

**RHODES UNIVERSITY
DEPARTMENT OF GEOLOGY**

**OVERVIEW AND COMPARISON OF BESSHI-TYPE
DEPOSITS : ANCIENT AND RECENT**

DISSERTATION

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ABSTRACT

Besshi-type deposits range in age from early Proterozoic to early Tertiary, of which the largest number are late Proterozoic, early Palaeozoic or Mesozoic in age. No Archaean examples of Besshi-type deposits are known, probably due to insufficient availability of sialic crust for erosion and clastic marine sedimentation before the start of the Proterozoic. All Besshi-type deposits are contained within sequences of clastic sedimentary rock and intercalated basalts in a marine environment. The basalts and amphibolites are principally tholeiitic in composition.

Besshi-type deposits characteristically form stratiform lenses and sheet-like accumulations of semi-massive to massive sulphide. The main ore assemblage consists dominantly of pyrite and/or pyrrhotite with variable amounts of chalcopyrite, sphalerite and trace galena, arsenopyrite, gold and electrum, barite being absent in general. The median Besshi-type deposit (n=75) contains 1.3 million tonnes (Mt) of massive sulphide with a Cu grade running at 1.43%.

It is suggested that Besshi-type deposits form by both exhalative and syngenetic replacement processes when considering geological features and comparisons with modern analogues in the Guaymas Basin, Middle Valley and Escanaba Trough. The currently forming metalliferous sediments in the Red Sea provide for a brine pool model explaining the lack of footwall feeder zones below sheet-like deposits. Where thick sulphide lenses are contained in some Besshi-type deposits, combinations of exhalative precipitation and sub-sea-floor replacement of permeable sediments and/or volcanic rocks, take place in the upper parts of submarine hydrothermal systems.

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1 INTRODUCTION

This dissertation consists of three main sections, i.e. the ancient Besshi-type deposits in Japan and their equivalents in Canada, Namibia and South Africa, the recent massive sulphide deposits on the sea floor in the Guaymas Basin, Middle Valley, Escanaba Trough and Red Sea, together with a final section dealing with genetic and descriptive features of these deposits that may assist in exploring for Besshi-type deposits.

The bedded cupriferous iron sulphide deposits in Japan have been studied since the 1890's. Workers such as Kochibe (1892) and Nakajima (1893) have studied the features of the Besshi mine. A textbook on ore deposits, published by Watanabe (1895), proposed a sedimentary origin on the conformable copper-pyrite deposits. Kato (1937) applied the term "Besshi-type deposits" to bedded cupriferous iron sulphide deposits of the Besshi district, Shikoku Island, Japan (Kanehira and Tatsumi, 1970). Mitchell and Bell (1973) adopted this term used by Kato (1937) and introduced it to the Western literature as a specific class of massive sulphide deposit. These authors recognised the association of Besshi-type deposits with mafic volcanic rocks and large volumes of carbonaceous mudstones, clastic limestones, quartz-rich sediments and clastic volcanic rocks typical of deepwater deposition. These rocktypes were proposed by Mitchell and Bell (1973) to originate on the flanks of island-arc volcanoes in convergent plate settings. Mitchell and Garson (1976) agreed with this finding.

Sawkins (1976) suggested that Besshi-type deposits developed in tensional tectonic environments. Derkmann and Klemm (1977) described their second type of Alpine copper-pyrite deposits which are contained within a metasedimentary-greenstone sequence as a Kieslager-type deposit. Klau and Large (1980) postulated that Kieslager- and Besshi-type deposits could be similar but the extreme structural deformation hinders a more refined classification. These two authors further pointed out the similarities between the metamorphosed basic volcanics and sediments of the Khetri copper belt in India (Das Gupta, 1970) and Besshi-type deposits. Hutchinson (1980) described Besshi- or Kieslager-type deposits as a

separate class under his exhalative volcanogenic group with its plate tectonic setting as a fore-arc trough or trench. Franklin *et al.* (1981) summarised Besshi- and Kieslager-type deposits separately, suggesting that the Kieslager-type deposits' association with ophiolitic volcanic rocks may be seen as a distal variety of Cyprus-type deposits.

Fox (1984) compiled the first detailed review of Besshi-type deposits with the emphasis on the type geology, metal contents, lead isotopes and relating their setting to modern, actively forming deposits in the Gulf of California. Koski (1990) compared the massive sulphide deposits from the sediment-covered spreading axes in the northeast Pacific Ocean with the Besshi-type deposits in the Sanbagawa metamorphic belt and the Shimokawa mine in the Hidaka belt, Japan. Slack (1993) gave a comprehensive overview on world wide occurrences of Besshi-type deposits, concentrating on their geological setting and ore composition, integrating it into a descriptive model and comparing these deposits to possible analogues from the northeast Pacific Ocean.

2 ANCIENT DEPOSITS

2.1 Japan

2.1.1 Besshi district, Shikoku, Japan

2.1.1.1 Tectonic evolution

The Besshi deposit and several other deposits are situated within or associated with the Sanbagawa metamorphic belt in central Shikoku, Japan. The Sanbagawa metamorphic belt comprises part of the Inner Zone of southwest Japan (Fig. 1). The northern border of the Sanbagawa metamorphic belt is defined by the Median Tectonic Line (MTL), which constitutes a regionally significant transcurrent fault which separates the Outer and Inner Zones of southwest Japan (Fig. 1). This belt is represented by the Oboke nappe and structurally overlying, internally imbricated, Besshi nappe complex.

The outcropping Sanbagawa metamorphic rocks in central Shikoku have been subdivided into four "formations". These include the Kawaguchi, Koboake, Minawa and Ojoin "Formations" in ascending tectonostratigraphic order (Dallmeyer and Takasu, 1991). Mapping and geochronological studies suggest that this segment of the Sanbagawa terrane actually consists of several internally imbricated nappe complexes namely the Oboke (lower) and Besshi (upper) nappe complexes and are separated by thrust contacts. The Oboke nappe complex is composed largely of psammitic schists (Koboake "Formation") with interfingering pelitic schists (Kawaguchi "Formation"). The Besshi nappe complex comprises pelitic, basic and siliceous schists (Minawa "Formation") and pelitic schists (Ojoin "Formation") (Fig. 3).

Three deformational episodes have been described within the Sanbagawa metamorphic belt in central Shikoku. The first phase (D1) involved ductile shearing (E-W stretching lineation) during peak metamorphic conditions. A second phase (D2) occurred after growth of peak metamorphic minerals and locally involved south-vergent thrusting. A third (D3) and final phase of deformation involved

upright folding along approximately E-W-trending axial surfaces with only local development of non-penetrative crenulation cleavages. Megascopic D3 folds are developed in central Shikoku and control the regional exposure of the Oboke and Besshi structural units (Dallmeyer and Takasu, 1991).

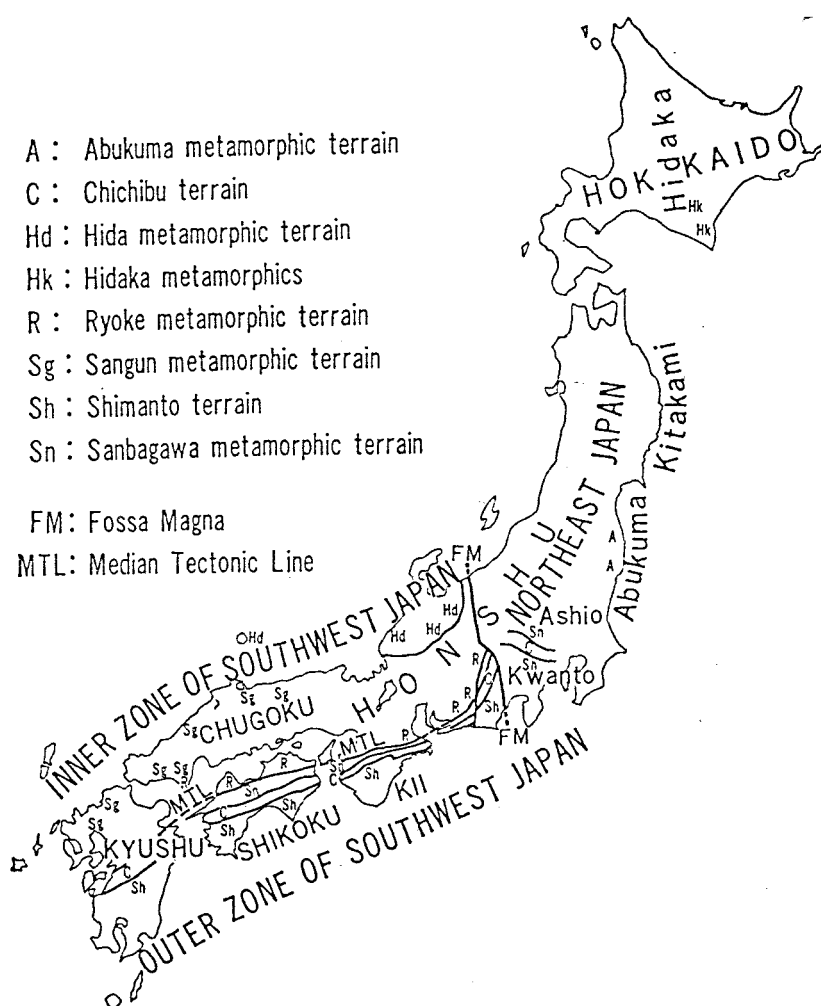


Fig. 1 Important geological terrains in Japan (After Tatsumi *et al.*, 1970).

The metallogenic epochs of the Japanese Islands are grouped into five major epochs by Tatsumi *et al.* (1970), i.e. Epoch 1 is the Permian-Carboniferous period in which the stratabound deposits such as the bedded cupriferous iron sulphide (Besshi-type) deposits were formed (Fig. 2). Epoch 2 comprises the Late Jurassic to Early Cretaceous period, in which the stratabound deposits mentioned above

2.1.1.2 Besshi deposit

The Besshi copper-pyrite deposit (Fig. 2) is conformable with the surrounding crystalline schists of basic and pelitic rock of the middle member of the Minawa Formation (Fig. 3). The ore body is tabular in shape and averages 3 metres in thickness along a strike length of 1800 metres. The ore body dips steeply northward and extends down dip for more than 2500 metres (Kase, 1988) (Fig. 4). It consists of two layers of massive sulphide ore and a layer of banded ore between them.

The ores are classified into three types, namely compact ore, banded ore and copper-rich ore. The compact ore consists mainly of pyrite, chalcopyrite, sphalerite and a small amount of gangue. The banded ore comprises thin alternations of sulphide-rich and quartz-chlorite-rich layers, while the copper-rich ore is essentially composed of chalcopyrite with minor pyrite and gangue minerals. The schistosity is conspicuous in the banded ore and compositional banding is found even in the compact ore (Kanehira and Tatsumi, 1970). The ore body is tightly folded in an isoclinal disposition together with the country rocks. Magnetite-garnet-quartz schists occurs in thin layers in both hanging wall and footwall of the sulphide ore body.

The Besshi deposit was subjected to glaucophanitic regional metamorphism in the late Mesozoic with a metamorphic facies transition from glaucophane-schist facies to epidote-amphibolite facies. Contact metamorphism was superimposed on the ore body from the 18th level downwards, which was probably caused by the thermal effect derived from a plutonic intrusion hidden in the deeper part of the mine (Kase, 1988). Pyrite was converted to pyrrhotite polytypes and the Fe contents of pyrrhotite change progressively with increasing depth from the monoclinic type (levels 18 to 25) to the Fe-rich hexagonal type with exsolved troilite (below 30 level) through a mixture of monoclinic and hexagonal types.

Formation			Lithology *	Thickness	Volcanic materials **	Sulphide deposits **
Yoshinogawa Group	Upper Sub-gr.	Ozyoin Formation	pelitic schist (Basic & quartzose sch.)	600-1080m		
	Middle Sub-gr.	Minawa Formation	Upper Member Pelitic & basic schists (Quartzose sch.)	820-1170m		7...
			Middle Member Basic schist (Quartzose & pelitic sch.)	500-2500m		73
			Lower Member Pelitic schist (Basic & quartzose sch.)	100-1600m		9
		Koboke Formation	Psammitic schist (Pelitic & quartzose sch.)	370-1180m		3
	Lower Sub-gr.	Kawaguchi Formation	Pelitic schist (Basic & quartzose sch.)	500-1500m		
		Oboke Formation	Psammitic schist (Schistose conglomerate)	300m		

Fig. 3 Stratigraphy of the Sanbagawa crystalline schists in Shikoku, Japan. The rocks in parenthesis are subordinate. Abundance of volcanic materials and sulphide deposits is schematically shown. Numerals indicate the number of sulphide deposits (After Kanehira and Tatsumi, 1970).

Intense Sn-As-Zn-Ag vein mineralisation was located at the 26th level of the Besshi mine which formed at different stages and temperatures. The previously mentioned hidden plutonic intrusion probably caused this mineralisation with a regional zoning of the vein mineralisation with Sn and As minerals forming in the centre and Sb minerals away from the district's centre.

Yamamoto *et al.* (1984b) established the range of $\delta^{34}\text{S}$ values for the Besshi deposit to vary from +0.8 to +3.4 per mil. These data which are consistently low, indicate that a substantial basaltic sulphide component with $\delta^{34}\text{S}$ values near 0 per mil, are incorporated in the ores of this Besshi-type massive sulphide deposit.

Sato and Sasaki (1980) established that the lead isotopic signatures of the Besshi-type deposits in Japan are variable ($\delta^{206}\text{Pb}/\delta^{204}\text{Pb}$ 17.47 - 19.38; $\delta^{207}\text{Pb}/\delta^{204}\text{Pb}$ 15.39 - 15.64; $\delta^{208}\text{Pb}/\delta^{204}\text{Pb}$ 37.25 - 38.78) and are consistently deficient in $\delta^{207}\text{Pb}$ relative to the conformable ore type, with a small but significant isotopic variation existing even within a single deposit. The data obtained from the Shimokawa area, the Outer Zone and Inner Zone Besshi-type deposits, show a large contribution of mantle-derived lead with a smaller contribution from the surrounding clastic sedimentary-derived lead during submarine hydrothermal alteration.

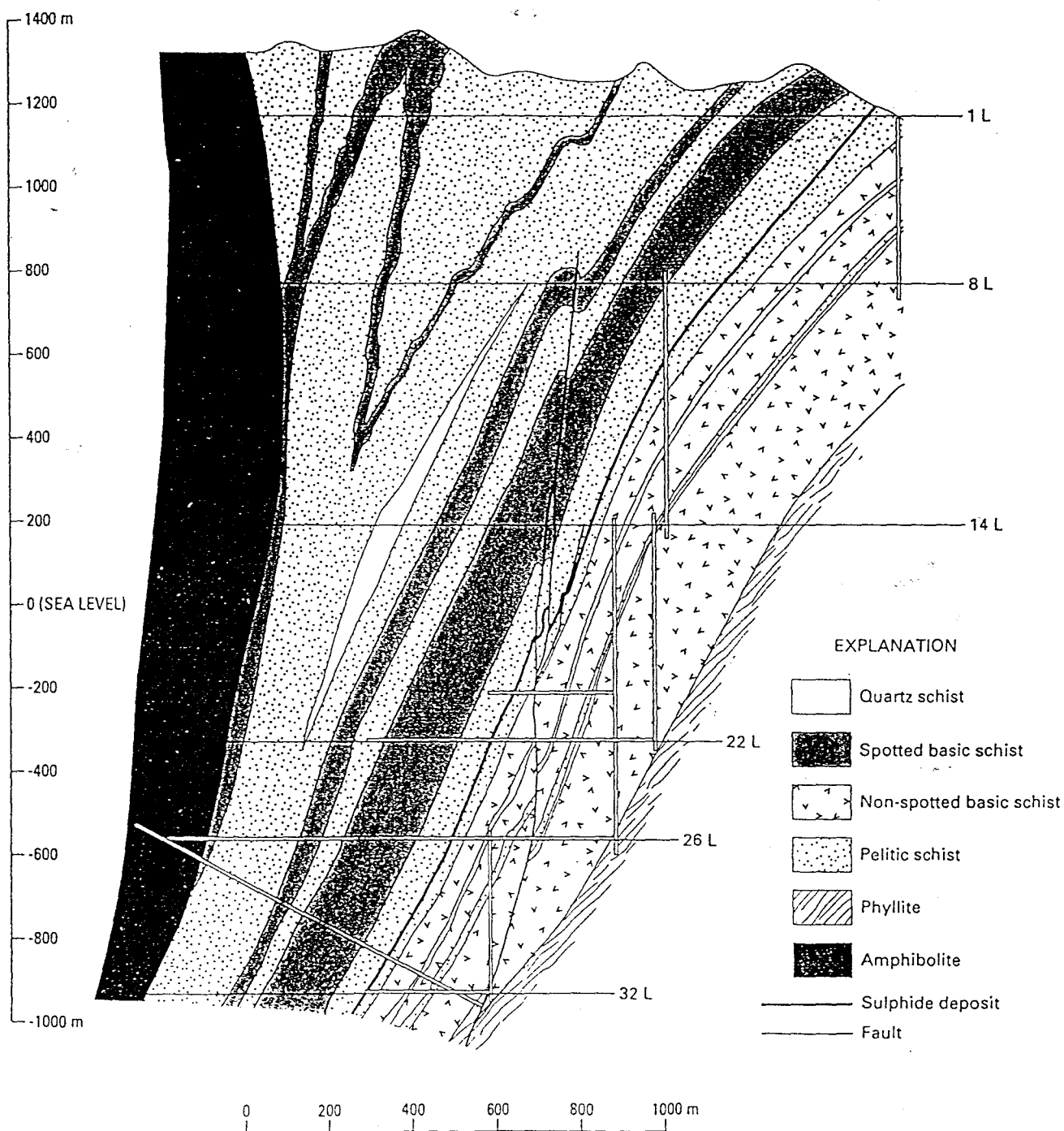


Fig. 4 Cross-section of the Besshi massive sulphide deposit (After Slack, 1993).

2.1.1.3 Shirataki deposit

The Shirataki mine is located 17 kilometres southeast of the Besshi mine (Fig. 2). The ore deposit lies in the Minawa Formation's upper member which is composed of basic, pelitic and garnetiferous quartzose schists, together with small bodies of serpentinite (Fig. 3). The ore body is associated with basic schist, stretching for more than 4 000 metres along the plunge of linear-structured wall rocks and down dip from 500 metres in the eastern part to 800 metres in the western part of the ore body. The ore body is closely related with the folded country rocks. Several ore shoots which are associated with fold axes are found in the Shirataki mine. Each shoot varies in size and form, probably due to various structural disturbances.

Several kinds of sulphide ores such as compact, banded, disseminated, chalcopyrite-rich and tongue ores are found in the Shirataki deposit (Takeda, 1970). The compact and banded ores are most abundant, consisting mainly of pyrite with small quantities of chalcopyrite, sphalerite, bornite and chalcocite. Banded ores exhibit the intraformational isoclinal folded structure of the area. The disseminated ore, which is developed in some marginal parts of compact ore bodies, consists of disseminated pyrite with considerable amounts of magnetite and hematite. The disseminated ore occasionally consists predominantly of chalcopyrite and bornite. Disseminated ores have chalcopyrite, sphalerite, quartz and chlorite frequently occurring as the pressure shadow around semi-idiomorphic crystals of pyrite and magnetite. The chalcopyrite-rich ores are often found in some compact ores and at the peripheries of ore bodies and mainly comprise chalcopyrite and pyrite with sphalerite and bornite. The tongue ores consist predominantly of bornite and chalcocite. This ore-type occurs as offshoots from the main ore bodies as veinlets and are oblique to the planar structure of the wall rocks. The tongue ore consists of bornite, chalcocite and chalcopyrite with pyrite, covellite, native silver and stromeyerite. The chalcopyrite-rich and tongue ores appeared to be formed through the metamorphic mobilisation of the compact ores. The gangue

minerals in the ores are quartz, soda amphibole, chlorite, epidote, muscovite, garnet and carbonate.

2.1.1.4 Sazare deposit

The Sazare mine is situated 11 kilometres northeast of the Shirataki mine (Fig. 2). The deposit lies in the Minawa Formation's green schist member which consists of basic, pelitic and quartzose schists (Fig. 3). The wall rocks comprise schists rich in chlorite, epidote and muscovite with soda amphibole which appears to be in equilibrium with chlorite and other minerals. The Sazare district belongs to the warped part of the southern wing of the Yakushi anticline and the main planar structure exhibits steep monoclinic dip to the south. The linear structures plunge variably from horizontal to eastward at a low angle, whereas the country rocks near the ore horizon are controlled by isoclinal folding together with high-angled lineations.

The Sazare mine comprises several ore bodies such as the Kinsha, Kinsen, Kongo, Kinritsu, Sazare and Shinsen ore bodies. The deposit extends for approximately 2 500 metres along strike and down dip for 2 000 metres. The characteristics of the various ores in this deposit are very similar to those of the Shirataki deposit. The pyrite grain size in the compact ores however is finer than that of the Shirataki deposit (Takeda, 1970). In general the grain size of pyrite is variable depending on the metamorphic grade in the stratified deposits of the Sanbagawa metamorphic terrain.

2.1.1.5 Iimori deposit

Massive copper-pyrite deposits of the Iimori area occur concordantly in the Sanbagawa crystalline schists (Fig. 2) which are divided into three formations in the area, namely the Ryumon, Iimori and Tomobuchi Formations. The Iimori Formation consists of abundant basic schists and subordinate amounts of pelitic and quartzose schists, where the other two formations comprise abundant pelitic, psammitic and some basic schists. The total thickness of the formations exposed in the district is estimated at about 3 500

metres. The district is bound in the north by the Median Tectonic Line and to the south the Sanbagawa crystalline schists grade into weakly metamorphosed Palaeozoic rocks. The localities of the deposits are largely confined to the Iimori Formation and all deposits are associated with basic schists.

The largest deposit of the area which is found in the Iimori mine is thin (from 0.2 to 2.8 metres thick), extends down dip for 250 to 300 metres and extends for more than 7 000 meters along its plunge which is nearly parallel to the linear schistosity of the country rocks. The ore body is gently undulating in the central area but was subjected to repeated cleavage folds in its western part (Kanehira, 1970). The other deposits are much smaller in dimensions and are all bedded or flat lenticular in form.

The ores are classified into massive and banded ores. The massive ores are composed of abundant sulphides such as pyrite, chalcopyrite, sphalerite and bornite with minor gangue minerals. The banded ores consist of pyrite and gangue minerals and minor amounts of chalcopyrite and sphalerite. Quartz, carbonate, albite and chlorite are common gangue minerals in most banded ores.

The grain size of pyrite from the deposits in the higher metamorphic grade (Iimori and Amano mines) is coarser compared with that of the lower metamorphic grade (Hosokawa mine). The quartz and silicates show a prominent preferred orientation similar to those of the surrounding schists. Pyrite crystals of both massive and banded ores are elongated parallel to the lineation imprinted on the ores and wall rocks.

2.1.1.6 Yanahara deposit

The Yanahara mine lies in the lower part of the middle formation of the Palaeozoic Maizuru Group in the Yanahara mining district (Fig. 2). The lower part of the middle formation mainly consists of massive slates with some acidic volcanic rock layers, which are 30 metres thick at the Yanahara mine. The Maizuru Group has been intruded by the Yakuno sheet-like intrusive rocks which parallel the

Palaeozoic terrain and consist of gabbroic, diabasic, granodioritic and felsic rocks (Mitsuno, 1988).

The Maizuru Group of the district has undergone three phases of folding. During the first phase of deformation, the Maizuru Group, the Yanahara deposit and the Yakuno intrusive rocks were shear folded in a NNW-SSE trend. During the second stage of deformation, the Yanahara ore deposit and its host rocks, together with the Lower to Middle Triassic Fukumoto Group, were shear folded with a component of flexural-slip along a NNW-SSE trend. During the third phase of deformation, the ore deposit and its host rocks with the Upper Triassic Nakaiso Formation were subjected to flexure-slip folding with an E-W orientation (Mitsuno, 1988).

The ore deposit of the Yanahara mine comprises the Main deposit which consists of the first, second, third, Lower and the Shinbu orebodies and a number of satellite ore deposits. The Main ore deposit is approximately 500 metres wide, extends down dip for 1 000 metres and varies in thickness from 30 to 100 metres (Fig. 5).

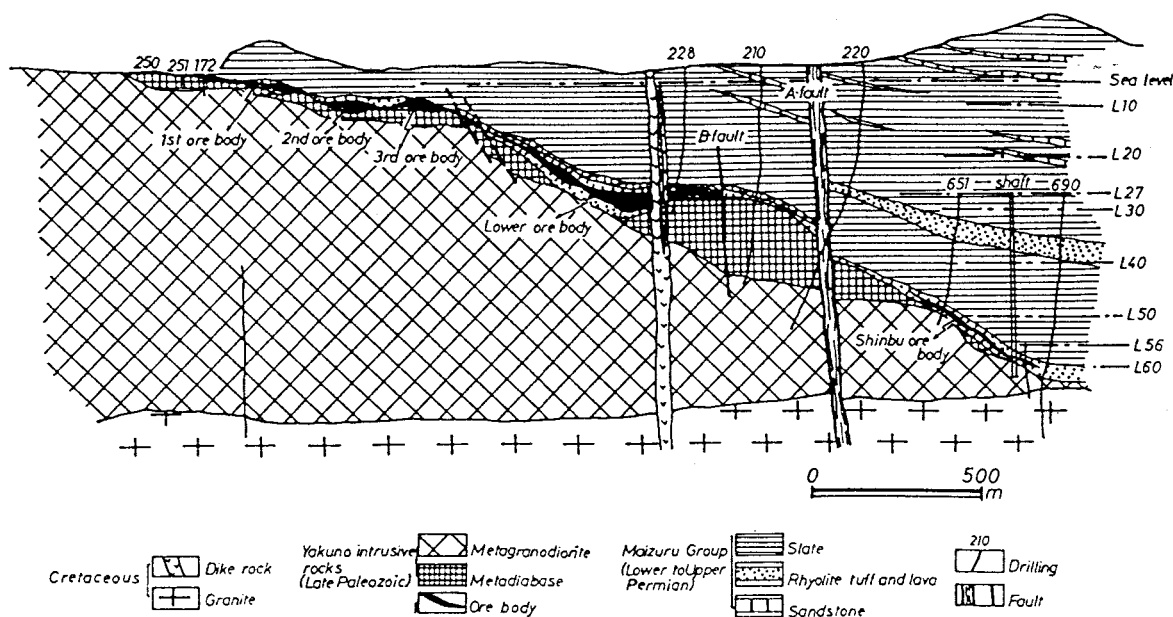


Fig. 5 Geological section of the Main ore deposit, Yanahara mine (After Mitsuno, 1988).

The satellite bodies are situated on the western and eastern side of the Main ore body. These ore bodies consist largely of pyrite with subordinate chalcopyrite ore. The copper content of the western satellite ore bodies is higher than that of the Main ore deposit. The ore body is composed of disseminated pyrite ore in rhyolitic lava, laminated pyrite ore alternating with thin layers of rhyolitic tuff and coarser-grained massive pyrite with fine- to medium-grained pyrite ore layers from the bottom to the top. The coarser grained pyrite ore, which forms the main component of the ore body, includes clearly laminated, fine- to medium-grained pyrite ore layers several centimetres thick. The strike and dip of these layers indicate that the ore body had been a single layer with the fold axis trending NNW-SSE, while being boudinaged into isolated lenses whose fold axis trends E-W.

The Yanahara ore deposit was influenced by contact metamorphism of Cretaceous granitic rocks and subsequent Late Cretaceous to Early Tertiary dykes. Granitic rocks comprise granodiorite and adamellite with hydrothermal mineralisation occurring after contact metamorphism took place. Ore minerals in these rocks are chalcopyrite and sphalerite.

