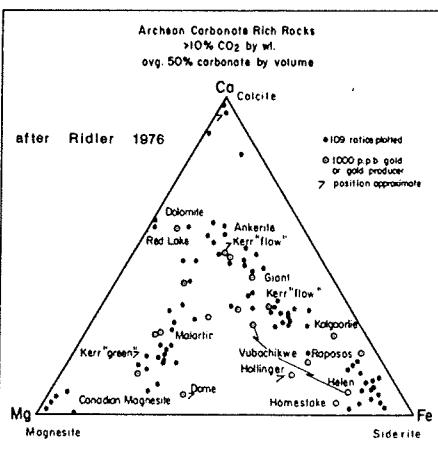


Figure 8B : (Upper): Compositions of vein carbonate minerals in ultramafic and quartz-feldspar porphyry host rocks, Dome Mine, plotted in Ca , Mg , Fe -co-ordinate space. Manganese contents are indicated by different symbols. (Lower): Summary diagram of carbonate mineral compositions in Archean carbonate-rich rocks (after Ridler, 1976).

Observations of Au-bearing vein deposits in Archaean greenstone belts, together with examination of the literature on this subject, reveal a broad empirical relationship between carbonate composition and the nature of the wall rock which the vein transects. In general, the predominant carbonate minerals are ankerite and ferroan dolomite in mafic rocks, magnesite and Fe-dolomite in ultramafic rocks, calcite in felsic rocks, and siderite and rhodochrosite in iron formation. Transitions in wall rock over the length of individual veins are reflected by corresponding changes in carbonate gangue composition (Bain, 1933). These observations support the deductions from chemical mass balance (see above) that the wall rock dominates the supply of the bivalent metal cations, and CO₂ is introduced by the hydrothermal fluids.

A summary of carbonate compositions for the Dome Mine is given in Fig. 8. Significantly, carbonates in gold-bearing veins are distinct in texture, chemistry and carbon isotope composition (see next section) from pre-existing carbonates associated with early silitization. Pearton (1982) has emphasized the fact that hydrothermal carbonates from the Sb-Au ores of Murchison possess different compositions from marine carbonates.

The disposition of carbonate-rich rocks and their spatial relationship to gold deposits of the Timmins District has been extensively discussed by Karvinen (1978, 1979, 1982), Fyon and Crocket (1979, 1980), Fyon *et al.* (1980, 1981, 1982), Roberts and Reading (1981), and Whitehead *et al.* (1980, 1981).

Karvinen considers that gold deposits in the Timmins camp are closely associated with two major 'carbonate-rich units' which can be traced along strike for a distance of 15 km, with most of the quartz-feldspar porphyry bodies of the area associated with the lower unit. Karvinen (1982) envisages a primary syngenetic enrichment of gold into the seafloor environment, during felsic volcanism, accompanied by 'exhalative' activity; followed by a second phase of Au enrichment into quartz-carbonate veins during regional metamorphism and deformation. However, Fyon and Crocket (1980) have clearly demonstrated that the 'carbonate units' are not continuous, but that a series of carbonate alteration zones exist with boundaries that trend from discordant to concordant with respect to the volcanic stratigraphy, with this stage of carbonate formation occurring shortly after extrusion. Similar relations of carbonate alteration to the volcanic stratigraphy and structures of the Virginiatown area have been described by Downes (1980). This alteration geometry may represent the passage of CO₂-rich fluids through permeable volcanic units acting as aquifers, with local breaching into successively higher permeable horizons and final debouching onto the seafloor, as suggested by Kerrich and Fryer (1979).

A. Carbon and Oxygen Isotope Geochemistry of Carbonate Minerals

The ubiquitous presence of hydrothermal carbonates, along with quartz, in mineralized fault zones, provides evidence that the fluid reservoir carried significant CO₂, raising the question as to the source of carbon. The oxygen and carbon isotope compositions of hydrothermal carbonates, principally from structurally hosted lode gold deposits, are summarized in Fig. 9. For the majority of data, carbonates occupy a restricted field, where δ¹⁸O spans 18 to 17, and δ¹³C = -8 to -0,5.

Fyon *et al.* (1980, 1981) report the carbon and oxygen isotope compositions of carbonate minerals from various settings in the Timmins District. Carbonates from a siliceous marble have δ¹³C (0,3 per mil PDB) and δ¹⁸O (+11,5 per mil SMOW), typical of other Archaean marbles (Schidlowski *et al.*, 1975). Dolomites from auriferous carbonate-rich chemical sediments are characterized by δ¹³C -3,2 per mil and δ¹⁸O +13 per mil. According to Fyon *et al.* (1980, 1981), the oxygen isotope compositions of the former may signify precipitation in a marine environment at temperatures of ~100°C, whereas the latter are higher temperature hydrothermal products associated with gold mineralization. Such high ocean-floor temperatures are in accord with high bottom-water temperatures independently deduced from the δ¹⁸O of cherts from various auriferous carbonate sediments by Kerrich and Fryer (1979).

Replacement dolomites in carbonatized volcanic flows at Timmins possess δ¹³C -1 to -3,5 per mil, and δ¹⁸O +10 to +16 per mil (Fyon *et al.*, 1980). The carbon isotope compositions are comparable to those of the two rock types described above, but according to Fyon *et al.* (1980, 1981) the lower δ¹⁸O signature of replacement carbonates probably signifies a higher formation temperature (and/or higher fluid δ¹⁸O), compatible with a slightly deeper level of precipitation compared to the marine floor sediments.

B. Source of CO₂

During normal prograde regional or burial metamorphism, fluid pressures are in the same order of magnitude as

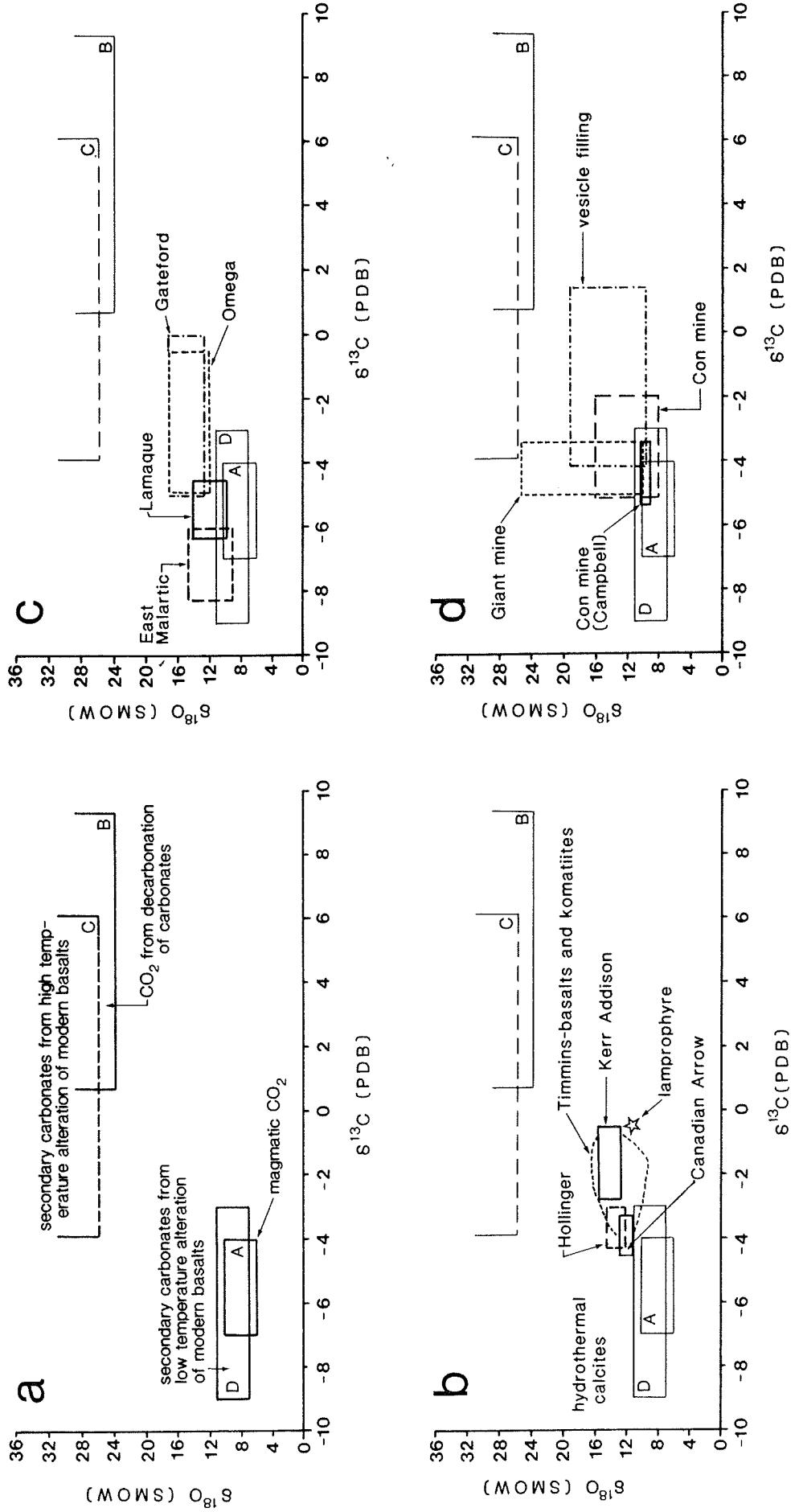


Figure 9(a) : Fields of different carbonates, and CO_2 derived from specified sources, plotted in $\delta^{13}\text{C}$ versus $\delta^{18}\text{O}$ co-ordinate space (modified from Fox, 1985; Fyon et al., 1983).

(b) : Isotopic composition of hydrothermal carbonates associated with gold deposits, from the Timmins area (after Fyon et al., 1983; Hollinger Mine, data this paper), Matachewan (Canadian Arrow) and Larder Lake (Kerr Addison) areas. Abitibi greenstone belt (see Fig. 9, Part I).

