GOLD MINERALIZATION IN ARCHAEAN CHERTS AND IRON-FORMATIONS

A review of the Economic Geology

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ABSTRACT

The distribution of gold in igneous rocks and minerals is described and discussed. Not all the gold in igneous rocks is contained within early formed crystal lattices. Evidence that gold can be associated with late stage crystallizing phases is described. It is concluded that some of the gold in hydrothermal and volcanogenic deposits may have come from a primary magmatic source rather than having been leached from solid country rock.

Gold is probably transported as chloride complexes at temperatures greater than about 300°C. At lower temperatures it is probably transported with other metals as sulphide and thio-sulphide complexes. The precipitation of gold from the transporting medium is brought about by changes in the physico-chemical conditions within that medium. Decrease in pressure is probably not a major cause of precipitation in volcanogenic environments.

The geology of volcanogenic iron-formations is described and discussed, relative to the development of greenstone belts. Oxide facies iron-formations were formed in shallow oxidizing environments. They are associated with volcanogenic and clastic sediments. Sulphide facies iron-formations were precipitated in the deeper parts of geosynclinal structures. They are associated with mafic and ultramafic rocks similar to modern oceanic volcanic assemblages. Carbonate facies iron-formations were deposited in the regions between oxide facies and sulphide facies. Other banded iron-formations are found associated with base metal massive sulphide deposits related to arc-type volcanic centres. These deposits are found in the regions where carbonate facies iron-formations were formed.

Exploration for and exploitation of gold deposits in Archaean iron-formations are discussed. Geochemical exploration programmes are aided by the association of gold with trace amounts of base metals. Geophysical exploration methods that can be employed include magnetometer, I.P. and E.M. surveys. The metallurgical treatment of the ores should include "roasting" because a large proportion of the gold occurs as submicroscopic grains within sulphide mineral crystals.

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ABSTRACT

INTRODUCTION

Gold has been the base of financial and commercial transactions for many centuries. Despite attempts in 1976 to demote gold to the status of a mere commodity it appears that this precious metal will remain a major element of international exchange (Tremblay and Descarreaux, 1978). This, plus the concern for inflation and the devaluation of national currencies, has resulted in renewed interest in gold.

The concern for energy supplies and costs, and for the capital costs of large projects (Tremblay and Descarreaux, 1978), suggest that smaller mining targets which require less capital outlay and have lower energy consumptions, may be the economical mining propositions of the near future. The presence of large numbers of gold deposits associated with Archaean iron-formations would indicate that in the present economical climate these deposits should be investigated further.

The purpose of this study is to compile a comprehensive description of gold deposits associated with Archaean iron-formations with a view to aiding the exploration for, and exploitation of these deposits in the future.

Historical Background

Gold has been produced intermittently from Archaean terrains for many centuries. In the Kolar Goldfield, India, gold has been won for more than 1000 years (Narayanaswami *et al*. 1960). Summers (1969) suggests that even earlier miners in other regions of Mysore State, India, provided the technical expertise for the development of the first ancient workings, in what is now Rhodesia, some 2000 years ago. These ancient workings were first discovered by the elephant hunter, Henry Hartley, in 1865. A year later Carl Mauch confirmed these findings and the Umfuli goldfield of Rhodesia and the Tati goldfield of Botswana were discovered (Phaup, 1972).

Brazil has a recorded gold production dating from the seventeenth century, most of which probably came from the Archaean terrain in the province of Minas Gerais (Anhaeusser, 1976). The earlier indian civilizations in South America probably obtained at least some of their gold from Archaean terrains on this continent before this time.

In Western Australia gold was first discovered in 1885 by the early European prospectors (Woodall, 1975). The Aborigines had no interest in gold so it is probable that this was the first discovery of gold in Archaean rocks on the continent.

The Superior and Slave Provinces of Canada have produced more gold than any of the other Archaean terrains, but this is because these Archaean terrains are aerially so much larger. The Archaean areas of Southern Africa have been the most prolific producers of gold per unit area with a production of 91 oz per km^2, followed by the Superior and Slave Provinces of Canada (79 oz per km^2) and Pilbara and Yilgarn Blocks, Western Australia (58 oz per km^2) (Anhaeusser, 1976).

Nomenclature

In past years there has been much confusion over the terms to describe iron-rich sediments whether they are of volcanogenic or sedimentary origin. Not only have different terms been used in different countries for the same rocks, but identical terms are sometimes used for different rocks as well. Terms that have been used include itabirite (Brazil), jaspilite (Australia) and banded-ironstone (Southern Africa).

Van Hise and Leith (1911) used the term "iron-bearing formation" to describe iron-rich rocks in the Lake Superior area of the United States. This term was adapted by James (1954) to "iron-formation". Iron-formation was defined by him as "... a chemical sediment, typically thin-bedded or laminated, containing 15% or more iron of sedimentary origin, commonly, but not necessarily containing layers of chert".

Brandt et al. (1972) suggest that the term "iron formation" be used as a generic lithologic term. They also suggest that individual workers define their own terms.

In this dissertation the term "iron-formation" will include all Archaean ferruginous and/or silicious chemically precipitated sedimentary rocks of volcanic origin regardless of facies-type and metamorphic grade, unless otherwise defined.

Facies of Iron-Formation

It is generally accepted now that iron-formations are chemically precipitated sediments. As is commonly known, iron occurs in nature in different oxidation states, which depend on the environment of deposition. The concept of facies variations in iron-formations has been discussed at length (Stanton, 1972; James, 1951, 1954, 1966; Goodwin, 1956; Borchert, 1960; Gross, 1965). James (1954) and Gross (1965) have emphasized the strong relationship between the composition and mineralogy of chemically precipitated iron-rich sediments and their oxidation states. Figures 1 and 2 illustrate some possible physico-chemical conditions under which the various iron minerals may be precipitated. It is interesting to note that at relatively higher pH conditions and lower Eh conditions, in the presence of amorphous silica, instead of magnetite, iron silicate is precipitated.

Stanton, 1972, "Ore Petrology", McGraw-Hill Book Co.

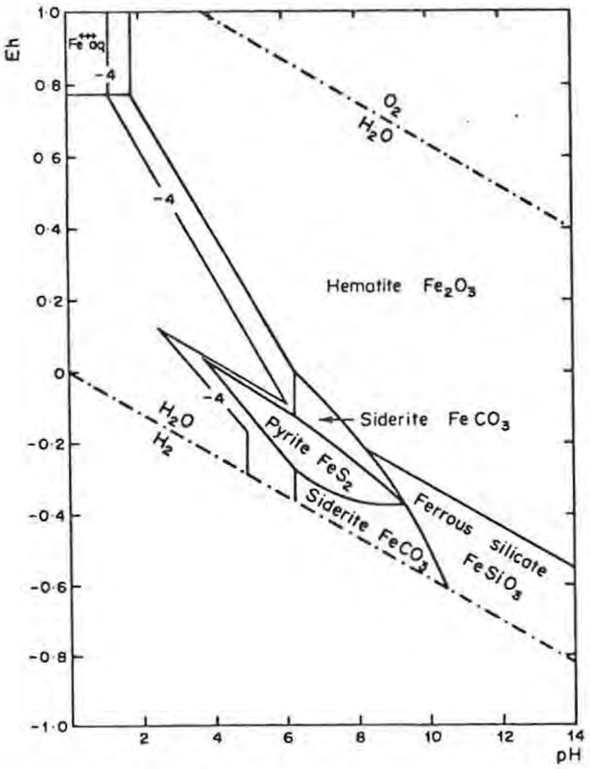


Fig. 1 Eh-pH stability relations among iron oxides, carbonate, sulphide, and silicate at 25°C, 1 atm total pressure, total $CO_2 = 10^{0}$ m, total sulphur = 10^{-6} m, and in the presence of amorphous silica. (Redrawn from Garrels, "Mineral Equilibria", after fig. 6.23 (p. 161), Harper & Brothers, New York, 1960).

Stanton, 1972, Ore Petrology, McGraw-Hill Book Co.

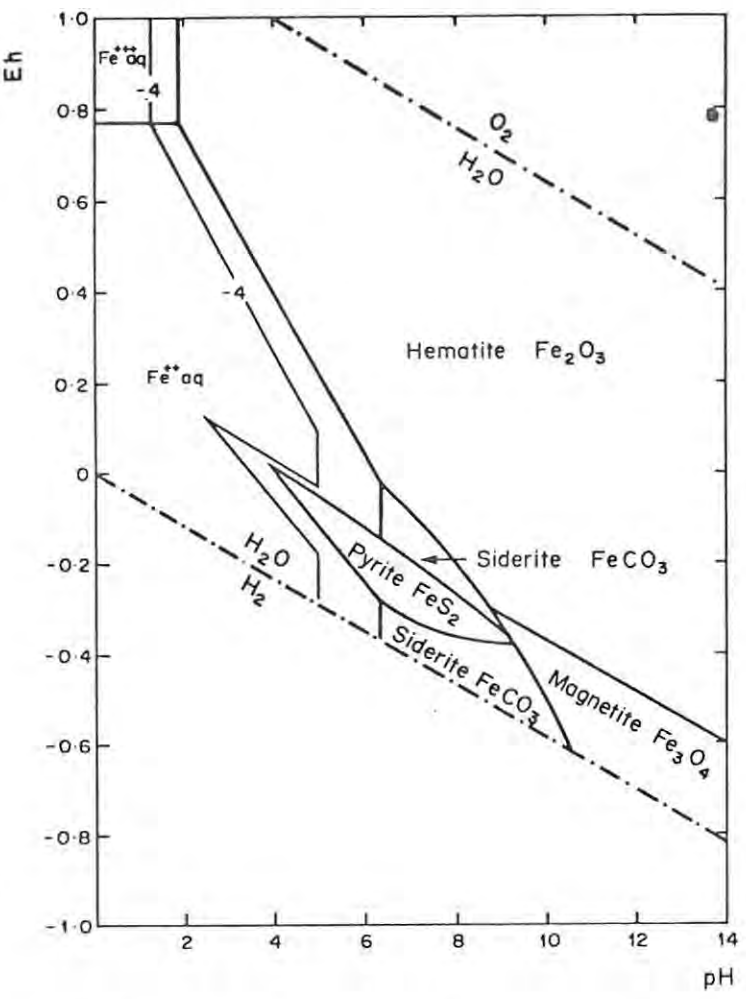


Fig. 2 Eh-pH stability relations among iron oxides, pyrite and siderite at 25°C, 1 atm total pressure, total dissolved carbonate = 10^{0} m, total dissolved sulphur = 10^{-6} m. Compare with Fig. 1 (Redrawn from Garrels, "Mineral Equilbria", after Fig. 6.21 (p. 157), Harper & Brothers, New York, 1960).

Harder (1919), because of the wide variety of chemical and mineralogical compositions of iron-formations, suggested using the end members of the groups of minerals; namely oxides, silicates, carbonates and sulphides as a means of classifying the rock types. He used primary unmetamorphosed minerals in his classification :

Oxide facies : hematite, magnetite, goethite
Carbonate facies : siderite
Silicate facies : greenalite, chamosite, minnesotaite, stilpnomelane
Sulphide facies : pyrite, marcasite, greigite, melnikovite ("pyrite gel")

Table 1 shows features typical of the facies of iron-formations.

Fripp (1974) working in a region of greenschist metamorphic grade used metamorphic equivalents of these minerals in his classification :

Oxide facies : magnetite, hematite
Carbonate facies : siderite, ferroan dolomite
Silicate facies : tremolite, grunerite, hypersthene, iron-chlorite
Sulphide facies : pyrrhotite, arsenopyrite, pyrite

Other mineral assemblages typical of metamorphosed iron-formations are given in Table 2.

Another type of facies variation occurs when the proportions of the constituent minerals change. At the one extreme where no iron minerals are present the rock is a plain or laminated chert; or in the case of the sulphide facies, a carbonaceous shale. At the other extreme the rock may be almost mono- or bi-mineralic, comprising the mineral(s) typical of that chemical facies type.

In metamorphic terrains the problem of nomenclature with respect to cherts and their metamorphic equivalents arises. Fripp (1974) overcame the problem by defining chert as a siliceous rock in which individual quartz grains could not be discerned with the aid of a hand-lens.

J. Eichler, 1976, in Wolfe, K.H. (ed.),"Handbook of Stratabound and Stratiform Ore Deposits", Pt II, Vol. 7. Elsevier Scientific Publishing Company.

TABLE 1 Special features of sedimentary facies of iron-formations (slightly modified from James, 1954).

	Oxide-facies		Silicate-facies	Carbonate-facies	Sulfide facies
	hematitic	magnetitic			
Lithology	evenly to maculate banded; alternating layers of hematite and quartz (chert)	evenly to maculate banded; layers of magnetite and silicate - carbonate- chert	± well banded granular green silicates, interlayered with quartz (chert) and magnetite	± well banded layers of carbonates and quartz (chert)	finely banded pyritic and carbonaceous slate with chert
Iron minerals	hematite (martite)	magnetite	hydrous iron silicates; greenalite; stilpnomelane; minnesotaite: (carbonates, iron oxides)	iron carbonates siderite. ankerite (magnetite, pyrite, iron silicates)	pyrite (iron silicates, carbonates)
Iron content (average % Fe)	30–42 (ϕ38)	25–35	20–30	20–35 (ϕ21)	15–25
Special features	no detrital dilution; (oolites)	granules, oolites	transitional to magnetite-oxide and carbonate-facies	stylolites, graphitic	commonly graphitic
Chemical environment	strongly oxidizing	weakly oxidizing to weakly reducing	weakly reducing	reducing	strongly reducing

J. Eichler, 1976, in Wolfe, K.H. (ed.),"Handbook of Stratabound and Stratiform Ore Deposits", Pt II, Vol. 7. Elsevier Scientific Publishing Company.

TABLE 2 Comparitive mineral assemblages in diagnetic and metamorphic iron-formation.

Facies	Oxide	Silicate	Carbonate	Sulfide
Diagenetic pre-metamorphic	Fe-hydroxides hydrohematite magnetite	Fe-SiO_2-gels (cryptocrystalline) greenalite, chamosite (magnetite, carbonates)	Fe-carbonates (Fe-silicates, magnetite)	Fe-sulfides in bituminous slate
	chert, cryptocrystalline quartz	chert (quartz)	chert (quartz)	chert
Low-grade metamorphic	hematite magnetite	chlorite minnesotaite stilpnomelane chamosite (?) (clay minerals)	siderite ankerite magnetite Fe-silicates (clay minerals)	pyrite graphite magnetite carbonates (clay minerals)
	quartz	quartz	quartz	quartz
Metamorphic	hematite magnetite (martite)	minnesotaite stilpnomelane cummingtonite grunerite riebeckite micas (olivine)	siderite ankerite magnetite clino-pyroxene ortho-pyroxene riebeckite	pyrrhotite pyrite magnetite graphite sericite
	quartz	quartz	quartz	quartz
Fe-content (%)	30–42 (ϕ38)	20–30 (ϕ26)	20–35 (ϕ21)	15–25 (ϕ20)

THE OCCURRENCE AND DISTRIBUTION OF GOLD IN IGNEOUS ROCKS IN THE EARTHS CRUST

The development of neutron activation analysis in the 1960's led to the accurate determination of gold at ppb levels. Until that time analysis by fire assay had been the major method of analysis for gold. In cases where gold occurred in the ppb range it had been necessary to extract and concentrate the metal before assaying.

The average gold concentration in igneous rocks is in the range of 1,0 - 5,0 ppb (Crockett, 1974; Tilling et al., 1973; Gottfried et al., 1972). Crockett (1974) has reviewed the literature and has come to the conclusion, supported by the majority of the authors, that there is a decrease in gold content with magmatic differentiation.

Gold Content of Felsic Plutonic Rocks

In the calc-alkaline suite the range is very narrow (less than 3,0 ppb), but in the alkaline suite the range is even narrower, being about 2,0 ppb (Gottfried et al., 1972; Tilling et al., 1973) (see Figure 3).

From their studies, Tilling et al. (1973) concluded that there is no significant difference between the gold content of the diorite and tonalite group (arithmetic mean = 2,7 ppb) and the gold content of the gabbro group (arithmetic mean = 3,9 ppb) (see Figure 4). However they did find that at the 99% confidence level there is a significant difference between the gold content of the diorite and tonalite group and the gold content of the granodiorite group (arithmetic mean = 1,2 ppb). Similarly the difference between the gold contents of the granodiorite group and the quartz monzonite and granite group (arthmetic mean = 0,75 ppb) is significant at the 95% confidence level. In the alkaline suite the three rock groups (Figure 4) could not be separated statistically on their gold contents (Tilling et al., 1973).

Modified from Tilling et al., 1973, Econ. Geol., vol. 68.

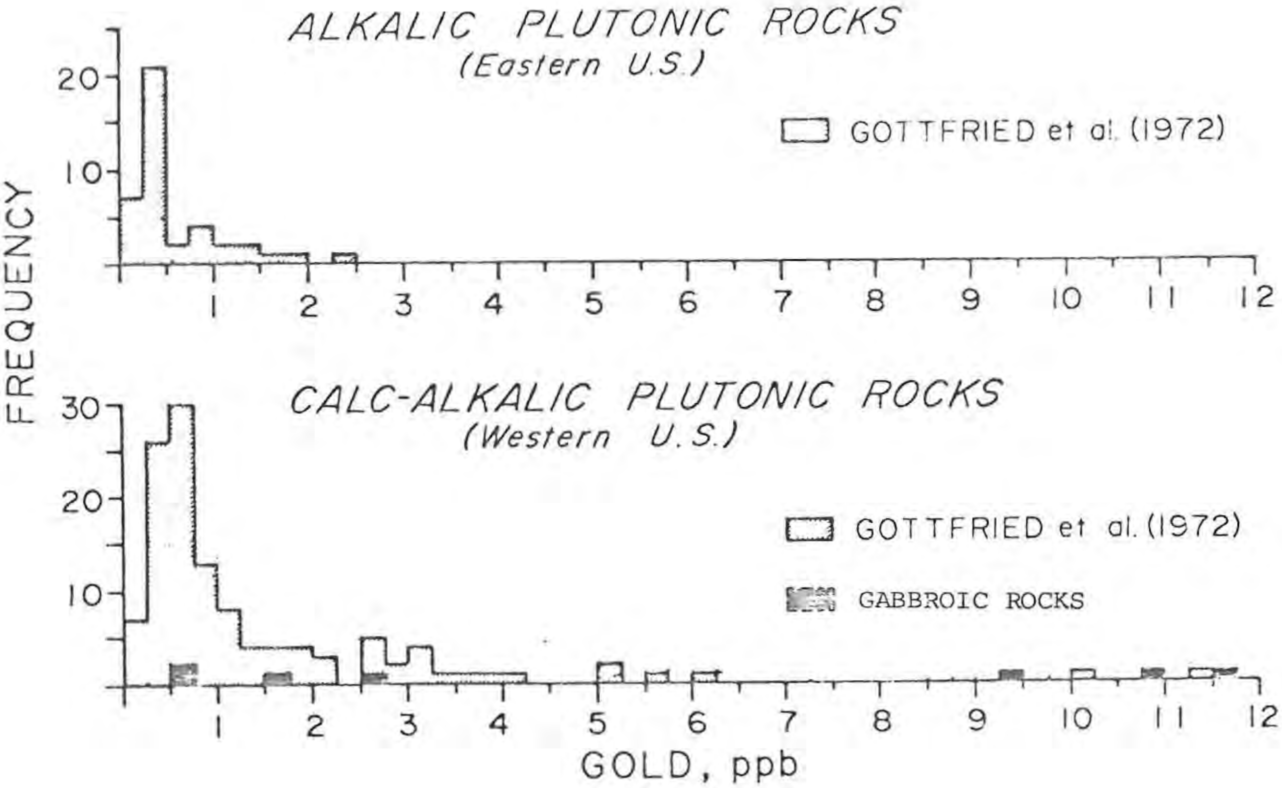


Fig. 3 Composite histograms of gold abundances of unaltered calc-alkalic and alkalic plutonic rocks. The data of Gottfried et al. (1972) for calc-alkalic rocks primarily represent samples from the Sierra Nevada, Southern California, Boulder and Idaho batholiths; data for gabbroic rocks are shown in black. Data for alkalic rocks are based mainly on samples from the White Mountain Plutonic Series, New Hampshire, but also include samples from North and South Carolina.

Tilling et al., 1973, Econ. Geol., vol. 68.

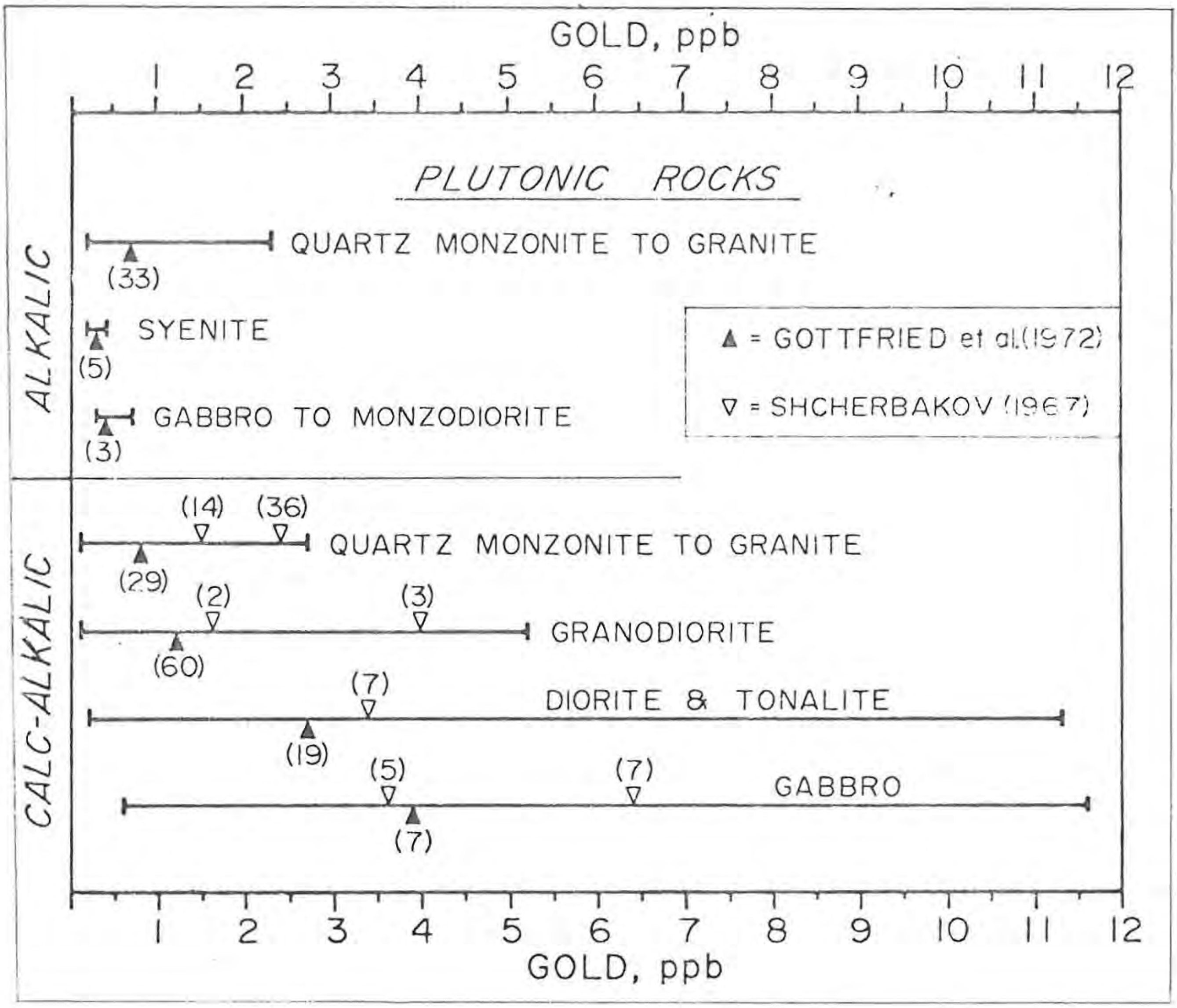


Fig. 4 Gold abundance of unaltered plutonic rocks according to rock type. The bars give the ranges of the observed gold values. The solid triangles give the arithmetic mean values for the data of Gottfried et al. (1972); the open triangles give the means reported by Shcherbakov (1967) for two regions in the U.S.S.R. (Altay-Sayan and Western Chukotka). Numbers in parentheses give the number of samples used in the calculation of mean.

Gold Content of Mafic Rocks

In their papers, Tilling et al (1973) and Gottfried et al. (1972) found that the gold contents of plutonic rocks appeared to be relatively uniform world-wide. One or two exceptions were noted, but their differences could be attributed to the possibility of contamination or to the analytical technique. However, they found that different basaltic provinces could be distinguished and separated by their gold content, and differences between basalts from different tectonic settings were noted. Thus at the 95% confidence level differences based on gold content were detected between oceanic tholeiites, continental tholeiites and circum-Pacific (orogenic) basalts (Tilling et al., 1973).

Tholeiites of the Hawaiian Islands have gold contents in the range 2,2 - 2,4 ppb, oceanic tholeiites from the Mid-Atlantic Ridge and East Pacific Rise have significantly lower gold contents (⩽1,0 ppb). The alkalic basalts from these areas have similar gold contents, less than 0,8 ppb (Gottfried et al., 1972). Tholeiites in continental settings also record large differences in gold content in basalts of similar tectonic setting and gross chemical composition.

The Columbia River basalt (3,7 ppb), the Snake River basalt (0,5 ppb) and the Watching basalt, New Jersey (4,7 ppb) are all statistically different (Tilling et al., 1973) (see Figures 5 and 6).

Tilling et al., 1973, Econ. Geol. Vol. 68.

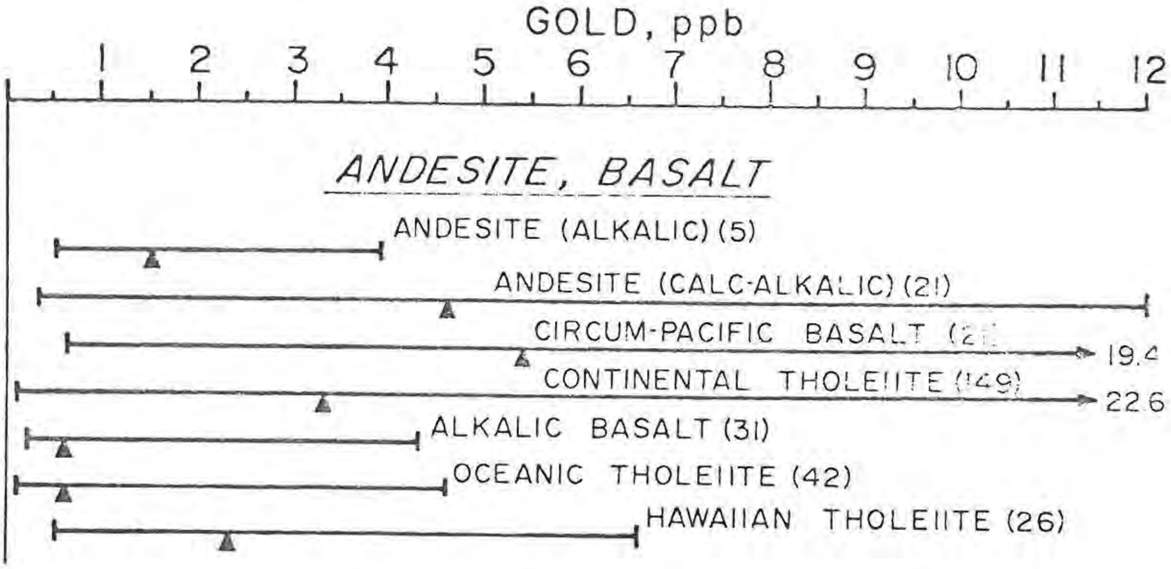


Fig. 5 Gold contents in unaltered volcanic rocks; ranges, means, and number of samples are denoted as in Figure 4. Data source : Gottfried et al. (1972).

Tilling et al., 1973, Econ. Geol., Vol. 68

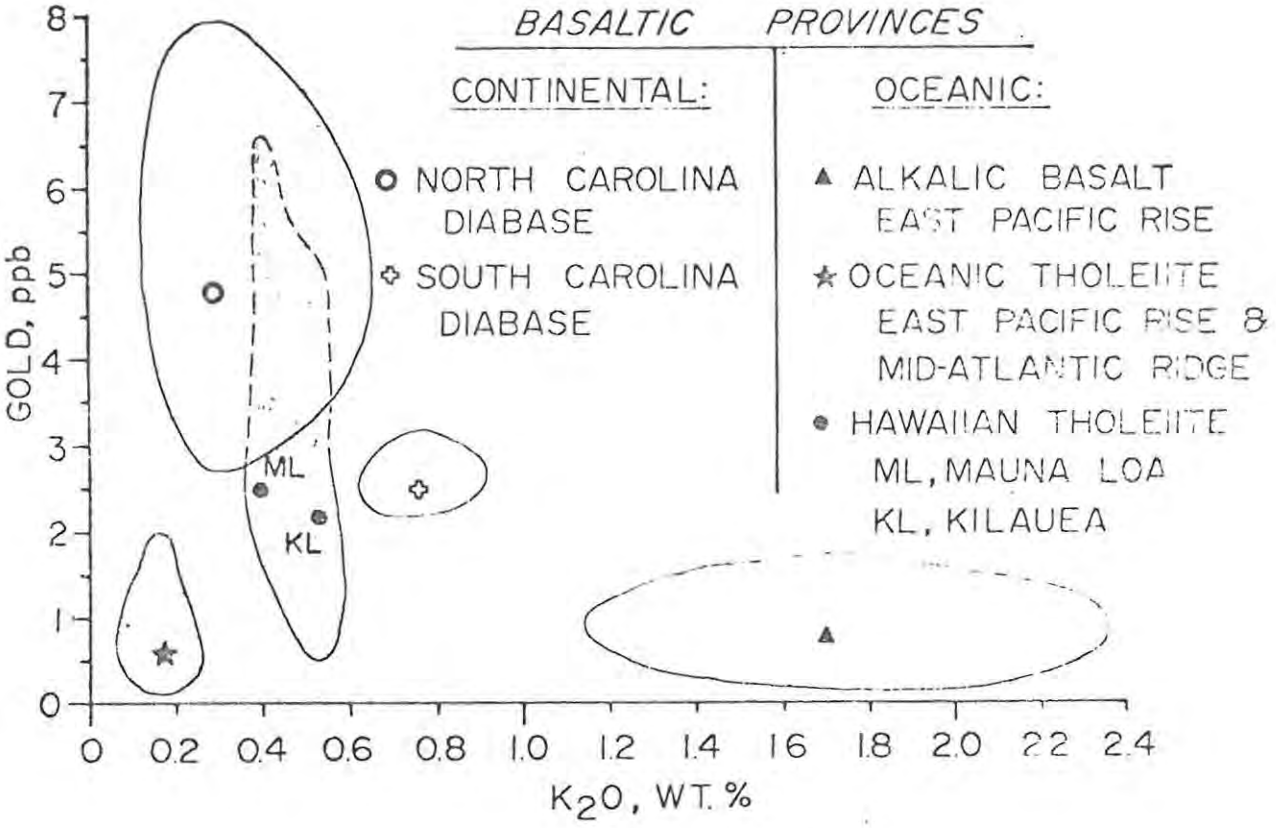
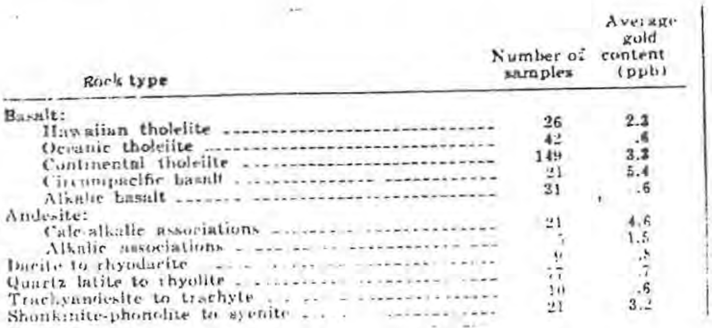


Fig. 6 Graph showing variation of gold content with K_2O for basalts from some oceanic and continental provinces. The mean values are given by the various symbols within each data field. The field for oceanic tholeiite excludes a value of 4,6 ppb gold for an altered sample, and the field for Hawaiian tholeiite is dashed above a concentration level of 3 ppb gold because of the paucity of data points above this level. Data source : Gottfried et al. (1972) and Gottfried and Greenland (1972).