Yamamoto *et al.* (1984a) found that notwithstanding the difference in age and acidity of the volcanic host rocks, the Besshi-type massive sulphide deposits of the Yanahara mining area show a remarkable uniformity in sulphur isotopes, all $\delta^{34}\text{S}$ values fall in a narrow range from +0.6 to +2.9 per mil. This implies that all these deposits have a common source of sulphur and sulphide, possibly dominated by sulphur derived from basaltic magmas in depth. Barite samples investigated from the Yanahara deposit itself yielded $\delta^{34}\text{S}$ values ranging from +12.1 to +15.1 per mil, which suggest that the coeval Permian seawater sulphate was the principal source for the sulphur in the barite samples.

2.1.1.7 Makimine deposit

In southern Kyushu, the Mesozoic sedimentary complex of the Shimanto terrain is divided into four geological units. The lower two groups

are those of the Kitagawa (Upper Triassic/Jurassic) and Kawabe Groups (Jurassic/Early Cretaceous). The correlatives of these groups in the Makimine mining district (Fig. 2) are the Yoshinomoto, Makimine and Hinokage Formations. Bedded cupriferous iron sulphide deposits of the Makimine mine occur in the Makimine Formation which consists of phyllites, schistose sandstone, alteration products of chert and phyllite, ferruginous phyllite and "green rocks", which were identified as products from diabase, gabbro and porphyrite. The general relation between the clastic sediments and "green rocks" seems to be concordant, but often shows local interfingering relationships. All these lithologies were regionally metamorphosed into low-grade greenschist facies.

The two major ore deposits arranged in parallel in the Makimine mine have dip extents of 2 800 metres, widths of 200 metres and thicknesses of 30 metres for the larger deposit and 1 300 metres, 500 metres and 30 metres for the smaller deposit, respectively. The individual parts of the ore bodies are a few tens to several hundred meters in length, 20 to 150 metres wide and 0.5 to 3 metres in thickness arranged in an overlapping *en échelon* pattern. The ore bodies occur in conformable relation to the phyllitic rocks, but local branchings and cross-cuttings are observed in many places. The principal ores are pyrite, chalcopyrite, pyrrhotite and sphalerite, associated with variable amounts of magnetite, cubanite, quartz, chlorite, green hornblende and garnet (Tatsumi *et al.*, 1970). The deposits are intruded by granitic rocks of Miocene age where the thermal metamorphism resulted in the formation of cubanite and other metamorphic minerals in the ores.

2.1.2 Shimokawa district, Hokkaido, Japan

2.1.2.1 Tectonic evolution

In Sikhote-Alin and Sakhalin, three tectonic units are recognised from the continent towards the sea, namely Eastern Sikhote-Alin volcano-plutonic belt, Western Sakhalin forearc basin and Eastern Sakhalin folded accretionary system (Fig. 6) (Miyake, 1988).

Such a zonation is similar to a modern convergent plate margin. Before the Cenozoic opening of the marginal Sea of Japan, each zone extended southward to Hokkaido. The southern extension of the Eastern Sakhalin folded accretionary system in Hokkaido widens to about 100 kilometres and is subdivided into three zones from west to east, the Kamuikotan zone, Sorachi zone and the Hidaka zone (Miyake, 1988). The Hidaka zone is underlain by the Cretaceous Hidaka Supergroup which is subdivided into the lower Shimokawa Formation and the upper Iwaonai Flysch.

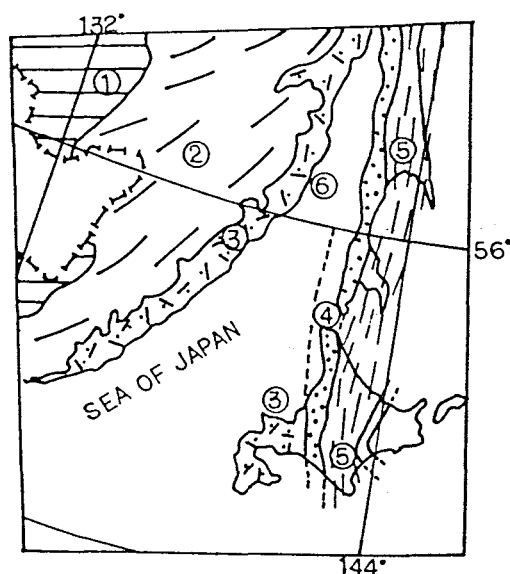


Fig. 6 Cretaceous tectonic zones of Soviet Far East and Hokkaido. 1 - Bureya-Khanka megablock, 2 - Megablock with Precambrian basement, 3 - Eastern Sikhote-Alin-Rebun-Kabato volcano-plutonic belt, 4 - Western Sakhalin-Ishikari-Teshio forearc basin, 5 - Eastern Sakhalin-Hidaka accretionary prism, 6 - Axes of the trenches (After Miyake, 1988).

The central and eastern parts of Hokkaido were the site of thick sedimentary accumulations of the Hidaka geosyncline, mainly in the Mesozoic. The oldest formations located in the geosynclinal sediments seemed to have been formed in the Permian. The Jurassic geosynclinal formations which form the Hidaka Complex comprise the Nakanosawa, Kamui and Sorachi Groups. The Sorachi Group consists of a lower part containing diabasic tuff, diabase, chert and limestone, while the upper part consists of chert with little diabasic tuff, containing fossil evidence of the Late Jurassic to early Cretaceous. The thickness of the Sorachi Group is about 2 300 metres. The

Nakanosawa Group is composed of slate, black sandstone and conglomerate, while the Kamui Group comprises slate and sandstone with intercalated chert, limestone and diabasic tuff. The Kamui Group is up to 6 000 metres thick.

The Hidaka orogenic terrain was developed in the Cretaceous and very intensely in the Latest Cretaceous to Early Palaeogene, when the glaucophanitic Kamuikotan metamorphic belt with intrusions of serpentinite was formed to the west of the central Hidaka zone and the Hidaka metamorphic zone to the southeast, in which schistose hornfelses and migmatites occur together with intrusive granitic rocks and gabbros. The sediments of flysch-type and of coal-bearing formations were deposited on either side of the Hidaka terrain in the Late Cretaceous and the Palaeogene respectively, and sediments of molasse type were deposited especially on the western side in the Miocene (Tatsumi *et al.*, 1970).

2.1.2.2 Shimokawa deposit

The Shimokawa deposits are located in the northern part of the Hidaka orogenic belt, Hokkaido (Fig. 7). The Hidaka belt can be divided from west to east into the Hidaka western greenstone belt, Greenstone-chert-turbidite facies belt, extending northwards to include the Shimokawa mining area, the Hidaka metamorphic belt and the Turbidite facies belt (Fig. 8) (Mariko and Kato, 1994).

The host rocks of the Shimokawa deposits are considered to be a part of the accretionary complex formed during the Cretaceous to Paleogene. The Shimokawa mining area, underlain by the Hidaka Supergroup, is subdivided into three parts by two faults trending NNW. The Hidaka Supergroup in the middle part of the area consists mainly of an apparent ophiolitic suite, which comprises serpentinite, pyroxenite, gabbro, diabase and massive and pillowed basalts. Numerous intercalations of slate, sandstone and rare limestone also occur. The boundary features between massive basalts or diabase and slate or sandstone appear to indicate that the mafic magma intruded into soft sediments.

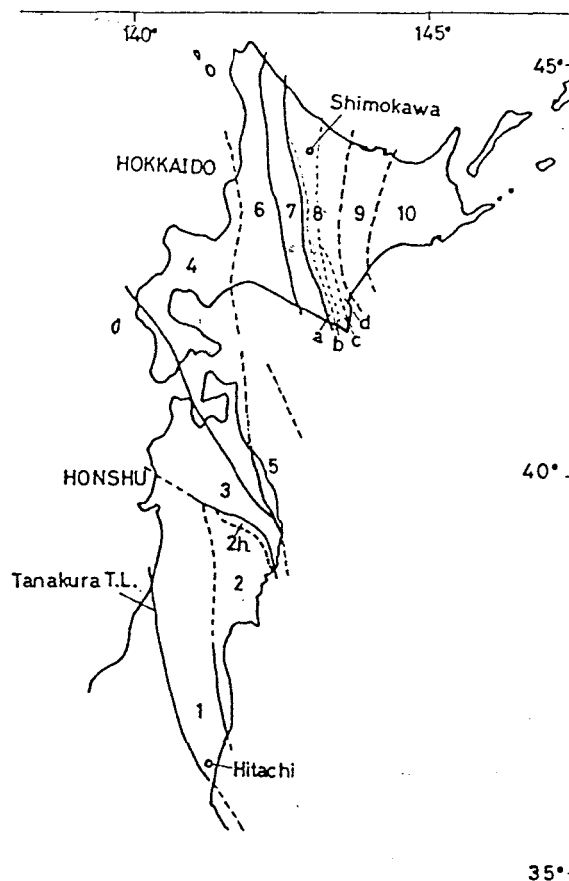


Fig. 7 Map of pre-Neogene tectonic division of northern Japan and locations of the Shimokawa and the Hitachi mines. 1 - Abukuma belt, 2 - South Kitakami belt, 2h - Hayachine structural belt, 3 - North Kitakami belt, 4 - Iwaizumi belt, 5 - Taro belt, 6 - Ishikari belt, 7 - Kamuikotan belt, 8 - Hidaka belt (a - Hidaka western greenstone belt, b - Greenstone-chert-turbidite facies belt, extending northwards to include the Shimokawa mining area, c - Hidaka metamorphic belt and d - Turbidite facies belt), 9 - Tokoro belt, Nemuro belt (After Mariko and Kato, 1994).

Gabbro, troctolite and granite of Miocene age intruded into the Hidaka Supergroup in the southern part of the area.

The mafic and sedimentary rocks are altered and the assemblage of replacing minerals ranges from prehnite-pumpellyite facies to a transitional greenschist-amphibolite facies. The chemical compositions of amphibole and chlorite formed by the alteration are analogous to those minerals formed at modern mid-oceanic ridges (Mariko and Kato, 1994).

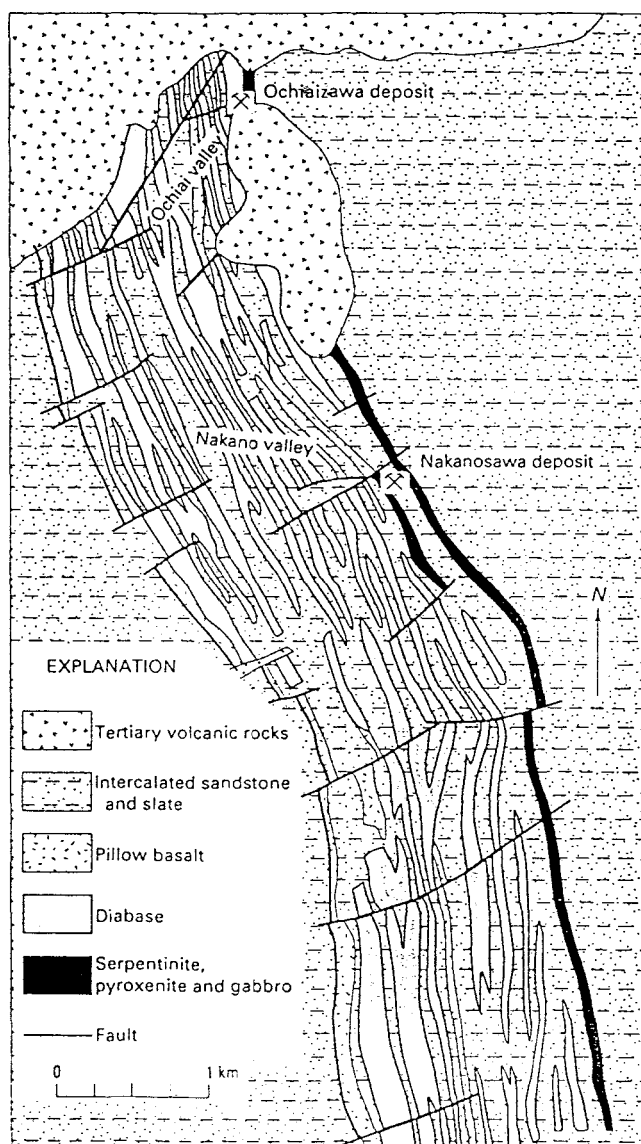


Fig. 8 Geological map of the Shimokawa mining area, Hokkaido, Japan (After Slack, 1993).

The Besshi-type massive sulphide Ochiaizawa and Nakanosawa deposits of the Shimokawa mining area occur in the middle part of the Hidaka Supergroup which was subjected to a transitional greenschist-amphibolite facies alteration. The Ochiaizawa deposit lies in the northern part of the mineralised zone and is composed of several ore bodies which are arranged in an *en échelon* pattern. Each of the ore bodies extends along strike and is lenticular in cross-section. In general the ore bodies occur at or near the

contact between slate and overlying diabase or pillow basalt. The Nakanosawa deposit, approximately 2.5 kilometres to the south, has not been exploited due to low copper prices experienced.

The mineral deposits of the Shimokawa mine, which occur as several shoots, have a reverse S-shape in plan and are steeply dipping, have a strike length of more than 2 800 metres, dip extent of 800 metres and an average thickness of about 7 metres with a maximum thickness of 25 metres.

The Shimokawa deposits consist of massive ore and banded ore. The massive ore is subdivided into high grade compact ore and low grade plain iron sulphide ore while the banded ore comprises sulphide chlorite-quartz laminated and disseminated ore (Fig. 9). The ores are dominantly composed of pyrite, marcasite, pyrrhotite, chalcopyrite, with minor sphalerite, magnetite, cubanite, valleriite, cobalt-pentlandite, electrum, quartz and carbonates (Miyake, 1988).

Bamba and Motoyoshi (1985) identified two phases of mineralisation, the early phase being cubic or colloform pyrite, whereas the subsequent phase is marked by the presence of chalcopyrite-pyrrhotite associated with sphalerite, worm-like pyrite, cubanite, mackinawite and cobaltpentlandite as accessory minerals. These authors established that a geochemical zonation is present at the mine, indicating that the later phase of mineralisation varied methodically from the base in the south towards the top of the orebodies in the north.

Mariko (1988) stated that the Shimokawa copper-pyrite deposits were formed from sub-sea floor circulating hydrothermal fluids which altered the surrounding country rocks. The country rocks were formed within a submarine environment associated with the development of new oceanic crust and an adequate supply of clastic material from the adjacent continental areas. Mariko (1988) saw the ore genesis as follows:

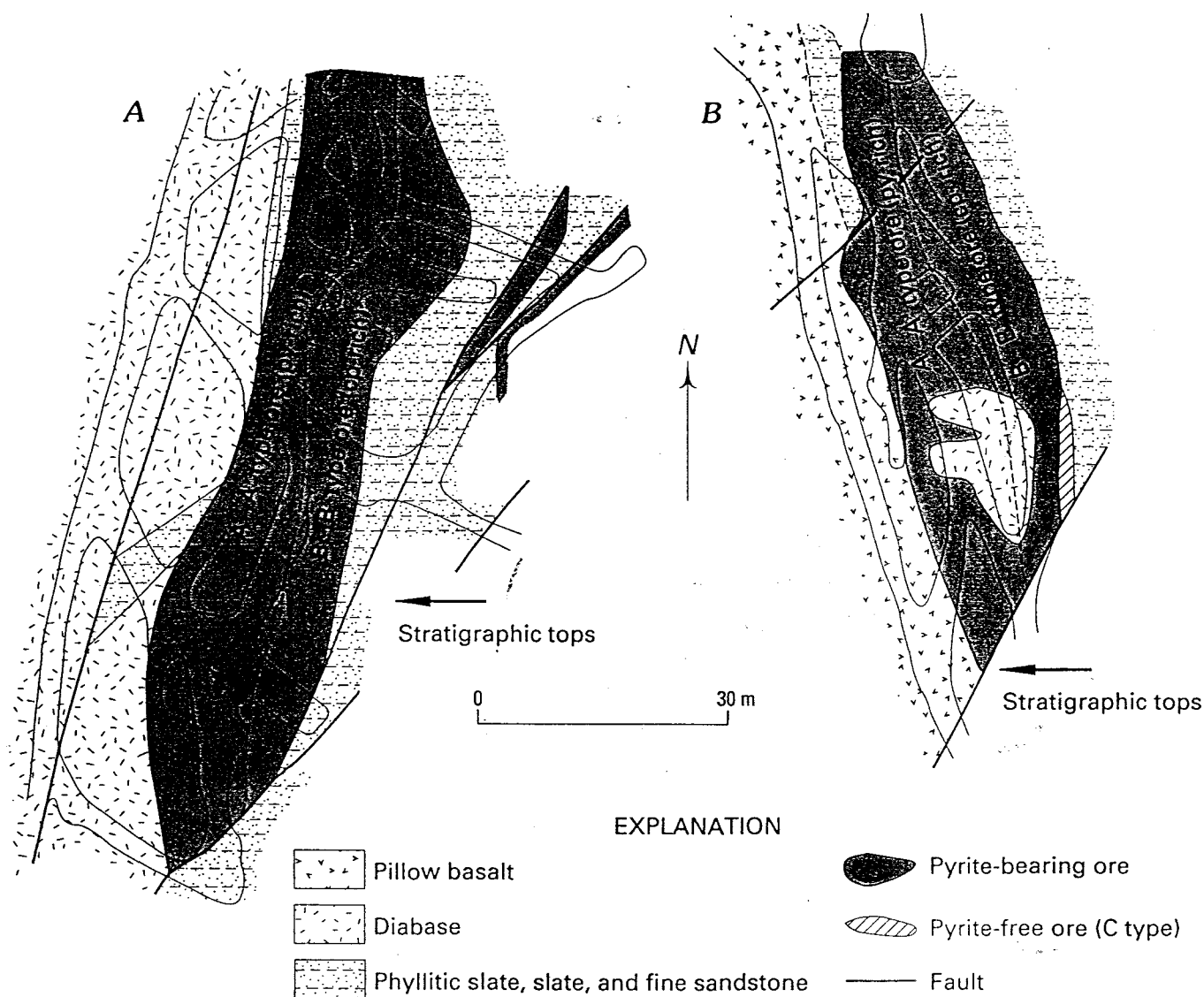


Fig. 9 Geological plan of the Shimokawa mine (Ochiaizawa deposit), indicating the geological relations of massive sulphide ores. A - U2 level of the Honpi ore body; B - Ad level of Uwabanhi ore body. Distribution of the overlying pyritic (A-A' type py-rich) and the underlying chalcopyrite-rich ores (B-B' type cp-rich) are indicated (After Slack, 1993).

Amorphous to microcrystalline iron sulphide precipitated on the sea floor. Extrusion of the overlying basalt protected the iron sulphide deposits from decomposition by seawater. Recrystallisation and growth of euhedral pyrite occurred within the iron sulphide and crystallisation of small amounts of pyrrhotite, chalcopyrite and sphalerite formed by subsequent hydrothermal activity. Sedimentation of clastic material, lava extrusion and sill intrusion

continued above the ore deposits. Replacement of pyrite by chalcopyrite took place due to the supply of hydrothermal fluids rich in copper and low in fS_2 to form high-grade copper ores.

2.1.2.3 Hitachi deposit

The conformable copper deposits of the Hitachi mine (Fig. 2) are situated in the southern end of the Abukuma andalusite-sillimanite metamorphic terrain (Fig. 7). This terrain forms a plateau which is occupied by regionally metamorphosed equivalents of Palaeozoic clastic and effusive rocks with granitic rocks intruded into them. The Hitachi area is underlain by the Hitachi series which comprises the Nishidouhira, Tamadare, Akazawa, Daiouin and Ayukawa Formations from the base upwards.

The Akazawa Formation is subdivided into the lower, middle and upper Akazawa members. The lower Akazawa member is composed of basic to intermediate schists and acidic schists, the middle member consists of massive quartzite (chert), pelitic schist and basic to intermediate schist, while the upper member consists of basic to intermediate schists interbedded with thin layers of acidic and pelitic schists. The Daiouin Formation is classified into two members of which the lower consists of basic to intermediate schists containing lenses of crystalline limestone with intercalations of pelitic schist whilst the upper member comprises tuffaceous psammitic to pelitic semi-schists. The coral from the limestone and fusulina from the pelitic rocks yielded a Middle to Lower Carboniferous age (Mariko and Kato, 1994). Basic, intermediate and acidic schists or semi-schists in the Akazawa and Daiouin Formations are derived from lava and/or pyroclastic rocks (Kase and Yamamoto, 1985).

The metamorphic grade declines south-eastward from amphibolite facies to greenschist facies through epidote-amphibole facies. Granodiorite has intruded in the northern area of the Hitachi mining district, causing the metamorphism of the pelitic schists to cordierite-andalusite-sillimanite-rich rocks and changing the sulphides in the ores to pyrrhotite (Kase and Yamamoto, 1985).

The Hitachi mine consists of more than ten deposits which are divided into two groups according to the stratigraphic horizon in which they occur (Kase and Yamamoto, 1985). The deposits of the Fujimi Group of deposits are interbedded between the lower acidic and upper basic schists of the Akazawa Formation. The deposits of the Fudotaki Group occur in a thick pile of basic schists of the Daiouin Formation.

The Fujimi deposits are located near the boundary between lower acidic schists such as siliceous biotite schists and quartz-muscovite schists and upper basic schists of the middle Akazawa member. Six ore bodies are arranged along the recumbent synclinal structure of the Akazawa Formation. The lithologies containing the sulphide layers of the Fudotaki deposits are a pile of basic schists, quartz-feldspar and/or quartz-muscovite schists, quartz-biotite schists, massive ores and again basic schists in ascending order, notably of the upper Akazawa member to the lower Daiouin member (Mariko and Kato, 1994). The quartz-biotite schists, which are located in the footwall of the deposits are thought to be metamorphosed equivalents of hydrothermally altered mudstones or tuffaceous mudstones, because of their fine compositional banding and aluminous composition (Kase and Yamamoto, 1985). These siliceous footwall schists are weakly mineralised in sulphides.

The massive ores of the Fujimi deposits comprise three types, sphalerite-predominant sphalerite-pyrite ore, pyrite-predominant ore with interstitial chalcopryrite and sphalerite, and chalcopryrite-predominant ore with subordinate amounts of pyrite and sphalerite. Pyrrhotite is often found in all three ore types, having formed from pyrite during contact metamorphism. The amount of pyrrhotite found is generally small, but sometimes pyrrhotite becomes the main iron sulphide instead of pyrite. Sphalerite-predominant ores are only found in the Fujimi deposits. These ores occur in the upper levels of the Fujimi deposit and appear to be concentrated in the hanging-wall of the ore body. The sphalerite-predominant ores always contain considerable volumes of barite. The amounts of gangue silicates are very small, but muscovite, chlorite and quartz are sometimes associated with barite.

No barite occurs in the sphalerite-poor ores, but cordierite, anthophyllite, magnesian-chlorite, biotite, muscovite and quartz are commonly seen. The gangue silicates of the sphalerite-poor ores of the Fujimi deposits are characteristically calcium-poor and magnesium-rich (Kase and Yamamoto, 1985).

The massive ores of the Fudotaki deposits comprise pyrite with interstitial chalcopyrite and sphalerite with the common gangue minerals being quartz, muscovite, biotite and chlorite, while magnesium-rich gangue silicates are absent. The nature of these ores is very similar to the Besshi-type ores in the Sanbagawa metamorphic belt (Kase and Yamamoto, 1985).

Mariko and Kato (1994) summarised the tectonic setting and genesis of the Hitachi mine as follows:

The Hitachi volcanic and sedimentary rocks, the protoliths to the metamorphic rocks, were formed in a back arc-arc trench system. The Hitachi accretionary complex was found not to contain any continental shelf-type limestone. The constituents of the complex are probably fragments of an island arc on the Farallon oceanic plate. The massive sulphide deposits were formed by a hydrothermal solution associated with acid and intermediate volcanic activity at the spreading axes in the back arc basin of the island arc. The authors are postulating that the island arc, including products of the back arc and fore-arc, collided with the Japanese arc. Considering the ore characteristics, tectonic settings and associated volcanic rocks, it is felt that the Hitachi massive sulphide deposits occupy rather an intermediate position between Besshi-type deposits and Noranda-type deposits, than Kuroko-type deposits.

Yamamoto *et al.* (1983) established the range of $\delta^{34}\text{S}$ values for the Hitachi deposit to vary from -1.6 to +6.0 per mil. These data which are consistently low, indicate that a substantial basaltic sulphide component with $\delta^{34}\text{S}$ values near 0 per mil, is incorporated in the ores of this Besshi-type massive sulphide deposit.

Table 1 Summary of Besshi-type deposits of Japan

Deposit	Size(Mt)	Cu%/Zn%/Ag%/Au(g/t)	Ore mineralogy	Host rock	Age
Besshi, Japan	30	2.5/0.3/7/0.2	Py,cpy,sph	Basic and pelitic schists, mag-garnet-quartz schists	Jurassic
Shirataki, Japan	5	1.3/0.3/4/0.1	Py,cpy,sph,brn	Basic-pelitic-garnet-quartz schists	Jurassic
Sazare, Japan	4.2	1.7/0.4/7/0.1	Py,cpy,sph,brn	Basic-pelitic-quartz schists	Jurassic
Iimori, Japan	4.9	1.2/n.a.	Py,cpy,sph,brn	Basic-pelitic-quartz schists	Jurassic-Cretaceous
Yanahara, Japan	25	1.2/n.a./58/0.6	Py,cpy,sph	Slates with acidic volcanics	Permian(?)
Makimine, Japan	5.2	1.8/n.a./6/0.4	Py,cpy,po,sph	Phyllites, schistose sandstone, diabase, gabbro and porphyrite	Jurassic-Cretaceous
Shimokawa, Japan	8	2.3/1.3/10/0.9	Py,marc,po,cpy	Massive basalts, diabase, slate, sandstone	Cretaceous
Hitachi, Japan	29	1.4/0.7/5/0.5	Py,cpy,sph	Siliceous biotite and quartz-muscovite schist, basic schist	Jurassic(?)

Note: n.a.- not available

2.2 Canada

2.2.1 Canadian Cordillera

The Canadian Cordillera is an 800 kilometre wide segment of the Circum-Pacific Orogenic Belt that forms the north-western margin of the North American continent. The Canadian part of the Cordillera is subdivided into two fundamental tectonostratigraphic regions, an eastern or "pericratonic" region, the Columbian Orogen, which comprises deformed miogeoclinal rocks interpreted as having been deposited at the margin of the craton, and a western or "suboceanic" region, called the Pacific Orogen consisting of rocks of eugeosynclinal character which may be allochthonous with respect to the craton (Thompson and Panteleyev, 1976).

The Columbian Orogen is an "alpine orogen" with a metamorphosed and intensely deformed and intruded core zone, the Omineca Crystalline Belt, flanked to the east by an unmetamorphosed foreland zone of décollement faulting and folding, called the Rocky Mountain Fold and Thrust Belt. Tectonic elements that form the Columbian Orogen are almost entirely a result of the last orogenic period of intense and protracted mountain building. Deposits in the Columbian Orogen are older, ranging from Early Proterozoic to Late Proterozoic, and occur in epiclastic and carbonate rocks, with often no apparent relationship to volcanism or volcanic rocks (Thompson and Panteleyev, 1976).

The Pacific Orogen is eugeosynclinal in character and comprises the Insular, Coast Crystalline and western part of the Intermontane tectono-physiographic zones. Deposition took place within island arc, ocean basin and successor basin environments, since the Late Palaeozoic along the westward (oceanic) side of the miogeoclinal sedimentary prism (Columbian Orogen) which was well developed by Late Palaeozoic time. The deposits of the Pacific Orogen are younger (mainly Mesozoic) and occur within or closely related to volcanic rocks (Fig. 10) (Thompson and Panteleyev, 1976).

