

**A REVIEW OF THE DEPOSITION OF IRON-FORMATION AND
GENESIS OF THE RELATED IRON ORE DEPOSITS AS A
GUIDE TO EXPLORATION FOR PRECAMBRIAN IRON ORE
DEPOSITS IN SOUTHERN AFRICA**

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SPINE: **EXPLORATION OF PRECAMBRIAN IRON ORE DEPOSITS IN SOUTHERN AFRICA**

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DISSERTATION

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ABSTRACT

Iron-formations are ferruginous sedimentary rocks which have their source from fumarolic activity associated with submarine volcanism, with deposition of iron as oxides, hydroxides, and hydrous oxide-silicate minerals in shallow and/or deep marine sedimentary systems. The Precambrian iron-formations of southern Africa have a wide age range, but are more prominently developed before 1.8Ga. These iron-formations occur in greenstone belts of the Kaapvaal and Zimbabwean cratons, in the Limpopo mobile belt, in cratonic basins and in the Damara mobile belt.

The Archaean-Proterozoic sedimentary basins and greenstone belts host iron ore deposits in iron-formation. Iron-formations have a lengthy geological history. Most were subjected to intense, and on occasions repeated, tectonic and metamorphic episodes which also included metasomatic processes at times to produce supergene/hypogene high grade iron ores. Iron-formations may be enriched by diagenetic, and metamorphic processes to produce concentrating-grade iron-formations. Uplift, weathering and denudation, have influenced the mineral association and composition of the ores, within which magnetite, haematite and goethite constitute the major ore minerals.

The iron resources of the southern Africa region include the Sishen deposits, hosting to about 1200 Mt of high grade direct shipping ore, at >63% Fe. Deposits of Zimbabwe have

more than 33 000 Mt of beneficiable iron-formation.

The evaluation of an iron ore prospect involves many factors which must be individually assessed in order to arrive at an estimate of the probable profitability of the deposit. Many of these are geological and are inherent in the deposit itself. Other factors are inherent aspects of the environment in which the ore is formed. Although the geological character of the ore does not change, technological advances in the processing techniques may have a great effect on the cost of putting the ore into marketable form.

Geochemical, geophysical and remote sensing methods would be used for regional exploration. Chip sampling and drilling are useful for detailed exploration. Purely geological exploration techniques are applicable on a prospect scale in the exploration of iron ore deposits. Regional exploration targeting should choose late Archaean greenstone belts containing oxide facies iron-formation or Early Proterozoic basins located at craton margins as they are both known to host high-grade haematite orebodies formed by supergene/hypogene enrichment.

Most types of iron ore deposits in southern Africa are described and classified. An attempt is made to emphasize the major controls on mineralisation, in the hope that these may be applicable to exploration both in the southern African region and within analogous settings around the world.

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PREFACE

This dissertation is largely a literature review on the controversies concerning deposition of iron-formation. Emphasis is placed on those iron-formations that may host economic iron ore deposits or those that may constitute sub-economic iron ore deposits. The authors views on the deposition of iron-formation are also expressed.

The review is tackled on a worldwide basis for three main reasons:

- i to establish the general tectonic settings and ages of economic iron ore deposits;
- ii the enormous volume of literature available as compared to the limited southern Africa data base;
- iii and the large amount of research that has been put into some deposits of interest worldwide.

The above data are modelled and are then applied to the southern African scene in order to aid exploration for iron ore deposits on the sub-continent. Emphasis is placed on those deposits that are pertinent to the Zimbabwean and South African situation. Only Archaean and Proterozoic age deposits are discussed although controversies on genetic concepts usually confuse the issue as regards the ages of deposits.

Southern Africa is considered to be that area of Africa south of the 16°S latitude and includes South Africa, Zimbabwe, Mozambique, Botswana, and Namibia. The iron ore deposits in South Africa and Zimbabwe will be discussed. Zimbabwe hosts the Archaean iron-formation hosted iron ore deposits while South Africa has both Archaean and Proterozoic iron ore deposits.

1.0 INTRODUCTION

1. GENERAL

Iron ore is an important commodity because it is the primary source of iron, the metal most used by man. It is produced on a large scale, being surpassed in bulk tonnage by coal, limestone and crude oil. Because the growth of the iron ore industry is intimately linked with the fortunes of the steel industry (over 95% of iron ore is consumed by the steel-makers), the expansion and movement of iron ore closely tracks steel industry developments. Three types of iron ores are used in marketing circles and these are listed below with their commercial factors:

a) haematite and haematite-goethite ores

Lumpy ore (+6-30mm): $\geq 62\% \text{Fe}$; $\leq 0.05\% \text{S}$; $\leq 0.06\% \text{P}$

Fines (-6mm): $\geq 62\% \text{Fe}$; $\leq 0.05\% \text{S}$; $\leq 0.07\% \text{P}$

b) Limonitic ores

Sinter fines: grade - $\geq 56.5\% \text{Fe}$; $\leq 0.05\% \text{S}$; $\leq 0.05\% \text{P}$

sizing - 100%: -9.5mm;

90%: -6mm;

10%: -0.149mm.

Iron ore is generally marketed as *lumpy ore (+6-30mm)*, *fines (-6mm)*, *directly reduced ore (DRO)* and *pellets (+9-16mm)*. Pellets are made from magnetite ores. Titaniferous magnetite is used in smelters as a nitrogen consumer and as a furnace lining protection usually in the *burden* ratio of 3:100.

The ore is won with very large units of earth moving equipment in large open pits, being selectively mined to avoid stratigraphic shale or phyllite bands, metadolerites, and the iron-formation host.

2. RESOURCES

Australia and Brazil account for more than two thirds of the world's seaborne trade because of their enormous reserves. There has never been an iron ore shortage, such is the natural abundance of the commodity although world resources are not truly known.

3. SUPPLY

World iron ore production has generally been on the decline (946Mt produced in 1990 and 1991) after peaking in 1989 (982Mt). Iron ore mines are commonly very large indeed. Sishen, in South Africa, with a capacity of 25Mt per annum, is not unique. Mt Newman in the Hammersley Province (Western Australia) produces up to 31Mt per annum. In 1991, 403Mt of iron ore were shipped on the international trade. South Africa produced 28.95Mt and exported 15.5Mt thus contributing 3.1% of the world's production and 3.8% of the world's exports, respectively. After three years of gradually increasing production up to 1990, output of iron ore in South Africa declined by 4% in 1991. Buchwa mine, the sole supplier of iron ore in Zimbabwe, produces close to 1Mt per annum, all of which is consumed by the Zimbabwe Iron and Steel Company Limited (ZISCO).

World steel production was 733Mt in 1991, nearly 35Mt lower than the output for 1990. Much of the decline in output occurred in Eastern Europe and the defunct U.S.S.R., because of weak local markets, considerable overcapacity, lack of finance for raw materials, energy, investment in plant modernisation and obsolete technology. Further fall in output is expected as restructuring of their industry continues, with closure of the much remaining open hearth steelmaking capacity, a certainty. An expansion of continuous casting and reversal of the long products to flat products ratio,

currently at 2:1, will also be given a higher priority.

The African steel industry, which is dominated by South Africa, grew strongly in 1991, up to 6.8% to 11.4Mt. South Africa increased its output by 10% to 9.5Mt in 1991 to account for eighty percent of the total continent's production and thus reinforced its dominant position. The leading producer, Iscor, increased output by 20% to 7.6Mt and its influence both in Africa and the world is expected to increase as the trade restrictions are eased. ZISCO produces for both the local and export markets.

4. USES

Iron is the backbone of modern civilisation. Iron is used in homes, farms, cities, machines, vehicles, trains and ships. To enumerate the various uses of iron would be to compile a history of the innumerable creations of modern civilisation and industry. Each of the main types of iron-steel, cast iron, wrought iron, and iron alloys - has its particular sphere of use; steel, of course, exceeds all others.

5. OUTLOOK

Because the iron ore industry so closely tracks its steel *daughter*, the "*chill winds*" of lower prices were felt in the ore prices established for 1992. The fall in prices ranged from 4-7% according to grade location. However the overall picture shows some small growth anticipated in global steel production over the next eight years or so. For 1992, though, there is unlikely to be any growth. This is due to the downturn in Japan and especially, the cutback in the former Comecon group, as rationalisation takes place. China, South Korea, Iran and certain other smaller producers will, in contrast, maintain some growth. Iron ore shipments will in

general follow these developments. Output in 1992 from most producing areas will be similar to 1991, but some increase in inventories may occur.

There are a number of incremental expansion projects announced in various countries (Brazil, Australia, Venezuela, Liberia etc) which should ensure the steady growth potential of supply. All these plans, however, will depend on ore prices at least matching inflationary pressures; any attempt by the consuming industry to depress prices could have negative effect on iron ore growth. If this is significant it may result in potential shortages of good quality ore. Consistent quality is one of the main aspects increasingly stressed by ore buyers, who in turn have to face similar demands from their steel customers. It should also be added that electric steel-making could show a relatively higher growth pattern which may impinge marginally on iron ore demand. However, expected growth in DRO production is likely to offset any such reduction in blast furnace operations. Major steps are being taken to further beneficiate iron ore products.

One other factor which affects the overall growth in steel consumption is technology; there is a trend toward less demand for use, for example, in the aviation and car industries, where the search for stronger and lighter materials will continue. This effect is likely going to be felt mostly in developed economies. There is still significant steel growth potential in other parts of the world where current consumption is very low, for instance Eastern Europe, Africa and Asia. However, growth in these areas will, at best, occur in line with the overall global picture unless an innovative investment structure can be engineered to meet the demands of these areas. Such investment could accelerate growth in these countries where per capita steel use is very low.

The overall picture for iron ore should be one of steady growth. The potential exists to meet expansion demands beyond the predicted trend, but this will require good long-term planning and a responsible approach to ore pricing. It is essential that iron ore prices, fixed annually, are sufficient to allow necessary investment programmes to proceed. With greater emphasis on quality and reliability, recognition will be given to the cost of providing materials which meet these demands. The future of the iron ore industry and therefore steel, is good.

B. IRON ORE DEPOSITS

1. WHAT CONSTITUTES AN IRON ORE DEPOSIT?

The definition of iron ore varies from place to place, country to country and continent to continent. It is largely dependent on the economy of the country of occurrence and technology. There was a transition, in the mid-1950s in Canada and the USA, from mining haematite-goethite ores to the production of concentrating-grade iron-formation (*taconite*) ores processed to pellets. The observation that '*the waste of one generation is the ore of the next*' is well exemplified by the i-f story. The Fe-content was thought to be too low and the silica content too high for the iron-formations to be considered ORE until the end of the Second World War. Iron-formation was only mined as ore where it had undergone some form of enrichment process whereby the silica was leached by weathering or other processes and the iron oxides (magnetite) oxidised to the ferric oxides, principally haematite. The ores were termed 'NATURAL' or 'DIRECT SHIPPING' (DRO) ore and the standard was called FIFTY-ONE-FIFTY in Canada and USA, to imply the 51.5%Fe-grade. These ores are called the THABAZIMBI-TYPE ores in South Africa; and the BUCHWA-TYPE ores in Zimbabwe after the mines where these haematitic-type ores were first mined. Both the Thabazimbi-type and the Buchwa-type ores have an iron tenor of $\geq 60\%Fe$ and $< 9\%SiO_2$. The same type of ore is mined at Mt Whaleback Mine, in the Hammersley Basin, Australia.

It became economic to work low-grade magnetite *taconites* and '*plain*' iron-formations because of technological advances which include the following:

- i jet-piercing drills that can drill through iron-formation at low cost;
- ii the beneficiation methods, pelletizing and sintering of the iron.

The result has been an increase in the local (internal) consumption of low-grade DRO's ($\leq 50\%Fe$) by large companies because of the competitiveness of the pellets on the world exports. The viability of an iron ore deposit is discussed further in section F.

2. IRON ORE MINERALOGY AND ASSOCIATIONS

The economic iron ore minerals are listed in table 1 below. They include magnetite, haematite, limonite, and siderite.

Table 1 : Economic Iron Ore Minerals			
Mineral	Composition	%Fe	Commercial Classification
Magnetite	$FeO \cdot Fe_2O_3$	72.4	Magnetic or black ore Red ore Brown ore Spathic, black band, clay ironstone
Haematite	Fe_2O_3	70.0	
<i>Limonite</i>	$Fe_2O_3 \cdot nH_2O$	59-63	
Siderite	$FeCO_3$	48.2	

Other iron-bearing minerals, such as pyrite FeS_2 , pyrrhotite $Fe_{1-x}S$, marcasite FeS and chamosite $Fe_2Al_2SO_5(OH)$ do not occur in significant amounts in large high-grade deposits and therefore are not considered as potential sources in the foreseeable future. Sulphides also create metallurgical problems in the realisation of the iron. The major iron ores are oxides. Of these minerals, magnetite is the richest but is of relatively minor abundance; haematite is in the 'mainstay' of the iron industry. With technological advances the relatively low-grade limonite and siderite ores will become important in the future. Limonitic ores are currently the major sources of iron in Europe; a large limonite deposit will come on stream in Zimbabwe in 1994.

Common impurities in iron ores which sometimes cause major metallurgical problems during smelting are silica, SiO_2 , calcium carbonate (in the form of calcite, $CaCO_3$),

phosphorus, manganese (especially in haematite and limonite), sulphur, alumina (Al_2O_3), water (of crystallisation as in limonite) and titanium (as TiO_2). The maximum allowable levels for various ores are discussed under section A.1 above.

3. REALISATION OF IRON

The realisation of iron is done in two steps namely the reduction of iron oxide to iron (called *pig iron*) in a smelter; followed by the treatment of the pig iron to make various types of wrought iron, cast iron and steel as per the customer's specification.

The ore, coke and dolomitic limestone are charged into the blast furnace. The chemical process involves the combustion of coke in air/oxygen blown at the bottom of the furnace through tuyeres to form carbon monoxide which then reduces the iron ore to the metal. Dolomitic limestone is used as a flux and largely slags off the silica, alumina, phosphorus and other impurities. Dolomitic limestone also controls the basicity of the blast furnace. It is sometimes necessary to add quartz in order to reduce the basicity so as to reduce the slag volumes. A '*rule of thumb*' concerning the relationship between the raw iron ore and the pig iron product, is two tonnes or less of *good* iron ore should produce one tonne of pig iron. Pig iron always contains carbon. The quality of the dolomitic limestone used determines the amount of carbon in the pig iron. The amount of carbon in the pig iron in turn determines the use to which the pig iron is put. Various types of pig iron are basic open hearth, foundry and Bessemer. The Bessemer pig iron is further classified as, low phosphorus, malleable and forge. Specialised pig iron includes that which contains high silicon and manganese. The increased use of iron pellets, sintering and beneficiation is signifying the death of the

Bessemer process.

Wrought iron is purer than pig iron and being malleable and ductile, it is sent to the *puddling furnace* where the impurities are slagged off by stirring. It is hammered into the required shapes in the hot state. Wrought iron resists corrosion.

Steel is an alloy of iron, carbon, vanadium, titanium and manganese so as to enhance its properties like compressibility, tensile and torsional strength, and resistance to corrosion. Steel usually contains 1% carbon but this may be increased to 1.6%. Mild steel contains low carbon usually approaching wrought iron in properties. The old method of making steel was in an open hearth furnace where pig iron, haematite and limestone are charged in a *tilting furnace* lined with refractory bricks. The haematite acts as a flux for collecting excess carbon and silicon; sulphur is vaporised and/or reacts with manganese to form manganese sulphide (MnS). The calcium in the refractory brick lining reacts with the excess phosphorus and this prevents the steel from being '*cold short*', a term used for brittle steel. Pig iron high in phosphorus is treated in a *basic* open hearth or one lined with basic bricks. On the other hand, an acid open hearth is used to treat pig iron low in phosphorus.

The *Bessemer process* is another method of producing steel which seems to have superseded the archaic *open hearth* method. The low-phosphorus and low sulphur pig iron is charged in a barrel shaped furnace through which a stream of oxygen is blown to slag off the impurities. Additives required to make various types of steel, for example, carbon and manganese, are introduced by air blowers. Various ferro-alloys are made by adding proper amounts of alloy metals to yield the desired steel.

The *taconite* or iron-formation is upgraded as follows:

- i crushing;
- ii magnetic separation of magnetite;
- iii adding 1% bentonite as a binder;
- iv rolling the mix to 1-2- centimetre pellets; and
- v baking to harden the pellets.

4. TYPES OF IRON ORE DEPOSITS, OCCURRENCE AND DISTRIBUTION

4.1 Introduction

Iron ore deposits are widespread in space and time. Based on genetic concepts, iron ore deposits can be classified into three major groups namely magmatic, sedimentary and residual deposits. This classification is also linked to the age of the deposits.

Ore deposits can also be broadly divided into those generated by *endogenetic* processes and those generated by *exogenetic* processes. The former, into which the magmatic iron ore deposits fall, are invariably associated with thermal processes and, in general, can be related more readily to magmatic and tectonic activity. The sedimentary and residual iron ore deposits fall into the latter group. These are deposits formed by surficial processes such as weathering and shallow- or deep-marine sedimentation, which have tenuous genetic relationships.

4.2 Magmatic deposits

The magmatic deposits are formed at the late stages of fractional crystallisation of layered complexes e.g. the Bushveld Complex. These layered complexes span the Archaean-Proterozoic boundary (Great Dyke of Zimbabwe; are of Proterozoic (Bushveld Complex of South Africa) and Phanerozoic (Stillwater Complex of USA) ages.

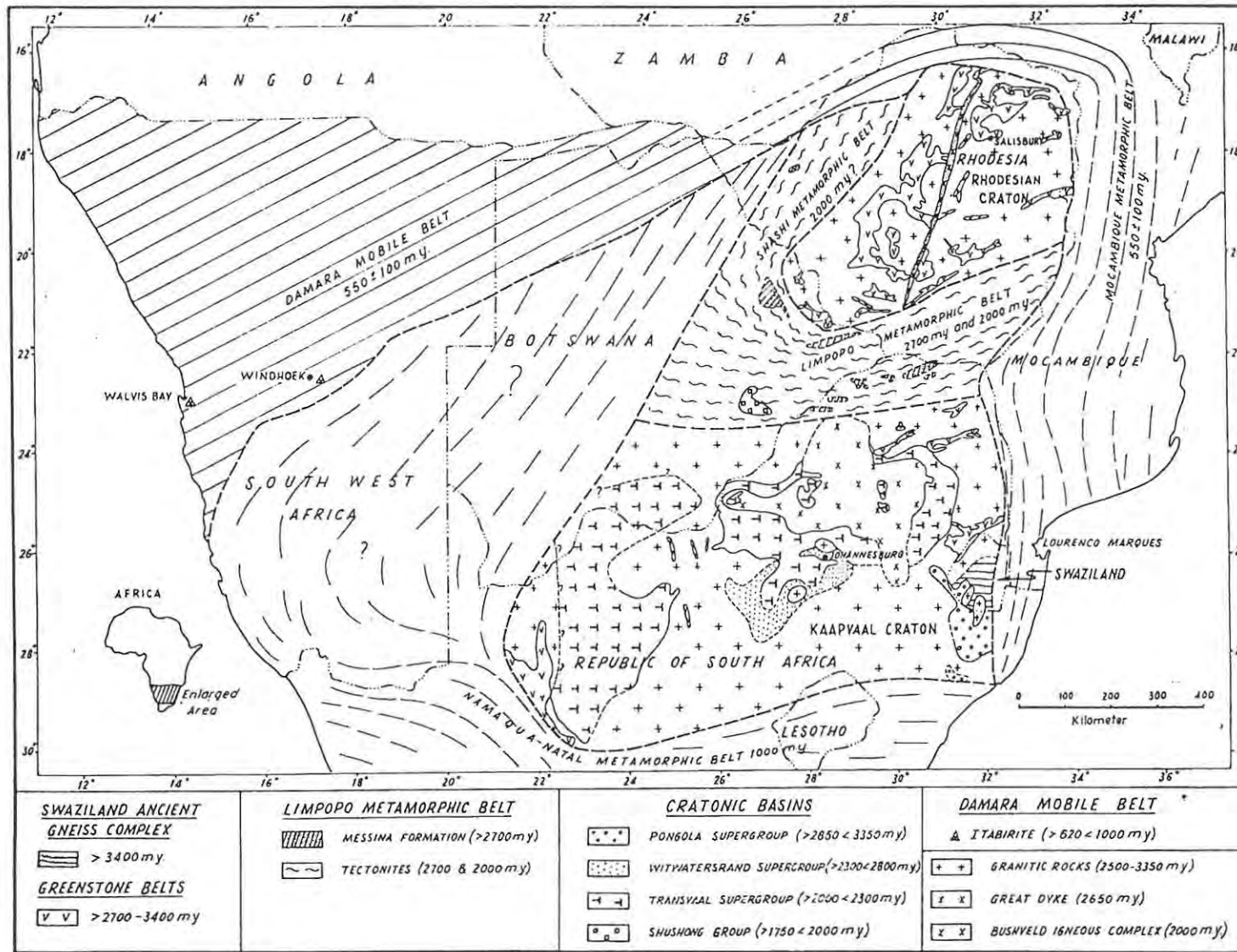


Figure 1: Geotectonic map of Southern Africa showing the distribution and classification of the tectono-sedimentary units in which the iron-formations occur (after Beukes, 1973).

The magmatic iron ore deposits occur: a) within anorthositic massifs, as at Kiruna iron ore mine in Sweden; and b) as upper differentiates forming discrete layers usually concordant with layering at the top of layered complexes. Examples of occurrences are at Mapoch mine that occurs in the Upper Zone of the Bushveld Complex (figure 1); and magnetite layers developed in the Stillwater Complex (USA), Skaergaard (Greenland), Duluth (USA), Muskox (Canada), Mambula (Zululand, South Africa), and Kapalaqula (Tanzania). The ages of these deposits are depicted by the age of the layered intrusion and therefore the magmatic deposits span the Archaean to Tertiary.

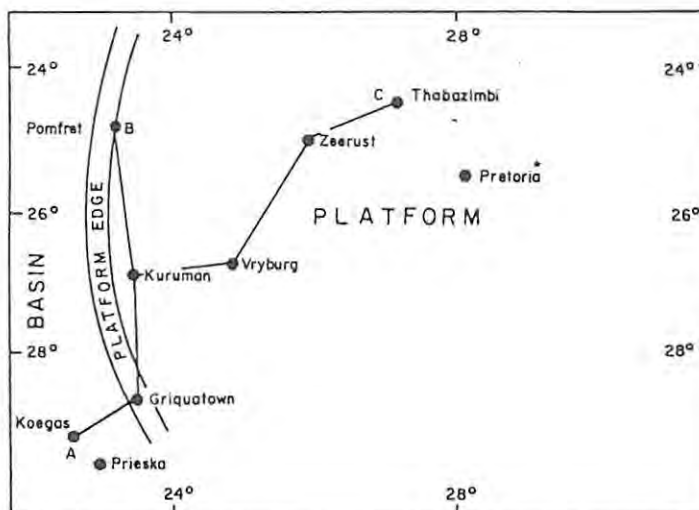
Magmatic deposits have a high Ti and V content (except the Kiruna deposit, Sweden which has high P instead) which causes problems in smelting. These deposits are, therefore, largely exploited for V and Ti, for example, the Highveld Steel and Vanadium Company owned Mapoch mine, Eastern Transvaal, South Africa. Iron ore is obtained as a by product.

4.3 Sedimentary deposits

The sedimentary iron ore deposits are hosted in iron-formations in Archaean ($\geq 2.5\text{Ga}$) greenstone belts and Early Proterozoic sequences. Because of their age, the sedimentary iron ore deposits are affected by metamorphism, deformation in the form of faulting, folding, rotation and subsequent supergene and/or hypogene enrichment. Sedimentary iron ore deposits form the bulk of the exploited deposits in the world.

The iron-formations of Archaean age (figure 1) are interbedded with thick successions of greywacke, phyllite and slate. Examples are, the iron ore deposits at Buchwa, Nyuni Hills, Black Mamba near Kadoma, and Manisi in Zimbabwe. The most spectacular deposits are those of Proterozoic age, deposited mainly between 1700 and 2500Ma. The best known of

these include formations in the Lake Superior district in the USA, the Superior Province of the Canadian Shield, the Hammersley Basin of Western Australia and the Transvaal basin of South Africa (figure 1). The



Hammersley Basin is in excess of 480 kilometres in length and contains the 915-

Figure 2a: Distribution of major tectono-sedimentary elements that controlled the deposition of the Transvaal Sequence (fig 2 section) (after Beukes, 1986).

metre-thick Brockman iron-formation. The Transvaal basin which is split into the Transvaal and the Griqualand West sub-basins contains two very thick iron-formations which host the Thabazimbi and the Sishen iron ore deposits, respectively (figure 2). Other well known Precambrian iron-formations occur in the Krivoy-Rog area of Russia, in India, Liberia, Ghana, and Brazil.

The distribution of the Precambrian iron-formations in southern Africa are shown in figure 1 (after Beukes, 1973). The genesis of the sedimentary ore deposits is linked to the controls of deposition of the iron-formation host rock.

4.4 Residual deposits

The residual iron ore deposits are those that are formed as a result of weathering processes and are generally Recent in age. These deposits include laterite which form in Equatorial

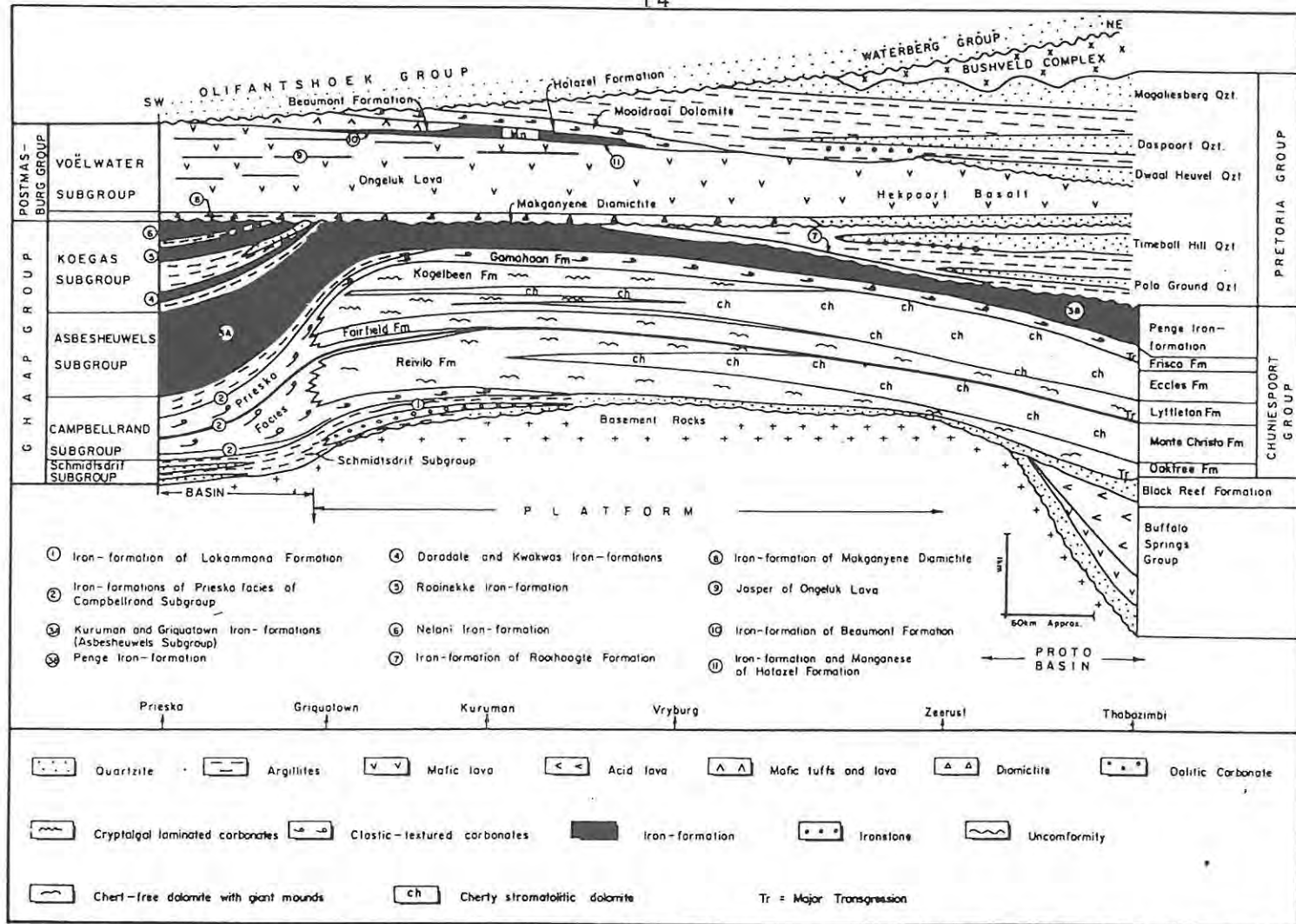


Figure 2: South-west to north-east stratigraphic correlations and sedimentary facies relationships in the Transvaal Sequence between the Griqualand West and Transvaal sub-basins. Line of section indicated in Fig 2a (after Beukes, 1986).

regions. The residual ore deposits are usually small in size and economically insignificant.