(c) : Isotopic composition of hydrothermal carbonates associated with gold deposits from the Larder Lake (Gateford and Omega properties, Foy, 1985), Malarctic, and Val d'Or areas (Lamaque Mine). Note the provinciality of $\delta^{13}\text{C}$, probably signifying variation in properties of various carbon donors in the source region.

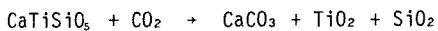
(d) : Isotopic composition of hydrothermal carbonates associated with gold mineralization from the Cona and Giant mines, Yellowknife. Also illustrated is the field for early vesicle filling carbonates, probably associated with spilitization of the volcanics.

lithostatic pressures (Fyfe *et al.*, 1978). Except in environments where partial melting occurs, total fluid pressure is close to the sum of partial pressures of H₂O and CO₂, the dominant volatiles. If the mixtures are considered ideal, the partial pressures are directly related to the mole fractions. As the average crustal content of carbon is around 200 ppm and the water content is close to 20,000 ppm, it is clear that water is the dominant volatile species in metamorphism (Fyfe *et al.*, 1978).

For given lithostatic pressures and temperatures, studies of the chemistry of fluids (Ellis and Mahon, 1977) and experimental studies of carbonate and hydrate equilibria (Winkler, 1976; Fyfe *et al.*, 1978) clearly show that for most temperatures up to the amphibolite facies P_{H₂O} is much greater than P_{CO₂}. Only under extreme conditions, as for the granulite facies (Janardhan *et al.*, 1979), does CO₂ become a major or even dominant volatile species. This observation is in accord with the great thermal stability of relatively pure carbonate rocks. Most work on metamorphic rocks shows that given the very low porosity and permeability of rocks at depth, massive degassing proceeds only where the geothermal gradient intersects the equilibrium vapour pressure curves when the latter approach the lithostatic pressure. Thus, for pure dolomite or siliceous marbles temperatures in excess of 800°C (migmatite temperatures) are required to achieve conditions of P_{CO₂} ~ P_{lithostatic}.

In the general case of regional metamorphism and dewatering of a pile of complex lithology including marine sediments and spilitic rocks, the primary carbonate content is likely to be significantly greater than 200 ppm. In Archaean sections where altered submarine volcanic rocks are dominant, again the CO₂ content is likely to be large via fixation of marine carbonate into cooling basalts. Jolly (1974) indicates that calcite is ubiquitous in the low-grade metavolcanics of the Archaean Abitibi belt in Ontario. Aumento *et al.* (1975) indicate that CO₂ may reach 3% in heavily altered sea-floor basalts, and Hallberg *et al.* (1976) report CO₂ contents of 0,1 - 0,3 wt % even for relatively unaltered Archaean basalts.

In such environments, given thermal gradients in the order of 30°C km⁻¹, P_{H₂O} will rise 280 bar km⁻¹. At 300°C (10 km), P_{fluid} will be in the order of 2800 bar and P_{CO₂} will be no more than 100 bar. At 500°C (17 km), P_{fluid} will approach 5000 bar with P_{CO₂} in the 1000-bar range. At Yellowknife, the presence of sphene in epidote-amphibolite-facies metabasalts, formed via the reaction:



sets the xCO₂ of metamorphic fluids to ≤0,5 at 500°C and 2 kb (Hunt and Kerrick, 1977).

It is clear that if this metamorphic fluid moves upward and remains in equilibrium with Ca-Fe-Mg-silicates, it will precipitate carbonates by wall-rock alteration until virtually all CO₂ is removed. Inspection of carbonate equilibria (Winkler, 1976; Fyfe *et al.*, 1978; Turner, 1981) indicates that maximum carbonate formation will occur in the greenschist-facies region if the source-rocks have moderate CO₂ contents. In summary, Au and quartz precipitation reflects diminished solubility related to temperature, and carbonate precipitation is related to the above-equilibrium CO₂ pressures relative to wall-rock alteration equilibria (a similar situation to reduction reactions related to excess H₂ pressures discussed above under redox reactions).

Additional reactions producing CO₂, together with reduced gas species such as CO and CH₄, in metamorphic fluids may include the oxidation and hydrolysis of reduced carbon or hydrocarbon compounds present in many interflow sediments. For instance, the reduction of hematite or magnetite linked to the oxidation of carbon yields CO₂; and the hydrolysis of carbon may produce CO₂ + CH₄.

In general, the δ¹³C and δ¹⁸O of carbonates form a data array between the marine and igneous carbonate reservoirs. On this basis, Fyon *et al.* (1980) interpret the results as mixing of juvenile and seawater carbonate reservoirs, or juvenile and lower crustal CO₂ derived from granulitization (Fyon *et al.*, 1983, 1984). However, the δ¹³C data also lie between the marine carbonate reservoir and either the δ¹³C range of carbon in a CO₂-CH₄-bearing fluid derived from decarbonation reactions of marine carbonates or the δ¹³C of organic hydrocarbons. Here, the data of Fyon *et al.* (1980) are interpreted in terms of mixing between carbon derived from the last three named reservoirs.

A distinct provinciality exists in the carbon isotope composition of hydrothermal carbonates. For instance, δ¹³C is uniform at -4 to -9 per mil for the East Malartic and Lamaque mines, whereas at Timmins carbonates occupy a restricted range of 0 to -4 per mil. These differences are interpreted to reflect areal variations in the relative proportions of carbon-bearing materials in the crust donating CO₂ to the hydrothermal fluids.

C. Source of Sulphur

Hydrogen sulphide is a minor volatile component of hydrothermal fluids involved in Archaean gold deposits, reflected in the presence of sporadic concentrations of pyrite + arsenopyrite. The sulphur isotope composition of pyrite from gold deposits of the Timmins camp occupies a restricted field given by $\delta^{34}\text{S} = 2,7 \pm 1,9$; and for Archaean lode gold deposits of Australia, Lambert *et al.* (1983) report a range in $\delta^{34}\text{S}$ pyrite of +1 to +4. These results signify a well-mixed reservoir and uniform redox conditions at precipitation. Such values may reflect sulphur derived from a magmatic sulphur reservoir, and/or leaching of sulphides generated by the redistribution of magmatic sulphur accompanying early marine water alteration of igneous rocks. A redox-dependent fractionation exists for the isotopic composition of hydrothermal pyrites (Ohmoto and Rye, 1979). At the Canadian Arrow deposit, located on a splay of the Porcupine Destor fault, pyrite is depleted in $\delta^{34}\text{S}$ ($\bar{x} = -11,2 \pm 1,0 \text{ ‰}$, $n = 18$) relative to examples at Timmins. Schwarcz and Rees (1984) suggest a well-mixed sulphur reservoir to account for the uniformity of sulphur isotope compositions, coupled with relatively oxidizing conditions to yield the isotopically light values.

VIII. THE SOURCE ROCK PROBLEM

It is interesting, in the context of the foregoing discussion, to consider the question of gold deposits located in rocks of ultramafic composition. Pyke (1975, 1976) has noted a relationship of Au mineralization and ultramafic volcanic rocks in the Porcupine camp, Ontario, claiming that ultramafic rocks are favourable "source beds" because of their intrinsically elevated Au abundance. Analysis of Au by sensitive modern techniques reveals that unaltered ultramafic rocks have abundances of 0.5 - 2 ppb Au, in the same range as other primary igneous rocks (Tilling *et al.*, 1973; Anhaeusser *et al.*, 1975; Kwong and Crocket, 1978); these data contraindicate the "source-bed" concept.

There are two further aspects to the source-bed problem; namely the regional geological association and local causal factors. Although prominent gold deposits are physically related to ultramafic rocks, a case can be made for gold associations with a multitude of "special" rock types. In the Porcupine and Malartic camps, at Mine Bousquet and Mine Doyon - Cadillac District, and at Agnico Eagle, Quebec, gold deposits are related to Na-enriched felsic intrusive-extrusive igneous rocks. At the Kirkland Lake, Larder Lake, and Matachewan camps, Ontario, gold is locally hosted by 'syenites'. Similarly, the presence of Au in granodiorites is notable at Val d'Or, Quebec, and Dryden, Ontario. However, as Hutchinson (1976) and Hodgson (1983) have pointed out, the overwhelming preponderance of Au produced from Archaean lode deposits is from mines located within mafic igneous rocks. There is thus no geological basis for claiming a special relationship of Au with ultramafic rocks.

At the local scale, ultramafic rocks may provide favourable chemical environments for gold precipitation. In a typical granite, basalt, and ultramafic rock, the sum of cations (Fe, Mn, Mg, and Ca) which may combine with CO_2 to form carbonates is 4, 21, and 32 at. %, respectively. From this it is clear that ultramafic rocks have the greatest capacity to undergo intense carbonate-replacement alteration, and thus to exert the greatest integrated chemical effect on fluids exchanging with the rocks via scavenging of CO_2 . If wall-rock reactions may locally mediate precipitation of Au [for which there is abundant empirical evidence, cf. Bain (1933)], in addition to temperature (Fyfe and Henley, 1973), then it is possible that some of the Au may be transported as a carbonyl (CO), COS or related complex, with carbonate-forming reactions mediating Au precipitation by reduction of P_{CO_2} and (or) decrease of acidity. It is well known (Ephraim, 1948, pp. 776-810) that many metals capable of forming carbonyl and carbonyl-chloride complexes are also enriched in lode Au deposits. The presence of "metamorphic" Au-bearing quartz veins in greenstone belts may also partly reflect the high capacity of mafic and ultramafic rocks to store volatiles during early sea-floor spilitization reactions, the volatiles being released subsequently during crustal heating, as discussed above.

A brief summary of the various rock types which have been proposed as 'source' for the chemical components concentrated in gold deposits is given in Table II.