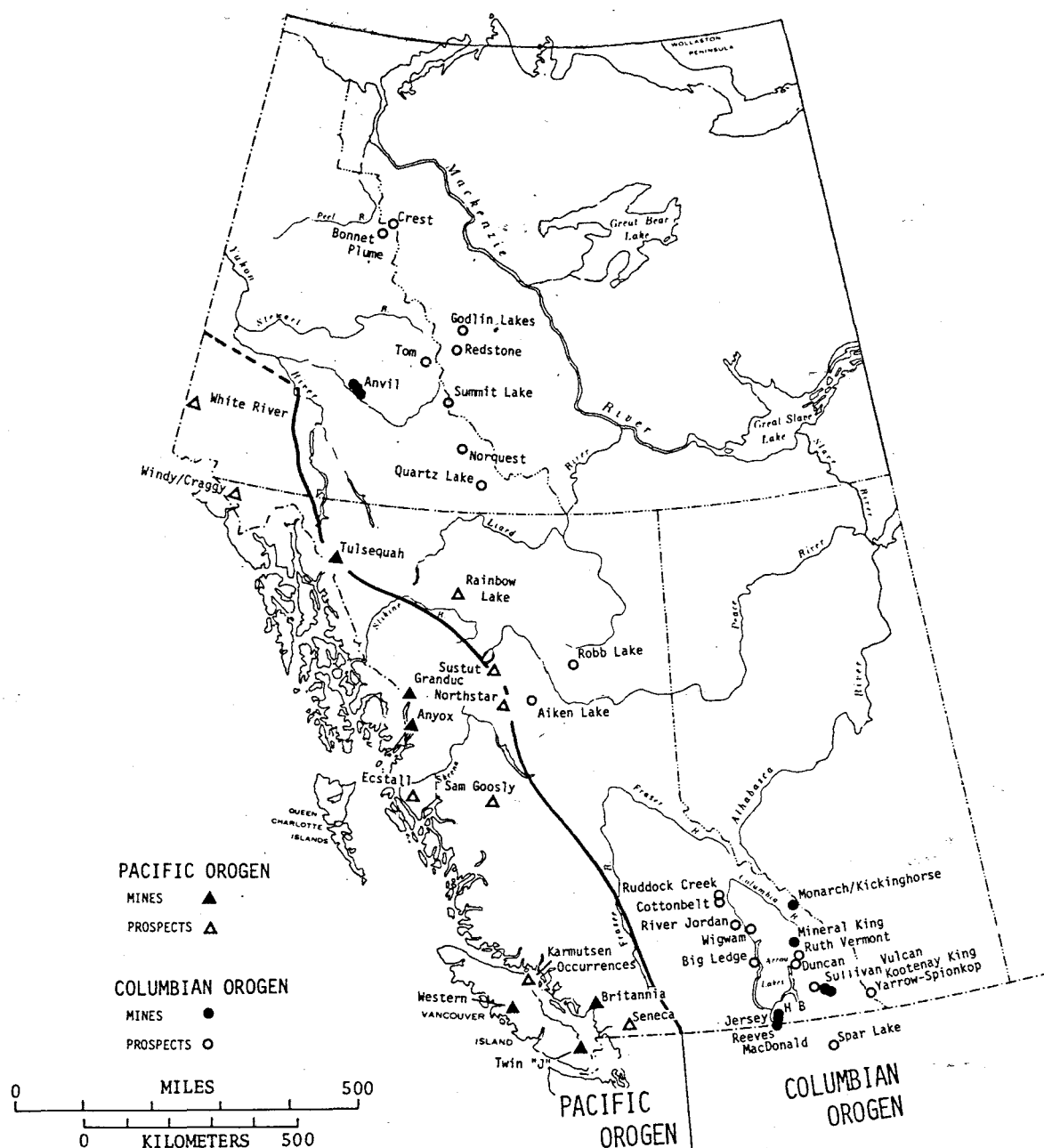


Fig. 10 Geographic location of stratabound metal deposits in the Canadian Cordillera (After Thompson and Panteleyev, 1976).

2.2.1.1 Pacific Orogen

2.2.1.1.1 Windy Craggy deposit

The Windy Craggy area is within the allochthonous Alexander terrain of the Insular tectonic belt in the extreme northwest of British Columbia (Fig. 11). This terrain includes a thick succession of complexly deformed Precambrian to Permian basinal and platformal carbonate and clastic rocks with a subordinate volcanic component.

The Windy Craggy deposit is located in the lower part of the Middle Tats Group which consists of interbedded graphitic and calcareous argillites, pillowed and massive mafic amygdaloidal to pillowed flows, tuff, agglomerate and limestone of Upper Triassic age (Peter and Scott, 1990). Mineralisation consists of at least two discrete massive sulphide lenses, the North and South Sulphide Bodies (NSB and SSB), each with an underlying stockwork/feeder zone with a possible third lens being investigated. The deposit remains open at depth and along strike (Peter and Scott, 1993). Pyrite was found to be the first sulphide to be deposited in primary structures such as framboids and colloform bands. Pyrrhotite, chalcopyrite, gold, electrum, marcasite, sphalerite, digenite and ankerite were deposited later. Gangue minerals such as stilpnomelane, chlorite, quartz and calcite occur as overgrowths on these minerals and rarely as intergrowths with the sulphides.

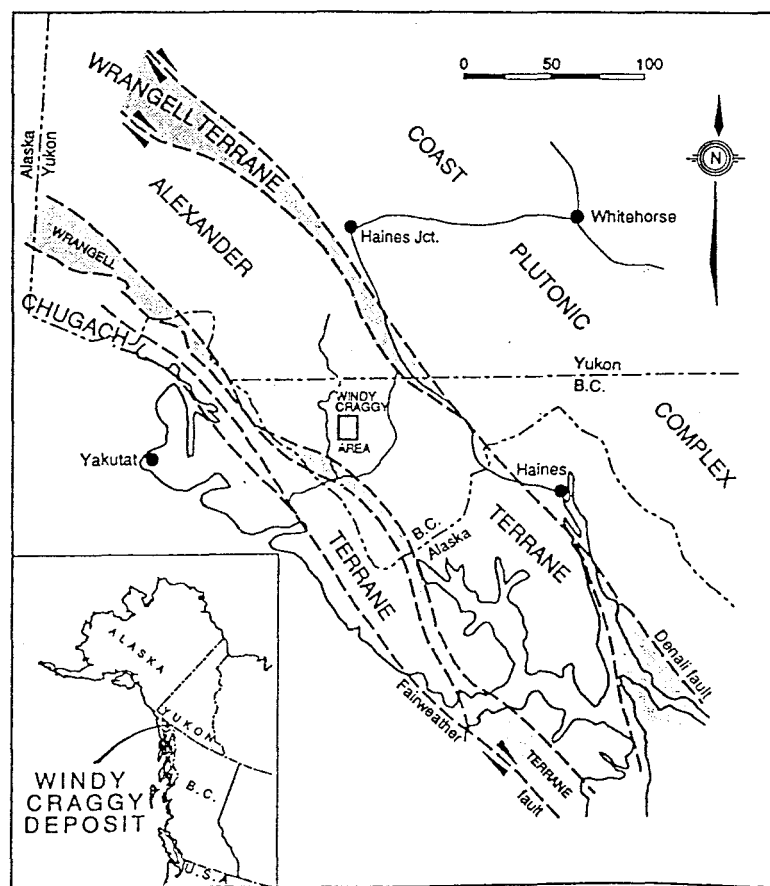


Fig. 11 Locality and tectonic setting of the Windy Craggy deposit, northwest British Columbia (After Peter and Scott, 1993).

The NSB is about 120-150 metres thick and 500 metres in diameter with a well developed stockwork/alteration zone within both the volcanic and the sedimentary rocks. The SSB is relatively more deformed and deformation/translation has altered the primary morphology of this lens. The NSB is mineralogically zoned from a pyrrhotite-rich core to a pyrite-rich outer/upper portion. Magnetite occurs at the transition from pyrrhotite to pyrite as a primary feature (Peter and Scott, 1990). Stockwork/stringer mineralisation crosscuts altered footwall basalt and argillite host rocks and extends at least 200 metres stratigraphically below the massive mineralisation in the NSB, while displaying distinct mineralogical zonation. Moderate to extreme bleaching and silicification of the host rocks in the stockwork/stringer zone grades into an adjacent zone of intense chloritisation with a distinct, bright, apple-green colour. This variety of chlorite grades laterally away from the stockwork zone to a darker olive-green coloured chlorite.

Although the host rocks have been altered to greenschist facies, their incompatible element and rare earth element (REE) chemistry relate the footwall and hanging-wall volcanic flows, geochemically to the footwall sills. Peter and Scott (1990) deduced from the geochemical characteristics and geological relationships that the Windy Craggy host basalts formed in a rift setting from an undepleted mantle source. According to these authors the footwall and hanging-wall flows and sills do not show any typical subduction zone geochemical signature, while the later dykes which cross-cut the mineralisation show an "arc" signature. Possible tectonic settings for this deposit include either a back-arc basin or a rifted continental margin/shelf setting. Peter and Scott (1990) found a linear trend extending from the field of Mid Oceanic Ridge Basalt to that of pelagic sediments in their $\delta^{207}\text{Pb}/\delta^{204}\text{Pb}$ vs $\delta^{206}\text{Pb}/\delta^{204}\text{Pb}$ isotope analyses. This is interpreted as the result of lead from a mantle source (footwall basalts) being mixed with lead from an upper crustal source (terrigenous sediments intercalated with the footwall basalts). It can therefore be said that the hydrothermal fluid from which the sulphides were precipitated had interacted with sediments before discharge onto the palaeo-ocean floor (Peter and Scott, 1990). Fluid inclusion studies indicated that the deposit

was probably formed under a water column of 1 950 metres or greater at a temperature of 350 degrees Celsius and 10-15 equivalent weight percentage NaCl (Peter and Scott, 1993). These findings are in accordance with many modern sea-floor sulphide deposits such as Guaymas Basin: 2 000 metres and Atlantis II Deep, 2 000-2 100 metres. The large size of massive sulphide lenses at Windy Craggy are thought to be indicative of a dense, bottom-hugging fluid, such as in the Atlantis II Deep, but the densities were found to be less dense than seawater, and therefore not bottom-hugging, but dispersing into the overlying water column similar to black smokers (Peter and Scott, 1993). The sulphur isotope compositions of the sulphides were found to indicate that the leached sulphides in the host rock basalts were the predominant source of sulphur, with a minor contribution derived from inorganically-reduced seawater sulphate. Carbon isotope analyses revealed that dissolved marine carbonate or seawater was the predominant source of carbon in the carbonates of the deposit. The hydrothermal fluid, calculated from carbon and oxygen isotope compositions, consisted of seawater that became enriched in ^{18}O upon reaction with sedimentary host rocks in the subsurface during hydrothermal circulation (Peter and Scott, 1993).

2.2.1.1.2 Granduc deposit

The Granduc deposit, near Stewart in north-western British Columbia, is hosted by a Lower Jurassic volcanic-sedimentary sequence at the eastern margin of the Coast Plutonic Complex (Fig. 10).

The host rocks form part of a shallow marine succession of thick andesitic pillow lavas, siltstone, crystal tuff, conglomerate, volcanic sandstone and some rhyolite and chert members which trend north and dip steeply to the west. The overlying sulphide-bearing sedimentary units comprise graphitic siltstones, gypsum-bearing limestones, conglomerate lenses, tuff, quartzite and chert horizons. The orebodies which occur in several zones are made up of overlapping and interconnected massive sulphide lenses that are localised predominantly within sedimentary rocks which in turn overlie a thick pillow-lava sequence.

The rocks in the area have been found to have undergone polyphase folding, faulting and, close to the mine area, cataclasis has resulted in the development of strongly laminated rocks in which a secondary foliation subparallels bedding. The finely laminated rocks are classified as cataclasites, mylonites and phyllonites with a quartz-biotite and biotite schist composition. The ore occurs as lenses of fine- to medium-grained intergrowths of pyrite, chalcopyrite and pyrrhotite with rare small lenses of sphalerite, galena and traces of arsenopyrite and cobaltite in a gangue of quartz, calcite, calc-silicates, magnetite and apatite. These sulphide lenses are separated by barren rocks or by weakly mineralised pyritic stockworks and low-grade stringer lodes. On small scale, individual ore bodies are highly irregular with feathery, diffuse boundaries and have been thickened in the crests of folds, attenuated along fold limbs with accompanying brecciated ore textures with evidence of sulphide flowage, crude banding, rotated ore and gangue fragments together with chalcopyrite-pyrrhotite segregation veins.

Thompson and Panteleyev (1976) favoured the volcanogenic theory which stated that massive sulphide lenses were deposited periodically in a shallow-water, reducing environment in a basin which is receiving marine sediments and andesitic volcanoclastic debris. Deposition of pillow lavas at the base of the mineralised succession was followed by intermittent deposition of sulphide-bearing volcanoclastic rocks and fine-grained sedimentary rocks. Ore bodies were modified by post-mineralisation deformation.

2.2.1.1.3 Anyox deposits

The Anyox deposits consist of twelve known ore bodies, formed over a distance of 10 kilometres in a roof pendant of volcanic and sedimentary rocks near the eastern margin of the Coast Plutonic Complex (Fig. 10) (Thompson and Panteleyev, 1976). The main deposit was the Hidden Creek mine where eight massive sulphide lenses were mined.

The host rocks are "greenstones" derived from a thick succession of pillow lavas, pillow breccias and possibly dykes with intercalated,

thinly bedded marine siltstone units of Middle Jurassic age. The rocks are intensely altered and andesitic in composition. The ore bodies are pipe- to sheet-like lenses of massive pyrite, pyrrhotite, chalcopyrite with minor sphalerite, galena, magnetite and arsenopyrite in a gangue of quartz, calcite, sericite and minor epidote and garnet. The sulphide bodies have been formed along the contact of pillow lavas and the overlying siltstones. The upper parts of the ore bodies consist of massive sulphides having sharp boundaries with the country rocks. The footwall zones are situated in pillow lavas and have more dispersed sulphide mineralisation with diffuse boundaries forming low-grade zones in silicified pyritic rocks (Thompson and Panteleyev, 1976). The ore bodies are commonly enveloped by thin zones of pyritic quartz-sericite schist. Polyphase deformation has resulted in complex fold patterns, but shearing is only developed locally.

Thompson and Panteleyev (1976) postulated that the stratiform massive sulphide lenses are volcanogenic deposits formed in sea-floor depressions during degassing of the volcanic pile at the end of a submarine volcanic cycle prior to marine sedimentation. These writers further stated that the Anyox deposits resemble Besshi-type ores in terms of ore mineralogy, host rock composition, lithology and environment of deposition.

2.2.1.2 Columbian Orogen

2.2.1.2.1 Goldstream deposit

The Goldstream deposit and several other small stratabound copper-zinc deposits occur in the geosynclinal rocks of the Kootenay Arc, between the Rocky Mountain Thrust and Fold Belt and the Shuswap Metamorphic Complex which all belong to the Omineca Crystalline Belt, south-eastern British Columbia (Fig. 12). The Goldstream deposit comprises a thin sheet of massive sulphides hosted by sericitic quartzites, calcareous, chloritic and graphitic schists, probably Early to Middle Palaeozoic in age. The massive sulphide layer is structurally overlain by a mangiferous, iron-rich chert.

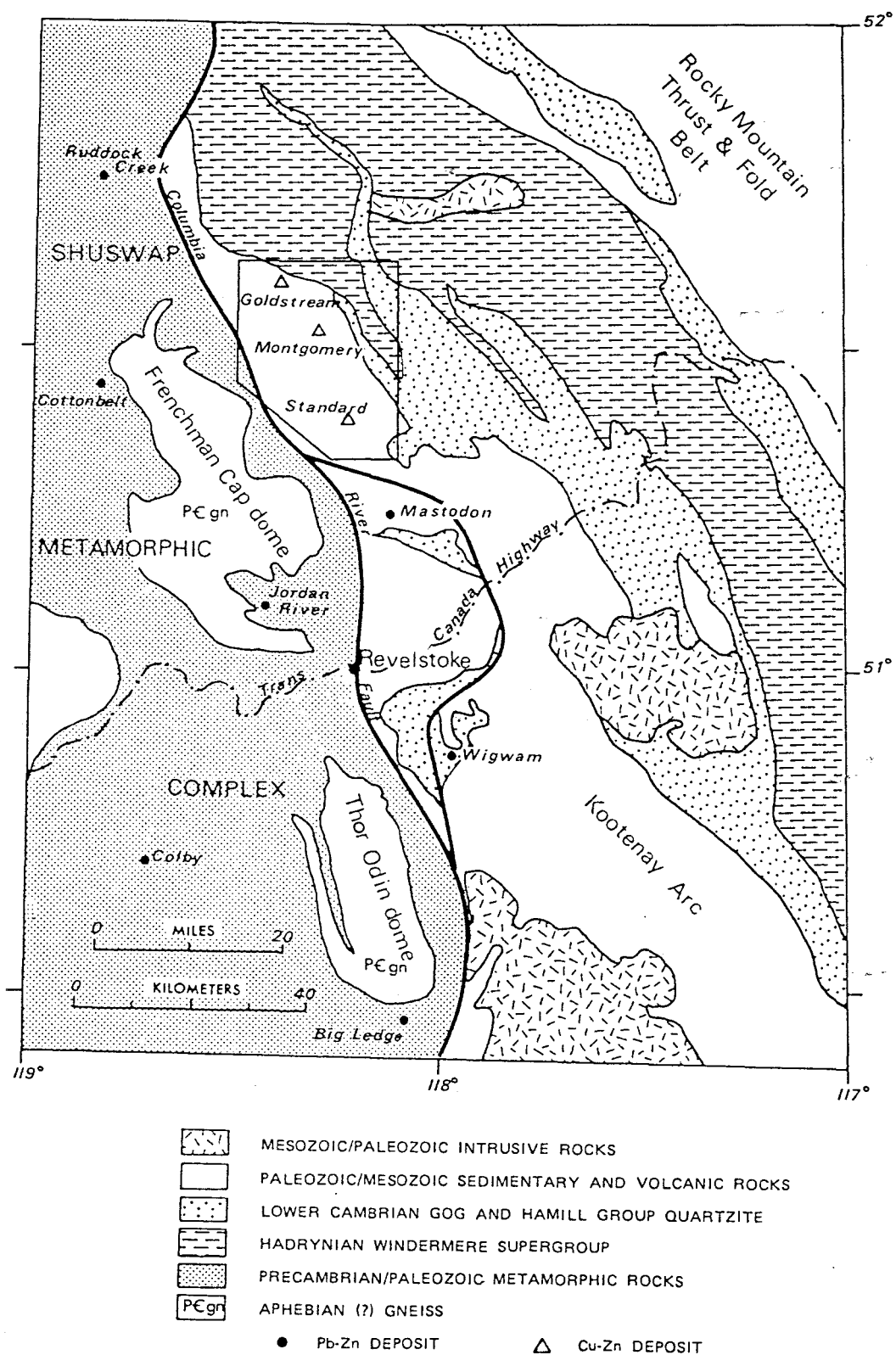


Fig. 12 Geological map of the northern Kootenay arc and the Shuswap Complex indicating the localities of copper-zinc and lead-zinc deposits. The Goldstream area is outlined (After Høy *et al.*, 1984).

Regional structures suggest that the deposit is overturned and that the chert layer is interpreted to have formed as a siliceous exhalite stratigraphically below the massive sulphide deposit (Høy *et al.*, 1984). Further features that tend to support structural inversion include a more gradational massive sulphide-hanging-wall contact than footwall contact, a thicker, mineralised hanging-wall section, extensive disseminated pyrrhotite in the hanging-wall rocks and pronounced dark chlorite alteration in the immediate hanging-wall rocks that might be the remnants of a sheared sulphide-rich footwall stringer zone.

The massive sulphide layer averages from 1 to 3 meters in thickness, stretches for at least 400 metres along strike and continues down plunge for at least 1 200 metres. The massive sulphide layer comprises closely intermixed pyrrhotite, chalcopyrite and sphalerite with numerous subrounded inclusions of quartz, phyllite and carbonate. The deposit is situated on the inverted limb of a large D2 isoclinal fold. The most conspicuous structures in the deposit are late folds and faults of D3 deformation, which deform both the layering and the D2 penetrative mineral foliation. The contacts of the sulphide zone with the hanging-wall and footwall vary from sharp to gradational, from smooth to highly contorted and brecciated; the structural footwall contact being more sharp than the structural hanging-wall contact. The distribution of copper, zinc and silver in the massive sulphide layer have a simple and regular disposition with increasing values towards the thicker, central parts of the layer. The deposit has a pronounced lateral metal zonation, with $Zn/(Zn + Cu)$ ratios increasing to the east. Høy *et al.* (1984) suggested that this zonation reflects either an original lateral deposit zonation or it is the result of extreme shearing of a vertically zoned massive sulphide lens that had a more copper-rich zone in the stratigraphically lower part of the lens and a relatively zinc-rich zone above. No consistent vertical zonation was observed by Høy *et al.* (1984).

Høy *et al.* (1984) equated the Goldstream deposit and several other small deposits in the vicinity to Besshi-type deposits on the grounds of their having similar basic volcanic host rocks, similar shapes, being bed-like or lenticular, similar mineral assemblages

with pyrrhotite (or pyrite), chalcopyrite and sphalerite and similar regional tectonostratigraphic setting in metamorphosed and highly deformed geosynclinal rocks.

2.2.2 Labrador Trough

2.2.2.1 Soucy deposit

The Labrador Trough is a well-preserved, thrust-imbricate Lower Proterozoic fold belt (2150-1850 Ma) that stretches for 850 kilometres from the Grenvillian metamorphic front to Ungava Bay. The trough has been divided into three litho-tectonic zones of which a central foreland fold-thrust belt, consisting of volcano - sedimentary sequences intruded by abundant gabbroic sills, hosts the Gerido Group. The Gerido Group consists of the basal Abner dolomite, the Baby Formation clastic-dominated sediments in the centre, with the Hellancourt tholeiitic basalts at the top. The Baby Formation consists of three members, the lower member consists of turbidites, conglomerates, quartzites, wackes, siltstone and mudstone of variable proportion and thickness. The middle Baby member is an iron-formation and consists of oxide-, carbonate-, silicate- and sulphide-facies sediments. The upper Baby member consists largely of rhythmically layered, thin-bedded distal turbidites and mudstones. The Soucy Number 1 massive sulphide deposit is located near the western margin of the foreland belt and is located, along with other massive sulphide deposits in the region, within iron-formations of the middle Baby member (Barrett *et al.*, 1988). Proterozoic Cu-Zn deposits do occur in geological settings that are similar to the Phanerozoic Besshi-type deposits of Japan, such as rift-related volcano-sedimentary sequences.

The Soucy Number 1 deposit lies on the eastern limb of the Lac Bourgault syncline (Fig. 13). The deposit outcrops as two sub-parallel north-northwest striking zones of gossan. During drilling, two corresponding massive sulphide zones were located, the A zone which is the main sulphide deposit is overturned, lensoid and extends along strike for at least 400 metres. The D zone, 200 metres to the west, is thinner, tabular, overturned and traceable for at least 500 meters along strike.

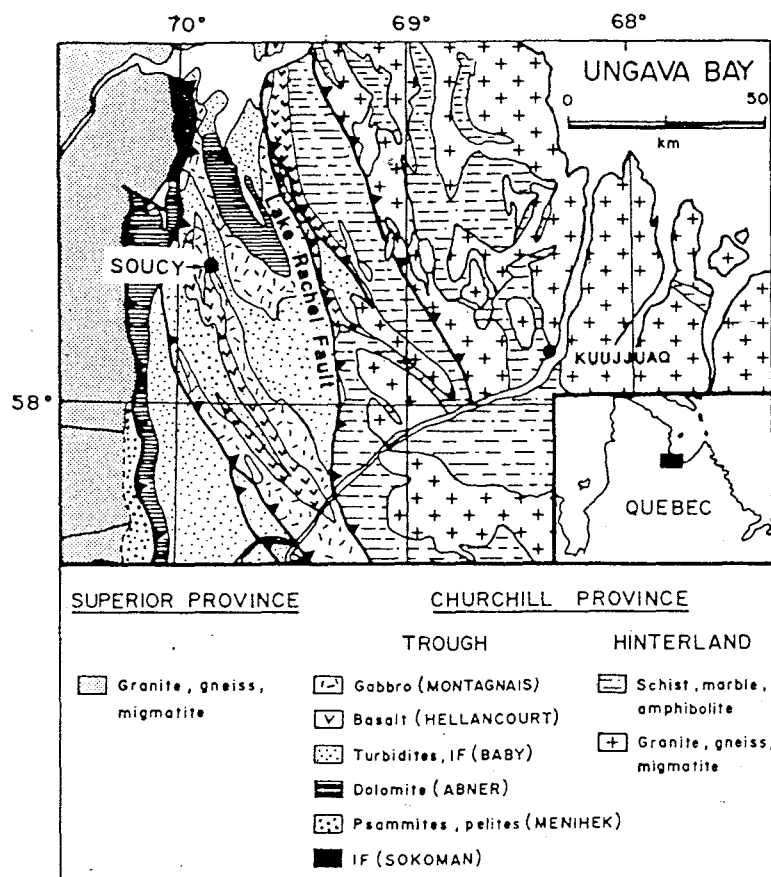


Fig. 13 Regional geology of the northern Labrador Trough and locality of the Soucy Number 1 massive sulphide deposit (After Barret *et al.*, 1988).

These features are interpreted to represent an original semi-continuous lens of sulphide that has been affected by two coaxial deformations.

The sediments underlying the ore lens are rhythmically layered, thin-bedded, light grey quartzo-feldspathic siltstones, dark grey mudstones and laminated argillites representing the unmineralised country rock. Graded bedding and load casts indicate overturned stratigraphy. The "feeder pipe" alteration zone occurs within these sediments. The lithologies comprising the pipe are dark green in colour and cut by irregular, millimetre-scale veins largely at high angles to bedding. The "alteration pipe" is strongly depleted in K, Na and Ba, while slightly depleted in Si and Al. Barrett *et al.* (1988) interpreted the meandering and curving shape of the veins as being part of a feeder system and indicative of the fact that the

sediments were unconsolidated at the time of veining. The footwall is altered over approximately 5 metres, directly below the ore lens, but lacks the veins of the "pipe". The contact with the underlying feeder zone is gradational over 0.5 metres. The contact between the footwall zone and the massive sulphide lens is sharp. The massive sulphide lens is made up of sub-spherical pyrite framboids in a matrix of fine-grained, recrystallised, granoblastic pyrrhotite, chalcopyrite, sphalerite and traces of galena. Pyrrhotite replaced the pyrite framboids and magnetite while chalcopyrite and sphalerite replaced the iron sulphides. The overlying laminated sulphides (<0.5 mm) consist of pyrite microspherules disseminated throughout a coarser grained, recrystallised pyrrhotite-iron silicate - quartz - muscovite matrix.

Barrett *et al.* (1988) compared the Soucy deposit with Besshi-type deposits with a few differences, such as having a footwall alteration zone, lower cobalt contents in the massive sulphides, a thick laminated sulphide blanket above the massive sulphides and a regional setting within iron-formations. Barrett *et al.* (1988) further postulated that the absence of these features from Besshi-type deposits could merely be a function of their higher metamorphic grade. These writers envisaged the Soucy sub-basin to be located on a foundered, deep-water continental margin block, bound by major faults. The presence of distal turbidite and pelitic interturbidite host sediments surrounding the ore lens, and the abundance of thick gabbroic sills, suggest a setting analogous to the early phase of spreading at the sediment-buried axis of the Gulf of California (Barrett *et al.*, 1988). One difference is that hydrothermal solutions at Soucy could have entered an anoxic bottom layer within a sub-basin, whereas in the Guaymas basin they were found to be discharging into circulating oceanic bottom waters. The hydrothermal circulation system responsible for the massive sulphide deposit is inferred to have been driven by a heat source related to rising asthenosphere during early rifting. More advanced stages of rifting eventually led to pervasive sill intrusion on the foundered margin with the extrusion of basalt on the ocean side of the Soucy site (Barrett *et al.*, 1988).

2.3 Namibia

2.3.1 Tectonic evolution

In southern Africa, late Precambrian orogeny resulted in the formation of a 3 000 kilometre chain of geosynclines following the western and southern coast of the subcontinent. This chain is preserved as the coastal branch of the Damara Province, the Gariep Province and the Saldanian Province (Tankard *et al.*, 1982). Development of the chain took place during the so-called Pan-African epoch ($\pm 1\ 000 - 450$ Ma).