In differentiated mafic bodies gold is either distributed uniformly over the range of rock types (Wager and Brown, 1968) or increases to a maximum in the intermediate stages of differentiation (Rowe, 1969). There is no indication that gold is actively concentrated in the final stages of magmatic crystallization.

The highest gold contents have been found in the basaltic and andesitic members of the circum-Pacific (orogenic belt) calc-alkaline suites. Rhyolites of calc-alkaline series as well as those of alkaline affinities are lower in gold content, regardless of geological setting (Table 3), (Gottfried et al. 1972).

Gottfried et al., 1972, U.S. Geol. Surv. Prof. Paper 727



Rock type	Number of samples	Average gold content (ppb)
Basalt:		
Hawaiian tholeiite	26	2.3
Oceanic tholeiite	42	.6
Continental tholeiite	149	3.3
Circumpacific basalt	21	5.4
Alkalic basalt	31	.6
Andesite:		
Calc-alkalic associations	21	4.6
Alkalic associations	5	1.5
Dacite to rhyodacite	9	.8
Quartz latite to rhyolite	17	.7
Trachyandesite to trachyte	10	.6
Shonkinite-phonolite to syenite	21	3.2

Table 3 Average gold content in selected rocks representative of different volcanic associations

In alkaline provinces, all rock types were found to be low in gold content (Gottfried et al., 1972). These rocks are low in CaO, and high in NaO, but the subsilicic high alkalic rocks from Central Montana (Gottfried et al., 1972) with high K_2O and CaO contents have higher gold contents.

Keays and Scott (1976) found a similar correlation in the mid-ocean ridge basalts in the Atlantic, where basalts with low K_2O contents contain the highest concentrations of gold, although this may be a result of the interaction between the extruded lavas and the sea water.

Gottfried and Greenland (1972) found that in oceanic and continental basalts, high gold values were associated with high K_2O contents and low gold values were associated with low K_2O (mid-oceanic ridge) basalts. They suggest that this is a result of gold being mobilized and redistributed prior to the generation of the magma.

In Archaean rocks it appears that there has been no apparent early enrichment of gold in supra-crustal rocks. Stephenson and Ehmann (1971) show that in the Price Lake - Beresford Lake district in Manitoba the average gold contents of igneous and metamorphic rocks are similar to their younger equivalents. Anhaeusser et al. (1975) concluded that in the Barberton Mountainland of South Africa the range of 1,0 - 1,5 ppb is typical of the tholeiitic basalts elsewhere in the world. They ascribe the earlier, much higher values attained by Viljoen et al. (1970) of 5 - 20 ppb, to less accurate analysis.

Sighinolfi and Santos (1976) working in granulite terrains concluded that the metamorphism apparently has no effect on gold content. They found that the rocks examined had gold contents in the same range as the unmetamorphosed equivalent rocks.

Gold in Minerals

Gold is mono-isotopic consisting of the single nuclide ^{197}Au. Like copper and silver, gold is in group lb of the periodic table. It contains a single 'S' electron outside a complete 'd' shell. Although these three metals have electronic similarities they behave differently in nature. Gold is strongly siderophilic and only slightly chalcophilic, but silver and copper are strongly chalcophilic. This means that in nature gold occurs mostly as the metal or an alloy, usually with silver. Because of their similar atomic radii, gold and silver form a continuous solid solution series. The chalcophilic nature of gold is largely limited to the formation of tellurides (Allmann, 1974).

The gold content of sulphide minerals is in the region of ppm but only at the ppb level in oxides and silicates. In sulphides gold occurs largely as the free metal. It is uncertain if it occurs in these minerals in true isomorphic substitution.

Crockett (1974) has reviewed the literature and concludes that as with the rocks, the mafic minerals have higher gold contents than do the felsic minerals. These conclusions are supported by Tilling et al. (1973) although they doubt the accuracy and the validity of results of other authors (Figure 7).

Wager and Brown (1968) concluded that in the Skaergaard intrusion, gold entered, with equal ease, into plagioclase, pyroxenes, olivines, ilmenite and magnetite. They concluded that the gold was present as uncharged atoms which became incorporated into the silicates and oxides in a random manner. They do note however that the ferrodiorite with a high copper sulphide content has 10 times the average gold content of the rocks. This feature of gold distribution was also noted by Rowe (1969) in the Great Lakes differentiated dolerite in Tasmania (Figure 8). In this study, Rowe (1969) found a systematic increase of gold with differentiation until the mafic index ($FeO + Fe_2O_3/FeO + Fe_2O_3 + MgO \times 100$) equalled 70. At this central part of the sill Greenland and Lovering (1966) found that the copper content reached a maximum.

Tilling <u>et al.</u>, 1973, Econ. Geol., Vol. 68

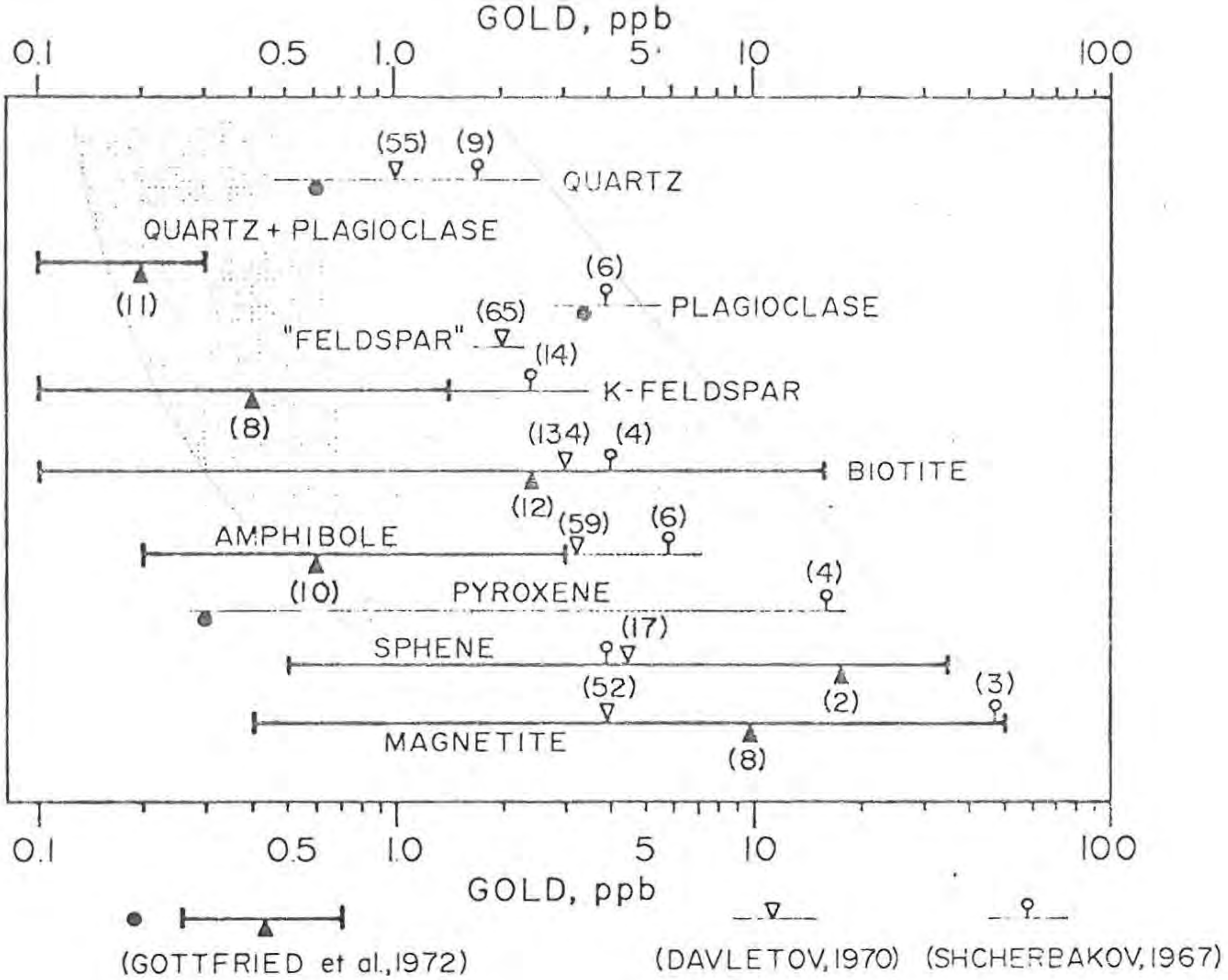


Fig. 7 Gold contents in rock-forming igneous minerals. Ranges, means, and number of samples are denoted as in Figures 4 and 5; solid circles indicate single samples. The shaded zone encompasses various estimates of the means for each mineral and demonstrates the tendency for the mafic minerals to be higher in gold than the salic minerals. Note the log scale for gold abundance.

Rowe, 1969, Chem. Geol., Vol. 4

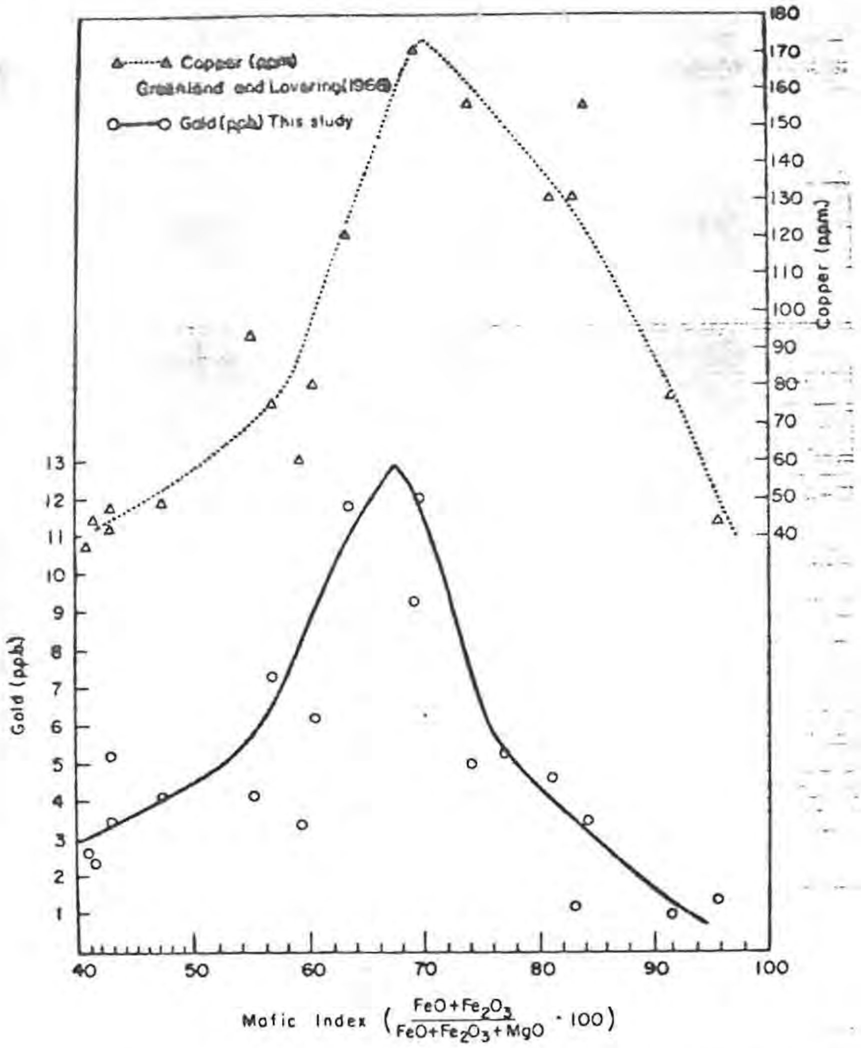


Fig. 8 Variation of gold and copper content with mafic index for drill core D.D.H. 5123, Great Lake sheet, Tasmania.

This systematic increase in gold content suggests that during crystallization the distribution coefficient for gold between crystals and liquid was less than one. This implication is only true if gold does not readily enter into the structures of plagioclase, pyroxene, olivine and iron oxides (Rowe, 1969).

The regular variation of gold suggests that it was in true solution rather than as a colloid trapped on crystal grains (Rowe, 1969). When copper sulphide separated out as an immiscible liquid, the chalcophilic tendency of gold became apparent and the gold entered the sulphide droplets.

In the Merensky Reef, Crockett et al. (1976) found that although gold had a weak and variable association with the sulphides, the sulphide fraction was a more efficient concentrator of gold than the spinel fraction. Spinel however made up 3,5% of the total rock by weight, relative to the 0,9% of the sulphide, thus the spinel was the greater contributor of gold to the total rock content, by mineral association.

At the Strathcona Mine, Sudbury, Ontario, Chyi and Crockett (1976) found that in the ore zone the gold was strongly fractionated towards the sulphides. They also found that gold was accepted by the minerals in the order pentlandite > chalcopyrite > pyrite ≫ magnetite.

Tilling et al. (1973) investigating the gold content of the mineral fractions observed that when comparing gold in minerals and rocks of the Maryville stock, there was often an unsatisfactory material balance. That is, the total gold content for the whole rock did not match the gold content of the individual mineral fractions. Similarly Crockett et al. (1976) observed that in the Merensky Reef only 5% of the total gold content was present in the spinels and sulphides. This indicates that gold is also to be found as discrete particles as the native metal or as an alloy.

It is also noteworthy that in the Merensky Reef Crockett et al. (1976) observed that the highest gold values were intimately associated with a zone of hydrously altered silicate minerals with sulphides replacing them. From the structure of the investigated area it was apparent that there was no 'pot-hole' in the immediate vicinity. It was thus concluded that the alteration was brought about by a primary magmatic fluid rich in water, that was also rich in sulphur and gold.

Keays and Scott (1976) suggest that there are two mineral phases that can host gold : (1) the early crystallizing silicate and oxide fractions; (2) the later crystallizing sulphide fraction. They propose that during solidification gold not taken into the early formed crystal lattices will be fractionated into a late stage fluid phase.

This is in contrast to the conclusion of other workers (Tilling et al., 1973; Gottfried et al., 1972) although they do admit that in the Butte Quartz-Monzonite the samples with gold contents higher than 5,0 ppb are found only in strongly hydrothermally altered rock.

It appears therefore that there is a concentrating process, albeit a passive one, allowing gold to be concentrated in later phases of a cooling magma. If the magma contains more gold than can be taken into the structures of the early crystallizing phases (e.g. olivines, pyroxenes and oxides) it may be taken in by sulphides when they form immiscible droplets.

It appears also that the higher concentrations of gold are likely to be associated with mafic rocks. Late stage fractionation products, particularly those with proportionately larger sulphide contents, derived from mafic rocks are therefore more likely to contain higher gold concentrations.

TRANSPORT AND DEPOSITION OF GOLD IN HYDROTHERMAL SYSTEMS

The Composition of Hydrothermal Fluids

From the studies of known hydrothermal deposits of gold and existing geothermal systems, chemical and geological controls of transport and deposition can be deduced. Barton (1959) has listed three source: from where information regarding the chemical properties of hydrothermal systems may be obtained :

1. Thermal springs and fumeroles,
2. Fluid inclusions,
3. Phase relations of minerals in the ore and the wall-rock.

Details of the physico-chemical properties of hydrothermal gold deposits and geothermal systems have been summarised by Foster (1977) (see Table 4).

The source of the hydrothermal fluids is still being debated. Samples of fluid obtained from fluid inclusions are now the only samples available from existing deposits. The fluids so far studied have all been aqueous with varying amounts of gaseous and dissolved components. Isotope studies on thermal springs and fumeroles indicate that a relatively small proportion of the water is magmatic. It has usually been diluted by meteoric water to the extent that only 5 - 10% of the water is magmatic (Barton, 1959). Brookes et al. (1969) suggest that in the case of the Red Sea Brines, a further possible source of water is that which is released from the transformation of gypsum to anhydrite.

White and Waring (1963) have studied volcanic emanations and have listed the major gaseous constituents. Steam is the major vapour that has been detected. In areas of porous rocks gaseous mixtures resembling air have been noted, but not so from areas where the rocks are still molten. CO_2 was found to be the dominant active gas. This is confirmed by the findings of Burnham (1967) who showed that the primary hydrothermal fluids, particularly those from more felsic magmas, are CO_2 rich. HCl was found to be absent in a large proportion of high temperature fumeroles. HF was found to be an active constituent in the temperature range $100^\circ - 300^\circ C$. H_2S was found in greater quantities in the higher temperature ranges, but never exceded 20% of the total gases. SO_2 was often found to be the dominant oxidized sulphur gas at high temperatures. SO_3 was often more abundant than HCl. It was usually detected at H_2SO_4 rather than SO_3.

TABLE 4 Physico-chemical parameters of the transport and deposition of gold by hydrothermal fluids. (Modified after Foster, 1977, Inst. Min. Res., Univ. Rhod., Report No. 25).

locality	total solutes wt %	dominant cations	dominant anions	pH	depositional temperature °C	pressure bars	mineralisation type	Reference
Ore deposits								
Tenmile, Nevada	2,1-1,0	Na>K>Ca		3,4	340-285	≃500	Au+qtz+adularia vein	Nash, 1972, U.S. Geol. Surv Prof. Pap. 800-C
Sunnyside, Colorado	3,6-0	Na>K>Ca			320-260	150	Au+tellurides vein	Casadevall and Ohmoto, 1976, 25th Int. Geol. Congr Abstr., Vol. 3
Idarado, Colorado	2,5-0,1				306-268		Au+cpy+py vein	Nash, 1975, Econ. Geol., Vol. 70
Finlandia, Peru					275-255	'low'	Electrum+sulpho-salts vein	Kamilli, 1974, Econ. Geol., Vol. 69
Jamestown, Colorado	3-5				270-205	500-125	Au+py, Au+tellurides vein	Nash and Cunningham, 1973, Econ. Geol., Vol. 68
Creede, Colorado	4-12	Na>K	Cl	5,4	268-190	40-50	Ag-Pb-Zn-Cu-Au veins	Barton *et al.*, 1977, Econ. Geol., Vol. 72.
Cortez & Carlin, Nevada	7,3-5,4				265-160		disseminated Au in cbt rocks	Nash, 1972, U.S. Geol. Surv Prof. Pap. 800-C
Yatani, Japan	1,2-0,6			≥6	250-200		Electrum+argentite +sl, gn, cpy	Hattori, 1975, Econ. Geol., Vol. 70
Geothermal Systems								
Salton Sea	33,2	Na>Ca>K			370-270		3,1 ppm Au in bore-hole ppt	White *et al.*, 1963, Sci., Vol. 139
Broadlands drillhole 2	0,49	Na≫K≫(Ca+Mg)	$Cl>\langle CO_3>HBO_2>H_2S$	6,4	276		55 ppm Au in dis-charge ppt	Weissberg, 1969, Econ. Geol. Vol. 64
Rotokawa drillhole 2	0,88	Na≫K>(Ca+Mg)	$Cl>\langle CO_3>H_2S>HBO_2$	≃5	215		70 ppm Au in dis-charge ppt	" " " " "
Steamboat Springs	≃0,2	Na≫K≫Ca	$Cl>CO_3>SO_4$		146-100		10 ppm Au in silica sinter	White, 1967, Ch 13, in Barnes (ed.) Geochemistry of Hydrothermal Ore Deposits
Ohaki Pool	0,31	Na≫K>(Ca+Mg)	$Cl>\langle CO_3>HBO_2$	7,0	95		85 ppm Au in silica sinter	Weissberg, 1969, Econ. Geol. Vol. 64
Champagne Pool, Waiotapu	0,44	Na≫K>Ca	$Cl>\langle CO_3>SO_4$	5,7	75		80 ppm Au in silica sinter	" " " "

B was not readily detectable as a gaseous phase, but it was known to be present in certain systems because of the presence of borates in the condensates. The Br content was found to be about the same as, or slightly less than, that of normal sea water (0,66 ppm; Tooms, 1970). Organic gases such as CO, CH_4, NH_3 and H_2 were often absent. The absence of chlorine as a gas was ascribed to the fact that it had complexed with available cations. Chloride is strongly concentrated in the aqueous phases of magmas (Kilinc and Burnham, 1972).

These aqueous phases are also CO_2 rich, and from the studies of White and Waring (1963) it appears that this CO_2 content remains constant, although Barton (1959) suggests that fluid inclusion studies have shown the later fluids to be CO_2 poor.

The composition of the solutes in hydrothermal fluids has been found to be relatively uniform (Foster, 1977). Barton (1959) has suggested from comparisons of data that a hypothetical ore forming fluid would contain 5 to 30% chlorides of Na and K, 20 ppm H_2S and 200 ppm SO_4. Examination of Table 4 shows very few gold bearing systems have concentrations greater than 5 wt%. Concentrations can be as high as 68 wt% (Roedder, 1971) but it appears that the fluids with higher salinities are those associated more with base metal deposits (Nash 1972).

Casadevall and Ohmoto (1977) however found that at the Sunnyside Mine, Colorado, the higher salinities were associated with both base metal sulphide and gold mineralization. The salinities were low (6,0 - 3,6 equivalent Wt% NaCl). Perhaps it is not the relative salinity that is pertinent, but the empirical salinity. Gallagher (1940) has correlated the presence of base metal sulphides with the albite content of the source magma. He concluded that those gold deposits associated with silver and large amounts of sulphides are related to K rich rocks, whereas the gold deposits with little silver and few sulphides are related to albite rich rocks. He further concludes that the K rich igneous rocks and the Na rich igneous rocks with their separate ore deposit types were derived from distinct magma types.

The temperatures of the fluids at the time of deposition vary, but seldom were they above $300^{\circ}C$ with the majority in the range $200^{\circ}-275^{\circ}C$. Geothermal systems show a much greater range of temperatures, but the greater concentrations of gold in the sinters are from those with temperatures below $100^{\circ}C$.

The pH of hydrothermal fluids is generally slightly acid. The quantification of the degree of alkalinity or acidity of natural hydrothermal fluids is problematical. The pH of a solution decreases with an increase in temperature (Figure 9). At the same time if a neutral solution is subjected to pressure, the pH value rises above 7.

Barton, 1959, in Abelson, ed., Researches in Geochemistry.

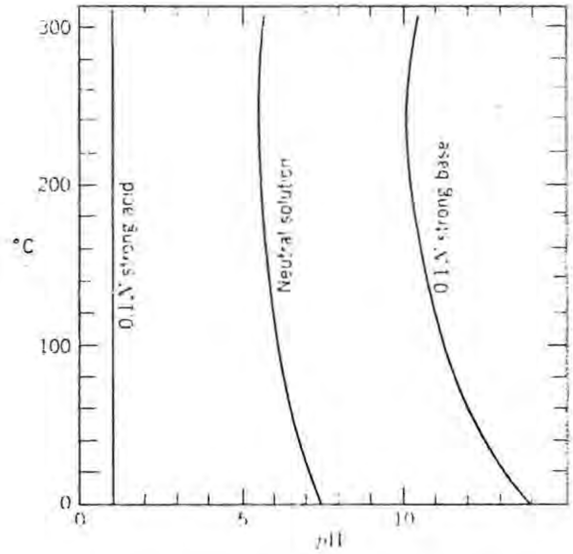


Fig. 9 The effect of temperature of pH in some unbuffered solutions.

To summarize, hydrothermal fluids associated with gold deposits are fairly uniform in their composition, temperature, and pH. They are aqueous phases with their main solutes being chlorides of Na, K and Ca. Their concentration is usually rather low, being in the region of 0,1 to 7,0 weight % NaCl equivalent. The temperatures of hydrothermal systems are thought to be approximately $150^{\circ}-275^{\circ}$C. Geothermal systems generally have lower temperatures (about 90°C). The pH's of both types of system are usually near neutral or slightly acid. It should be emphasized that this information relates to the state of the fluid at the time the minerals were precipitating. The earlier physico-chemical state of the fluid is open to conjecture.

Transport and Deposition of Gold

Transportation of material implies not only a transport medium and energy, but also a source and a depository (Fyfe and Henley, 1973). A large proportion of the world's hydrothermal gold deposits are associated with mafic and ultramafic rocks of Archaean Greenstone belts, which closely resemble oceanic basalts in composition. Keays and Scott (1976) suggest that gold not incorporated into silicate and oxide crystal lattices is concentrated in the residual aqueous phases. Evidence of this has been reported in the Merensky Reef of South Africa (Crockett et al., 1976). Keays and Scott (1976) suggest that the gold not taken up by early crystallizing phases is released into oceanic sediments after extrusion of basalts. They propose that when the lava extrudes the outer portion solidifies quickly and contracts. The contractions cause cracks and fractures which provide access for the heated sea water to enter the slower solidifying centres of the pillows and flows. The chilled margin protects those minerals on the outside from the effects of the sea water. The interior part however is not so protected. Sea water invades and alters the rock releasing that gold that has not been incorporated into the early crystallizing phases. These solutions enrich the surrounding pelagic sediments in gold.

Fyfe and Henley (1973) suggest that it is this pelagic material enriched in gold, that is the source of gold, for hydrothermal deposits. The gold in the sediments is readily accessible, but some of the gold in the silicates may be released during metamorphism. These sediments also contain chlorides which during metamorphism may become the solute in the hydrothermal liquid.

At the sites of the oceanic basalt extrusion, as at mid-ocean ridges, there is a high heat flow. It is possible that this energy drives a convection cell which supplies the fluid transport medium as interpreted in the Red Sea (Brookes et al., 1969; Craig, 1969).

At temperatures associated with hydrothermal systems (i.e. 100°-$400^{\circ}C$) the solubility and transport of gold is still a controversy. Ogryzlo (1935) demonstrated that gold is soluble in acid chloride and alkaline sulphide solutions up to $300^{\circ}C$. Frondel (1938) experimented with colloidal gold. He showed that up to $150^{\circ}C$ the stability of gold sols increased with increased temperature and with decreased concentration of the solution. The presence of colloidal silica was shown to prevent the coagulation of the gold sols at temperatures higher than $150^{\circ}C$. In the experiments gold sols were stable up to $410^{\circ}C$ when no electrolyte was present. Frondel demonstrated that in acid solutions gold could be transported as a true solution. He also showed that gold could be transported as particles adsorbed on to other dispersed colloidal particles. With a decrease in acidity the gold was found to become a sol when protected by colloidal silica. However it is unlikely that much gold is transported as a colloid in hydrothermal systems because of the presence of electrolytes in natural systems.

Barnes and Czamanske (1967) and Krauskopf (1951a) suggest that gold should be soluble as chloride complexes in the presence of strong oxidizing agents (e.g. MnO_2), up to a few hundred degrees centigrade in acid solutions. Helgeson and Garrels (1968) suggest that because calculations show that enough gold can be carried as gold chloride complexes in acid solutions above 175°C, this is probably the main means of transport at these temperatures. They further suggest that the richer deposits formed above 175°C were all probably formed as a result of gold transport as chloride complexes. However, it appears that this model is based only on transport by chloride complexes.

A large proportion of hot springs depositing gold and other elements (including As, Ag, Sb), are near neutral or alkaline (Boyle, 1954; Weissberg, 1969). This suggests that in these neutral to alkaline environments gold is more likely to be transported as complexes with these elements.

At temperatures and pressures greater than those experienced in most epithermal deposits, gold can be dissolved in large quantities under certain conditions (Henley, 1973). In KCl solutions at or near supercritical temperatures where the ratio of HCl to KCl is buffered by a granitic assemblage (quartz-muscovite-potash feldspar), and the oxygen fugacity is maintained by the assemblage magnetite-hematite, the solubility rises almost exponentially (Figure 10). Similarly, a rapid increase of the molality of HCl relative to KCl in a solution buffered by the same assemblage is seen (Figure 11). The gold is thought by Henley (1973) to complex :

$$\mathrm{Au} + {}_a\mathrm{HCl} + {}_b\mathrm{O}_2 \leftrightarrows \left\{\text{Au-Cl complex}\right\} + {}_c\mathrm{H_2O}$$

Above 500°C hydrothermal solutions behave as ideal mixtures, allowing equilibrium solubilities to be calculated for gold on the basis of the known gaseous gold chlorides :

$$\mathrm{Au\,Cl},\ \mathrm{Au\,Cl_2},\ \mathrm{Au\,Cl_3},\ \mathrm{Au_2\,Cl_6}.$$

This last molucule is the most stable gaseous form at temperatures exceeding 500°C (Henley, 1973). At these temperatures other components have little or no effect on the solubility of gold unless they involve solvation by chloride ligands, as do most of the ore metals.

Henley, 1973, Chem. Geol., Vol. 11

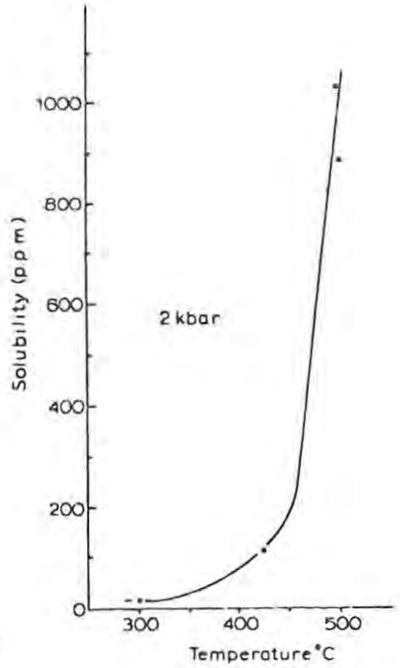


Fig. 10 Some experimental data for gold solubility in the range 300-500°C at 2000 bar total pressure in 2 M (at N.T.P.) potassium chloride solutions, buffered by the assemblages quartz-muscovite-K-feldspar and hematite-magnetite.