IX. AGE RELATIONSHIPS

A tightly constrained chronological framework exists for volcanic-plutonic activity in the Timmins and Kirkland Lake areas of the Abitibi greenstone belt, based on the classic U-Pb isotopic model age determinations conducted by Krogh, Nunes, and coworkers. However, the timing of fluid transport through the major structures and the source of solute components have remained a controversial issue. Attempts have been made to place limits on

TABLE II
Potential Gold Source Rocks

1. Quartz - feldspar porphyries (Trondhjemitic)	Carter, 1948 Ferguson <i>et al.</i> , 1968 Griffiths, 1968 Boyle, 1976
2. Ultramafic rocks	Pyke, 1976 Keays, 1981
3. Kenoran granodiorite batholiths	George, 1967 Jones, 1948
4. Metasedimentary rocks	Hodgson and MacGeehan, 1982
5. Any host lithology LATERAL SECRETION	Boyle, 1961, 1978
6. Chemical sediments	Fripp, 1976 Hutchinson, 1976 Ridler, 1976 Karvinen, 1978, 1980

TABLE III

Age - Constraints, Abitibi Belt

TIMMINS		(Nunes and Pyke, 1981)
Lower Supergroup (top)	2725 ± 2	
Upper Supergroup (top)	2703 ± 2	<u>GOLD</u>
Kidd Creek felsic host	2717 ± 4	
KIRKLAND LAKE		(Nunes and Jensen, 1980)
Hunters Mine Group	2710 ± 2	
Blake River Group	2703 ± 2	
Timiskaming		<u>GOLD</u>
NORANDA		
Lake Dufault granodiorite	2,701	
ABITIBI BATHOLITHS	2,675 - 2,685	(Krogh <i>et al.</i> , 1982)
MATACHEWAN DYKES	$2,690 \pm 93$	0.700 ± 0.001 (Gates and Hurley)

Ore Deposits - Closure Ages

 $^{87}\text{Sr}/^{86}\text{Sr}$ initial

Hollinger	2.440 ± 26	$^{40}\text{Ar}/^{39}\text{Ar}$	(muscovite)	0.70262 ± 0.0014
Macassa	2.575 ± 3	$^{40}\text{Ar}/^{39}\text{Ar}$	(actinolite)	
Kerr Addison	2.510 ± 30	$^{40}\text{Ar}/^{39}\text{Ar}$	(muscovite)	0.70249 ± 0.0009
Dumagami	2.420 ± 19	$^{40}\text{Ar}/^{39}\text{Ar}$	(muscovite)	0.70406 ± 0.0001
E. Malartic	2.504 ± 16	$^{40}\text{Ar}/^{39}\text{Ar}$	(biotite)	

the age of fluid discharge, using incremental $^{40}\text{Ar}/^{39}\text{Ar}$ and Rb-Sr isochrons on some of the prominent mineralized fault domains.

In the Kirkland Lake area, the Hunter Mine Group, located near the top of the Lower Supergroup, or first volcanic cycle, has been dated at $2,710 \pm 2$ Ma. The Blake River Group, near the top of the Upper Supergroup, or second volcanic cycle, has an age of $2,703 \pm 2$ Ma (Nunes and Jensen, 1980). These are upper concordia intercept model ages on zircons, which probably represent minima given the absence of xenocrystic zircons. Collectively, these data, which span the cumulative stratigraphy of volcanic cycles with an apparent thickness of ~ 35 km, signify a time interval of 7 ± 3 Ma for accumulation of much of the Kirkland Lake section of the Abitibi greenstone belt (Table III). In the Timmins District, volcanic activity spanned ~ 25 Ma from 2725 to 2703 Ma (Table III).

Major granitic batholiths are thought to be broadly contemporaneous with volcanic activity. Krogh *et al.*, (1982) report that the majority of batholiths have ages in the narrow interval 2,675 to 2,685 Ma (U-Pb isotopic ages on zircons). The Round Lake batholith has a Rb-Sr isochron age of 2,730 Ma (Ontario Geological Survey, unpublished data). A lower age constraint on Abitibi belt igneous activity is provided by a Rb-Sr whole rock isochron age of $2,690 \pm 93$ Ma on the Matachewan diabase dyke swarm (Gates and Hurley, 1973), which transects all rock types, batholiths, and some of the structurally hosted mineralization in the Timmins area.

Existing age determinations on gold mineralization in structural lineaments such as the Kirkland Lake - Larder Lake fault zone are principally Stacey-Kramers model Pb isotopic ages on galenas and leached coexisting pyrite, which define ages of 2,656 to 2,700 Ma (Franklin *et al.*, 1983). Rb-Sr isochron and incremental $^{40}\text{Ar}/^{39}\text{Ar}$ spectra yield closure times of 2,550 to 2,700 Ma (Table III, Fig. 12; Kerrich *et al.*, 1984). The least distribution isotopic systems are in good agreement with model Pb ages, whereas the younger ones reflect resetting during continued displacement along the major structures into the Proterozoic. For example, the Kirkland Lake fault zone locally displaced Proterozoic Cobalt Group sediments, which are in the order of 2,250 Ma.

The hydrothermal muscovite from Kerr Addison, on the Kirkland Lake fault zone, is characterized by a clear diffusion loss configuration, which can be modelled by assuming that there have been two separate events affecting the structure since formation of the muscovite (Fig. 12a). The best fit age of growth is $2,750 \pm 50$ Ma: in this model 57 per cent of the ^{40}Ar was lost at 2,030 Ma, with a further 3 per cent loss at $1,280 \pm 240$ Ma. The former event may reflect the last significant displacement increment on the fault zone, whereas the 1,280 Ma plateau may correlate with Grenvillian activity.

Based on the above results in conjunction with field relations, dominant hydrothermal activity in Abitibi belt gold deposits is interpreted to have occurred during an episode of major fracturing and ductile deformation that accompanied emplacement of granitic batholiths into the supracrustal sequences: hydrothermal mineralization was thus an integral part of the thermal and mechanical energy of the greenstone belt (see also Colvine *et al.*, 1984) for a summary of age relationships).

X. SOURCE OF THE HYDROTHERMAL SOLUTES

The origin of fluid reservoirs which were involved in Archaean gold systems has been considered above on the basis of light stable isotope evidence. However, the source rock reservoir for hydrothermally transported solutes, such as Si, CO_2 , K, Rb, Li, Cs, Sr, and rare metals including Au, Ag, As, Sb, B, Se, Te, Pb, Bi, W (etc.) is a more subtle problem. Here, the question is explored by means of radioisotope tracers, namely $^{87}\text{Sr}/^{86}\text{Sr}$ and $^{207}\text{Pb}/^{204}\text{Pb}$, $^{206}\text{Pb}/^{204}\text{Pb}$, which may be diagnostic of the specific rock reservoir from which the solutes were derived.

Initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratios of mineralized domains in major structures may be estimated using Rb-Sr isochrons, and from analyses of Sr-rich, Rb-poor minerals which have plausibly acted as closed isotopic systems.

Whole rock Rb-Sr isotopic systems have been disturbed subsequent to the time of mineralization, probably due to continued displacement on major structures into the Proterozoic. However, the Rb-Sr isotopic data may be extrapolated backwards to estimate the $^{87}\text{Sr}/^{86}\text{Sr}$ ratio at the time of fluid transport, by rotating isochrons about the mean $^{87}\text{Sr}/^{86}\text{Sr}$ of mineralized rocks, to the correct age as independently determined by model Pb ages, and lower age limits imposed by U-Pb dates on zircons (see above). For instance, using a value of $0,70249 \pm 0,0009$ for $^{87}\text{Sr}/^{86}\text{Sr}$ at the time of closure, namely 2,510 Ma, and an estimated age of 2,690 Ma for mineralization, the $^{87}\text{Sr}/^{86}\text{Sr}$ of fluids would have been 0,7014 ($^{87}\text{Rb}/^{86}\text{Sr} = 0,0367$) to 0,7023 ($^{87}\text{Rb}/^{86}\text{Sr} = 0,057$) (Table III, Fig. 10a).