C. IRON-FORMATIONS

1. NOMENCLATURE

The vast majority of the world's iron ore requirements comes from Precambrian-age cherty *banded iron formations* (BIFs).

BIFs are thin- to medium-bedded interlamination of iron oxide, iron carbonate, or iron silicate materials with chert or jasper. The terms *iron-formation* and *banded iron-formation* are used inconsistently by geologists which led to confusion and misunderstanding. A generalised definition by James (1954) for *iron formation* states:

"a chemical sediment, typically thin-bedded or laminated, containing 15% or more iron of sedimentary origin, commonly but not necessarily, containing layers of chert".

Most economic iron-formations contain 25-35%Fe and the essential characteristics of the iron-formation are the presence of thin layers or nodules of chert. It was proposed by Kimberley (1978) to use *ironstone* to refer to James's (1954) *iron-formation* and to restrict use of the term *iron-formation* to describe a mappable rock unit or 'package' dominated by HIS (i.e. Kimberley, 1978) *ironstone* and with ironstone layers defining its top and bottom. Because all iron-formations are banded it is proposed that '*banded*' be left out in *banded iron formation* and that 'iron formation' be linked by a hyphen so that James's (1954) banded iron formation becomes '*iron-formation*' (i-f). Mappable rock units are usually assigned names that bear no resemblance to the rock type. For example mappable sequences of sandstone or granite or limestone are not respectively referred to as sandstone formation or granite formation or even limestone formation. They are in turn given names like Lannes Formation, or Penge Formation or Longwe Formation. Therefore Kimberley's argument of wanting to differentiate between rock unit and mappable unit for i-f is baseless.

There are two broad classes of i-f units namely the *ALGOMA-type* and the *SUPERIOR-type*. The Algoma-type is Archaean (≥ 2.5 Ga) in age and is specifically relatable to submarine volcanic processes. The Superior-type is Early Proterozoic (2.5-1.8Ga) in age and does not necessarily include volcanic input. It will be argued that the Superior-type i-f is related to distal submarine volcanism (Section C.9).

2. MINERALOGY AND PETROGRAPHY

The iron minerals in i-fs are shown in table 2.

Table 2: Original i-f precipitate and metamorphic equivalents		
Compound	Inferred Initial Precipitate	Now Observed
SiO ₂	Amorphous	Chert
Fe ₂ O ₃	Amorphous Fe ₂ O ₃ .nH ₂ O	Haematite
Fe ₃ O ₄	Hydro-magnetite Fe ₃ O ₄ .nH ₂ O	Magnetite
FeCO ₃	Siderite	Siderite
Fe ₃ Si ₂ O ₅ (OH) ₄	Amorphous ferrous silicate	Fe ₃ Si ₂ O ₅ (OH) ₄ greenalite
		Fe ₇ Si ₈ O ₂₂ (OH) ₂ grunerite
Fe sulphide	FeS	FeS ₂
Na-Fe silicate	Na-Fe silicate gels	Riebeckite Stilpnomelane Chert
		Fe ₃ Si ₄ O ₁₀ (OH) ₂ minnesotaite Fe ₂ SiO ₄ fayalite

(after Garrels et. al., 1973)

The minerals mostly found in i-fs include granular magnetite, Fe₃O₄; haematite, Fe₂O₃, as granular black material, as crystalline specularite, as soft red powdery haematite, or as reniform '*kidney ore*', and massive earth limonite (Fe₂O₃.nH₂O, Fe[OH]₃, or FeO.OH). Siderite, FeCO₃; chlorite, Fe₆Si₄O₁₀[OH]₈; greenalite (the Fe²⁺ analogue of kaolinite, [Fe,Mg]₆Si₄O₁₀[OH]₈); chamosite (the Fe³⁺ analogue of kaolinite or greenalite, Fe₄Si₄O₁₀[OH]₈); minnesotaite (the Fe²⁺ analogue of talc [Fe,Mg]₃Si₄O₁₀[OH]₂); grunerite (the amphibole, [Fe,Mg]₇Si₈O₂₂[OH]₂); stilpnomelane, [Ca,Na,K]_½[Fe²⁺,Mg,Al]₄Si₄O₁₀[OH]₂.2H₂O; the olivine fayalite,

Fe_2SiO_4 ; the ferruginous chert '*jasper*' and the sulphides, pyrite, FeS_2 ; and pyrrhotite, Fe_{1-x}S have variable abundances.

I-fs consist essentially of alternating chert and iron oxide meso-bands¹ ranging from a fraction of a millimetre to 4cm. Siliceous i-f contain thinner iron oxide meso-bands than the chert mesobands. Microbanding¹ is apparent in both chert and the iron oxide meso-bands.

Silicate facies i-f are generally sub-ordinate. Grunerite is present in i-fs adjacent to granitic plutons, and for this reason, it is relatively more abundant in those i-fs that are interbedded with ultramafics. Because grunerite is mostly confined to the magnetite-quartz meso-bands or to the contacts between magnetite and quartz meso-bands, it is thought that the grunerite is of metamorphic origin and represents a reaction product between magnetite and quartz (Phaup, 1933; Harrison, 1970). Grunerite can also be formed from the weathering of minnesotaite or greenalite.

Carbonate facies i-f consist of iron carbonate-rich mesobands alternating with chert (Harrison, 1970). Carbonaceous shales and slates intimately associated with i-fs, represent the sulphide facies.

An i-f whose mineralogy is dominated by magnetite, iron silicates, chert and minor to accessory haematite and siderite is called a *taconite*. When granular quartz and haematite become the dominant minerals as a mixture then the i-f is termed an *itabirite*. Itabirites are found in Brazil.

In the field i-fs are thin-bedded, with strong colour contrast between dark bands of haematite or magnetite, red bands of jasper, and light to white bands of chert. In polished sections the thin banding is accentuated and there

¹Mesobands are normally 1mm-5cm thick; microbands are anything less than 1mm.

is an abundance of pull-apart and apparent slump features. These features which are a result of soft-sediment deformation, are typical of i-fs. The layers and nodules of silica are chert. Because of metamorphism and later weathering the silica has recrystallised to resemble quartzite or sandstone. Examination of many specimens has led to the conclusion that the material is predominantly the result of chemical sedimentation with minor localised detrital input (Guilbert and Park, 1986). Tyler (1949) describes silica which has a mosaic texture as indicative of in-situ crystallisation. Another observation in support of chemical sedimentation followed by in-situ crystallisation is the lack of heavy accessory minerals in i-fs, usually expected and observed in clastic sediments. The stratigraphic accumulations in a basin indicate a lateral gradation from local littoral accumulations of mechanically deposited grains to chemical precipitates basinward (figure 3). Therefore, the

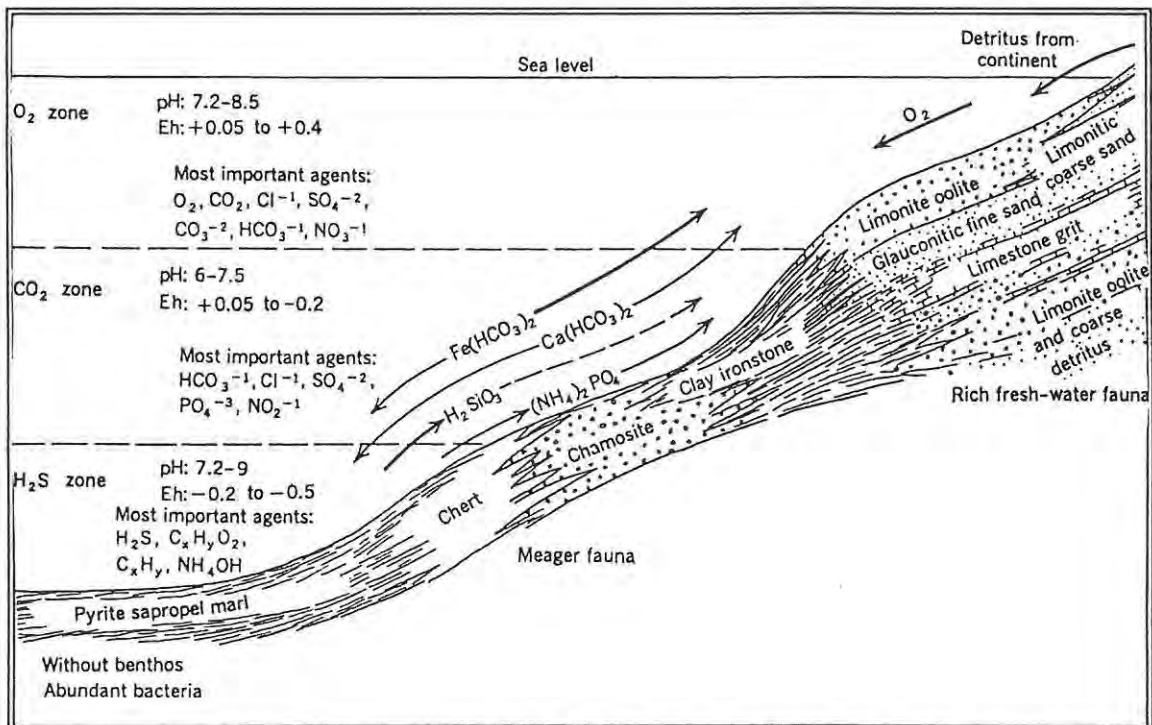


Figure 3: Schematic Section showing the relationship between iron-formation facies and physico-chemical conditions (after James, 1966 in Pettijohn, 1975).

bulk of the silica in i-f is chemically precipitated chert.

Lateral facies change is reflected in the grading of i-fs along strike into banded ferruginous chert, and ultimately into banded chert or *quartzite*, as a result of the lateral variation in the iron and silica. Quartzite is considered to be recrystallised banded chert or a residual rock on which enrichment processes acted.

3. SEDIMENTARY STRUCTURES

Stowe (1968) described some relict oolitic textures in the i-fs in the Kwekwe region of the Midlands greenstone belt. Ripple marks, cross-bedding, graded bedding, scour and fill structures, rill marks and small depressions have been reported by various authors. Other structures of major importance which include post-depositional, pre-lithification slump folds, bordered on either side by undisturbed i-f (Macgregor, 1928; Bliss, 1962), indicate that the material was in a plastic state at the time of deformation. These features are characteristic of i-fs in greenstone belts.

4. TEMPORAL AND SPATIAL DISTRIBUTION

I-fs are widely distributed in the world. The largest and most abundant ones are found around the Atlantic and Indian Oceans. There are minor occurrences in the Pacific Basin. The major iron ore producing districts are Brazil, Venezuela in South America; the Lake Superior region in Canada; the Hammersley Basin in western Australia; the Transvaal Basin in South Africa (figure 1); the west coast of Africa, particularly Gabon, Liberia, and Mauritania; and in several districts in Russia, India, Manchuria. Large deposits were discovered in 1960 in Australia, and in 1967, 16 billion tonnes at 67%Fe in the Carajas Range, 700km south of the mouth of the Amazon river in Brazil.

The time distribution of i-fs is shown in figure 4 (after James, 1983) and is meant to be illustrative rather than quantitative. It is apparent from figure 4 that most of the world's i-f falls within four general age groupings: middle Archaean (age 3.5-3.0Ga), late Archaean (age 2.9-2.6Ga), early Proterozoic (age 2.5-1.9Ga) and late Proterozoic-early Phanerozoic (0.75-0.45Ga) (James, 1983).

In terms of the quantity of i-f initially deposited and preserved, the sedimentation peak of early Proterozoic age is by far the most significant. This is the age of most great deposits of the Lake Superior type, with

90% of all known i-f is assigned to this depositional epoch. In terms of number of stratigraphically distinct occurrences, the (quantitatively) minor peak of late Archaean age is most notable, with deposits numbered in the thousands. Virtually all of these i-fs are of the Algoma type, deposited in orogenic environments and closely associated with the volcanic rocks that now define the late Archaean greenstone belts in many parts of the world (James, 1983).

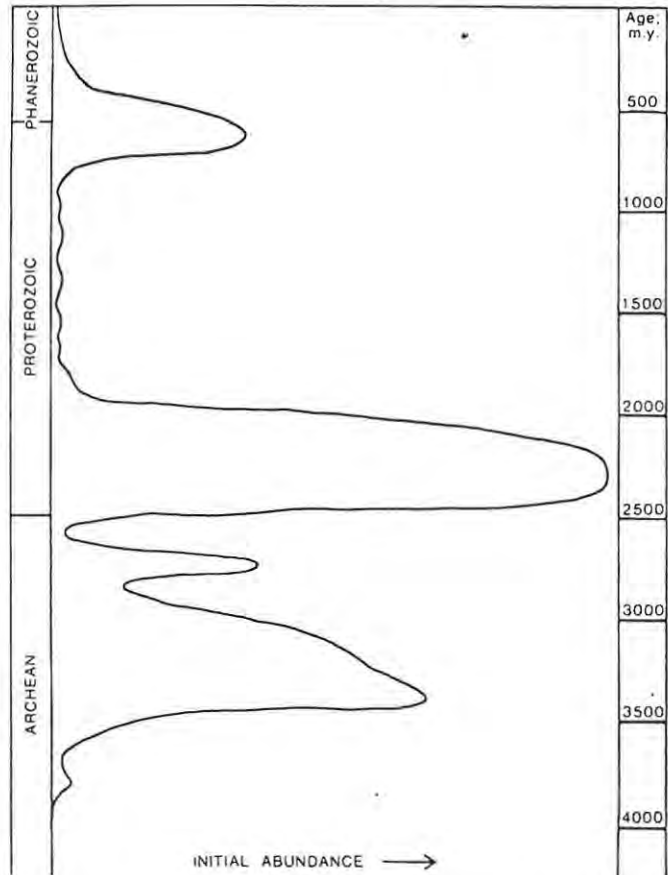


Figure 4: Estimated abundance of iron-formation deposited through geologic time. Horizontal scale is logarithmic; range 0-10¹⁵ tonnes (after James, 1983).

Deposits of the 3.5-3.0Ga age occur within greenstone belts, such as the minor deposits in the Swaziland Sequence of southern Africa and can be regarded as Algoma type, similar to the Late Archaean greenstone belts. However, some of these older greenstone belts are not readily classified due to poor preservation of their stratigraphies.

The late Archaean (2.9-2.6Ga) was an epoch of cratonisation in most parts of the world, with widespread volcanism (much of it submarine), contemporaneous volcanogenic sedimentation (including i-f), and igneous intrusion in localised belts or basins within (presumably) areas of thin sialic crust. Within individual areas the complex events associated with what are now recognised as *greenstone belts* typically occupied no more than about 50Ma. The cratonisation process surely was not synchronous even within a single shield, but it seems to have reached a peak in many parts of the world at about 2.7Ga ago. However, in some regions the crust was stabilised much earlier - the greenstone belts on the Kaapvaal craton of South Africa, for example have ages that approach 3.4Ga (Anhaeusser and Button, 1976).

The late Proterozoic-early Phanerozoic epoch (750-450Ma) of iron sedimentation is not well understood, perhaps because the assigned deposits may include several genetically unrelated types. Some appear to be of the Algoma class, associated with contemporaneous volcanism; others bear an ill-defined relation to late Proterozoic glaciation. These deposits do not occur (or have not been reported) in southern Africa and therefore are beyond the scope of this report. A whole volume, special publication No. 46, edited by Young, T.P. and Taylor, W.E.G. (1989) dedicated to the Phanerozoic iron-formations was published by the Geological Society of London and interested readers are referred to it.

5. SIGNIFICANCE OF PEAKS IN THE DEPOSITIONAL RECORD

Only speculations as to the meaning of the great variation in the rate of i-f deposition through time can be offered.

The middle Archaean interval is the least sharply defined of all maxima in the depositional record. No depositional peak appears to exist. However geological data is fragmentary and therefore, it is possible that the distribution pattern may be related more to fortuitous preservation after deformation than to a time-controlled cycle of deposition. This, if so, leads to the inference that during this early period of Earth history, sea water at all times contained abundant iron and silica in solution derived from submarine volcanism, and that i-f was deposited whenever and wherever environmental conditions favoured precipitation rather than retention (James, 1983). For example, an abundance of oxygen-producing microbiota may have induced precipitation of iron.

The late Archaean interval is coincident with the widespread volcanism that produced the great greenstone belts of this age. Early Proterozoic volcanism is largely manifested as lavas in clastic sequences. Individual units of i-f are numerous, but the total quantity is not particularly large. It seems evident that despite undoubted fumarolic contributions of iron and silica on a major scale to a seawater already virtually saturated, the continued crustal instability and the periodic flooding of depositional basins with volcanogenic debris input is not compatible with the quiet environmental conditions necessary for long-continued '*clean*' chemical sedimentation. Significant accumulations of i-f were possible only during relatively brief periods of quiescence. Chemical deposition was halted or heavily diluted by increased input of clastic materials (James, 1973). There was also a lack of abundant oxygen producing bacteria to raise the Eh necessary for precipitation of iron. Consequently, greenstone belts i-fs are numerous relatively

thin and discontinuous units.

A large time gap of about 1Ga exists in which no i-fs were deposited before the onset of deposition again during the late Proterozoic-early Phanerozoic interval. However, this epoch is one of uncertain significance because of the lack of data. Some of the i-fs are NOT i-f in the true sense of the definition of James (1954). The 'i-fs' lack the banding and are associated with coarse clastic sediments of possible glaciogenic origin. The glaciogenic association would cause a problem for the Griqualand West subgroup because of the diamictite-associated Manganore formation. This indicates that '*laminated ore*' of the Sishen deposits might be Phanerozoic in age. Some ore deposits like those found in the USSR resemble the Algoma-type i-f associated with regional orogeny and volcanism.

6. DEPOSITION OF IRON-FORMATION

6.1 The deposition process: theories and models

The period of greatest iron-sedimentation in the Earth's history is in the early Proterozoic interval (2.5-1.9Ga). The depositional maximum is believed to be due to coincidence of a number of more-or-less independent variables - some structural, some geochemical and some biological. James (1983) advanced a model as an explanation for this extraordinary epoch presented here verbatim:

1. ***"During Archaean time, and continuing into early Proterozoic time, the oceans, below a thin surface layer in equilibrium with a weakly oxygenic atmosphere, constituted a major reservoir for dissolved iron and silica derived from diverse sources.***
2. ***"The almost worldwide era of orogeny and cratonisation of late Archaean time was followed by a long period of crustal stability - the Eparchaean interval of older concepts - during which continental blocks were reduced to surfaces of low relief, with***

consequent very low levels of clastic input to basins of deposition.

3. *"During the early Proterozoic, but probably at different parts of the world, the previously stable cratons were subjected to weak structural disturbances, perhaps in part, epeirogenic but mainly extensional, with consequent development of shallow intra-continental troughs and marginal basins.*
4. *"Encroachment of seawater into the newly formed troughs and basins with continued replenishment by nutrient-rich deeper ocean waters, triggered a series of events, including rapid growth and evolution of biota, some of which were (or became) oxygen producers, and precipitation of iron and silica to form i-f. The precipitation of iron and silica was induced in part, by greater availability of oxygen (including that locally generated biologically by photosynthesis), in part, by increased concentration of iron and silica due to seawater evaporation, and perhaps in part directly by biological processes. The nature of the iron precipitates would depend upon local conditions. In broad, unrestricted shallow basins, oxide facies would be produced, whereas in deeper, barred basins and troughs, conditions might favour precipitation of silicate, carbonate, and sulphide facies, none of which require an excess of oxygen.*
5. *"Geochemical transfer of dissolved iron and silica, contained in the constantly upwelling deep ocean waters, to marginal basins and troughs, to be trapped out as chemical precipitates, was an irreversible process. It was halted locally, only by change in configuration or filling of the basin of deposition, or ultimately, in a worldwide sense, only when the oceans were essentially depleted of at least their iron content. In essence, the existence of sites appropriate for upwelling and precipitation provided, not just the means whereby large-scale*

chemical equilibrium could be established between deep oceanic waters and evolving oxygenic atmosphere, but in fact the necessary milieu for the biological evolution and biological processes that were responsible for those changes in the composition of the Earth's atmosphere'.

This model is very much dependent on the structural conditions, i.e., the development of basins on stable but generally a weak, and therefore, young craton. The general observation is that these cratonic basins of major i-f development are *sub-basins* of the major oceanic basins which are our present-day oceans. The sub-basins could also represent the proto-oceans since oceans have opened and closed many times since the Early Proterozoic. The Atlantic ocean is the major basin for the Minas Gerais Basin in Brazil which broke up from the western coast of Africa; so is the Indian ocean the major basin for the Hammersley-Nabberu basins, in Australia. The Transvaal basin is central to both the Indian and Atlantic major basins. The weaker the craton, the larger the sub-basins formed and therefore the thicker the i-f horizons. The Pilbara craton must have been weaker than the Kaapvaal craton which explains why the Hammersley-Nabberu basins hosts the thickest i-f in the world, about 15 000m (Trendall, 1983) as compared to 300m and 250m for the Transvaal (Beukes, 1973) and Minas Gerais (Dorr, 1973) basins, respectively. The larger the basin, the more complete the i-f types (or *facies* ?) ranging from oxide to silicate to carbonate and finally sulphide facies at the deepest portion of the basin, for example, the Hammersley basin. The independent development of basins/troughs results in different ages of the i-fs within the epoch.

Six theories on the mode of i-f deposition are:

1. *The source of iron and silica was volcanism at the seafloor which formed springs of iron and silica rich fluid of magmatic origin (Trendall, 1965, 1968;*

Gross 1980).

2. *Iron and silica were transported in solution having been leached from nearby landmasses and were rhythmically deposited as chemical sediments in water, probably in response to seasonal variation of the water composition. The mode of deposition was direct inorganic precipitation of silica and iron or one of several biochemical processes (Baarghorn and Tyler, 1965; Eugster and I-Ming, 1973).*

3. *The deposition of i-f involved emplacement of thickly bedded fine grained ferruginous tuffs and other iron-rich sediments which were subsequently diagenetically oxidised and silicified under the influence of solutions that were partly volcanic in origin which resulted in the segregation of the sedimentary pile into alternate finer beds of banded chert/jaspers with more iron-rich layers (Dunn, 1935, 1941).*

4. *I-fs were deposited as end members/final products of carbonate sedimentary cycles (Button, 1976a).*

5. *Deposition of i-f occurred after an accumulation of iron concentration in a primitive Archaean acidic sea with a pH of ≤ 6 , an Eh of about 0, and with seawater in equilibrium with an atmosphere rich in CO_2 . Iron was sourced from erosion of landmasses and volcanism and under these conditions would remain in the ferrous form in the sea. The depletion of CO_2 in the atmosphere, as time went by, led to an increase in the pH of the sea with removal of H_2CO_3 by the cyanobacteria and ultimately condition conducive to the precipitation of FeCO_3 and Fe_3O_4 which led to the wholesale deposition of iron in seawater as magnetite, haematite and siderite (Jolliffe, 1966 in Guilbert and Park, 1986).*

6. *The upwelling of cold, deep seawater saturated in CO_2 , Ca^{2+} , and Fe^{2+} onto the warm continental shelf resulted in the formation of i-f. Increased temperature, oxidation potential, and CO_2 loss would lead to the oxidation of Fe^{2+} and the precipitation of Fe^{3+} -iron oxides, hydroxides or silicates (table 2). The same water would be supersaturated in amorphous silica and would precipitate chert (Holland, 1973).*

All the above proposals revolve around the idea that the source of the iron was seafloor volcanism which vented out iron-rich fluids under low Eh and high pH conditions, a proposition which is also supported by Fryer (1983) using REE data. Fryer (1983) found that there is a characteristic Eu enrichment in most Early Proterozoic and Archaean i-fs which can be interpreted as implying a significant hydrothermal input into the water into which they precipitated. Anomalous Ce behaviour in i-fs of Early Proterozoic age suggest the existence of strongly oxidising conditions in the marine environment, at least locally, at this time with obvious implications for iron transport and oxygen budgets in the oceans (Fryer, 1983). The Archaean and Proterozoic ocean was hotter and depleted in ^{18}O compared to the modern ocean (Perry, 1983).

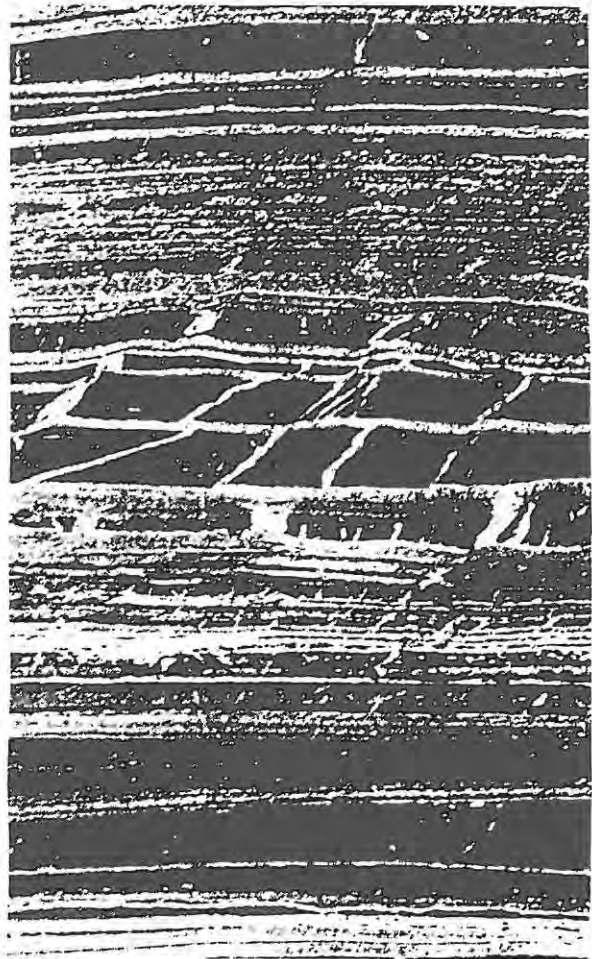


Figure 5: Close-up of i-fs showing the quartz (white) slumping into and forming veins in haematite-magnetite (black) and jasper (grey)-Sherman mine, Canada (Guilbert & Park, 1986).