Modified after Henley, 1973, Chem. Geol., Vol. 11

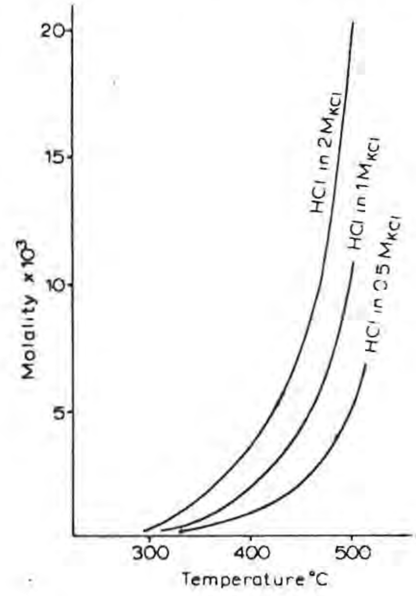
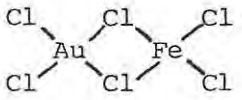
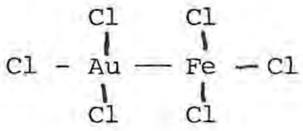


Fig. 11 Variation of the molality of H^+ and HCl in potassium chloride solutions buffered by the assemblage quartz-muscovite-K-feldspar.

The probable transport mechanism involves the formulation of a metal - gold molecule (Foster, 1977) such as :



or



$$\mathrm{Cl - \overset{\overset{Cl}{|}}{\underset{\underset{Cl}{|}}{Au}} - \overset{\overset{Cl}{|}}{\underset{\underset{Cl}{|}}{Fe}} - Cl}$$

At lower temperatures (<400°C) gold is soluble in aqueous sulphide solutions. This is dependant on the formation of stable sulphide complexes (Barnes and Czamanske, 1967). At neutral or near neutral pH's where the dominant sulphide species is HS^-, gold is most soluble (Barnes and Czamanske, 1967; Weissberg, 1969). Krauskopf (1951a) suggests that the gold-sulphide complex is formed :

$$\mathrm{Au + HS^- \longrightarrow AuS^- + \tfrac{1}{2}\, H_2O}$$

In more acid solutions and alkaline solutions where the dominant sulphide species are H_2S and S^{--} respectively gold is less soluble (Weissberg, 1969).

Because the above reaction involves the oxidation of gold to the aurous state, it is consequently dependent on the reduction of ionic hydrogen to hydrogen. Therefore the solubility of gold in a natural hydrothermal system could be dependent on the hydrogen fugacity. The fH_2 is controlled mainly by the equilbria between the minerals pyrrhotite, pyrite, magnetite and hematite.

Experimenting with thiocomplexes of gold Seward (1973) has related the solubility of gold to the concentration NaHS, pH, and temperature :

1. solubility increases with temperature increases;
2. solubility increases with a decrease in pH to neutrality;
3. solubility increases with increase in HS^- content at a constant pH.

The effect of pressure on the solubility of gold-thiocomplexes is rather complicated. Seward (1973) showed that the effect of pressure on these complexes depended on the temperature and the pH of the system. In alkaline solutions a decrease in pressure increased the solubility, especially at temperatures exceeding 250°C. When the temperature was below 200°C the effect of pressure was minor. In solutions at or near neutrality a decrease in pressure from 1500 to 500 bars increased the solubility when the temperature exceeded 250°C. At temperatures below 250°C a pressure decrease caused a decrease in solubility.

H_2S, being an uncharged molecule, has weak complexing properties. In the pH ranges investigated, Seward (1973) found the concentration of S^{--} to be very low. He therefore concluded that three thio-complex species existed :

$Au_2(HS)_2S^{--}$	in alkaline solutions;
$Au(HS)_2^-$	in neutral solutions; and
AuHs	in acid solutions.

By analogy the transport of gold and other metals may be possible by means of complexes of the type $Au(AsS_2)$, $Au(AsS_3)$, $Au(Sb_2S_4)^-$ (Foster, 1977). Tellurium may also be involved in the transport of gold in hydrothermal solutions. The lack of information on gold tellurides in banded iron-formation deposits suggest that they have not been found in these deposits. This could be because they are unstable at the lower temperatures associated with the deposition of iron-formations (Stillwell, 1953). It appears therefore that HS^- is the most versatile sulphur species, forming sulphide complexes (Krauskopf, 1951) and thio-complexes (Seward, 1973).

Boyle (1969) says that no single complex is responsible for transporting gold. From observations he has deduced that in different settings different complexes are responsible. Arsenic and antimony appear to have been responsible in several instances. Sulphur and tellurium complexes are apparently responsible in yet others.

The precipitation of gold from hydrothermal solutions is also a controversy. Each worker using one type of complex has concluded that certain conditions are necessary for precipitation. Henley (1973) has shown that at low temperatures gold-chloride complexes are precipitated by a lowering of temperature (Figure 12, area A). In area B a decrease in temperature changes gold from the auric to the aurous form while an increase in temperature changes gold from the aurous to the auric form. In area C the solution resembles a fluid in a granitic environment. Here the gold is precipitated from a saturated solution by a fall in temperature.

Henley, 1973, Chem. Geol., Vol. 11.

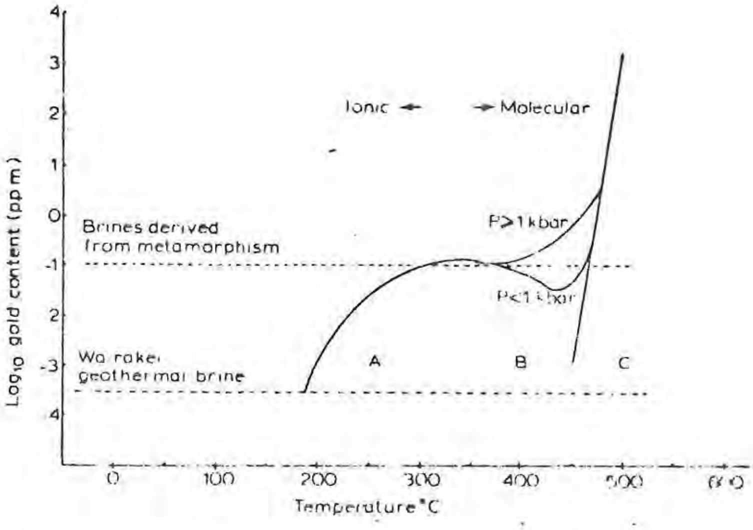


Fig. 12 General solubility curve for gold in 3 M potassium chloride solution. The curve below 300°C is that of Helgeson and Garrels (1968) for solutions in contact with quartz and pyrite. The curve for $T > 300^{\circ}C$, P = 1 kbar is calculated from solubility data for solutions where m_{HCl} is buffered by the quartz-muscovite-K-feldspar assemblage. Oxygen fugacity is buffered at values intermediate between those on the quartz-fayalite-magnetite and hematite-magnetite buffers.

Where gold is complexed with sulphur other major depositional mechanisms are involved. Changes in pH are important (Seward, 1973; Weissberg, 1969; Ewers and Keays, 1977). These changes can be brought about in several ways. Oxidation of H_2S to H_2SO_4 lowers the pH. This causes a decrease in the reduced sulphur activity which in turn causes precipitation. The precipitation of sulphides (perhaps as a result of temperature decrease) also lowers the activity of reduced sulphur (Seward, 1973).

Boyle (1969) says that changes in temperature do not appear in nature to be major precipitation initiators. The fact that so many gold veins occur in dilatant zones in rocks covering great vertical ranges suggests that pressure changes, particularly decreases, are more important. Ewers and Keays (1977) show that H_2S is lost very rapidly as a result of 'boiling' when the pressure is released. This loss of H_2S reduced the activity of reduced sulphur with a concomittant precipitation of gold. However, Weissberg (1969) suggests that when CO_2 is lost by 'boiling off' the activity of S^{--} and HS^{-} increases, thus increasing the solubility of As, Sb, Hg, and Au. However this 'boiling' action would certainly cause the precipitation of the other solutes, including the chlorides (Ridge, 1973).

Thus the pressure changes must surely be controlled if they are that important as causes of precipitation.

Ridge (1973) working from the calculations of Haas (1971) has shown that massive sulphide deposits could not have formed from gaseous volcanic emanations. He states that with the release of pressure at the surface, the resultant 'boiling' action would cause the precipitation of the chlorides. However, no massive halite deposit associated with massive sulphide deposits are known. Based on this concept he has shown (Ridge, 1973), that hydrothermal solutions at elevated temperatures containing dissolved halides need hydrostatic pressures to prevent the sudden precipitation of halides at the spring or fumerole vent.

Thus, given the temperature of the ore forming liquid and its halide content, one can calculate a minimum depth of deposition (Table 5 gives the vapour pressure of fluids of selected concentrations of Na Cl at given temperatures and the minimum depth of water to prevent their boiling).

At atmospheric temperatures and pressures gold is soluble as auric chloride (Cloke and Kelly, 1964) but the conditions needed for this appear to be extreme ($pH<1$, Eh 0,1-0,5υ, Goleva et al., 1970). In sea water gold is soluble also as AuO_2^- (Krauskopf, 1951a). The limiting factor appears to be the amount of oxygen dissolved in the sea. Goleva et al. (1970) say that the presence of pyrite and other sulphides in a solution will precipitate colloidal gold. They also say that calcium carbonate is a common co-precipitant of gold. These facts may explain why gold in iron-formations is generally only associated with sulphide and mixed sulphide-carbonate facies zones (Fripp, 1974).

Foster (1977) has diagrammatically represented a possible sequence of events through which gold will pass in a hydrothermal system (Figure 13). At supercritical temperatures the gold is transported as volatile chloride complexes. As the temperature decreases the complexes change and the gold combines with other metals and thiosulphides and sulphides, depending on pH and availability of metals and the hydrogen fugacity. As the metals are precipitated as sulphides or sulpho-salts, gold may be retained in the system. This gold may form sols, protected by colloidal silica, to be precipitated on the surface, subaerially, or in sub-marine environments.

Haas, 1971, Econ. Geol., Vol. 66

TABLE 5 Thermal Profiles for Selected Compositions of NaCl-H_2O Liquids with Corresponding Vapour Pressures and Densities

The temperature at the water-table (P_o=1.013 b) is given at the bottom of each profile.

Temp. (°C)	2.0 Wt Per Cent NaCl			5.0 Wt Per Cent NaCl			10.0 Wt Per Cent NaCl		
	Depth (meters)	Pressure (bars)	Density (g/cm³)	Depth (meters)	Pressure (bars)	Density (g/cm³)	Depth (meters)	Pressure (bars)	Density (g/cm³)
T_b	0.0	1.0	0.972	0.0	1.0	0.992	0.0	1.0	1.027
90.	-3.3	0.7	0.979	-3.4	0.7	1.000	-3.5	0.7	1.035
100.	-0.1	1.0	0.972	-0.3	1.0	0.993	-0.7	0.9	1.029
110.	4.2	1.4	0.965	3.9	1.4	0.986	3.2	1.3	1.022
120.	10.0	2.0	0.957	9.4	1.9	0.978	8.4	1.9	1.014
130.	17.6	2.7	0.949	16.7	2.6	0.970	15.2	2.5	1.007
140.	27.3	3.6	0.941	26.1	3.5	0.962	23.8	3.4	0.999
150.	39.7	4.7	0.932	37.9	4.6	0.953	34.8	4.4	0.990
160.	55.1	6.1	0.922	52.7	6.0	0.944	48.5	5.8	0.982
170.	74.3	7.8	0.913	71.0	7.7	0.935	65.6	7.4	0.973
180.	97.7	9.9	0.902	93.4	9.7	0.925	86.2	9.4	0.963
190.	126.0	12.4	0.892	120.5	12.2	0.915	111.3	11.7	0.954
200.	160.1	15.4	0.881	153.1	15.1	0.904	141.4	14.5	0.944
210.	200.8	18.9	0.869	191.9	18.5	0.893	177.2	17.8	0.934
220.	248.9	22.9	0.857	237.8	22.5	0.882	219.4	21.7	0.923
230.	305.5	27.7	0.845	291.7	27.1	0.870	269.0	26.1	0.912
240.	371.7	33.1	0.831	354.7	32.5	0.857	326.7	31.2	0.901
250.	448.6	39.3	0.817	427.8	38.6	0.844	393.6	37.1	0.889
260.	537.9	46.4	0.803	512.3	45.5	0.830	470.7	43.8	0.876
270.	640.5	54.4	0.787	609.6	53.3	0.816	559.3	51.3	0.863
280.	58.7	63.4	0.771	721.2	62.2	0.801	660.6	59.9	0.850
290.	894.3	73.6	0.753	849.0	72.1	0.784	776.0	69.4	0.835
300.	1049.5	84.9	0.734	994.9	83.2	0.767	907.3	80.0	0.820
310.	1227.2	97.5	0.714	1161.3	95.6	0.748	1056.2	91.9	0.804
320.	1430.7	111.5	0.691	1351.1	109.3	0.727	1225.0	105.1	0.787
330.	1664.3	127.1	0.666	1567.8	124.5	0.704	1415.3	119.1	0.768
	T_b = 100.3°C			T_b = 100.9°C			T_b = 101.9°C		

	15.0 Wt Per Cent NaCl			20.0 Wt Per Cent NaCl			25.0 Wt Per Cent NaCl		
	Depth (meters)	Pressure (bars)	Density (g/cm³)	Depth (meters)	Pressure (bars)	Density (g/cm³)	Depth (meters)	Pressure (bars)	Density (g/cm³)
T_b	0.0	1.0	1.063	0.0	1.0	1.101	0.0	1.0	1.139
90.	-3.7	0.6	1.072	-3.9	0.6	1.110	-4.1	0.6	1.150
100.	-1.0	0.9	1.065	-1.4	0.9	1.104	-1.9	0.8	1.144
110.	2.6	1.3	1.059	1.9	1.2	1.097	1.1	1.1	1.137
120.	7.3	1.8	1.051	6.2	1.7	1.090	5.0	1.6	1.129
130.	13.6	2.4	1.044	11.9	2.3	1.082	10.2	2.1	1.122
140.	21.6	3.2	1.036	19.2	3.1	1.075	16.8	2.9	1.114
150.	31.7	4.3	1.028	28.5	4.0	1.067	25.2	3.8	1.106
160.	44.3	5.5	1.020	40.0	5.2	1.058	35.7	4.9	1.098
170.	59.9	7.1	1.011	54.3	6.7	1.050	48.6	6.3	1.089
180.	79.0	9.0	1.002	71.7	8.5	1.041	64.4	8.0	1.081
190.	102.0	11.2	0.993	92.7	10.6	1.033	83.5	10.0	1.072
200.	129.6	13.9	0.984	117.9	13.2	1.024	106.4	12.4	1.063
210.	162.4	17.0	0.974	147.8	16.2	1.015	133.6	15.2	1.055
220.	201.1	20.7	0.964	183.1	19.7	1.005	165.7	18.5	1.046
230.	246.4	25.0	0.954	224.3	23.7	0.996	203.1	22.3	1.037
240.	299.1	29.9	0.944	272.2	28.4	0.986	246.6	26.7	1.028
250.	360.1	35.5	0.933	327.6	33.7	0.977	296.8	31.8	1.019
260.	430.2	41.9	0.922	391.1	39.7	0.967	354.3	37.5	1.010
270	510.5	49.1	0.910	463.8	46.6	0.957	420.0	44.0	1.002
280	602.1	57.2	0.899	546.4	54.3	0.947	494.7	51.3	0.993
290.	706.1	66.3	0.887	639.9	62.9	0.937	579.0	59.5	0.985
300.	823.8	76.5	0.874	745.5	72.6	0.927	673.9	68.6	0.977
310.	956.8	87.8	0.861	864.2	83.3	0.916	780.4	78.7	0.969
320.	1106.7	100.3	0.847	997.3	95.2	0.906	899.3	90.0	0.962
330.	1275.3	114.2	0.832	1146.2	108.4	0.896	1031.7	102.4	0.954
	T_b = 103.2°C			T_b = 104.7°C			T_b = 106.7°C		

In submarine environments, gold is not adsorbed by, and precipitated with, iron and manganese hydroxides. It is probably carried away from the vent area by currents, (Keays and Scott 1976) and precipitates in more distal neutral or alkaline reducing environments.

Modified after Foster, 1977, Inst. Min. Res., Univ. Rhod., Report No. 25

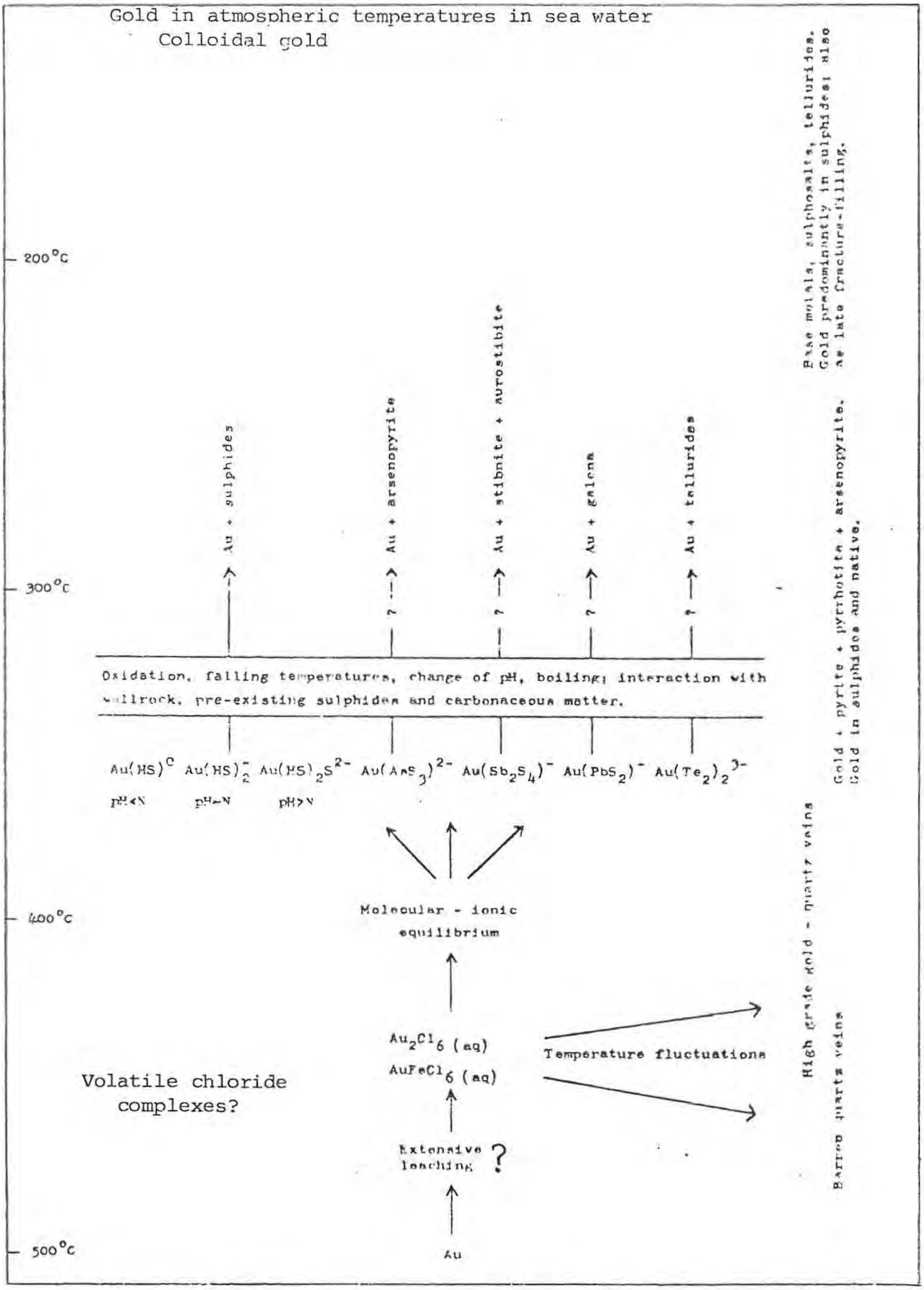


Fig. 13 Aqueous gold - bearing species in a cooling hydrothermal fluid

DISTRIBUTION OF ARCHAEAN IRON-FORMATION

Iron-formations of Archaean age are found mainly in greenstone belts in continental cratonic areas. The greenstone belts occur as 'islands' in 'seas' of granite-gneiss. Remnants of greenstones are also found in some metamorphic mobile belts, e.g. Limpopo mobile belt (Mason, 1970).

Spatial Distribution

Goodwin (1973) has described the distribution of Archaean iron-formations in Canada according to their facies types. Figure 14 shows their distribution relative to Archaean greenstone tectonic 'basins'. The main areas in which they are developed are the Slave, Churchill and Superior Provinces.

In Southern Africa greenstone belts containing iron-formations are found mainly on the Rhodesian and Kaapvaal cratons (Beukes, 1973). Iron-formations are also to be found in the highly metamorphosed Messina formation in the Limpopo Mobile Belt (Figure 15). In this area both massive and banded iron-formations are present. In the Pongola Supergroup, which is considered to be the oldest cratonic basin on the Kaapvaal craton (Beukes, 1973), the iron-formations are almost wholly confined to the upper sedimentary Mozaan Group. Only one iron rich sedimentary horizon has been found in the lower volcanic Insuzi Group (Du Toit, 1931). This is in contrast to the Swaziland Supergroup of South Africa (Viljoen and Viljoen, 1969a) and the Basement Complex of Rhodesia (Bliss and Stidolph, 1969) in which all the major units contain iron-formations.

Other regions in Africa where Archaean greenstone belts are known to occur are Tanzania and Kenya (Shackleton, 1946), and Central Africa (Woodtli, 1961).

In South America (Figure 16) very old iron-formations (greater than 3,0 by) are known in the granulite terrain round Imataca, Venezuela (Dorr II, 1973). In Brazil, iron-formations older than 2,7 by are known in the Minas Gerais area of Quadrilatero Ferrifero. Dorr II (1973) reviewing earlier literature describes them as being algoman-type iron-formations in paragneisses which were derived from sedimentary and volcanic sequences. Archaean iron-formations are also known in Uruguay. Dorr II (1973) describes these as being early to middle Precambrian in age, so they may not be true Archaean iron-formations.

Goodwin, 1973, Econ. Geol., Vol. 68

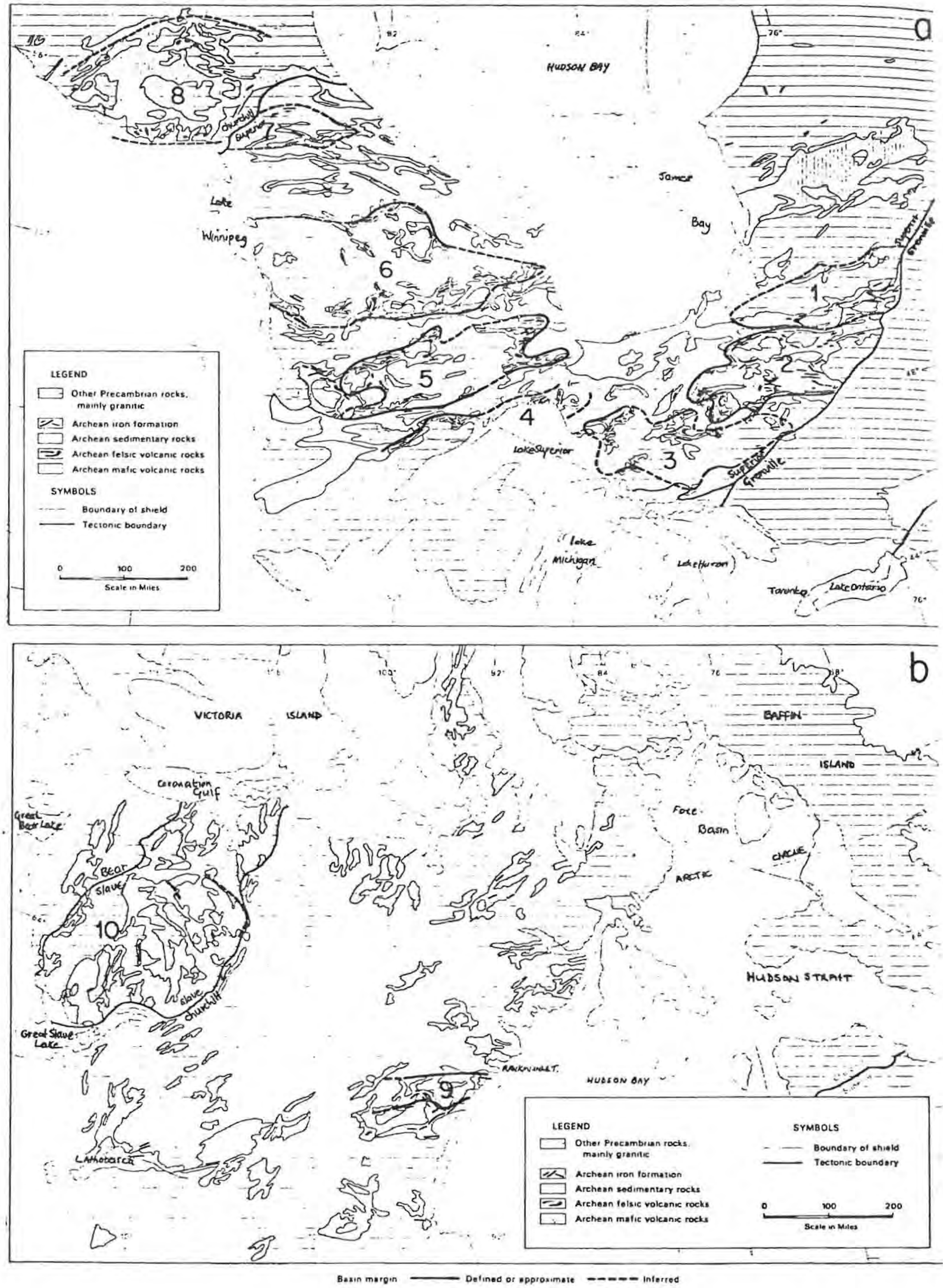


Fig. 14 Distribution of Archaean tectonic basins in Canada : (a) southern shield; (b) northern shield. Basin margins have been defined on the basis of the triple lithofacies association of 1) oxide-sulphide facies transition, 2) arc-type felsic volcanic piles and, 3) proximal conglomerates (not illustrated). 1-Matagami Basin; 2-Abitibi Basin; 3-Algoma Basin; 4-Superior Basin; 5-Keewatin Basin; 6-Berens Basin; 7-God's Lake Basin; 8-Kisseynew Basin; 9-Kaminak Basin; 10-Slave Basin.

Beukes, 1973, Econ. Geol., Vol. 68

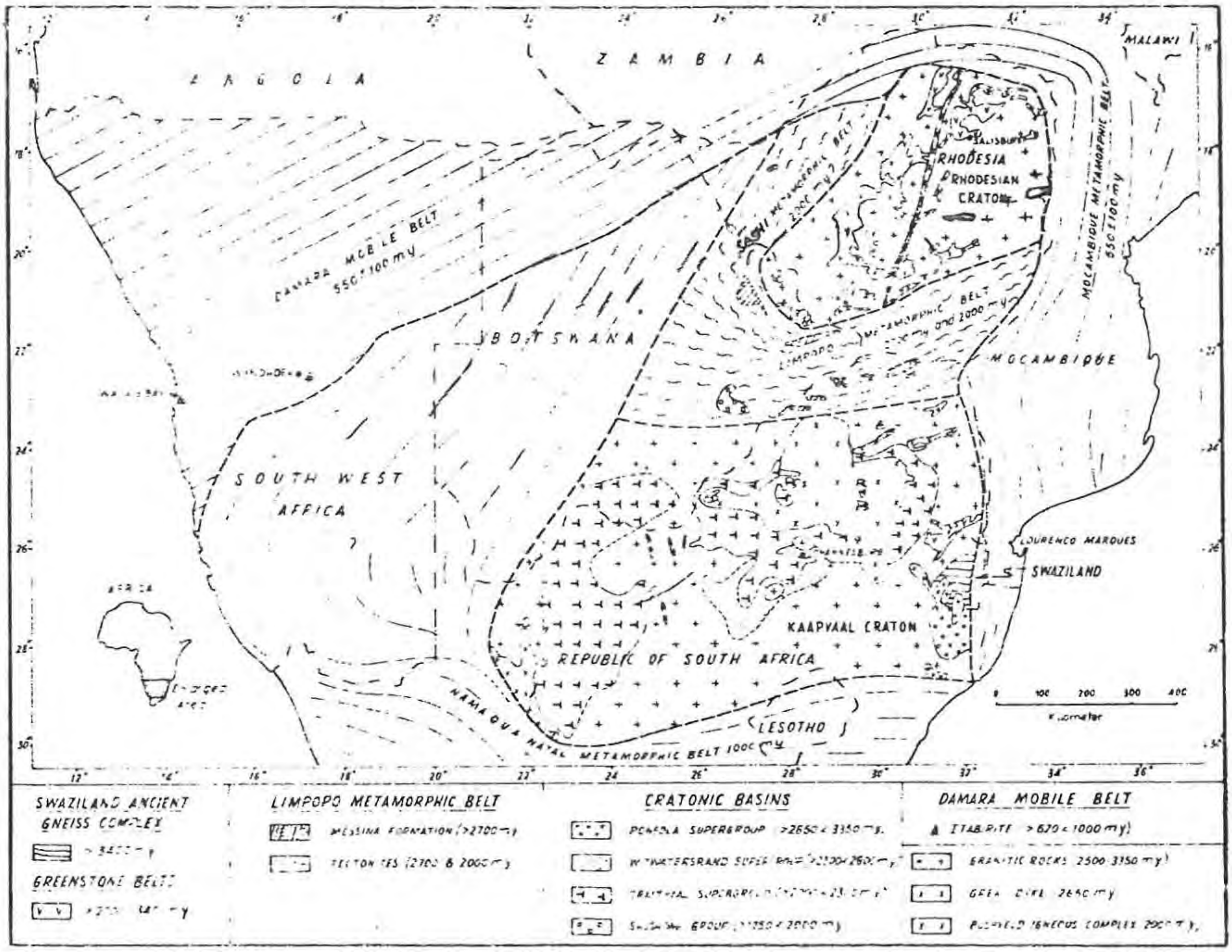


Fig. 15 Geotectonic map of southern Africa showing the distribution and classification of the tectono-sedimentary units in which the iron-formations occur.

Dorr II, 1973, Econ. Geol., Vol., 68

Fig. 16 Map showing the general location of Archaean iron-formations in South America

English literature on Russian iron-formations appears to be lacking. Most iron-formations seem to be early Proterozoic (2,0-2,6 by) in age (Alexandrov, 1973). The Korkian-Belozerka iron-formations of the Greater Krivoy Rog, Ukrainian Shield (Figure 17) are thought to be Archaean. They are usually associated with volcano-sedimentary cycles (Alexandrov, 1973).