The coastal branch of the Damara Province comprises a northern and southern part which is divided by an extensive oblique offshoot known as the intracontinental branch of the Damara Province. The intracontinental branch comprises, from north to south, of several different zones i.e. a Northern Platform (NP), the Northern Zone (NZ), Central Zone (CZ), Okahandja Lineament Zone (OLZ), Southern Zone (SZ), Southern Margin Zone (SMZ) and the Southern Foreland (SF) passing further southward into an undeformed Southern Platform (SP). The OLZ and SZ are together referred to as the Khomas Trough. The Khomas Trough in the inland branch of the Damara Orogen comprises thick multiply-deformed metagreywackes and metapelites of the Kuiseb Formation (Fig. 14). Also part of the Kuiseb Formation is the Matchless Member, a 350 kilometre-long and up to 3 kilometer-wide zone of metamorphosed mafic volcanic and intrusive rocks which are interposed with the metasediments (Kukla, 1992).

The Matchless Amphibolite Member is structurally emplaced within the sequence and contains tholeiitic, mid-ocean ridge-type metavolcanic rocks, including pillow lavas and breccias, as well as ultramafic lithologies and metagabbroic lenses. Pelagic graphite schists, pelitic schists and a marble unit are intercalated within the metabasic rocks. Original sedimentary structures, turbidite facies, the vertical facies distribution of progradational and retrogradational cycles, and the lateral extent of major sedimentary units indicate that psammitic parts of the sedimentary protoliths have been deposited as turbidites on an elongate submarine fan.

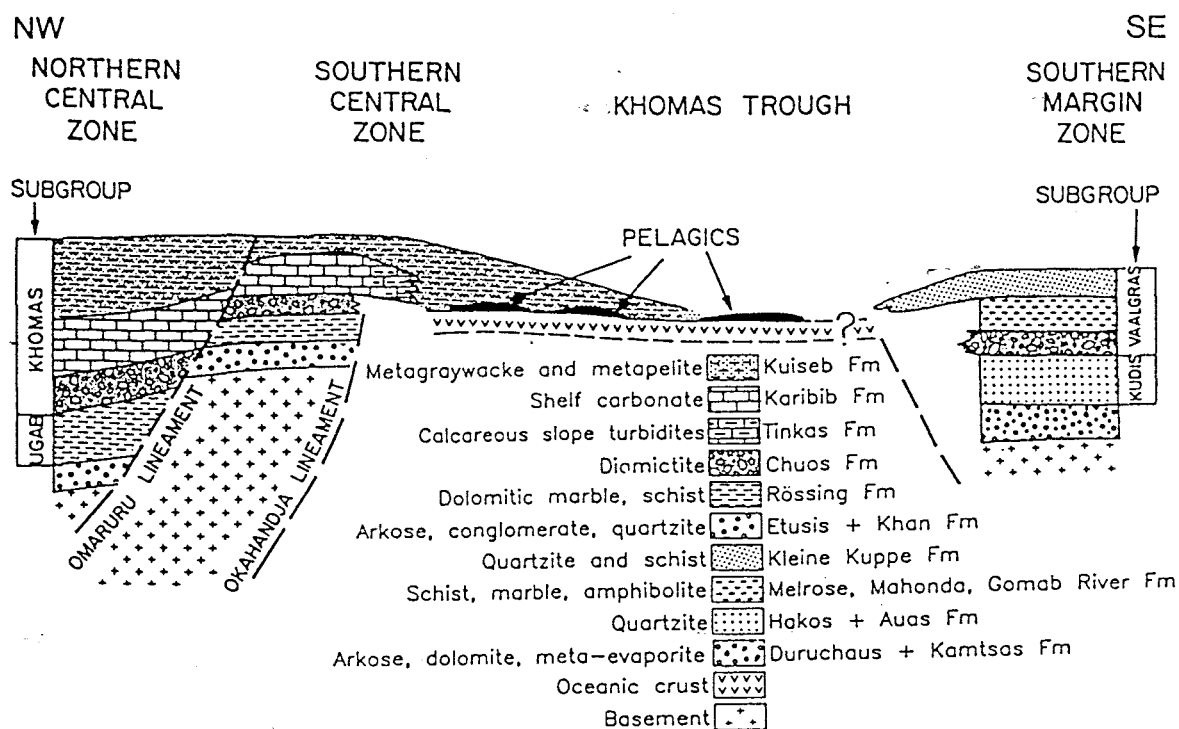


Fig. 14 Rock-relationship diagram of the Khomas Trough and adjacent areas in the Damara Orogen. Stratigraphic contacts are emphasised in their palinspastic context (After Kukla, 1992).

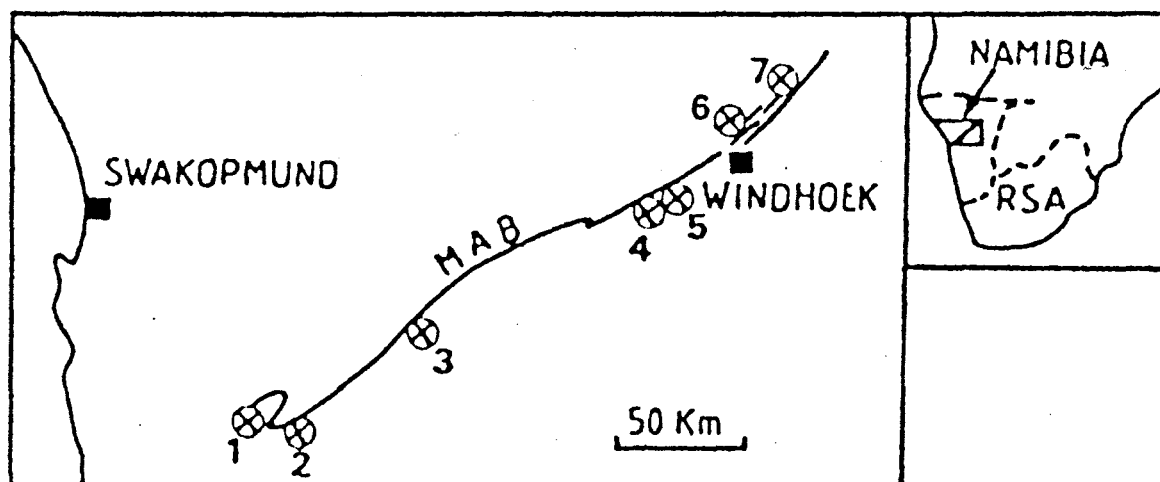


Fig. 15 Schematic representation of the Matchless Amphibolite Belt in the intracontinental branch of the Damara Orogen, Namibia. The location of associated massive sulphide deposits are indicated: 1-Hope, 2-Gorob, 3-Niedersachsen, 4-Matchless and Matchless West, 5-Kupferberg, 6-Otjihase, 7-Ongeama and Ongambo (After Pirajno, 1992).

Geochemical signatures suggest that an active continental margin source to the northeast of the present-day Khomas Trough supplied the vast amounts of clastic sediment. Radiometric dating of Kuiseb Formation rocks in the Khomas Trough has mostly produced ambiguous results which are probably due to isotopic re-equilibration during prolonged amphibolite facies regional metamorphism.

2.3.2 Gorob deposit

The Gorob deposit is located near the south-eastern extremity of the Matchless Amphibolite Member (Fig. 15). There are eight discrete sulphide lenses in the vicinity of Gorob which itself is separated by approximately 200 metres of quartz-biotite schists from the base of the Matchless Amphibolite Member (Fig. 16) (Killick, 1983).

The surface expression of the deposit is a gossanous stratiform magnetite quartzite, sericite quartzite or quartz-muscovite schist stained by secondary copper minerals such as chrysocolla and atacamite.

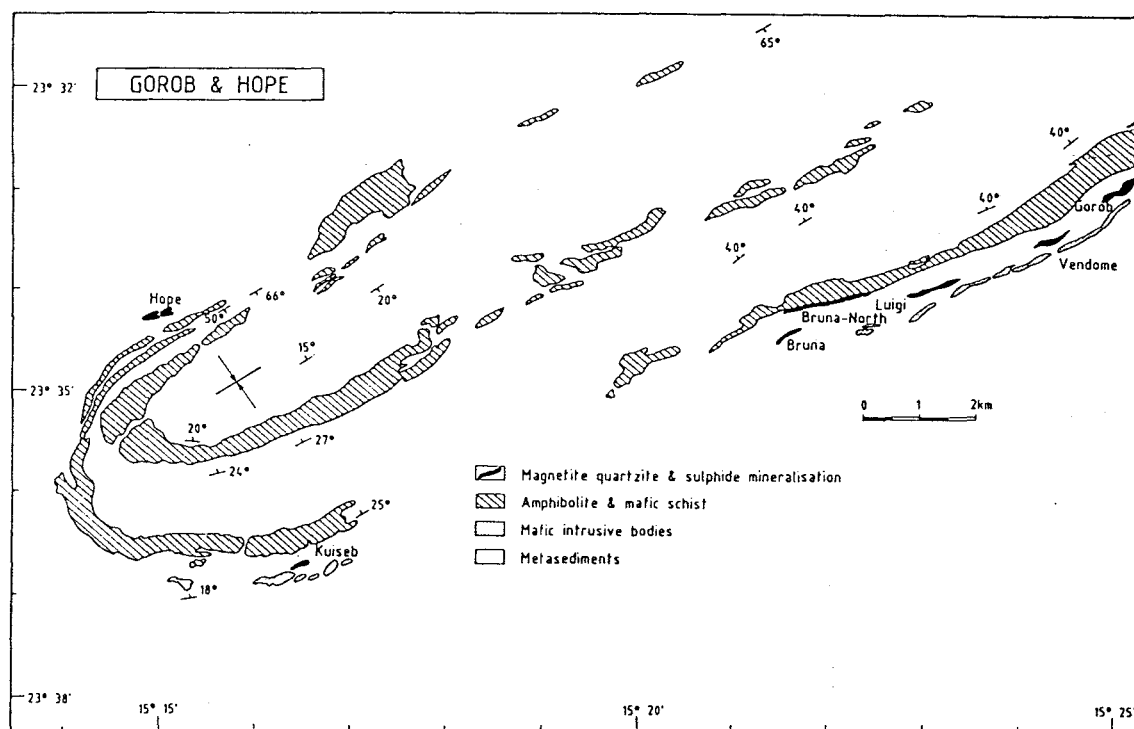


Fig. 16 Geological map of the Gorob and Hope area (After Breitskopf and Maiden, 1988).

Below the level of oxidation, pyrite is the dominant sulphide, followed by chalcopyrite, pyrrhotite, sphalerite, galena and rare molybdenite (Killick, 1983). A marked metal zonation is recognised, from a copper-rich hanging wall to a comparatively zinc-rich footwall.

Preussinger (1990) summarised the complex structural history of the Gorob area stating that the mineralised zone between the amphibolites and the graphitic schists is characterised by zones of "intraformational", S-shaped F2-folds and strongly foliated pelitic schists with intrafolial folds, in contrast to almost undeformed areas where even sedimentary structures are preserved. These alteration zones comprise Fe-Mg-rich chlorite schists (chloritic alteration), Al-rich staurolite-kyanite-sillimanite schists (argillaceous alteration) and micaceous felsic rocks (potassic alteration). During the ore-forming process, the amounts of Fe, Mn, Mg, Ba, Cu and Zn generally increased and Na, Ca, K, Rb, Sr and Ce decreased according to Häussinger *et al.* (1993). These authors drew up an alteration index that could be used as an in mine exploration tool when doing diamond drilling.

Discordant relationships are preserved at the margins of the Gorob alteration zone by the progressive obliteration of sedimentary structures. This indicates that these sediments must have been deposited before the hydrothermal activity evolved. The sea floor-seawater interface at the time of deposition is indicated by concordant layers of magnetite-quartzite, thus the alteration of the underlying metasediments must have occurred shortly after their deposition.

2.3.3 Matchless deposit

The Matchless mine is situated approximately 40 kilometres southwest of Windhoek, Namibia (Fig. 15). This deposit is associated with the Matchless Amphibolite Belt in the Kuiseb Formation of the Damara Sequence. The lithologies in the mine area consist of a thick (400 metres) horizon of amphibolite overlying quartz-mica schist with intercalated amphibolite bands. In addition there are layers of quartzitic rock, chlorite-rich schist, and amphibole-bearing

schist, which are particularly well developed in the immediate vicinity of the orebody. This major lithologic layering is considered to represent bedding. The sequence strikes generally northeast and dips at about 40 degrees to the northwest (Klemd *et al.*, 1987).

The Matchless copper deposit is contained within a zone of sulphide-bearing rocks about 500 metres south of the main amphibolite horizon. The deposit consists of three ore shoots of massive sulphide rock, plunging to the west at 30 degrees to 35 degrees, which are separated from one another by sulphide-bearing schist (Fig. 17). Each ore shoot is about 3 metres thick and 120 metres in strike length, and has been traced down plunge by drilling for 600 metres (Adamson and Teichmann, 1986).

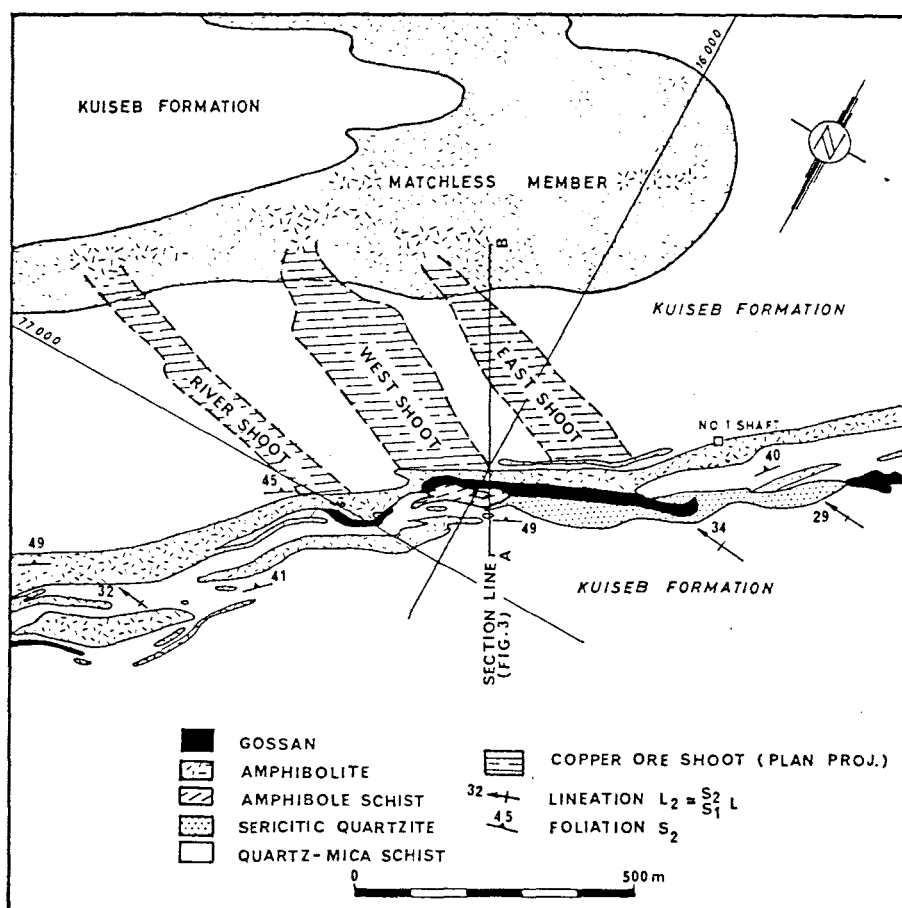


Fig. 17 Surface geological map of the Matchless deposit area (After Adamson and Teichmann, 1986).

The main sulphide minerals are pyrite and chalcopyrite, with smaller amounts of pyrrhotite and sphalerite, and accessory marcasite, galena, bornite, cubanite and molybdenite. Layering is variably developed. In massive pyrite rock, vague layering is due to subtle changes in the proportion of quartz, magnetite, chalcopyrite, and sphalerite.

Data collected by Klemd *et al.* (1987) indicate that the present Matchless mineralisation is the result of a complex geologic history involving initial sea-floor exhalative concentration from seawater-derived hydrothermal fluids to produce a stratigraphic zone of pyrite mineralisation within an envelope of hydrothermally altered rock; syntectonic reconcentration by movement of metamorphically derived fluid to produce ore shoots plunging down F2 fold hinges; and minor remobilisation related to late deformation.

In their study of the Matchless ore lenses, Klemd *et al.* (1989) have proposed that the chlorite-quartz $\pm$ ankerite schist, with intercalated bands of magnetite quartzite and lenticels of sulphides at the immediate structural hanging wall of the orebody, may represent an original footwall feeder zone, and therefore the sequence would be overturned. In their model, the magnetite quartzite is interpreted as the metamorphosed equivalent of a ferruginous cherts deposit around veins on the sea-floor. By contrast, the pyrite-quartz-muscovite (or sericite) rocks, which occur in the structural footwall are interpreted as exhalites forming as a result of waning hydrothermal activity, and they appear to decrease in number away from the mineralisation, that is, upward in the stratigraphic hanging wall. Geochemically, the assumed footwall feeder zone rocks show depletions in Na, Ca and Sr, and enrichment in Rb, Ba, Fe and S as well as Cu, Zn and Pb. However (Smalley 1990), do not support the model proposed by Klemd *et al.* (1989), but instead regard the magnetite quartzite bands as the top of the mineralised system, in which case the sequence would be the right way up. Maiden (1993) proposed that a syntectonic pre- or syn D2 hydrothermal event introduced additional copper superimposed onto a pre-existing pyritic stratiform volcanogenic deposit relatively richer in Zn and Pb. All minerals were subsequently recrystallised during the metamorphic peak.

2.3.4 Matchless West deposit

The Matchless West cupriferous pyrite deposit consists of 1 million drilled out tonnes of 2.3 per cent Cu. It lies 1 kilometre southwest of and on strike with the now closed Matchless Mine in the structural footwall of the Matchless Amphibolite Belt within the Khomas Trough - Kuiseb Formation metasediments of the upper Proterozoic Damara Orogen, Namibia (Smalley, 1990).

The deposit is arranged in a narrow, shoot-like, westward plunging linear zone and is associated with lithotypes of volcanothermal origin (Fig. 18). These lithotypes are overlain by a 40 - 50 metres thick coarse-grained massive amphibolite (metagabbro). Three distinct zones were recognised during detailed surface mapping (Smalley, 1990), i.e. the ore zone is composed of massive gossan (after pyrite) with lenses and pods of clean fine-grained magnetite quartzite and gossanous magnetite quartzite forming a thin (less than 1 metre thick), foliation parallel horizon 80 metres long. The footwall zone lithotypes are zoned: an outer pyrite quartz sericite schist and an inner sericite pyrite quartzite.

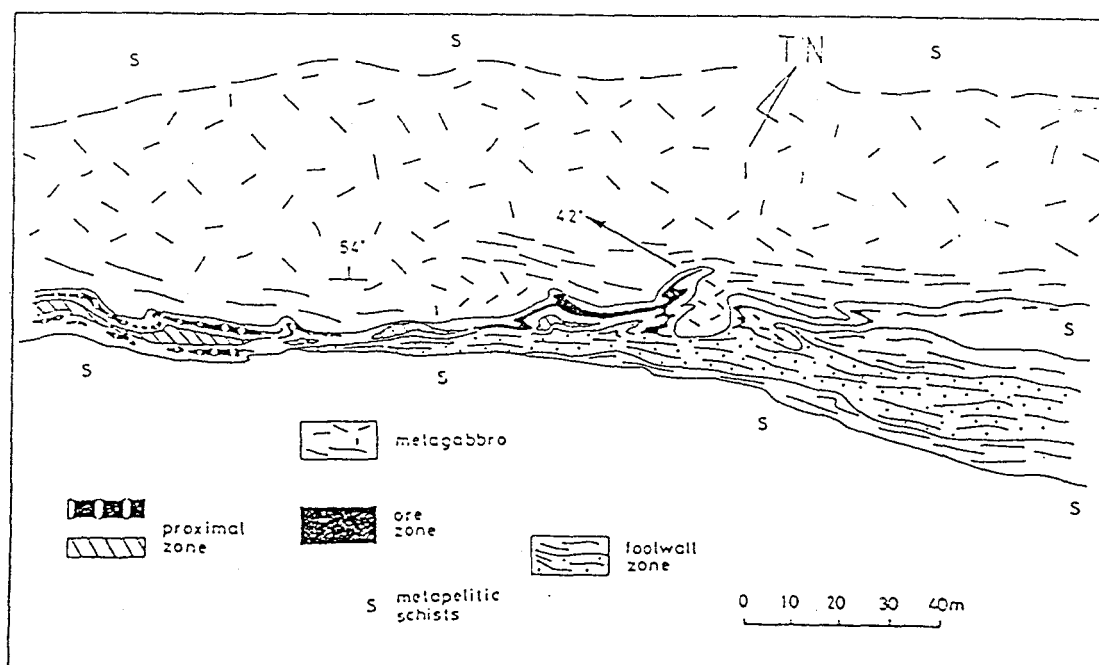


Fig. 18 Surface geological map of the Matchless West deposit, identifying the main lithological rock types (After Smalley, 1990).

Pyrite occurs as stringers and thin bands within the quartzite. This footwall assemblage is orientated at an angle of 12 degrees to the ore zone and intersects the latter at the thickest, best mineralised, central part. The proximal zone overlaps structurally below the ore zone and extends westwards on strike for 120 metres.

This zone consists of an inner assemblage of coarse-grained ultramafic lithotypes (talc schists and variable talc carbonate chlorite amphibole schists) bounded by a lower and an upper, clean (sulphide poor), fine-grained magnetite quartzite. Copper mineralisation (after chalcopyrite) is apparent on surface as secondary chrysocolla and malachite in the gossanous lithotypes of the ore zone and gossanous areas within the ultramafic assemblages of the proximal zone. The economically important mineralisation is located in the central part of the ore zone where refolding has caused duplication and structural thickening.

In summary it can be said that the footwall zone represents footwall alteration of the host quartz plagioclase biotite schists (Kuiseb Formation metasediments) which have been rotated by sinistral second phase deformation into subparallelism with the ore zone. This implies that this deposit is the right way up. Secondly, the ore zone is the result of metalliferous chemical precipitation at the seawater/sediment interface immediately above the feeder zone. Highest grade copper mineralisation is coincident with the largest amount of pyrite at the intersection of this ore zone and the divergent footwall zone indicating the approximate position of the vent. Thirdly, the various ultramafic lithotypes and the sulphide poor magnetite quartzites of the proximal zone are typical products, after metamorphism of primary chemical precipitates resulting from the reaction of seawater with hot hydrothermal fluids. Fourthly, the Mg, Ca, Si and OH-rich assemblages suggest lower temperature of precipitation and less interaction with hydrothermal fluids than the ore zone and may be function of simple seawater heating, resulting from the thermal venting system, rather than any direct fluid contribution from it.

2.3.5 Otjihase deposit

The Otjihase deposit is located 20 kilometres north of Windhoek, Namibia (Fig. 15). Ore reserves are approximately 16 million tonnes of 2 percent Cu, 0.3 - 0.6 percent Zn, 9 - 10 ppm Ag and 0.35 - 0.5 ppm Au. The mineralisation is hosted in quartz-biotite schists of the Kuiseb Formation, but in the vicinity of the orebodies, in the host rocks are also included garnetiferous quartz-biotite schist and chlorite schist, talc-carbonate schist and actinolite schist. The lithologies strike northeast, dipping northwest at about 15 degrees north and three northeast trending faults occur in the area and have divided the deposit into different fault-bounded blocks. The downthrow of these faults to the west is between 60 and 120 metres (Thomson, 1989). Surface expressions of the mineralisation are characterised by four distinctive gossanous magnetite quartzite outcrops which can be traced for a distance of at least 1.5 kilometres (Fig. 19). These outcrops are up to 12 metres thick and have a downdip extent of about 4 kilometres. The gossan material is red-brown in colour and consists of leached limonite, goethite and siliceous material in which Cu values are only about 0.1 per cent. Oxidation of the sulphide zones extends down to a depth of about 50 metres in places, where there is an abrupt change to sulphide without a zone of secondary enrichment. The hypogene mineralogy comprises pyrite and chalcopyrite with erratic sphalerite and very minor pyrrhotite. The supergene mineralogy consists of malachite, chrysocolla and chalcocite. The mineralisation is situated some 200 - 300 metres structurally, and possibly also stratigraphically, above the Matchless Amphibolite Member. The main orebody is narrow, elongate and lenticular (Fig. 20).

An important feature both of the Otjihase deposit (and of Gorob) is that the plunge of the long axis of the ore shoots is defined by a prominent D2 mineral lineation, which parallels the direction of minor fold axes seen on the surface in the footwall magnetite quartzite. The hanging wall contact is usually transitional, whereas the footwall contact is sharp.

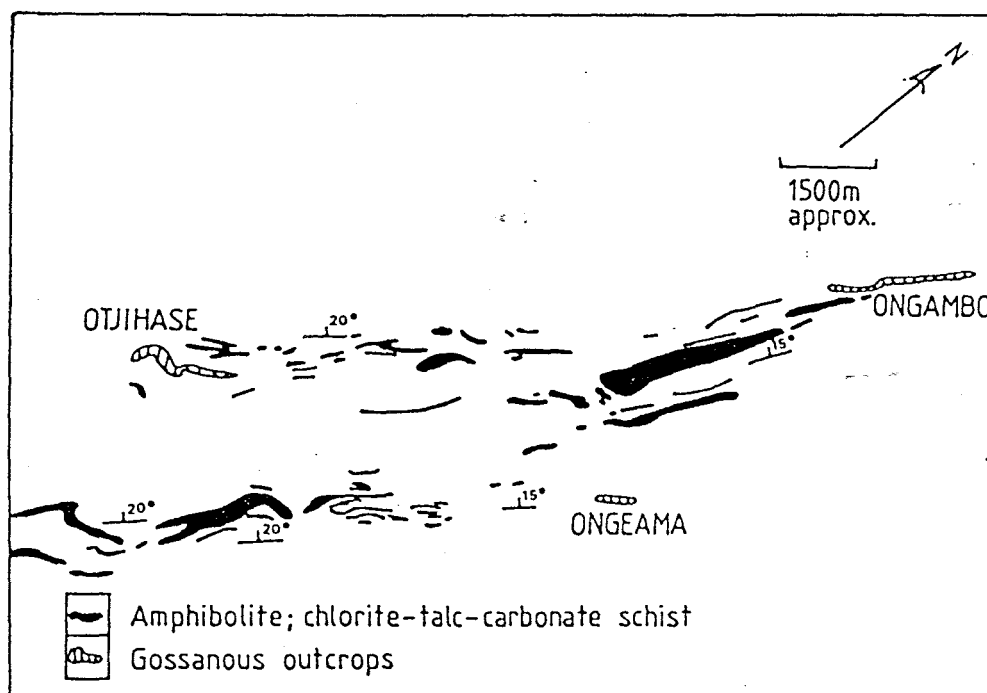


Fig. 19 Distribution of the Matchless Amphibolite and associated lithologies in the Otjihase, Ongeama and Ongambo area (After Pirajno, 1992).