6.2 The Role of Bacteria on the deposition of i-f

THE FIRST OXYGEN PRODUCERS

The role of bacteria in the deposition of i-f is not fully understood hence, it is not possible to determine their

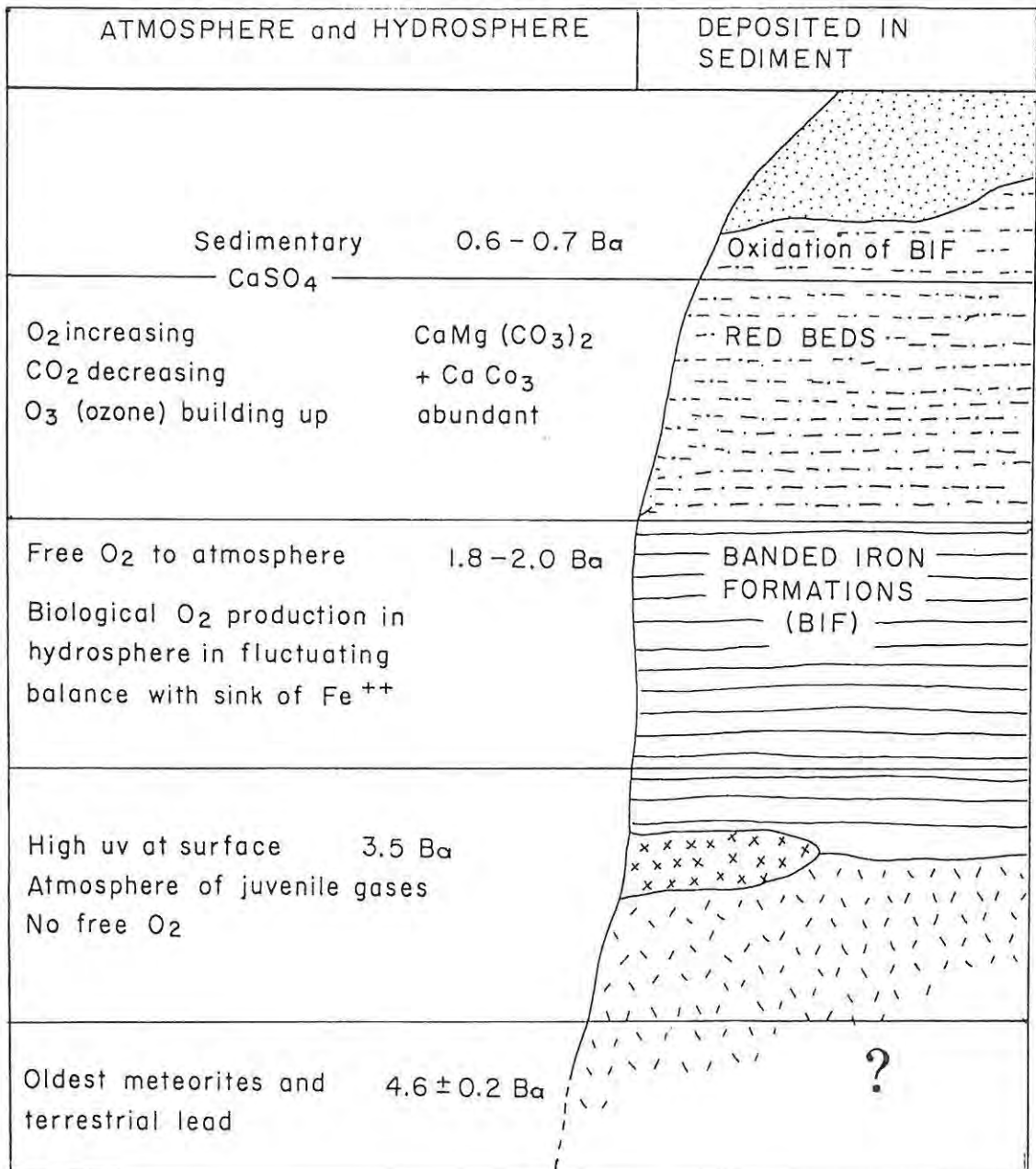


Figure 6: Postulated main features of lithospheric and atmospheric history on the primitive earth (after Cloud, 1968 *in* Margulis, 1970).

importance although the probable relevance cannot be overruled or side-lined. Iron-secreting bacteria exist in the present environment (Cloud, 1973, Margulis, 1981). What is not clear is whether the Archaean and/or Precambrian environment was conducive to their existence. Under favourable conditions these micro-organisms could also have played a major role in the deposition of iron by catalysing

an organic reaction that would produce the same results if given enough time. The occurrence of stromatolites in the Transvaal Dolomites ($\approx 2.3\text{Ga}$) in South Africa, the Bulawayan Dolomites (2.7Ga) in Zimbabwe indicate that biogenic oxygen accumulated during the Precambrian on a worldwide level and there is no evidence to argue against the idea suggesting similarity to the metabolism of the stromatolites and the oncolites found in the Precambrian carbonate sequences and their Palaeozoic and modern analogues (figure 6). The first organisms to generate oxygen photosynthetically were the cyanobacteria, traditionally called '*blue-green algae*'. These cells had their days of abundance, diversity, and domination of the landscape, long before the Cambrian. They first appeared in the Archaean, and by late Proterozoic many species had spread over the Earth. In the Proterozoic they built major rock formations, like extensive reefs and may be responsible for the widespread deposition of the i-fs. These algae became less abundant towards the end of the Proterozoic (1.8Ga). Several thousand species of the cyanobacteria exist today forming scums on ponds, puddles, rice paddies, and shower curtains - where ever they can find light, water and security from scavengers. Their decline from dominance is considered to be due to several factors the major one being the immense success of the eukaryotic algae and plants that succeeded them (Margulis, 1984).

THE TRANSFORMATION OF THE ATMOSPHERE

The rise of aerobic photosynthesis was a global catastrophe in as far as these microbes were concerned. This is due to the fact that oxygen is very reactive and thus '*grabs*' electrons to produce '*free radicals*' which are highly reactive, short-lived chemical species that wreak havoc with organic compounds by forcing fermenting bacteria and photosynthetic sulphur bacteria to retreat to oxygen-deficient zones. Oxygen produced by the first cyanobacteria

did not accumulate in the atmosphere because it must have combined quickly with other gases, including hydrogen, ammonia, carbon monoxide, and hydrogen sulphide. This hypothesis is supported by the observation of hydrogen sulphide, produced by present-day lake-bottom bacteria, reacting with oxygen before it reaches the surface of the water giving hydrogen sulphide no chance of accumulating (Margulis, 1984). Oxygen also reacts with elements and minerals dissolved in oceans, rivers, and lakes.

The formation of sedimentary rocks is linked to processes involving the reworking and transportation of surface materials, which are thus brought into extensive contact with the atmosphere. Therefore sedimentary rocks formed in an atmosphere with abundant oxygen will contain minerals with the constituent elements in the oxidised form and vice versa. The minerals found in the older Archaean rocks contain reduced forms of sulphur, uranium and iron. Iron occurs mostly in the reduced/less oxidised Fe^{2+} , ferrous form rather than in the more oxidised Fe^{3+} , ferric form. The possible Eh-pH variations possible are shown in figure 7. Similarly, uranium is found in the less oxidised uraninite form in sediments of 2Ga or older thereafter oxidised $\text{U}^{5+}/\text{U}^{6+}$ forms like carnotite², tyuyamunite, torbernite and autunite become common. In the Archaean, sulphides are common and sulphates are rare.

From the foregoing discussion, the transition to the oxidising atmosphere is further highlighted by i-fs. The Archaean i-fs are made largely of layers containing more reduced iron (magnetite) alternating with layers containing more oxidised iron (haematite) indicating varying concentrations of atmospheric oxygen. The Ghoko-Longwe i-f, near Somabhula, the Manisi i-f, near Chivhu and the Buchwa i-f, south of Zvishavane (Zimbabwe localities) are excellent

²The chemical formulae are shown in appendix 1

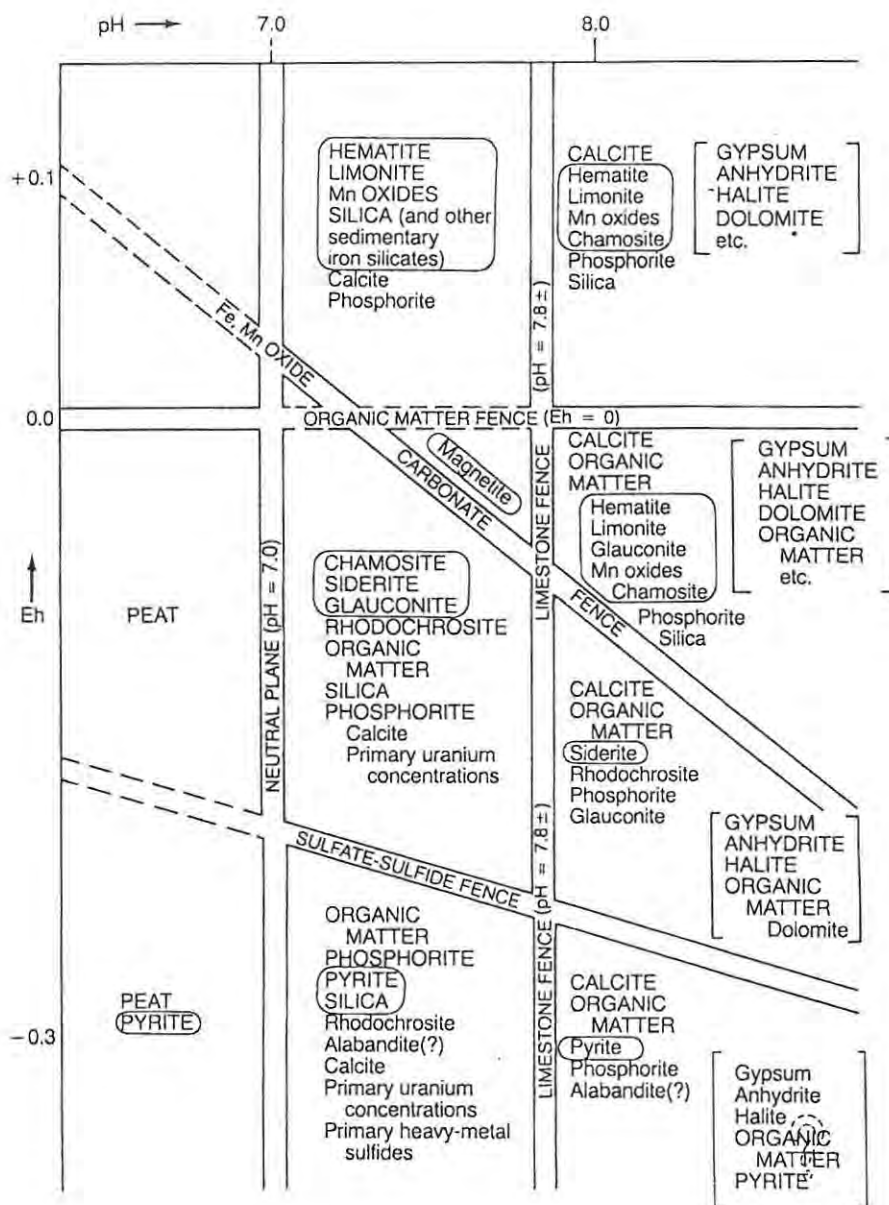


Figure 7: Fence diagram showing the Eh-pH fields in which chemical end members of the non-clastic sediments are formed under normal seawater conditions. (after Krumbein and Garrels, 1952 in Guilbert and Park, 1986).

examples of i-fs deposited under reducing and oxidising conditions. The darker coloured ferrous (reduced) iron was transported and deposited when oxygen was absent and the 'rust-like' ferric (oxidised) iron was deposited when oxygen was present.

However, for rocks laid down in the mid to late Proterozoic ($\leq 2\text{Ga}$) nearly all the iron deposited was '*fully-oxidised*' and exhibits the '*rusty*' appearance characteristic of sediments formed in the presence of oxygen. These are exemplified by the Penge i-f and the Manganore i-fs of the Transvaal Sequence, South Africa, the Hammersley Brockman i-f in Western Australia.

The deposition of i-f terminated abruptly at 1.8Ga, to be replaced by sediments known as '*red beds*' suggesting that the earth had accumulated a significant concentration of oxygen.

In summary it can be said that although the photosynthesisers started generating oxygen earlier in the Archaean, for millions of years other molecules, namely reduced sulphur, reduced carbon, and volcanic gases such as hydrogen constituted '*sinks*' for the oxygen and as a result no net accumulation of oxygen in the atmosphere occurred as the molecules became oxidised. Only when most of the *sinks* had been oxidised, did oxygen begin to accumulate, and once it began to accumulate, it did so rapidly because there was nothing to prevent it (Margulis, 1981).

7. TYPICAL IRON-FORMATION DEPOSITS

The i-fs of southern Africa occur in four types of tectono-sedimentary units, namely greenstone belts of the Kaapvaal and Zimbabwean cratons; the Limpopo mobile belt; cratonic³ (?) basins of the Pongola, Witwatersrand, and the Transvaal Sequences and the Shushong Group; and the Damara belt (figure 1). The Algoma-type i-f occur in greenstone belts and the Limpopo mobile belt. The i-f that form the Buchwa and Manisi

³The term *cratonic* is used here to mean a basin on the edge of a craton, open to the sea and floored by basement rocks including greenstones. The mode of basin formation can be rifting, thermal subsidence etc.

Ranges typifies the Algoma-type and will be described. Canadian deposits will also be described as a comparison since it is the only place in the world where these i-fs are currently being exploited as sources of iron ore. The cratonic basins host the Superior-type i-fs. Only the Transvaal basin will be discussed.

7.1 The Algoma-type iron-formations

The Adams and Sherman Deposits, Ontario, Canada

The Adams and the Sherman are concentrating-grade i-f deposits currently being mined. They are part of the Archaean Abitibi Greenstone belt. The Adams orebody which is situated south of Kirkland Lake, Ontario consists of thinly laminated magnetite-quartz beds of the Boston i-f. The i-f varies in thickness up to several hundred metres and covers a strike length of 10 km. The in-situ grade averages 27%Fe. The i-f is magnetically upgraded to provide pellets at 62-63%Fe. Mining scale averages 1Mt per annum which is railed 2000 km to the smelter at Pittsburgh in Pennsylvania.

The ore contains minor locally developed concentration of pyrite, haematite and jasper all of which have been deformed in the form of mild contortions and soft sediment slump structures. Chemical analysis shows minor amounts of *exhalite-related elements* - Ni, Cr, Mn, Ba and trace amounts of Au indicating that the source of the elements was fumarolic activity associated with submarine volcanism.

ADAMS MINE

Geology

The i-f rests on massive flows of *pillowed* tholeiitic basalt containing minor lenticular i-f intercalations. An

inconsistent well-bedded felsic tuff overlain by a layer of nodular and fine-grained disseminated sulphides sometimes separates the i-f from the mafic volcanics. The tuffs are very variable in thickness sometimes reaching tens of metres thick and also containing i-f intercalations. The sulphides range in thickness from several centimetres to 10 metres. The i-f is *lean*, cherty and of low iron and largely ferruginous clastics at the base developing into finely laminated interbedded chert and magnetite i-f up the sequence. The amount of magnetite and thickness of the magnetite layers, in comparison to chert, increases up the sequence. Minor amounts of haematite occur in the magnetite bands to give them their reddish colour. Garnet, tremolite, actinolite (a bluish amphibole), pyrite and chlorite also occur in the i-f. The top of the i-f grades back into *lean* i-f again, averaging 5%Fe.

A graphitic, sulphidic tuff horizon, up to 3 metres thick rests on the i-f; and is overlain by a 200-300 metre thick *chert-derived* quartzite. The quartzite forms the floor for the fissure-volcanism-derived mafic and ultramafic flows that gave rise to *komatiites* with spinifex texture. This last volcanism is not related to the previous felsic volcanism (Guilbert & Park, 1986).

Mining

There are eight orebodies at the Adams Mine. In the Peria Pit most of the i-f layers whose true thickness ranges 30-50 metres have been thickened considerably by regional folding and brecciation due to slumping during and after deposition. The largest deposit whose dimensions are 1 km long by 200 metres thick is mined at South Pit. Guilbert & Park (1986) proposed a relationship between the Adams i-f (ore) and the proximal submarine volcanism.

SHERMAN MINE

Geology

Sherman Mine occurs at Temagami (Boyum & Hartviksen, 1970). It consists of well developed alternate grey and white bands of magnetite and chert. The i-f is magnetite-quartz-rich, contains more jasper, less pyrite and is mildly metamorphosed to greenschist facies in comparison to the Adams i-f. The i-f lies stratigraphically between layered flows and volcanoclastic rocks. The basal rocks include andesitic rocks that have been fractured to the verge of brecciation and silicified by hot spring fluids, presumed to have entered the basin at the time of deposition. The i-f are finely laminated displaying soft sediment deformation (Guilbert & Park, 1986).

Mining

The i-f averages 60 metres in thickness reaching a maximum of 100 metres for a total strike length of 2 km. Like Adams deposit the i-f is also intercalated with felsic tuff which have now been metamorphosed to chlorite and stilpnomelane. However, the degree of deformation is higher. The rocks have been folded into a synclorium with the limbs dipping at 75° towards each other. The ore grade averages 25-30%Fe and is beneficiated by magnetic separation to pellets grading 65%Fe. One million tonnes are exported annually.

7.2 The Superior-type iron-formations

The Transvaal and Griqualand West sub-basins, South Africa

The Transvaal basin is made up of the Transvaal and the Griqualand sub-basins. The gross stratigraphic sequences in the two sub-basins are similar consisting of a basal mixed siliclastic and volcanic unit unconformably overlain by a

chemical sedimentary unit which is in turn unconformably overlain by a mixed chemical-volcanic-siliclastic rock unit (Beukes, 1983).

The i-fs of the Transvaal-Griqualand West belts of South Africa occur in the lower part of the Transvaal Sequence, overlying a thick dolomite sequence (Beukes, 1973) and andesitic lavas in places. The total length of outcrop, including the gap between the north end of the Griqualand West belt and the Transvaal Basin, is about 1200 km.

There is an apparent better development of the i-fs in the Griqualand West area than in the Transvaal. The i-fs occur from near the base of the chemical sedimentary Ghaap Group to near the top of the Postmasburg Group. The Kuruman and Griquatown i-fs which make up the Asbesheuwels Subgroup are the thickest, reaching 1 000 metres in places. The only i-f known in the Transvaal sub-basin is the Penge iron-formation (up to 650 metres thick) which is a correlative of the Asbesheuwels Subgroup (figure 7), and a very thin unit of restricted lateral extent near the base of the Pretoria Group in the western Transvaal (Beukes, 1983).

Shale from the sequence has yielded a Rb-Sr age of 2263 ± 85 Ma (Hamilton, 1977), which is close to the mean of the maximum of 2643 ± 80 Ma determined on zircon from the underlying lavas (Van Niekerk and Burger, 1978) and the minimum of 2095 ± 24 Ma determined by the Rb-Sr for a post Transvaal intrusive (Hamilton, 1977). Deposition of the Transvaal Sequence and its i-fs probably began around 2460-2500Ma ago and ended prior to about 2100Ma ago (Beukes, 1983).

Stratigraphy

The Kuruman and Penge i-f are intimately associated with and form an integral part of a larger and more significant period

of chemical sedimentation within the Transvaal basin.

The Kuruman i-f overlies the Campbell Rand Dolomite with a gradational contact consisting of intercalations of lithologies characteristic of each (figure 8). The transitional zone consists of alternating layers of

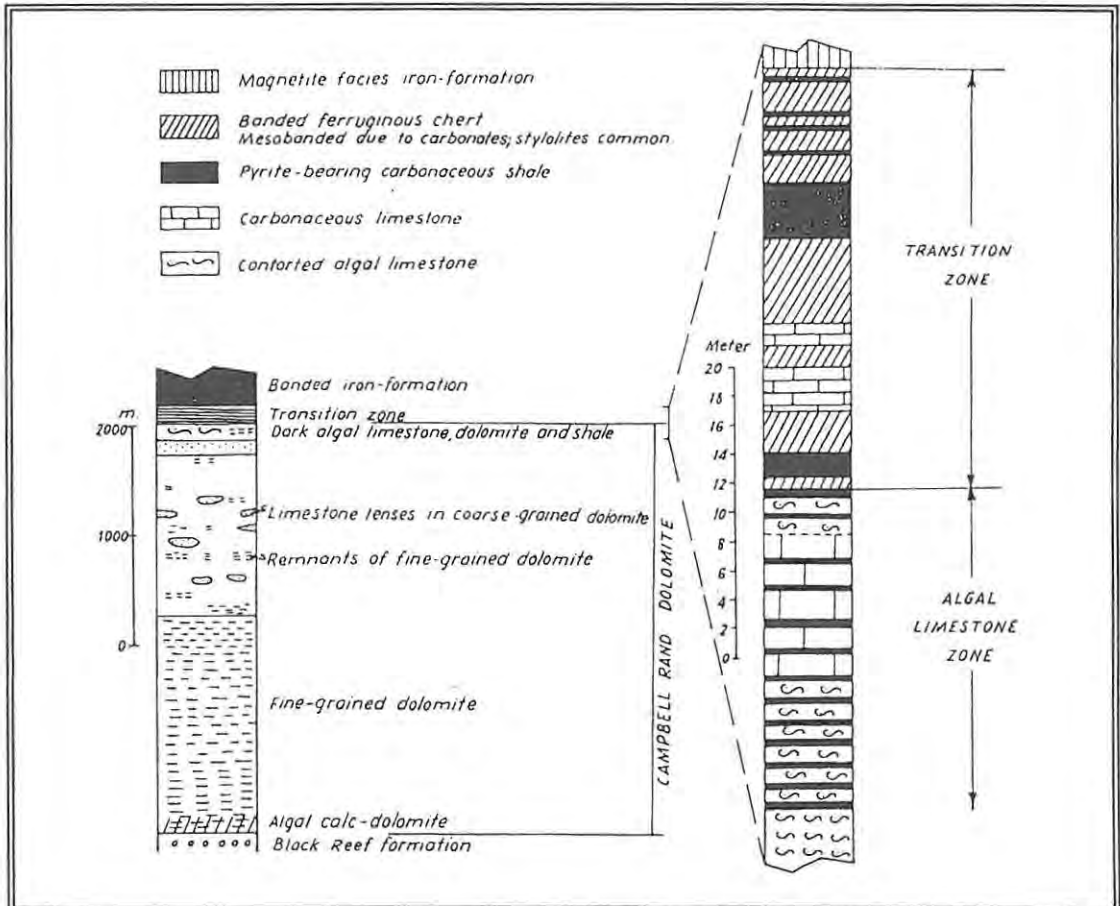


Figure 8: The Relation of the Kuruman iron-formation to the Campbellrand Dolomite (after Beukes, 1973).

carbonaceous limestone, banded ferruginous chert, and pyrite-bearing carbonaceous shale. Chert layers become more abundant progressively upward through the transitional zone, and concomitantly their iron content increases and the limestone interlayers become less abundant (Beukes, 1973). The presence of sulphides and carbonaceous shale indicates deposition in a stagnant reducing environment. The occurrence of chert indicates felsic volcanism which gradually changed to being

mafic which culminated in the Ongeluk lavas. The carbonaceous shales was probably initially deposited as a tuff.

The top of the Campellrand Dolomites is marked by a zone of dark algal limestone and carbonaceous shale. The algal bands in the limestone are highly contorted and the interstitial spaces are filled with white calcite. The information on the Penge i-f-Malmani Dolomite is sparse but contorted algal limestones occur near Zeerust and the top of the Malmani Dolomite in the Penge area is characterised by a carbonaceous shale which is followed by a mixed zone (or transition zone) consisting of dolomite, limestone, chert, i-f, and carbonaceous shale (Button, 1972).

According to Beukes (1973), the Malmani and Campbellrand Dolomites are typical shelf-carbonate deposits. To him (Beukes, 1973), an important application of the carbonates is that the environment of deposition of dolomites and that of the associated algal structures can be directly correlated with post-Cambrian and even present day environments of carbonate deposition. This then might imply that there is no reason to seek a totally different environment from that of the present day to explain the deposition of the overlying i-fs. This supposition is not feasible. Precambrian terrains and therefore basins were very unstable characterised by regressions and transgressions. The deposition of algal dolomites indicates a regression while the deposition of cherts and carbonaceous shale indicates a transgression. Intercalations correspond to minor transgressions before the onset of the major transgression/s which gave rise to the deposition of ferruginous cherts and finally i-f as the basin deepened with the progression of transgression/s.

Kuruman i-f consists of evenly banded i-f which grades into a *Main Marker bed* characterised by pinch and swell structures within jasper and chert mesobands which average 5cm in thickness. It reaches a maximum thickness of 15m where it

grades into a sedimentary breccia consisting of flat lying, angular to subrounded discs of jasper and chert in a fine grained jasper-magnetite matrix (Beukes, 1973). This might indicate the position of an old fault which was reactivated by sedimentary loading.

The main marker is overlain by evenly banded magnetite i-f which grades upward into banded, brown jasper, by a progressive decrease in the magnetite content.

Structure and metamorphism of the strata

On the Kaapvaal craton, the Transvaal strata are gently folded. The folding is related to large open dome and basin structures in the Archaean granite-greenstone basement (Beukes, 1983). Some of the domes and basins affected deposition of the Transvaal strata, especially during the early stages of its development. As deposition progressed, the basement topography and structural highs and lows tended to become smoothed out (Tyler, 1979). In the Transvaal area, the structural configuration was modified by the intrusion of the Bushveld Complex (Beukes, 1983). Sagging of the Transvaal strata took place below the Bushveld Complex resulting in a saucer shaped Transvaal basin. The floor of the Bushveld Complex cuts across the Transvaal sequence (Button, 1976b) and stoping of the roof led to the incorporation of large xenoliths of Transvaal strata into the complex. The degree of metamorphism decreases away from the floor. The Penge i-f outcrops within the limits of the metamorphic aureole of the Bushveld Complex and has been altered to a grunerite-bearing i-f (Beukes, 1973). Outside the Bushveld metamorphic aureole, the grade of metamorphism is of the lower greenschist facies related to burial metamorphism. Along the western margin fold belt there is an increase in the grade of metamorphism to upper greenschist and lower amphibolite facies.

However, most of the i-fs of the Griqualand West area have

been very little affected by metamorphism or structural deformation (Beukes, 1983).