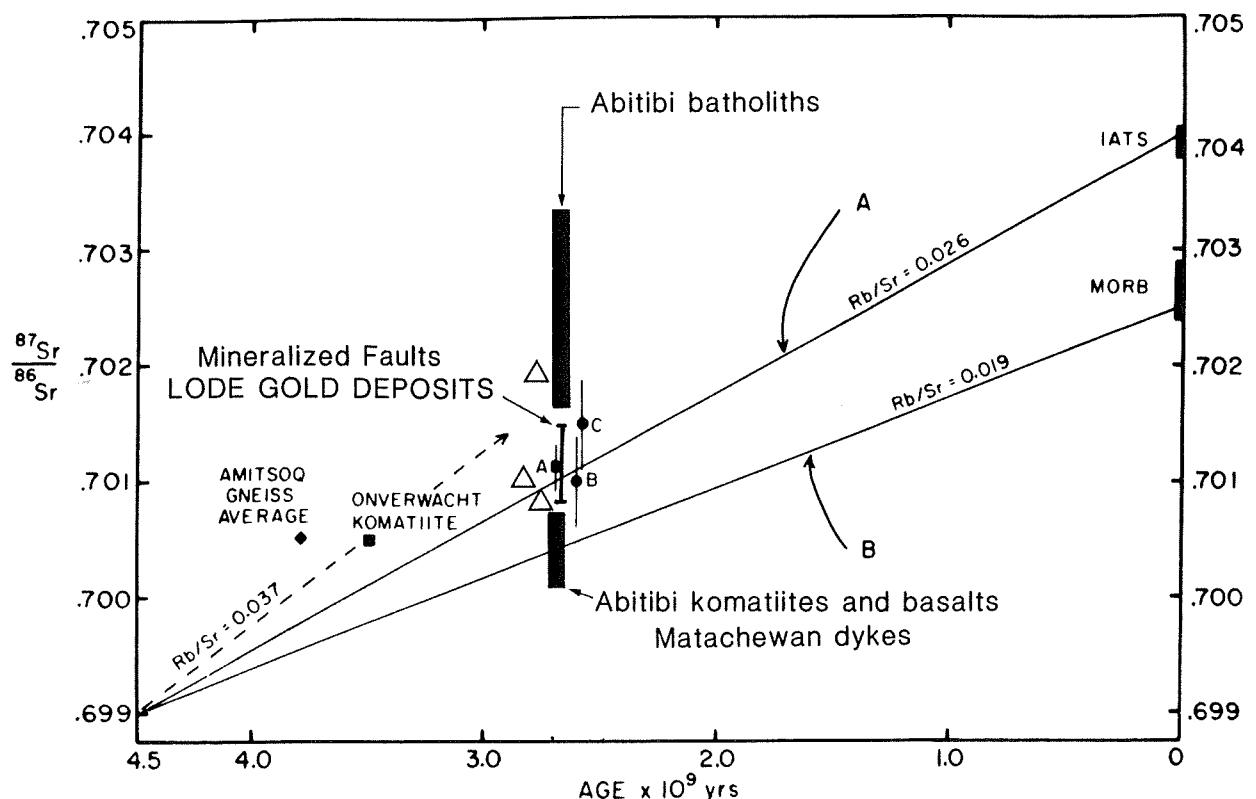


Figure 10A : Contemporaneous $^{87}\text{Sr}/^{86}\text{Sr}$ of mineralized linear structures, compared to mantle, komatiites, and granodiorite batholiths, for the Abitibi greenstone belt at 2.7 Ga.

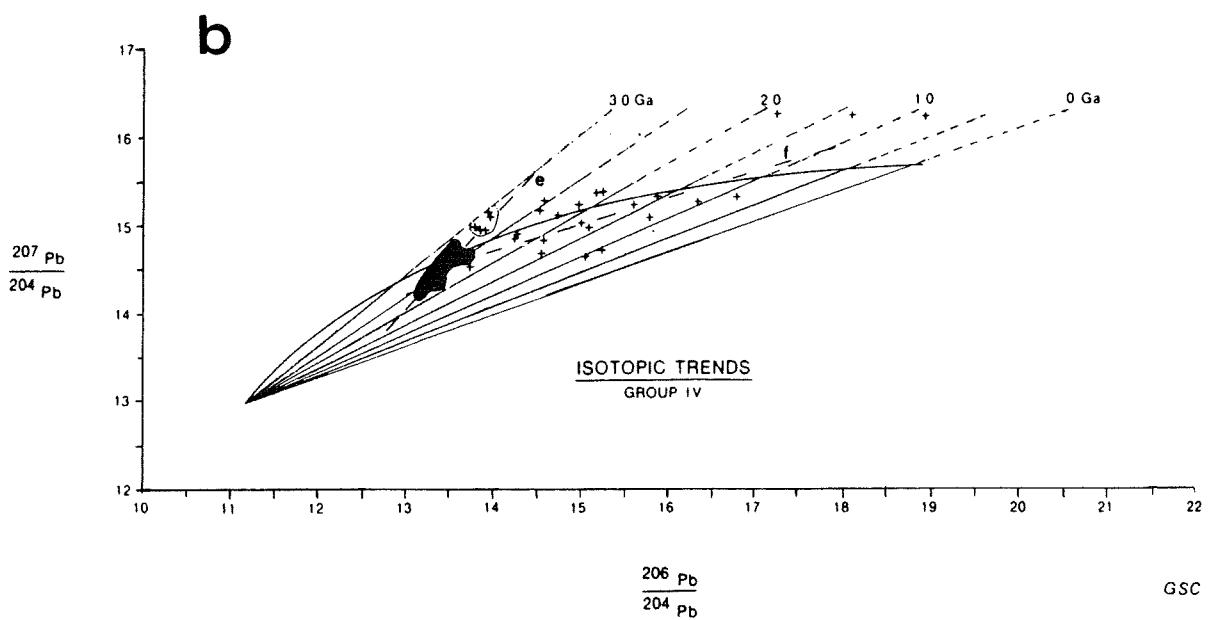


Figure 10B : Lead isotopic composition of massive sulphides (solid) and mineralized structures (crosses), after Franklin et al., (1981)

Contemporaneous $^{87}\text{Sr}/^{86}\text{Sr}$ ratios of the hydrothermal mineralization have been independently estimated from analyses of tourmaline, scheelite, and piedmontite, all of which possess $^{87}\text{Rb}/^{86}\text{Sr} \leq 0,004$, and therefore have negligibly small radiogenic strontium evolution over the elapsed time since formation. Moreover, these minerals, unlike ferrodolomites present in the fault zones, appear to have behaved as isotopically closed systems. Strontium isotope data for selected deposits are compiled in Table III. Initial strontium isotope ratios for the fault mineralization are likely those of the hydrothermal fluids which donated Sr, and this may also reflect the $^{87}\text{Sr}/^{86}\text{Sr}$ of the rock reservoir with which the fluids equilibrated.

If the estimated range of 0,7010 to 0,7023 for the source rock reservoir of hydrothermal solutes has validity, then this reservoir was more radiogenic than the upper mantle at 2,690 Ma ($0,700 \pm 0,001$), the Matachewan diabase dyke swarm ($0,701 \pm 0,001$), or contemporaneous Abitibi belt volcanics of mafic to ultramafic composition (0,7000 to 0,7012) (Table III; Faure, 1977; Gates and Hurley, 1973; O'Nions and Pankhurst, 1978; Fig. 4; Hart and Brooks, 1974). An inferred radiogenic character for the source may implicate contributions from sialic basement to the greenstone belt, and/or felsic volcanic rocks of the Lower Supergroup, in addition to volatiles released from the mafic to ultramafic volcanic sequence (Fig. 10a).

Franklin *et al.* (1983) have shown, on the basis of lead isotopic work, that, in general, fault-hosted gold mineralization in the Abitibi greenstone belt possesses a characteristically more radiogenic signature and higher μ than contemporaneous mafic volcanic rocks (Fig. 10b).

Thus the deduced Sr and measured Pb-isotopic character of the fault mineralization are more evolved than their contiguous mafic volcanic bounding media, collectively implicating a deeper level, more radiogenic source for these components. This source is likely the same one as that donating Si, CO₂, K, Rb, Li, Cs, Au, Ag, As, Sb, B (etc.). Long range communication to a deep crustal source is corroborated by the presence of the major structures themselves, as well as independent estimates of transport distances based on enrichments of rare elements in the structures (see above; Kerrich and Fryer, 1979; Kerrich, 1983).

XI. ALUMINOUS ROCKS ASSOCIATED WITH GOLD DEPOSITS

Unusually aluminum-rich volcanic rocks containing exotic aluminum-oxide, -hydroxide and -silicate minerals have been identified in proximity to several gold-bearing vein systems. Altered pillow basalts enveloping auriferous laminated ferroan dolomite chert units at the Dome Mine contain up to 27 wt % Al₂O₃, compared to initial abundances of ~12 wt % in the unaltered precursors (Fryer *et al.*, 1979). TiO₂ and Zr undergo an increase in aluminous rocks over primary parental abundances in approximately the same proportion as Al₂O₃. The high-alumina rocks are characterized by low SiO₂, MgO, CaO, and Na₂O: they contain abundant chloritoid, stilpnomelane and muscovite.

Overall compliance of the inter-element ratios of TiO₂, Al₂O₃ and Zr between the least altered pillow basalt precursors and highly aluminous products (see Kerrich, 1983) signifies that the suite of least to most altered rocks are from an initially chemically uniform population of basalts. Furthermore, these consistent inter-element ratios indicate that the observed proportional increases of TiO₂, Al₂O₃ and Zr in aluminous rocks are due to their concentration via isochemical behaviour during leaching of soluble components, rather than to absolute introduction of the elements.

As described in the preceding sections, relatively aluminous rocks of mafic and ultramafic parentage are present at the Dome, Hollinger and Kerr Addison mines. In all cases, initial alteration of the volcanic flows appears to have proceeded at low temperatures in the presence of low-salinity, CO₂-rich, high- ^{18}O fluids, at or near the seafloor, resulting in massive loss of SiO₂, Fe₂O₃, MgO, CaO, and Na₂O with concomitant proportional concentration of TiO₂, Al₂O₃, Sc, Zr, and Hf (see Part I). Reactions between CO₂-rich aqueous fluids and the volcanic flows may have involved hydrolysis of igneous Fe, Mg, Ca, Na-silicates, with removal of these elements in solution, leaving a low- to intermediate-temperature residual hydrothermal assemblage, or its metamorphosed product, such as clay minerals, corundum (Al₂O₃), diasporite (Al₂O(OH)₃), gibbsite (Al(OH)₃), mullite (3Al₂O₃.2SiO₂), pyrophyllite (Al₂[Si₆O₁₀](OH)₄), chloritoid, muscovite and one or more of the three aluminosilicate polymorphs (Al₂SiO₅).

It is interesting to note that in most of the settings where aluminous rocks and/or exotic aluminosilicates have been reported in association with ore deposits, notably at Mattabi (Franklin *et al.*, 1975) and Michipicoten (La Tour *et al.*, 1980), abundant carbonate minerals are also present, or there is evidence for CO₂-rich fluids. If

this association is valid, then participation of bicarbonate in hydrolysis reactions may catalyze production of an aluminous residue (cf. Krauskopf, 1967, Chapter 4). Further, if boiling occurs in a hydrothermal discharge plume, evolution of an acid vapour phase may act to promote the production of alumino-silicate minerals such as pyrophyllite, corundum and paragonite (cf. Burt and Rose, 1979). Boiling fluids would signal relatively shallow water conditions. Because CO₂ dissolved in water acts to suppress the solubility of quartz, immiscible separation of CO₂ from a hydrothermal fluid may leave the residual fluid undersaturated with respect to quartz, and thus promote its solution. As immiscible CO₂ rises toward the surface, its condensation in cool shallow aquifers would generate acid solutions, that could enhance leaching of the more soluble silicate minerals.