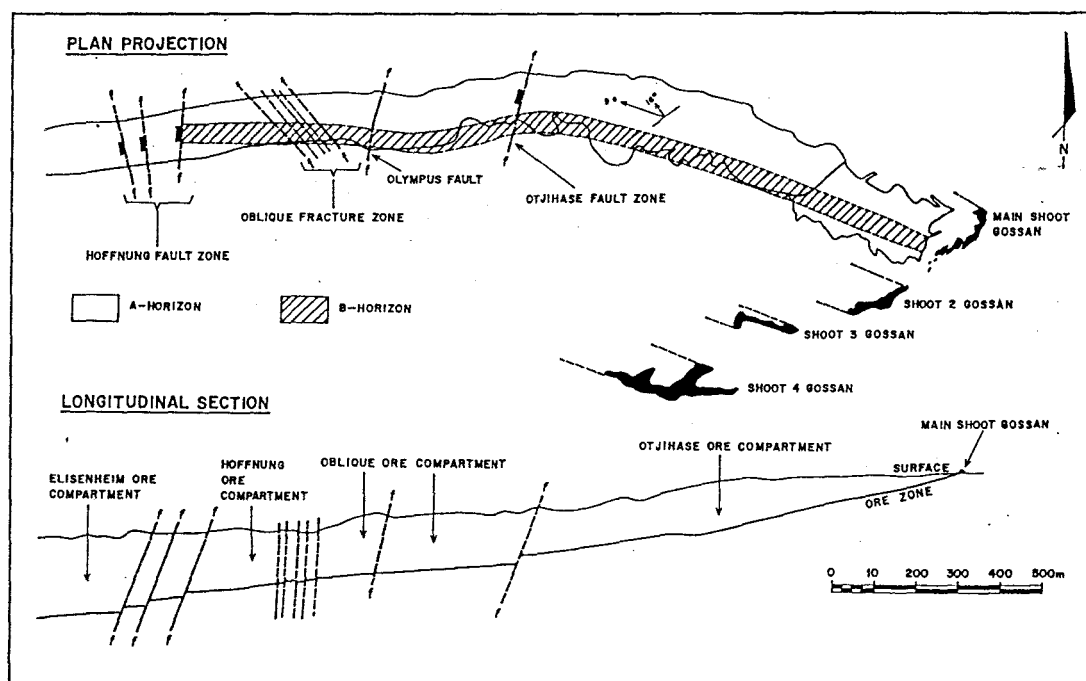


Fig. 20 Plan projection and longitudinal section of the main shoot, Otjihase mine, Namibia (After Thomson, 1989).

The orebody averages 2.5 metres in width, but may locally thicken to 7 metres. Goldberg (1976) divided the ores into massive, disseminated and stringer. The massive ore is composed of more than 80 per cent sulphides and 20 per cent magnetite quartzite gangue. Sulphides are fine- to coarse-grained and form massive bands composed of pyrite and chalcopyrite or alternating bands of pyrite and chalcopyrite. Disseminated ore contains up to 50 per cent sulphides in a gangue of sericitised chlorite-biotite schist and magnetite quartzite. Stringer sulphides form narrow veins of either pyrite or chalcopyrite.

2.3.6 Ongambo and Ongeama deposits

The Ongambo and Ongeama copper deposits are situated about 20 and 30 kilometres northeast of Windhoek, respectively (Fig. 15). Mineralisation at Ongeama is exposed over a strike distance of 900 metres and consists of two discrete, sinuous, gossanous bands which are conformable with the schists (Fig. 19). Mineralisation at Ongambo outcrops as a gossanous, magnetite quartzite for a total strike length of 4 900 metres. The width of the mineralised zone is estimated to be of the order of between 1 to 2 metres.

These two orebodies are both deformed and metamorphosed to low to medium grade greenschist-amphibolite facies. The sulphide bodies at Ongambo and Ongeama are elongate in shape, similar to Matchless and Otjihase, suggesting structural control. Moroni (1990) noted that the mineralisation normally consists of variably developed massive sulphide portions, either quartz-talc- or amphibolite bearing and overlying the zone of mineralisation is an extensive horizon of sulphide- and barite-bearing exhalative. Some orebodies have a metamorphosed and deformed alteration pipe developed in association with the mineralisation. Moroni (1990) also recognised the formation of quartz-muscovite and chlorite alteration envelopes and the subtle mineralogical changes in the Ongambo wallrocks, suggesting metasomatic hydrothermal alteration accompanied by extensive redistribution, leaching and introduction of elements from outside.

2.4 South Africa

2.4.1 Areachap deposit

The Areachap copper-zinc massive sulphide deposit occurs in the Proterozoic metamorphosed volcano-sedimentary sequence of the Copperton-Areachap Formation, Gordonia District, South Africa.

The magnetic pattern of the sequence is characteristic of a dominantly volcanic sequence with thin lava flows of different composition, alternating with beds of tuff. The amphibole gneisses and amphibolites are probably metamorphosed intermediate lavas. Texturally, quartz-feldspar rocks, interbedded with the amphibole gneisses and amphiboles, closely resemble recrystallised pyroclastics. These beds are extremely irregular and cannot be correlated from borehole to borehole (Voet and King, 1986). The biotite-garnet schists may represent pelitic sediments, but similar rocks elsewhere in the Northern Cape area are believed to be metamorphosed acid tuffs. These schists are the most easily recognisable and continuous beds in the area and their characteristic magnetic signatures can be traced over considerable distances. Petrological examination of the host schist and quartzite indicates that they probably represent shale, impure sandstone and arkose. No clear wall rock alteration was observed by Voet and King (1986). The ubiquitous chlorite development in the host rock can probably be ascribed to retrograde metamorphism. Anthophyllite was identified in the immediate wall rock of the massive sulphide body.

The lithologies at Areachap strike north-westerly direction, dipping steeply to the southwest at 85 degrees. There is evidence that only the western limb of a syncline is present.

The sulphide body is hosted by a biotite-garnet schist which is enclosed mainly by amphibole gneiss and amphibolite. The massive sulphide is coarse grained and probably completely recrystallised. The main ore minerals are pyrite and pyrrhotite, with chalcopyrite, sphalerite and iron oxides being present as important additional

constituents. Pyrite and pyrrhotite generally form a matrix of interlocking crystals. Rarely, chalcopyrite may form a matrix, but otherwise chalcopyrite is interstitial to pyrite and pyrrhotite. Sphalerite is present interstitially, enclosed in the other sulphides, or associated with the silica gangue. A prominent gossan is developed on surface. The gossan consists mainly of silicified, massive hematite with nests of goethite, veins and films of copper carbonates and oxides.

Voet and King (1986) proposed the genesis of the sulphide mineralisation as follows:

Accumulation of sediments and volcanic deposits commenced in a basin of which the present Areachap area was possibly the central part. Subsequent outpouring of intermediate lavas occurred which are indicated by the amphibole gneisses and amphibolite units. Flows were generally limited in size and the outpouring was frequently interrupted by more explosive events, causing the deposition of rhyolite tuffs, represented by the quartz-feldspar intercalations. During the deposition of the host rock clays, sandstones and arkoses, little volcanism was experienced. The concomitant basin subsidence accelerated, causing fracturing and faulting of the floor, allowing metal-rich brines, derived from a deep-seated magma chamber, to enter and spread out as fumaroles within the still unconsolidated sediments. It was concluded on a field trip to the Areachap deposit that it has many similarities to a Besshi-type deposit.

2.4.2 Bokspits deposit

The Bokspits copper-iron sulphide deposit occurs in the Proterozoic amphibolite belt of the Areachap Group, along the eastern margin of the Namaqua mobile belt, northwestern Cape, South Africa. The metavolcanic belt, dated at 1 200 - 1 300 million years old, separates the Archaean Kaapvaal Craton in the east from the Korannaland high-grade metamorphic gneiss terrain in the west (Geringer *et al.*, 1987).

The main rock types hosting the mineralisation are massive amphibolite and pyroxene amphibolite, feldspathic amphibole gneiss, amphibole schist, biotite-quartz-feldspar gneiss with mica schist and amphibolite. The lithostratigraphic succession is highly complicated by intensive structural deformation and late-stage shearing as well as the intrusion of granite and diorite bodies. Copper-iron mineralisation occurs as thin, disseminated, strata-bound layers within a metavolcanic succession, called the Jannelspan Formation, between a basal sequence of metapsammitic rocks and a sequence of metapelites at the top. Along the western limb of the Kraalkop antiform the mineralisation is associated with a highly ferruginous chert and muscovite schist near the top of the massive amphibolite. Along the eastern limb of the Kraalkop structure the mineralisation occurs in an amphibolite schist within a feldspathic amphibole gneiss. Malachite staining indicates the presence of sulphide mineralisation in outcrop. The mineralised zone is highly chloritised and marked by intensive calcite and quartz vein fillings over an interval of approximately 40 metres (Geringer *et al.*, 1987).

The most important sulphides present in the ore are pyrite, chalcopyrite, sphalerite and small quantities of pyrrhotite. The oxides present are magnetite, hematite and ilmenite. The sulphides occur predominantly as disseminated grains and thin layers of massive sulphides parallel to the foliation of the host rock. The high-grade, upper-amphibolite-facies metamorphism which the host rocks have suffered, can be seen as fractured and disrupted pyrite.

Geringer *et al.* (1987) reported that the mid-oceanic-ridge basalt normalised trace element distribution patterns for the amphibolites correspond with island-arc volcanics, being distinctly different from that of oceanic tholeiite or within-plate continental tholeiites. The rare earth element chemistry indicates that the amphibolite host rock is moderately enriched in light rare earth elements with positive Eu anomalies and lower than normal mid-oceanic-ridge basalt heavy rare earth element concentrations. These data support the previous findings of the host rocks being calc-alkaline lavas of island-arc affinity.

According to Geringer *et al.* (1987), features such as the association of predominantly pyrite and chalcopyrite ores with basaltic rocks of low-K tholeiite composition, the absence of galena and the rare earth element distributions, all point to a possible correlation to Besshi-type deposits. These authors concluded that according to the model proposed by Hutchinson (1980), the Bokspits deposit resembles a Besshi-type deposit which formed in a front-arc environment associated with low-K arc tholeiites along the eastern margin of the Namaqua mobile belt.

2.4.3 Prieska deposit

The Prieska ore body occurs within close proximity to the north-eastern boundary of the strongly deformed Namaqualand mobile belt, northwestern Cape, South Africa.

Rock types in the area can be grouped into two varieties, banded gneiss and homogenous gneiss. The homogenous gneiss is deeply weathered and constitutes a chlorite-rich, quartz-sericite schist which on fresh exposure contains amphibole as the main mafic constituent. The banded gneiss varies from feldspathic quartzite to feldspathic amphibolite. Numerous cross-cutting pegmatite veins are related to later fracturing and is thought not to be connected to an intrusive stock (Wagener and Van Schalkwyk, 1986).

The sulphide-bearing strata is made up of an outer, quartz-rich zone, containing trace amounts of copper, zinc and lead in sharp contact with an inner, silica-poor zone hosting the economic sulphides (sphalerite, chalcopyrite and galena) in a matrix of pyrite, carbonates, sulphates and calc-silicate minerals. Pyrite is the dominant sulphide and on average amounts to 45 percent by weight. Sphalerite favours localities rich in carbonate and barite where it sometimes reach a contents of 40 percent. Traces of molybdenite, silver and gold are present.

The tabular ore body has a strike length of more than 2 000 metres and a depth extent of at least 1 000 metres, plunging at 45 degrees.

Thickness of the ore body varies from less than half a metre to 30 metres, the average being 7 metres. The dip is generally steeper than 70 degrees, dipping at 30 degrees at the northern extremity and steepening to a vertical dip 600 metres further south-east. The ore body is made up of numerous lenses of varied composition. These lenses appear to be randomly distributed and the only obvious trends are the increased abundance of pyrrhotite near the structural hanging-wall and the distribution of a tourmaline-rich ore type with pale sphalerite along the lateral fringes of the ore body.

Wagener and Van Schalkwyk (1986) summarised the genesis of the deposits in the following way:

Sedimentary textures in the ore and hanging-wall gneiss have survived granulite grade metamorphism, indicating that the ore was precipitated as a chemical sediment. Textures previously considered to be of volcanogenic origin were reinterpreted by these authors as a metamorphic texture. No explosive volcanic activity is evident and the deposit can be regarded as a distal deposit from the volcanic vent area. A progressive variation in depositional environment, commencing with clastic sedimentation and culminating with the precipitation of the sulphides is proposed by the authors. The variety and contrast of mineral associations in the chemical sediments of the ore zone are interpreted as having formed in a sea-floor environment associated with an interface between anoxic brine and normal water in a stratified basin.

Table 2 Summary of Besshi-type deposits in Canada, Namibia and South Africa

Deposit	Size(Mt)	Cu%/Zn%/Ag%/Au(g/t)	Ore mineralogy	Host rock	Age
Windy Craggy, Canada	297	1.4/0.3/4/0.2	Py,po,cpy,gold	Graphitic, calc argillites, pillowed mafics,tuff,agglom	Triassic
Granduc, Canada	39	1.7/n.a./7/0.1	Py,cpy,po,sph	Andesitic pillow lavas, siltstone, tuff, volcanic sandstone	Jurassic(?)
Anyox, Canada	23.9	1.5/n.a./9/0.01	Py,po,cpy,sph	Pillow lavas, pillow brecc, thinly bedded marine siltstones	Middle Jurassic
Goldstream, Canada	3.2	4.5/3.1/20/n.a.	Po,cpy,sph	Sericitic quartzites, calc, chloritic and graphitic schists	Late Proterozoic
Soucy, Canada	5.4	1.5/1.8/14/1.6	Py,po,cpy,sph	Oxide-, carbonate-, silicate-, sulphide facies sediment	Early Proterozoic
Gorob, Namibia	1.1	2.7/n.a	Py,cpy,po,sph	Magnetite, sericite, quartz-muscovite schist	Late Proterozoic
Matchless, Namibia	2.7	2.3/n.a	Py,cpy,po,sph	Quartz-mica schist, chloritic schist, amphibolite-bearing schist	Late Proterozoic
Matchless West, Namibia	1	2.3/n.a	Py,cpy	Magnetite quartzite, quartz-sericite, sericitic quartzite	Late Proterozoic
Otjihase, Namibia	16	2/5/10/0.4	Py,cpy,sph,po	Quartz-boitite schist, chlorite schist, talc-carb schist, actinolite schist	Late Proterozoic
Ongombo & Ongeama, Namibia	n.a.	n.a.	Po,py,cpy	Quartz-talc- and amphibolitic schist	Late Proterozoic
Areachap, South Africa	8.9	0.4/2.24/4.6/0.07	Py,po,cpy,sph	Biotite-garnet schist, amphibole gneiss,amphibolite	Proterozoic
Bokspits, South Africa	n.a.	n.a.	Py,cpy,sph,po	Massive amphibolite, amphibole gneiss	Proterozoic
Prieska, South Africa	47	1.7/3.8/n.a.	Py,sph,cpy,po	Pyritic metasediments, laminated gneiss	Proterozoic

Note: n.a.- not available

Table 3 Summary of Besshi-type deposits in the United States, Norway and Australia

Deposit	Size(Mt)	Cu%/Zn%/Ag%/Au(g/t)	Ore mineralogy	Host rock	Age	Reference
Elizabeth, USA	2.9	1.8/0.5/40/0.3	Po,cpy,py,sph	Phlogo-tremolite-carb rock	Devonian(?)	McKinstry & Mikkola (1954)
Ducktown, USA	163	1/0.9/3/0.3	Po,py,cpy,sph	Metagraywacke, graywacke congl, mica schist, chlorite-garnet schist	Late Proterozoic	Magee (1968)
Gossan Lead, USA	40	0.5/1.5/8/<0.1	Po,cpy,sph,py	Metasediments, amphibolite, actinolite-chlorite schist	Late Proterozoic	Henry <i>et al.</i> (1979) Gair & Slack (1984)
Ore Knob, USA	1.5	2.5/0.5/16/<0.1	n.a.	n.a.	Late Proterozoic	Slack (1993)
Stone Hill, USA	1.3	0.6/0.8/n.a.	Po,py,cpy,sph	Chlorite-rich mafic schist, schist, gneiss and quartzite	Late Proterozoic	Meathery & Hollister (1984) Nui & Leshner (1991)
Denali, USA	4.5	2/n.a.	Cpy,py,bn,cc	Tuffaceous ash, black graphitic argillite, shale	Triassic	Seraphim (1975)
Green Mountain, USA	3.6	0.8/1.4/n.a.	Po,sph,cpy,gn	Pelitic, arenaceous seds, subvolc mafic dykes, sills	Jurassic	Mattinen & Bennett (1986)
Grey Eagle, USA	10	1/n.a.	n.a.	n.a.	Jurassic	Slack (1993)
Mountain City, USA	2	n.a.	n.a.	n.a.	Ordovician	Slack (1993)
Beatson, USA	6	1.7/n.a.9/0.1	Py,po,cpy,sph	Intercalated slate and graywacke	Early Tertiary	Crowe <i>et al.</i> (1992)
Tverfjell, Norway	19	1/1.2/13/n.a.	Py,sph,cpy,po	Dark to light gray phyllites, metabasalts	Ordovician(?)	Fox (1984)
Killingdal, Norway	3	1.9/5.9/23/0.9	Py,sph,cpy,po	Dark to light gray phyllites, metabasalts	Ordovician(?)	Rui (1973)
Folldal, Norway	3	1.5/3/n.a.	Py,sph,cpy,po	Dark to light gray phyllites, metabasalts	Ordovician(?)	Fox (1984)
Hersfjo, Norway	3.2	1.4/1.4/n.a.	Py,sph,cpy,po	Dark to light gray phyllites, metabasalts	Ordovician(?)	Fox (1984)
Grassmere, Australia	n.a.	n.a.	Cpy	Metasediments and metamafic volcanics	Late Proterozoic(?)	Degeling <i>et al.</i> (1986)
Peak Downs, Australia	n.a.	n.a.	Cpy in qtz-mag-py	Sericite-quartz schist, chlorite-quartz schist, siliceous ironstone	Cambrian(?)	Cox (1975), Murray (1986)
Kanmantoo, Australia	12	1.1/n.a.	Cpy,po,mag	Garnet-andalusite schist, quartz-feldspar schist, quartz-mica schist	Cambrian	Verwoerd & Cleghorn (1975)
Girilambone, Australia	n.a.	n.a.	Cpy	Banded quartzites, metaseds, metamafic volcanics	Ordovician	Degeling <i>et al.</i> (1986)
Mt Molloy, Australia	n.a.	n.a.	Py,cpy,sph,gn	Spilitised basalts	Silurian-Devonian	Gregory & Robinson (1984)

Note: n.a.- not available

Table 4 Summary of Besshi-type deposits in China, Philippines, Uganda, Europe and Chile

Deposit	Size(Mt)	Cu/%Zn%/Ag%/Au(g/t)	Ore mineralogy	Host rock	Age	Reference
Tongchangjie, China	n.a.	1.9/0.8/n.a.	Py,po,cpy,mag	Basalts, basaltic tuff, tuffaceous siltstone, siliceous shale, limestone	Palaeozoic	Yang & Mo (1990)
Liwu, China	20	1.8/n.a./30/0.3	n.a.	n.a.	Triassic	Slack (1993)
Hixbar, Philippines	1.1	1.2/1.1/14/2	Cpy,sph,sgene cc	Quartz sericite schist in chlorite schist	Ordovician (?)	Bryner (1969)
Rapu-Rapu, Philippines	3.2	1.8/2.5/40/2.5	n.a.	n.a.	Early Tertiary	Slack (1993)
Kilembe, Uganda	19	1.8/n.a.	Cpy, cobalt py	Chlorite-biotite schist, amphibolitic schists	Early Proterozoic	Warden (1985)
Saladipura, India	112	n.a.	Py,arpy,po,sph	Carbon phyllites, marbles, quartzites, amphibolites	Middle Proterozoic	Das Gupta (1970) Sarkar <i>et al.</i> (1980)
Arinteiro, Spain	23	0.6/n.a.	Po,cpy	Garnetiferous amphibolite, siliceous metasediments	Early Palaeozoic	Badham & Williams (1981)
Gairloch, Scotland	2.5	1/1/n.a.	Py,po,cpy,sph	Quartz-carbonate schist, chl hornblende schist, quartz-magnetite schist	Early Proterozoic	Williams <i>et al.</i> (1985) Jones <i>et al.</i> (1987)
Kizil-Dere, Georgia	40	2/n.a.	n.a.	n.a.	Jurassic	Slack (1993)
Tisova, Czech Republic	1.5	0.7/n.a.	Py,cpy,po	Phyllites, quartzites, metabasites	Ordovician (?)	Pouba & Ilavsky(1986)
Hermioni, Greece	n.a.	2.9/0.6/20/0.4	Py,cpy,sph	Schists, sandstones and diabase	Cretaceous	Marinos (1982)
Altin-Tepe, Romania	12	n.a.	n.a.	n.a.	Late Proterozoic(?)	Slack (1993)
Gjegjan, Albania	n.a.	n.a.	Py,cpy,hema,mag	Argillo-siliceous-hematitic schists, diabase	Triassic	Pumo <i>et al.</i> (1982)
Walchen, Austria	n.a.	1/n.a.	Py,cpy,mag,po	Serpentinite, greenstone, calcareous-mica-schist, black phyll	Ordovician(?)	Derkmann & Klemm (1977) Tarkian & Garbe (1988)
Pirén Alto, Chile	<1	1/<0.1/6/<0.1	Py,po,cpy,sph	Quartz-alb-chl-epid schists, alb-musc-epid schists	Palaeozoic	Alfaro & Frutos (1988) Alfaro & Collao (1990) Schira <i>et al.</i> (1990)

Note: n.a.- not available

3 RECENT DEPOSITS

Close similarities in the occurrence (wide distribution within a sediment-covered sea-floor environment), associated lithologies (fine-grained turbidites and basalts with mid-oceanic-ridge basalt chemistry), sulphide assemblage (Fe-Cu-Zn) and composition (presence of carbonaceous material) of Besshi-type deposits in Japan and the massive sulphide deposits at Middle Valley, Escanaba Trough, Guaymas Basin and the Red Sea generally support an earlier proposal that the ancient and modern deposits are metallogenic analogues.

Major differences include the greater thickness of sediment; the presence of both alkalic and mid-oceanic-ridge basalts and the absence of an underlying mafic complex in the Sanbagawa terrane; the sheet-like nature of Besshi-type deposits compared to the mound-like forms at Middle Valley, Escanaba Trough and Guaymas Basin; the lower Cu/Zn ratio and higher Pb and Ba contents in the modern deposits; and the greater amount of pyrite relative to pyrrhotite and the Co enrichment in the Besshi-type deposits. The abundant sulphate and carbonate gangue mineralisation in the modern sites are absent in the Besshi-type deposits (Koski, 1990).

3.1 Tectonic evolution

The following tectonic model developed for the northeast Pacific spreading centres by Kappel and Franklin (1989), is an attempt to explain the many diverse morphological and structural attributes of medium-spreading ridges and may provide a basis for explaining the distribution and composition of their associated base metal massive sulphide deposits.

Stage 1 is represented by a period of excessive, extrusive axial volcanism that builds the axial crest, with abundant inflated pillow lavas present, while an elongate summit depression (axial valley) is narrow or absent during this stage. Sulphide deposits during this stage occurs immediately adjacent to the principal bounding faults of the elongate depression. The deposits are large in comparison with those of the East Pacific Rise and are composed of pyritic mounds containing copper and zinc. Stage 2 is characterised by rare

to absent volcanism and the development of the elongate summit depression, which widens by collapse of the summit of the crestral ridge, as a result of tectonic stretching. Sulphide deposits form at similar positions as during stage 1. Stage 3 has renewed volcanism along axial fissures in the floor of the summit depression. Deposits are very small relative to those in stage 1 and stage 2, very zinc-rich, copper and iron poor.

Deposits in heavily sedimented failed rifts, such as Middle Valley are the largest (tens of millions of metric tons) and consist of pyrrhotite, zinc-rich mounds. The heat for these deposits is more effectively conserved in the high-temperature reaction zone, since the impermeable sediments inhibit cooling of the heat source by inflow of cold seawater. The largest volcanic-hosted deposits therefore occur in those segments that formed closest in time to a period of major constructional volcanism. Upwelling hydrothermal fluid is focused along the principal marginal faults of the elongate summit depression and rises from the reservoir without significant cooling or mixing as a result of the relatively impermeable cap rock (Kappel and Franklin, 1989).

3.2 Middle Valley

Middle Valley is a 15 kilometre-wide, sediment-covered axial rift within the Juan de Fuca Ridge system (Fig. 21). Middle Valley was recently an active mid-oceanic spreading centre until 200 000 years ago, when spreading shifted west to the now-active, sediment buried, intermediate-rate spreading centre in West Valley (60 millimetres/year). Middle Valley is still undergoing minor readjustment, as indicated by the presence of fissures and faults within the sediment fill in the valley, which controls the position of present-day hydrothermal activity.

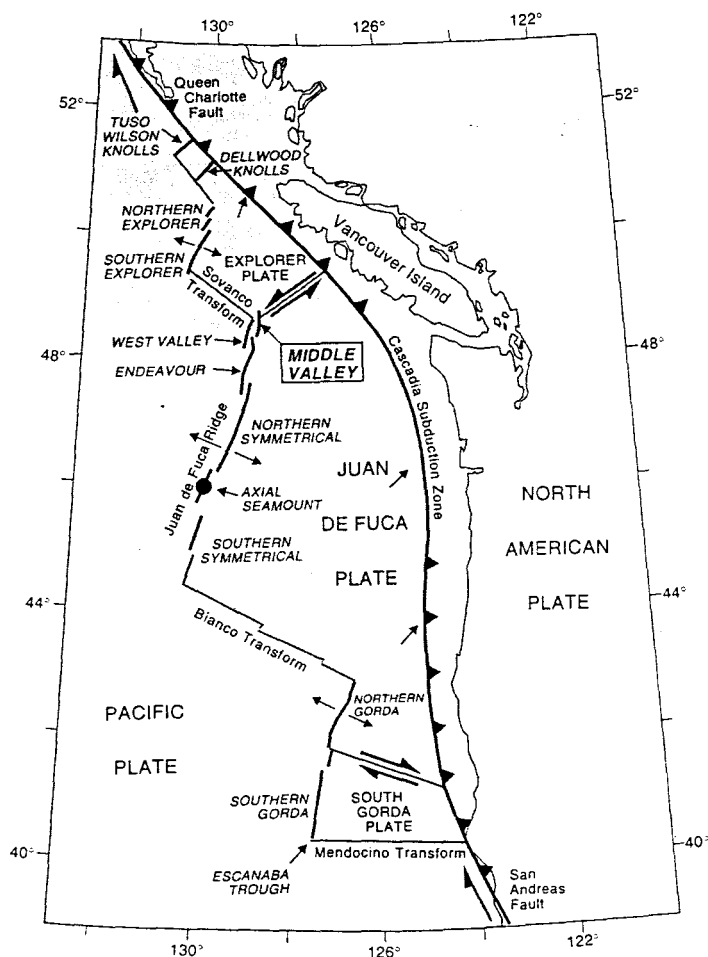


Fig. 21 Map of the Juan de Fuca Ridge system, indicating the main tectonic elements and the location of Middle Valley, northern Juan de Fuca Ridge (After Goodfellow and Franklin, 1993).

Seismic profiles across the valley indicate a series of downstepping block-faults, with the deepest part of the valley along the axis. The surface expression of these faults indicates that most are parallel to the regional north-south trend of the rift-propagating regional faults, but several oblique NW-SE structures are also present (Ames *et al.*, 1993).

Middle Valley is filled mostly by turbiditic and hemipelagic sediments because of its proximity to the continental margin during the Pleistocene, dominantly from Queen Charlotte Sound. The thicknesses of the sediments increase from the margins to the centre of the basin and from south to north, near the Sovanco transform (Fig. 21) (Goodfellow and Blaise, 1988). The composition of the basement at Middle Valley is unknown, but diamond drilling intersected a variety of sills and extrusive rocks, including normal mid-ocean-ridge basalt at the basin margins along the eastern boundary fault.

Two known sites of hydrothermal activity, Bent Hill and the Area of Active Venting (AAV) were identified by geological and geophysical observations. These two sites are situated on the eastern side of the valley, 3 and 5 kilometres, respectively west of the eastern boundary fault, at a water depth of 2 460 metres. Sediment thickness ranges from about 300 metres near the North Sulphide Mound to more than 1 500 metres north of the AAV. The more northerly AAV is defined by anomalously high acoustic reflectance indicating aprons of hydrothermally indurated sediment surrounding over 20 sites (Ames *et al.*, 1993). Another prominent zone of reflectance, south of Bent Hill, is a prominent mound of sulphide, with a single moderate-temperature vent (264 degrees Celsius). A further large pyritic massive sulphide deposit (150 x 150 x 96 metres) has been documented about 100 metres south of Bent Hill and 230 metres north of the active vent.