8. DISCUSSION

The absence of stromatolites is consistent with the interpretation that the i-fs were deposited in water deeper than about 100 metres (Walter and Hofmann, 1983). I-fs lack any evidence of the former presence of benthic photoautotrophs, and with their finely banded nature and the lack of evidence for reworking (a characteristic of peloidal i-fs), Archaean-Early Proterozoic i-fs can be interpreted as having been deposited below both wave base and the photic zone. The big question on how the well developed alternate banding between silica and iron compounds came about is avoided. '*Symbiotic*' growth of the iron compound and silica on the ocean floor is proposed by the author, in which the iron compound (magnetite/siderite) was the host and silica, the guest. The source of iron and silica is assumed to be seafloor volcanism. Density stratification of the fluid into silica-rich and iron-rich layers is assumed. The iron-rich layer would underlie the silica-rich layer for obvious reasons. Quiescent conditions are assumed to have prevailed at the ocean floor although minor turbulence cannot be discarded. Cyanobacteria is assumed to have been concentrated in both layers. Silica would precipitate under anaerobic conditions ($\text{pH} \approx 7.5$; $\text{Eh} \approx -0.3$) forming a cap to the iron-rich layer. Under the same conditions the cyanobacteria produces oxygen as the waste product which only raises the Eh while maintaining the pH with subsequent precipitation of iron compounds (figure 7). The cyanobacteria would be destroyed during the precipitation of the iron. This whole process is assumed to be syn-volcanic. The process should probably be called anti-symbiotic growth because the *silica cover* shields the cyanobacteria from further CO_2 supply from the atmosphere which it so desperately needs for its metabolism and the precipitation of iron oxides suffocates the bacteria since the bacteria would be stuck in the iron oxide '*mush*'. This

theory, which involves cyanobacteria in i-f, deposition finds support from the observation that the dominance of the bacteria came to an end at 1.8Ga, at a time when i-f deposition terminated. When there is insufficient time for the two layers to separate completely, then haematite-jasper-silica banding (gradation) results. For low concentrations of iron within the *volcanically derived fluid*, ferruginous chert (jasper)-silica layers result. Soft-sediment slumping associated with i-fs, mostly if not always, involves the disturbance of silica and not the iron compound. In other words, silica '*fingers out*' and slumps into the iron mineral bands and not vice versa (figure 5).

9. CONCLUSIONS

Three great periods of Precambrian i-f development are >2.6Ga, 2.6-1.9Ga and 1.9-0.6Ga (Bayley & James, 1973).

I >2.6Ga i-f: These constitute the oldest group and are called the *Algoma type*. The group represents several periods of orogeny, volcanism, and plutonism and the i-f deposition is generally enclosed in and closely related to volcanic rocks. The Algoma-type is also called the **SHALLOW-VOLCANIC-PLATFORM (SVOP) i-f** (Kimberley, 1978). Typical examples include the Buchwa, Ghoko-Longwe, Manisi, Nyuni, Ripple Creek Ranges in Zimbabwe; the i-f ranges in Barberton and Murchison Belts in South Africa; i-f ranges in the Abitibi belt in Canada; and i-f ranges in the Yilgarn Block, Australia. All the above mentioned examples are associated with greenstone belts. Thicknesses range between 50-100 metres. The magnetite/haematite ratio is >1. The oxide and carbonate facies iron minerals are interlaminated with chert, jasper, and finely granular quartz. The silicate facies are rare

with the mineralogy spanning greenalite-chamosite-minnesotaite. Carbonaceous sulphide facies are not uncommon. The compositions of the i-f do not vary much at 40-55%SiO₂ and 27-37%Fe. Minor elements (%) include Mn and Ba, trace elements(ppm) Co, Ni, Cu, Cr, As, Sr and Au(ppb) are associated and might be considered to be consistent with the geochemistry of volcanic exhalations (Guilbert & Park, 1986). The i-f mostly occur with greywackes (which are now metamorphosed to chloritic rocks and phyllites) containing a considerable volcanoclastic component. Pyroclastics and lava flow units which range in composition from deeply subjacent basalts to more abundant intermediate to felsic lithologies are also part of the package. The immediate footwall rocks may include altered-fractured, sometimes brecciated zones indicative of pipes or solution conduits. Bostrom and Peterson (1966) showed that iron and anomalously high Cu, Cr, Ni, Pb and Ba are related to volcanic processes thus giving evidence for the postulated Archaean submarine hot springs as sources of iron and other elements for i-fs. The abundance of iron compounds in the thermal brines of the Red Sea Atlantis II Deep and the Salton Sea land further evidence for a volcanically related source.

1.9-2.6Ga: These are referred to as the *Superior-type* i-fs (Bayley & James, 1973) and are the most abundant and economically very significant. They are also termed METAZOAN-POOR-EXTENSIVE-CHEMICAL-SEDIMENT-RICH-SHALLOW-SEA (MECS) i-f (Kimberley, 1978). They are typically cherty

i-f which are characteristically very thick, commonly several hundreds of metres. The magnetite/haematite ratio is still >1 but the proportion of haematite is comparably higher than in the Algoma-type i-f. Iron is considered to be concentrated in the magnetite-iron silicate or as *steel-blue* haematite-iron silicate assemblages rather than as simple oxide-silicate units. The Earth's atmosphere became more oxidising at about 2.0Ga (Lepp & Goldich, 1959; Cloud, 1973; Garrels et.al., 1973; Margulis, 1984). This is coincidentally, the period of major deposition of magnetite-chert; siderite-chert; and carbonate-chert i-fs presumably in reducing Precambrian environments. The *blue-green-algae* and subsidiary stromatolites and algal reefs and colonies make-up the fossil record in the i-fs. The source of iron may still be volcanic because a close affinity to volcanism is demonstrated by the Ongeluk lavas that underlie the Asbesheuwels (or Asbestos Hills) formation in the Griqualand sub-basin, Transvaal Sequence, South Africa (figure 9). Guilbert & Park (1986) are of the opinion that the deposition of the i-fs is not related to volcanism because of the absence of volcanic units at Hammersley Basin in Australia, and in the Lake Superior Basin in Canada and the USA. When the volcanic package is absent, the i-f is usually underlain by carbonaceous shale resting on basement. The carbonaceous shale is usually pyritiferous - an indication of reducing conditions, e.g., Mt Newman in the Hammersley basin, Australia. Absence of underlying volcanics does not rule out the volcanic

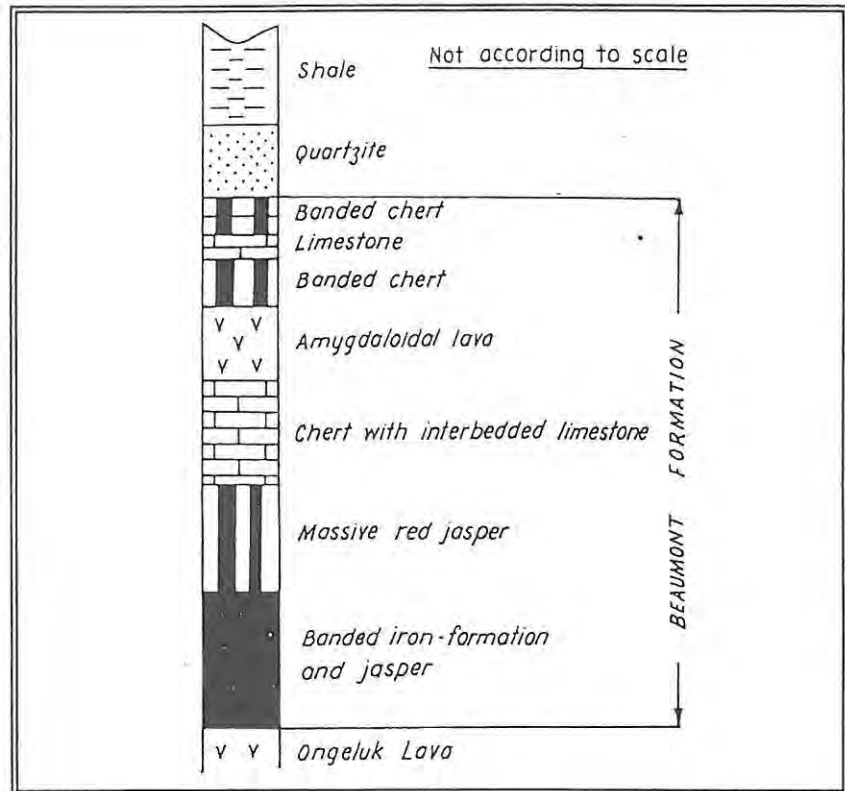


Figure 9: The i-f overlies the Ongeluk lava conformably in the Griqualand West sub-basin (after Visser, 1944 *in* Beukes, 1973).

activity as the source of iron. Sawkins (1984b) in his *episodic dewatering model of basin brines*, shows that hydrothermal fluids can travel for miles before deposition of minerals takes place. This view-point is also shared by Bonatti (1984) who reports that elements which are relatively insoluble in sea water (e.g. Fe, Mn, Ni, Cu and Co) will tend to precipitate on the sea floor after some residence time (estimated at ≥ 50 years for Galapagos) in sea water; implying that hydrothermal plumes impregnated with Fe and/or Mn may be able to travel horizontally for distances up to 1 000 km from the source prior to deposition. The residence time for a given element depends on its solubility, its concentration in the solution, the

temperature of the solution, the rate of admixture and the dilution of the hydrothermal liquid with the sea water. The influence of possibly contemporaneous, but distant volcanism cannot be over-ruled because it has not been reported in the literature. It is possible that the occurrence of volcanics underlying the i-fs has been overlooked. The only difference therefore between the Algoma-type i-f and the Superior-type i-f is the presence of large amounts of oxygen in the system produced by the *blue-green algae* during the formation of the latter. Kimberley (1978) alludes to the fact that the evolution of tectonic and magmatic environments was of more significance than the biological or atmospheric-hydrologic evolution. Deep-water volcanic environments dominated in the Archaean, while continental shelf and inland sea environments prevailed in the Middle Precambrian and Phanerozoic, respectively. All the three tectono-sedimentary regimes influenced i-fs in the Late Precambrian times. Evidence of plate tectonics is present in the Proterozoic to explain the large size of depositional basins. But minimal deposition would occur if the correct Eh and pH conditions within the depositional basin are not met. The right Eh conditions were set by the oxygen produced by the *blue-green-algae*. The oxidation was so effective that most of the sulphides were oxidised. Sweeney & Kaplan's work (1973) on pyrite framboid formation showed that pyrite framboids only form on spherical nuclei. The path of pyrite formation begins with the reaction of

hydrogen sulphide and dissolved iron to produce amorphous iron sulphide or mackinawite ($\text{FeS}_{0.9}$) which can transform to hexagonal pyrrhotite ($\text{FeS}_{1.1}$) under oxygen-deficient conditions. Hydrogen sulphide and iron is sourced from the seafloor volcanism; the *blue-green-algae* serve the dual purposes of providing the spherical nuclei as well as producing the oxygen that oxidises the amorphous iron sulphide to iron oxide (haematite?). Therefore, biological and atmospheric-hydrological evolution also play a major role in the precipitation of i-fs as demonstrated by the occurrence of the *blue-green algae* in the i-fs. Benatti (1984) also mentions that organisms are absent from hot ($>350^\circ\text{C}$) fast flowing brines (e.g. black smokers have flow rates of m/s^{-1}) while cooler ($<300^\circ\text{C}$) slower flowing ones are densely populated by encrusting worms, crabs and other organisms (e.g white smokers). This serves to indicate the presence of blue-green-algae indicates that *near-quiescent* conditions prevailed during the deposition of i-fs.

D. GREENSTONE BELTS

1. INTRODUCTION

Greenstone belts are confined to the Archaean ($\geq 2.5\text{Ga}$). The Archaean rocks, exposed in small areas on all the continents (figure 10), comprise crustal provinces (or cratons) which are roughly equidimensional in plan view and range in size from $<0.1-2.6 \times 10^6$ square kilometres with most falling between 0.25 and 0.50×10^6 square kilometres (Condie, 1981). Archaean provinces exhibit three rock associations: granite-greenstone association, the high-grade association, and the cratonic(?) basin association.

The granite-greenstone association is the most important in iron ore exploration in that it is characterised by supra-crustal successions comprising dominantly of mafic volcanic rocks, engulfed in a sea of granitic rocks. This association dominates in the Archaean provinces in North America, southern Africa, and Australia (Condie, 1981). The Granite-greenstone association will be discussed further.

The high-grade association, which dominates in the Archaean in central and northern Africa, Greenland, and in the Soviet Union, is characterised by gneiss-migmatite-granulite complexes, layered intrusions and high-grade supra-crustal remnants (Condie, 1981). The i-fs found in this association are highly metamorphosed such that the iron oxide mineral is largely magnetite. There are no high grade iron ore deposits associated with these i-fs. The i-fs can be mined and beneficiated by magnetic separation. This association tends to be the oldest of the three associations.

Greenstone belts are linear- to irregular-shaped, synformal supra-crustal successions which range in width from 5-250 km and up to several kilometres in length. Most belts are 10-50 km wide and 100-300 km long, containing 10-20 km thick

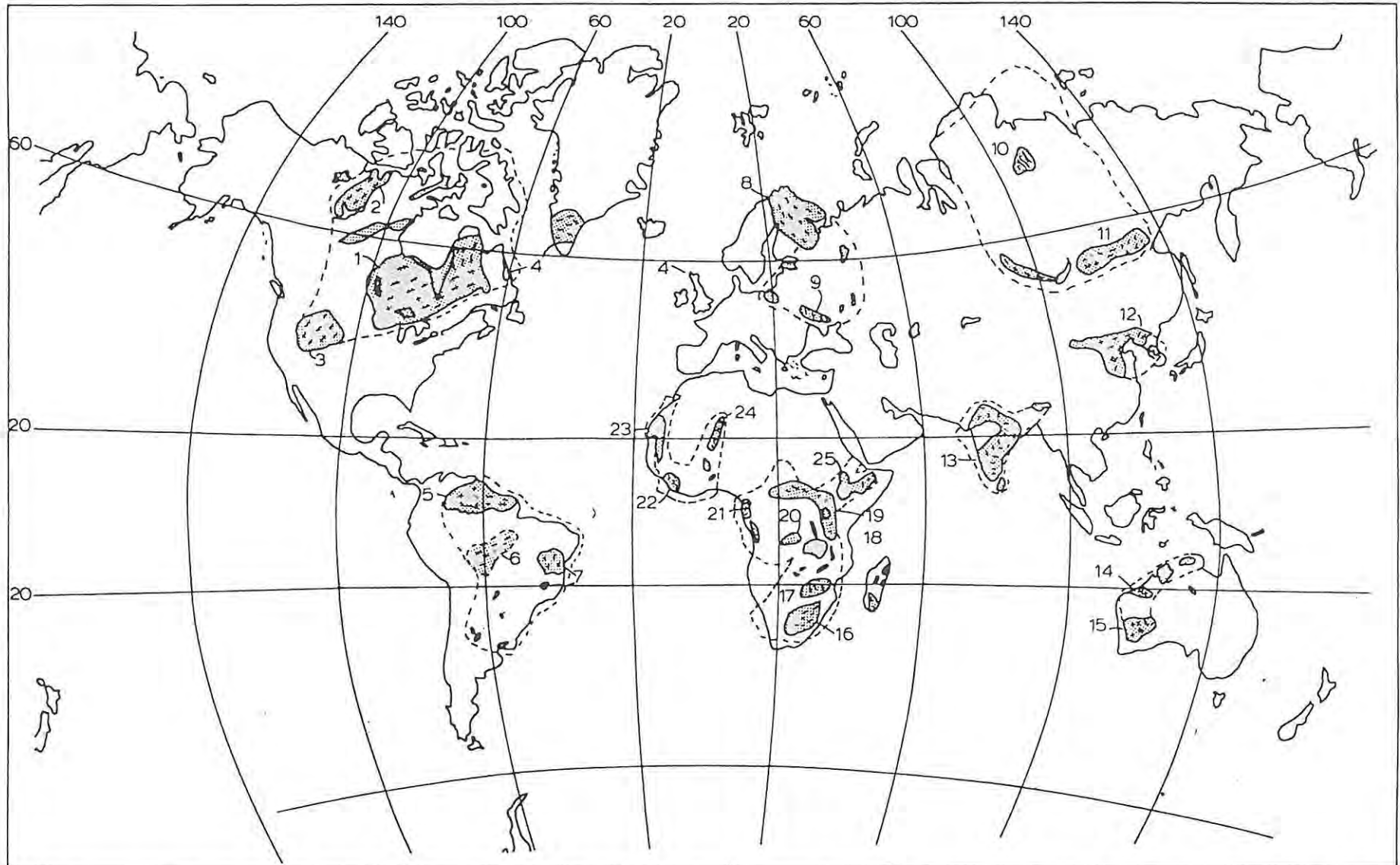


Figure 10: Distribution of Archaean Provinces (exposed - shaded; inferred - dashed line) The key to major provinces (terrains) is shown on the next page (after Condie, 1981).

**KEY TO MAJOR PROVINCES (TERRAINS) REFERRED TO IN FIGURE 10 ON
PREVIOUS PAGE**

<u>Granite-greenstone</u>	<u>High-grade</u>	<u>Granite-greenstone & high-grade</u>
2 Slave	4 North Atlantic	1 Superior
3 Wyoming	5 Guiana	7 Sao Francisco
14 Pilbara	6 Guaporé	8 Kola
16 Kaapvaal	10 Anabar	9 Ukrainian
17 Zimbabwean	11 Aldan	13 Indian
	12 Chinese	15 Yilgarn
	18 Zambian	19 Central African
	20 Kasai	22 Liberian
	21 Cameroons	
	23 Mauritanian	
	24 Ouzalain	
	25 Ethiopian	

(Note: North Atlantic terrain includes Nain, Godthaab and Lewisian).

exposed stratigraphies. The oldest known greenstone belts, the Isua in Greenland, Barberton in South Africa, and Ghoko-Longwe and Buchwa belts in Zimbabwe, are 3.5-3.8Ga old. However, most of the world's greenstone belts average 2.6-2.7Ga in age. Greenstone belts are typically metamorphosed to the greenschist and amphibolite facies with the metamorphic grade generally increasing towards the contact with plutons. The effect of metamorphism is to transform the haematite (which would be more expensive if not impossible to beneficiate), to magnetite, which is easily amenable to beneficiation. Supergene enriched deposits are generally associated with belts of low metamorphic grade. The structure of most greenstone belts tends to be faulted synforms whose parasitic fold axes and major faults parallel the synformal axis. The fold axial region are areas of large orebody formation by supergene or hypogene enrichment. Older greenstone belts (3.5-3.8Ga), although highly metamorphosed, tend to be less deformed as compared to the younger ones (2.5-2.7Ga) and therefore the magnetite and chert are coarse grained, form well developed micro- and/or meso-bands which makes the i-f easily amenable to concentration.

The i-fs range in thickness from a few centimetres to 300m,

averaging 6-40 metres and tend to be more persistent than the enclosing rocks, like carbonates and argillites. Periods of sedimentation are terminated by lava flows, a result of an unstable tectonic environment, and therefore lava flows played a major role in controlling the lateral extent and thicknesses of i-fs. I-fs commonly extending beyond the limits of clastic and pyroclastic horizons. This extraordinary preservation exhibited by i-fs indicates that their lateral distribution is basically controlled by the size of the basin in which they were deposited. Thus i-fs constitute valuable stratigraphic marker beds.

2. STRATIGRAPHY

The stratigraphy of greenstone belts consists of successions of pillowed, mafic volcanic rocks at the base with calc-alkaline volcanic suites increasing in abundance up the sequence in some belts. Some greenstone belts contain an abundance of ultramafic and komatiitic lavas in their lower parts (Condie, 1981) and these usually comprise a minor component of sediments consisting of greywacke-argillite with smaller amounts of chert and other clastic sediments. These greenstone belts with thick ultramafic and komatiitic lavas do not usually exhibit well developed i-f horizons and therefore do not carry iron ore deposits.

Many greenstones belts contain 5-10 major cycles comprising ultramafic or mafic flows overlain by felsic volcanic rocks, capped with sedimentary rocks characterised by an upward decrease in the grain size of greywackes complimented by an increase in the amount of shale, chert and an introduction of chemical sedimentation to form i-fs at the top of the cycle. The thicknesses of the i-fs varies from greenstone belt to greenstone belt. The Belingwe greenstone belt in Zimbabwe has very thin horizon of i-f as compared to the Buchwa greenstone belt which host the largest haematitic-type ore deposit in that country. Both the Barberton and the Murchison greenstone

belts in South Africa, have poorly developed i-f horizons. The thicker the i-f, the more likely it is to host a high-grade *economic* deposit. The thickness of the i-f is inversely proportional to that of the ultramafic/mafic component. Younger sequences, like the Transvaal Sequence in South Africa and the Superior Province in Canada/USA, whose reference as greenstone belts is still controversial, contain very little or no ultramafic/mafic components at the base, and host some of the largest deposits in the world (table 3, after Condie, 1981).

Table 3: Percentages of rock types in Archaean greenstone belts (after Condie, 1981).

	Superior Province		Zimbabwean Province (Western successions)	Karoo Province (Barberton)	Slave Province (Yellowknife)	Yilgarn Province (Kalgoorlie system)
	Abatibi	Wabigoon				
ultramafic- <i>mafic</i>	4	3	10	16	≤1	23
Mafic to felsic	72	81	75	51	73	48
Sediments	24	16	15	33	26	29

Mafic rocks in this category are tholeiite and basaltic komatiite flows closely associated with ultramafic flows.

Of primary economic importance is James' (1954) oxide facies i-f characterised by alternating bands of chert and magnetite and/or haematite. I-fs may grade laterally into banded ferruginous and non-ferruginous cherts (Beukes, 1973). Lenticular dolomites and limestones, carrying stromatolitic structures, ferruginous in places and interbedded with oolitic or nodular cherts, overlie and underlie the i-f, locally. The i-fs in the Tati greenstone belt are underlain by impure limestones and quartzitic rocks, the latter grading along strike into chlorite-tremolite schists which are believed to represent mafic tuffs (Mason, 1970 *in* Beukes, 1973). Tremolite-talc-chlorite schists and tremolite-chlorite schists occur intricately intercalated with i-fs in the ultramafic units of greenstone belts in Zimbabwe. Beukes (1973) concludes that the close association of banded chert, banded ferruginous chert and i-f (*triad*) indicates that they have a genetic relationship. This triad terminates both mafic-felsic and ultramafic units although the triad tends to

be more developed in the latter. Because of the cyclical nature of the volcanism, the triad sometimes forms the base to the volcanics, for example, the Buchwa belt. Those units of the triad that are iron deficient are associated with felsic volcanism (i.e. the tuffs), for example, the *Chert Ridge* at Ripple Creek; and the iron-rich ones are associated with mafic-ultramafic volcanics.

3. GENESIS

It is generally observed that i-fs are persistent litho-stratigraphic units associated with Archaean volcanism in *greenstone belts*. I-fs are thus part of a volcanic ore-forming process and are a result of exhalations. James (1954) who articulated the facies concept of i-fs, recognised that high-energy littoral environments were relatively well oxygenated and characterised by haematite stability and magnetite meta-stability, and that progressively basinward, deeper water, lower energy, more reducing environments were characterised by magnetite, siderite, and pyrite-pyrrhotite stability (figure 13). This representation helps to explain the presence of clastic sediments, ferruginous sediments and volcanoclastic rocks. This setting is well represented at

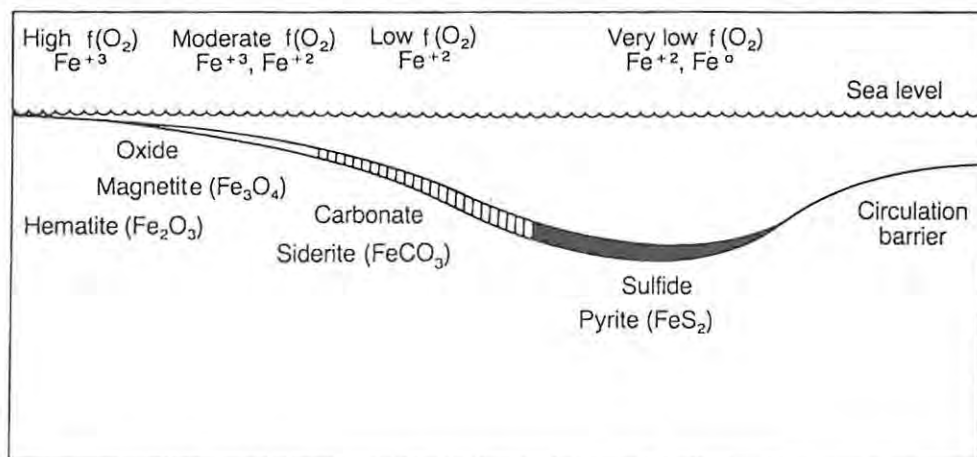


Figure 11: Depositional zones in a hypothetical basin in which iron compounds are being precipitated (after James, 1954).

Ripple Creek, which is in the 2.7Ga old Midlands Greenstone Belt in Zimbabwe, where the whole sequence from the deep water pillow lavas through, tuffs, dolomite, massive sulphides, to the shallow water clastics and finally the i-f exists (figure 12). The conglomerate and sandstone intercalations in the volcanoclastics (tuffs) indicate intermittent short-lived relatively high energy environment conditions - basin margin. Deposition of i-f indicate

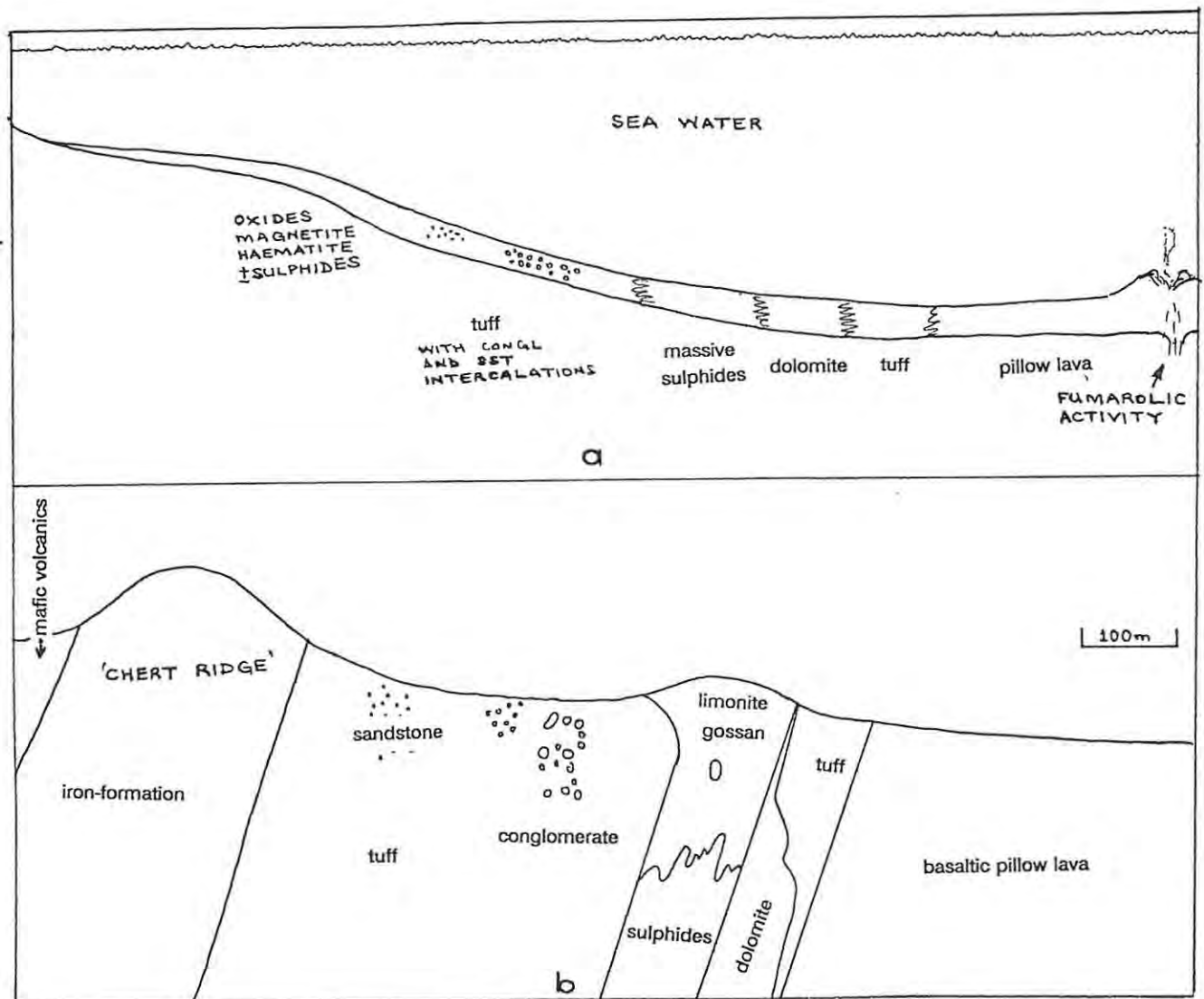


Figure 12: Diagrammatic East-west section of the Ripple Creek area, Redcliff, Zimbabwe showing :-
 a) hypothetical depositional basin;
 b) stratigraphical and lithological relationships.

reverting back to quiescent conditions, i.e., relatively deep water. Deep water conditions are further indicated by the association of the i-f with clastics lacking sedimentary structures other than normal laminar bedding. The i-f contains some pyrite, an indication that currents were present to transport the sulphides further afield. The precipitation of sulphides also indicate low oxygen fugacity, i.e., reducing conditions. The reducing conditions partly serve to explain the iron oxide-poor nature of the i-f. The i-f is therefore referred to as a *Chert Ridge* in the Buchwa Iron Mining Company (BIMCO) literature.