Valliant *et al.* (1982) describe alumina-rich rocks in three distinct settings at the Thompson Bousquet Mine, Quebec. These are: (i) a hangingwall tuffaceous unit, composed of andalusite, corundum, paragonite, kaolinite, muscovite, quartz, and kyanite; (2) a unit footwall to the stratabound pyritic ore; and (3) a second unit in the footwall, composed of chloritoid, quartz, muscovite, carbonate, magnetite, and sulphides. For the first two, production of aluminous rocks is attributed to alteration of tuffs by a ponded hot sea floor brine and ascending hydrothermal brine, respectively. The third setting is interpreted as a chemically precipitated aluminous subfacies of iron formation. However, these rocks have similar Al₂O₃/TiO₂ ratios as the contiguous mafic ash tuff located to the west, and an origin in terms of extensive leaching of tuffs is more plausible than coprecipitation of aluminum and titanium in a chemical sediment fortuitously in the same proportion as in the contiguous volcanics.

XII. SODIC AND ALKALINE IGNEOUS ROCKS

A characteristic feature of the major fault zone hosted gold deposits are sporadically distributed colinear arrays of sub-volcanic stocks along with their extrusive equivalents, of highly sodic or alkaline composition. Prominent examples include trondhjemites of the Malartic, Matachewan and Timmins areas, and alkaline rocks at Kirkland Lake.

The sodic rocks conform in most respects to the definition of a trondhjemite based on major element oxides, according to Barker (1979); namely SiO₂ > 68 % usually < 75 %, Al₂O₃ typically > 15 % at 70% SiO₂ and < 14 % at 75% SiO₂, and Na₂O 4,0 - 5,5 % (etc.). For instance, a population of six trondhjemites at Canadian Arrow possess SiO₂ = 65,9 ± 1,8 1σ, Al₂O₃ = 16,7 ± 0,22, and Na₂O = 8,32 ± 1,16 (McNeil and Kerrich, 1985). Most such trondhjemites have elevated primary Na₂O contents, but this is incremented during pervasive spilitic alteration in the presence of marine water, under conditions of low temperature, as evidenced by reversed quartz-feldspar fractionations ($\delta^{18}\text{O}$ quartz = 9 to 11, $\delta^{18}\text{O}$ albite = 10 to 16, Fryer *et al.*, 1979; Kerrich and Fryer, 1979; Kerrich, 1983). A general concensus exists that such magmas form by partial melting of mafic granulites at the base of the greenstone belt sequence (Collerson and Fryer, 1978; Barker *et al.*, 1979).

Groups of alkaline igneous intrusions and tuffs at Kirkland Lake are characterized by SiO₂ in the range 48 - 56 %, and K₂O 5,5 to 9,8 %. Relatively abundant ilmenite, hornblende, and/or augite is reflected in elevated TiO₂, Fe₂O₃, MnO, CaO, and MgO contents, along with high concentrations of mafic-affiliated elements such as V, Cr, Co, Ni, and Sc.

Formerly termed syenites these rocks are, in fact, potassic basalts and intrusive equivalents. Specifically the low absolute Zr and Nb abundances result in high Ti/Ar (13-35) for syenites, where Ti/Zr is typically about 7; and Nb/Y ratios of 0,2 to 0,5 which are smaller by a factor of 5 to 10 than is typical of syenitic rocks (Kerrich and Watson, 1984). Moreover, the Ti/V, Zr/Y, K/Rb, and Rb/Sr ratios are lower, and the Cr/Ni ratios higher, than is the case for most rocks of syenitic composition. Rare earth element distributions for unaltered samples of the alkaline rocks are remarkably uniform (Fig. 11), defining a coherent array with La 100 to 200, and the REE 10 times chondritic abundances.

A general concensus exists that alkalic mafic rocks are the products of low degrees of partial melting of asthenosphere, or 'enriched' asthenosphere, and such fusion products have a preferred spatial association with linear fracture zones, such as rifts. Along with the alkaline magmas, are lamprophyric dykes and domains of fenitization, highly enriched in large ion lithophile elements. Representative chondrite-normalized diagrams are depicted in Fig. 11a-d.

The spatial association of trondhjemites and alkaline magmas with major linear structures is interpreted in terms of trans-lithospheric fractures, which tap magmas of sodic composition generated at the base of the crust, and deep asthenospheric partial melts of alkaline composition. Such magmatic activity is broadly coeval with

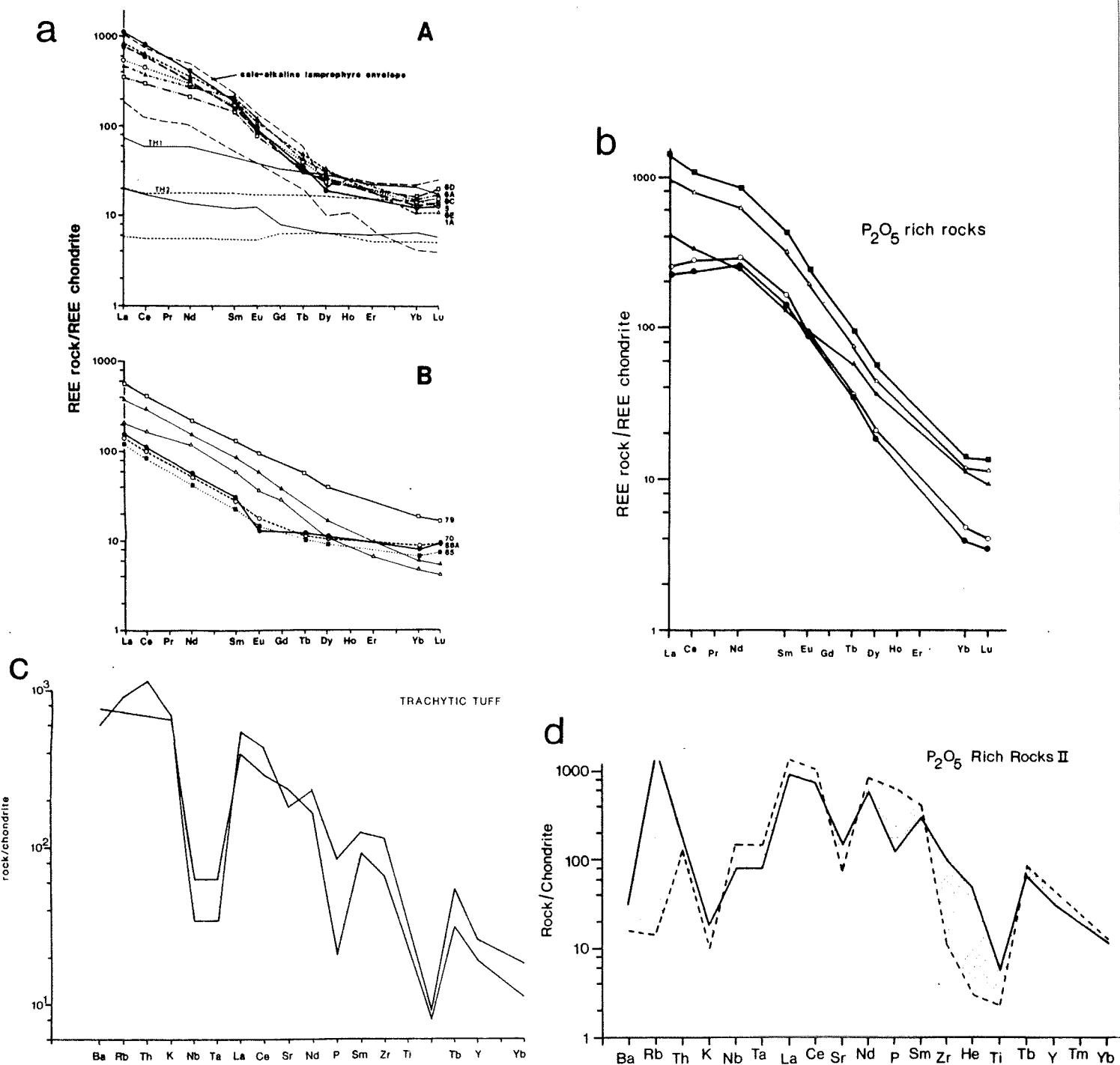


Figure 11 : Chondrite normalized element distributions for specified suites of alkaline magmas. (a) lamprophyres (after McNeil and Kerrich, 1985), (b) fluorapatite-bearing fenitized trondhjemites (King, 1983), (c) trachytes from Kirkland Lake, (d) fluorapatite-rich fenites (King, 1983).

venting of hydrothermal fluids, probably from a mid-crustal reservoir. Thus the "exotic" magmas, and gold deposits, collectively reflect the tectonic environment, and any genetic link between deposits and magmas remains controversial.

XIII. MODELS FOR GENERATION OF LODE GOLD DEPOSITS

A. Boundary Conditions

Any hypothesis which encompasses geological and chemical characteristics of lode gold deposits in greenstone belts must satisfy the following constraints:

1. concentration of gold in major transcrustal fracture systems, independent of lithology or stratigraphic position;
2. precipitation of gold over a significant range of ambient redox conditions, from sulphate-pyrite to pyrrhotite-graphite-methane;
3. separation of base metals from rare elements at all stages of gold enrichment;
4. Au concentration factors in the order of 10,000 and therefore source rock volumes of hundreds of km³;
5. source of abundant carbon dioxide;
6. relatively high, uniform fluid temperatures of about 360° to 450°C in veins;
7. hydrothermal solutions with δ¹⁸O 6 to 8 per mil, δD -50 to +5 per mil, high K/Na, high CO₂, low salinity, and the presence of reduced gas species such as H₂, CO and CH₄;
8. hydraulic pressure ≥ lithostatic pressure in veins;
9. water/rock ratios sufficient to cause complete oxygen isotope exchange of host rocks with the aqueous reservoir along flow conduits, and to extensively modify primary host-rock chemistry;
10. mineralizing hydrothermal solutions originating under highly reduced conditions, leading to extensive reduction of wall rocks along hydrothermal conduits; and
11. a mechanism which is intrinsically efficient under Archaean crustal conditions, specifically 2.7 Ga ± 1 Ga greenstone belts.