Bent Hill forms one of a chain of sediment mounds that are oriented approximately north-south and are parallel to extensional faults bounding the Middle Valley Rift. Bent Hill is a 60 metre-high by 400 metre-wide sediment structure that is abruptly truncated to the

west by a north-south fault. Near the southern margin of Bent Hill is a 35 metre high by 100 metre wide dormant sulphide mound. A nearby anhydrite chimney is actively venting clear fluids of 265 degrees Celsius (Goodfellow and Franklin, 1993). The primary mineral assemblage in the massive clastic sulphides of Bent Hill consists of an open interlocking network of hexagonal pyrrhotite platelets that is partly filled in by black subhedral to euhedral sphalerite with cores and lamellae of Cu-Fe sulphide minerals. The Bent Hill massive sulphide deposit contains up to 11.2 percent Zn, 1.2 percent Cu and 0.26 percent Pb.

The AAV is an 800 by 400 metre vent field which is located immediately east of a right-lateral offset of an extensional rift-parallel fault. The earlier mentioned 20 vent sites and associated hydrothermal mounds in the AAV are oriented northwest along what may be a conjugate set of orthogonal extensional faults. These mounds are up to 15 metres high and 100 metres wide, with anhydrite chimneys and rubble derived from collapsed chimneys near their summits. Hydrocarbon-rich barite-silica chimneys that formed at about 265 degrees Celsius occur on flat surfaces between mounds. Hydrothermal sediment in the AAV consists of Mg-rich smectite, Cu-Fe sulphide (similar to isocubanite), pyrite, sphalerite and galena (Goodfellow and Franklin, 1993).

Average $\delta^{34}\text{S}$ values for pyrrhotite and pyrite from Bent Hill are +8.1 and +4.1 per mil, respectively (Fig. 22). The consistently lower $\delta^{34}\text{S}$ values reflect the common replacement of pyrrhotite by pyrite within sulphide chimneys. Barite in the Bent Hill massive sulphide deposit has an average value of +10.1 per mil, whereas gypsum has values similar to modern seawater, of about +20.8 per mil (Goodfellow and Blaise, 1988). Both Bent Hill and the AAV sulphides have $\delta^{34}\text{S}$ values more positive than those of sulphides formed at unsedimented ridges and of magmatic sulphur from mid-ocean-ridge basalts (Goodfellow and Franklin, 1993). This suggests that a major component of sulphur at Middle Valley originated from the reduction of seawater sulphate. Barite with $\delta^{34}\text{S}$ values considerable less than modern seawater probably formed by the mixing of Ba in vent fluids with sulphate formed by the oxidation of pre-existing sulphide,

either within the chimney structure or in the underlying sulphide mound. This mechanism is consistent with the common occurrence of barite in the interstices of interlocking pyrrhotite crystals (Goodfellow and Franklin, 1993). Ames *et al.* (1993) came to the conclusion that at Middle Valley, two completely different types of deposits are defined by their contrasting assemblages of minerals, geochemistry and source compositions.

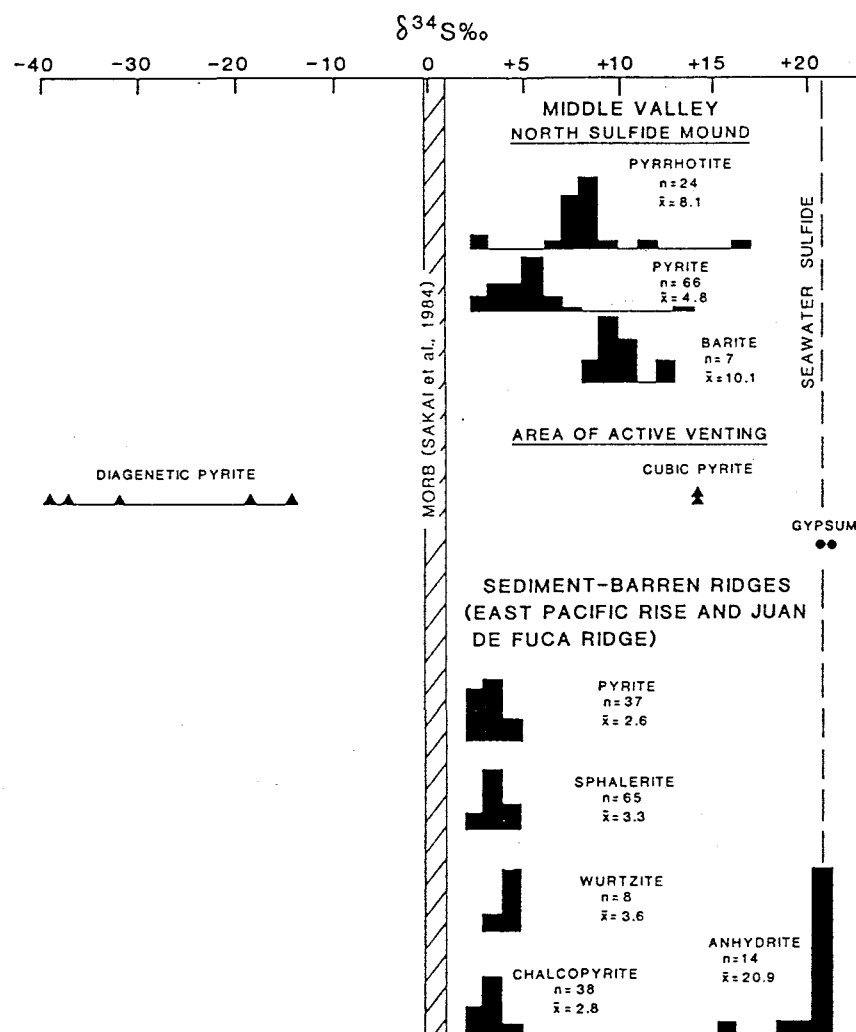


Fig. 22 Histogram of $\delta^{34}\text{S}$ values for pyrrhotite, pyrite and barite from the North sulphide mound and pyrite and gypsum from the Area of Active Venting. Histograms of $\delta^{34}\text{S}$ values for pyrite, wurtzite, sphalerite, chalcopyrite and anhydrite from the sediment-barren East Pacific Rise and Juan de Fuca Ridge, together with the range of $\delta^{34}\text{S}$ values in oceanic basalts and seawater (After Goodfellow and Blaise, 1988).

Mineralisation at the AAV and the active vent at Bent Hill is sulphur-poor and formed from fluids indicative of seawater-sediment interaction (Figs. 23 and 24). The older sulphide deposits at Bent Hill formed from fluids with a higher S_2 activity and O_2 activity together with a much greater degree of fluid-basalt interaction.

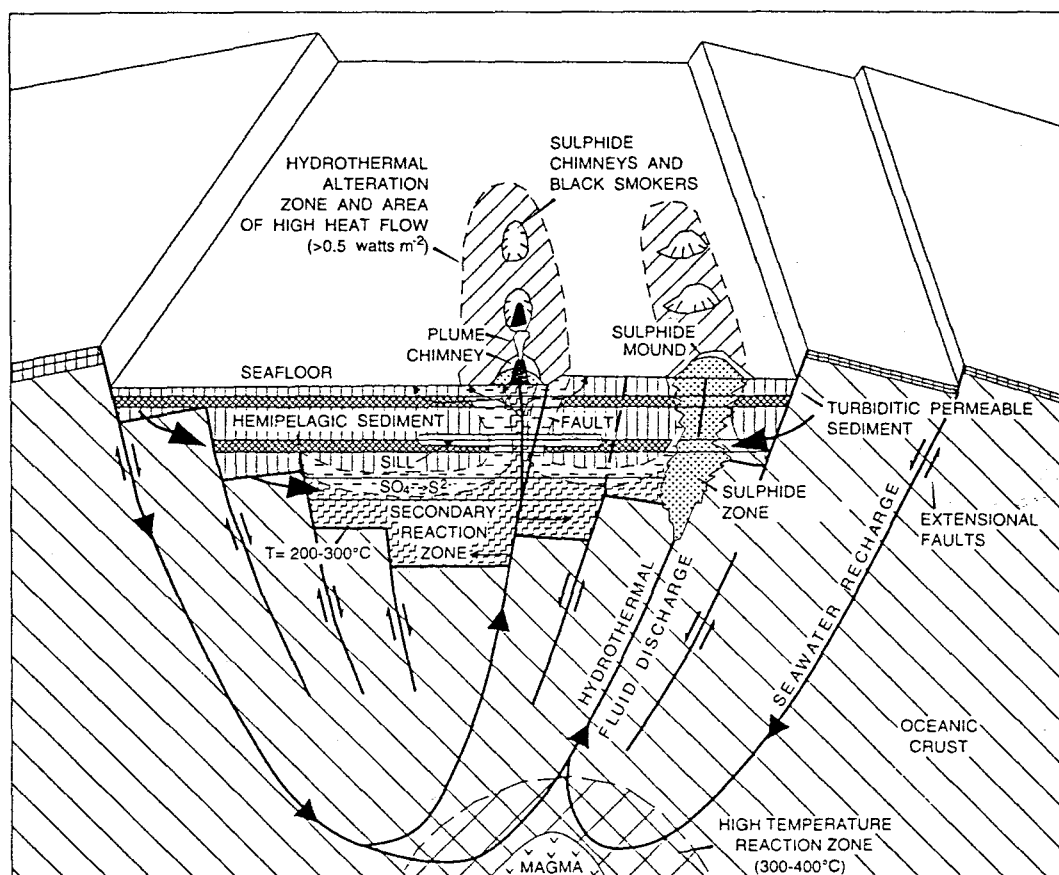


Fig. 23 Schematic diagram of the Middle Valley sedimented rift indicating the distribution of sulphide deposits and hydrothermally active anhydrite chimneys. Hydrothermal alteration associated with fluid upflow, possible hydrothermal circulation patterns and a hydrothermal reaction zone in oceanic crust adjacent to a magma body are also indicated. High-temperature, metal-rich fluids which formed the Bent Hill sulphide deposit, were generated in a lower temperature reaction zone in the lower part of the sedimentary sequence in Middle Valley (After Goodfellow and Franklin, 1993).

Indicators of a significant contribution of sedimentary material to the venting hydrothermal fluids are: low levels of total base metals and high levels of Pb compared to unsedimented ridge deposits, a bulk enrichment in Sr, a characteristic pyrrhotite-dominant sulphide assemblage, Pb and Sr isotope data (Goodfellow and Franklin, 1993) and high-pH fluids with low S_2 activity and O_2 activity based on mineral chemistry and associated assemblages. Indicators of a significant basaltic component to the hydrothermal fluid that precipitated the Bent Hill massive sulphide deposits includes: high copper and zinc content, low Pb and Sr isotopic values and abundant primary pyrite and pyrrhotite (Goodfellow and Franklin, 1993).

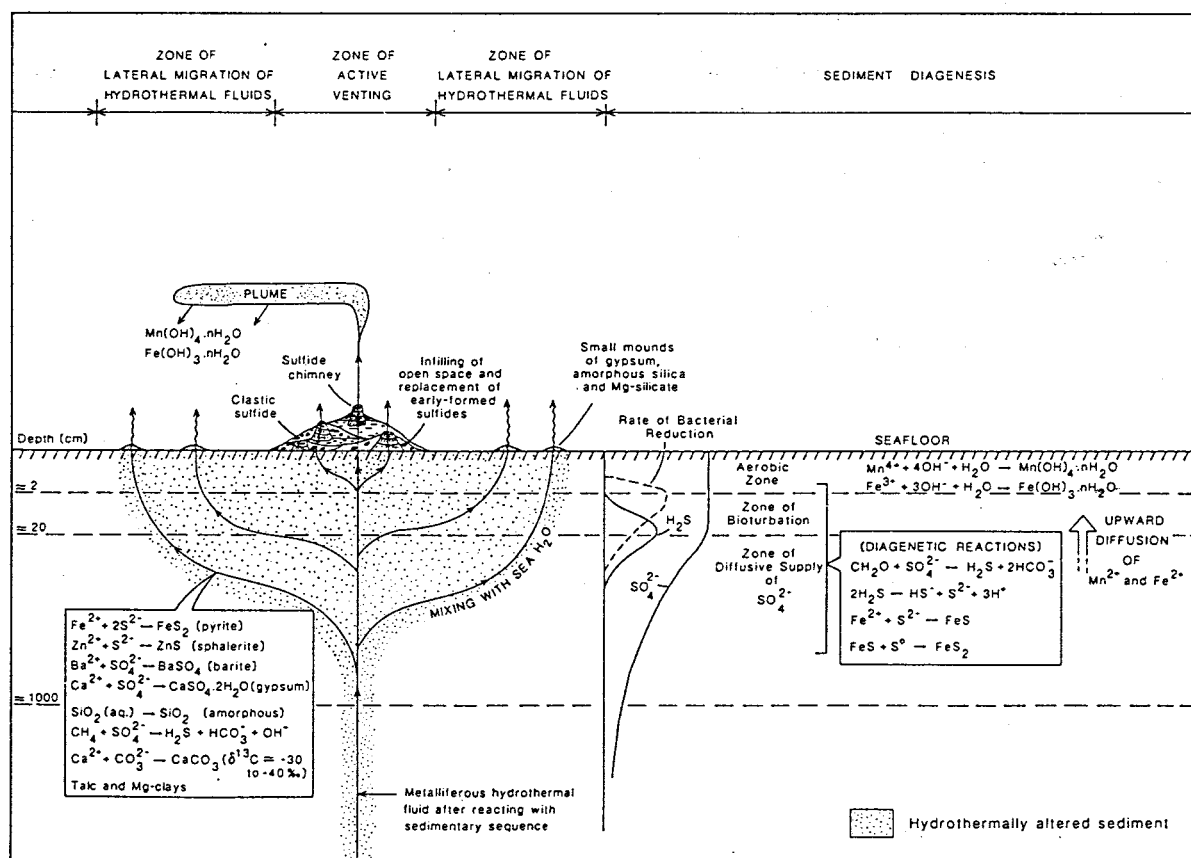


Fig. 24 Schematic model for the formation of hydrothermal mounds, Middle Valley. Basic chemical reactions in the hydrothermal alteration zone and relative changes in H_2S and SO_4^{2-} contents in pore waters are also indicated (After Goodfellow and Blaise, 1988).

Late-stage hydrothermal fluids have affected the original sulphide assemblage at Bent Hill, as indicated by the recrystallisation of sulphides, the formation of magnetite beneath Bent Hill, the dissolution of zinc and gold at depth and their subsequent precipitation near the mound surface. This resulted in a zone-refined deposit typical of many ancient deposits on land (Franklin *et al.*, 1981).

Ames *et al.* (1993) postulated that the two hydrothermal events, represented by the active vents (AAV and Bent Hill) and the massive sulphide deposits at Bent Hill, may represent a continuum process of crustal cooling within a recently active rift environment or two independent events with a coincidental spatial relationship.

3.3 Escanaba Trough

The Gorda Ridge spreading centre is located in the northeast Pacific Ocean and is bound by the Mendocino fracture zone in the south and the Blanco fracture zone to the north (Fig. 25). The northernmost spreading segment trends north-northeast and has a full-spreading rate of about 55 millimetres per annum, decreasing to the south with the Escanaba Trough opening at a total rate of about 24 millimetres per annum (Zierenberg *et al.*, 1993). The tectonic setting of Escanaba is therefore an open-ocean ridge near a continental margin (Koski *et al.*, 1990).

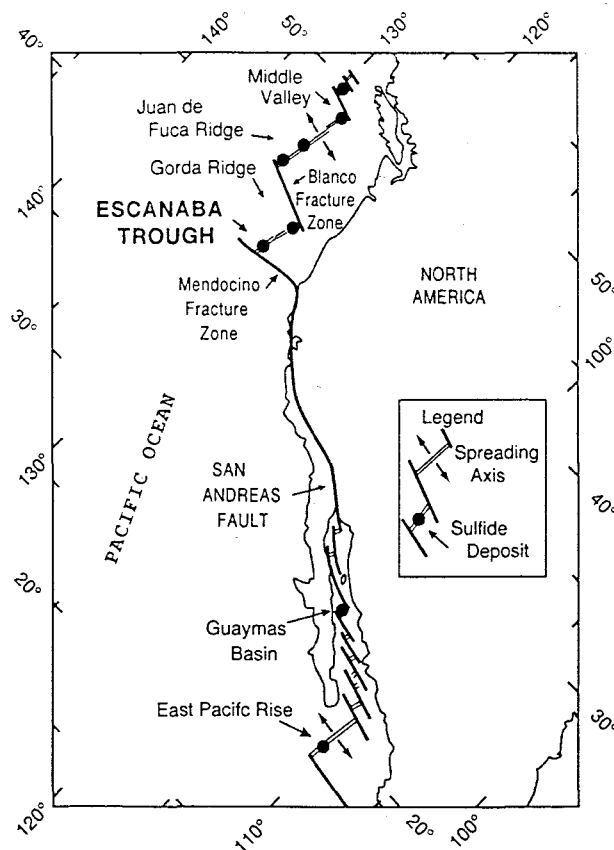


Fig. 25 Locality map indicating the tectonic setting of the Escanaba Trough at the southern end of the Gorda Ridge spreading centre close to the intersection with the Mendocino fracture zone (After Zierenberg *et al.*, 1993).

The morphology of the Escanaba Trough is consistent with this slow-spreading rate and it therefore consists of a wide axial valley floor bounded by steep normal faults, along which the valley floor has been dropped down from 900 to 1 500 metres with respect to its flanking ridges. The axial valley, which is situated at a depth of 3 300 metres, increases in width from 5 kilometres in the north to more than 15 kilometres close to the intersection with the Mendocino fracture zone.

Proximity to the North American continent has resulted in the burial of the southern two thirds of the trough by hemipelagic and turbiditic sediment of between 300 and 700 meters thick over the spreading axis, thinning towards the north. Sedimentation was relatively rapid at 25 millimetres per annum during low stands of sea level in the Pleistocene and the entire sediment fill of the trough was probably deposited within the past 100 000 years (Zierenberg *et al.*, 1993).

Recent igneous activity is confined to discrete volcanic centres 3 to 5 kilometres in diameter, with thinner sediment cover where sills intrude and disrupt the sediment and basalt flows which extruded locally onto the sea floor. Detailed geological mapping and sampling by surface ship and submersible at these two centres together with reconnaissance at two other sites have revealed six major massive sulphide deposits and numerous sulphide occurrences. The sulphide deposits are associated with the volcanic edifices, but are all hosted within the sediments. A variety of hydrothermal sulphides and sulphates were deposited during large-scale fluid discharge through the sediment-cover such as pyrrhotite-rich sulphide deposits, polymetallic sulphide assemblages and sulphate-dominant deposits (Koski *et al.*, 1990). Pyrrhotite-rich sulphide deposits include mounds and chimneys, cement in sedimentary breccia and clastic layers (Fig. 26). Pyrrhotite-rich mounds and chimneys are porous sulphide aggregates containing subordinate amounts of isocubanite, chalcopyrite, sphalerite, arsenopyrite, loellingite and barite, trace amounts of native bismuth, bismuth telluride and alabandite.

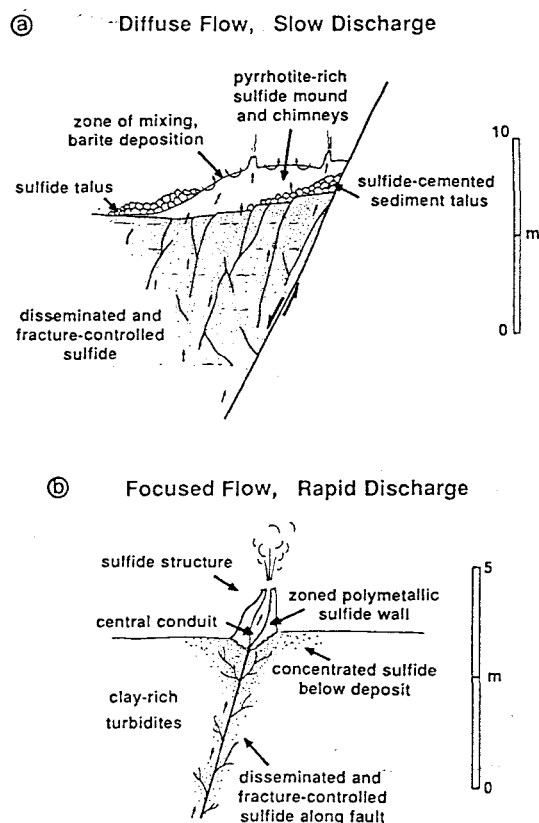


Fig. 26 Generalised models for the formation of (a) a pyrrhotite-rich sulphide mound, and (b) a zoned polymetallic sulphide structure (After Koski *et al.*, 1988).

Pyrrhotite-rich samples contain large amounts of Cu (21 percent), As (1.6 percent), Bi (820 ppm), Co (3 000 ppm) and Au (10.1 ppm). Sulphur isotopes from pyrrhotite and marcasite (average $\delta^{34}\text{S}$ of +6.1 per mil) are enriched in $\delta^{34}\text{S}$ relative to sulphides at unsedimented ridges (Koski *et al.*, 1990). Less abundant polymetallic sulphide assemblages occur as zoned chimney-like structures and cement in sandy turbidite (Fig. 26). Polymetallic sulphides consist of abundant sphalerite, isocubanite, pyrrhotite, galena and arsenopyrite, and traces of loellingite, tetrahedrite and stannite. This sulphide type is enriched in Zn (43 percent), Pb (14 percent), As (2.8 percent), Ag (680 ppm) and Sn (1 500 ppm) (Koski *et al.*, 1990).

Sulphate-dominant deposits comprise barite-rich crusts and chimneys formed on sulphide mounds, isolated barite chimneys on sediment and sulphate-silicate (barite + anhydrite + stevensite + amorphous silica) sinter deposited around active vents at 220 degrees Celsius.

Sulphur isotope ratios for barite and anhydrite range from +19.7 to +21.5 per mil, similar to values for seawater sulphate. The sulphur isotope data from pyrrhotite-rich massive sulphides indicate a mixture of basaltic and seawater sulphate (as seen above), but the depth of fluid penetration is unknown (Fig. 27).

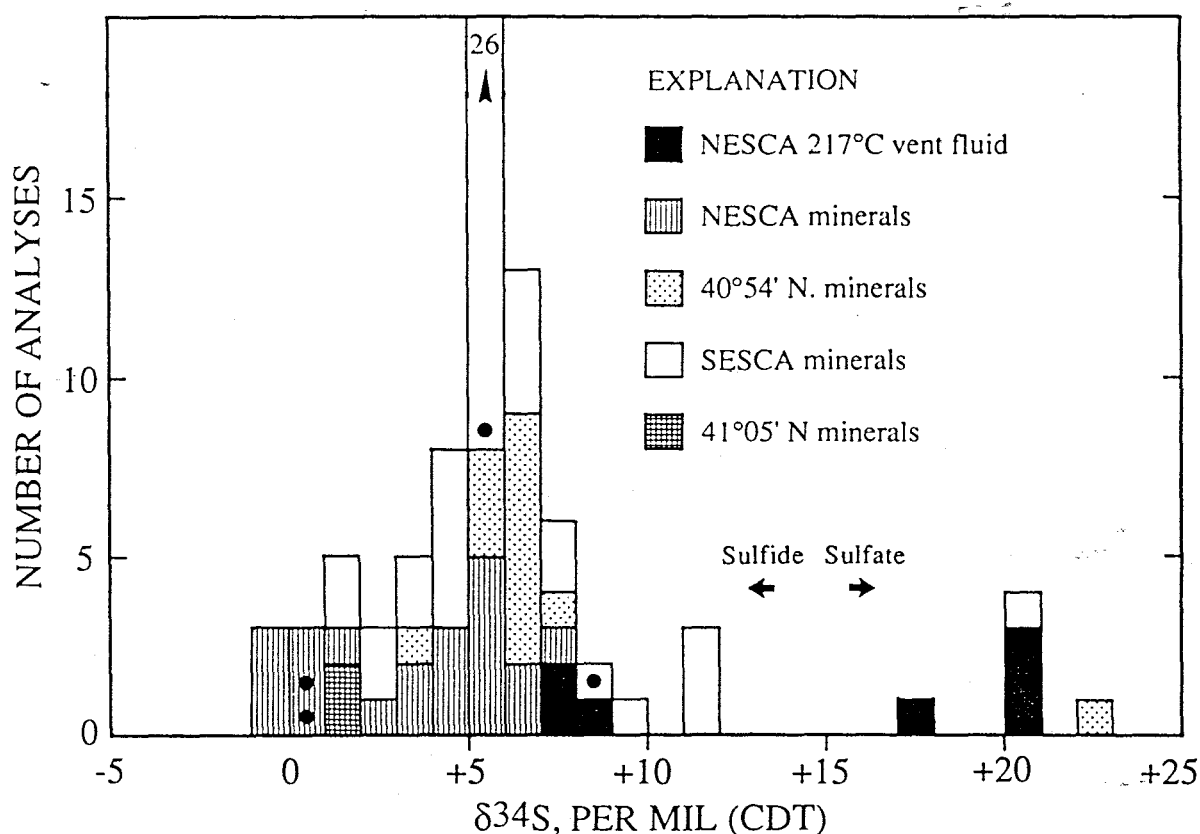


Fig. 27 Histogram of $\delta^{34}\text{S}$ values for Escanaba Trough sulphide and sulphate minerals. Dots indicate samples from which hydrocarbons have been reported (After Böhlke and Shanks, 1994).

The Pb isotope ratios of pyrrhotite-rich and polymetallic massive sulphides from the Escanaba Trough (LeHuray *et al.*, 1988) are nearly identical to nearby sampled turbidite, so no component of basaltic Pb is evident (Koski *et al.*, 1988). The abundance and isotopic composition of sulphides, sulphates, Mg silicates in the sulphide mounds and subjacent sediment require the near-surface mixing of

seawater with large volumes of hydrothermal fluid at temperatures near 200 degrees Celsius (Koski *et al.*, 1990).

The relatively impermeable sediment cover acts as a thermal blanket, preventing large-scale convective cooling by circulating sea water through the ridge crest. Hydrothermal circulation is restricted to areas where tectonic and volcanic processes provide cross-stratal permeability. Large-scale control on the location of massive sulphide deposits is provided by the volcanic centres, which serve as topographic and thermal foci for the discharge of hydrothermal fluids. This focusing, combined with permeable aquifers at depth in the sediment column, results in long-lived hydrothermal discharge within a limited area and formation of large massive sulphide deposits (Zierenberg *et al.*, 1990). The surface location of individual vent sites and deposits is primarily controlled by small igneous intrusions into sediment that overlies the volcanic centres. Massive sulphide deposits form at the base of these hills by hydrothermal discharge along faults that bound the uplifted areas (Fig. 28).