The i-fs consist of two sub-facies of the oxide zone: one characterised by haematite and the other by magnetite, both as primary sediments. The haematite sub-facies normally consists of fine-grained haematite inter-layered with chert or jasper, with common oolitic structures indicating accumulation in a strongly oxidising near-shore environment, like the *Chert Ridge*. The magnetite facies consists of magnetite inter-layered with silica, carbonates, iron silicates, or a combination of these minerals - its mineralogy and associations suggesting weakly oxidizing to moderately reducing conditions, are absent or weakly represented in the Ripple Creek area. The same applies for the silicate facies, composed of hydrous ferrous silicates (greenalite, stilpnomelane, or minnesotaite) commonly associated with either carbonate- or magnetite-bearing-rocks, suggesting that optimum conditions for deposition ranged from slightly oxidising to slightly reducing, probably with mild post-depositional metamorphism. The carbonate facies which is supposed to contain interbedded siderite, or iron-rich ankerite, and chert is usually contaminated by clastic sediments (polymictic conglomerates, sandstones). The carbonate facies forms in an environment of high pH. The Ripple Creek sulphide facies consists of nearly 90% massive pyrite with about 10% pyrrhotite and accessory carbonaceous material. The massive nature of the sulphides and the carbon

content are evidence for the ultra-stagnation and reducing conditions that prevailed during deposition. The top 180 metres of the sulphides have now been oxidised to massive earthy limonite/goethite/stilpnomelane with kaolinite and pyrolusite pockets and intercalations and capped by a one metre thick haematite crust (figure 12).

4. PLATE TECTONICS

Plate tectonics, considered by Sclater et al., 1980, in a fundamental way, as a mechanism by which excess thermal energy from the mantle is dissipated was mainly operational in the Phanerozoic, having probably developed initially in the Proterozoic (figure 13). There is no evidence of plate tectonics in the Archaean despite the occurrence of lithologically similar packages to those found in the Proterozoic. Burkes et al., (1976) postulates the granite-greenstone terrain (craton) formation are related to the ancient subduction processes. The lack of evidence is due to the low preservation potential of the craton since the supracrustal (volcanic) portions of neutral and compressional arcs are rapidly lost to the forces of erosion. Plate tectonics as is evidenced today, started with the break-up of *Gondwanaland* about 180Ma ago.

5. SUMMARY

Goodwin (1973) applied the facies concept with success to study the Abitibi Greenstone Belt and related terranes in Canada. It is also used here to explain the Ripple Creek geology. The differences between the Archaean (Algoma type-) and the Proterozoic (Superior type-) i-fs can also be explained in terms of the facies model. The main observations and conclusions can be summarised thus:

i Archaean (>2.5Ga) i-fs are intimately associated with

greenstone belt volcanic or segments. Those i-fs that occur individually, are commonly associated with the upper horizons of the pyroclastic phases of dominantly tholeiitic to calc-alkaline, mafic to felsic volcanic sequences and nearby turbidite assemblages. Therefore the source of iron can be attributed to chemical components of the volcanic processes.

- ii The common worldwide deposition pattern of iron from shallow to deep water is oxide, carbonate and sulphide occurring on Archaean palaeo-slopes which are part of major Archaean basins (figure 13, after Gross, 1980). The preservation potential is in the order: oxide (the Chert Ridge), carbonate (contaminated by clastic sediments) and sulphide (oxidised to limonite and now occur at a depth >180m from surface) as exemplified by Ripple Creek (Appendix 2). Therefore oxide facies i-fs are the most common, prominent and the most easily recognised in Archaean terrains. On the other hand, the carbonate and the sulphide facies i-fs are scarce although widely distributed, thinner, discontinuous and inconspicuous in comparison to the oxide facies i-fs.

- iii Goodwin (1973) facies plot of known Archaean i-fs in the greenstone belts of the Canadian Shield revealed an existence of several large Archaean basins, which are elliptical in outline - a feature which he attributed to structural deformation. The basins exhibit a triple lithofacies association of oxide-carbonate-sulphide i-f transition, arc-type felsic volcanic rocks and proximal volcanoclastic conglomerates. He assumed that the basins were initially circular and of 750-1000 km in diameter. Whatever the shape of a basin, the facies laid down in a basin are largely a function of the diameter and depth of the basin which in turn determines the depositional conditions (e.g. reducing or oxidising, quiescent or turbulent).

E. ORE GENESIS

1. INTRODUCTION

Volcano-sedimentary or sedimentary deposits contribute more than 95% of the close to one billion tonnes of iron ore mined in the world per annum. About 5% of the world's iron ore production comes from intrusive magmatic segregation ores and extrusive volcanic iron ore.

2. MAGMATIC DEPOSITS (MIOD)

The MIOD are associated with the great calc-alkaline orogenic belts of the world. There is a general relationship between mountain building and ore deposits. The calc-alkaline rocks are generally young and there was a rapid increase in the mass ratio of calc-alkaline rocks to other igneous types - komatiites, tholeiites and alkalic rocks - in the last 200 to 300 Ma.

The MIOD owe their existence to the segregation of a magmatic iron-rich fluid, normally with 4-5% phosphorus pentoxide, 1-2% vanadium pentoxide and 12-20% titanium dioxide from a differentiated complex of mafics and anorthosites. If the ferruginous melt precipitates iron ores at the top of a layered complex (e.g. Magnet Heights, Bushveld Complex, South Africa) or is injected into country rocks as intrusive, pod-like masses of magnetite (e.g. Kiirunavaara mine, Kiruna, Sweden) then the result is intrusive magmatic segregation ores. On the other hand if the ferruginous fluid vents out to surface to be laid down as stratiform, generally stratabound iron-rich flows and tuffs with layered volcanic rocks, typically andesites and latites, as foot-wall and/or hanging-wall, then these are called extrusive volcanic ores (e.g. Pea Ridge and Iron Mountain, Missouri; El Laco, Chile).

The differentiation that produces calc-alkaline rocks, geochemically affects iron enrichment in the same melt. The Mg/Fe ratios decrease and the total iron increases along with other differentiation indices like Ti, V and P. The conditions that work in concert for the Fe-enrichment to occur are:

- i High sodium content leading to the '*alkali-iron effect*', the formation of sodium-iron-oxygen complex ions in late melts, which increase the melt's iron content and fluxing;
- ii Low early fugacity of oxygen, which restricts Fe³⁺ activity in the melt and prevents the progressive extraction of iron by steady magnetite (FeO.Fe₂O₃) separation;
- iii High phosphorus content, which fluxes the melt and permits late, low-temperature mobility of normally high-temperature phases; and
- iv Combination of two or more of these effects to move the melt into a temperature-composition field of two-liquid immiscibility wherein a silica-rich and an alkali-phosphorus-magnetite(-haematite)-rich liquid separate from one another, the later being separately intruded or extruded or the sequence can cool in place soon after segregation (Guilbert and Park, 1986).

Kiruna, Sweden

The high-grade ore deposits of the Kiruna District in northern Sweden (figure 14) of which the Kiirunavaara deposit (figure 15) is the largest, are products of magmatic segregation and are the world's largest deposits of this type. The Kiirunavaara ore is *sandwiched* between Precambrian syenite porphyry foot-wall and quartz porphyry hanging-wall. Highly altered flows and silicified tuffs and sediments overlie the quartz porphyry. The strike of the sequence

ranges from north to north-east and dips at between 50° and 60° east. The strike length of Kiirunavaara deposit is 5 km and averages 90 metres in thickness. About 400Mt, which is equivalent to 25% of the initial reserve prior to the commencement of mining in 1903, were mined at an average grade of 62%Fe by 1980 in order to supply the Swedish steel-making companies. The ore

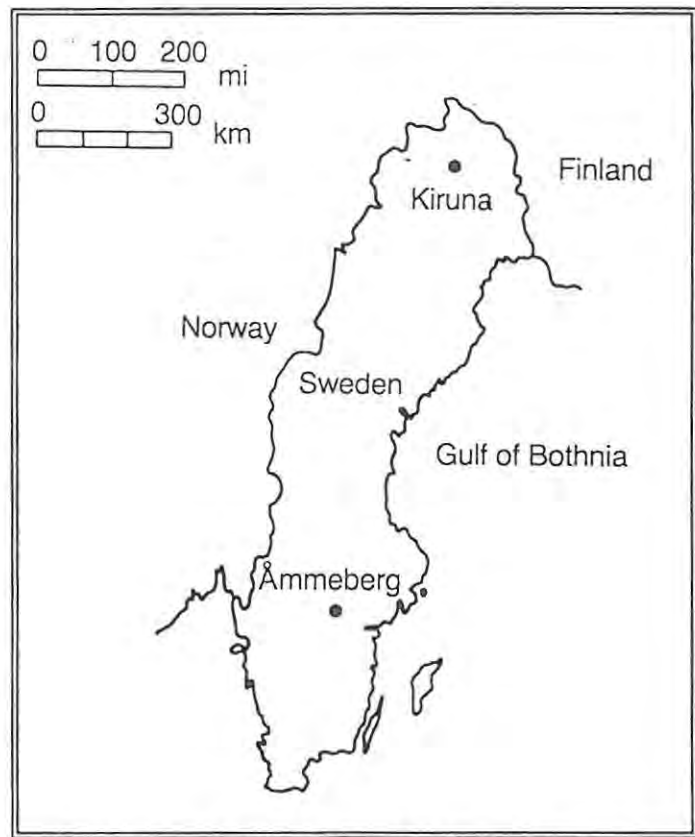


Figure 14: The Kiruna Iron Ore District, Sweden

mineralogy consists largely of fine-grained magnetite with subsidiary fluoro-apatite and accessory haematite which display a well-defined layering/banding. Because of the presence of apatite, the phosphorus content is greater than 2% hence the universal term *Kiruna-type ore* which means *high-phosphorus* iron ores. Gangue minerals include actinolite and diopside. The texture exhibited by apatite prisms is trachytic. Pure apatite and magnetite-apatite are sometimes inter-layered - which is considered by Geijer (1910) to be flow banding.

There is a sharp contact relationship between the ore and the *enclosing* rocks although a thin amphibolite skarn is developed in places and veins or apophyses of ore (locally called *ore breccia*) branch out into both the hanging-wall and foot-wall rocks. Porphyry dykes whose composition and texture are intermediate between the hanging-wall and foot-wall porphyries, intrudes the foot-wall and spread out beneath the

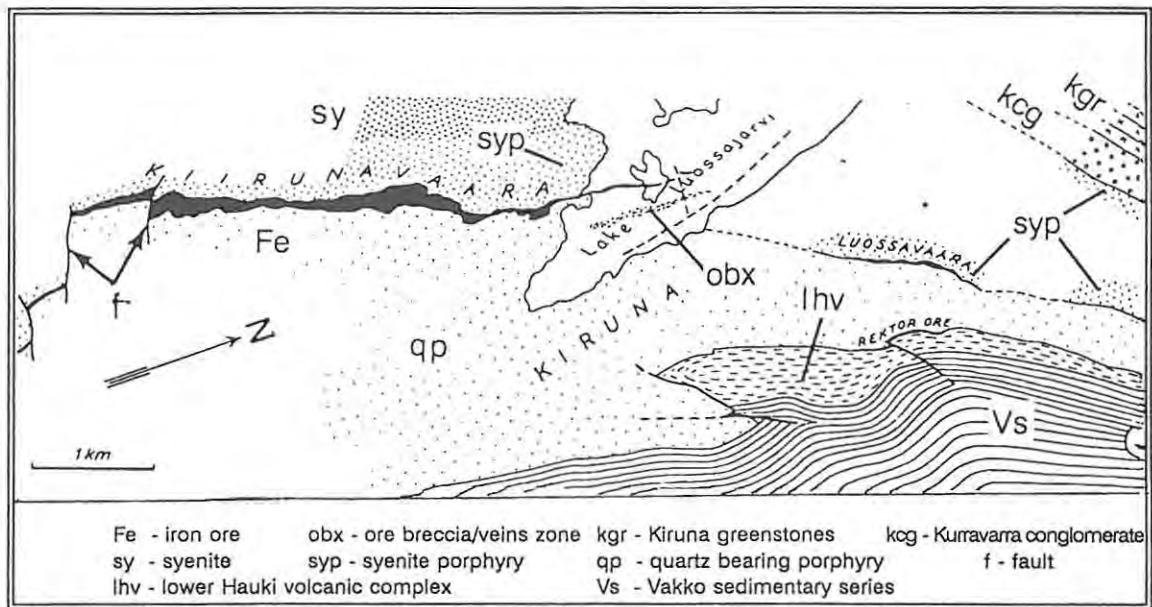


Figure 15: The Kiirunavaara Iron ore deposit (after Geijer, 1910)

hanging-wall. The dykes are also broken and engulfed in the ore. One dyke has been found to intrude the magnetite. The observations show that the porphyry dykes were intruded before the ore and the ore was intruded during the waning phases of the igneous activity that produced the porphyry dykes. The ore has suffered late strike-slip faulting and attendant granophyre dyke intrusion.

The Kiirunavaara deposit was formed as a result of the intrusion of a highly mobile magnetite-apatite fluid ore magma, which became concentrated during magmatic differentiation as an immiscible fraction in the parent magma and was separated deep within the Earth's crust, prior to the intrusion. It is pointed out that the volatile substances played a major role in the forceful injection during emplacement of the magnetite-apatite fluid. The orebodies invade porphyries; magnetites show liquid flowage (trachytic texture) and contact zones underwent slight metasomatism (skarns) indicating presence of volatiles (Guilbert and Park, 1986).

3. HYDROTHERMAL DEPOSITS

Hydrothermal iron deposits (HID) are magnetite-rich iron deposits associated with calc-alkaline plutonism. The mineralisation process involves a magma chamber and therefore the HID form a subset of MIOD. The HID are commercially of minor importance as a source of iron, and have variable modes of occurrence. They are found associated with intermediate composition intrusive units around the Pacific Basin, in a belt 600 km long by 30 km, most notably Chile, central America, Australia, and Japan; they manifest themselves as magnetite skarns; and as replacements in non-carbonate wall rocks.

4. SEDIMENTARY DEPOSITS

4.1 Introduction

Iron ore deposits associated with i-fs are classified according to the dominant mode of their formation. The classification is subjective, and cannot accommodate the ideas of all investigators on the genesis of these deposits. The classification below is an attempt to synthesize best the available data on the iron ore deposits. Three classes are recognised:

1. *the syn-genetic deposits*, to produce the oolitic iron ores, which are generally Phanerozoic in age;
 2. *the supergene/hypogene deposits*, to form the high grade haematitic iron ores, which are currently the major source of iron;
- and 3. *the metamorphogenic deposits*, to form the magnetite quartzite (or itabirite), which are suitable for upgrading.

4.2 The syn-genetic deposits

The syn-genetic iron ore deposits are associated with the i-f deficient, high energy *Upper Clastic Unit* of greenstone belts or Phanerozoic sequences; usually at the base of the unit. The oolitic iron ore deposits occur as lenticular beds composed of closely packed oolites, generally consisting of a quartz core engulfed in concentric layers of chamosite, ankerite, kaolinite, magnetite, haematite, lepidocrocite and goethite (Schweigart, 1965); which are extensively developed in shaly horizons and often with orthoquartzite associations. Apparent sedimentary structures include cross-bedding, ripple marks, and rarely, mud cracks. Oolitic iron ore deposits are deposited in an intermediate environment marking the transition from quiescent deep water conditions which favour the deposition of chemical sediments like dolomite and i-f, to high energy shallow water environment, which is accompanied by detrital input. Under these conditions the oolites will form in the zone of mixing of reduced, iron-rich, deep oceanic waters and oxygenated, near-shore waters.

Examples of these deposits are best developed at the base of the Pretoria Group in the Potchefstroom area where lenses of up to 8 metres in thickness were observed (Wagner, 1928). Quoted opencast mineable reserves stand at >4 500 Mt at 45%Fe and 20-25%SiO₂ (Wagner, 1928). The high silica content make it impossible to be a direct blast furnace feed. The smelter maximum tolerable silica level is 9%. The nature of the ore (that is, near-total mixture of iron minerals, haematite and magnetite, with the gangue mineral, silica) makes it economically impracticable to beneficiate, because of losses experienced. For example, Wagner (1928) found that 4½ tonnes would yield one tonne of concentrate at 55%Fe and 10.7%SiO₂. Oolitic iron ores have, therefore, lost their exploitability to the high grade, low silica haematitic type ores. Blending of the oolitic and the high grade haematitic type ores may provide a solution to the workability of oolitic iron ores.

4.3 Supergene/hypogene deposits

a. Introduction

I-f host the high grade haematitic iron ores, which represent the most economically viable type of deposits. The deposits are generally developed at the lower portion of the i-f. The attitudes of the deposits vary from shallow dip ($\approx 10^\circ$) to quite steep ones ($\approx 70^\circ$). The ore horizons are usually repeated by strike slip faulting. With increasing depth, the ore tends to finger out to barren i-f or the underlying rock-type.

b. The enrichment process

The enrichment is considered to be due to the leaching out of silica bands in the i-f by meteoric waters - *supergene enrichment*. The leaching of silica bands is supported by the collapse structures in the thinly laminated haematite induced by the reduction of the rock volume on dissolution of the silica. The general *carrot* shape of haematite supergene bodies indicates depletion of the leaching power of the meteoric waters with depth as they become saturated with silica (figure 16). The porosity of the i-f due to faulting, dolerite dyke intrusion or recrystallisation due to metamorphism are considered to play major roles in channelling the meteoric waters and therefore, determine the size and position of supergene enriched orebody.

'Massive' haematite *pockets* within the supergene-enriched body are considered to be due to rising iron-rich fluids (or connate waters) to *cement* the porous, banded, silica-leached i-f, with subsequent increase in the iron tenor, a decrease in the gangue mineral (silica) component, and an increase in the specific gravity of the ore. This is *hypogene enrichment* - a process which is considered to complement supergene enrichment rather than hypogene enrichment being primarily

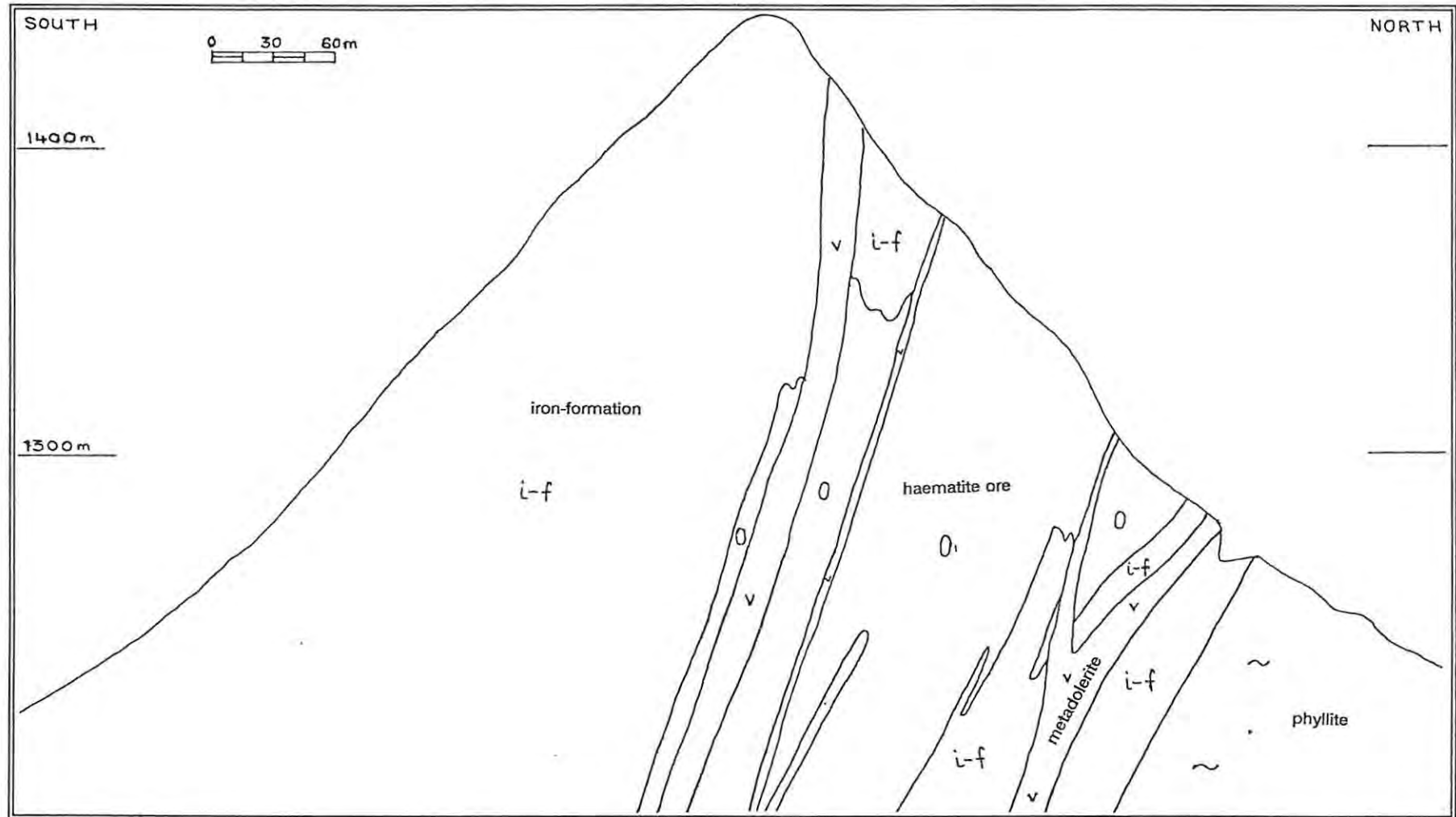


Figure 16: Cross-section of the Buchwa West iron ore deposit, Zimbabwe (COURTESY OF BUCHWA IRON MINING COMPANY).

responsible for the enrichment.

In South Africa, haematitic orebodies are developed in the Penge i-f at Thabazimbi (Van Deventer et al., 1986) and in the Manganore i-f (equivalent to the Asbesheuwels i-f) at Sishen (Van Schalkwyk and Beukes, 1986). In Zimbabwe, the Buchwa (Worst, 1962a; Bourhill, 1978) and the Manisi (Worst, 1962b; Gapara, 1989) haematite deposits are both hosted in i-fs. A brief resumé of the regional geology for each of the deposits is presented.

c. The Thabazimbi Deposits, South Africa

The haematite iron ores at Thabazimbi, occur at the base of a Superior-type i-f, the Penge Formation. The four deposits rest on the lowermost shale unit, which in turn, is underlain by the Malmani sub-group dolomites (Figure 17). Both the host rocks and the deposits dip at about 50°S. The haematite orebodies grade into the low grade deposits comprising talc-haematite and carbonate-haematite along the dip direction (Van Deventer et al., 1986). Tectonic activity caused the

Fe ₂ O ₃	SiO ₂	Al ₂ O ₃	MnO	K ₂ O	TiO ₂	CaO	MgO	P ₂ O ₅	S	Na ₂ O	Total
89.3	4.85	0.76	0.35	0.09	0.06	0.85	1.00	0.046	0.009	0.01	97.6

Fe = 62.3% ; max Al₂O₃ = 1.48

duplication and/or triplication of the Penge Formation in the Thabazimbi area (Strauss, 1964b). Tectonism is also manifested in the ores as haematite breccia consisting of haematite fragments set in a matrix of haematite of probably secondary origin (Strauss, 1964b; Van Deventer, et al., 1986).