Alexandrov, 1973, Econ. Geol., Vol. 68

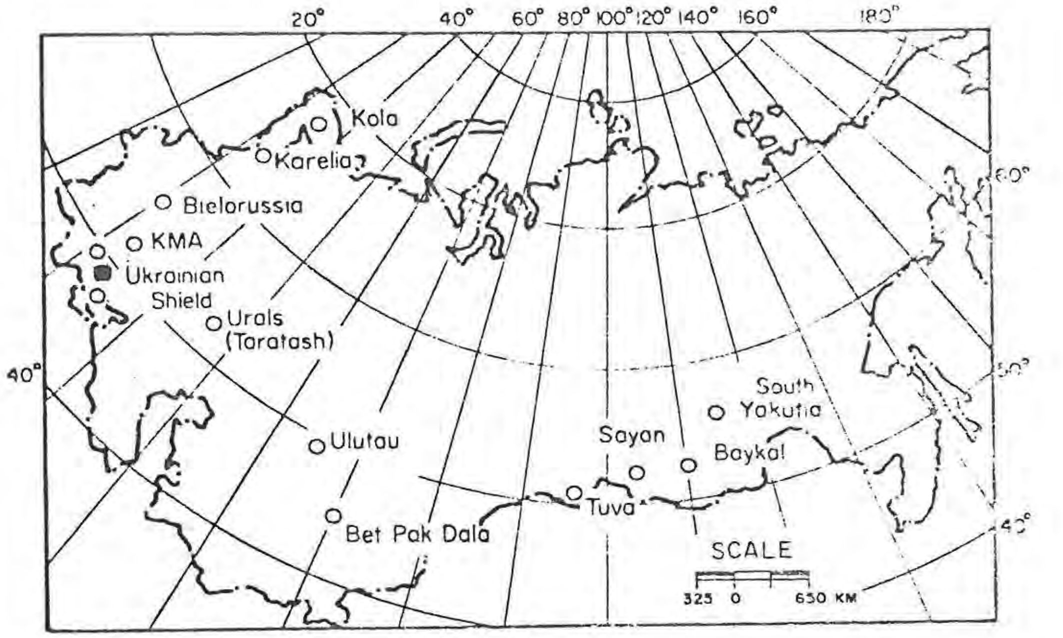


Fig. 17 Deposits of Archaean banded iron ores of the Soviet Union

In Australia the iron-formations of Archaean age are all to be found in the Yilgarn and Pilbara Blocks of Western Australia (Figures 18, 19). Greenstone belts cover about ¼ of the area, the rest being underlain by granite-gneiss.

In India the Archaean Dharwar rocks are found closely associated with Archaean gneisses in many parts of the Peninsula (Wadia, 1926). There are three main areas in the Peninsula where these rocks are exposed :

1. Southern Deccan including the type area of Dharwar and Bellary, and a major portion of Mysore State;
2. the areas of Carnatic, Chata Nogpur, Jabalpur, Nagpur, Behar, Rewah and Hazaribagh;
3. the Arovalli region extending as far north as Delhi and far south as north Gujarat.

Trendall, 1973, Econ. Geol., Vol. 68

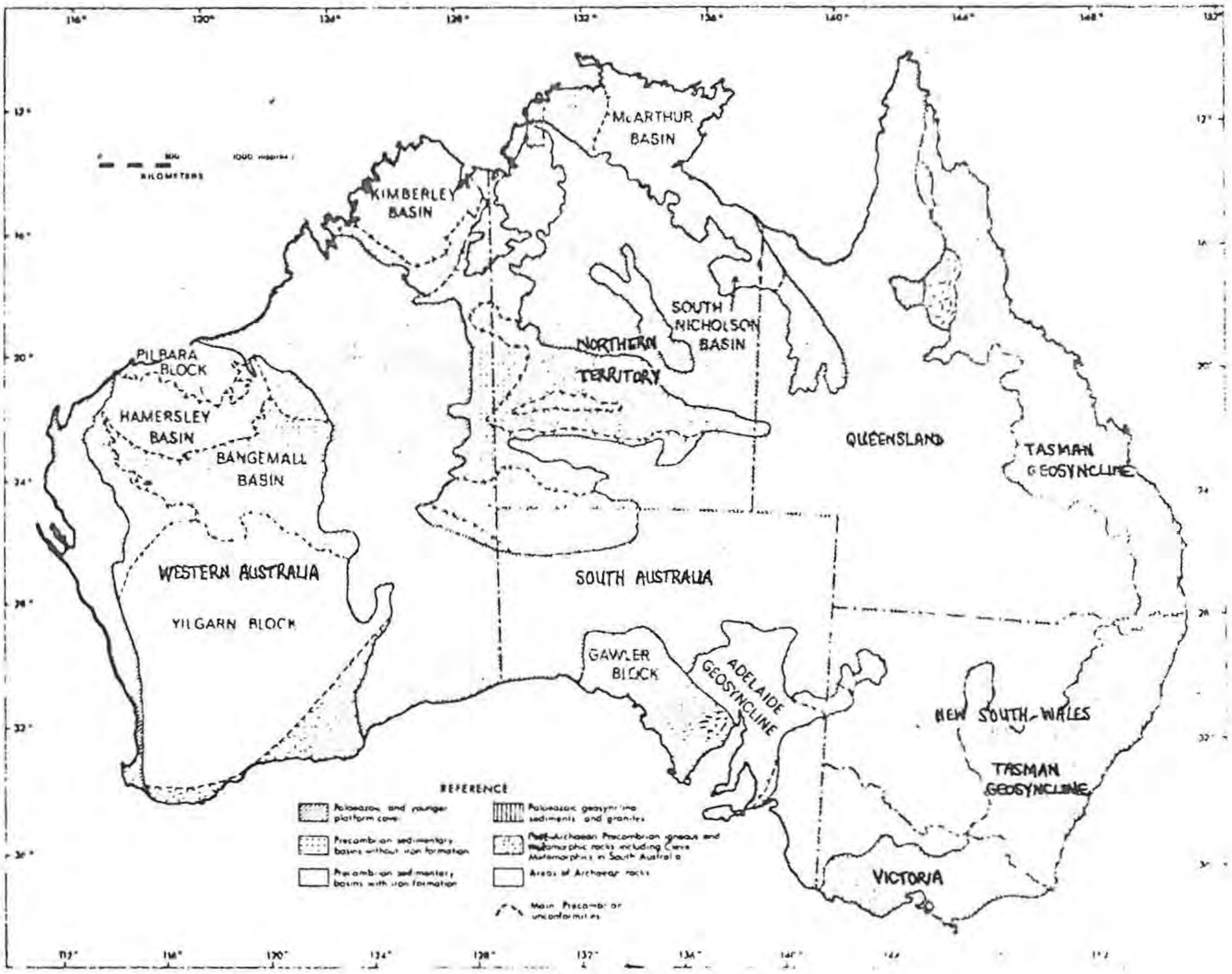
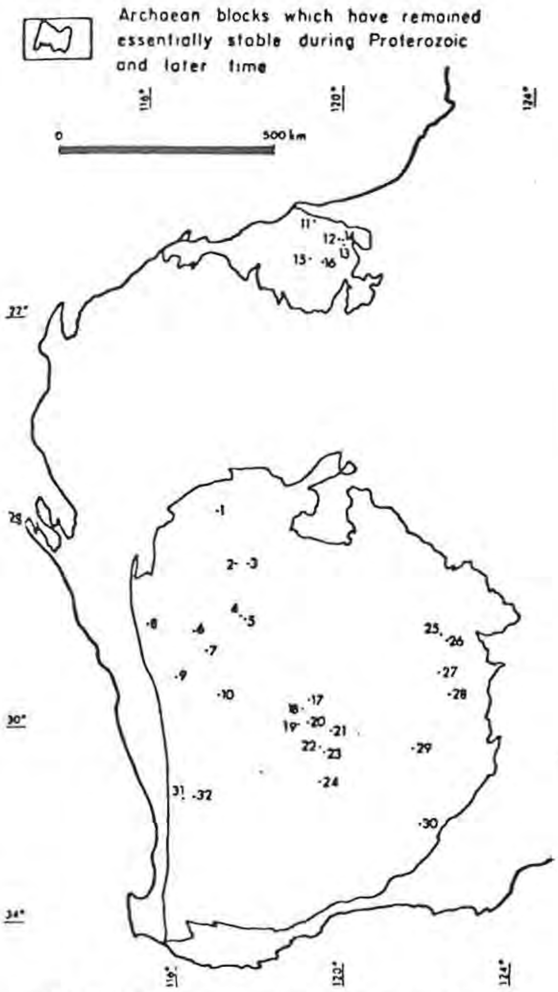


Fig. 18 Geotectonic sketch map of the Australian continent.



1 JACK HILLS, 2 WELD HERCULES GOLD MINE, 3 WELD RANGE, 4 BOOGARDIE, 5 MT MAGNET, 6. YALGOO, 7 GNOWS NEST RANGE, 8. TALLERING RANGE, 9. KOOLANOOKA HILLS, 10 MT GIBSON, 11 MT GOLDSWORTHY, 12 SHAY GAP, 13. KENNEDY GAP, 14. CATTLE GORGE, 15. NORTH POLE, 16 MARBLE BAR, 17 PIGEON ROCKS, 18 NORTH WINDARLING, 19 MT JACKSON, 20 MARDA, 21 BUNGALBIN 22 DOWDS HILL, 23 KOOLYANOBBING RANGE, 24. PARKERS RANGE, 25 MT WINDARRA, 26 LANCEFIELD, 27. MT FLORENCE, 28 EDJUDINA RANGE, 29 LAKE YINDARLGOODA, 30 NORSEMAN, 31 CLARKELINE 32 NORTHAM

Fig. 19 Archaean BIF localities in Western Australia

The Dharwar rocks are also exposed in the Shillong plateau of the Assam ranges. They are also probably well represented in the central and northern zones of the Himalayas.

In the United States iron-formations of three major periods are known :

0,6 - 1,9 by

1,9 - 2,6 by

> 2,6 by

The pre 2,6 by iron-formations are algoman-type associated with volcanic mafic to felsic cycles (Bayley and James, 1973). The main areas of Archaean iron-formation development are (Figure 20) north Minnesota, Michigan, Wyoming and Montanna. In the Black Hills area of South Dakota, the Homestake formation contains ferruginous cherts with which is associated the Homestake gold deposit.

Bayley and James, 1973, Econ. Geol., Vol. 68

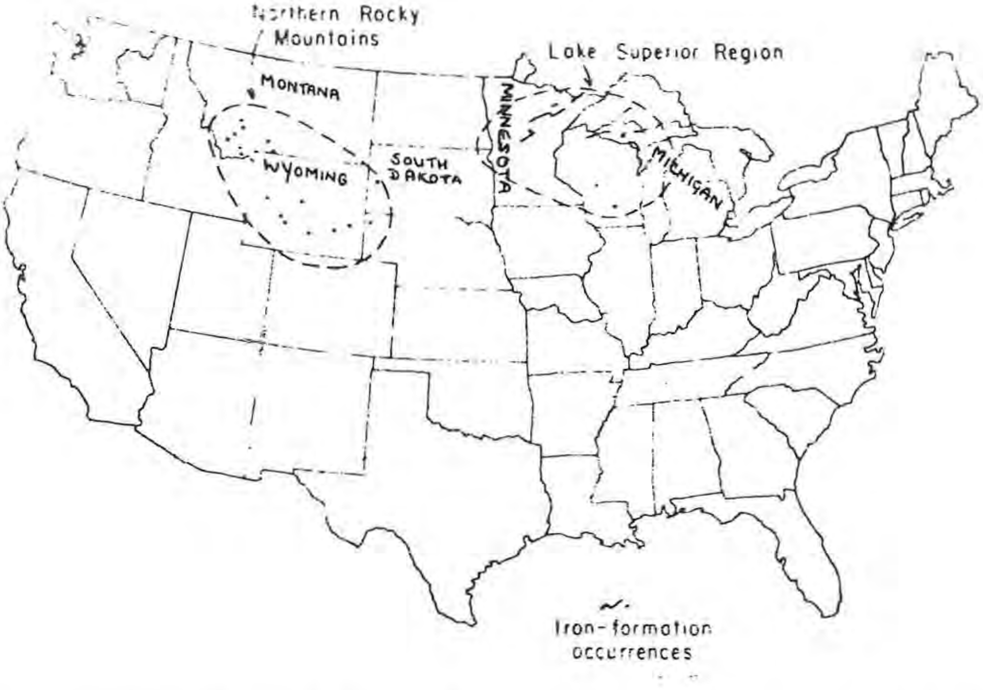


Fig. 20 Distribution of Archaean iron-formation in the United States

Stratigraphic Distribution

A study of relevant literature reveals that throughout the world Archaean sequences have remarkably similar features. Local variations have resulted in a multitude of stratigraphic terms and it is generally impossible to correlate these smaller units over any great distance. Also, in some areas only certain portions of a lithologic sequence may be preserved, so only lithologic correlations can be attempted.

Viljoen and Viljoen (1969a) attempted to correlate greenstone successions around the world with that of the Barberton mountain land. Oversimplification of the stratigraphy has resulted in a rather 'layer cake' model (Williams and Furnell, in prep.). Anhaeusser et al. (1969) have correlated the major lithological units in Archaean greenstone terrains from around the world (see Table 6). The lithological characters by which these units have been correlated are listed in Table 7. By the very nature of the rock types and their age, it is possible that not all these features are to be found in any one sequence. A very noticeable example is the absence of a predominantly ultramafic unit at the base of the sequences except for those in Rhodesia and South Africa.

Even in these two countries this lower unit is not ubiquitous. In the Murchison range, north east Transvaal, the lowermost rocks of the greenstone belt appear to be mafic in composition rather than ultramafic (Van Eeden et al., 1939). Because of the high degree of deformation and alteration, this is not definite. In the Barberton greenstone belt this unit is well developed. Similar lithological units are to be found at the base of the greenstone belts in the Pietersburg and Klein Letaba regions in the northern Transvaal.

In Rhodesia, a lower, predominantly ultramafic unit has been recognised since 1932 when MacGregor first mapped a now, rather controversial unconformity. This unconformity separates a 'Magnesian series' from an overlying mafic greenstone sequence. Later this ultramafic unit was incorporated into MacGregor's (1947, 1951) Sebakwian system. Harrison (1968), who has remapped MacGregor's type area, regards this lower ultramafic sequence as an intrusive into the Bulawayan greenstone succession. What Harrison regards as Sebakwian rocks are metamorphosed greenstone-type relics found in the granite-gneiss as mega-xenoliths, particularly in the Rhodesdale batholith between the Que Que and Selukwe areas. These occurrences are also regarded as Sebakwian rocks by Viljoen and Viljoen (1969a). Stowe (1968, 1969) regards these and lower portions of the Selukwe greenstone belt as pre-Sebakwian.

In Australia, there has been no formal recognition of a predominantly ultramafic zone at the base of the greenstone assemblages (Table 8).

Goodwin (1973) and Goodwin and Ridler (1970) have not described significant zones of ultramafic rocks at the base of the Canadian greenstone belts. Boyle (1961) mentions no major zone of ultramafics in the Yellowknife formation.

Anhaeusser et al., 1969, Geol. Soc. America Bull., Vol. 80

TABLE 6 The suggested Lithostratigraphic Correlation of Greenstone Belts of Shields around the World

GROUP	AFRICAN SHIELD			WESTERN AUSTRALIAN SHIELD		INDIAN SHIELD	CANADIAN SHIELD		
	South Africa	Rhodesia	East Africa	Yilgarn Block	Pilbara Block		Superior Province	Churchill Province	Slave Province
SEDIMENTARY GROUP	Moodies Fig Tree	Shamvian	Kavirondian	Yilgarn "Whitestone"	Mosquito Creek	Upper Dharwar	Timiskaming Knife Lake	Sickle Missi	Yellowknife Group (B)
GREENSTONE GROUP	Onverwacht	Bulawayan	Nyanzian	Kalgoorlie "Greenstones	Warrawoona	Middle and Lower Dharwar	Keewatin-Coutchiching Abitibi-Pontiac?	Wasekwan Amisk	Yellowknife Group (A)
ULTRAMAFIC GROUP	Onverwacht	Sebakwian	?	?	?	?	?	?	?

Modified after Anhaeusser et al., 1969, Geol. Soc. America Bull., Vol. 80

TABLE 7 The Primary Lithologies and Frequency of Occurrence of Rocks developed in Greenstone Belts

Group	Primary Lithological Types	Frequency of Lithological Types
SEDIMENTARY GROUP Predominantly sediments—subordinate volcanics	(a) *Essentially shallow water sediments* Minor development of volcanics and pyroclastics. Banded ironstones, jaspilites, cherts, limestones, sub-graywackes, shales, quartzites (clean and impure quartzites, feldspathic and calcareous quartzites). Boulder beds, conglomerates, grits, arkoses.	All sediments listed are frequently encountered. Chemical precipitates often present but may be poorly developed or absent entirely.
	(b) *Possibly deeper water sediments* Minor development of volcanics-tuffs, agglomerates, lavas.	Usually present
	Banded ironstones, jaspilites, cherts.	Usually present
	Graywackes, shales, conglomerates, grits.	Usually present
GREENSTONE GROUP Predominantly volcanics and pyroclastics-subordinate sediments	Banded ironstones, jaspilites, cherts	Often present
	Carbonaceous shales and cherts, graphitic sediments.	Often present
	Mafic and salic tuffs and agglomerates, ash beds.	Often present
	Siliceous and mafic graywackes, conglomerates, quartzites, limestones, grits (sediments interlayered with volcanics).	Often present Limestones sometimes present
	Minor development of ultramafites	Sometimes present
	Porphyries	Often present
	Rhyolites	Sometimes present
	Dacites	Often present
	Andesites	Always present—abundant
	Basalts (tholeiitic)	Often present
ULTRAMAFIC GROUP Predominantly volcanics and pyroclastics-subordinate sediments	Siliceous, aluminous tuffs, agglomerates. Minor interlayered sediments (cherts, graphitic shales, banded ironstones, fine-grained quartzites).	Often present but may be poorly developed or absent entirely
	Tholeiitic basalts (lava)	Usually present
	Dunites, peridotites (some extrusive) pyroxenites, gabbros, anorthosites	Not often encountered. Absent in many greenstone belts?
	Early crust—in part granitic?	

Gemuts and Theron, 1975, in Knight, C.L., Austr. Inst. Min. Metall. Monogr. No. 5

TABLE 8 Archaean stratigraphy and correlation in the Norseman - Coolgardie region

Gemuts & Theron (1975) Coolgardie-Norseman Area	McCALL (1969) Lake Lefroy-Kambalda Area	GLIKSON (1971, a) Coolgardie-Mungari Area	WILLIAMS (1969) Kurnalpi-1:250 000 Sheet		WOODALL (1965) Kalgoorlie Area	HALL & BEKKER (1965) Norseman Region
Sequence 8 Polymictic conglomerate and pebbly greywacke	KURRAWANG CONGLOMERATE	KURRAWANG BEDS		CYCLE 3		
Sequence 7 Acid tuffaceous rocks and acid volcanic breccia. Some acid extrusive rocks		MUNGARI BEDS BLACK FLAG METASEDIMENTS	*Association V* KALPINI FORMATION Basic to intermediate extrusive and intrusive rocks, ultramafic intrusive rocks, minor clastic rocks and chert			
Sequence 6 High-Mg basalt, ultramafic rocks, minor chert, black slate and tholeiitic basalt	YILMIA OPHIOLITE BELT CAVE ROCKS OPHIOLITE BELT	RED LAKE OPHIOLITES				
Sequence 5 Conglomerate Arkosic greywacke and argillite with minor tholeiitic basalt. Basal conglomerate	LAKE DAM BEDS WANDA WANDA BEDS CAVE ROCK BEDS MANDILLA BEDS TRIANGLE ISLAND BEDS	BROWN LAKE METASEDIMENTS MT. ROBINSON OPHIOLITES	*Association IV* GUNDOCKERTA FORMATION Turbidite sequence, Clastic sequence, conglomerate and sandstone	CYCLE 2		
Sequence 4 Top of sequence gradational to sequence 5 Greywacke, minor chert and black slate. Acid extrusive rocks and feldspar-porphyry intrusives. Minor tholeiitic basalt	MEROUGIL CREEK BEDS CAUSEWAY BEDS	GUNGA META/ARGILLITES	Acid volcanic complexes		BLACK FLAG BEDS Tuff, agglomerate, acid to intermediate lavas, slate, shale, greywacke, quartzite	MT. THIRSTY BEDS Slate, acid tuff, quartzite, chert ABBOTSHALL BEDS Chert with interbedded basalt, acid volcanic and sedimentary rocks
Sequence 3 Black chert marker Tholeiitic basalt and 4 horizons of ultramafic rocks. Intercalated slate and chert bands. Minor high-Mg basalt		COOLGARDIE OPHIOLITES	*Association III* MULGABBIE FORMATION Basic intrusive and extrusive rocks. Ultramafic rocks, intermediate and acid extrusive rocks minor cherts		Golden Mile Dolerite Paringa Basalt Williamstown Dolerite Kapai Slate Devon Consols Basalt Hannans Lake Serpentinite	WOOLYEENYER GROUP Tholeiitic pillow lavas, slate, minor high-Mg basalt
Sequence 2 Equivalent and exactly the same as the Noganyer Group			*Association II* GINDALBIE FORMATION Clastic sequence	CYCLE 1		NOGANYER GROUP Banded iron formation, greywacke, shale, conglomerate sandstone
Sequence 1 Equivalent and exactly the same as Penneshaw Beds			Acid volcanic complexes *Association I* MORELANDS FORMATION Tholeiitic and ultramafic lavas			PENNESHAW BEDS Greywacke and acid lithic tuff. Tholeiitic basalt

In India neither Wadia (1926, 1957) nor Narayanaswarmi et al. (1960) noted an ultramafic zone in the Dharwar greenstone sequences.

Bliss and Stidolph (1969) have reviewed the lithological associations of banded iron-formations in Rhodesia. They have found that there are two major environments in which they occur (Figure 21) : volcanic and sedimentary. The iron-formations in the volcanic environments are almost wholly confined to the Bulawayan and Sebakwian rocks. Iron-formations associated with sedimentary rocks are found in both the Bulawayan and Shamvaian Groups.

In the volcanic environment the iron-formations are association with both mafic and felsic rocks. Often there is a band of sediments (limestone, phyllite or arkose) separating the iron-formation from the underlying volcanic rocks (Figure 22). No signs of turbulent conditions (e.g. current bedding, scour and fill structures) during deposition have been reported. Indications of minor slumping are common.

The rock types associated with iron-formations in the sedimentary environments include arkose, quartzite, phyllite and limestone. The sediments are well differentiated suggesting that the conditions under which they developed were quiet, allowing the chemical sediments to precipitate undisturbed.

Tomich (1978) has noted that in the Yilgarn Block of Western Australia, where iron-formations are prolific in a greenstone sequence, the lower horizons are associated with conformable ultramafic rocks. Higher in the stratigraphic sequence where other iron-formations occur, the uppermost horizons are associated with sediments and rhyolitic rocks. He suggests from evidence in the Marda area that the stratigraphically higher horizons of iron-formation are the richer deposits of iron ore. That is, they are oxide facies iron-formations and thicker than those associated with ultramafic rocks.

Beukes, 1973, Econ. Geol., Vol. 68

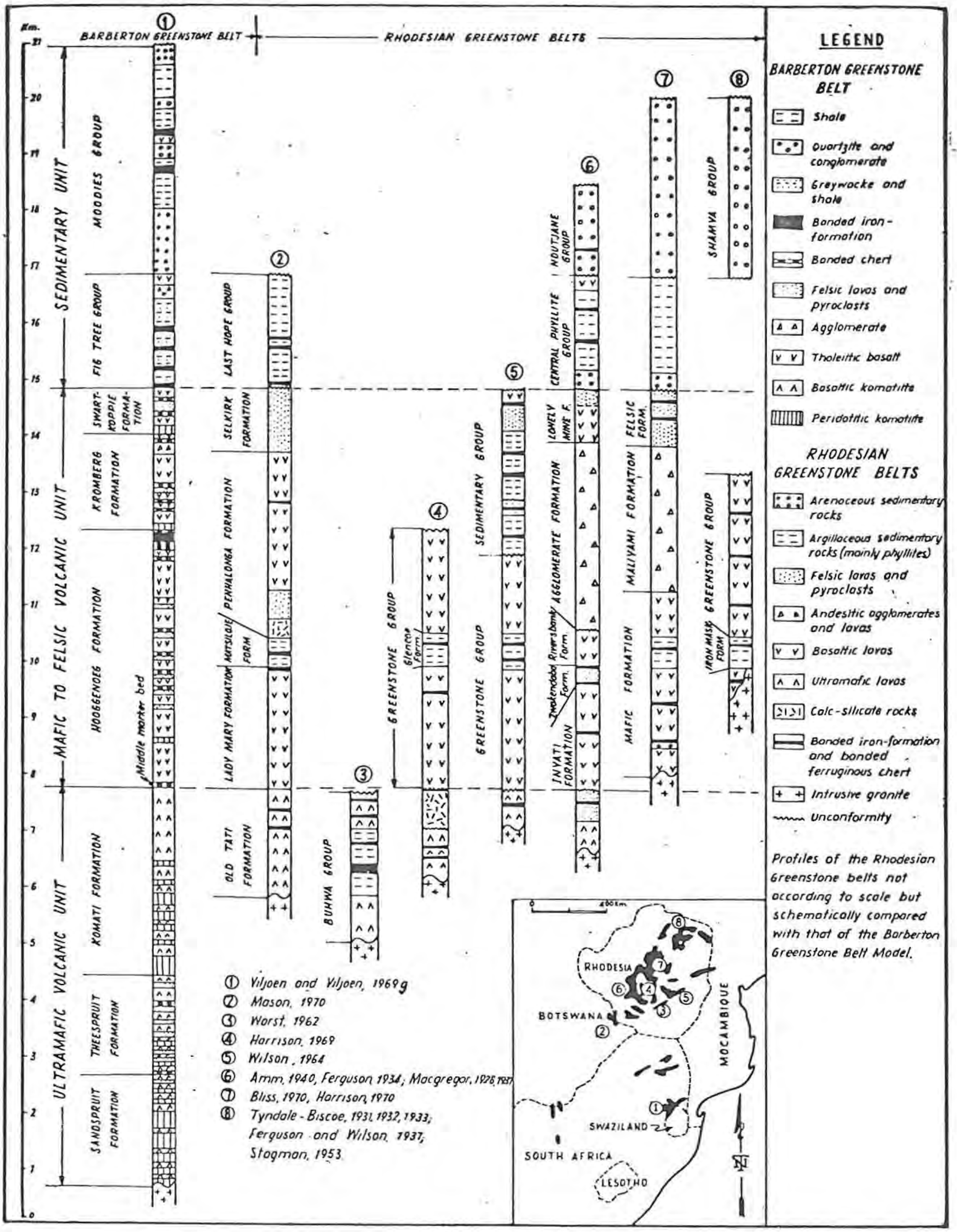


Fig. 21 Stratigraphy of the greenstone belts of south Africa, showing the distribution of iron-formations and chert

Beukes, 1973, Econ. Geol., Vol. 68

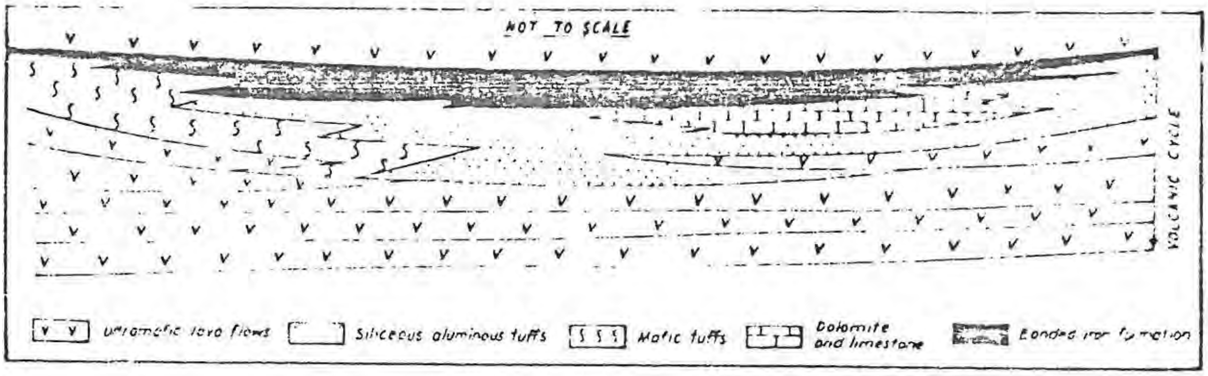


Fig. 22 The relationship between banded iron-formation and volcanic rocks in the volcanic units of the greenstone belts

Genesis of Iron-Formations in the Context of Greenstone Belt Development

It is generally accepted that greenstone belts were formed in geosynclinal type depositories on a primitive crust. The nature of this original crust is still being debated. Goodwin and Ridler (1970) imply that the original crust comprised a thin layer of oceanic type basalts with sialic nuclei distributed on it. A similar theory is proposed by Anhaeusser (1973). Other workers, notably Stowe (1968), Wilson (1973), Anhaeusser et al. (1969) and Hunter (1973) suggest that the basement on which the greenstones were deposited was a sialic complex comprising gneisses and migmatites. The notable similarities that the komatiitic rocks in the lower parts of the greenstone belts have to modern oceanic basalts suggests that at least in part, the crust was similar to oceanic basalts (Viljoen and Viljoen, 1969d; Anhaeusser, 1973).

It is possible that a thin sialic proto-crust (5 - 10 km thick) existed at this time. This crust would easily be broken up by a highly turbulent zone of melting below it, as suggested by Fyfe, 1974). This in turn could result in the fractures suggested by Anhaeusser et al.(1969) and Wilson (1973) (Figure 23). Fyfe (1974) suggests that further convective turbulence could cause these proto-cratonic plates to drift apart. If these hypotheses are accepted, the development of greenstone belts as envisaged by Goodwin (1973) and Goodwin and Ridler (1970) would be feasible.

These Canadian workers suggest that the individual basins or loci of greenstone development represent manifestations of thermal plumes in these fractures. Goodwin (1973) has divided the development of an Archaean 'basin' into three stages.

1. A spreading stage during which the sialic plates separated leaving a zone of primitive oceanic type crust between them. At the edges of the 'rift' arc type volcanics accumulated as the plates moved over 'hot spots'. These volcanic centres produced the massive sulphide deposits with the associated iron-formations, intercalated among the felsic extrusives and pyroclasts.

2. An exhalative stage coincided with the waning stages of thermal activity. It is probable that there was not one single period of exhalative activity, but a series, resulting in the intercalated chert and iron formation typically associated with greenstone belt mafic volcanics (Figure 24).