B. Models for Lode Gold Deposits

A number of hypotheses have been advanced to account for the source of Au and generation of lode deposits in greenstone belts. In summary, these are:

1. derivation of Au from stocks of quartz-feldspar porphyry considered to be of subvolcanic origin (Carter, 1948; Ferguson *et al.*, 1968; Griffis, 1968; Boyle, 1976);
2. lateral secretion of vein material from the enclosing rocks (Hurst, 1935; Boyle, 1961, 1976, 1976);
3. derivation of Au from granitic batholiths of Kenoran age (George, 1967; Jones, 1948);
4. mobilization of Au from ultramafic rocks (Pyke, 1975, 1976) or sediments (Hodgson and MacGeehan, 1982) postulated to have a high primary abundance of gold;
5. multistage origin (a): primary syngenetic enrichment of Au into gold-bearing chemical sediments generated from circulation of marine water, or from 'volcanic exhalations', during periods of volcanic quiescence. Precious metals and gangue constituents are considered to be remobilized into fractures from the primary chemical sediments during subsequent deformation and metamorphism (Hutchinson, 1976; Fripp, 1976; Ridler, 1976; Karvinen, 1978, 1980; and Fig. 25). Alternatively, gold-bearing veins are presumed to be discharge sites of a marine-water geothermal system (MacGeehan and Hodgson, 1982; Hodgson and MacGeehan, 1982; MacGeehan *et al.*, 1982; Fyon and Crocket, 1981, 1982).
6. multistage origin (b): solution and transport of gold, by fluids generated during crustal dehydration, with a possible initial concentration in the near seafloor environment, and main phase into hydrofractures; and
7. deposition of gold in veins from fluids containing a significant component of low-Cl, low-δ¹⁸O meteoric water in some felsic batholiths that emerged above sea level (Kennedy and Kerrich, 1982). Such deposits are neither abundant nor large producers, and are not commented upon further here.

These models and their variants are critically discussed below.

1. *Derivation of Gold from Quartz-Feldspar Porphyries*

A number of earlier workers referenced above, and Hodgson (1983), have pointed out the apparently anomalous abundance of felsic intrusive rocks with lode gold deposits of the Superior Province. Various possible explanations that have been proposed for this spatial association are:

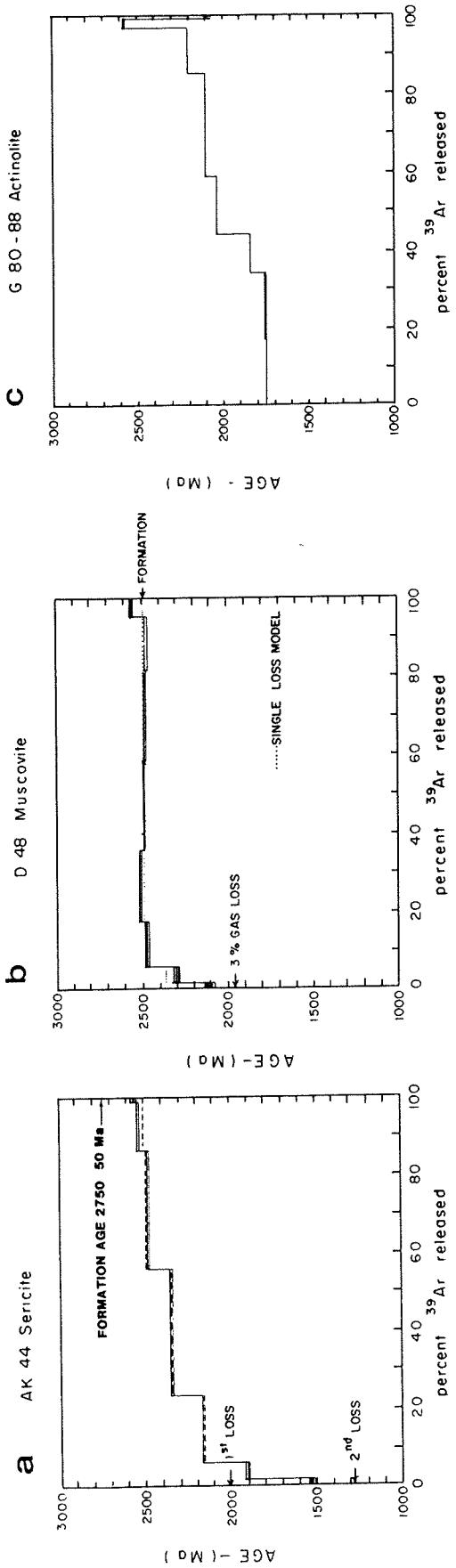


Figure 12 : Incremental $^{40}\text{Ar}/^{39}\text{Ar}$ spectra from fault-related hydrothermal minerals.
 (a) Kerr Addison muscovite, (b) Dumagami muscovite, (c) Macassa actinolite
 (after Kerrich, 1985).

1. magmas were a source for metalliferous hydrothermal solutions;
2. the intrusions provided heat; or
3. acted as conduits for the mineralization process.

Fryer *et al.* (1979) and Kerrich and Fryer (1979) suggested that the preferred siting of veins at margins of felsic intrusions at the Dome Mine, is a result of a structurally disharmonic style generated by folding of a volcanic sequence containing an irregular intrusive body. The various possible genetic factors have been re-evaluated by Marmot (1983).

Calculations of mass balance based on the volumes and measured precious metal abundances of relatively unaltered porphyry (Table I, Fryer *et al.*, 1979) do not support the hypothesis that these igneous bodies contributed a significant fraction of gold to the ore bodies. In addition, Davies (1977), Roberts (1980) and Gorman *et al.* (1981) point out that many of the major auriferous quartz veins post-date emplacement and spilitization of the porphyries.

Karvinen (1978, 1980) has convincingly demonstrated that many of the porphyries in the Porcupine Camp were extruded at a horizon contiguous with some of the auriferous sediments, and coevally with their precipitation, although the timing of gold introduction to the sediments is equivocal. Given their 'spilitic' character, the quartz-feldspar-porphyries almost certainly provided energy to drive seawater convection, as discussed for East Malartic (Kerrich *et al.*, 1980; Munha *et al.*, 1980; Kerrich and Hodder, 1982)); but this geothermal regime was probably not the one responsible for gold enrichment, based on the reducing character, low salinity, high $\delta^{18}\text{O}$, high CO₂, and low Na/K of fluids in the auriferous system.

Hodgson and MacGeehan (1981) have conjectured that 'magmatic' hydrothermal fluids play an essential role in transporting heat and at least some dissolved constituents from deep magma sources to near-surface hydrothermal systems. This postulate is difficult to test inasmuch as the near-surface geothermal systems may dilute any 'magmatic' fluid signature. However, the most abundant mafic magmas in greenstone belts, namely basalts, and the most pronounced gold-associated felsic magmas, the quartz-feldspar-porphyries of trondhjemite composition, are notably anhydrous in contradistinction to the 'granitic' magmas associated with copper and molybdenum porphyries which may have relatively high volatile contents.

It should be emphasized that the intrusives encompass a wide range of compositions, from diorite through granodiorite to high K-granites, as well as syenites, alkali gabbros and trondhjemites (Kerrich, 1983; Kerrich and Watson, 1984; Colvine *et al.*, 1984). Moreover, the association of 'felsic' intrusives with gold deposits is not apparent in the southern African and Australian greenstone belts (cf. Groves and Batt, 1984). If Archaean lode gold deposits, worldwide, formed via a similar mechanism, as indicated by uniformity of $\delta^{18}\text{O}$ quartz, $\delta^{13}\text{C}$ carbonate (etc.), then the role of felsic intrusions as a structural focus seems more likely than as a metal donor.

2. Lateral Secretion

The possibility of lateral secretion of material into veins from their enclosing volcanic rocks (Hurst, 1935; Boyle, 1961, 1976) may be evaluated with reference to the rate equations which describe possible independent mechanisms of chemical transport. Chemical transport through the crust may occur by means of (1) solid-state diffusion through crystalline structures; (2) grain-boundary diffusion; and (3) hydrothermal transport in solution.

An approximation of the characteristic transport distance (X) in a diffusive process is given by: $X = (2Dt)^{0.5}$, where D is the coefficient of diffusion at a specified temperature and t is time in seconds. The coefficients of solid-state diffusion, and grain-boundary diffusion through an inter-granular fluid, at temperatures of 500°C are of the order of $10^{-20} \text{ cm}^2\text{s}^{-1}$ and $10^{-6} \text{ cm}^2\text{s}^{-1}$, yielding characteristic transport distances of 800 m and 80 m, respectively, over a time interval of 10⁶ years (Fletcher and Hofmann, 1974; Fisher and Elliott, 1974; Ildefonse and Gabis, 1976). This distance of grain-boundary diffusion is too small by a factor of 100 in view of the calculations on gold transport given by Helgeson and Garrels (1968), Fyfe and Henley (1973), and Kerrich and Fryer (1979). The driving force for lateral diffusion is presumed to be gradients of chemical potential from wall rocks into dilatant zones of lower pressure and chemical potential. However, as discussed in a previous section, it can be deduced from structural relations that many auriferous vein systems underwent incremental opening under conditions of anomalously high fluid pressure, where $P_{\text{fluid}} > P_{\text{wallrock}}$. If the inferred hydraulic pressure relations are correct, then the very basis of lateral secretion is invalidated for the case of auriferous veins. Anomalously high fluid pressures in veins, with limited penetration into wall rocks, is entirely

consistent with the gradients in element abundances, redox, and ^{18}O in a direction orthogonal to the vein boundaries.