The ore grades are >60%Fe and <15%SiO₂. The average chemical

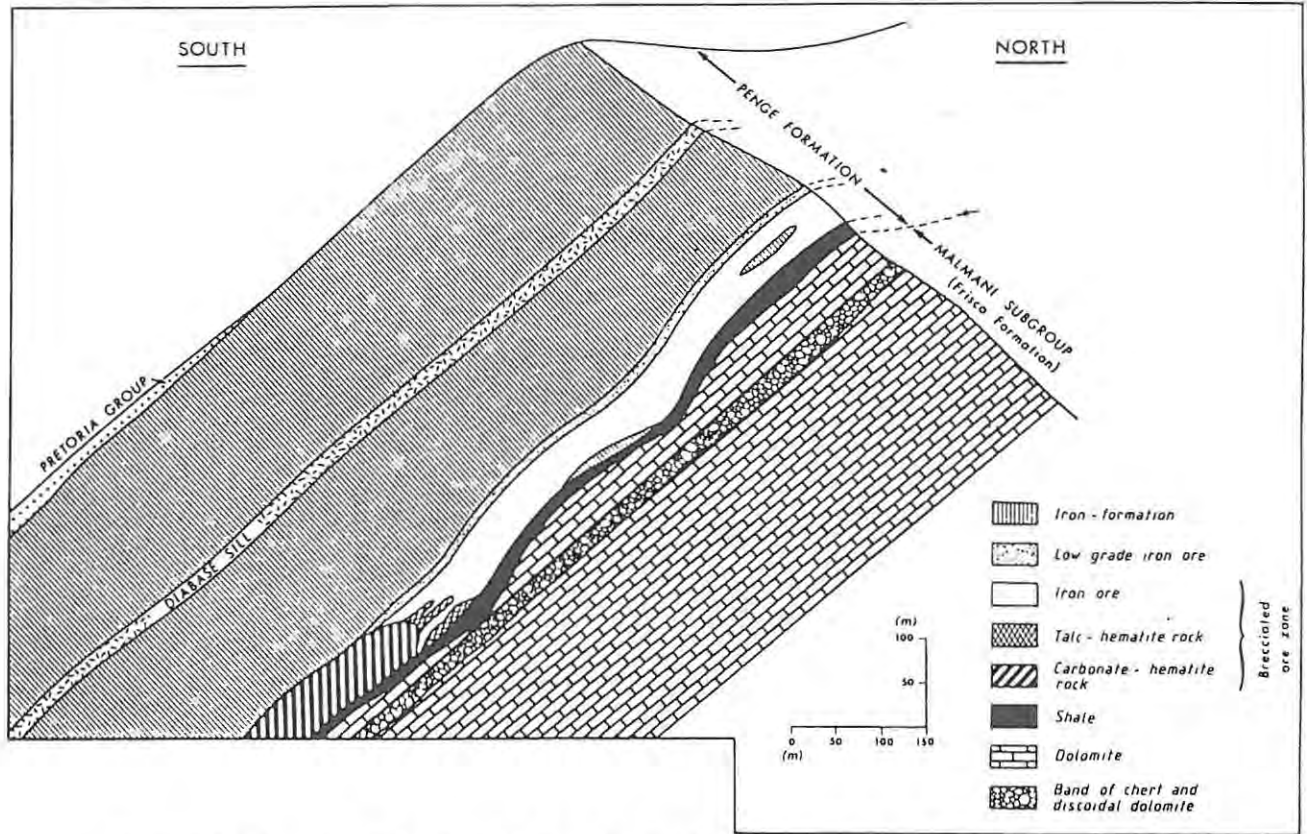


Figure 17: Profile through the haematite orebody in the Northern Range, Thabazimbi, showing the geology (after Van Deventer, et al., 1986).

composition of iron ore is shown in table II, below. The four orebodies which occur over a strike length of 12 km are separated by tracts of i-f. The width ranges 2-100 metres, averaging 20 metres.

The genesis of the deposits involved supergene enrichment and subsequent metamorphism of the i-f. Contact metamorphism and the tilting of the strata to the south were caused by the intrusion of the Bushveld Complex. Tectonic activity, ascribed to the 'Waterberg event' caused contact metamorphism of i-f to produce the talc-haematite and carbonate-haematite suites. The underlying dolomite underwent karstification above the water-table giving rise to collapse breccias of the overlying i-f. Both the carbonate-haematite and the talc-haematite were then subjected to epigenetic ferruginisation which involved slow supergene replacement of chert, calcite and talc by goethite and haematite (Van Deventer et al., 1986). It is considered that the basal shale horizon (figure

17) acted as an impermeable layer to downward percolating iron-rich meteoric and ground waters which resulted in the deposition of iron above it. It is considered by Van Deventer (1986) that Post-Karoo fault zones filled with iron ore indicate a second period of supergene/hypogene ferruginisation. Dolerite (diabase) sills which are related to this post-Karoo extensional deformation, have hardened the ore where they dissect the orebodies.

d. The Sishen Deposits, South Africa

The Sishen iron ore deposit, like Thabazimbi, also occurs in the Transvaal Sequence. Situated in the Northern Cape Province, the Sishen deposit, whose reserves stand at over one billion tonnes at $\geq 65\% \text{Fe}$, occurs in the Manganore i-f (of the Asbesheuwels subgroup) which rests on a dolomite sequence, the Campbellrand subgroup. The Asbesheuwels subgroup is unconformably (?) overlain by a clastic sequence, the Gamagara formation of the Olifantshoek Group (Strauss, 1964a; Van Schalkwyk and Beukes, 1986).

Three types of haematite deposits are recognised in the area. They include:

- a) the haematised Manganore i-f;
- b) haematised basal shales of the Gamagara Formation;
- c) haematised conglomerate of the Gamagara Formation.

The first type of orebody was formed by supergene enrichment of the Manganore i-f (similar to the Thabazimbi ore genesis) to form a 25 metre thick and 450 metre long, lenticular orebody which tapers out in all directions within the i-f. The ore is massive to thinly laminated with minor breccia structures developed. Several million of tonnes are present.

The second type of ore consists of basal shales haematised to high grade, hard, finely laminated haematitic type ore

(called *laminated ore*). Its thickness averages 25 metres. The third type ore, termed the *conglomeritic ore*, consists of conglomerates which have also been haematised. The conglomeritic ore, whose thickness averages 10 metres, conformably overlies the laminated ore with a sharp contact. The ore is of such a high grade that i-f (*i.e. low grade or waste material*) material is added to the Sishen ore in order to obtain the right slag volumes in the blast furnaces. Total opencast mineable reserves are over a billion tonnes, thus dwarfing the underlying supergene/hypogene enriched haematitic-type ore of the Asbesheuwels Subgroup.

The haematisation of the clastic sediments of the Gamagara Formation, to give rise to the laminated and the conglomeritic ores, is considered by the author to have been syntectonic precipitation of iron derived from fumarolic activity related to the volcanic activity that gave rise to the Ongeluk lavas. The Ongeluk lavas are pillowed indicating that they were extruded under water. They are characterised by extreme iron oxidation with calcium depletion and alkali enrichment. Ca-depletion and geochemical evidence suggest an extreme case of sea-floor spilitisation which involved the development of a large convection system involving large-scale movements of material (Schütte and Cornell, 1992). Although the main source of iron was fumarolic activity, iron was also leached from the Ongeluk lava at an early stage of sea-floor alteration and deposited in an oxygenated marine environment after expulsion from a hydrothermal vent. These iron-rich waters, driven by the hydrothermal circulation, haematised the Gamagara shale and conglomerate. The convection cells were depositing manganese proximal to the vent in the deeper and alkaline areas of the basins (in the Kalahari Manganese field in the north) and iron in the distal and slightly acidic areas of the basin (in the Sishen area in the south). The silicified chert breccia which rests on the dolomite at the base of the Gamagara indicates the existence of faults in the basin. The chert breccia is viewed as a

thrust breccia (Strauss, 1964). The gradual decrease in ferruginisation of the overlying shale, flagstones and quartzite indicate the depletion of the iron in the deep waters or regression of the sea prior to the onset of glaciation which gave rise to the diamictites of the Dwyka. The haematisation is considered to have occurred during or soon after extrusion of the Ongeluk lavas ($\approx 2.2\text{Ga}$) but after the deposition of the Manganore i-f and the dolomites (2.5-2.3Ga). It is proposed that the deposition of the Gamagara Formation was therefore syn-tectonic - in order to explain the rapid facies change in the basal units and the angular unconformity between the Transvaal and the Gamagara strata.

e. The Buchwa Deposits, Zimbabwe

The Buchwa iron ore deposits occur in the Buchwa greenstone belt belonging to the Bulawayan Group and consisting mainly of ultramafic rocks which are poorly exposed in the region. The belt is in contact with the northern zone of the Limpopo Mobile Belt. Metamorphism is in the greenschist facies. The general geology consists of a basal sedimentary suite overlying spinifex textured *greenstones* and a volcanic suite of ultramafic composition. The whole package rests on Archaean basement granite and has a southerly younging direction.

The sedimentary suite consists of quartzite overlain by phyllite, which in turn, is overlain by i-f. A thin veneer of phyllite separates the i-f from the overlying volcanics. The i-f forms the Buchwa Range of mountains, a topographical expression of the extremely hard and resistant i-f rocks. The quartzites are white coloured, fine to medium grained and show cross-bedding. They are confined to the northern portion of the range where they form 'foot' hills around the mountain. The quartzites are separated from the range by the phyllites, a fine grained, finely laminated, metapelite which

varies in colour from white to dark green to brown and reddish colour, where stained by i-f. The phyllite appears to be a chlorite-talc schist.

The ultramafics consist of silicified serpentinite, talc, tremolite, and talc carbonate schists. Late dolerite dykes, which are now metamorphosed, make a network of veins in the i-f.

The Buchwa i-f hosts eight high grade haematite orebodies, which are steeply dipping tabular bodies, confined to the same stratigraphic horizon, and separated by i-f patches. The deposits are namely, West Primary, Middle Primary, Whikwi, Summit, Bicycle, Chief, Manyanga and Neck. The West Primary deposit, the largest of them all, is currently being mined at a rate of close to 1Mt/annum at a grade of $\geq 60\% \text{Fe}$ and $\leq 10\% \text{SiO}_2$. It contained $\pm 30\text{Mt}$ opencast mineable reserves before the commencement of mining in the *seventies*. Whikwi Primary deposit was mined for some time and was stopped because of its high silica and alumina contents coupled with a large fraction of fines that were generated during comminution.

The mineralised portion of the West Primary orebody is about $\frac{1}{2}$ km long on surface with an average width of about 100 metres. The dip of the orebody, 63°S , is steeper than that of the enclosing i-f at $45^\circ - 50^\circ$. The orebody footwall contacts with the phyllite are concordant with the bedding and are shallowly dipping. The hanging-wall contacts with the i-f host are discordant, being steeper than the dip of the i-f.

The ore mineralogy consists largely of haematite with subsidiary magnetite and accessory specularite, pyrite, and malachite. The dolerite dykes, which are older than the Mashonaland dolerites, ubiquitous in Zimbabwe, divide the mineralisation up into a large orebody with subordinate lenses (figure 18). Ore genesis is thought to be largely

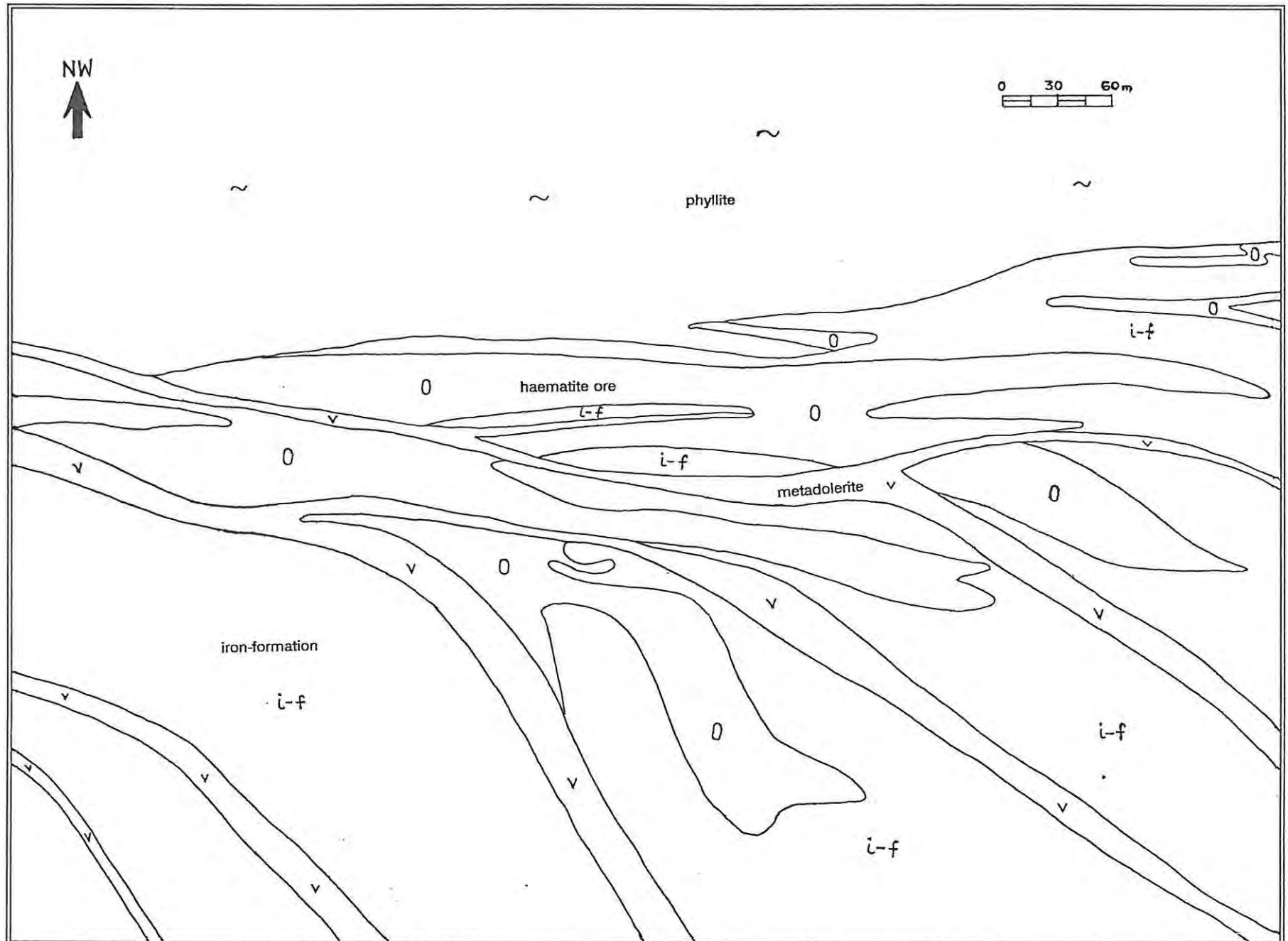


Figure 18: Surface plan of the Buchwa West Primary deposit, Zimbabwe (COURTESY OF BUCHWA IRON MINING COMPANY, 1992)

supergene followed by hypogene enrichment of the i-f with the foot-wall phyllite acting as a barrier to the iron rich fluids. The hypogene fluids are considered to have been generated by the late dolerite dykes. The same enrichment process described at Thabazimbi is envisaged to have produced the Buchwa deposits. All the necessary facets - faulting, shale barrier and i-f host, are present although the rocks are much older ($\approx 2.7\text{Ga}$) as compared to 2.3Ga for the Thabazimbi deposit host.

f. The Carajas Deposits, Brazil

Brazil hosts both the '*natural*' ores and the '*taconites*'. It is natural that the taconites will only be mined after the exhaustion of the '*natural*' ores. The Carajas deposit hosts 16 billion tonnes of haematitic-type ore at 66.5%Fe, 2.2%(Al₂O₃+SiO₂) and 1.1%P₂O₅. Pure haematite contains 70%Fe. The ore varies from *hard, semi-hard, mill, and soft* depending on how much fines are produced after crushing. These are respectively $\leq 25\%$, 25-50%, 50-75%, and $\geq 75\%$ fines. Just as *taconite* is a Canadian/USA term for i-f; so is *canga*, in Brazil, defined as a ferruginous breccia or conglomerate composed of fragments of haematite and itabirite cemented by limonite or haematite, and sometimes by other lateritic constituents. Canga, or iron-rich, silica-poor laterites accumulate over taconites, itabirites, or in boulder piles at the bases of slopes. The complete removal of silica is an enrichment process which results in recrystallisation and compaction resulting in a *hard-blue-haematite* mass - a very high grade ore. Pods and lenses of canga cap i-f sequences in many districts and these are a result of leaching of silica from weathering - a supergene process.

5. METAMORPHOGENIC DEPOSITS

Metamorphism usually accompanied by tectonism plays a very major role in the genesis of iron ore deposits. Haematitic-type ores are a result of high-grade, silica-leached and oxidised i-fs. The i-fs generally grade laterally, down-dip and stratigraphically downward into a mildly metamorphosed i-f - usually a beneficiable taconite. The low-temperature oxides, hydrous oxides (limonite), carbonates (siderite), and silicates (greenalite), originally deposited on the sea-floor may undergo:-

- i diagenesis which included dehydration and upgrading;
- ii regional metamorphism; and
- iii locally contact metamorphism

around major plutonic systems which are usually found proximal to these i-fs to produce the corresponding higher-rank layer lattice oxides magnetite, the silicates minnesotaite and stilpnomelane, the amphiboles grunerite and cummingtonite, the pyroxenes hedenburgite and ferrohypersthene (Bayley and James, 1973; Klein, 1973) which are amenable to beneficiation. Most Archaean Algoma-type i-f are beneficiable. They consist of well developed alternate bands of magnetite/haematite and chert. Metamorphism, usually related to tectonic activity enhances the fabric in the banding which causes 'clean' separation of the oxides and chert during comminution, usually along partings. The favourable deposits are discussed under i-fs (section C.9).

5.1 The Manisi Range, Zimbabwe

Owing to the hardness and differential weathering, the i-f of the Manisi Range builds the high ground in the Manisi-Mvuma greenstone belt, 200km south of Harare. The range, whose plan in outcrop resembles a compressed letter 'C', is ±75 km long. Its average thickness is ±2 km and rises to ±120 metres above the general elevation of the surrounding country. Karoo

basalt occupy the break in the letter 'C'.

The Manisi range belongs to the Upper Bulawayan Group. The rocks of the greenstone belt are folded into a closed syncline, of which the axis is almost horizontal. The general dip is between 50° and 80° E, i.e. towards the axis of the syncline. The i-f is overlain by clastics. Faulting of the i-f is limited.

The i-f is a reddish-brown coloured, hard, banded rock, consisting of alternate bands of haematite and silica. The individual bands vary from <1mm to 3cm in width, the more common range is 3-9mm. Fracturing of the rock and minute displacements of the bands are very common. The fissures and cracks are filled with silica. The i-f usually breaks along the bedding planes and fine striations are displayed in the haematite bands parallel to the banding.

Four high grade ore deposits are exposed as a series of small hills near the south-eastern margin of the i-f. The overlying clastics (grits) conceal the orebodies as well as the i-f so that only the high points are exposed. Developed over a total distance of 3 km, the orebodies grade laterally into i-f. The total tonnage of the high-grade deposits is so small as not to warrant opening up a mining. However, the i-f is of '*concentrating-grade*'. It is of consistent grade ($\geq 40\% \text{Fe}$ and $\pm 32\% \text{SiO}_2$, contains trace amounts S and P) and the reserves, at >30 000 Mt, are large. The ore types indicate haematite replacing magnetite on surface grading into a higher proportions of magnetite with depth.

F. EVALUATION OF IRON ORE DEPOSITS

1. INTRODUCTION

The geology of iron ore is so diverse that it entails use of a lot of geological and geophysical techniques in their exploration and evaluation. Also their geographical distribution is so diverse that every continent has important productive areas. As a result, exploration must be done under a wide variety of conditions that are dependent on the climate and infrastructure.

The quality of material demanded by the blast furnace operators has generally been changing since the sixties from the high-grade direct shipping ore (DRO) to the high grade iron concentrates and pellets made from low grade iron-formations. This has resulted in the diversion of exploration from DRO to ore amenable to concentration. The new emphasis on beneficiable ores has caused a need for unusually detailed geological study of deposits to provide the information which is of vital importance to proper evaluation and profitable exploitation. Important characteristics including mineralogy, texture, concentratability, grindability, distribution of ore types are investigated together with the more *traditional* factors such as transportation and fuel costs, taxes, and water and power rates.

The evaluation process is a team effort whose objective is to come up with a profit margin estimate. The geologist gathers the basic data, and is usually the only one who has the hands on information since he may indeed be the only one who will have seen the deposit. The engineers and metallurgists develop facts while working on data and samples gathered by the geologist. A general comprehension of all mine related disciplines is of utmost importance for the practising professional economic geologist. The geologist is customarily the member of the team with the best knowledge of the basic

environment and orebody morphology. As such, he is in a position to see that the talents of his colleagues are used to the best advantage and often he serves as the general coordinator of the project up to the engineering and construction stage.

2. EVALUATION PARAMETERS

The most important characteristics (appendix 3) of an iron ore deposit that enter into evaluation are split into those factors that are strictly geological and those that are non-geological. The geological factors must be very favourable for an iron ore deposit to be an economic orebody since without tonnage, grade, treatability etc., the availability of market, costs of power, taxes are only incidental. After passing the first list, the deposit should pass the second list. It should be pointed out that deposits are not perfect in every regard and therefore a compromise is usually the order of the day. The *in-situ* grade may be good but the grindability difficult or the tonnage may be enormous, but the shape and attitude of the orebody such that more expensive underground mining is required. Ultimately, it is the offsetting of potential revenues versus potential costs which determines the profitability and hence the viability of an iron ore deposit.

The evaluation process is an ongoing process and should be continuously reviewed. No economic estimate is likely to be valid for a long period, largely because of technological changes, changes in accessibility, rates, government policies, taxes etc., which cause a constant juggling in the relative economic standings. A good example of technological ingenuity is the development of autogenous mills which led to a large amount of saving on grinding cost. The construction of a railway line to a particular deposit may make another potential deposit more accessible and possibly, an economic

proposition. The minimum tonnage necessary for the development of a second mine is less than that for the first, in the same region.

Government policies determine the taxes, although the problems of foreign exchange and repatriation of capital are also important. Farsighted manoeuvres by some political entities have tended to attract risk capital. Iron ore deposits have a widespread geographical distribution. Therefore, mine developers have an unusually wide choice of location. The wide choice of location makes the economic estimate of an iron ore deposit, to be largely dependent on tax rates and other political factors. Consequently, the chances of developing a deposit will depend largely on tax rates and political factors.

Because profitability estimates are in constant review, it is imperative for exploration companies to maintain an inventory of worldwide iron ore resources which are constantly revised in the light of changes in the factors that affect the profit margins. Examples are the enormous reserves of titaniferous iron ores (both hard rock and beach sands) which are anathema to blast furnaces. A breakthrough in technology on smelting titaniferous ores will have a large impact on iron ore trade.

The discussion that follows is mainly directed to beneficiable iron-formation although some factors apply to high grade ores.

2.1 Ore-type and grade

Ore type and grade are inter-related. Ores are classified according to the method of beneficiation used. Examples are 'Wash' Ore, and 'Heavy Media' Ore. The grade of the ore is the iron content. The iron tenor which warrants an iron-formation to be called ORE depends on the economics of other factors in the evaluation. Some ores at 12-25%Fe may give

profit because they are easily grindable and concentrated. On the other hand, ores at 40%Fe can be so hard that the grinding cost will outweigh the high iron tenor resulting in the rejection of the deposit.

'Structure' which refers to the particle size or screen analysis, plays a major role in the beneficiation process. Structure is linked to the '*tumbling factor*' or the '*Q*' *index*. The '*Q*' index is the product of the percentage of the +6.4mm (or $\frac{1}{4}$ ") material before and after tumbling in a laboratory mills. It is related to the hardness, toughness, and grain size of the ore.

Assays of iron should quote the *total Fe content*, the *acid soluble iron*, and the *amount of magnetic iron in both the wet (natural) and dry conditions*. This is explained by the following example: total Fe is 30% but magnetic iron is only 25%. Hence a *straight* magnetic plant cannot be expected to recover >75% of the total Fe although it may be recovering (80% of the iron present as magnetite. On the other hand, a magnetic roasting plant will recover a much high percent of total Fe, say 90% (if iron silicates are not too abundant), because goethite, haematite and iron carbonates are all converted to artificial magnetite.

The amount of impurities may effect the viability of a deposit even though the Fe grade might be favourable. The total amount of minor elements (other than Mn, SiO₂, Al₂O₃, CaO and MgO) should not be >1%, otherwise they cause problems in the blast furnace. This figure varies from furnace to furnace and company to company. Limestone and coke also contribute to the trace element contamination. Assays of iron-formations from various parts of the world are shown in Appendix 4.

The grade should therefore encompass the weight percent recovery that can be made of a concentrate of a given assay.

The weight percent recovery then defines the amount of saleable product that will result and the value per tonne of that product. The raw crude assay only gives an indication of the amount of iron that theoretically is available, just as the iron unit recovery is a measure of plant effectiveness.

2.2 Tonnage

Tonnage determines the practicable size of a mining operation and therefore is of major importance. A mine life range of 20-25 years is used for amortisations, as a shorter life will increase the capital charge per ton sharply. On the other hand, a longer life (>25 years) does not affect the economics of depreciation but obviously influences plant expansion considerations. The world average plant life is 50 years. Working remote deposits necessarily requires the building of a town coupled with the transport costs for the product. Very large plant capacities, usually set by the market, tend to reduce the construction cost per unit of capacity, for remote deposits with an '*infinite*' tonnage.

2.3 Grain size and grindability

Evaluation of iron-formation for amenability to beneficiation centre around the grain size and grindability. The size of the mineral grains determine the amount of comminution necessary to liberate the ore minerals from the gangue so as to make an acceptable concentrate. The hardness and the toughness determine the amount of grinding required to achieve this liberation. Grinding is the largest single concentrating cost item. Therefore, geological factors or any other process that reduces the amount of grinding required are most welcome.

The level of grinding size at which there is a sharp improvement in the quality of the concentrate is related to

the average grain size. The aim is to produce a concentrate with $<7\%SiO_2$ at reasonably large screen size so as to have good marketability. Grindability index is an empirical measurement of the amount of power required to accomplish the liberation. On average a grind of 85-90% of -325 mesh is normal for Superior-type iron-formations. Sometimes 14 mesh is good enough. Texture is a function of metamorphism.

Pelletising requires that the coarse concentrate be reground to ± 325 mesh. Therefore there is no advantage in getting the 30-40% of raw ore which is recovered as concentrate because it still has to be ground to agglomerating size. Nonetheless coarse textured ore has a distinct advantage to that company with a large sinter capacity compared to pelletising capacity.

Autogenous mills, although cheaper tend to liberate the minerals at a coarser grain size without over-grinding. The reduced amount of grinding results in reduced consumption of mill liners as well as power. The coarser product presents less filtering problems with resulting benefits in balling and pelletizing.

The liberation also largely depends on the ore fabric or texture. Precambrian iron-formations are banded, consisting of alternate laminae and bands that are iron-rich and silica-rich (iron-poor). The layers should be such that the iron-rich ones should be solid iron oxide rather than concentrated disseminations (figure 19). The layered ore produces a high grade concentrate with much less grinding than does the finely laminated and disseminated ore. Thus these textural qualities affect the liberation that is achieved at various grind sizes.