3. The final stage is interpreted as the orogenic stage by Goodwin (1973). In this stage he envisages the final deformation of the greenstone basin. Rather than an orogenic occurrence in the Alpine sense, involving plate tectonics, as suggested by Anhaeusser (1973), it is more probable that the convection cell collapsed and the granitic forelands moved back towards each other. This would result in the compression of the material between them causing the synclinal features so typical of greenstone belts around the world.

Later deformation by late stage intrusive granitic plutons resulted in the arcuate shapes of the greenstone belts.

Anhaeusser <u>et al.</u>, 1969, Bull. Geol. Soc. America, Vol. 80

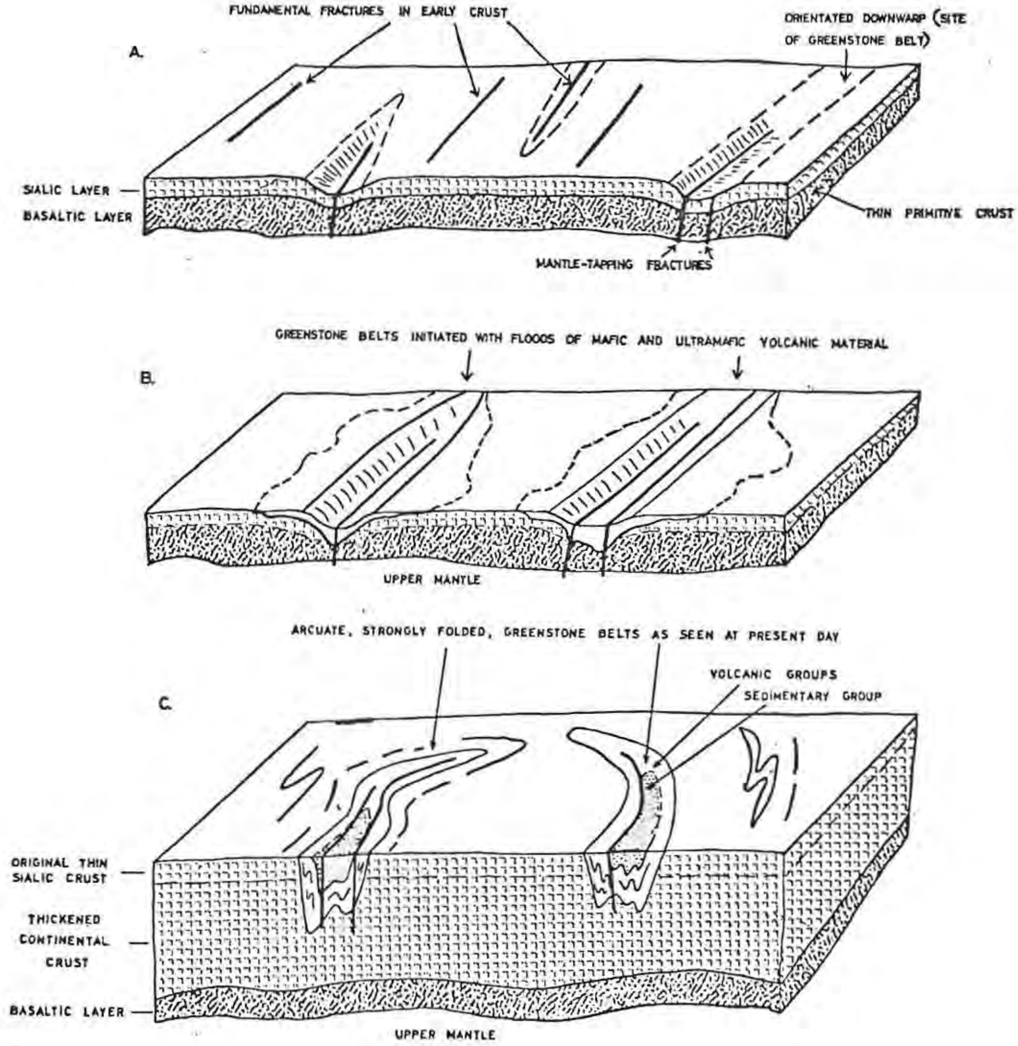


Fig. 23 Diagrammatic illustrations showing the suggested evolution of greenstone belts.

Beukes, 1973, Econ. Geol., Vol. 68

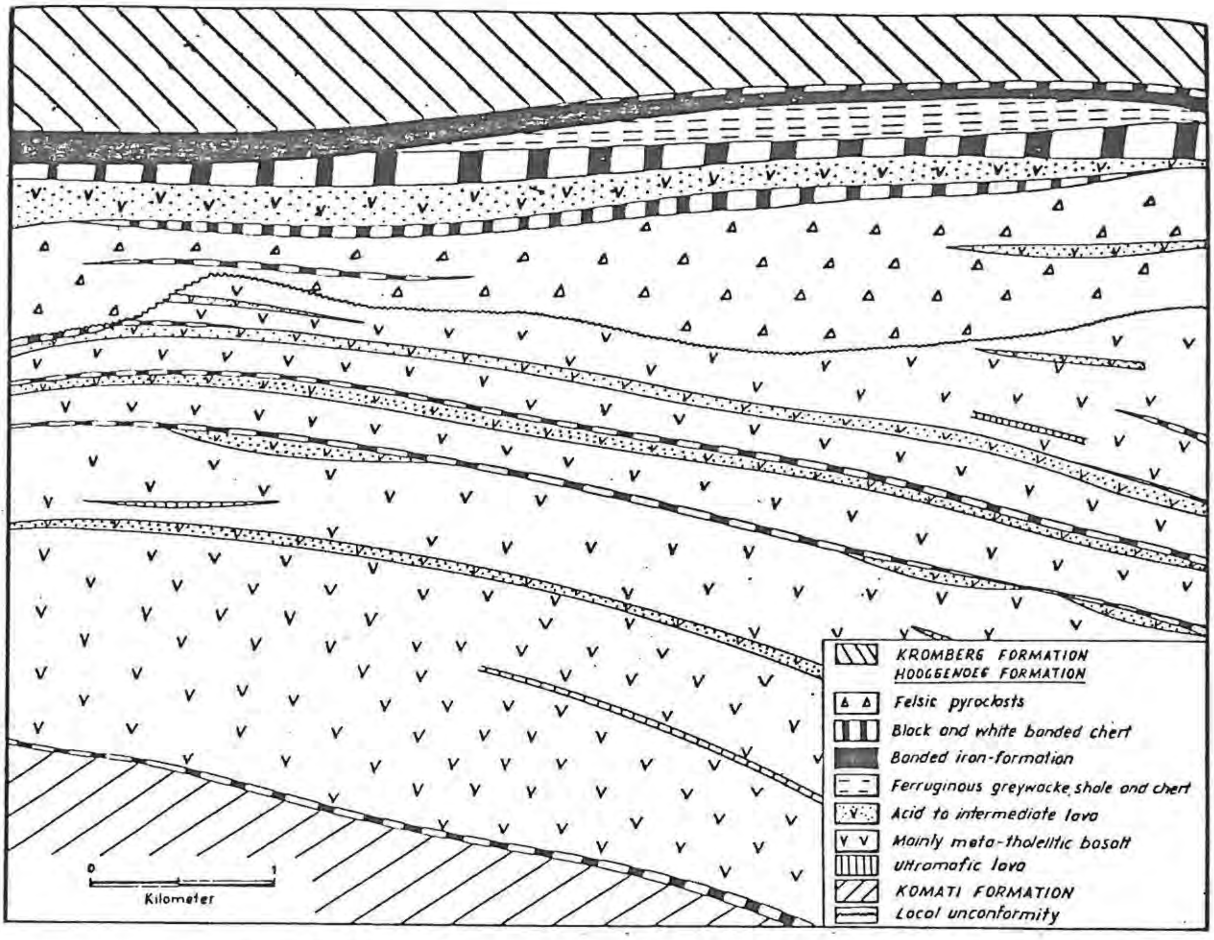


Fig. 24 Generalized map of the type area of the Hooggenoeg Formation in the Barberton greenstone belt showing the relation between minor cycles of volcanism, banded-chert, and banded iron-formation. Modified after Viljoen and Viljoen, 1969f.

Depositional Environments of Iron-Formation

Deposition environments of iron-formations within the context of Archaean basins in Canada have been described and discussed by Goodwin (1973), Goodwin and Ridler (1970), and Goodwin and Shklanka (1967). Ten such basins have been identified in the Slave, Churchill and Superior provinces (see Figure 14). The features of Canadian Archaean supracrustal assemblages are the result of a sequence of volcano-tectonic events (Goodwin, 1973). This model can probably be applied to Archaean terrains found elsewhere in the world.

The stratigraphic associations of iron-formations in the Michipicoten area are depicted diagrammatically in Figure 25. This diagram shows the lenticular nature of the rock units, although certain units do cover most of the area. The lensoid nature of the stratigraphy illustrates how even lithological correlations can be incorrect.

Goodwin, 1973, Econ. Geol., Vol. 68

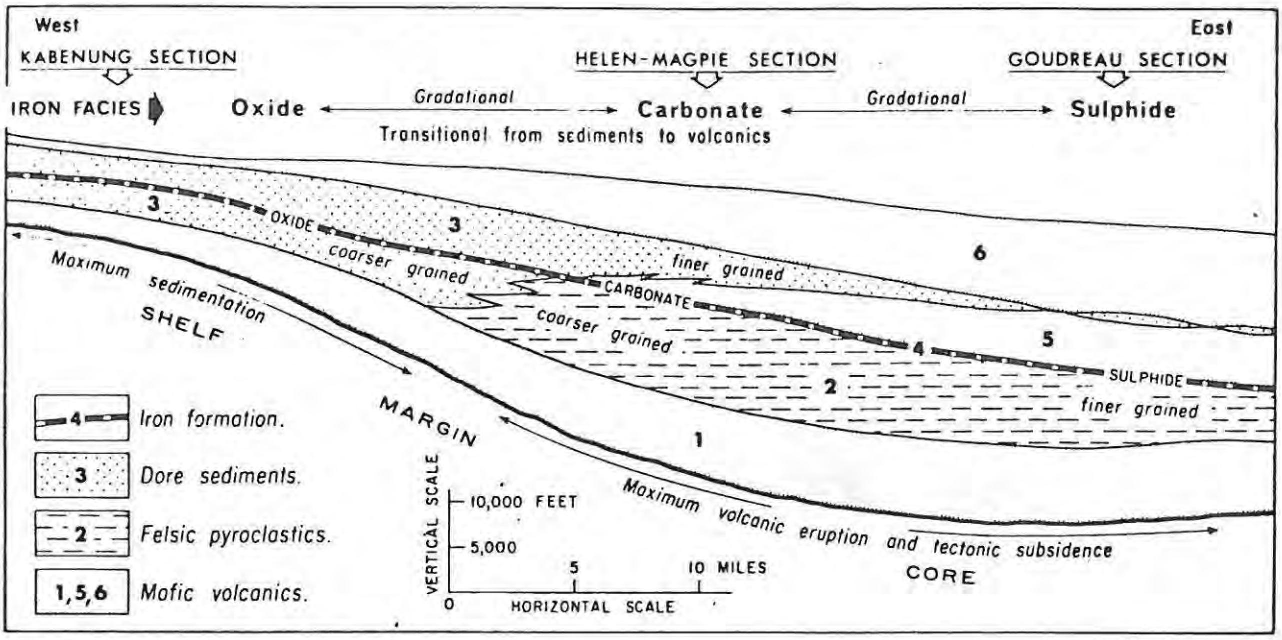


Fig. 25 Reconstructed stratigraphic section of the Michipicoten basin. The oxide-carbonate facies transition marks the slope-edge of the shelf upon which rests conglomerate-bearing sediments. The carbonate-sulphide facies transition overlies felsic pyroclastic piles within the margin of the basin with sulphide facies in the deeper water direction. Many Archean paleoslopes lack substantial carbonate facies in which case the oxide-sulphide facies transition together with felsic volcanic piles and proximal conglomerates serves to indicate the margin of the basin.

The stratigraphic sequence consists of six units (Goodwin, 1973). Unit 1, at the base, comprises mainly mafic lava flows and hypabyssal intrusions. In the deeper parts of the original basin these are overlain by felsic volcaniclastics (unit 2), iron-formation (unit 4), more mafic volcanics, and then thin lenses of clastic sediments (unit 3). At the top is a young sequence of mafic volcanics (unit 6).

In the shallower parts of the basin, overlying the mafic base is a thick sequence of coarse clastic sediments (unit 3) with intercalated iron-formation (unit 4). The clastic sediments are also overlain by unit 6 mafic volcanics.

The volcanic rocks, particularly the felsic volcaniclastic deposits in unit 2, have been compared to island arc sequences by Goodwin (1973). The pyroclastic deposits are usually to be found as discrete piles measuring 15 to 25 km long by eight km wide and about three km thick. They comprise tuffs, breccias and lava flows of intermediate to acid composition. Abrupt lithological alterations are typical of these rocks. Cycles of this type are well documented (see for example Spence and de Rosen-Spence, 1975, Econ. Geol., Vol. 70). These piles are thought to mark original centres of volcanic activity.

The iron-formations in the sequence, as in Rhodesia, have two main lithologic associations : volcanic and sedimentary. The sedimentary association is found in the shallower regions of the basin, typically towards the edges.

In the Abitibi belt, Goodwin and Ridler (1970) have divided the sedimentary facies into two distinct types : volcaniclastic and flyschoid (Figure 26). The volcaniclastic sediments comprise greywackes, shales, sandstones, conglomerates, breccias and iron-formations. The clastic deposits have the same chemical composition as the local volcanic rocks. In the Michipicoten area of Goodwin (1973), these sediments are probably included in Unit 2 of that stratigraphic sequence. The flyschoid facies consists of rythmically bedded greywacke-argillite sequences, derived from a tectonically stable craton.

Mason (1969) has pointed out that in Rhodesia there is a similar distinction in sediment types. The pre-Shamvaian sediments are generally composed of reworked volcanic material. The Shamvaian sediments are arkosic rocks derived from a granitic source, with minor intercalations of greywackes and shales derived from volcanic rocks.

In South Africa, particularly in the Barberton greenstone belt a similar division of the sedimentary sequences has been noted (see Viljoen and Viljoen, 1969c). The Fig Tree Group pelitic sediments have mineralogical and geochemical features that indicate they were derived from a terrain with a large ultramafic component. The overlying Moodies Group sediments are generally rudaceous to arenacious. The rudaceous horizons are polymictic with pebbles of a wide variety of compositions enclosed in a matrix of arkosic sandstone or sandy shale (Viljoen and Viljoen, 1969c).

Goodwin and Ridler, 1970, Geol. Surv. Canada, Paper 70-40

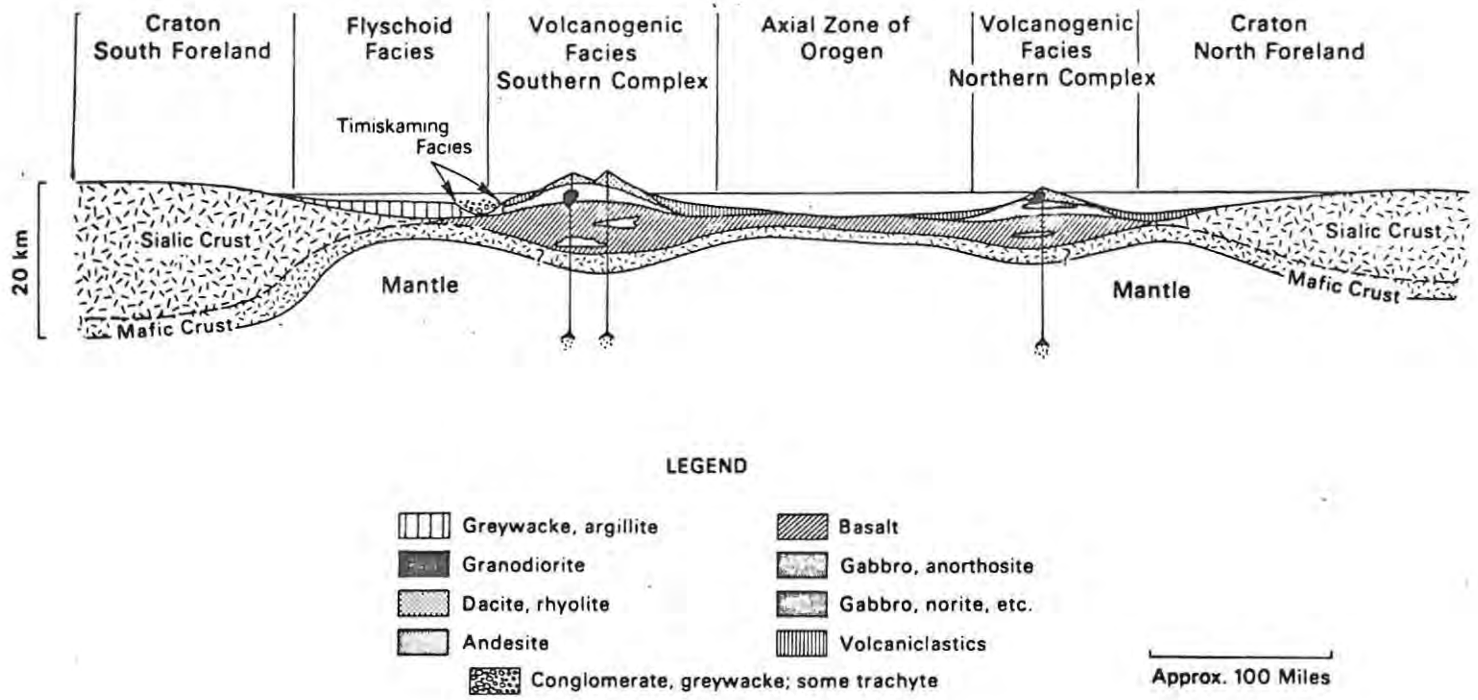


Figure 26 Hypothetical tectonic reconstruction of Abitibi orogen. Vertical cross-section. Length of the orogen is approximately 500 miles. Distance between southern and northern volcanogenic facies may have been in the order of 200 miles. Vertical scale is approximate only. The line of the section is intended to cross Noranda in the south and Matagami in the north. The smaller size and relief of the northern sialic foreland is indicated by the absence of flyschoid facies along its southern margin. The nature and thickness of mafic crust underlying the orogen is conjectural. In the absence of direct evidence of sialic contribution to the volcanic and sedimentary assemblages of the orogen (except for the flyschoid facies, i.e. Pontiac, at the south margin), no subjacent sialic crust is shown as being present, at least during the early development stages of the orogen.

Iron-formations are widespread in the greenstone belts of the Canadian shield. Oxide facies iron-formation are found in the shallower regions of original basins investigated by Goodwin (1973) and Goodwin and Ridler (1970). In the deeper regions, carbonate and sulphide facies iron-formations are characteristic (Figure 27). This supports the conclusions reached by Tomich (1978) from studies in the Marda area. Typical sections through portions of the Michipicoten area can be seen in Figure 28.

Goodwin, 1973, Econ. Geol., Vol. 68

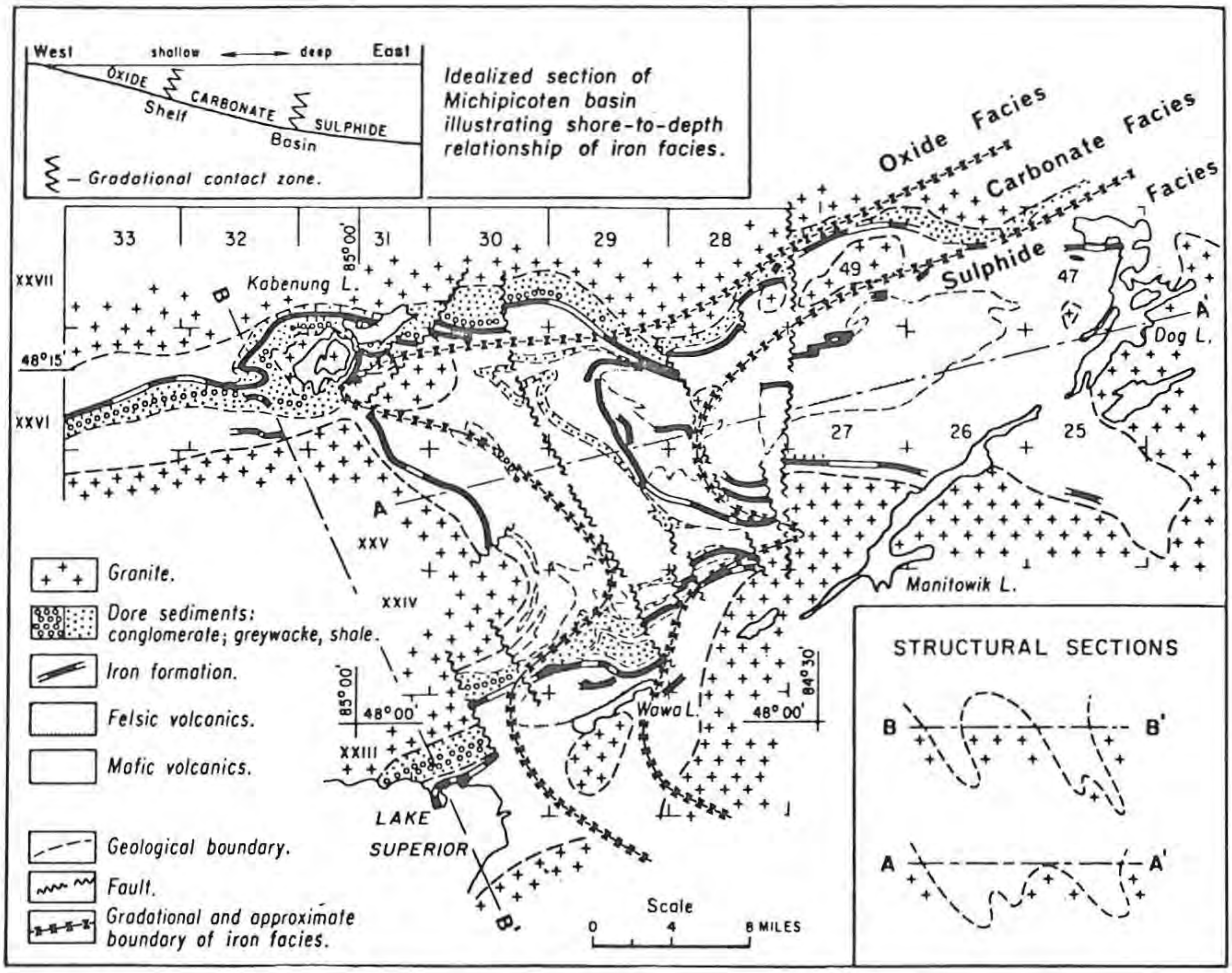


Fig. 27 General geologic map, Michipicoten area, Ontario, illustrating distribution of Archaean iron-formation by facies relative to mafic and felsic volcanic rocks and clastic sediments. The triple lithofacies association of 1) oxide-carbonate-sulphide facies transition, 2) arc-type felsic pyroclastic piles, and 3) proximal conglomerates, indicates the presence of an Archaean paleoslope as illustrated in the inset. The structurally deformed state of the supracrustal assemblages is indicated by the structural sections.

Goodwin, 1973, Econ. Geol., Vol. 68

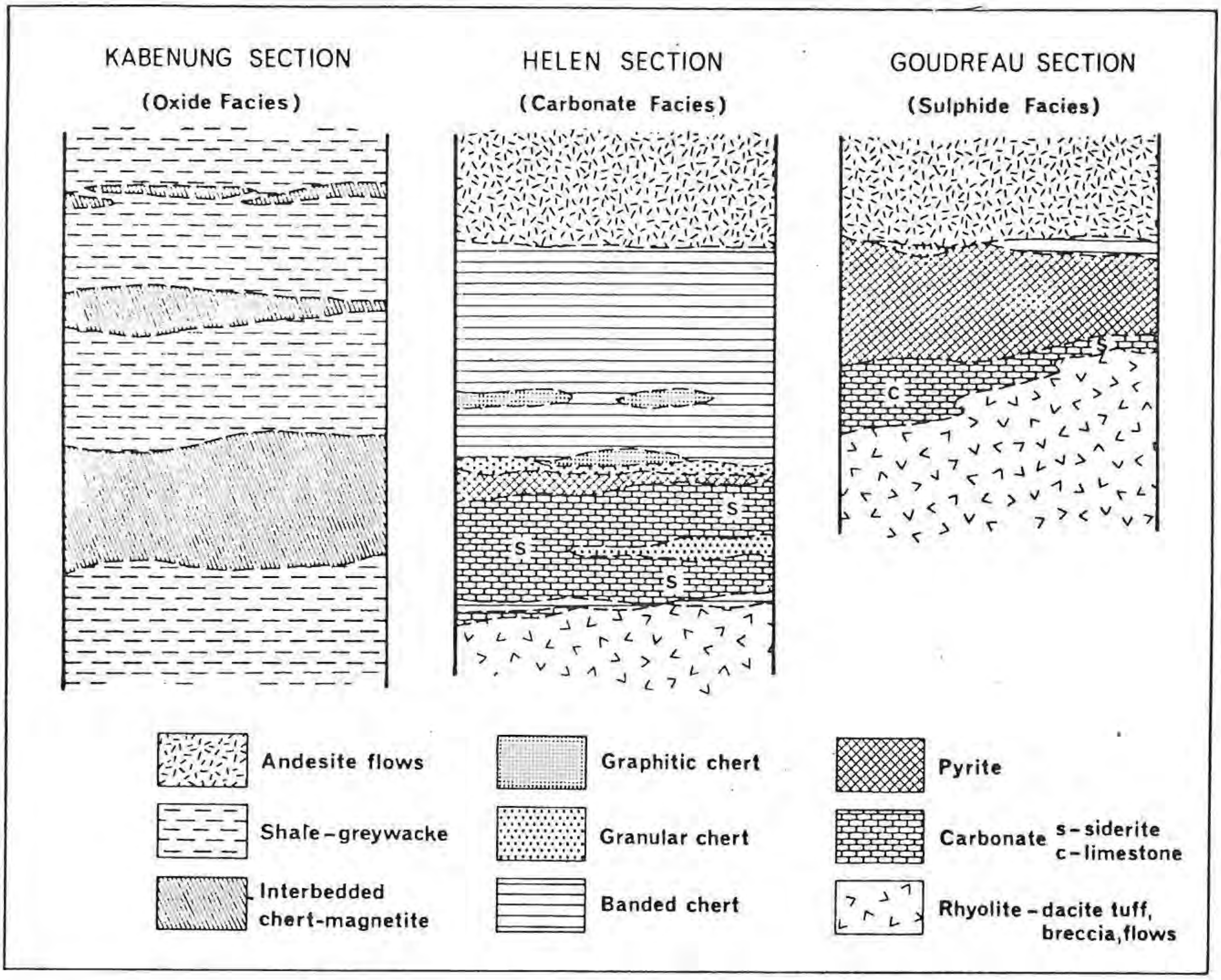


Fig. 28 Stratigraphic columns of Michipicoten oxide, carbonate, and sulphide facies of iron-formation. The oxide facies, as illustrated, is mainly enclosed in sedimentary rocks; however, it is locally enclosed in volcanic rocks. The three facies are intergradational across the area in the order shown from shallow water on the left (oxide facies) to deeper water on the right (sulphide facies). Note that the proportion of interbanded chert decreases markedly in the direction of deeper water and is absent to rudimentary in the sulphide facies iron-formation. The sulphide facies as illustrated was probably originally transitional, in turn, to still deeper water sulphidic, carbonaceous shale characteristic of the euxinic environment. Vertical scale : 1 inch equals approximately 200 feet.

Iron-formations associated with volcanic rocks is made up of three main facies types. In descending order they are banded chert, sulphide facies and carbonate facies. The increase in sulphide content is thought to be a function of depth. In the deepest parts of the basin sulphide argillaceous iron-formation is present indicating that euxinic conditions existed at the time of deposition.

THE FORMATION OF MINERAL DEPOSITS IN ARCHAEAN BANDED IRON-FORMATION

The close association of Algoman-type Archaean iron-formations with volcanic rocks has been demonstrated by many workers (e.g. Fripp, 1974, 1976a, 1976b; Goodwin, 1973). Other workers (e.g. White, 1967; White and Waring, 1963) have shown that metals and silica are contained in volcanic emanations. Weissberg (1969) and Ferguson and Lambert (1972) have described how, although the concentration of metals in fumerolic brines may be low, the precipitates can be considerably richer. Thus it appears that these metal deposits are probably volcanogenic.

The precipitation of the metals, sulphur and silica could be caused by physico-chemical mechanisms as discussed earlier, or by biochemical activity. LaBerge (1973) discusses at length microfossil forms found in iron-formations. These micro-organisms are thought to have been possible oxygen donors for the precipitation of ferric oxides. Cloud (1973) and Huber (1959) suggest that seasonal controls of the organisms could have resulted in the fine banding. Pflug (1966) has described similar structures in the Middle Marker horizon of the Onverwacht Group, South Africa. He suggests that these micro-organisms may be responsible for the enrichment of metallic trace elements found there.

Fripp (1974) has proposed a model for the development of sulphide facies iron-formation. He has since modified the original figures for temperature and depth of formation (Fripp, 1976b). In relation to the fumerole vent, iron-formations can be precipitated in three situations. In Archaean times because the crust was much thinner, its temperature was greater than it is today. In shallow water (0-20 m), the water would be near neutral with respect to pH, and oxygenated (Drever, 1974). Brines reaching the surface of the crust would have temperatures in the region 95°-130°C. The solubility of gold in a brine entering a shallow water environment would be lowered by the decrease in temperature. The silica, iron-sulphide and carbonate minerals in solution would also maintain a delicate balance with the sea water. Iron-formation precipitated in this environment would contain layers of chert, oxides, sulphides and carbonate. The tenor of the gold would be low because of the dilution effect of the other minerals.

In deeper water (20-200 m) the brines escaping from the crust would be hotter (130°-205°C). Precipitation of the sulphides and gold would be induced by the changes in temperature and pH. Carbonate minerals would be the first to precipitate, being less soluble in warmer solutions than cold solutions, followed by the sulphides and the gold.

In still deeper water (>200 m), the brines reaching the surface would possibly be even richer in sulphides, thus suppressing the solubility of gold, so that the amount of gold available would be minimal. However, base metals and iron would be more soluble in the brine, thus a deposit with a higher base metal content and a lower gold content would be formed.

Sato (1972) working with Kuroko type massive sulphide deposits has proposed a deposition model that may be applied to volcanogenic iron formations. He suggests that the site of precipitation is dependent on salinity and temperature (Figure 29). High density brines (types I, IIa and IIb) would not be widely dispersed, they would tend to precipitate their solutes relatively close to the vent and in depressions. Deposits rich in sulphides are likely to result from such brines.