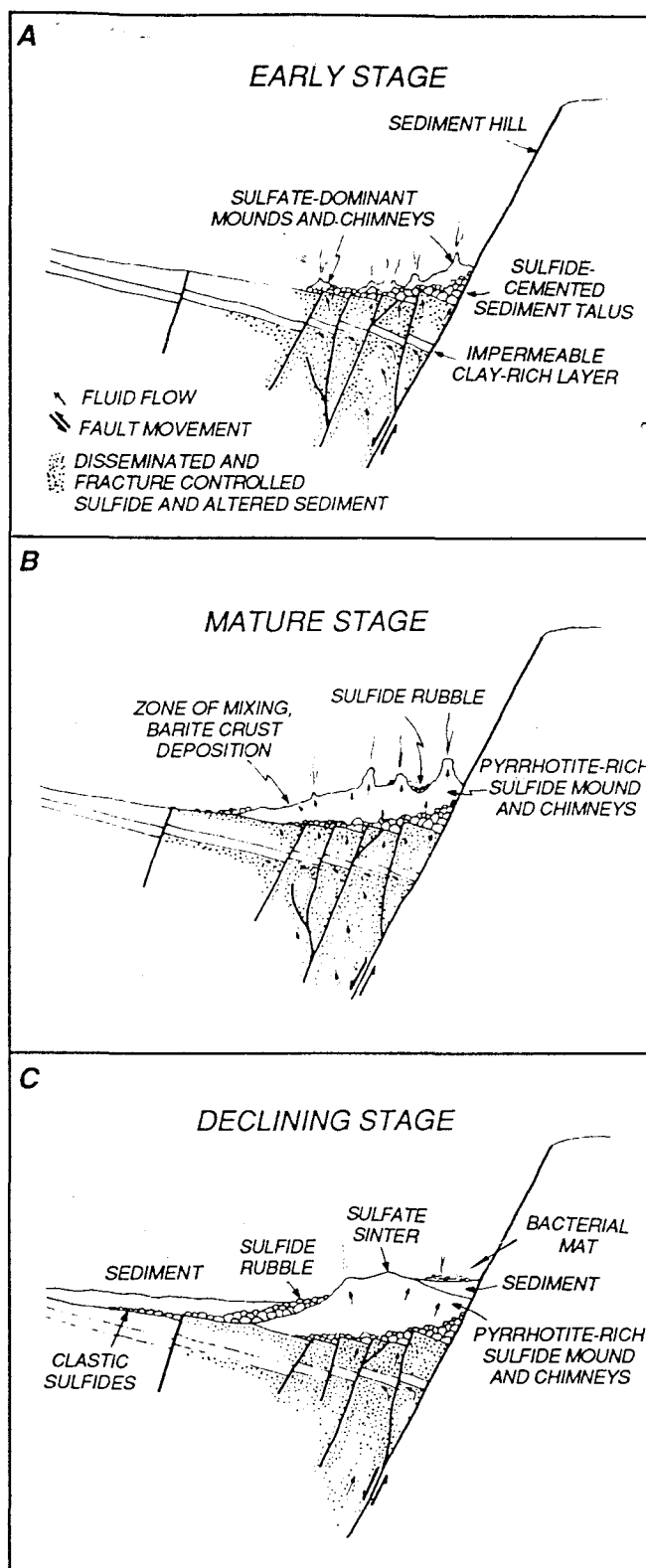


Fig. 28

Generalised model for hydrothermal discharge and deposition of sulphide mounds at the margin of an uplifted sediment hill. The mound is shown at all three stages of development. The bounding fault structure and sediment breccia deposits provide cross-stratal permeability for hydrothermal fluids. Fluids migrate laterally along porous sandy units in the turbidite section. Subsurface mixing of hydrothermal fluid and seawater along faults and within permeable beds results in the formation of disseminated and fracture-controlled sulphide mineralisation, magnesium metasomatism and formation of chlorite in sediment (After Koski *et al.*, 1994).

3.4 Guaymas Basin

Guaymas Basin is part of a complex rift zone in the Gulf of California and is transitional between the purely divergent boundary between the Pacific and North American plates at the East Pacific Rise and the transform boundary separating these plates, represented by the San Andreas fault (Fig. 29) (Koski *et al.*, 1985). The basin consists of two northeast-trending grabens, termed the Northern and Southern Troughs, each about 40 and 20 kilometres long and 3 to 4 kilometres wide, respectively. Oblique separation of Baja California from mainland Mexico is taking place at between 5 and 6 centimetres/year along a series of *en échelon* spreading axis

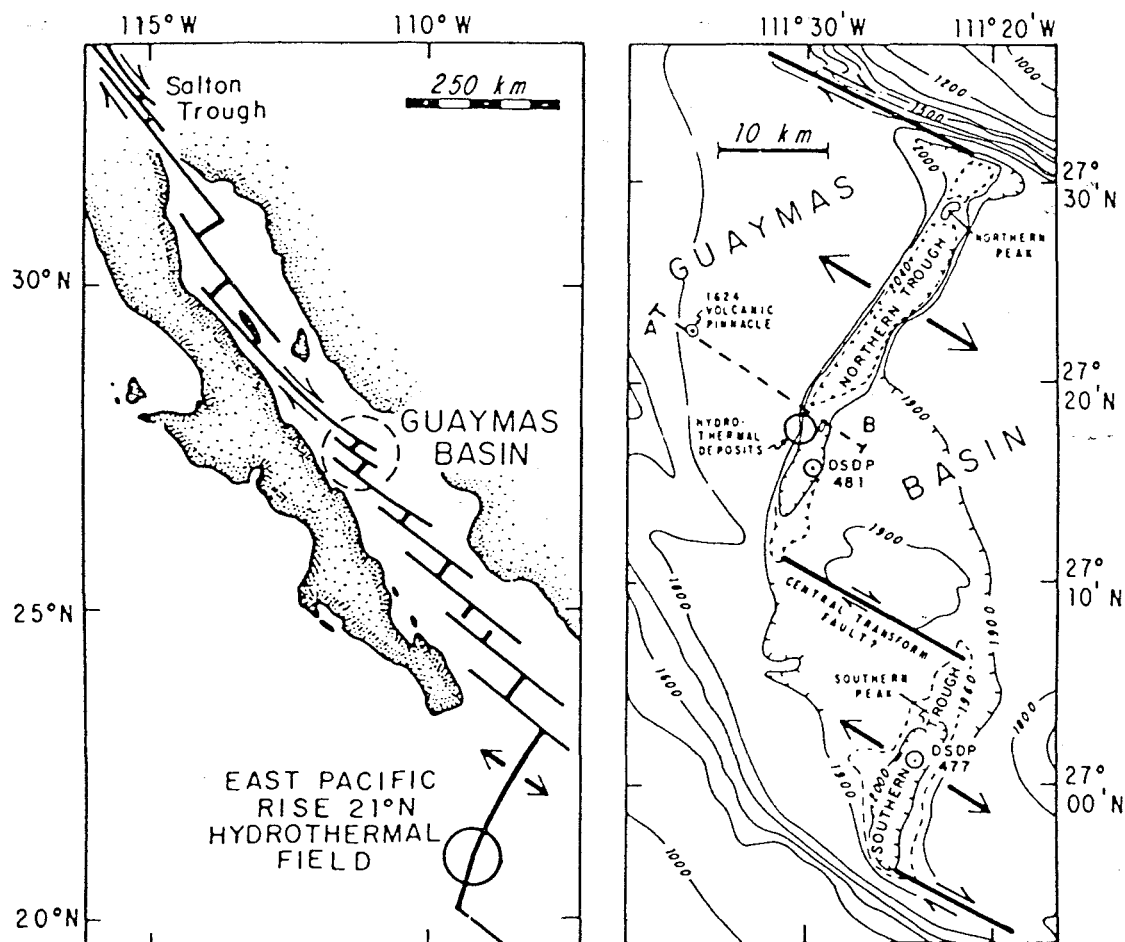


Fig. 29 Locality map of the Guaymas Basin, indicating both the northern and southern troughs (After Von Damm *et al.*, 1985).

segments, offset by transform faults of considerably greater length.

The basin floor is covered with biogenic and terrigenous sediments of at least 500 metres thick, and a high sedimentation rate of 1 to 2 metres per thousand years (Lonsdale and Lawver, 1980). The terrigenous component is derived predominantly from Tertiary volcanics of the Mexican mainland (Peter and Scott, 1988). Ocean-plate accretion occurs through dyke and sill intrusions which have uplifted parts of the rift valley floor (Einsele, 1985) (Fig. 30).

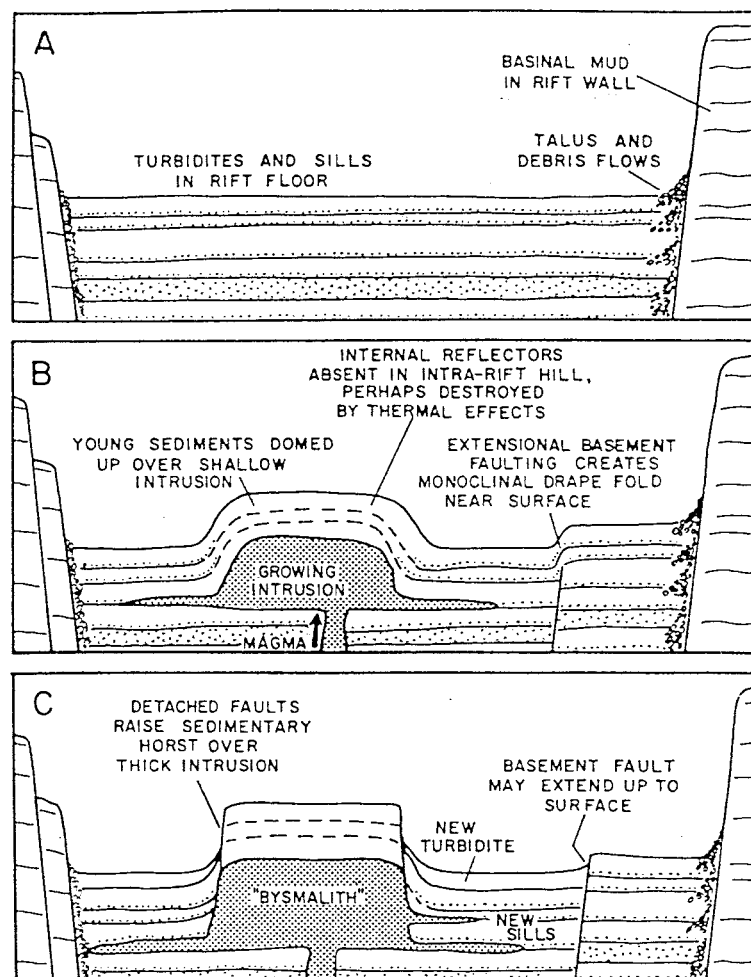


Fig. 30 A mechanism of intrusion, faulting and folding in the development of an immature axial rift valley of the Guaymas Basin (After Lonsdale and Lawver, 1980).

The areas of high conductive heat flow identified in the Southern Trough, are the only places where hydrothermal deposits and plumes have been located, comprising an area of 2 by 6 kilometres. Individual deposits are generally elongate in a northeast-southwest orientation and cluster in lines parallel to faults, which are thought to provide the main plumbing system that supplies the hydrothermal fluids to the sediments (Lonsdale and Becker 1985). Hydrothermal mounds are generally 5 to 25 metres high and 10 to 50 metres across. Chimneys and spires are columnar structures that vary greatly in height from a few centimetres to more than 30 metres (Peter and Scott, 1988). These structures are classified into carbonate- and sulphate-rich and sulphide-rich. The highest exit temperature measured for vent fluids in the Guaymas Basin was 359 degrees Celsius at a sea-floor pressure of 200 bars (Peter and Scott, 1988).

Samples which were collected, consist predominantly of sulphide containing open boxworks and dendritic aggregates of interpenetrating pyrrhotite tablets with porosities of between 40 and 50 percent. Sphalerite, chalcopyrite and galena are much less abundant, together comprising less than 25 percent of the total sulphides. Oxidation of pyrrhotite-rich samples is widespread and occurs as two forms: pyrrhotite-destructive replacement by lepidocrocite and crusts of goethite and amorphous Fe oxyhydroxide coating pyrrhotite and other sulphide grains (Koski *et al.*, 1985).

Sulphur isotope values obtained by Koski *et al.* (1985) from sulphide minerals in massive sulphide samples yielded $\delta^{34}\text{S}$ values ranging from +0.35 to +2.92 per mil, indicating that the dominant sulphide sulphur source is remobilised basaltic sulphide. Lonsdale *et al.* (1980) identified pyrrhotite from a talc-rich mound in the Northern Trough with a sulphur isotope value of -3.94 per mil, suggesting some remobilisation of bacterial sulphide from the sediment.

Figure 31 depicts a model for the growth of chimneys, spires and mound structures.

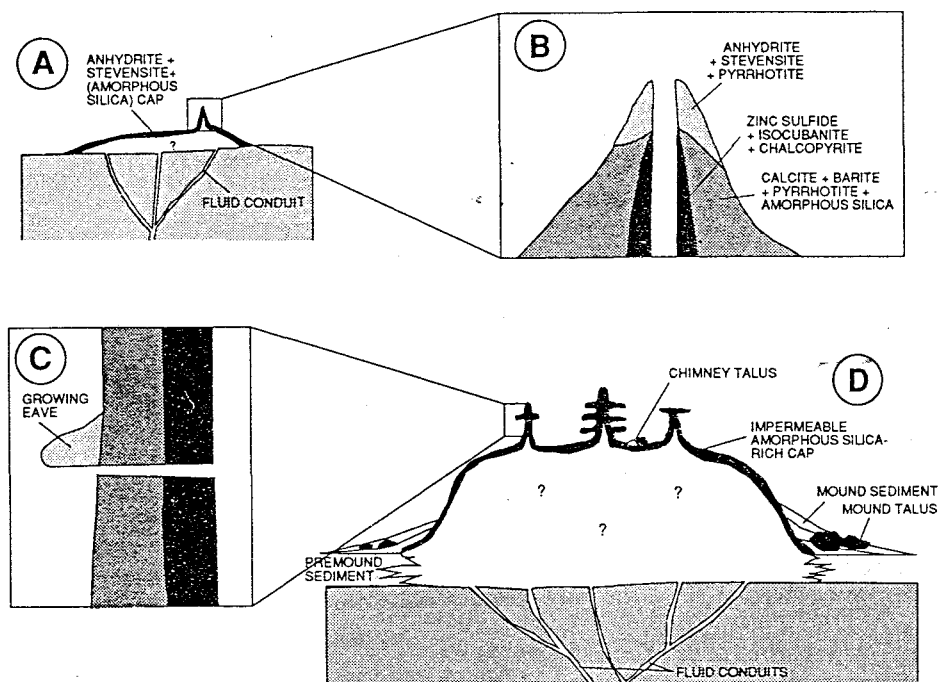


Fig. 31 Schematic development of chimneys, spires and mound structures in the Guaymas Basin (After Peter and Scott, 1988).

Earlier workers summarised evidence for two distinct types of hydrothermal systems encountered during deep sea drilling. The first system is envisaged as the emplacement of shallow sills in the sediment column producing relatively low-temperature, short-duration hydrothermal cells created by expulsion of ambient pore water (Koski *et al.*, 1985). A second and volumetrically more important, circulation system is related to deeper and inferred larger intrusions that involve fluid temperatures >300 degrees Celsius and recharged by circulating seawater. The larger system produces a greenschist-facies metamorphic assemblage (albite - epidote - chlorite - quartz) in the altered sediments and interstitial water compositions (high alkalinity, moderate pH and low Eh) similar to the vent fluid composition. Strontium and oxygen isotope investigation of interstitial waters and mineral phases at the deep sea drilling site, together with high helium 3 concentrations of $^3\text{He}/^4\text{He}$ of between 60 and 70 percent higher than atmospheric helium, indicating injection from the mantle into Guaymas Basin seawater (Lupton, 1979), and the sulphur isotope data mentioned above, all indicate that the fluids have penetrated and reacted with underlying basalt and the hydrothermal system is recharged by down-ward

percolation of seawater (Fig. 32) (Koski *et al.*, 1985). Chen *et al.* (1986) found that the Pb isotopic composition of a hydrothermal fluid from the Guaymas Basin resembles that of the sediments of the Gulf of California. This was found to be consistent with gathering of Pb from the volcanogenic detritus of which the sediment column consists and through which the hot fluids pass.

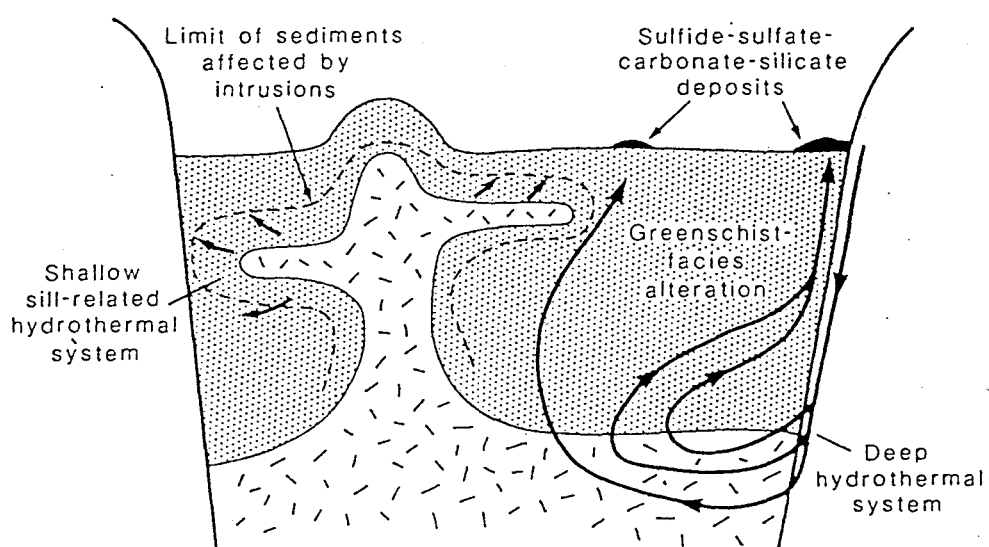


Fig. 32 Schematic illustration of the hydrothermal systems active in the Guaymas Basin. The small-scale circulation related to diabase sills and plugs causes expulsion of pore fluid, reduction in porosity and recrystallisation at temperatures <200 degrees Celsius. Deep circulation of seawater penetrates the igneous basement beneath the sediment pile to a depth of 500 metres, forms greenschist facies (albite - epidote - chlorite - quartz) mineral assemblages during extensive interaction at high temperatures (>300 degrees Celsius) with sediment and exits as reduced, weakly acid ($\text{pH} = 5.9$), but highly alkaline, metal- and sulphide depleted solutions at temperatures ≤ 325 degrees Celsius onto the sea-floor. The large-scale convection is fed by recharge of Guaymas Basin bottom water through the sediments and fault zones bordering the trough (After Koski *et al.*, 1985).

3.5 Red Sea

After the initial discovery of metalliferous hydrothermal sediments in the Atlantis II Deep of the Red Sea in 1965, several organisations became interested in the economic importance of these deposits and their implications for ore genesis. An area in the south was delineated which contains between 150 to 200 million short tons of sediment averaging, on a dry, salt-free basis, 5-to 6 weight percent Zn and 0.8 weight percent Cu (Pottorf and Barnes, 1983).

The Red Sea may have evolved tectonically through a single stage of continuous sea-floor spreading starting about 25 million years ago or may have developed through at least two distinct spreading stages (Pottorf and Barnes, 1983). The first stage occurred from about 41 to 34 million years ago, a questionable interval which has been mainly based on magnetic data. During this period, the African and Arabian shields rifted apart to form a major portion of the Red Sea. A tectonically inactive period of 30 million years followed, during which a sedimentary sequence of up to 5 kilometres thick was deposited. These sediments consist predominantly of evaporites interbedded with minor zinc-rich shales. A second period of spreading was initiated between 4 and 5 million years ago, rifting apart the older oceanic crust and the overlying sediments up to today. The rifting forms a central, northwest-southeast axial trough where new tholeiitic basalt is being added to the crust.

Up to 17 deeps have been identified within the median valley of the Red Sea (Fig. 33). The Atlantis II Deep contains the most extensive deposits and is found in an area where hot brine are actively discharging onto the sea floor. This deep has dimensions of about 5 x 14 kilometres and is roughly elliptical in shape (Fig. 34). The bottom of the pool is approximately 2 200 metres below sea level and is flanked by thick Miocene evaporites. Metalliferous sediments of between 10 to 30 metres thick are found at the bottom of the pool and directly overlie oceanic basalts (Pottorf and Barnes, 1983).

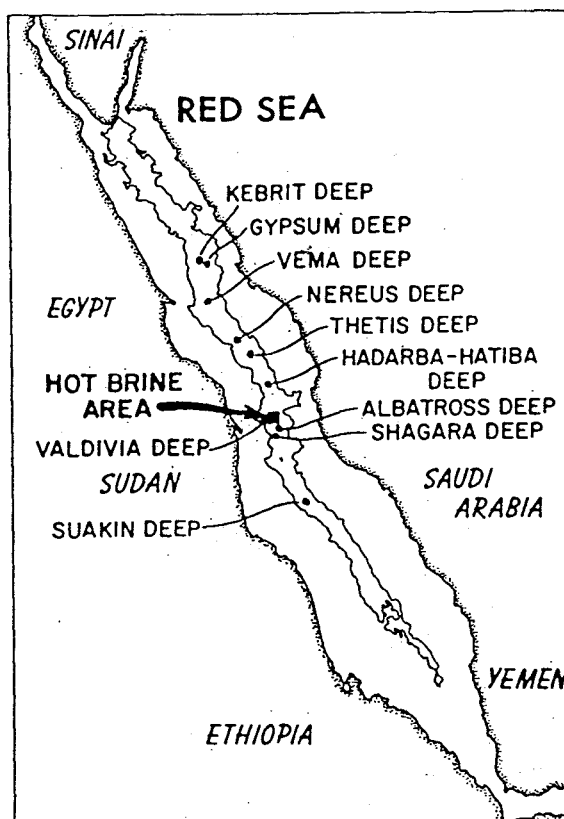


Fig. 33 Locality map indicating several deeps in the axial rift zone of the Red Sea which contain metalliferous sediment and/or saline brines. Hot brine area includes the Atlantis II Deep, Discovery Deep and Chain Deep (After Shanks and Bischoff, 1980).

The Atlantis II Deep contains about 5 cubic kilometres of brines which are stratified into two distinct layers of differing temperature and chlorinity (Fig. 35). The lower brine has a volume of 1.8 cubic kilometres and is about 150 metres thick at a depth of 2 040 metres and the sea-floor. This brine was at 61.5 degrees Celsius in 1977 with a chlorinity of 156 per mil and it contained no free oxygen. The upper brine is 50 metres thick and was at 50 degrees Celsius and had a 82 per mil chlorinity. This brine is partially oxygenated and forms the transitional zone of mixing between the lower brine and the overlying seawater. Detailed temperature-depth profiles indicated that the inflowing brine comes from the southwest basin, where upon discharge, the hot brine convects upward towards the lowermost 50 degree Celsius brine and spreads laterally into the west, east and north basins, respectively. When the 1971 and 1972 observations are combined with an estimate of the volume increase in the lower brine, the

temperature of the inflowing fluid was calculated at 210 degrees Celsius with an average rate of discharge of 24 000 litres per minute. The lower brine in the Atlantis II Deep is enriched relative to seawater in Na, K, Ca and Cl and depleted in Mg, SO_4^{2-} , I, F and NO_3^- . Extremely high volumes of Fe, Mn, Zn, Pb and Cu remain in solution, while H_2S has not been detected in the lower brine (Pottorf and Barnes, 1983).

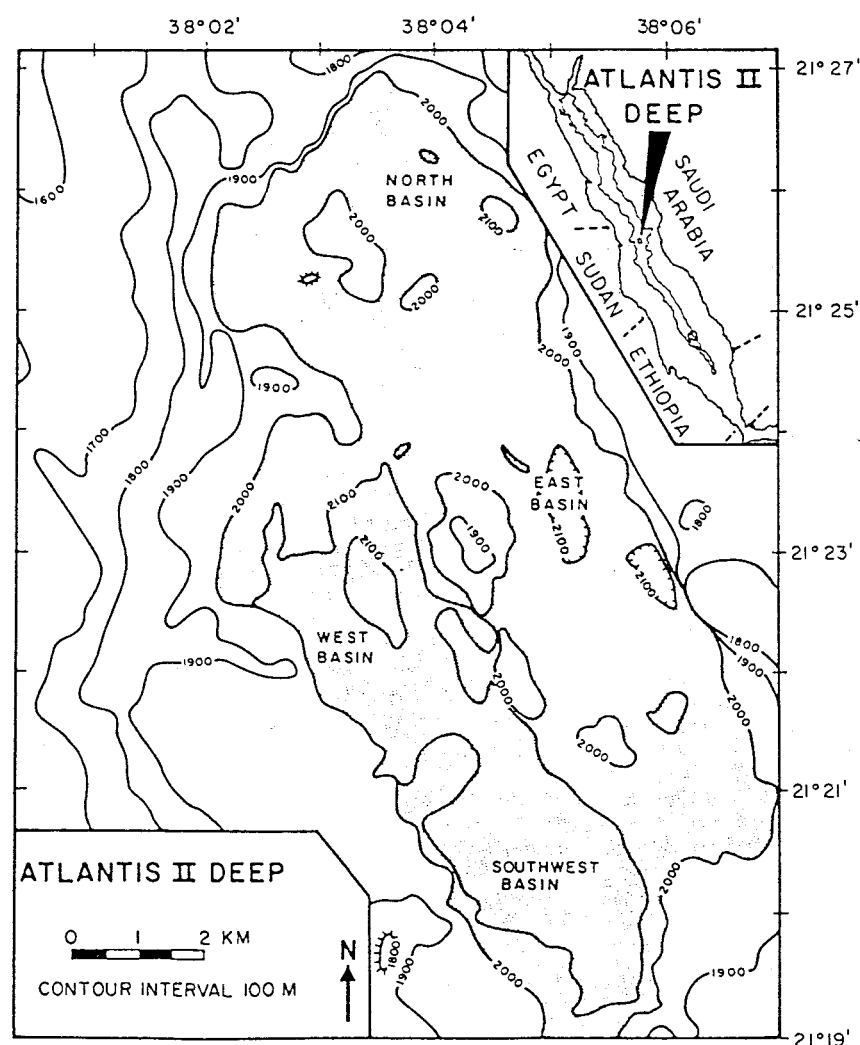


Fig. 34 Insert map indicates the locality of the Atlantis II Deep with the 500 metre contour defining the main trough of the Red Sea. The bathymetric map of Atlantis II Deep graben shows the location of the different basins. The area below 2 000 metres is covered by brine. (After Zierenberg and Shanks, 1988).

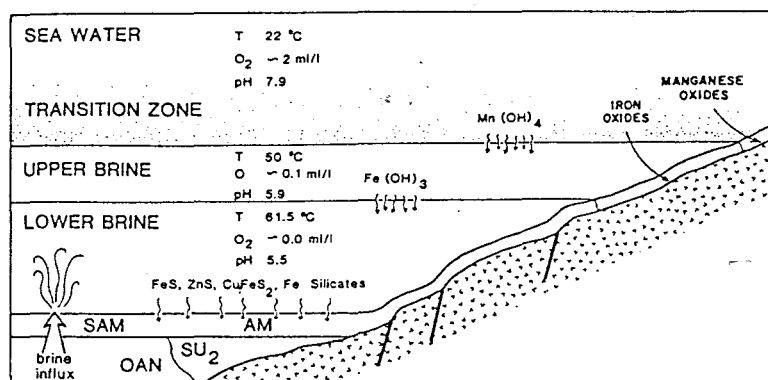


Fig. 35 Schematic representation of precipitation reactions within the density stratified brine pool of the Atlantis II Deep. See Fig. 36 for the explanation of SAM, AM, OAN and SU₂ (After Zierenberg and Shanks, 1988).

The metalliferous sediments of the Atlantis II Deep have been divided into five lithostratigraphic facies that can be roughly correlated throughout the basin (Fig. 36) (Pottorf and Barnes, 1983). Pottorf and Barnes (1983) summarised the stratigraphy based on previous workers as follows: The detrital-oxidic-pyritic (DOP) zone was deposited directly on tholeiitic basalts, which commenced about 25 000 years ago and continued to about 13 000 to 10 000 years ago (Shanks and Bischoff, 1980). The DOP zone consists of detrital carbonates, clays and other silicates with limonitic material. Minerals of hydrothermal origin are quite minor in this zone compared to other units. The lower sulphidic zone (SU1) is the first period of stable, widespread hydrothermal activity. Sphalerite, pyrite, chalcopryrite, hydrothermal clays, manganosiderite and anhydrite are the dominant assemblage. The central oxidic (CO) zone is characterised by oxidised, chaotically bedded sediments suggesting tectonic activity, in contrast to the conditions present during sulphide deposition. Goethite, hematite and silicates are most abundant. Outside the west basin, however, this zone has a large detrital component (Pottorf and Barnes, 1983). The upper sulphidic zone (SU2) represents a return to tectonically stable conditions with a mineralogical assemblage similar to the lower sulphidic zone. Sphalerite occurs with lesser amounts of pyrite and chalcopryrite.