Banded magnetite ores have the added advantage of being amenable to magnetic cobbing at various sizes. Coarse cobbing eliminate those grains that require hard grinding thus saving

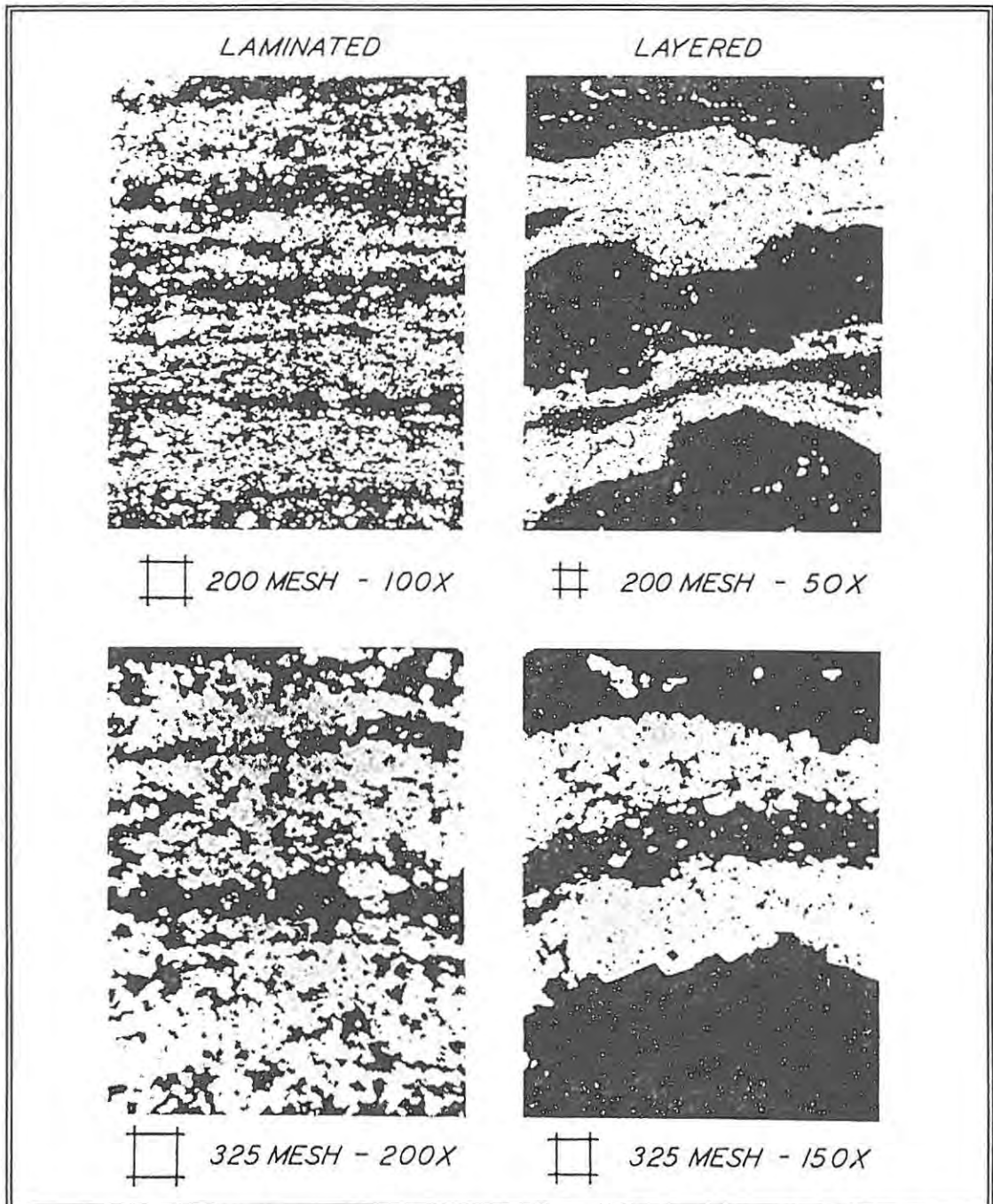


Figure 19: Photomicrographs in reflected light showing typical varieties of iron ore texture and structure (after Gunderson and Schwartz, 1962 *in* Ohle, 1972)

on the power and mill lining replacement. Coarse cobbing also eliminates the necessary selective mining required to isolate dykes. This also saves on the large mining equipment which would be precluded, and large magnetite blocks which would be lost due to excessive dilution would be recovered.

Iron silicates harden the ore thus making it very difficult to grind. Iron is also lost or eliminated with silica during beneficiation. Some silicates float with iron oxides in the flotation concentrator thus decreasing the product quality.

2.4 Mineralogy

Treatment characteristics and economics of an iron deposit are influenced by mineralogy. Ore minerals include haematite, magnetite, siderite, and goethite. Iron silicates are not classified as ore minerals.

Magnetite is concentrated by the cheap magnetic separation methods, while haematite, goethite and siderite are concentrated by the more expensive roasting and flotation methods. Haematite, when coarse, can be concentrated by low cost spirals.

Magnetite is cheaper to pelletise because of the exothermic oxidation of magnetite to haematite which releases a considerable amount of energy that saves on power requirements. Impurities are linked to the mineralogy. Phosphorous is bound to haematite and magnetite lattices and is always present in the concentrate. Titanium is a grating mixture in haematite and magnetite and is difficult to remove. However, the concentrating process removes a lot of impurities. Other impurities include copper, nickel, chromium, vanadium, cobalt and arsenic. The elimination of impurities in the concentration process depends on whether the impurities form separate entities, in which case they will be removed, or are incorporated in the mineral lattices of minerals of interest, in which case they will be troublesome.

2.5 Distribution of ore types

Iron-formations are inhomogeneous. Variations are both along and across bedding. The variations include crude grade, mineralogy, grindability, liberation size, etc. On the other hand, concentrating plants operate more efficiently on a uniform feed in all characteristics. This makes the knowledge of the distribution of ore types important, otherwise costs might be incurred in the ever changing plant setting to suite fluctuations in the properties of the plant feed. The more complicated the flowsheet, the more important it is to have a thorough knowledge of the geology and the more difficult is the geologist's job of grade control.

A significant contribution is called for on the mine geologist's part during the development and mining stages of an iron ore project which could result in an economical and efficient operation. During the early exploration period, detailed mapping and sampling can indicate the various ore types present, their distribution and therefore, selective mining or controlled blending can be recommended in order to produce a higher quality product. A higher quality product will ensure a higher profit and therefore, a higher internal rate of return and funds for more rapid payback of outlay.

2.6 Orebody depth and overburden

The depth of the orebody determines the amount of overburden that has to be stripped in order to mine the deposit. The costs of stripping vary according to the type of overburden. Overburden can be rock, sand, gravel, calcrete etc. Underground mining is planned when the stripping ratio is 6-8. The mining cost per tonne of open pit ore is considerably less than the costs for underground mining. Therefore feasibility studies should always be carried for both options even in the remotest possibility of an open pit being

feasible.

2.7 Orebody shape and attitude

The shape and the attitude of a deposit determine the mining method. A near equidimensional orebody will have a lower stripping ratio than a steeply dipping tabular orebody. Irregular orebodies are also very costly to mine due to highly selective mining required as compared to bulk mining methods applied to regular orebodies.

2.8 Location

Remote, rugged and inaccessible areas render potential iron ore deposits unattractive. Climate plays a role on the plant operating efficiency and therefore harsh climates (both hot and cold) are not favourable. Harsh climates and rugged and remote terrains only delay the working of a deposit. An inland location is less favourable than one near the coast because of high transport costs. A recent development is the pipeline transport of concentrate, a procedure which will eliminate the high rail haulage charges. However the development is only suitable in areas of rugged topography and where weather conditions permit.

2.9 The Market and commodity price

The wide distribution of both sources of ore and potential markets make the evaluation of market and price factors complicated. Development of iron ore deposits is done after prior arrangements on marketing contracts. In most cases, some of the partners are the consumers. Generally, iron is a low value mineral commodity. Consequently, in order to achieve profit, very efficient production is imperative and,

in the evaluation of new deposits, accurate estimates are of utmost importance.

The raw material requirements vary from company to company and this, therefore, determines the acceptability of an iron ore deposit type. For example, an iron ore deposit with more fines (-10mm) would be attractive to a steel company with a large sinter plant capacity than one without. On the other hand, a steel plant with other ore sources will accept an ore deposit that has impurities.

3. PROFITABILITY: A DISCUSSION ON THE ROLE OF THE GEOLOGIST IN EVALUATION

Ore deposits are confronted with various physical, geographical and political situations. In evaluation not all the cost estimates for individual items have the same degree of validity. The figures simply represent the best data available at a given time. Problems in data acquisition are that some factors are known precisely but others have to be estimated from fragmentary data or by comparison with known costs at other deposits that are similar. A fair knowledge of most of the purely geological factors, usually, is on hand by the time a project has had some widespread reconnaissance drilling and bulk sample testing.

As his contribution to the evaluation team effort, the geologist should provide the best possible interpretation of good data obtained from an orebody, especially its composition, and general environment. Participation in the revisions and modifications of mine layout and flowsheet, until the best combination is found is expected of the geologist, as engineering and metallurgical data are developed.

"This interplay of disciplines in a smoothly co-ordinated

organisation can be a most satisfying activity for a geologist who sees his field observations and scientific understanding of the orebody translated into profits. A new dimension is added to his thinking. For the experienced exploration geologist, it should be second nature to gather the answers to many engineering questions as he goes about his job of assembling the pertinent scientific data" (Ohle, 1972).

G. EXPLORATION

1. INTRODUCTION

Greenstone belts of $\geq 3.0\text{Ga}$ age contain relatively small metamorphosed i-fs which are suitable for beneficiation, for example, Bomvu Ridge iron ore deposit on the Barberton Greenstone belt (Bursill, et al., 1964), Manisi Range (Worst, 1962) and the Ghoko-Longwe Range. The i-fs can be described as *itabirites*, that is, well developed alternate bands of magnetite and chert/quartzite. Magnetite concentrate can be realised by crushing the rock to the right size and separating the magnetite magnetically. Greenstone belts of 2.9-2.5Ga in age, generally have suffered the vagaries of deformation and therefore are mildly metamorphosed and usually contain iron ore bodies hosted in i-f. The size of the orebody depends largely on the conditions and settings conducive to some form of iron enrichment being met.

Oxide facies i-fs are highly distinctive in the field both in outcrop and geophysically. The i-fs rich in silica tend to be very resistant to erosion, producing ridges or mountain ranges. Those rich in magnetite can be mapped out and detected using a magnetometer particularly when they are masked by cover like, vegetation, Karoo and/or Kalahari sands, swamps, tundra, muskeg or glacial debris.

Tectonic deformation is normally accompanied by greenschist facies metamorphism and produces penetrative and regular chevron folding and systematic crenulation. The deformed i-fs generally host high grade deposits in the fold hinges. The bedding which is usually well preserved is commonly wavy to crenulated in outcrop. A study and understanding of the structure of a deformed i-f is important.

2. CONTINENTAL SETTING OF THE BASINS

In southern Africa, the strata of the Archaean basins and Late Archaean-Early Proterozoic Basin are, respectively, preserved within the limits and at the craton margins of the Archaean tectonic units, namely the Zimbabwean and the Kaapvaal cratons. The Archaean basins are characteristically greenstone belts and are well developed on the Zimbabwean craton. The Late Archaean-Early Proterozoic basin is typified by the Transvaal basin and the Magondi Belt, Zimbabwe (Master, 1991).

The greenstone belts are preserved in larger volumes of granitic rock, the ages of which are generally in excess of 3.0Ga. The Archaean blocks of both cratons are surrounded by younger metamorphic mobile belts which usually contain economic amounts of beneficiable metamorphosed i-f, for example, the Moonlight Magnetite Quartzite in the Limpopo Mobile Belt, Northern Transvaal (Badenhorst, 1990). A suite of younger granitic intrusives is generally observed in the basins responsible for i-f metamorphism.

3. AGE OF BASINS

The strata range in age from Early Archaean (3.8Ga) to Early Proterozoic (1.8Ga)

4. SIZE OF THE BASINS

The basins vary tremendously in size depending on tectonic setting that prevailed at the time of formation. The Griqualand West and the Transvaal sub-basins which constitute the Transvaal basin, have a combined area of 250 000 km². The Transvaal basin is filled by sediments and volcanics up to 12 000 metres thick (Button, 1976c). The basins can also be quite small, for example the greenstone belt in which the Buchwa deposits occur in Zimbabwe is <1 000km² in area.

Geometrically, the cratonic basins are gentle warps on a sub-continental scale. They represent gently subsiding continental regions with plan-view dimensions of about 100 times their maximum downwarp dimension.

5. STRUCTURAL STYLES OF BASINS

The Early-Proterozoic successions are homoclinal and very gently deformed over large areas, generally along the margins of cratons where metamorphic/mobile belts are developed. The amount of deformation increases towards the margins of the cratons. The fold axes of the craton-marginal tectonic belts tend to parallel the margins of the craton. The craton interiors are characterised by locally developed structural disturbances which are usually related to the granite greenstone basement. It has been observed by Button (1973) in the Transvaal Basin that the axes of major synclinal features are frequently situated over Archaean greenstone belts.

6. GROSS STRATIGRAPHIC SUB-DIVISION

Three sub-division are apparent. Proterozoic units commence with a basal volcanic and clastic unit which grades into a chemical sedimentary unit, the latter being unconformably overlain by an upper clastic unit. The whole package rests on Archaean granite-greenstone terrains. Greenstone belt stratigraphy is similar to that described above. The basal volcanic unit may be poorly developed or absent, as in the Buchwa Belt (Zimbabwe) and is overstepped by the clastic sedimentary unit. The chemical sedimentary unit may also be absent being overstepped by the upper clastic unit, as at Sishen in the Griqualand sub-basin (Northern Cape).

Of most significance for iron ore exploration is the recognition of the chemical sedimentary unit which is almost always, characterised by total absence of allochthonous

coarse clastic rocks and the predominance of chemical sediments which include carbonates, chiefly dolomite, iron-formation and chert.

I-f generally contain minerals such as haematite, magnetite (oxides), siderite (carbonate), greenalite, stilpnomelane, minnesotaite, riebeckite (silicates) and quartz. Stilpnomelane may have formed from a montmorillonite precursor, which in turn, may have been derived from a volcanic glass (Grubb, 1971). Magnetite forms during diagenesis from haematite and by decomposition of siderite (Ayres, 1972). Minnesotaite rosettes grow from greenalite during low-grade burial metamorphism (Grubb, 1971; Ayres, 1972). Riebeckite forms by soda metasomatism of pre-existing stilpnomelane or quartz-iron mesobands, and is frequently found as radiating needles around magnetite grains. Grubb (1971) found experimentally that crocidolite growth is favoured by an abundant supply of Na^+ and Fe^{2+} cations combined with low pH and Eh conditions. From Ayres' (1972) conclusion that temperatures around 300°C and pressures ranging 4-6kb are required for crocidolite formation, it can be inferred that the occurrence of crocidolite in i-f indicates metamorphism of the i-f at the foregoing conditions. I-fs that contain crocidolite do not host haematitic-type high grade ore deposits.

7. EXPLORATION TECHNIQUES IN SOUTHERN AFRICA

Magnetic methods of geophysical prospecting are used in thickly vegetated or sand covered regions. These magnetic techniques are achieved by use of a dip needle and a magnetometer in airborne surveys and ground follow-up and have helped in the mapping of the bed-rock structure beneath the Kalahari and Karoo cover. However most of the high-grade ore are related to structures, notably folds and the result has been the discovery of considerable amounts of ore.

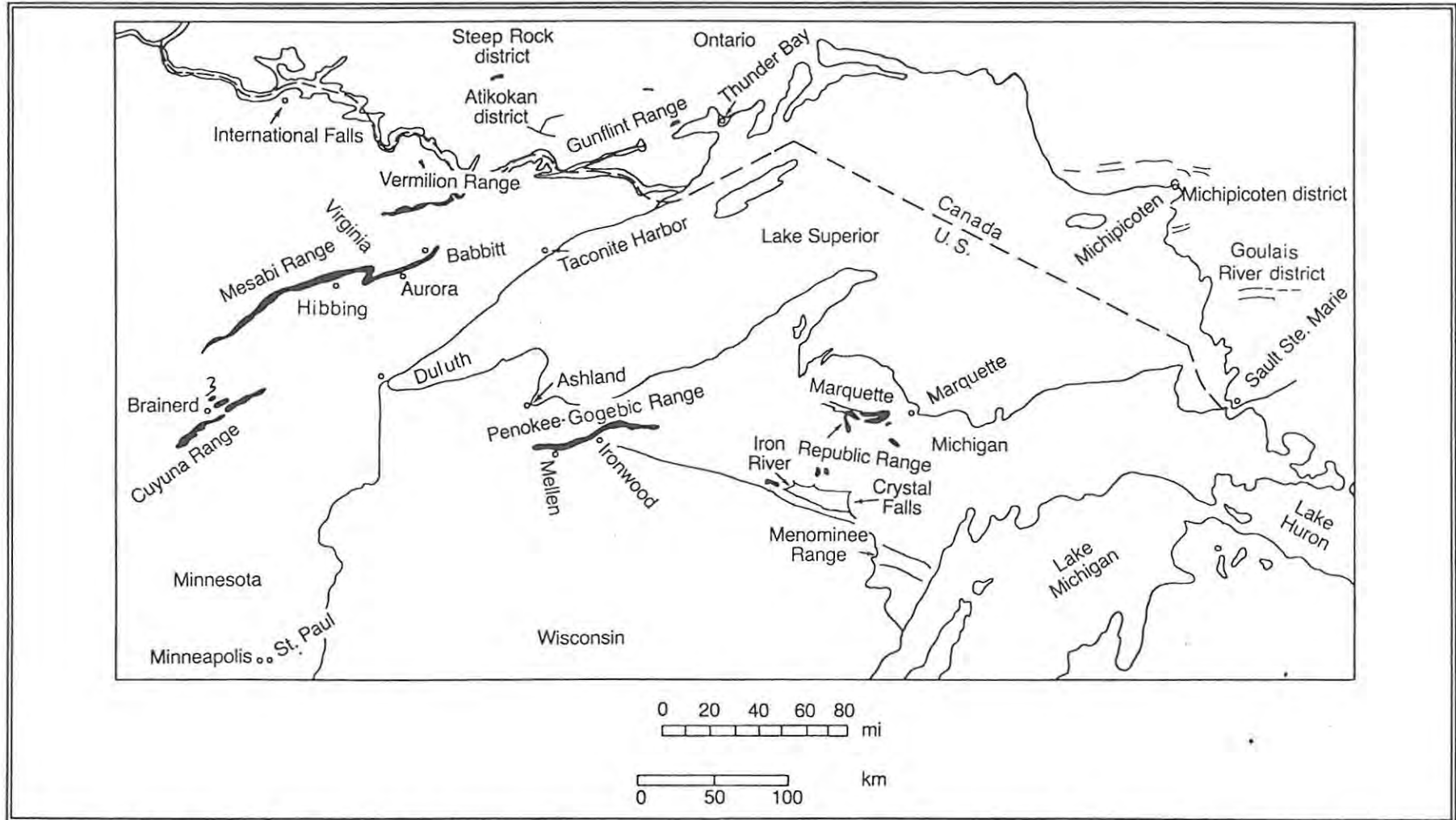


Figure 20: Map of the Lake Superior region showing the location of major iron ore districts (from Lake Superior Iron Ore Association, 1952 *in* Guilbert & Park, 1986).

Generally high grade '*natural*' ores occur as small discontinuous pods along and in the i-f. Geophysical exploration has to be oriented towards delineating the i-fs. Geophysical exploration was carried out at Moonlight Range Prospect (Badenhorst, 1990) to pick up the i-f (magnetite quartzite) under Karoo cover. The haematitic-type high grade ores define magnetic '*lows*'. Valuable information regarding the structure and depth extension of the magnetite quartzite orebodies under calcrete or soil cover can be obtained from geomagnetic surveys (Badenhorst, 1991) especially where the success of aerial and detailed geological mapping is limited due to scarcity of outcrop and occurrence of thick sand cover in some areas. Interpretation of geology in trenches although cheap can be disadvantageous or problematic where big boulders and scree occur around outcrops and sub-outcrops.

Small percussion drills may be used to delineate relatively shallow ore and to determine continuity of sub-outcrops. The i-fs (taconites), i.e., magnetite-rich, define magnetic *highs*. The best start for an exploration program is at *old direct-shipping-ore* pits since the ore is usually hosted in i-f and therefore the pits always bottom out in *taconite* or show a transition from '*natural ore*' to '*oxidised ore*' through '*oxidised taconite*' to '*taconite*'.

The Manisi, Buchwa and Ghoko-Longwe Ranges have simple structures. Sishen and Thabazimbi have similar successions. The i-fs are affected by faulting and folding and thus a detailed knowledge of the stratigraphy and structure is essential to exploration.

The best '*natural*' ores are generally concentrated along mafic dyke intrusives, for example Buchwa and Thabazimbi Ranges. The dykes are guides to the mineralising fluids, whether connate or meteoric.

In summary, exploration for iron ore deposits in southern

Africa entails identifying the type of ore required, i.e., whether high grade haematitic ore or concentrating grade iron-formation. Exploration for the high grade deposits entails identifying a deformed Archaean to Early Proterozoic oxide facies iron-formation. Ore deposits occur at the base of the iron-formation proximal to the fold hinges. Exploration for concentrating grade iron-formation involves identifying $\geq 2.7\text{Ga}$ old undeformed metamorphosed oxide/carbonate facies iron-formation, usually, in old (2.7-3.8Ga) metamorphic/mobile belts. The suitable iron-formation occurs on the fold limbs.

Airborne and ground geophysical methods are used to identify the iron-formations which are concealed under late cover sequences. Detailed geological field mapping and/or percussion drilling is required to pick up the high grade (DRO) ores.

H. CONCLUSIONS

1. The sedimentary basins of the world of Early Proterozoic (2.5-1.9Ga) age, to a large extent, and those of Mid-Archaeon (3.5-2.5Ga) age, to a less extent, are considered to be the major repositories of iron in addition to a wide range of other valuable minerals and metals (gold, uranium, manganese, copper, lead, zinc, crocidolite and amosite). Prominent among these basins are those situated on the ancient shield areas of southern Africa, Western Australia, Brazil, Canada and the United States of America.

Basins that host i-fs are divided into a basal volcanic and clastic unit, a chemical sedimentary unit, and an upper clastic unit. Stratigraphic relations, lithologies, and depositional environments within each of these subdivisions are similar. Larger iron ore deposits are hosted in the Superior-type i-f. The Algoma-type i-f hosts smaller deposits of iron formed by supergene/hypogene processes.

Generally, the basins that contain i-fs which host large deposits of iron ore are correlatable with one another as concerns basin development and tectonic history in the context of the evolution of Gondwanaland. The above concept must form the basis of any genetic model that is required to help direct exploration for iron ore deposits from continent to continent.

2. A continuum of deposition of i-fs existed from early Archaean time until approximately 1800Ma ago. The major early Proterozoic development of i-fs is not a synchronous event (Gole and Klein, 1981). Sawkins (1984a) attributes the widespread early Proterozoic i-fs

to the presence of increasing levels of oxygen in shallow marine environments. The increase in oxygen availability caused the ferrous iron budget of the world's oceans to be essentially used up by about 1800Ma ago (Holland, 1973) with the waning of submarine volcanism and the associated iron-silica-supplying fumarolic activity. Button (1976a) considers local evaporation to have aided the deposition of i-f, a very minor contribution in the authors opinion.

3. The high grade (DRO) ores are formed by supergene/hypogene enrichment of the oxide facies iron-formation. Metamorphism accompanied by minor deformation of oxide/carbonate facies iron-formation in >2.7Ga greenstone belts or metamorphic/mobile belts produces '*concentrating grade*' iron-formation.
4. Major factors that determine the viability of iron-formation as ORE are:
 - i its distance to the nearest rail head and port;
 - ii distance to the beneficiation plant;
 - iii geological setting - i.e whether is can be mined by *cheap* opencast methods and whether it can be cheaply and easily handled; and
 - iv its amenability to concentration, which include grain size, grindability, mineralogy and impurities.

Weathered taconite contains unusually amounts of silica and is beneficiated to an economic grade by removing the silica through washing and gravity methods. Jaspilites or hard jasper-haematites are beneficiated by flocculation and/or flotation, for example, the Marquette Range ore in the USA (figure 20).

Taconite ore evaluation is complex and is based on geological and mineralogical data (Ohle, 1972). The grade is not very important in so far as the total iron is concerned. The mineralogy plays an important role. Higher concentrations of magnetite are required at a grain size liberation of 200-300mesh (75-50 μ m) and is critical. The beneficiated product should have 3-6%SiO₂ content. A silica content of <3% will prevent slag formation in the blast furnace and therefore, is undesirable. Low levels of Na and K, which are typical of *taconites*, are appreciated. High levels of K and Na are *anathema* to blast furnace linings. The acceptable levels of titanium are $\leq 0.7\%$ otherwise a *pasty* slag and therefore, poor separation would result. Higher levels of phosphorus affect the steel-making properties of the ore. The textures, grade of metamorphism, and the mineralogy of gangue material also play a major role in the viability of a deposit, in as far as amenability to beneficiation is concerned. Non-geological factors include transportation costs, infrastructure costs, and the overall energy cost per tonne of finished product.

5. Deposition of i-fs occurred in deep marine basins. The iron and silica was supplied by fumarolic activity associated with submarine volcanism. These deep marine basins may have been the proto oceans to the present ones. Therefore plate tectonics probably started in the Archaean.

BIBLIOGRAPHY

- ANHAEUSSER, C.R. and BUTTON, A., 1976, A review of southern African stratiform ore deposits - their position in time and space: *In: Wolf, K.H. (ed.), Handbook of Stratiform and Stratiform Ore Deposits*, Elsevier, Amsterdam, p257-319.
- AYRES, D.E., 1972, Genesis of iron-bearing minerals in banded iron formation mesobands in the Dales Gorge Member, Hammersley Group, Western Australia: *Econ. Geol.*, **67**, p1224-1233.
- BADENHORST, J.C., 1990, The Precambrian iron-formations in the Limpopo Belt as represented by the Magnetite Quartzite Deposits at Moonlight, Koedoesrand area, Northern Transvaal: *MSc Diss.*, Rhodes University, Grahamstown, 99pp.
- BARGHOORN, E.S., and TYLER, S.A., 1965, Microorganism from the Gunflint Chert: *Science*, **147**, p563-577.
- BAYLEY, R.W. and JAMES, H.L., 1973, Precambrian iron-formations of the United States: *Econ. Geol.*, **68**, p934-959.
- BERKMAN, D.A., 1989, Field Geologist's Manual: *The Australian Institute of Mining and Metallurgy Monograph No. 9*, 3rd edition, Parkville, Victoria, 382pp.
- BEUKES, N.J., 1973, Precambrian Iron-Formations of Southern Africa: *Econ. Geol.*, **68**, p960-1004.
- , 1986, The Transvaal Sequence in Griqualand West: *In: Anhaeusser, C.R and Maske, S. (eds.) Mineral Deposits of Southern Africa, Vol I, Geol. Soc. S. Afr.*, p819-828.

- BLISS, N.W., 1962, The geology of the country around Umvuma and Felixburg: *Rhodesia Geol. Survey Bull. No. 56*, Salisbury, 103pp.
- BONATTI, E., 1984, Hydrothermal metal deposits from the oceanic rifts: a classification: *In: Hydrothermal Processes at Seafloor Spreading Centres*, Plenum Publishing Corp., p491-502.
- BOOKSTROM, A.A., 1977, The Magnetite Deposits of El Romerol, Chile: *Econ. Geol.*, 72, p1101-1130.
- BOSTROM, K., and PETERSON, M.N.A., 1966, Precipitates from hydrothermal exhalations of the East Pacific Rise: *Econ. Geol.*, 61 p1258-1265.
- BOURHILL, P.E., 1978, Holes A2 and B2 diamond drilling and revised ore reserves between sections A and C - final report: *Buchwa Iron Mining Company internal report* (unpubl), 13pp.
- BOYUM, B.H., and Hartviksen, R.C., 1970, General geology and ore-grade control at the Sherman mine, Temagami, Ontario: *Can. Inst. Min. Metall. Trans.*, 73, p267-276.
- BURKE, K., DEWEY, J.F., and KIDD, W.S.F., 1976, Dominance of Horizontal Movements, Arc and Micro-continental Collisions during the later Permian Regime: *In: Windley, B.F. (ed.) The Early History of the Earth*, Wiley, London, p113-129.
- BURSILL, C., LUYT, J.F.M. and URIE, J.G., 1964, The Bomvu Ridge Iron Ore Deposit, p405-414: *In: Haughton, S.H., (ed.), The Geology of Some Ore Deposits in Southern Africa, Vol. II*, Geol. Soc. S. Afr., Johannesburg, 739pp.