The 'lateral diffusion' or 'lateral secretion' mechanism has been supported by Wanless *et al.* (1969), and independently advanced by Roslyakova and Roslyakova (1972) to account for a large number of Russian lode gold deposits. However, Ames (1964) and Ridge (1968) argue that none of the data reported by Boyle (1961) in support of 'lateral diffusion' are incompatible with a hydrothermal origin for the deposits.

In addition, the hypothesis of lateral diffusion is difficult to reconcile with the following observations:

- (a) adjacent to vein margins the profiles of gold and silica abundances, representing the putative chemical potential gradient to drive diffusion, are opposed according to the diagrams of Boyle (1961, 1979), and yet these two components are both interpreted to have diffused into the veins in response to potential gradients;
- (b) the inferred depletion of Al_2O_3 in dilatant zones would require diffusion of this low-mobility component out of the system coupled with diffusion of silica;
- (c) lateral diffusion is not consistent with the oxygen isotope data, which indicate isotopic equilibrium of veins and immediate wall rocks, but disequilibrium of these with country rocks at distances of > 20 m from veins;
- (d) this proposed diffusion mechanism does not account for the highly reduced state of wall rocks;
- (e) nor for the observed separation of base metals from rare elements; nor (f) for the predominance of lode gold deposits in greenstone belts vs. Phanerozoic and Proterozoic metamorphic terrane.

Finally, if lateral diffusion operated, a relationship might be anticipated between the distinctive trace-element suites of various rock types and their abundances in veins, but this is not the case (Kerrich, 1983, Tables I and II). For instance, lead does not exhibit preferential enrichment in veins traversing felsic igneous rocks, nor is copper more abundant in mafic-hosted veins.

3. Kenoran Batholiths as a Source

Derivation of gold from granitic batholiths of Kenoran age is difficult to evaluate in the absence of adequate geochemical tracers.

4. Au Donation from 'Special' Source Rock Types

The source of gold in lode deposits has been a long-standing question. Such deposits display field associations with both mafic-ultramafic volcanic rocks and/or sediments or the lower stratigraphic levels of greenstone belts (Viljoen *et al.*, 1970; Anhaeusser, 1976; Hutchinson, 1976; Pyke, 1975, 1976); and with felsic intrusive rocks emplaced later in the overall development of the volcanic complexes (Goodwin, 1965; Ridler, 1970; Goodwin and Ridler, 1977; Hodgson, 1983).

A direct genetic link between ultramafic rocks and lode-gold deposits has been proposed by Viljoen *et al.* (1970) and Pyke (1975, 1976), based on the assumption of source rocks having an enhanced primary Au content. The elevated gold abundances quoted to substantiate this were obtained by fire assay, which is now known to yield anomalously high values at low abundance levels. Studies of precious-metal contents in igneous rocks determined by more reliable analytical techniques have conclusively demonstrated that ultramafic rocks have lower Au (0.75 ppb) and Ag (40 ppb) compared to primary basalts, which contain an average 2 ppb Au and 100 ppb Ag (Table I; Tilling *et al.*, 1973; Frueh and Vincent, 1974; Anhaeusser *et al.*, 1975; Kwong and Crocket, 1978). The studies of an Archaean greenstone belt show that all unaltered igneous and sedimentary rocks have < 2 ppb Au, and carbonated ultramafic rocks are only weakly anomalous (4 ppb). For the Timmins District, Fyon and Crocket (1982) report that the mean Au abundance of Mg-tholeiitic basalts of 2.4 ppb is greater than that of ultramafic rocks (mean 1.9 ppb); and Saager *et al.* (1982) did not detect a differentiation trend of Au from ultramafic to mafic rocks in greenstone belts of southern Africa.

The conclusion that emerges from these studies is that the spatial association between lode-gold deposits and specific rock types does not provide unequivocal evidence for a genetic link between host rock and lode. On the contrary, the magnitude of transport distances, as derived in a previous section, contradicts close spatial relationships being correlated with genetic factors. Given that ultramafic rocks do not represent an intrinsically gold-enriched source rock, then the observed local association of gold deposits with rocks of ultramafic affinity had been attributed to several factors, including:

- (i) preferential gold release during carbonatization (Fyon *et al.*, 1982);

- (ii) ultramafic rocks signal the onset of new volcanic cycles, with quiescent episodes favouring seawater circulation and attendant chemical sedimentation (Fyon *et al.*, 1982);
- (iii) ultramafic rocks provide effective sinks for gold from hydrothermal solutions (Kerrich and Fyfe, 1981); and
- (iv) ultramafic rocks represent the base of a submarine rift environment, of high heat flow, in which mafic and ultramafic rocks initially store hydrates, carbonates and hydrocarbons during reaction with seawater, and subsequently undergo continuous dehydration and decarbonation during progressive burial (Kerrich, 1983; Groves and Batt, 1984).

Hodgson and MacGeehan (1982, p. 225) have proposed, as one of several possible alternatives, that the spatial association of gold deposits in greenstone belts with detrital sedimentary rocks may be due to preconcentration of gold and/or the gold-mobilizing constituents into the sediments. The putative gold enrichments in detrital sedimentary rocks can be discounted for Abitibi belt gold deposits from the available data on gold abundances in this rock type (Kwong and Crocket, 1978; Table I), although it is not possible to comment on the unspecified 'gold-mobilizing constituents' of Hodgson and MacGeehan (1982). However, Saager *et al.* (1982) report that sedimentary rocks (mean Au 130 ppb) are significantly enriched in gold over volcanic rocks (mean Au 10.8 ppb) in southern African greenstone belts. Thus, several authors have concluded that the bulk of rocks in Archaean greenstone belts are not better gold source beds than equivalent rock types of the Proterozoic or Phanerozoic eras, except for the volumetrically small iron formations (Kwong and Crocket, 1978; Fryer *et al.*, 1979a; Kerrich and Fryer, 1979; Kerrich, 1980; Fyon *et al.*, 1982; Saager *et al.*, 1982).

The preferred spatial association of lode gold deposits, of both chemical sedimentary and vein types, with the boundary between volcanic and sedimentary domains in greenstone belts is attributed here to the controls of structure and permeability on discharge of ponded hydrothermal reservoirs.

5. Multistage Origin

(a): Seawater Transport of Au, With or Without Remobilization

There exists a large body of field, sedimentological, textural, isotopic, and REE evidence for some enrichment of gold into chemical sediments (Hutchinson, 1976; Fripp, 1976; Ridler, 1976; Karvinen, 1978, 1980; Fryer *et al.*, 1979; Fyon *et al.*, 1980, 1981). Saager *et al.* (1982) report that ferruginous chemical sediments in greenstone belts of southern Africa are distinctly more auriferous (mean Au 172 ppb) than Superior iron formations. However, it is not known with certainty if the origin of the auriferous chemical sediments was primary, secondary replacement, or a combination of the two processes.

Several authors tacitly have presumed that convecting seawater hydrothermal fluids form syngenetic sedimentary gold enrichment (Fripp, 1976; Fyon and Crocket, 1981; Fyon *et al.*, 1982; Hodgson and MacGeehan, 1982; MacGeehan and Hodgson, 1981, 1982; MacGeehan *et al.*, 1982), although no direct evidence is presented for seawater involvement. However, the balance of available evidence suggests that the auriferous fluids implicated in gold deposits were different from seawater-geothermal systems generating base metal deposits, in terms of temperature, $\delta^{18}\text{O}$, fluid pressure, Na/K, CO₂/H₂O, salinity, Mg/Fe, redox, relative metal solute concentrations, and reduced gas species. These differences in fluid properties are reflected in divergent metal distributions and patterns of alteration of the two deposit types (see Figs. 1-3).

Furthermore, any model that involves seawater cannot readily account for the predominance of lode gold deposits in Archaean greenstone belts coupled with their relative paucity in the Proterozoic and Phanerozoic marine environment. Base-metal deposits which are generally considered to have been formed from seawater-hydrothermal fluids exist in sequences of most age from the Archaean to the present. However, Archaean-type lode gold deposits are not prominent in Phanerozoic mid-ocean-ridge, island arc, or ocean island environments. Extensive carbonate alteration, LIL element enrichments, massive quartz veins and lode gold deposits are known neither from observations of the modern ocean-floor basalts subjected to seawater alteration nor to obducted fragments of ocean crust. If, however, the Archaean auriferous sediments were deposited from metamorphic fluids debouching into the near-seafloor environment, then the secular variation is possibly explicable: in the Archaean the products of greenstone-belt dehydration would have debouched back onto the ocean floor, whereas in the Phanerozoic dehydration of the ocean floor proceeds in subduction zones.

It has been suggested that gold may locally be mobilized out of chemical sedimentary rocks with a postulated primary Au enrichment during crustal deformation (Hutchinson, 1976; Fripp, 1976; Ridler, 1976; Karvinen, 1978,

1980; Fyon *et al.*, 1980, 1981), leading to vein deposits in the vicinity. However, most vein deposits do not appear to have acquired their precious-metal complement from local remobilization of gold in the sediments. The putative remobilization does not account for separation of rare from abundant elements in the veins, for K₂O enrichments coupled to Na₂O depletions around veins, or for the pronounced differences in REE distributions, carbonate mineral chemistry, δ¹⁸O, and redox between sediments and veins.