Sato, T. 1972, Min. Geol. 22

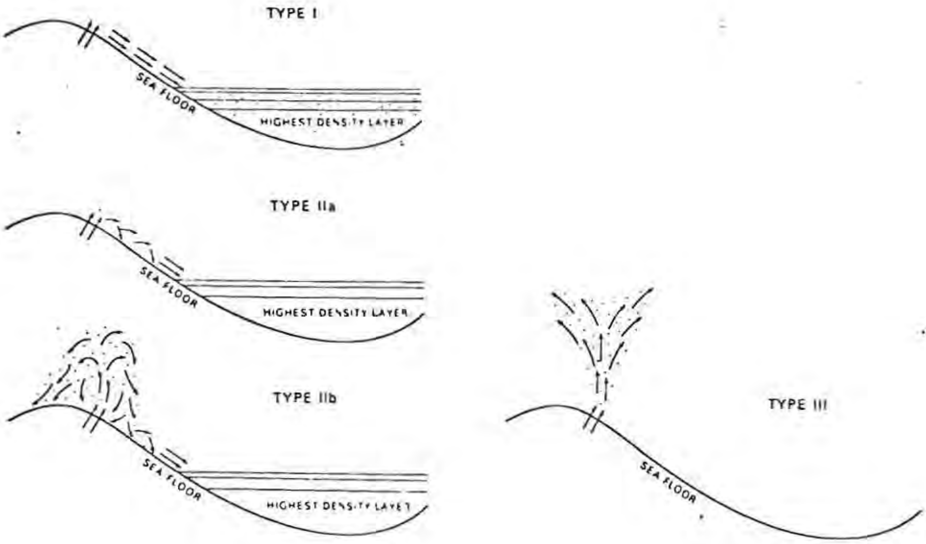


Fig. 29 Probable behaviours of ascending ore-forming solutions welling up onto the sea floor, according to the four basic types of temperature-salinity relations. Dotted areas are sites of mineral precipitation, assuming that minerals are precipitated during early stages of mixing with sea water.

Low density brines (type III) are likely to be more widely dispersed. As they ascend to the upper levels of the sea, their contents will be oxidized, and oxide facies deposits will probably form over a wide area.

The cause of the banding in iron-formations is highly speculative. Lambert and Bubela (1970) have shown that banding can be produced experimentally in sulphide ores. Goodwin (1956) suggested that rhythmic banding could be the result of precipitation from exhalations alternatively rich in silica and iron. Lepp and Goldich (1964) proposed that the chert layers resulted from selective replacement of carbonates by the silica. Similar selective replacement mechanisms have been put forward to explain sulphide facies iron-formation. Gair (1962) and Tolbert (1964) suggest that the sulphides have selectively replaced the iron oxide minerals. Sawkins and Rye (1974) and Rye and Rye (1974) have shown that at the Homestake deposit, South Dakota, the sulphur in the sulphides associated with the gold is a primary constituent of the deposits.

It is suggested here that the banding may be a function of the density of the minerals. From a volcanic exhalation, the less soluble products would be precipitated. The denser iron minerals would sink faster to form a band followed by the less dense silica. This sequence would be repeated for successive exhalations.

THE GEOLOGY OF GOLD DEPOSITS IN IRON-FORMATION

A recent study of auriferous iron-formation is that of Fripp (1974, 1976b). He investigated the stratigraphic position of these deposits and suggested that, based on the Barberton model of Viljoen and Viljoen (1969a), these deposits in Rhodesia occur in the Sekakwian Group (Fripp, 1974, 1976a, 1976b).

Vubachikwe Gold Mine

The Vubachikwe mine is in the Gwanda schist belt in the south of Rhodesia. The country surrounding the mine is underlain largely by mafic to felsic volcanic rocks. Minor amounts of ultramafic material,and metasediments including iron-formation, are also present (Fripp 1974, 1976b; Tyndale-Biscoe, 1940). The rocks have been metamorphosed to greenschist facies.

Stratigraphy at the Vubachikwe Mine

The stratigraphy at the Vubachikwe mine is relatively simple (Figures 30 and 31). The rocks strike at approximately 150° and dip at 70° to the south west. At the base is a thick (1000 m) felsic unit comprising schistose quartz-feldspar porphyry rocks with interbedded crossbedded meta-aluminous felsic tuffs. Thin layers of felsic pyroclasts, chert and quartzite are also present.

Above this felsic unit is a 2000 m thick mafic unit comprising meta-volcaniclastic rocks and basaltic lavas. Fripp (1974) has divided this unit into four sub-units.

(i) Sub-unit A at the base starts with a thin layer of mafic schists overlain by three or four horizons of oxide facies iron-formation. The rest (about 2/3 of the 90 to 400 m thick sequence is mafic lavas and tuffs, overlain by ultramafic rocks, which are mainly talc-actinolite schists and meta-peridotitic lavas.

(ii) Sub-unit B contains the ore horizons of the Vubachikwe mine. The base of this sub-unit is the first iron-formation horizon above the ultramafics at the top of sub-unit A. Sub-unit B comprises thin iron-formation horizons intercalated with mafic lavas and tuffs. Some pyrrhotite-rich argillites and iron-rich limestones are also present. This unit is 150 to 350 m thick.

(iii) Sub-unit C is about 1500 m thick comprising mafic pillow lavas with intercalated calcareous or marly beds. Along strike near the Blanket mine, these beds are associated with gold-arsenopyrite mineralization.

(iv) Sub-unit D consists of about 900 m of porphyritic andesite tuffs and agglomerates intercalated with pillowed actinolite schists. Metagabbroic horizons and a cross-bedded calc-silicate unit are also present.

The ore Bearing Zone

The ore zone lies about 500 m above the base of the mafic unit. Two sections through the sequence are depicted in Figure 31, through AT and CV shafts.

Fripp, 1976b, Econ. Geol., Vol. 71

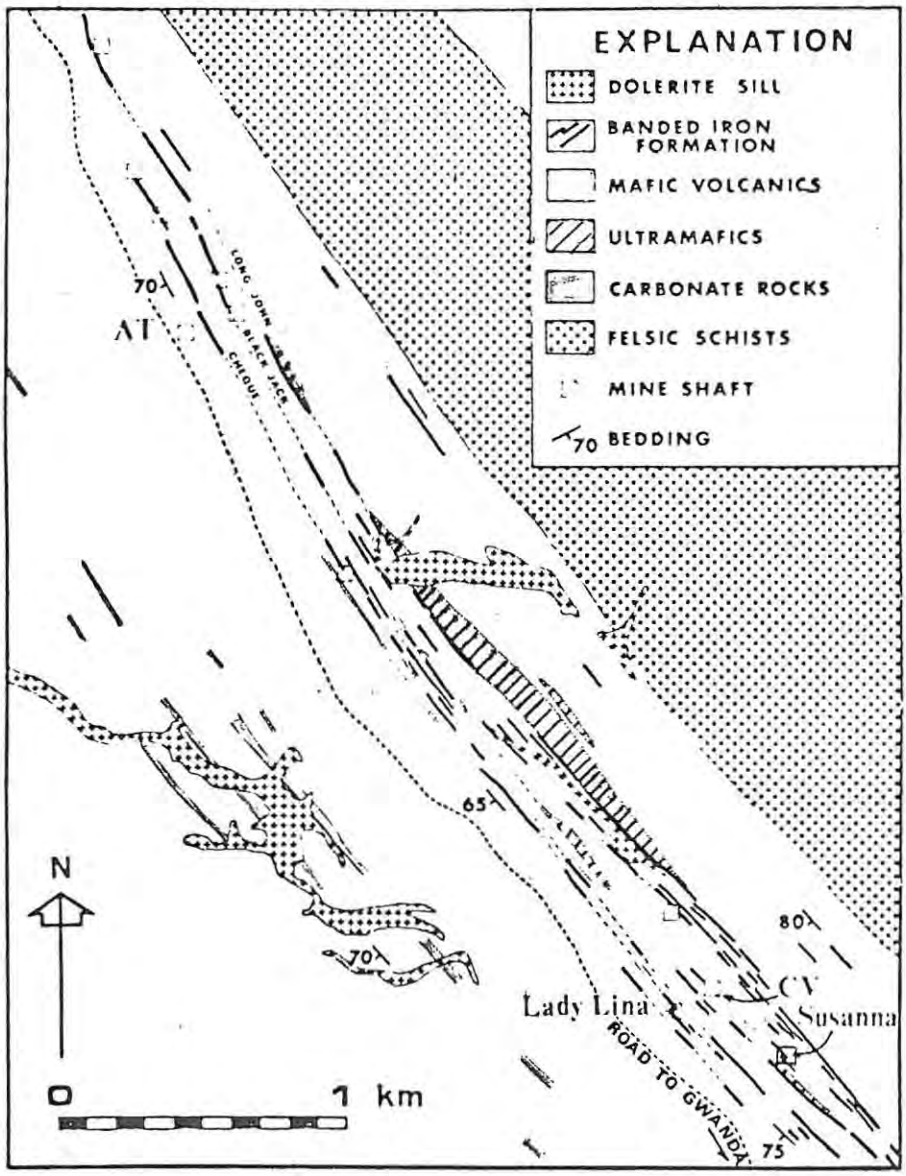


Fig. 30 Simplified surface geology of the Vubachikwe mine area. The total succession is shown in Figure 31.

Fripp, 1976b,Econ. Geol., Vol. 71

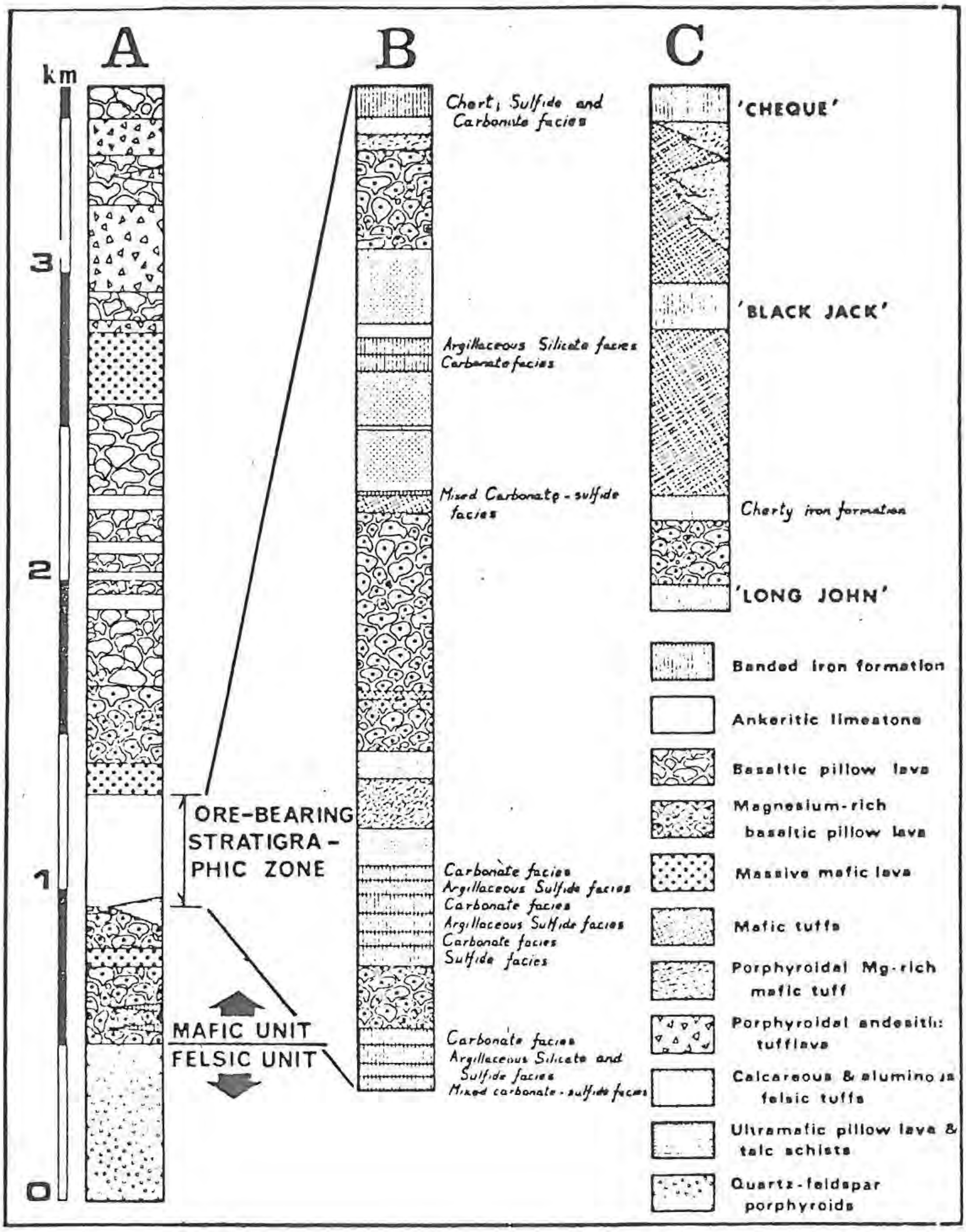


Fig. 31 A, Stratigraphic succession in the Vubachikwe mine area
B, Detailed succession at the CV shaft
C, Detailed succession at the AT shaft

In the At shaft the succession comprises three auriferous horizons of banded iron-formation interlayered with mafic and felsic tuffs and lavas. The lowermost iron-formation, the "Long John", is overlain by tremolitic pillow lavas, which in turn are overlain by a thick succession of mafic tuffs. The "Black Jack" iron-formation overlies these tuffs. Between the "Black Jack" iron-formation and the stratigraphically higher "Cheque" iron-formation is a sequence of porphyroidal mafic tuffs.

At CV shaft the sequence is about twice as thick as at AT shaft. The main difference between the two sequences, is the presence of thick beds of felsic tuffs and more iron-formation horizons at CV shaft.

Petrography of the Iron-Formation

The dominant facies of the iron-formations present is carbonate facies, followed by sulphide and silicate facies, with very little oxide facies present.

The oxide facies iron-formations are closely associated with the silicate facies material. Often the oxide facies contains considerable amounts of biotite, tremolite and pennine.

The silicate facies occurs as two varieties :

(i) An indistinctly layered melanocratic argillaceous rock consists of layers rich in pennine, biotite and tremolite alternating with dark cherty layers. Garnet-rich layers occur occasionally.

(ii) A chert-rich variety consists of distinct layers of tremolite and/or biotite alternating with lighter coloured silica. This facies variant was found in the "Long John", "Black Jack" and "Cheque" horizons. Small amounts of carbonate minerals and garnets were also found.

The sulphide facies also occurs as two varieties :

(i) A more argillaceous variety comprises alternating layers of sulphides (pyrrhotite mainly), quartz and silicates (usually pennine rich). Carbonates and garnets occur as accessories. This rock-type frequently grades into the argillaceous variety of the silicate facies.

(ii) The less argillaceous variety of sulphide facies makes up the major portions of the "Long John", "Black Jack" and "Cheque" horizons at AT shaft. The rock is well layered; the layers being sulphide (usually arsenopyrite) rich, quartz rich, quartz and silicate rich, or quartz-carbonate rich. This rock-type contains a large amount of ankerite, often to the extent that it could be called a mixed sulphide-carbonate facies iron-formation.

The carbonate facies iron-formation consists largely of alternating layers of iron-rich carbonate (siderite, ferroan dolomite, ankerite) and quartz. This is the dominant facies of iron-formation at CV shaft. Siliceous ferroan carbonate beds in the sequence often grade into carbonate facies iron-formation.

The Distribution of Ore

The gold ore is confined to relatively thin beds of iron-formation. These beds occur in what Fripp (1974) has described as "mega-boudins". The gold-bearing sulphides are found as concordant interlayers and lenses. Even when folded the lenses and layers are concordant with the banding in the iron-formation. The zone of sulphide mineralization may be thick or thin. In one bed of iron-formation there maybe several zones of sulphide mineralization seperated by thick zones of silicate or carbonate facies material. Where the sulphides are sufficiently rich in gold, and in mineable quantities, they constitute the ore. The gold is found only in the sulphide facies iron-formation.

The gaps between the "mega-boudins" contain crushed and brecciated iron-formation and greenstone, with irregular quartz veins.

Mineralogy of the Ore

The gold is contained in the sulphides in the sulphide facies, or mixed sulphide-silicate facies, or mixed sulphide-carbonate facies iron formation. Arsenopyrite occurs as well formed crystals with anular zones of inclusions. The gold is found mainly within the crystals of arsenopyrite. The gold grains are irregular or rounded in shape, and up to 25μ in diameter. Occasionally where the arsenopyrite has been crushed, the gold occurs as grains between the larger fragments of arsenopyrite, or between the silicate and sulphide fragments. About 85% of the gold is found associated with the arsenopyrite.

Neardy Silver-Lead Deposit

Fripp (1974) has described the geology of this deposit demonstrating the close association of base metal sulphide deposits with iron-formations. The deposit is in the Umtali-Odzi schist belt in the eastern part of Rhodesia. The rocks of the region are metamorphosed mafic, ultramafic and felsic volcanics and volcaniclastics with some sediments, including banded iron-formation. In the vicinity of the workings the rocks are mainly ultramafics comprising peridotites and serpentinites which have been metamorphosed to greenschist facies. Two sequences, each about 500 m thick, of mafic and felsic rocks with tuffs, argillites and banded cherts and iron-formations are present. The northern mafic to felsic sequence contains the Neardy deposit, and along strike several other minor base metal deposits.

Petrography of the Neardy Deposit

Beds of banded chert and iron-formation are interlayered with mafic and felsic rocks. The iron-formations occur in two groups, each of four horizons, separated by about 50 m of tuffs and tuff-lavas. In each group only one horizon has been investigated as they contain most sulphide facies material. The horizons are 10 to 25 m thick.

The iron-formations show well developed mineralogical layering. The iron-formation beds that are plumbiferous comprise several varieties of sulphide rich facies.

(i) Angillaceous varieties have two major components : dark cherty layers, and sulphide-rich layers containing some silicate and carbonate material. The sulphide layers are usually pyrrhotite rich, but there are some rich in galena and chalcopyrite.

(ii) The massive sulphide variety consists of alternate layers of pyrrhotite and pyrite with lesser amounts of galena and chalcopyrite. Cherty layers may also be present.

(iii) Other varieties of sulphide facies iron-formation comprising alternate layers of dark and light coloured rock are also present. The dark layers consist of pyrite, pyrrhotite and magnetite. They usually occur as almost monomineralic layers. In the sulphide rich layers chalcopyrite and galena occur as minor constituents. The light coloured layers consist of quartz with some opaque ore minerals, chlorite, carbonate and tremolite.

The interbedded tuffs are poorly layered schistose rocks consisting of quartz, hornblende, tremolite, pennine, phlogopite, biotite and zoned plagioclase. Tuff-lavas are a variety of rocks that show no signs of original bedding. They are schistose rocks containing plagioclose, hornblende and tremolite, with minor amounts of quartz and chlinochlore. Small amygdales are sometimes present. Ultramafic schists with spinifex textures are also present. The olivine blades in these rocks have been altered to antigorite.

Other Rhodesian Deposits

Other deposits investigated (Fripp, 1974, 1976b) include Beehive, Camperdown, Connemara, Empress, Marvel, Nelly 404, Pickstone and Sherwood Starr. At all these deposits gold ore has been won from sulphide facies and mixed sulphide facies iron-formations. The iron-formations are interbedded with magnesium-rich mafic and ultramafic rocks.

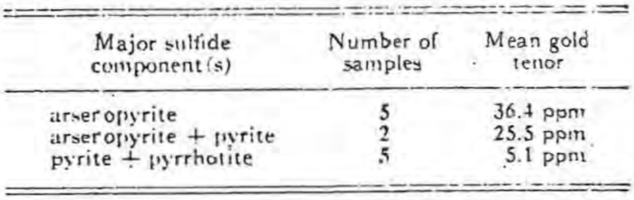
The sulphide iron-formations constituting the ore at these deposits are all well layered rocks with the layering ranging from 1 mm to 40 mm in thickness. The quartz-rich layers are generally milky white or dark and translucent, and alternate with sulphide rich layers. At the Beehive, Marvel and Sherwood Starr, the sulphide is almost exclusively arsenopyrite. At Empress and Pickstone, both pyrite and arsenopyrite are present, while at Camperdown, Connemara and Nelly 404, the dominant sulphides are pyrite and pyrrhotite.

Iron-rich carbonate minerals are present at all the deposits. Often they form layers or zones in the iron-formations. In the arsenical ores, the gold was found to occur as microscopic inclusions within the arsenopyrite and are therefore "free-milling ores". The pyritic or pyrrhotitic ores however are "roasting ores".

The relationship between the gold content of the iron-formation samples and the associated sulphide minerals is shown in Table 9. It appears from this table that there is a direct correlation between arsenic and gold.

Fripp, 1976, Econ. Geol., Vol. 71

TABLE 9 Relationship between Gold Content of Iron-Formation Samples Analyzed and Their Sulphide Mineralogy



Major sulfide component(s)	Number of samples	Mean gold tenor
arsenopyrite	5	36.4 ppm
arsenopyrite + pyrite	2	25.5 ppm
pyrite + pyrrhotite	5	5.1 ppm

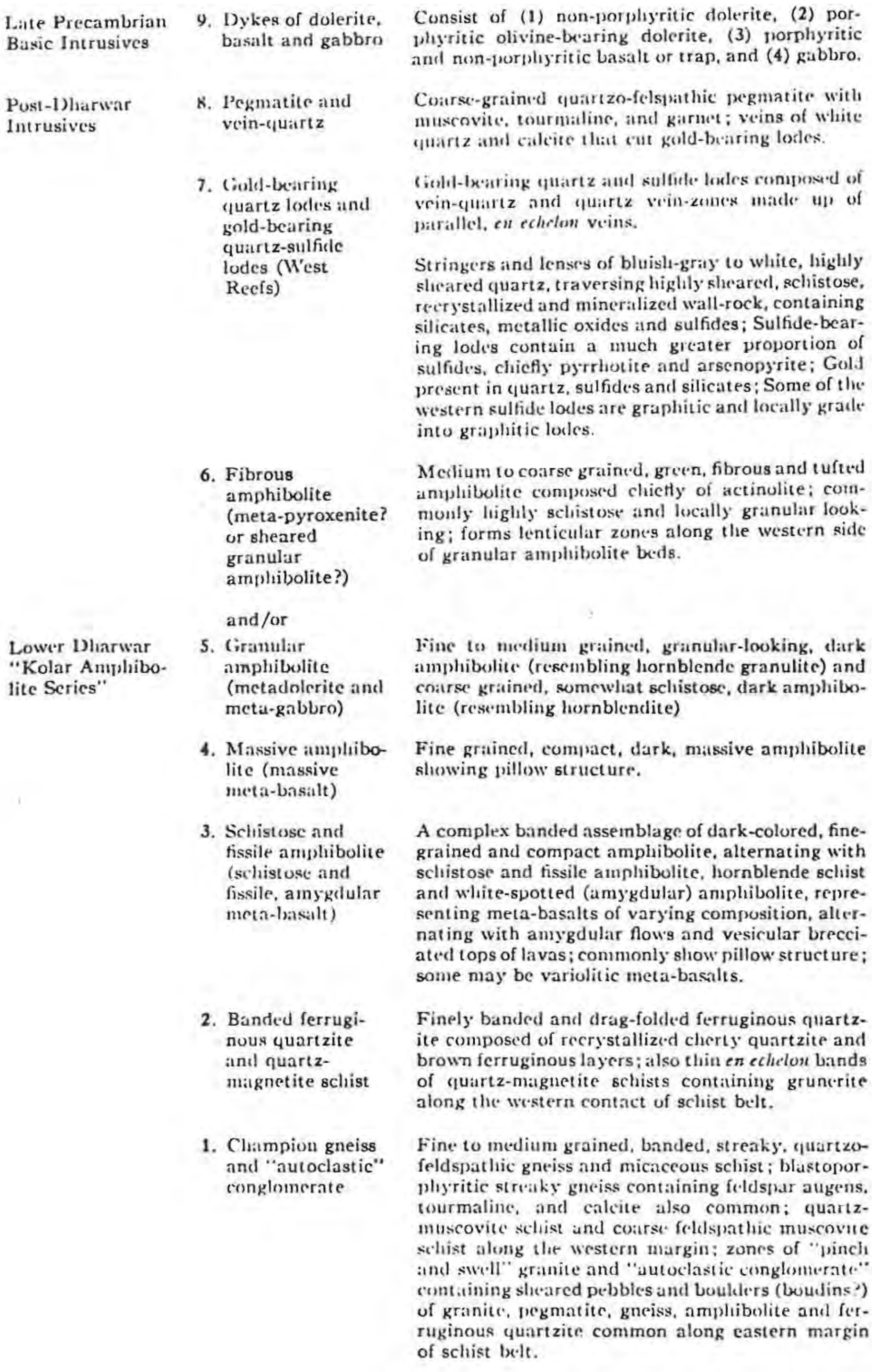
Deposits in the Kolar Goldfield, India

In the Kolar Goldfield in India (Narayanaswami et al., 1960), the Oriental, McTaggart and other sulphide bearing West Reefs appear from their descriptions to be sulphide facies iron-formations although the authors suggest that ".... sulphides are seen clearly to replace the silicates and quartz" (p. 1448). In these lodes "The chief sulphides are, in order of abundance, pyrrhotite, arsenopyrite, pyrite, chalcopyrite and minor galena. Pyrrhotite and arsenopyrite together make up more than 90% of the sulphides. The pyrrhotite, which is the most abundant and forms nearly 75% of the sulphides occurs as disseminate grains and granular aggregates forming parallel layers, bands and lenses following the foliation and banding of the lodes".(Narayanaswami et al., 1960 pp. 1447-1448).

The stratagraphic control of these deposits is well demonstrated. The stratagraphic sequence of the Kolar Amphibolite rocks is shown in Table 10, and in diagrammatic section in Figure 32. The deposits are always associated with the contacts between a sequence of mafic to felsic volcanic and volcaniclastic rocks. It has been shown that some of these "reefs" are duplicated because of folding, and thus give a mirror-image repetition.

Narayanaswami <u>et al</u>., 1960, Econ. Geol., Vol. 55

TABLE 10 Geological Succession of rocks within Kolar Amphibolite Series



Late Precambrian Basic Intrusives	9. Dykes of dolerite, basalt and gabbro	Consist of (1) non-porphyritic dolerite, (2) porphyritic olivine-bearing dolerite, (3) porphyritic and non-porphyritic basalt or trap, and (4) gabbro.
Post-Dharwar Intrusives	8. Pegmatite and vein-quartz	Coarse-grained quartzo-felspathic pegmatite with muscovite, tourmaline, and garnet; veins of white quartz and calcite that cut gold-bearing lodes.
	7. Gold-bearing quartz lodes and gold-bearing quartz-sulfide lodes (West Reefs)	Gold-bearing quartz and sulfide lodes composed of vein-quartz and quartz vein-zones made up of parallel, *en echelon* veins. Stringers and lenses of bluish-gray to white, highly sheared quartz, traversing highly sheared, schistose, recrystallized and mineralized wall-rock, containing silicates, metallic oxides and sulfides; Sulfide-bearing lodes contain a much greater proportion of sulfides, chiefly pyrrhotite and arsenopyrite; Gold present in quartz, sulfides and silicates; Some of the western sulfide lodes are graphitic and locally grade into graphitic lodes.
	6. Fibrous amphibolite (meta-pyroxenite? or sheared granular amphibolite?)	Medium to coarse grained, green, fibrous and tufted amphibolite composed chiefly of actinolite; commonly highly schistose and locally granular looking; forms lenticular zones along the western side of granular amphibolite beds.
	and/or	
Lower Dharwar "Kolar Amphibolite Series"	5. Granular amphibolite (metadolerite and meta-gabbro)	Fine to medium grained, granular-looking, dark amphibolite (resembling hornblende granulite) and coarse grained, somewhat schistose, dark amphibolite (resembling hornblendite)
	4. Massive amphibolite (massive meta-basalt)	Fine grained, compact, dark, massive amphibolite showing pillow structure.
	3. Schistose and fissile amphibolite (schistose and fissile, amygdular meta-basalt)	A complex banded assemblage of dark-colored, fine-grained and compact amphibolite, alternating with schistose and fissile amphibolite, hornblende schist and white-spotted (amygdular) amphibolite, representing meta-basalts of varying composition, alternating with amygdular flows and vesicular brecciated tops of lavas; commonly show pillow structure; some may be variolitic meta-basalts.
	2. Banded ferruginous quartzite and quartz-magnetite schist	Finely banded and drag-folded ferruginous quartzite composed of recrystallized cherty quartzite and brown ferruginous layers; also thin *en echelon* bands of quartz-magnetite schists containing grunerite along the western contact of schist belt.
	1. Champion gneiss and "autoclastic" conglomerate	Fine to medium grained, banded, streaky, quartzo-feldspathic gneiss and micaceous schist; blastoporphyritic streaky gneiss containing feldspar augens, tourmaline, and calcite also common; quartz-muscovite schist and coarse feldspathic muscovite schist along the western margin; zones of "pinch and swell" granite and "autoclastic conglomerate" containing sheared pebbles and boulders (boudins?) of granite, pegmatite, gneiss, amphibolite and ferruginous quartzite common along eastern margin of schist belt.