- BUTTON, A., 1972, The Transvaal Basin Project: *Econ. Geol. Res. Unit, Univ. Witwatersrand, 13th Ann. Rep.*, p21-27.
- BUTTON, A., 1973, The depositional history of the Wolkberg proto-basin, Transvaal: *Trans., Geol. Soc. S. Afr.*, 76, p15-25.
- 1976a, Iron-formations as an End-member in Carbonate Sedimentary Cycles in the Transvaal Supergroup, South Africa: *Econ. Geol.*, 71, p193-201.
- , 1976b, Stratigraphy and relations of the Bushveld floor in the eastern Transvaal: *Trans., Geol. Soc. S. Afr.*, 79, p3-12.
- , 1976c, Transvaal and Hammersley basins - review of basin development and mineral deposits: *Miner. Sci. Eng.*, 8, p262-293.
- CHESHIRE, P.E., LEACH, A. and MILNER, S.A., 1980, The Geology of the Country between Gwelo and Redcliff: *Zim. Geol. Surv. Bulletin No. 86*, Salisbury, 300pp.
- CLOUD, P., 1973, Palaeo-ecological Significance of the Banded Iron Formation: *Econ. Geol.*, 68, p1135-1143.
- CONDIE, K.C., 1981, Archaean Greenstone Belts, *Developments in Precambrian Geology 3*, Elsevier, Amsterdam, 434pp.
- , 1982, Early and middle Proterozoic supracrustal successions and their tectonic settings: *Am. J. Sci.*, 282, p341-357.
- COXON, B.D., 1992, Pers. Comm.
- DORR, J.V.N., 1973, Iron-formations of South America: *Econ.*

Geol., 68, p1005-1022.

DUNN, J.A., 1935, The origin of iron ores in Singhbhum, India: *Econ. Geol.*, 30, p643-654.

-----, 1941, The origin of banded haematite ores in India: *Econ. Geol.*, 36, p355-370.

ELLISON, D., 1992, Outlook for iron ore and base metals: *The Chamber of Mines Journal*, Thompson Publications, Harare, vol.34, No. 7 (July 1992 issue), p15-23.

EUGSTER, H.P., and I-MING, C., 1973, The Depositional Environments of Precambrian Banded Iron-formations: *Econ. Geol.*, 68, p1144-1168.

FRIPP, R.E.P., 1976, Stratabound Gold Deposits in Archaean Banded Iron Formations, Rhodesia: *Econ. Geol.*, 71, p58-75.

FRYER, B.J., 1983, Rare earth elements in iron-formation, p345-358: *In: Trendall, A.F. and Morris, R.C., (eds.): Iron-formation: Facts and Problems, Developments in Precambrian Geology 6*, Elsevier, Amsterdam, 558pp.

GAAL, G., 1992, Global Proterozoic tectonic cycles and Early Proterozoic metallogeny: *S. Afr. J. Geol.*, 95 (3/4), p80-87.

GAPARA, C.S., 1988, The Geology of the Ripple Creek Iron Ore Deposit and the Surrounding Country: *B.Sc. Hons. Thesis (unpubl)*, University of Zimbabwe, Harare.

_____, 1989, The Manisi Iron Ore Deposits: *Buchwa Iron Mining Company Internal Report (unpubl)*, 38pp.

GARRELS, R.M., PERRY (Jr.), E.A., and MACKENZIE, F.T., 1973,

Genesis of Precambrian iron-formations and the Development of Atmospheric Oxygen: *Econ. Geol.*, **68**, 1173-1179.

GEIJER, P., 1910, Igneous Rocks and Iron Ores at Kiirunavaara, Lousavaara, and Toulavaara, Sweden: *Econ. Geol.*, **5**, 699-718.

GOAL, M.J., and KLEIN, C., 1981, Banded-iron-formations through much of Precambrian time: *J. Geol.*, **89**, p169-184.

GOLDICH, S.S., 1973, Ages of Precambrian banded iron-formations: *Econ. Geol.*, **68**, p1126-1134.

GOODWIN, A.M., 1973, Archaean iron formations and tectonic basins of the Canadian Shield: *Econ. Geol.*, **68**, p915-934.

GROHMANN, G., 1990, Iron Ore: *South Africa's Mineral Industry - 1989*, Dept. of Mineral & Energy Affairs, Minerals Bureau, Braamfontein, p97-101.

GROSS, G.A., 1965, Origin of Precambrian iron formations, discussion: *Econ. Geol.*, **60**, p1063-1065.

-----, 1980, A classification of iron formations based on depositional environments: *Can. Mineral.*, **18**, p215-222.

GRUBB, P.L.C., 1971, Silicates and their paragenesis in the Brockman Iron Formation of Wittenoom Gorge, Western Australia: *Econ. Geol.*, **66**, p281-292.

GUILBERT, J.M. and PARK, Jr., C.F., 1986, *The Geology of Ore Deposits*, W.H. Freeman and Company, New York, 985pp.

- HAMILTON, P.J., 1977, Isotope trace element studies of the Great Dyke and Bushveld mafic phase and their relation to early Proterozoic magma genesis in southern Africa: *Econ. Geol.*, 12, p24-52.
- HARRISON, N.M., 1970, The geology of the country around Que Que: *Rhodesia Geol. Survey Bull. No. 67*, Salisbury, 125pp.
- HOLLAND, H.D., 1973, The Oceans: a Possible Source of Iron in Iron-formations: *Econ. Geol.*, 68, p1169-1172.
- JAMES, H.L., 1954, Sedimentary Facies of Iron Formation: *Econ. Geol.*, 49, p235-293.
- , 1983, Distribution of banded iron-formation in space and time, p471-490: *In: Trendall, A.F. and Morris, R.C., (eds.): Iron-formation: Facts and Problems, Developments in Precambrian Geology 6*, Elsevier, Amsterdam, 558pp.
- KIMBERLEY, M.M., 1978, Palaeo-environmental Classification of Iron-formation: *Econ. Geol.*, 73, p215-229.
- KLEIN, Jr., C., 1973, Change in mineral assemblages with metamorphism of some banded Precambrian iron-formations: *Econ. Geol.*, 68, p1075-1088.
- LEPP, H., and GOLDICH, S.S., 1959, The Chemistry and Origin of iron formations: *Econ. Geol.*, 54, p1348-1349.
- MACKIN, J.H., 1954, Geology and Iron Ore Deposits of the Granite Mountain Area, Iron County, Utah: *USGS Map MF 14*.
- MACGREGOR, A.M., 1928, The geology of the country around the Lonely Mine, Bubi district: *Southern Rhodesia Geol.*

Survey Bull. No. 11, Salisbury, 98pp.

- , 1968, Iron Ore Deposits of the Iron Springs District, South-west Utah, p992-1019: *In: Ridge, J.D. (ed.), Ore Deposits of the United States 1933/1967, Graton-Sales Vols., New York, AIME, 1880pp.*
- MALLINSON, C.A., 1985, Proterozoic Basins in Southern Africa and some Related Mineral Deposits; Field Trip to the Transvaal/Griqualand West and Witwatersrand Basins-with a Chapter on Kimberlites (unpubl): *Rhodes University, Grahamstown, 154pp.*
- MARGULIS, L., 1970, *Origin of Eukaryotic Cells: Evidence and Research Implications for a Theory of the Origin and Evolution of Microbial, Plant, and Animal Cells on the Precambrian Earth*, Yale University Press Ltd.(publ), London, 349pp.
- , 1981, *Symbiosis in Cell Evolution; Life and Its Environment on the Early*, Freeman (publ), San Francisco, 419pp.
- , 1984, *Early Life*, Jones and Bartlett Publishers, Inc., Boston, 160pp.
- MARSDEN, R.W., 1968, Geology of the iron ores of the Lake Superior region in the United States, p489-506: *In: Ridge, J.D., (ed.), Ore Deposits of the United States 1933/1967, Graton-Sales Vols., New York, AIME, 1880pp.*
- MARSH, J., 1985, Geochemical characterisation of lavas of the Ongeluk Formation, Northern Cape, (unpubl. report), 11pp.
- MASTER, S., 1991, Stratigraphy, Tectonic Setting and Mineralisation in the Early Proterozoic Magondi

Supergroup, Zimbabwe: A Review: *Econ. Geol. Research Unit*, Univ. Witwatersrand, Inf. Circular 238.

MAZHINDU, J, 1985, Middle Primary ore deposit diamond drilling - Interim report: *Buchwa Iron Mining Company Internal Report* (unpubl), 21pp.

METAL BULLETIN, 3 August 1992, No.7703, p10.

METAL BULLETIN, 6 August 1992, No.7704, p11.

METAL BULLETIN, 10 August 1992, No.7705, p11.

METAL BULLETIN, 13 August 1992, No.7706, p19.

METALS AND MINERALS, Annual Review, 1992.

MILLER, L.J., 1976, Corporations, ore discovery and the geologist: *Econ. Geol.*, 71, p836-847.

MINING ANNUAL REVIEW, 1992.

MORRIS, R.C., 1980, A textural and mineralogical study of the relationship of iron ore to banded iron-formation in the Hammersley Iron Province of Western Australia: *Econ. Geol.*, 75, p184-209.

-----, 1983, Supergene alteration of banded iron-formations, p513-534: *In: Trendall, A.F. and Morris, R.C., (eds.): Iron-formation: Facts and Problems, Developments in Precambrian Geology 6*, Elsevier, Amsterdam, 558pp.

OHLE, E.L., 1972, Evaluation of iron ore deposits: *Econ. Geol.*, 67, p953-964.

- O'ROURKE, J.E., 1961, Palaeozoic banded iron formations: *Econ. Geol.*, 56, p331-361.
- PERRY, (Jr.), E.C., 1983, Oxygen isotope geochemistry of iron-formations, p359-371: *In: Trendall, A.F. and Morris, R.C., (eds.): Iron-formation: Facts and Problems, Developments in Precambrian Geology 6*, Elsevier, Amsterdam, 558pp.
- PETTIJOHN, F.J., 1975, *Sedimentary Rocks, Third Edition*: Harper & Row Publishers, New York, 628pp.
- PHAUP, A.E., 1933, The geology of the Lower Gwanda gold belt: *Rhodesia Geol. Survey Bull. No. 24*, Salisbury, 74pp.
- PRETORIUS, D.A., 1981, Gold and Uranium in quartz-pebble conglomerates: *Econ. Geol.*, 75th Anniv. Vol., p117-138.
- SAGE, A.M., 1992, Vanadium: *Metals & Minerals Annual Review - 1992*, Mining Journal, London, p73.
- SAWKINS, F.J., 1984a, *Metal Deposits in Relation to Plate Tectonics*, Springer-Verlag, Berlin, 325pp.
- , 1984b, Ore genesis by episodic dewatering of sedimentary basins: Applications to giant Proterozoic lead-zinc deposits: *Geology*, 12, p451-454.
- SCHÜTTE, S.S. and CORNELL, D.H., (1992), Extreme sea-floor alteration as the source of the Kalahari Manganese Deposits: *24th Geocongr. Abstracts*, Geol. Soc. S. Afr., p354-356.
- SCHWEIGART, H., 1965, Genesis of iron ores of the Pretoria Series, South Africa: *Econ. Geol.*, 60, p269-299.

- SCLATER, J.G., JAUPART, C., and GALSON, D., 1980, The Heat Flow through Oceanic and Continental Crust and the Heat Loss of the Earth: *Rev. Geophys. Space Phys.* 18, p269-311.
- STAGMAN, J.G., 1978, An outline on the Geology of Rhodesia: *Rhodesia Geol. Surv. Bulletin No. 80*, Salisbury, 126pp.
- STANISTREET, I.G. and McCARTHY, T.S., 1991, Changing Tectono-Sedimentary Scenarios Relevant to the Development of the Late Archaean Witwatersrand Basin: *Econ. Geol. Research Unit, Univ. Witwatersrand, Inf. Circular 233*.
- STOWE, C.W., 1968, The geology of the country south and west of Selukwe: *Rhodesia Geol. Survey Bull. No. 59*, Salisbury, 209pp.
- STRATTON, P., 1992, Steel: *Metals & Minerals Annual Review - 1992*, Mining Journal, London, p73.
- STRAUSS, C.A., 1964a, The iron ore deposits in the Sishen area, Cape Province, p393-403: *In: Haughton, S.H., (ed.), The Geology of Some Ore Deposits in Southern Africa, Vol. II*, Geol. Soc. S. Afr., Johannesburg, 739pp.
- , 1964b, The iron ore deposits in the Thabazimbi area, Transvaal, p383-392: *In: Haughton, S.H., (ed.), The Geology of Some Ore Deposits in Southern Africa, Vol. II*, Geol. Soc. S. Afr., Johannesburg, 739pp.
- SWEENEY, R.E., and KAPLAN, I.R., 1973, Pyrite Framboid Formation: Laboratory Synthesis and Marine Sediments: *Econ. Geol.*, 68, p618-634.
- THE INTERGOVERNMENTAL GROUP OF EXPERTS ON IRON ORE (IGE), 1992, Iron Ore: *Metals & Minerals Annual Review - 1992*,

Mining Journal, London, p57-59.

THE JSE HANDBOOK, August 1992.

THE SUNDAY TIMES, October 11, 1992.

TRENDALL, A.F., 1965, Origin of Precambrian iron formations, discussion: *Econ. Geol.*, **60**, p1065-1069.

-----, 1968, Three great basins of Precambrian banded iron-formation deposition: a systematic comparison: *Geol. Soc. Amer. Bull.*, **79**, p1527-1544.

-----, 1973, Precambrian iron-formations of Australia: *Econ. Geol.*, **68**, p1023-1034.

TRUSWELL, J.F., 1990, The Transvaal and Griqualand West Sequences: Some Current Issues: *Econ. Geol. Research Unit, Univ. Witwatersrand, Inf. Circular 232*.

TYLER, N., 1979, The stratigraphy of the Early-Proterozoic Buffalo Springs Group in the Thabazimbi area, west-central Transvaal: *Trans., Geol. Soc. S. Afr.*, **82**, p215-226.

TYLER, S.A., 1949, Development of Lake Superior Soft Ores from Metamorphosed Iron Formation: *Geol. Soc. Amer. Bull.*, **60**, p1101-1124.

VAN DEVENTER, J.L., ERIKSSON, P.G. AND SYNMAN, C.P., 1986, The Thabazimbi iron ore deposits, north-western Transvaal: *In: Anhaeusser, C.R. and Maske, S., (eds.), Mineral deposits of Southern Africa Volume I*, *Geol. Soc. S. Afr.*, p923-929.

VAN NIEKERK, C.B. AND BURGER, A.J., 1978, A new age for the Ventersdorp acidic lavas: *Geol. Soc. S. Afr. Trans.*,

81(2), p155-163.

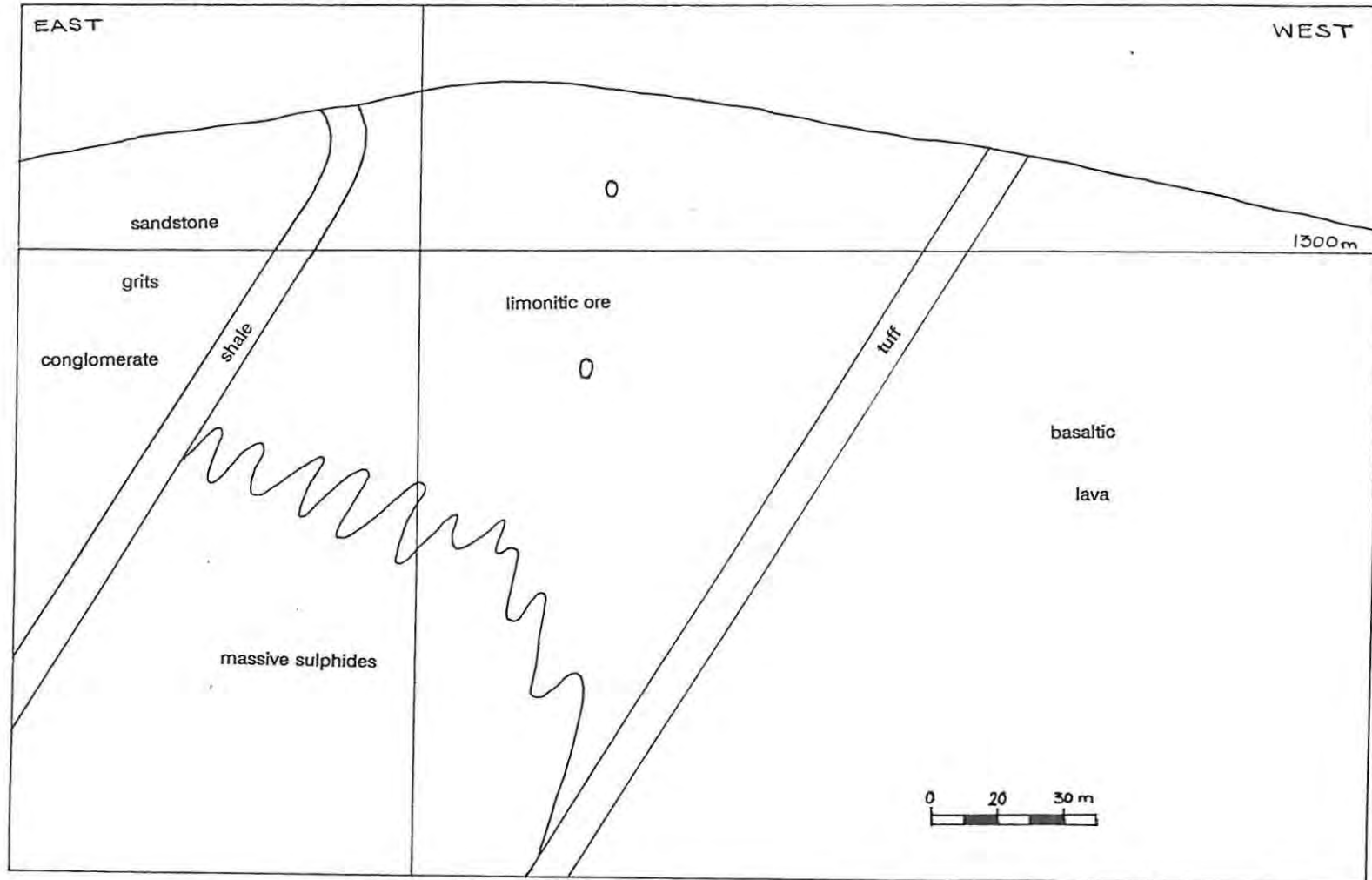
- VAN SCHALKWYK, J.F. AND BEUKES, N.J., 1986, The Sishen iron ore deposit, Griqualand West: *In: Anhaeusser, C.R. and Maske, S., (eds.), Mineral deposits of Southern Africa Volume I, Geol. Soc. S. Afr., p931-956.*
- WAGNER, P.A., 1928, The iron deposits of the Union of South Africa: *Geol. Surv. S. Afr. Mem., 26, 268pp.*
- WALTER, M.R. and HOFMANN, H.J., 1983, The Palaeontology and Palaeoecology of Precambrian Iron-formations, p373-394: *In: Trendall, A.F. and Morris, R.C., (eds.): Iron-formation: Facts and Problems, Developments in Precambrian Geology 6, Elsevier, Amsterdam, 558pp.*
- WINKLER, H.F.G., 1967, *The Petrogenesis of Metamorphic Rocks: Springer-Verlag, New York, 237pp.*
- WORST, B.G., 1962a, The geology of the Buchwa Iron Ore Deposits and adjoining country: Belingwe District: *S. Rhodesia Geol. Surv. Bulletin No. 53, Salisbury, 114pp.*
- , 1962b, The geology of the Mwanesi Range and the adjoining country: *S. Rhodesia Geol. Surv. Bulletin No. 54, Salisbury, 63pp.*
- YOUNG, T.P. AND TAYLOR, W.E.G., (eds.) 1989, Phanerozoic Ironstones: *The Geological Society Special Publication No. 46, London, 251pp.*

APPENDIX 1

<u>Mineral</u>	<u>Chemical Formula</u>
Autunite	$\text{Ca}(\text{UO}_2)_2(\text{PO}_4)_2 \cdot 10-12\text{H}_2\text{O}$
Carnotite	$\text{K}_2(\text{UO}_2)_2(\text{VO}_4)_2 \cdot 3\text{H}_2\text{O}$
Torbernite	$\text{Cu}(\text{UO}_2)_2(\text{PO}_4)_2 \cdot 8-12\text{H}_2\text{O}$
Tyuyamunite	$\text{Ca}(\text{UO}_2)_2(\text{VO}_4)_2 \cdot 5-10\text{H}_2\text{O}$

APPENDIX 2

East-West section of the Ripple Creek deposit, Zimbabwe, showing the relationship of the massive sulphides to the overlying limonite - a product of sulphide oxidation (COURTESY OF BUCHWA IRON MINING COMPANY)



APPENDIX 3

Parameters used in the evaluation of iron ore deposits
(after Ohle, 1972).

A. Geological Factors

1. Type and Grade: Direct shipping, wash, heavy media, taconite, etc.; Rice ratio, impurities, and associated elements.
2. Tonnage: Crude and product, effect on capital cost and recuperation schedules; weight recovery.
3. Grain-Size: Grind size, texture, liberation of ore minerals, elimination of impurities.
4. Grindability: KWH per ton to reduce the ore to concentrating and agglomerating sizes.
5. Mineralogy: Magnetite, hematite, goethite, silicate or carbonate. Impurity mineralogy—effect on the ability to separate the impurities in processing.
6. Distribution of Ore Types: Grades, textures, mineralogies—can selective mining be done?
7. Depth and Nature of Overburden: Sand and gravel, or rock. Open pit versus underground mining.
8. Shape and Attitude of the Ore Body: Tons per vertical foot. Effect on stripping ratio.
9. Location: Topographic effects, climate.

B. Non-Geologic Factors

1. Market and Price: Individual company variations in requirements and locations of furnaces.
 2. Politics and general business climate.
 3. Transportation cost.
 4. Labor and Housing: Availability and Cost.
 5. Construction cost.
 6. Power: Availability and Cost: Fuel
 7. Water Supply.
 8. Taxes.
 9. Royalty Rate.
 10. Inflation Factors.
 11. Tailings disposal.
 12. Environmental Factors: Ecology.
 13. Financing.
-

Appendix 4 :

Summary of Elements Found in Iron Ores Used by the American Iron and Steel Industry*

Source	Analyses, weight percent																		
	Fe	P	SiO ₂	Al ₂ O ₃	Mn	S	CaO	MgO	Ti	V	Cr	Zn	Mo	As	Pb	Sn	Co	Ni	Cu
Cuyuna	55.83 32.75	0.367 0.104	10.61 24.82	4.60 0.94	12.92 0.47	0.093 0.008	0.84 0.15	0.35 0.07	0.090 0.005	0.009 *	0.019 *	0.006 *	0.005 *	0.006 *	— —	— —	— —	0.002 *	0.011 *
Marquette	61.57 37.13	0.180 0.014	5.93 44.98	4.66 0.59	2.46 0.06	0.210 0.008	0.93 0.28	1.53 0.16	0.325 0.03	0.020 0.004	0.010 0.001	0.005 0.002	0.005 0.0018	0.006 0.002	* *	* *	* *	0.006 0.002	0.012 0.006
Menominee	59.60 37.11	0.465 0.015	4.33 44.57	3.97 0.89	5.69 0.05	0.259 0.006	1.49 0.24	4.20 0.18	0.235 0.065	0.035 0.016	0.015 0.002	0.041 0.012	0.004 0.0012	0.014 0.005	* *	* *	* *	0.020 0.009	0.025 0.007
Mesabi	63.20 49.07	0.118 0.028	3.18 21.68	11.48 0.45	5.48 0.07	0.073 0.005	0.72 0.09	0.21 0.07	0.21 0.008	0.010 *	0.018 *	0.007 *	0.005 *	0.008 *	* —	— —	* —	0.005 —	0.019 —
Steep Rock	60.69 50.80	0.033 0.019	2.97 10.35	2.36 0.98	0.28 0.10	3.60 0.016	0.23 0.08	0.22 0.04	0.30 0.01	0.012 0.006	0.072 0.013	0.007 *	0.004 *	0.03 *	— —	— —	0.10 —	0.02 *	0.10 *
Michipicoten	51.54 49.31	0.027 0.016	9.81 11.79	2.91 1.59	2.90 2.79	0.098	3.02 2.98	8.40 7.46	0.075 0.056	0.009 0.005	0.004 0.001	0.010 0.005	0.003 *	0.024 0.020	* *	0.075 *	* *	0.004 0.002	0.025 0.005
Eastern Penna. (Cornwall)	62.83	0.005	5.10	1.52	0.072	0.031	1.40	1.60	0.08	0.011	0.013							0.013	0.021
Texas (Lone Star)	47.52	0.18	15.34	6.90	0.32	0.10	0.15	0.030	0.026	0.032	0.020	—	0.004	0.013	—	—	—	0.008	0.004
California (Eagle Mt.)	55.00	0.080	11.00	1.60	0.08	0.40	1.80	3.50	0.13	0.003	0.005	—	0.001	0.005	—	0.003	0.020	0.005	0.05
Wyoming (Sunrise)	52.40	0.090	12.68	0.751	0.09	0.036	1.25	0.36	0.07	0.020	0.013	0.058	0.004	0.011			0.002	0.046	0.005
Utah	57.81	0.310	4.29	1.17	0.06	0.070					0.001			0.019				0.005	0.014
Tenn. C & I (Red Ore)	37.70	0.32	15.50	3.00	0.15	0.30	13.35	0.60	0.080	0.018	0.002	0.001	0.001		0.006	0.001	0.001	0.003	0.004
Tenn. Copper Co. (Sinter)	69.00	0.005	1.50	0.45	0.10	0.07	0.10	0.60		0.008		0.20	0.01				0.05	0.01	0.12
Venezuelan	64.85 63.00	0.13 0.05	0.98 0.80	2.45 1.80	0.10 0.08	0.043 0.020	0.017 *	0.084 *	.20 0.09	0.006 —	0.009 —	— —	* —	* —	— —	— —	* —	— —	0.003 *
Chile (Tofo)	60.30	0.042	7.49	1.76	0.09	0.021	1.41	2.00	0.59	0.28	0.008	—	—	—	—	—	—	0.15	0.010
Labrador	58.00	0.058	8.40	0.84	1.30	0.012	0.17		0.005	—	0.001	0.008	—	0.001	—	—	0.001	—	0.005
Brazilian (Itabira)	68.50	0.035	0.39	0.76	0.05	0.008	0.01	0.03	0.014	—	0.008	0.005	0.004	—			—	0.006	0.018
Swedish (Kiruna)	66.35	0.485	1.97	0.50	0.11	0.006	2.24	0.94	0.040	0.074			0.010				0.022	0.009	0.004
Liberian (Bomi Hills)	69.50	0.055	0.11	0.19	0.16	0.011	0.03	0.15	0.04	—	0.012	0.002	0.003	—			—	0.002	0.004

— None
*Trace

* (Modified from Jacobs et al. (1954), Table 1. in *Ore*, 1972)