Hodgson and MacGeehan (1982) have likened Archaean gold deposits in proximity to epizonal intrusions to precious metal deposits associated with Tertiary epizonal stocks of the western USA. However, the latter typically possess base metal and precious metal domains, unlike the Archaean examples (cf. Sunnyside Mine, Casadevall and Ohmoto, 1977). The only feature in common with the precious metal zones of the two ages of epizonal deposits is a low-chloride hydrothermal reservoir: low-Cl, high-¹⁸O (metamorphic) fluids in a submarine environment for the Archaean; and low-Cl, low-¹⁸O (meteoric) fluids in a continental environment for the Tertiary.

Macdonald (1983), Phillips *et al.* (1984), and Pearton (personal communication, 1985) have provided genetic field and mineralogical evidence for an epigenetic origin of iron-formation hosted gold deposits.

6. Multistage Metamorphic Fluid Systems

The remaining hypothesis presented to account for the origin of lode gold deposits involves focussed discharge of metamorphic fluids. Single-stage models for veins have been proposed by Fyfe and Henley (1976) and Henley *et al.* (1976), and multistage models by Kerrich and Fryer (1979), Kerrich and Fyfe (1981), Kerrich and Hodder (1982), and Kerrich (1979, 1980, 1981, 1983). During burial of hydrated greenstone-belt assemblages, devapourization reactions generate significant fluid volumes (Fyfe, 1973, 1974; Fyfe *et al.*, 1978). Fluids are continuously released within the greenschist facies; and at the greenschist-amphibolite transition, which takes place at 450–500°C, about 5 wt % structural water and volatiles, including CO₂, are released from a hydrated rock of mafic composition. Large fluid volumes are also generated from the dehydration of zeolites and clays at lower grades of metamorphism.

The volume relations of dehydration reactions are such that metamorphic fluids are evolved under conditions where p fluid ≥ P lithostatic. A second factor contributing to elevated hydraulic pressures is differential thermal expansion of pore fluids (Norris and Henley, 1976; Knapp and Knight, 1976). This effect is enhanced under conditions of steep geothermal gradients. Hence, a metamorphic fluid reservoir satisfies the observed temperatures, δ¹⁸O and δD of the mineralizing solutions, together with the low salinity, and elevated CO₂/H₂O plus K/Na. In addition, a metamorphic reservoir accounts for the requirement of fluids generated under high temperature reducing conditions, at elevated hydraulic pressures and inferred low water/rock ratios. As stated above, a magmatic hydrothermal fluid, or mixtures of magmatic and metamorphic are also possible based on present data.

A common argument raised against the involvement of metamorphic fluids in Archaean lode gold deposits is the lack of evidence for prograde regional metamorphism in the vicinity of mines. However, fluids are continuously released from hydrated minerals during prograde metamorphism (Fyfe *et al.*, 1978). Thus, the presence of metamorphic isograds indicating higher-grade conditions around gold deposits is not a necessary condition to make the hypothesis viable, as asserted by Hodgson and MacGeehan (1982). Rather, the presence of intrusive-extrusive quartz-feldspar porphyry bodies which are commonly associated with, and emplaced coevally with, lode gold deposits attests to higher-temperature conditions at lower crustal levels. These quartz-feldspar-porphries, which are typically of trondhjemitic composition, are generally considered to be the product of partial melting of mafic granulites (Condie, 1980), requiring the presence of higher-grade metamorphic rocks between the deposit and deeper-level regions of partial melting. Hence, the postulated metamorphic fluid reservoirs and trondhjemitic magmas are temporally related expressions of a mafic crust undergoing dehydration at higher levels and melting at deeper levels.

The residence and accumulation of metamorphic fluids in the crust is determined by the balance between the rates of fluid generation and fluid expulsion. In the Archaean crust where thermal gradients are generally accepted to have been high, particularly in provinces of continuous volcanic activity, the rate of fluid generation by metamorphic dehydration reactions must have been faster and occurred over narrower vertical crustal intervals, leading to an increased probability of generating and ponding extensive metamorphic fluid reservoirs. When the hydraulic pressure exceeds the sum of the crustal horizontal stress plus the tensile strength of the rock, subvertical hydraulic fracturing must ensue, providing focused discharge for the ponded reservoir, and high fluid/rock ratios in the conduit. High boron and CO₂ concentrations in many gold-bearing veins and chemical

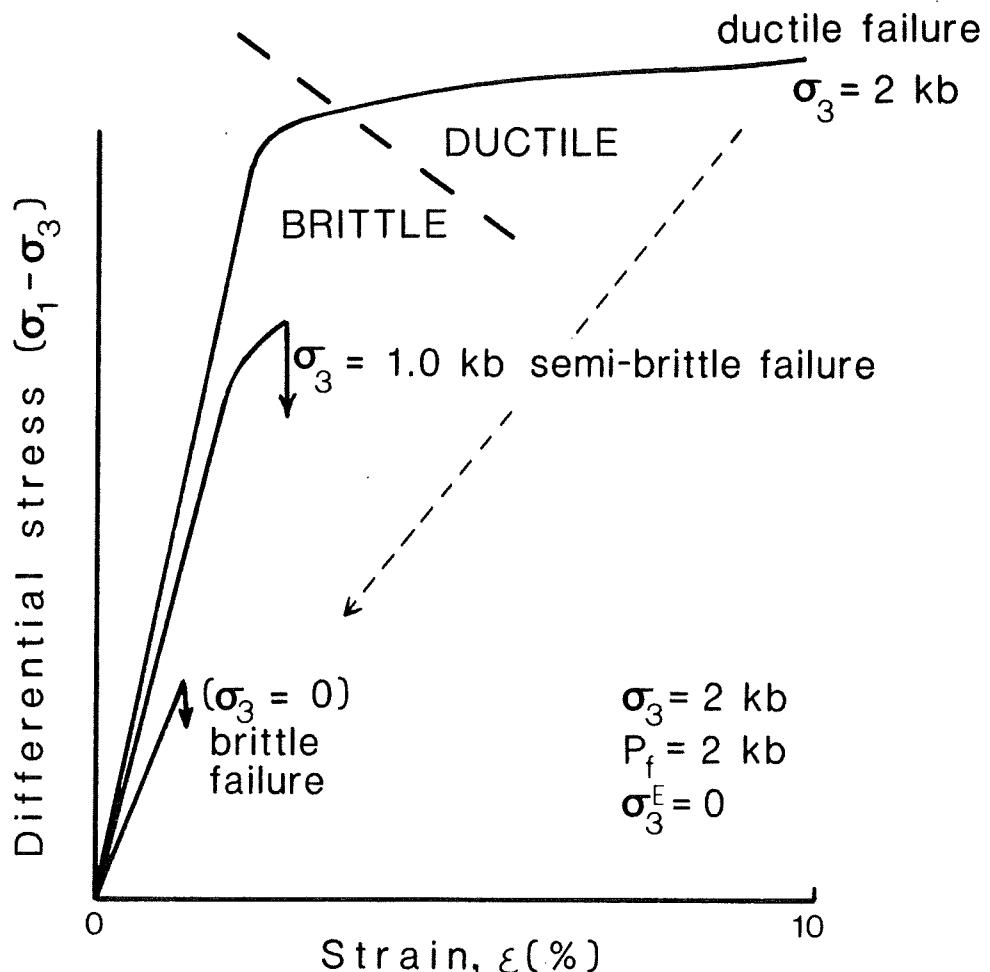


Figure 13 : Generalized rheological behaviour of rock, illustrated in differential stress ($\sigma_1 - \sigma_3$) versus strain, co-ordinate space. Note the transition from brittle to ductile behaviour at elevated values of confining stress (σ_3). Switching from ductile to brittle behaviour can be induced by transient increases of fluid pressure which act to diminish the effective confining stress ($\sigma_3 - P_{\text{fluid}}$).

sediments may be a consequence of ponding hydrothermal reservoirs beneath impermeable sediments in source regions; inasmuch as sediments have greater abundance of B than igneous rocks and typically contain hydrocarbons which may oxidize to CO₂ (Turekian and Wedepohl, 1961). Haak *et al.* (1984) has demonstrated the loss of many trace metals, especially the LIL elements - Tl, Ba, Pb, Bi, Hg, Zn, and Cd, from pelites during regional metamorphism, and most granulites are known to be depleted in LIL elements (cf. Barbey and Cuney, 1982).

Subsequent generations of vein deposits may form in response to additional burial and heating of rocks arising from superimposed thrust sheets. The ubiquitous conjunction of brittle fracturing and ductile deformation in lode gold deposits attests to cycling through the brittle-ductile transition at depths of ~ 8-12 km in response to fluctuations of fluid pressure (Fig. 13).

It is suggested that rift environments, with high heat fluxes, are appropriate for efficient gold concentration (Kerrich, 1983; Groves and Batt, 1984). Here, listric normal faults initially control extension along with ultramafic volcanism, provide pathways for mafic and felsic magmas, are responsible for scarp-dominated sedimentary rocks, and act as conduits for hydrothermal discharge. At late stages of greenstone belt development, reverse motion on the faults accommodates crustal compression, along with continued hydrothermal discharge.

The structural setting of Archaean lode gold deposits has been emphasized by many workers from the earliest times, such as Knopf (1929), Bain (1933) and authors of 'The Structural Geology of Canadian Ore Deposits' (1948). Gunning and Ambrose (1940) have emphasized the association of gold deposits with 'breaks', where this term carries the connotation of linear domains including gradients of deformation intensity in conjunction with the juxtaposition of diverse rock types, typically volcanic and sedimentary.

Structural control of gold deposits, and the role of structure in generating large plumbing systems has subsequently been endorsed by Fyfe and Henley (1975), Fryer *et al.* (1979), Kerrich and Allison (1978), Kerrich and Fryer (1979), Kerrich (1983), this paper, and recently by Colvine *et al.* (1984). Thus, structure is of paramount importance for both chemical sedimentary and vein-type lode gold deposits.

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