Narayanaswami et al., 1960, Econ. Geol., Vol. 55

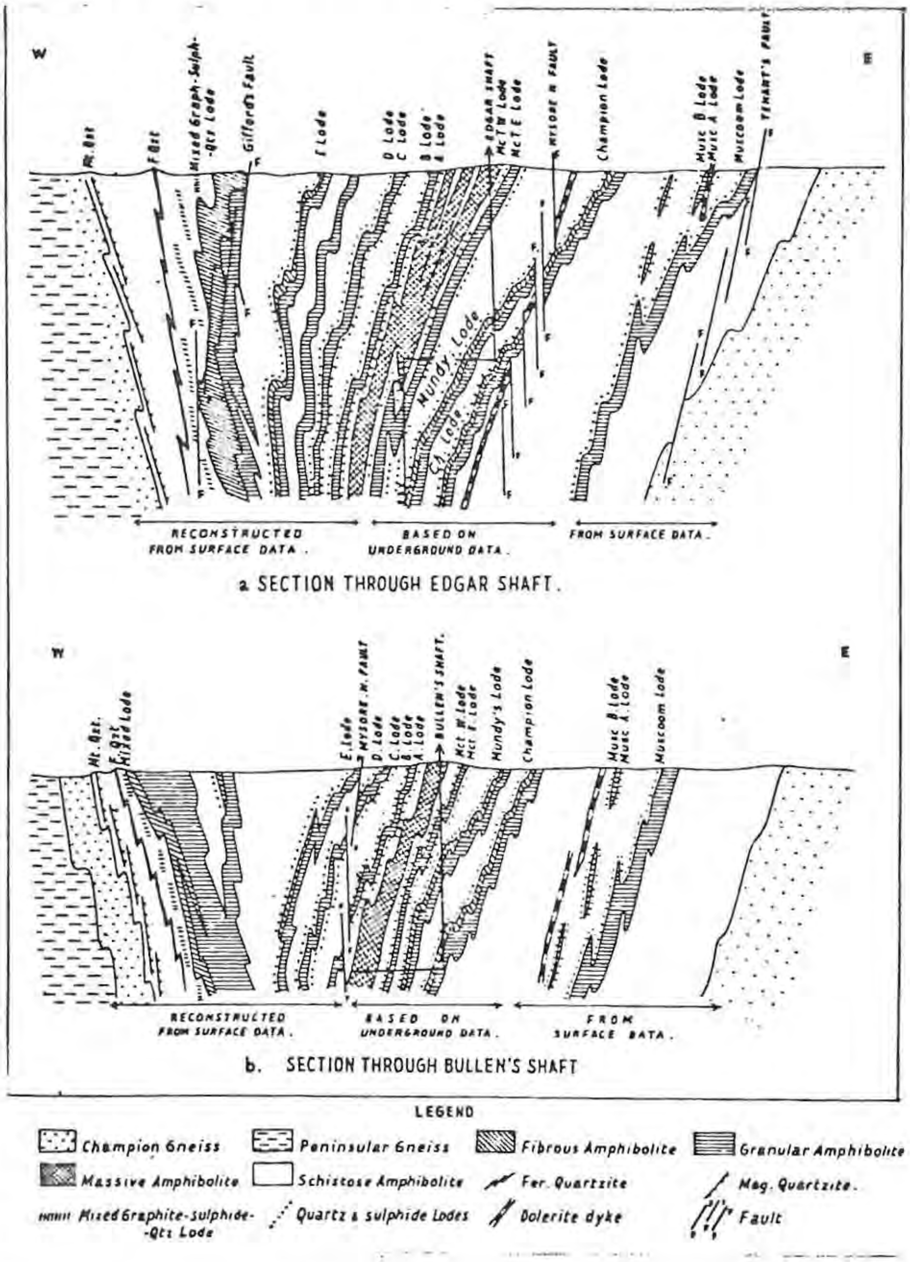


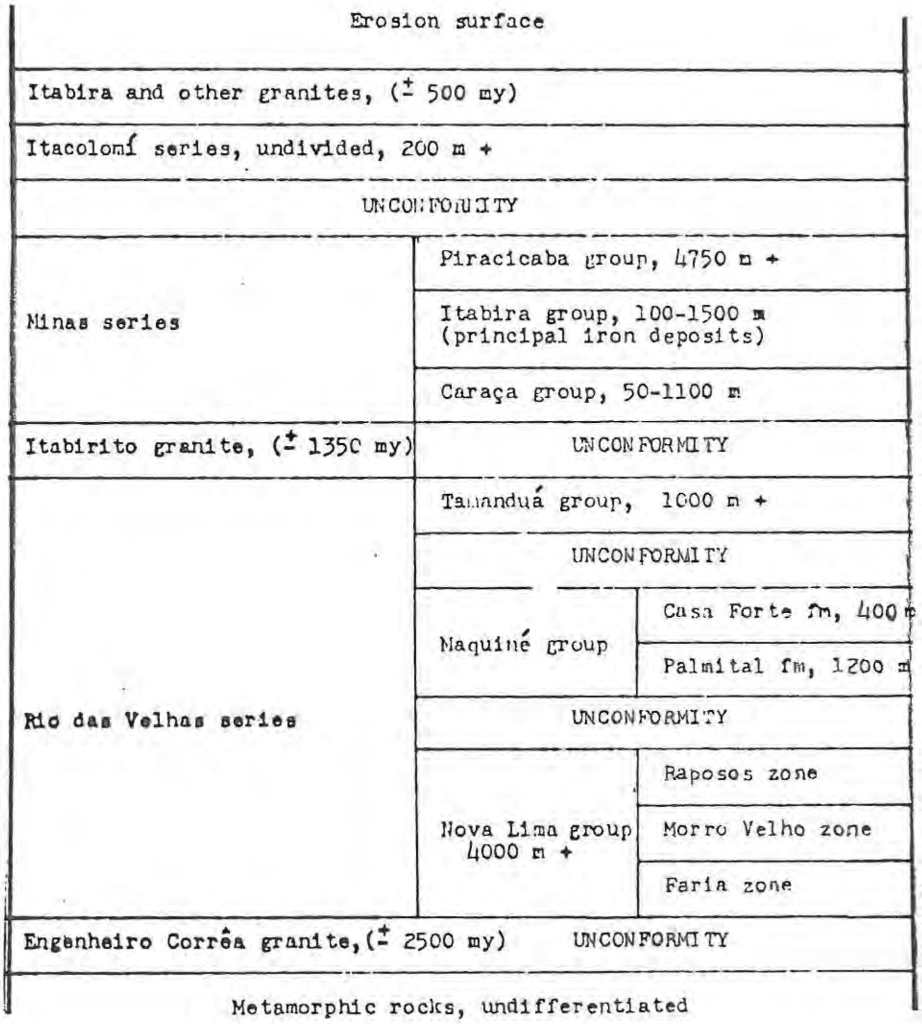
Fig. 32 Geological Cross-section across Kolar Mining Belt. (Showing stratigraphic control of lodes)

Raposos Gold Mine, Brazil

At the Raposos gold mine in Brazil (Tolbert, 1964; Gair, 1962), the principle host rock for the gold is mixed carbonate-sulphide facies iron-formation, although oxide facies and sulphide facies occur in the mine area as well. The main rock types in the Raposos Zone (see Table 11) apart from the iron-formations, are chloritic schists and phyllites with subordinate carbonate-sericite schists, graphitic schists, dolerite dykes, amphibole-garnet hornfelses and quartzite. The iron-formations occur as two distinct belts about 150-200 m apart. Each belt contains two or three parallel iron-formation horizons which are continuous over several hundred meters although they pinch and swell locally. A number of varieties of chlorite schist envelop the iron-formations (Figure 33).

Tolbert, 1964, Econ. Geol., Vol. 59

TABLE 11 Generalized Precambrian stratigraphy of the northern part of the Quadrilátero Ferrifero, Minas Gerais



<table>
<tr><td colspan="3">Erosion surface</td></tr>
<tr><td colspan="3">Itabira and other granites, (± 500 my)</td></tr>
<tr><td colspan="3">Itacolomí series, undivided, 200 m +</td></tr>
<tr><td colspan="3">UNCONFORMITY</td></tr>
<tr><td rowspan="3">Minas series</td><td colspan="2">Piracicaba group, 4750 m +</td></tr>
<tr><td colspan="2">Itabira group, 100-1500 m
(principal iron deposits)</td></tr>
<tr><td colspan="2">Caraça group, 50-1100 m</td></tr>
<tr><td>Itabirito granite, (± 1350 my)</td><td colspan="2">UNCONFORMITY</td></tr>
<tr><td rowspan="8">Rio das Velhas series</td><td colspan="2">Tamanduá group, 1000 m +</td></tr>
<tr><td colspan="2">UNCONFORMITY</td></tr>
<tr><td rowspan="2">Maquiné group</td><td>Casa Forte fm, 400 m</td></tr>
<tr><td>Palmital fm, 1200 m</td></tr>
<tr><td colspan="2">UNCONFORMITY</td></tr>
<tr><td rowspan="3">Nova Lima group
4000 m +</td><td>Raposos zone</td></tr>
<tr><td>Morro Velho zone</td></tr>
<tr><td>Faria zone</td></tr>
<tr><td>Engenheiro Corrêa granite, (± 2500 my)</td><td colspan="2">UNCONFORMITY</td></tr>
<tr><td colspan="3">Metamorphic rocks, undifferentiated</td></tr>
</table>

Tolbert, 1964, Econ. Geol., Vol. 59

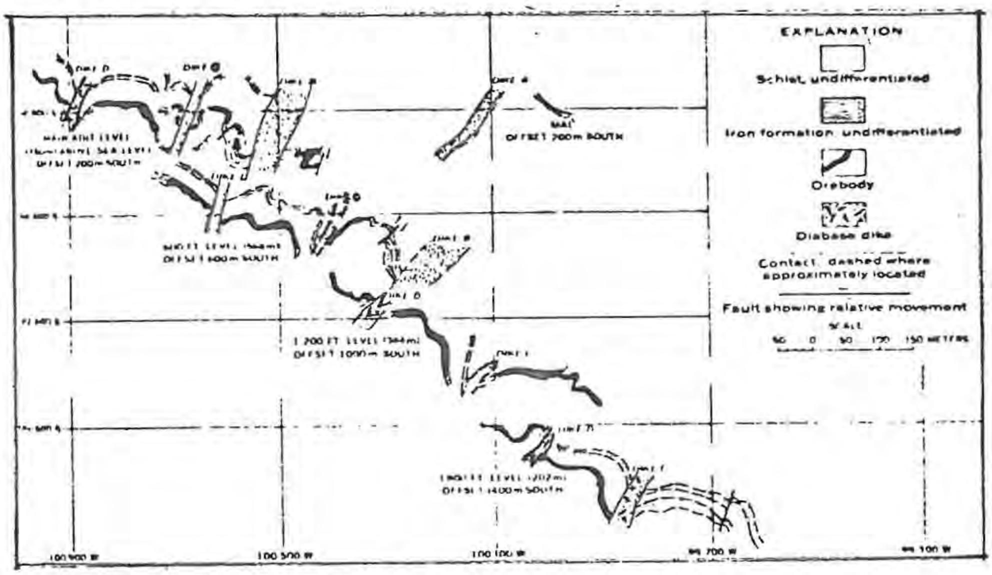


Fig. 33 Composite plan of selected levels in the Raposos mine, showing the configuration of orebodies at various depths and their relation to the dykes. Levels are offset to the south to avoid superposition.

The beds of iron-formation average three to five meters in thickness, but the range is from less than one meter to more than 30 m. The layers in the iron-formation are 5 - 15 mm in thickness. The main sulphide minerals present, in order of abundance, are pyrite, pyrrhotite and arsenopyrite. Chalcopyrite is rare. Microscopically visible gold was found to be associated more often with the pyrrhotite than with any other sulphide minerals.

Where the pyrrhotite had been deformed caticlastically gold was seen on the contacts between pyrrhotite and other sulphide minerals.

Morro Velho Mine, Brazil

The Morro Velho mine (Gair, 1962) like the Raposos mine is in the Nova Lima Group of the Ria Das Valhas Series of Minas Gerais, Brazil. Because of the lack of marker beds and the depth of weathering very little detail of the regional geology could be descerned. However, some drilling and mapping done in the Gaia tunnel indicated that metavolcanics are interbedded with metasediments. In the tunnel, mapping showed that the metavolcanics were interlayered with carbonate and oxide facies iron-formation.

The ore bodies of the Morro Velho mine are in carbonate facies iron-formation (lapa sêca). There is one major iron-formation horizon and several other thinner ones. Gair (1962) says it is possible that only one iron-formation horizon is present but that it has been structurally deformed to produce the pattern. The ore consists of sulphides with gold and silver in a gangue of mainly quartz and dolomite or ankerite. In some places quartz, carbonate and sulphides fill tension gashes that cross the layering of the iron-formation. The principle sulphides, in order of abundance, are phyrrhotite, arsenopyrite, pyrite and chalcopyrite.

Hill 50 Mine, Western Australia

The Australian Archaean shield contains numerous gold deposits in iron-formation (jaspilite) hosts (Woodall, 1975; Baxter, 1975; Campbell, 1965; Williamson and Barr, 1965). The Hill 50 mine in the Murchison Goldfield in the north east part of the Yilgarn Block is in rocks typical of the Kalgoorlie group (Lewis, 1965).

The Hill 50 mine is just north of the town of Mt. Magnet and the ore is hosted by the Boogardie group of greenstones and iron-formations. The stratigraphy of the Boogardie group has been investigated by Foreman (1960). He has divided the rocks in the Boogardie area into two units: the Lennonville Beds, overlain by the Boogardie Group. The Lennonville beds consist of 800-2500 m of intercalated schists and iron-formations, with a conglomeratic bed near the base.

The Boogardie Group (Figure 34) is a sequence of iron-formations and greenstones of Archaean age composed of nine recognisable units. Foreman (1960) suggests that the stratigraphy in the area be divided accordingly.

The lowest sequence of rocks is the Poverty Flat formation. This consists of a sequence of magnetite-sericite-quartz-chlorite schists which may or may not be carbonated. Interbedded within this sequence are four or five thin (0,1-0,3 m) iron-formation horizons.

The Jupiter Jaspilite is a 1,5 to 10 m thick iron-formation sequence containing minor schistosé lenses. This formation supported gold mining operations at the Jupiter deposit.

The next sequence is the Mt. Magnet Greenstones. This 100-220 m thick sequence consists of banded magnetite-plagioclase-carbonate-chlorite rock and quartz-carbonate-sericite-chlorite rock with minor interbedded iron-formations.

Foreman, 1960, J.R. Soc. West. Aust., Vol. 43

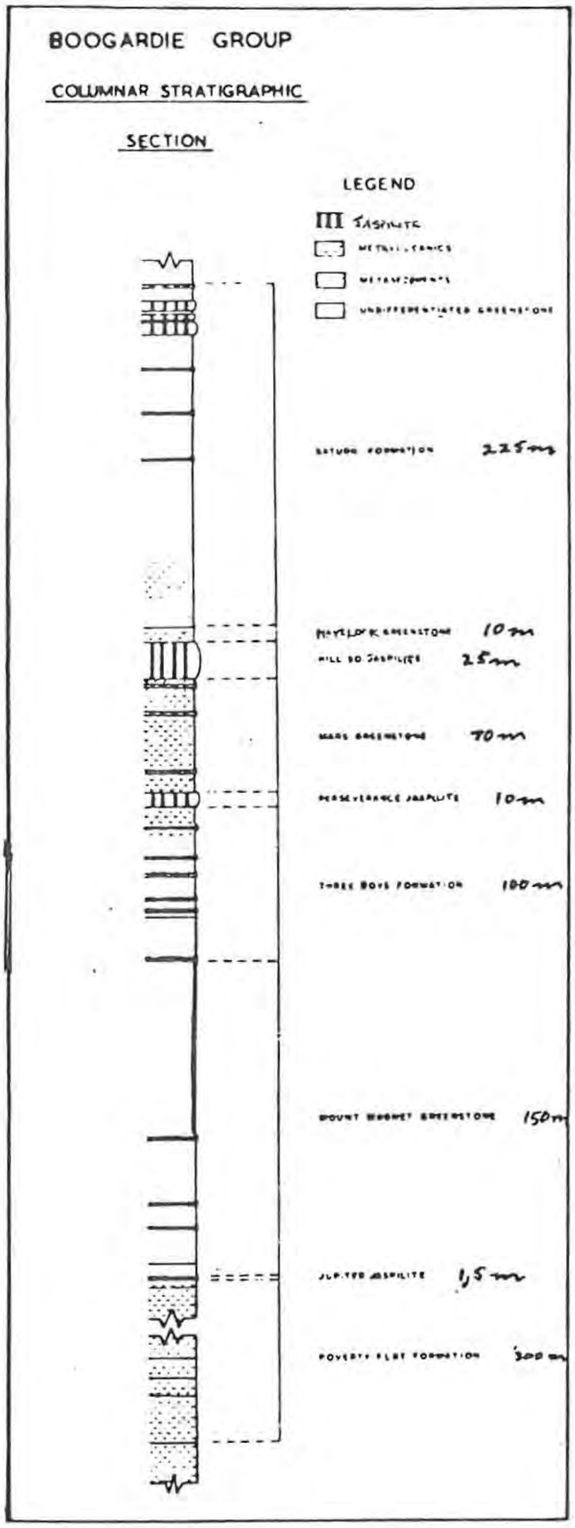


Fig. 34

The Three Boys Formation is 80-100 m thick. It comprises carbonated quartz-sericite-magnetite schists which may or may not have plagioclase. Interbedded with these schists are seven iron-formation horizons each between 0,3 and 3 m thick.

The Perserverence Jaspilite is about 6-12 m thick.

The Mars Greenstones are a 40-80 m thick sequence of quartz-chlorite schists, interbedded with iron-formations and chloritized quartz dolerite.

The Hill 50 Jaspilite is about 15-25 m thick. It hosts the Hill 50 gold mine.

The Havelock Greenstone is a chloritized quartz dolerite/basalt underlying a 0,3 m thick iron-formation. It is about 5 - 13 m thick.

The top sequence of the Boogardie Group is the Saturn Formation. This is a thick (225 m) sequence of chloritized basalts and seven interbedded iron-formation horizons. The top of the formation is marked by the uppermost iron-formation. The greenstones overlying the Boogardie Group are apparently devoid of iron-formation horizons.

Petrography

Lavas make up portions of the Mars Greenstone sequence and the Mt. Magnet greenstone sequence. The Havelock Greenstone unit consists almost entirely of basaltic flows. The lavas otherwise are a mixture of basic to intermediate rocks. The rocks are generally massive and have been wholly chloritized. Sedimentary greenstones make up most of the Poverty Flats formation and the upper half of the Mars Greenstones. Typical sedimentary structures (bedding, cross-bedding, ripple marks and scour-and-fill structures), have been found and used to identify the genesis of these beds. The rocks resemble a combination of chemical and fine grained volcaniclastic sediments.

The iron-formations consist of alternating bands of red chert (from hematite flakes) and hematite and/or magnetite. Bands that are carbonate rich are also to be found. In the ore horizons the bands are alternately carbonate- and sulphide-rich. The sulphides are mainly pyrrhotite with minor pyrite. The gold is consistently associated with the sulphides, being almost completely absent from areas lacking in them.

Some Auriferous Iron-Formations in North America

Sawkins and Rye (1974) investigated the Homestake deposit in South Dakota, U.S.A., and concluded that the gold was genetically related to the deposition of the host Homestake formation. The detailed stratigraphy of the area is shown below :

Sawkins and Rye (1974)

TABLE 12

Cambrian and later rocks

Precambrian intrusives	Gabbroic intrusives and some surface flows
Grizzly formation >1000 m	phyllites and meta-greywackes
Flag Rock formation >1500 m	phyllites, quartzites, meta-basalts, tuffs, pyroclastics, graphitic schists and phyllites - all interbedded
North Western formation 1300 m	acid schists and phyllites
Ellison formation 1000-1500 m	detrital quartzites and banded phyllites
Homestake formation >100 m	carbonate rich graphitic schists metamorphosed to cummingtonite quartz schists
Poorman formation >700 m	carbonate rich graphitic schists and phyllites

The ore bodies are restricted to a thin stratigraphic section of ferruginous carbonate rocks and their metamorphosed equivalents. They consist of chloritized portions containing short irregular veins and conformable masses of silica, pyrite and arsenopyrite.

Stable isotope studies (Rye and Rye, 1974) indicate that the gold and other constituents of the ore bodies are primary and have not been introduced. Localization of the ore in folds is probably due to migration during later deformation.

McConnell (1964) observed the similarity between the Homestake deposits and gold deposits in the Contwoyto Lake area, Northwest Territories, Canada. These deposits are confined to amphibole rich beds in a Precambrian sequence of phyllites, slates, siliceous tuffaceous horizons and iron-formation. McConnell (1964) suggests that amphibole rich zones are metamorphosed iron rich sediments similar to those in the Homestake deposits. The gold-bearing portions are generally restricted to heavy sulphide mineralized areas particularly where coarse arsenopyrite and fine banded pyrrhotite occur.

In the Abitibi Belt, Canada, Hutchinson et al. (1971) have shown a relationship between magmatic differentiation in volcanic complexes and the metallogeny of the area. The 'exhalite' iron-formations contain gold deposits as at Errington Township, Ontario. The suggested relationships between tectonism, magmatism and metallogeny are depicted in Figures 35 and 36.

Hutchinson et al., 1971, CIM Transactions, Vol. 74

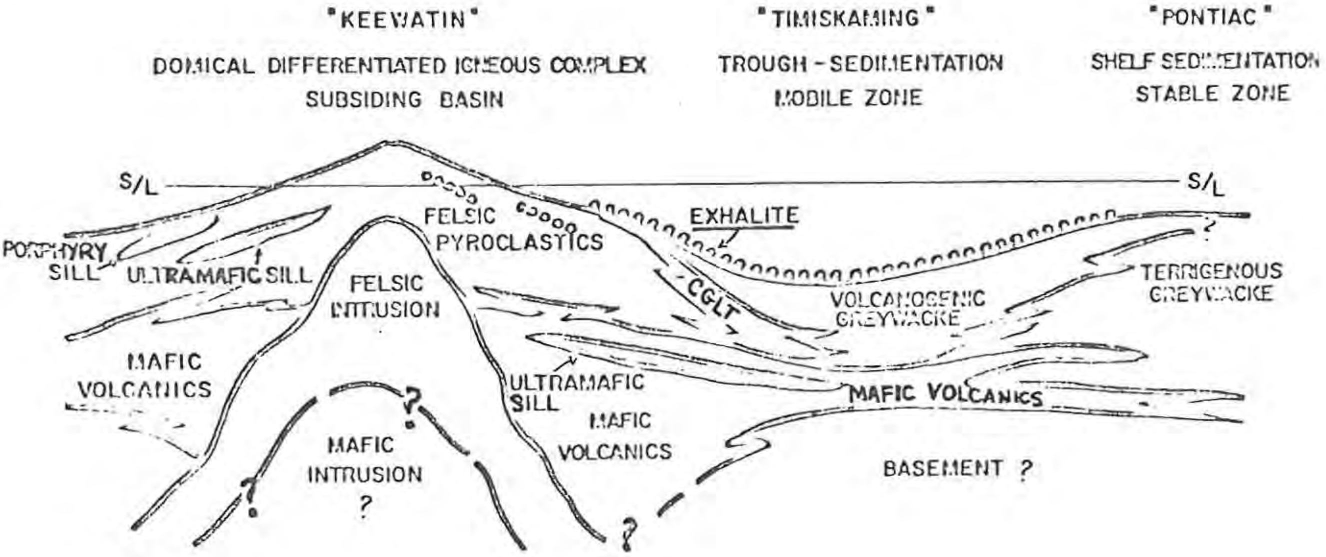


Fig. 35 Archaean tectonic-stratigraphic relations

Hutchinson et al., 1971, CIM Transactions, Vol. 71

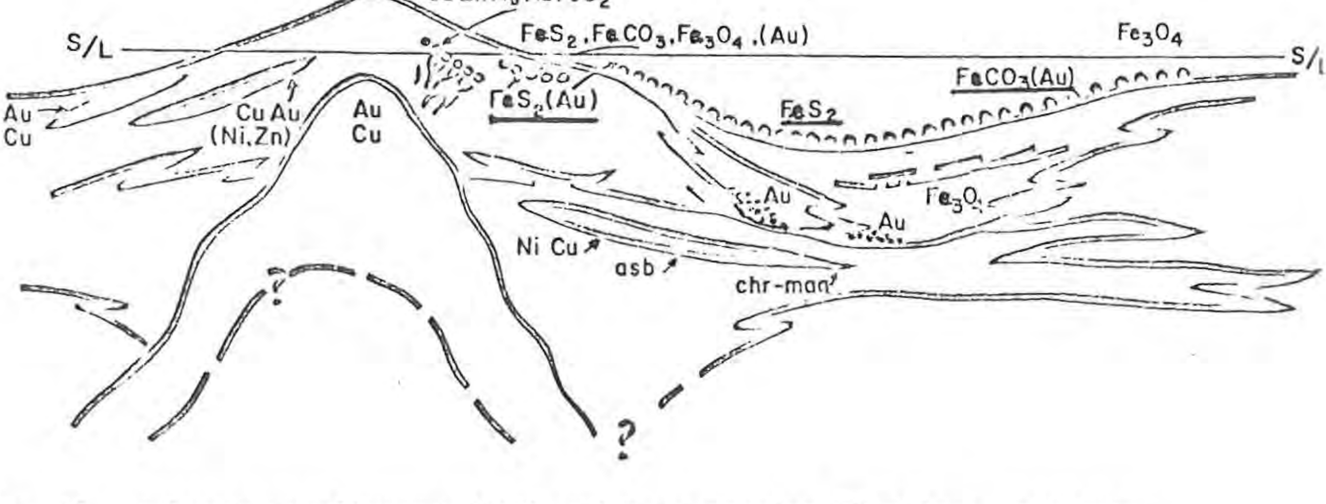


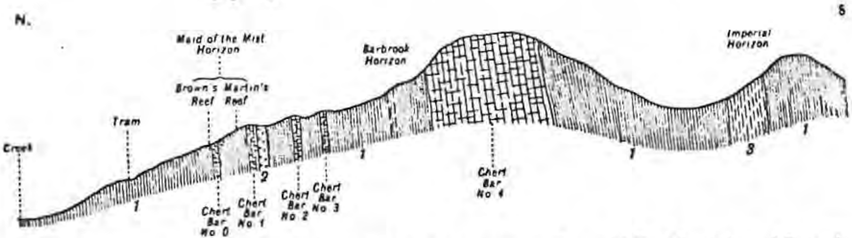
Fig. 36 Archaean metallogeny (full width of section 50 miles; maximum vertical thickness, 10 miles)

Hutchinson et al. (1971) suggest that like the related massive sulphide deposits in the volcanic complexes, the auriferous iron-formations are primary volcano-exhalative deposits. This suggestion is quite contrary to most earlier workers. The origin of the gold associated with chert and carbonate horizons was attributed by them to the selective replacement of iron oxide layers by auriferous iron sulphides resulting in the "impregnations" and replacement deposits.

South African Deposits

In South Africa the Barbrook - Maid of the Mist gold belt in the Barberton gold mining district contains several ferruginous chert or iron-formation ('calico rock') horizons associated with gold deposits (Figure 36) (Hall, 1918). The country rocks are a variety of slates and shales.

Hall, 1918, Geol. Surv. South Africa, Mem. No. 9



1. Slates, etc. 2. Basic sill. 3. Soft slaty ferruginous rocks. N.B. Martin's and Brown's Reef belong the workings of the Louws Creek G.M. Co. (French Bobs).

Fig. 36 Section across the Barbrook-Maid of the Mist Gold Belt showing the approximate horizons of the Barbrook, Imperial, French Bobs, and Maid of the Mist Mines.

From the Barbrook Mine, Hall (1918) describes what is obviously a gossan : " .... a very remarkable weathered rock resembling a skeleton cast, full of rudely rectangular cavities, defined by thin walls of quartz ..." (p. 282). This indicates that the rock was probably a weathered sulfide facies iron-formation.

"Black Reefs" or mineralized iron-formations are described by Mendelssohn (1939) from the Weigel mine in the Murchison area of north east Transvaal. The country rock is a series of carbonate-rich chlorite-sericite schists which were probably originally tuffs. The iron-formation changes along strike to the west, to a ferruginous quartzite. The mineralized portion of the "reef" varies from a "stringer" to a zone nearly 2,5 m in width. At its greatest development the orebody consists of roughly parallel bands of quartz with sulphide bands separated by mineralized country rock. The contacts are gradational - the iron-formation becoming less sulphide rich gradually. The iron-formation is described by Mendelssohn (1939) as a quartz-carbonate rock very similar to the "iron-stones" in the Lake Superior area. The main sulphide present is arsenopyrite with minor pyrite. Minute inclusions of rutile were described by Mendelssohn (1939). The gold is associated with the arsenopyrite, and Mendelssohn (1939) suggested that it was replacing the earlier arsenopyrite. Gold is also associated with the very minor amounts of chalcopyrite present.

STRUCTURAL CONTROLS OF GOLD DEPOSITS IN BANDED IRON-FORMATION

Three factors must be considered when examining the structural control of gold mineralization in Archaean banded iron-formation. These are :

(i) syn-depositional controls of mineralization,

(ii) the style of the structural deformation associated with Archaean rocks, and

(iii) the competence of the iron-formation relative to the enclosing rocks and how the iron-formation will behave when deformed.

Primary Controls of Mineralization

Syngenetic gold mineralization in Archaean iron-formation is associated with sulphide facies horizons, and to a greater or lesser extent with mixed sulphide-carbonate facies horizons. The development of these facies types depended on the physico-chemical conditions prevailing at the time of deposition. Sulphide facies horizons developed in deeper water areas, while the other facies types developed in shallower water (Goodwin, 1973). Therefore, the original shape and structure of the depository floor defined the sites of deposition of the auriferous iron-formation.

Imposed Structural Controls

The structural deformation often associated with Archaean rocks can usually be divided into two styles. The first style of deformation resulted in typically steep-sided almost isoclinal synformal structures. There may have been up to three periods of deformation but the folds are generally co-axial. This makes it difficult to separate the individual periods on a regional scale. The regional greenschist metamorphism of these areas was imposed at this time.

The second style of deformation seen in these areas resulted from the vertical emplacement of younger granitic plutons. This caused the rather linear greenstone belts to become "wrapped around" the plutons to give the arcuate forms to these belts (Macgregor, 1951). These forms are well displayed in Canada (Figure 14) and Rhodesia (Figure 15).

On a more local scale, these two different styles of deformation have affected the iron-formations rather differently.

Mineralization Controlled by Folding

The earlier periods of deformation resulted in the originally horizontal or sub-horizontal horizons becoming steeply inclined. Minor parasitic folds on the limbs of the synforms are common and well displayed by the iron-formation and chert horizons. Stretching on the limbs of these parasitic folds has resulted in boudins being developed. Fripp (1974) has described these features at the Vubachikwe mine. Similar features can be seen at the Crown-Lennox mine near Mashaba, Rhodesia. At these deposits the low pressure zones between the boudins contain later hydrothermal quartz, which is usually barren. Minor amounts of the syngenetic mineralization may have been remobilized, but it is only when the quartz veins brought their own gold that they are mineralized well enough to warrant serious investigation.

Further evidence of stretching on the limbs of folds can be seen on the contacts between the iron-formation and the enclosing rocks. Fripp (1974) describes shearing and faulting that occurs on the contact between the iron-formation and the greenstones at Vubachikwe mine. At the Crown-Lennox mine similar features can be seen. There the contact is marked by a narrow (up to 10 cm wide) zone of fine grained strongly sheared greenstone.

In the hinge zones of the folds the iron-formation horizons are often thicker. This is probably a tectonic feature which has resulted from flow folding. It is also noticeable that the hinge zones often coincide with strong sulphide mineralization.

At the Raposos Mine (Tolbert, 1964) (Figure 38), there is a system of eastwardly plunging assymmetrical isoclinal folds. The axial planes strike north to north-east and their dip varies from east to south-east. The fold axes plunge at about right-angles to the strike, to the east or south-east. The angle of plunge varies from 45° at the surface to about 30° underground. The strike of the axial planes changes from north on the surface to north-east on the lower levels.

The Copperhead mine in the Southern-Cross Bullfinch Belt, Yilgarn goldfield (Clappison and Zani, 1953) also shows this relationship between fold hinge zones and sulphide mineralization in banded iron-formation. There are two favourable positions for ore (Figures 39 and 40):

Tolbert, 1964, Econ. Geol., Vol. 59

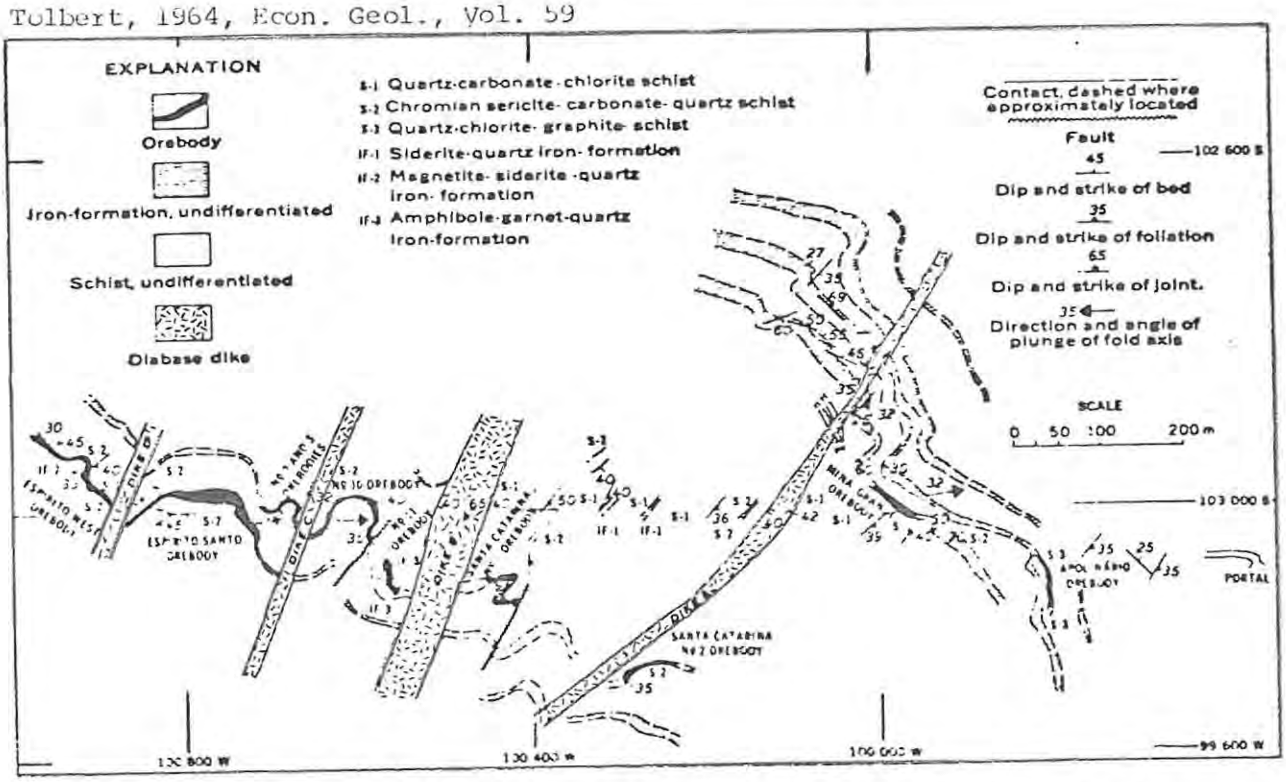


Fig. 38 Geology map of the main abit level, Raposos mine. Geological mapping by A.F. Matheson, 1953. G.E. Tolbert, J.D. Hagen, 1958.

(i) in the main isoclinal anticline (A 10) and its limbs. This is most noticeable on the lower of the two iron-formation horizons, and

(ii) in the smaller contorted folds on the east limb of the A 10 anticline. Ore in these "traps" is found in both the upper and lower iron-formations.

Clappison and Zani (1953) suggest that collapse of the folds seem necessary for the occurrence of ore. Numerous smaller ore shoots are found in favourable structures near the collapsed folds. It appears that the gold bearing sulphide mineralization has been remobilized into the low-pressure zones in these collapsed folds and nearby related structures.

In the Norseman area of the Dandas goldfield, Western Australia, the gold rich lodes occur in folded or brecciated structures within the iron-formations (Hall and Becker, 1965). The mineralization is best developed along the crests of parasitic folds where it forms "pipes" of pyritic ore which follow the plunge of the fold crests. A good example of this localization of mineralization is to be found at the Lady Miller lode (O'Driscoll, 1953). Here the ore is in the footwall of the iron-formation. Major concentrations of pyritic ore are found as pipe-shaped bodies in the crest of "en echelon drag folds".

Clappison and Zani, 1953, in Edwards, D.B. (ed.) Geology of Australian Ore Deposits.

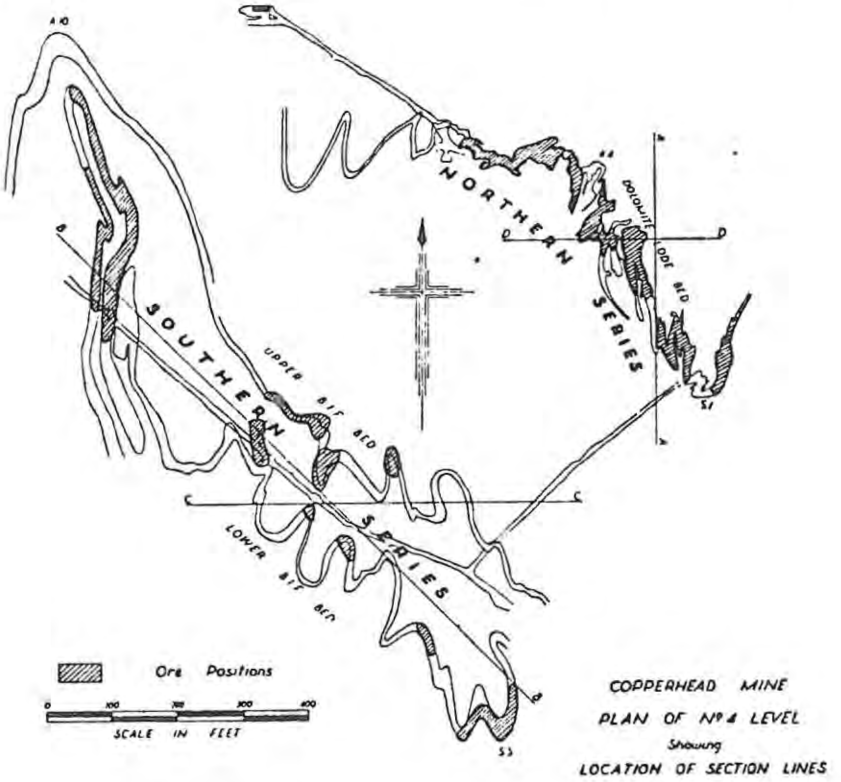


Fig. 39 Plan of the 410-ft level of the Copperhead Mine

Clappison and Zani, 1953, in Edwards, D.B. (ed.), Geology of Australian Ore Deposits

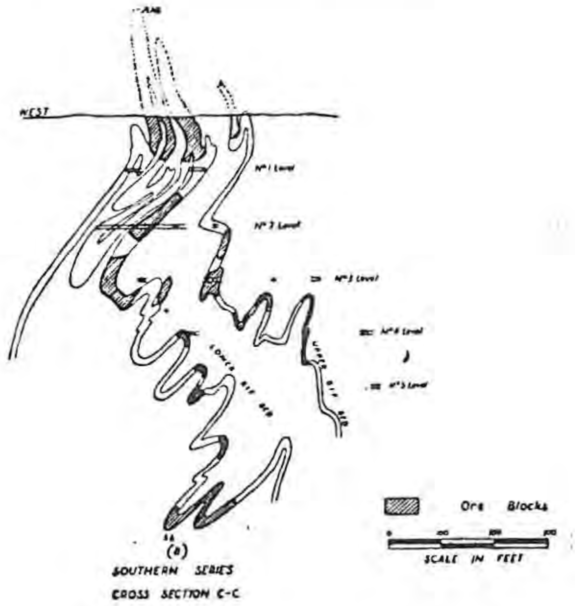


Fig. 40 East-West Sections through the Copperhead Mine, along the line CC of Fig. 39.

At the Crown-Lennox mine strong evidence of remobilization of the auriferous pyrrhotite into a low-pressure zone can be seen in the north part of A reef, Lennox Section (Figure 41). Here the folding is a result of dragging against a fault with a strong sinistral tear component. In the hinge zone of the fold, massive pyrrhotite is associated with large clots and thin stringers of quartz. There is no regular layering as is typical of the iron-formation away from the fault, although in places it can still be recognised. There is no evidence of quartz and pyrrhotite having been introduced along the fault. On the outer side of the fold, minor structures containing quartz go from the iron-formation to the greenstones in the hangingwall. These can be seen to stop less than a meter from this contact. It is unlikely that these minor quartz lenses were the 'feeders' for this massive pyrrhotite mineralization.

It is probable that the remobilization of the sulphide material is a relatively local phenomenon, but there is a definite tendency for the sulphide material to be concentrated in the regions near folds. This may be due to the competence differences between the rocks of the different facies types. The point at which the initial folding is most likely to occur, is at the junction of the competent sulphide facies material and the less competent rock of other facies types. Continued deformation would remobilize the sulphides into the low-pressure dilatant zones in the fold closures.

Mineralization Controlled by Faulting

Faults are often to be found displacing iron-formation horizons. Epigenetic gold bearing quartz veins have sometimes been emplaced in these openings. They seldom affect the primary mineralization of the iron-formation, but can be an added 'bonus' to the ore reserves of the deposit. At the Crown-Lennox mine the original workings were started on such a quartz vein. Later, the miners found that the gold in the iron-formation was more consistent, so it was exploited instead.

The deformations causing these faults have been discussed relative to any associated mineralisation by Anhaeusser (1976). As the mineralization associated with them is epigenetic, they will be discussed no further here.

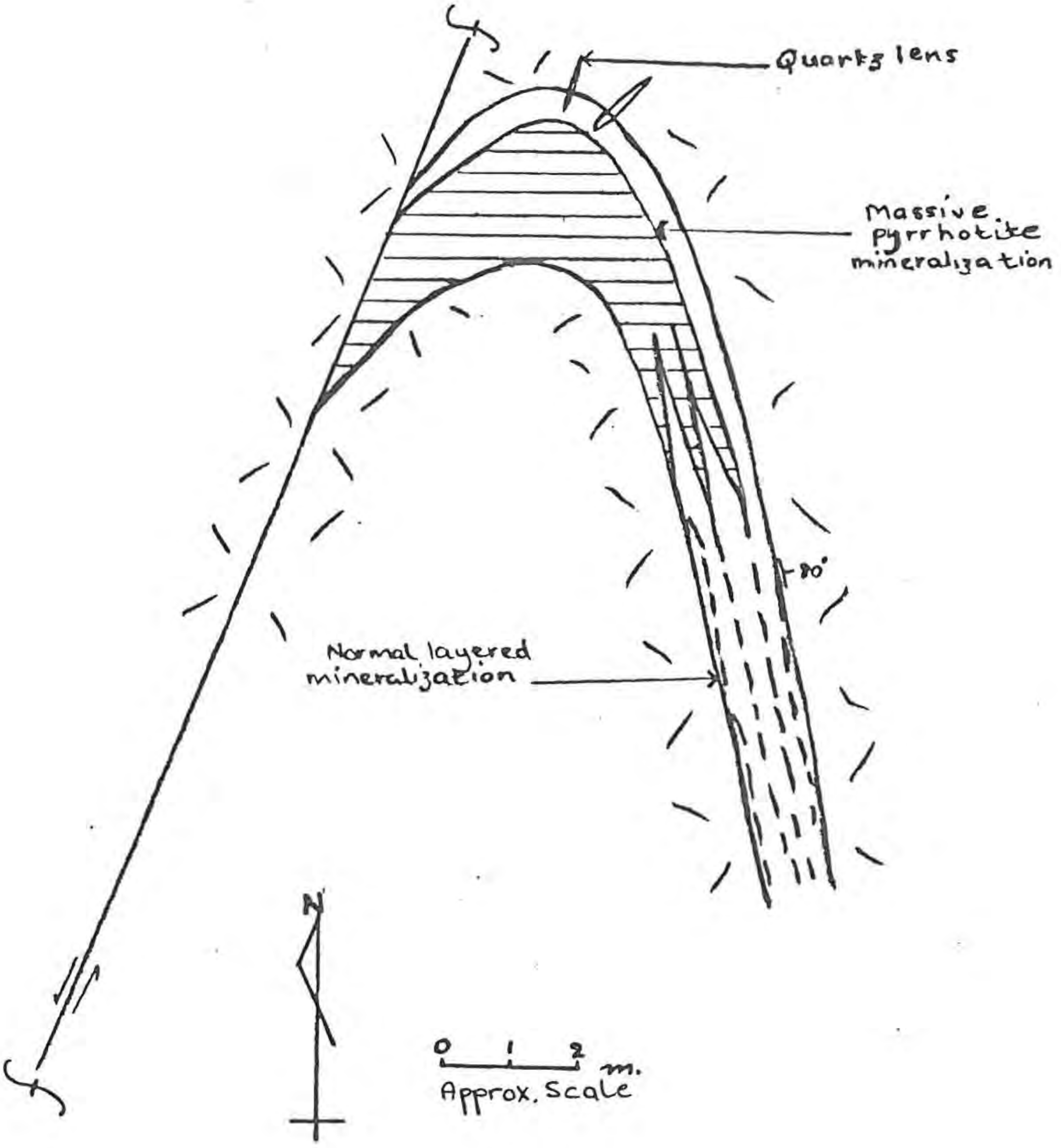


Fig. 41 Sketch map of geology on 2 level, Lennox Section, Crown-Lennox mine, showing massive sulphide mineralization in hinge zone of drag fold.

Exploration and Exploitation of Archaean Iron-Formation Hosted Gold Deposits

Exploration

Several geological factors provide parameters in the exploration for gold deposits hosted by Archaean iron-formation :

(i) the tendency of iron-formations to form topographically high features,

(ii) the stratigraphic control and associations of the host iron-formation,

(iii) the association of the gold with other elements, especially base metals.

(i) The silica in the iron-formation, because of its relative hardness and chemically inert nature, resists erosion to a greater degree than the surrounding rocks. Thus ridges of iron-formation, which can be traced for great distances, are formed. These ridges of iron-formation have been great aids in unravelling the complex structures of some Archaean terrains (Tomich, 1978; MacLeod, 1965).

(ii) The iron-formations are chemical sediments precipitated during breaks in volcanism. There are two major lithological associations of gold bearing iron-formations and cherts :

(a) the ultramafic-mafic association, and
(b) the mafic-felsic association.

In the ultra-mafic-basic association there is no strong evidence of cyclic volcanicity, with respect to chemical or mineralogical changes. The iron-formations usually overlie tuffaceous deposits (e.g. Fripp, 1974) suggesting that the exhalative volcanicity resulting in the iron-formation was preceded by a more violent episode of volcanism. The strong association of carbonate rocks with these deposits cannot be too strongly emphasized. The carbonatization is thought by several workers (e.g. Tomich, 1976; Karvinen, 1978) to be the result of carbon-dioxide rich emanations of a volcanic source. Thus carbonate facies iron-formation may be precipitated, and the surrounding rocks may also be carbonated.

In the field volcanic rocks with iron-carbonate will weather brown, while those without iron-carbonate will weather dark green or black. In this way the carbonated horizons can be distinguished.

In fresh exposures, either underground or in diamond-drill core it may be necessary to use a simple staining technique to identify the carbonated horizons. Several examples of stratabound gold deposits associated with carbonated volcanic rocks have been described.

Fripp (1974) mentions that along strike from the iron-formations in unit C at Vubachikwe, gold occurs with carbonated extrusives.

In the Southern Cross-Bullfinch belt of the Yilgarn goldfield, Western Australia, Clappison and Zani (1953) have described two similar occurrences. At the Copperhead mine, the northern series of deposits is in a dolomitic quartz greenstone. Higher ore grades are associated with strong dolomitization. These deposits are on the same stratigraphic horizon as the southern series of deposits which are typical iron-formation hosted sulphide-gold deposits. The Southern Cross mines have similar features. The Frasers group of deposits are disseminations in a dolomitic tuffaceous horizon. At the same stratigraphic position, but along strike, are the Three Boys group of deposits which are in iron-formation.

Where the iron-formations are associated with the mafic to felsic igneous cycles, the iron-formation forms the topmost layer in the sequence (Spence, 1975; Spence and de Rosen-Spence, 1975). Immediately below it are volcanogenic arkosic sediments and rhyolites. Immediately overlying it is the mafic rock at the base of the next volcanic cycle. Spence and de Rosen-Spence (1975) have described how all the base metal volcanogenic sulphide deposits are all located on or near one stratigraphic horizon. This situation appears to prevail elsewhere in Archaean terrains.

Gold mineralization in iron-formations is always associated with sulphide or mixed sulphide facies iron-formations. The sulphides can be identified in the weathered outcrop as gossan (Hall, 1918).

All these features can be determined in detailed mapping. Modern technological aids to exploration which use the geological and geochemical features of iron-formation deposits can also be used.

Geochemistry is probably the easiest exploration programme to carry out in the field, as it needs no special or expensive field equipment. For techniques it is suggested that authoritative works on the subject be consulted (e.g. Hawkes and Webb, 1962). It is not necessary to confine the geochemical programme to gold alone. Because of the association of these deposits with other elements (especially base metals) they can be used as 'indicator' elements. Typical elements that can be used include : Ag, As, Ba, Bi, Co, Cu, Mn, Pb, S, Sb, Zn.

Geophysics can be a major technical aid in the search for gold deposits in banded iron-formations. Magnetometer surveys can be used to distinguish magnetite rich oxide facies iron-formation from the other facies types, when they are covered by over-burden.

Electrical geophysical techniques can be used if the style of sulphide mineralization is known. Where the mineralization is 'massive' electromagnetic (E.M.) systems can be used. If the mineralization is relatively sparse or disseminated induced-polarization (I.P.) systems should be employed.

Evaluation

This is perhaps the most important phase in the life of a mine. It is necessary to have all the relevant geological facts at ones disposal before one can make the correct decision as to whether or not a mineral deposit is to be mined.

Important factors to be considered with respect to gold deposits include :

(i) the distribution of the gold mineralization in the iron-formation horizon,

(ii) whether or not the mineralization is economical below the oxide and secondary enrichment zone, and

(iii) detailed mineralogy of the ore zone.

(i) The distribution of sulphide facies horizons within an iron-formation bed must be ascertained. The sulphide facies horizon may consist of a series of lenses distributed irregularly through a much thicker iron-formation bed. This creates problems in mine layout and development. The stopes would be small, and irregularly shaped and spaced. The sulphide facies horizons may however be continuous on either the hangingwall or footwall portions of the iron-formation, or even make up the entire unit in places. The sulphide rich horizon could even be a combination of all these. At the Crown-Lennox mine, Rhodesia, examples of all these instances are to be found.

The structural controls of the mineralization must be defined as far as possible. Parasitic folds or slight "rolling" of the ore body may show directions of preferential mineralization. Soft sediment slumping can indicate small basins where sulphides may have accumulated. It is imperative that both primary and secondary or superimposed structures on all scales be noted.

The future potential of the area should be taken into account. If the iron-formation horizon is small, with no others present nearby, then it is unlikely to have much potential at depth or along strike.

(ii) Secondary enrichment in the ore zones above the water table can be misleading. Many ancient workings and old prospects started on these zones of supergene enrichment, but were abandoned when the gold values decreased with depth. The Iron Duke mine in Rhodesia was originally a gold mine, but now the pyrites is produced purely as a source of sulphur (Fripp, 1976a).

(iii) Detailed studies of the mineralogy must be made to aid testing of the ore before a plant can be designed. In this context a true representative of the ore should be studied. If the ore varies, then the various types should be tested and analysed separately, and not as a collective sample. Further details are discussed later.

Exploitation

The mining of these deposits is often controlled, and can be aided by their structural and mineralogical nature. The thickness of the beds and their attitude, which is often near vertical, are two major controls of the mining methods. The steep dip aids mining in that the broken ore in a stope does not have to be mechanically or manually moved to draw-points. Intelligent mine planning based on adequate knowledge of the geological controls of the body allows the broken rock to be gravity fed to draw-points in the footwall of the stope. The narrowness of the auriferous horizons has two major effects :

(i) it is unlikely that the mine will become mechanized to the extent of involving large volume trackless machinery, and

(ii) in narrower portions, for removal of the ore it may be necessary to incorporate low grade or barren material to allow adequate working space. A width of about one meter is usually a minimum. Therefore detailed knowledge of the distribution and nature of the ore material is necessary for ore reserve and grade estimations before mine layout and planning can be started.

The constant association of the gold with sulphide mineralization is also a factor that aids the mining and reduces the running costs. 'Visual sampling' can be carried out on a routine basis. Chemical assays need only be done as a check. In this way laboratory costs can be reduced to a minimum. Where the gold is strongly associated with a particular mineral in the sulphide facies (e.g. with arsenopyrite in mixed arsenopyrite and pyrrhotite), a stronger control on the day to day running of the mine can be maintained.

The contacts between the ore zones and the country rock are nearly always sharp. Therefore, the determination of the ore body contacts by chemical assay need not be done on a routine basis.

A definite tendency for auriferous sulphides to be associated with plunging parasitic fold hinges has been noted (e.g. Clappison and Zani, 1953). These folds often increase the thickness and grade of the ore, with leaner zones on the fold limbs. This results in a series of small pipe-like bodies along the pitching fold closures separated by zones of low-grade material. Therefore extensive development may have to be done in waste. This is expensive mining to a small worker. Alternatively, several shafts (one for each ore body) may have to be sunk.

Extraction

The treatment of gold ores from iron-formations is controlled by the fact that the gold is "locked" in the sulphide minerals (Fripp, 1974, 1976b) as small (up to 30μ) diameter particles. Very little free native gold is present except where later intrusive quartz veins have brought in new low-sulphide ore, or remobolized the syngenetic stratiform ore. Liebenberg (1972) has classified these ores as 'unoxidized complex sulphide ores'.

Henley (1975) does not describe Archaean auriferous iron-formations as a class. However he does discuss the treatment of gold ores where the gold is intimately associated with sulphides. For details on the treatment of these ores it is suggested that a standard text be referred to.

Important stages in the treatment of these sulphidic ores are : gravity concentration, amalgamation, flotation and cyanidation (Figure 42) (Henley, 1975).

It is necessary to make a detailed study of the gold ore in order that the most efficient and economical method of extraction be determined before the plant layout is designed. Details that must be investigated include (Henley, 1975) :

(i) the size of the gold granules,

(ii) the identity of the gold bearing minerals (if any) and the ganque,

(iii) the sites of the gold granules in the ore material,

(iv) the surface textures of the gold granules,

(v) the presence and type of coatings on the surface of the gold granules.

In the case of the sulphidic ores of iron-formations where the gold is fine-grained the plant circuit should include fine grinding, flotation, roasting, and cyanidation.

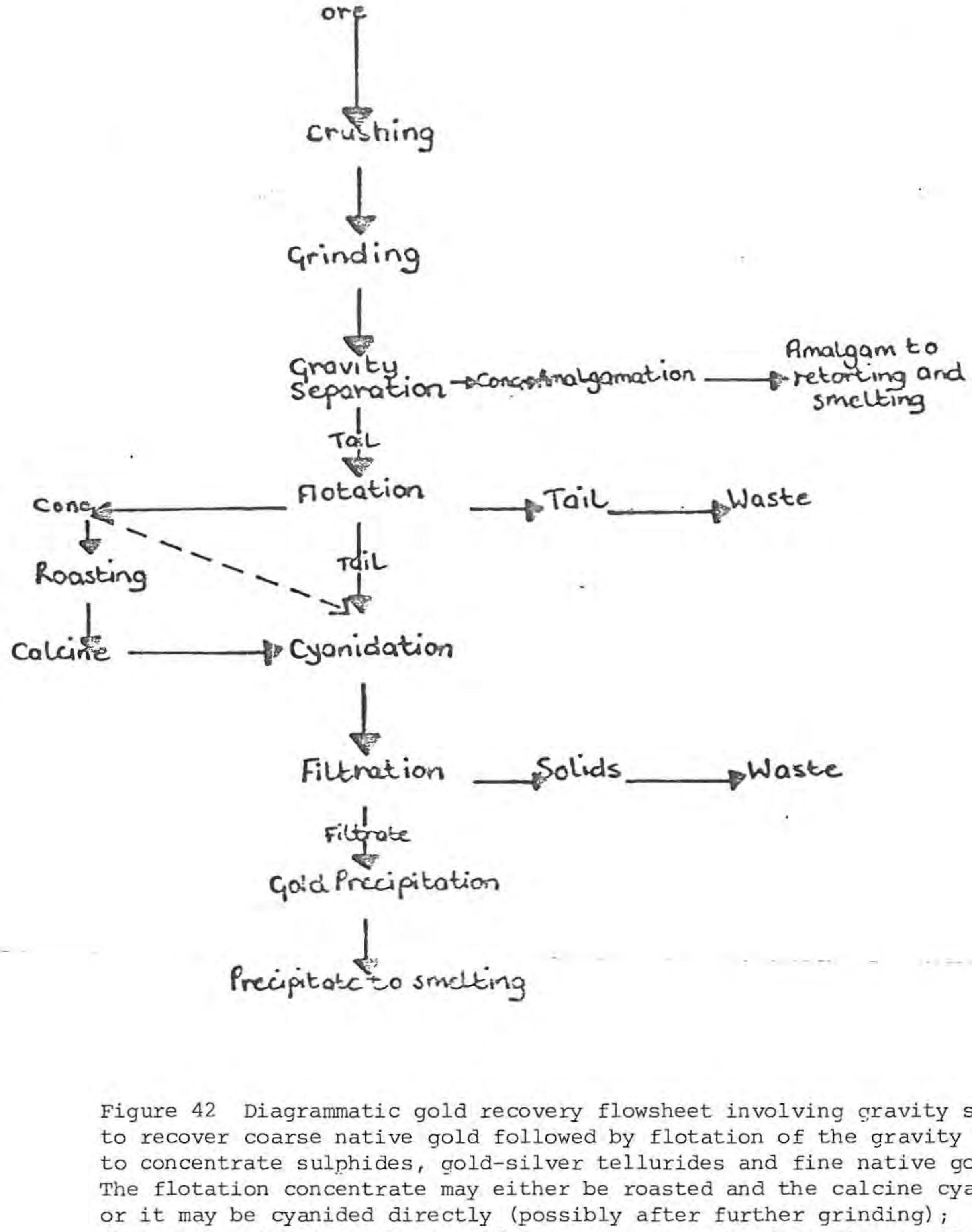


Figure 42 Diagrammatic gold recovery flowsheet involving gravity separation to recover coarse native gold followed by flotation of the gravity tailing to concentrate sulphides, gold-silver tellurides and fine native gold. The flotation concentrate may either be roasted and the calcine cyanided, or it may be cyanided directly (possibly after further grinding); the flotation tailing may be cyanided or sent to waste depending on its gold content.

SUMMARY AND CONCLUSIONS

The Geochemistry of Gold

The average concentration of gold in igneous rocks has been found to be in the range 1,0-5,0 ppb. Mafic rocks are found to have slightly higher concentrations than felsic rocks. This is because gold can be accommodated more readily into early crystalizing silicate and oxide mineral crystal lattices than into salic mineral crystal lattices.

Comparisons of the gold contents of mineral fractions, and the total gold content of the whole rock have shown that not all the gold in a rock is in crystal structures of other minerals. It is thought that this gold is in the form of discrete particles disseminated in the rock. Work on the Merensky Reef (Crockett et al., 1976) and other differentiated mafic bodies has shown that gold is associated with concentrations of late forming immiscible sulphides. Also in the Merensky Reef higher gold concentrations were found to be associated with zones of the reef that had been hydrously altered by a primary magmatic fluid.

It can be concluded from this that excess gold in a magma that is not taken into early formed crystal lattices can be concentrated in late stage aqueous magmatic fluids. This in turn suggests that not all gold found in hydrothermal and volcanogenic deposits need be derived by the leaching of surrounding solid rock.

The mechanisms by which gold is transported are still uncertain. At higher temperatures (>300°C) gold is probably transported as complexes with base metals and chloride ions. At lower temperatures gold is most likely to be transported as sulphide and thio-sulphide complexes. The close association of gold with arsenic, copper and other metals in known deposits suggests that the complexes involve these other elements as well.

The precipitation of gold from a transporting medium is the result of the reduction of the solubility of gold in that medium. The solubility of gold in chloride solutions decreases appreciably at subcritical temperatures. Precipitation of gold at lower temperatures (less than 250°C) is controlled by changes in pH, Eh, fo_2 and total sulphur content of the transporting medium. Boyle (1969) suggested that a major reason for the precipitation of gold is decrease in pressure. With respect to volcanogenic deposits Ridge (1973) has shown that this is not a major mechanism. Any large decrease in pressure would result in the precipitation of vast amounts of chloride minerals, but no such deposits have been found associated with volcanogenic base metal deposits. Submarine deposits are therefore thought to have been formed at depths at which the hydrostatic pressure has prevented the precipitation of chlorides by "boiling".

The Geology of Auriferous Iron-Formations

Gold deposits in Archaean iron-formations are found in the sulphide facies iron-formation. These rocks are associated with mafic and ultramafic rocks similar to modern oceanic volcanic assemblages. It is concluded that these gold deposits were formed in deeper parts of geosynclinal structures. Gold is also associated with base metal massive sulphide deposits. These deposits are related to cyclic mafic to felsic arc-type assemblages near the edges of the geosynclinal structures.

The term "structural controls of mineralization" with respect to the deposits under discussion, is a misnomer. The structures associated with these deposits have not really controlled the emplacement of the mineralization. To a large extent well mineralized portions of iron-formation are associated with tight parasitic folds. Folding is thought to have started at the junction between the more competent sulphide facies rocks, and the less competent material of the other facies types of iron-formation. Continued deformation resulted in the remobilization of the sulphides into the low pressure zones of the folds.

The mineralogy of the gold ore is relatively simple. The gold is wholly associated with the sulphide minerals. Where metals besides iron are present, noteably arsenic and to a certain extent copper, the gold is more often associated with their sulphides.

The exploration and exploitation of gold deposits in iron-formation is controlled and aided by the geology and the associations of the deposits. The stratigraphic positions of the sulphide facies iron-formation dictate the regions for exploration. The associated elements can be used as indicator elements in geochemical surveys. The presence of iron sulphides allows the use of geophysics in their exploration. Magnetometer and electrical geophysical methods in particular can be employed to some advantage.

The mining of the deposits is controlled by their geology. The tendency for the deposits to be small and relatively narrow prevents large scale mechanical methods from being employed. Because of the strong association of sulphides with the gold, 'visual sampling' can be done to reduce operating costs.

The metallurgical treatment of the ores is controlled almost entirely by the fact that a large proportion of the gold occurs as submicroscopic grains in the sulphides. The ores therefore have to be "roasted" during their treatment.

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