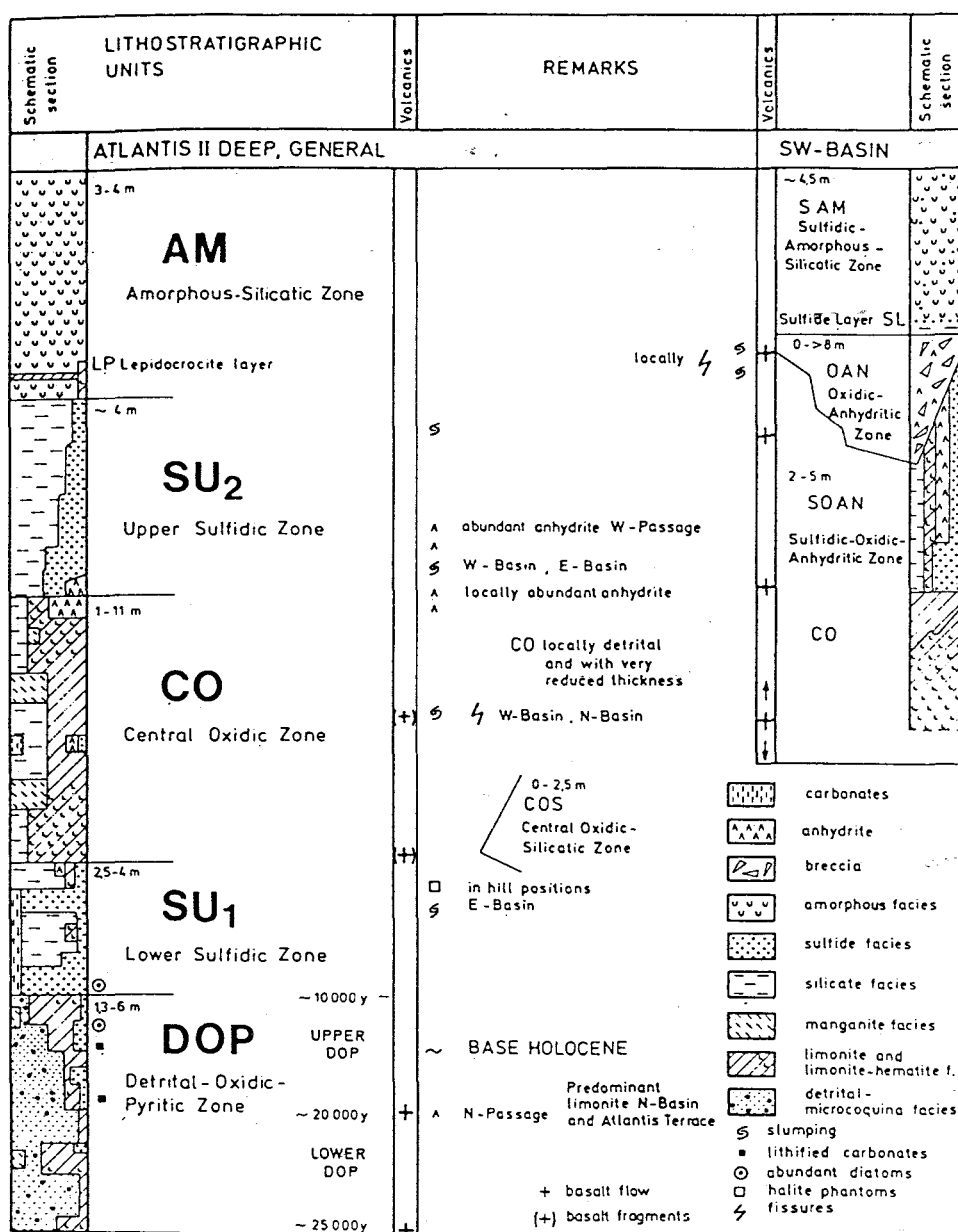


Fig. 36 Generalised lithostratigraphic sequence of the Atlantis II Deep metalliferous sediment (After Zierenberg and Shanks, 1988).

Silicates, light green in colour, carbonates and locally anhydrite are also prevalent. The amorphous silicic (AM) zone is still precipitating and consists mainly of poorly crystalline silicates and oxides with minor sulphides.

The Atlantis II Deep sediments are only the uppermost features of a much larger geothermal system at depth. In the absence of

appreciable drilling data to investigate the characteristics of the rocks and fluids directly below the thin (10 - 30 metres) veneer of sediments on the sea-floor, subsurface processes are inferred, predominantly from geochemical and geophysical data.

The following well-documented evidence was collected by several earlier workers in relation to the lower brine pool (see also Fig. 37):

Oxygen and hydrogen isotope studies indicate that the brines of the Atlantis II Deep originated ultimately from the Red Sea palaeo-waters. The brines are depleted in Mg^{2+} and SO_4^{2-} and enriched in Ca^{2+} with respect to seawater. This chemical trend is identical to the exchange processes observed in all experiments involving basalt-seawater interactions at elevated temperatures. The brines display no $\delta^{18}O$ shift, as is commonly observed for fluids that have undergone high-temperature, water-mineral interactions in response to the water/rock ratio, their isotopic difference, their kinetic response and temperature. No appreciable mixing has been observed between the lower brine and overlying seawater based on trace element and isotopic contents. As previously mentioned, no H_2S was detected in the lower brine. The sulphate content is 840 parts per million and combined Fe + Mn + Cu + Zn + Pb is equal to 169 parts per million, in comparison to Red Sea water which contains 3 140 parts per million sulphate and 0.14 parts per million maximum combined metals. Helium isotopes in the brine require a basaltic (mantle) source, while rare earth and trace element studies suggested an ultramafic source, possibly peridotites. Very reduced minerals, including cubic cubanite, pyrrhotite and high iron sphalerite, were precipitated with anhydrite in veins at about 250 degrees Celsius, indicating SO_4^{2-} - H_2S disequilibrium (Pottorf and Barnes, 1983). Large 100-micron grains of cubic cubanite + chalcopyrite were physically transported by hydrothermal fluids, quenched from above 210 degrees Celsius and deposited on the sea floor (Pottorf and Barnes, 1983).

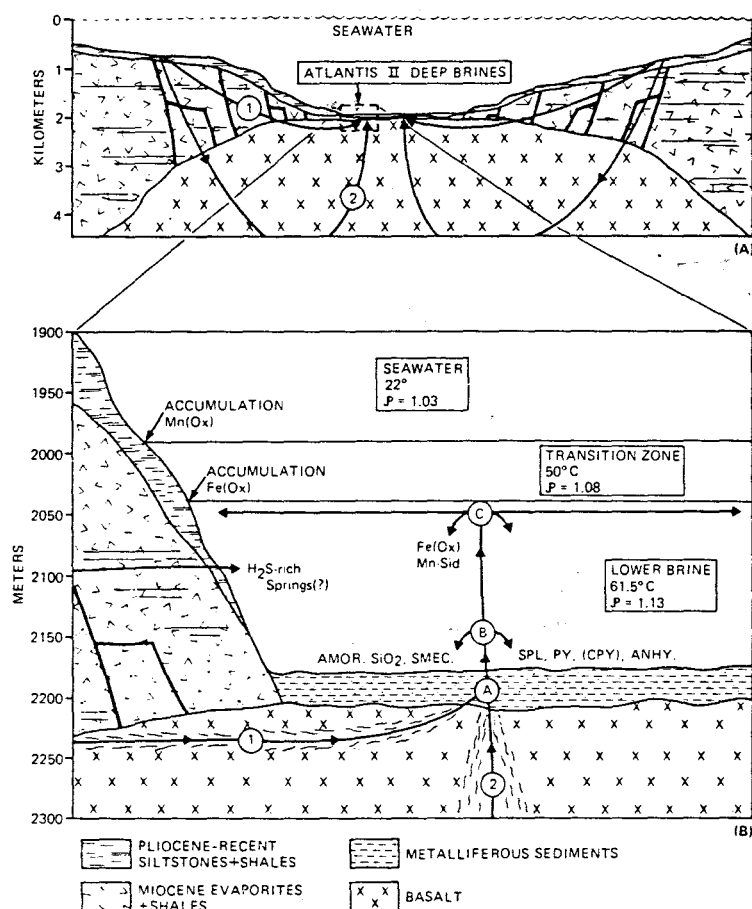


Fig. 37 Summary of processes in the genesis of Atlantis II Deep deposits (After Pottorf and Barnes, 1983).

- A. A schematic northeast-southwest cross section through the axial trough of the Atlantis II Deep and the Red Sea. Palaeo-seawater circulates through fractures in the evaporite-shale sequence on the basin flanks and is heated by and reacts with basalts. Fluids 1 and 2 represent shallow and deep circulation patterns, respectively.
- B. A cross section through the discharge area. Shallow-circulating fluid 1 is relatively oxidised with $\Sigma SO_4 > H_2S$, below 250 degrees Celsius and highly saline. Fluid 2, minor in volume compared to fluid 1, is very reduced with $H_2S > \Sigma SO_4$ and above 334 ± 17 degrees Celsius, highly saline and evolved through deep circulation. Both fluids alter basalt locally to chlorite (wavy lines). Reduced minerals precipitate from fluid 2 in veins in the basalts directly beneath the metalliferous sediments. At A, fluids 1 and 2 mix where fissures or veins of sulphates, sulphides and silicates occur in the sediments. This composite fluid discharges onto the sea-floor at a temperature of 200 - 250 degrees Celsius and a density of about 0.98 to 1.03, whereupon it rises, cools and precipitates phases, shown at B, until it is deflected by the less dense, partially oxygenated transition zone at C. This fluid circulates across the Deep to transport fine-grained sulphides.

Table 5 Summary of sea-floor deposits equated to Besshi-type deposits of Japan

Deposit	Ore mineralogy	Host rock	Tectonic setting
Middle Valley	Po, py, marc, sph, talc > isocub, cpy > cov, gn	Fine-grained terrigenous turbidite overlying probable MORB igneous basement, high heat flow areas have hydrothermally altered sed with carb cement	Mid-oceanic ridge near continental margin, medium-rate spreading axis
Escanaba Trough	Pyrrhotite-rich sulphide: po, marc > isocub, arspy > lollingite, gn > native bismuth Polymetallic sulphide: sph, po, gn > isocub, arspy > lollingite	Arkosic silt, sand and mud turbidite overlying pillow basalt (low-K MORB)	Open-oceanic ridge near continental margin, slow-rate spreading axis
Guaymas Basin	Mounds: calc, barite, amorph silica, po, marc, talc, stevensite, anhydrite > py, sph, cpy, isocub > gn Chimney: calcite, barite, anhydrite, amorph silica, talc > po, sph, wurtzite, isocub > gn, cpy	Mud turbidite and interbedded diatomaceous ooze intruded by diabasic sills of MORB composition	Rifted continental margin, medium-rate spreading axis
Red Sea	Sph, py, cpy, po, manganosiderite, anhydrite, goethite, hematite	Metalliferous sediments containing detrital carbonates, clays, silicates and anhydrite overlying oceanic basalts (tholeiites) and adjacent Tertiary evaporites and shales	Two stage intracontinental spreading rift

Note: n.a.- not available

4 ORE DEPOSIT MODEL

4.1 Genetic model

A genetic model for a mineral deposit type is an attempt to give a rational and consistent explanation of the characteristics of the deposit type in terms of physical and chemical processes (Lydon, 1988). The following sections will cover the most pertinent processes and related geological aspects which have been prominent in the recent literature: source of metals and heat, generation and contents of ore fluids, types of fluids, sulphide accumulation, palaeotectonic setting and isotopic characteristics.

4.1.1 Source of metals

Until about 1967 there existed an assumption that the exhalations from which the sulphide precipitates were derived were products of a crystallising, differentiating, subvolcanic melt which means that the gasses and solutions were assumed to be products of active sub-sea-floor magmatic activity (Stanton, 1991).

From 1968 to 1971, interpretation of sulphur, oxygen and hydrogen isotope ratios, trace element abundances and fluid inclusion data indicated that the aqueous component of the exhalations was largely seawater (Stanton, 1991). Corliss (1971) stated that most of the ore metals tended to concentrate in the residual glass rather than in the crystalline phases of sea-floor basalts and that such glass would be highly susceptible to hydrothermal leaching. He continued to propose that heated seawater, percolating through the sub-sea-floor basaltic pile, leached the metals from the glassy phases, transported them as chloride complexes and conveyed them to the sea-floor, where the solutions appeared as hot springs from which the metals precipitated as sulphide sinters. At that time, volcanism was not thought to play any active role in the generation of the ore solutions, but it simply provided the accumulated basaltic lavas of the ocean floors, which long after their emission, became passive providers of metals for leaching (Stanton, 1991).

Sawkins (1986) proposed that the evidence from isotopic and other geochemical data together with experimental results on hydrothermal rock interaction, indicate a process more intricate than seawater leaching alone. He put forward a hybrid process which involves magmatic derivation of the metals, followed by mixing with convecting seawater at higher levels in the sub-sea-floor basaltic pile and proposed the operation of relatively long-lived seawater convection systems on which are superimposed short-term additions or pulses of metal-rich late magmatic fluids.

Stanton (1991) found that nickel occurs at about 100 parts per million in the pyroxenes of fresh sea-floor basalts from an intraoceanic arc, and at 65 parts per million in the amphiboles, with a range of 100 to 3 500 parts per million in the olivines. Lead occurs at between 2 to 3 parts per million in most ocean-floor and arc basalts and is incorporated in significant amounts only where these basalts contain potassium. Copper is incorporated at about 30 parts per million in all major silicates. Copper reaches a peak abundance when plotted against SiO_2 at around 52 to 55 percent SiO_2 and then fall away again at higher percentages of SiO_2 (Stanton, 1991).

Stanton (1991) used, amongst others, nickel, lead and copper to indicated how magmatic crystallisation and differentiation may affect the availability and the timing of availability of the different metals for exhalation and ore formation. Nickel is subtracted from the melt by olivine very early in the process of lava evolution and hence is rarely available for exhalation, and then only in association with the most mafic phases. Lead is incorporated in significant quantities only in potassium-rich phases, and hence, in the absence of K feldspar, is almost entirely segregated into the late-stage melt. Copper builds rapidly from a high base to reach concentrations in the melt on the order of 2 000 parts per million by the time basaltic andesites begin to develop. Copper is therefore present in high concentrations to enter the volatile phase and exhale relatively early in the evolution of the lava series (Stanton, 1991).

4.1.2 Source of heat

The nature of the heat source responsible for driving the submarine hydrothermal systems is still an unresolved problem according to Slack (1993).

In his review, Slack (1993) mentioned that most authors invoke a basaltic magma at depth, which is well documented beneath actively forming massive sulphide deposits on modern spreading ridges. Large mafic intrusions are generally unknown in the vicinity of the Japanese Besshi-type deposits and the volume of exposed metabasaltic rocks, which are mainly flows and/or tuffs rather than sills, are clearly too small to have served as the heat sources for the mineralising systems (Campbell *et al.*, 1981; Lydon, 1988). Kunugiza *et al.* (1986) and Takasu (1989) reported that the large metagabbroic intrusions present near some deposits in the Besshi district, were tectonically emplaced after the formation of the deposits.

Metadiabase bodies, which are volumetrically minor and which are widely interpreted as sills derived from basaltic magma at depth, are quite common in the area. These metadiabase/metabasalt sills, as well as dykes of similar composition, could not have provided adequate heat to sustain the hydrothermal systems that formed the sulphide deposits, when considering the heat flow considerations from modern analogues in Guaymas Basin and Escanaba Trough (Koski *et al.*, 1994). The required heat in these deposits is generated by basaltic magmas at the axis of the spreading ridges where new oceanic crust is forming and the sediments overlying the magmas are acting as a thermal blanket, preventing sudden heat loss to the overlying seawater. The absence of remnant oceanic crust in nearly all of the Japanese Besshi-type deposits does not require a different thermal model, since these sulphide-bearing sedimentary sequences could have been detached from their oceanic substrate during continental accretion (Slack, 1993). Miyake (1988) suggested such a process for the Shimokawa deposits in Hokkaido, where vestiges of a sheeted diabase dyke system are present.

Slack (1993) pointed out that where no remnants of oceanic crust are known to occur near sediment-hosted Besshi-type deposits, mineralisation may have occurred by means of downward-penetrating convection in which heat and hydrothermal fluids are generated during crustal extension and transmitted to surface along synsedimentary growth faults (Russell, 1983; Strens *et al.*, 1987).

4.1.3 Generation of ore fluids

It is widely excepted that the hydrothermal fluid that emanates from modern submarine spreading ridges is seawater which has reacted with mostly basaltic rocks as it is circulated through oceanic crust by a convection cell.

The chemical evolution of the seawater (Table 6), as it is heated in the presence of basalt, is dominated in the early stages by the removal of sulphate, which precipitates as anhydrite due to the temperature increase. The major cationic exchange is dominated by the loss of magnesium to the basalt in exchange for calcium. Magnesium is also removed above 250 degrees Celsius as magnesium hydroxysulphate-hydrate (MHS) causing high acidity in the fluid. High acidity may also be created by epidote formation and by sodium fixation during retrograde cooling when silica activities are high. Potassium is lost to basalt at low temperatures, but leached from it above 150 degrees Celsius. The result is a sulphurous Na-Ca-Cl fluid containing several parts per million concentrations of Fe, Mn, Zn and Cu (Table 6) (Lydon, 1988).

The elevated salinities of ore fluids generated within oceanic crust can be explained by a variety of processes. Rock hydration reactions by consuming water will concentrate chloride in the pore fluid i.e. a rock of 4 percent porosity absorbing 4 weight percent water by clay alteration would increase the salinity of the pore fluid by a factor of 2.7 (Cathles, 1983). Phase separation of the NaCl-H₂O fluid under high temperature and pressure conditions produces a brine with salinities two to three times that of seawater. This chloride brine may be formed and discharged during active hydrothermal venting or may collect in the sub-surface to

form a brine reservoir that is tapped later (Von Damm and Bischoff, 1987). Ore fluids of massive sulphide deposits associated with sedimentary rocks could attain high salinities by the dissolution of evaporites, as in the case of the Red Sea brines or by a process of ion filtration by argillaceous lithologies. A summary of the reactions that could take place in the upper layers of the sub-sea sediment pile is given by Goodfellow and Blaise (1988) (Fig. 24).

Table 6 The composition (in milligram/gram) of some natural metalliferous hydrothermal solutions compared with modern seawater (After Lydon, 1988).

	Red Sea	Kuroko	SJFR*	Seawater
Temperature(°C)	60	320	224	2
pH	5.5	4.5	3.2	8
Na	92 600	17 500	18 300	10 790
K	1 870	5 000	2 020	395
Ca	5 150	4 400	3 860	413
Mg	764	510	<1	1 280
SiO ₂	60	?	1 400	10
Cl	156 000	40 400	38 640	19 355
SO ₄	840	?	-50	2 745
H ₂ S	?	?	63	<1
ΣCO ₂	140	8 800	?	103
Fe	81	6	1 045	<1
Mn	82	?	197	<1
Zn	5.4	3	59	<1
Cu	0.3	5	<0.1	<1
Pb	?	3	?	<1
Ba	0.9	?	?	<1

Note: * - Southern Juan de Fuca Ridge

4.1.4 Contents of ore fluids

The leaching of trace metals during interaction of the hydrothermal fluids with the sub-sea-floor rock column is still thought to be a major component of trace metal supply to the ore-bearing fluid.

Trace metals occur in rocks as either labile or bound components. Labile components are those which are loosely bound at the surfaces of mineral grains or those held by a grain coating of a material with high adsorption capacities. Bound components occur as ionic substitutions for essential elements within the crystal structures of rock-forming minerals (Lydon, 1988). The largest proportions of trace elements in most lithologies occur as bound components and can only be leached during host mineral destruction i.e. alteration into secondary minerals.

The conversion of basalt into a chlorite - epidote - actinolite - albite - quartz assemblage (typical greenschist) requires only hydration. The reactive component of the solution is the water itself and does not depend upon the supply of dissolved components. The reaction could proceed to completion at a water : rock ratio of 1 : 30 and result in more than 1 500 parts per million metal in the pore fluid if 50 parts per million metal were leached from the basalt (Fig. 38 and Table 7).

The most favourable circumstance for leaching the bound metal component of a rock and reaching maximum concentrations of the trace metals in the pore fluids, is achieved during thermal or barometric metamorphism of a stratal aquifer. The mineral-destructive reactions are transforming one mineral assemblage into another due to changes in temperature and/or pressure, allowing the bound trace metals to partition into the pore fluid. Pore fluids in a geopressured stratal aquifer are not continually replaced, leaching will continue until the pore fluid becomes saturated with respect to the metals (Lydon, 1988). Both the reactive capacity and leaching ability of a solution increases with increasing salinity. The reactive capability of a pore solution is largely due to the

cationic concentration and the leaching ability, *via* metal complexing, is due to the anionic concentration.

The Cu-Zn-type association, of which Besshi-type deposits fall under, may be explained if the metals were transported mainly as bisulphide complexes, thereby suppressing the mobility of Pb (Fig. 39). The solubilities of both chalcopyrite and sphalerite as bisulphide complexes are about 100 times greater than that of galena under comparable conditions (Lydon, 1988).

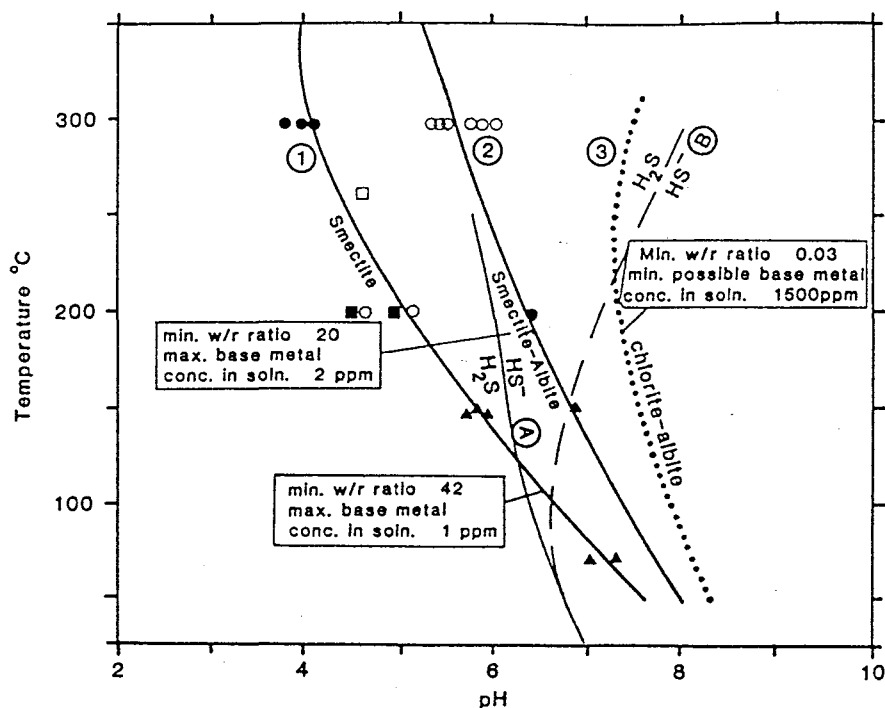


Fig. 38 Summary of experimental data on seawater-basalt interaction indicating quenched pH and alteration mineral assemblages for experiments of more than 14 days duration (filled squares, circles and triangles and circles from various authors, after Lydon, 1988). Bars join pH measurements at the end of 14 days and at the end of the run. Curve for chlorite - albite - (+ quartz) calculated assuming 1 molar NaCl and an activity of Mg^{2+} equal to the smectite-albite stability at the same temperature. Reactions, mineral compositions and leached metal concentrations are as indicated in Table 7 (After Lydon, 1988).

Table 7 Some reactions involving the alteration of basalt to hydrous secondary mineral assemblages. The concentration of trace metal in the product solution is calculated assuming that a total of 50 parts per million trace metal is lost from the weight of basalt which reacts with the solution. The minimum seawater to basalt ratio is calculated from the weight of seawater (containing 1 350 parts per million Mg) required to supply the weight of magnesium and/or water required to supply the reactants (After Lydon, 1988).

REACTION 1 $0.076 \text{ Basalt} + 0.963 \text{ Mg}^{2+} + 2.212 \text{ H}_2\text{O} + 0.84 \text{ O}_2 \rightarrow 1.0 \text{ Smectite} + 0.785 \text{ Ca}^{2+} + 0.202 \text{ Na}^+ + 0.024 \text{ Fe}^{2+} + 0.088 \text{ H}^+ + 0.587 \text{ H}_4\text{SiO}_4$ Minimum seawater : basalt ratio = 42 : 1 <1 parts per million in product solution
REACTION 2 $0.104 \text{ Basalt} + 0.663 \text{ Mg}^{2+} + 1.781 \text{ H}^+ + 1.187 \text{ H}_2\text{O} + 0.149 \text{ O}_2 \rightarrow 1.0 \text{ Smectite} + 0.50 \text{ Albite} + 1.260 \text{ Ca}^{2+} + 0.326 \text{ Fe}^{2+} + 0.525 \text{ H}_4\text{SiO}_4$ Minimum seawater : basalt ratio = 20 : 1 <2 parts per million in product solution
REACTION 3 $1.0 \text{ Basalt} + 0.009 \text{ Mg}^{2+} + 10.379 \text{ H}_2\text{O} + 0.179 \text{ O}_2 \rightarrow 4.90 \text{ Smectite} + 1.604 \text{ Chlorite} + 2.075 \text{ Actinolite} + 3.776 \text{ Epidote} + 1.893 \text{ Quartz} + 0.009 \text{ Ca}^{2+}$ Minimum seawater : basalt ratio = 1 : 30 >1 500 parts per million in product solution
MINERAL FORMULAE Smectite: $(\text{Na}_{0.17}\text{Ca}_{0.09}\text{Mg}_{1.73}\text{Fe}^{2+}_{0.32}\text{Fe}^{3+}_{0.43}\text{Al}_{0.39})\text{Al}_{0.90}\text{Si}_{3.10}\text{O}_{10}(\text{OH})_2$ Chlorite: $(\text{Mg}_{0.51}\text{Fe}_{0.49})_5\text{Al}_2\text{Si}_3\text{O}_{10}(\text{OH})_8$ Actinolite: $\text{Ca}_{2.0}(\text{Mg}_{0.60}\text{Fe}_{0.4})_5\text{Si}_8\text{O}_{22}(\text{OH})_2$ Epidote: $\text{Ca}_{2.0}(\text{Fe}_{0.60}\text{Al}_{2.4})\text{Si}_3\text{O}_{12}(\text{OH})$ Albite: $\text{NaAlSi}_3\text{O}_8$ Basalt: $\text{Na}_{4.9}(\text{Ca}_{11.71}\text{Mg}_{10.27}\text{Fe}^{2+}_{8.83}\text{Fe}^{3+}_{1.55})\text{Al}_{17.18}\text{Si}_{49.33}\text{O}_{160}$

Volcanogenic massive sulphide deposits, other than Besshi-type deposits, exhibit a pronounced temperature dependent Cu-Zn zonation which requires precipitation from a solution in which metal chloride complexes are dominant (Fig. 40). Since Besshi-type deposits do not exhibit a pronounced Cu-Zn zonation, the contribution of a chloride complexed fluid is envisaged as minor since Lydon (1988) reported that a previous worker found the progressive desulphidisation of a fluid above 200 degrees Celsius would cause precipitation of copper as chalcopyrite, while zinc remains as a zinc chloride complex (compare Figs. 39 and 40). Sphalerite would only precipitate on subsequent cooling of the solution, indicating a temperature dependant zoning which Besshi-type deposits do not possess.

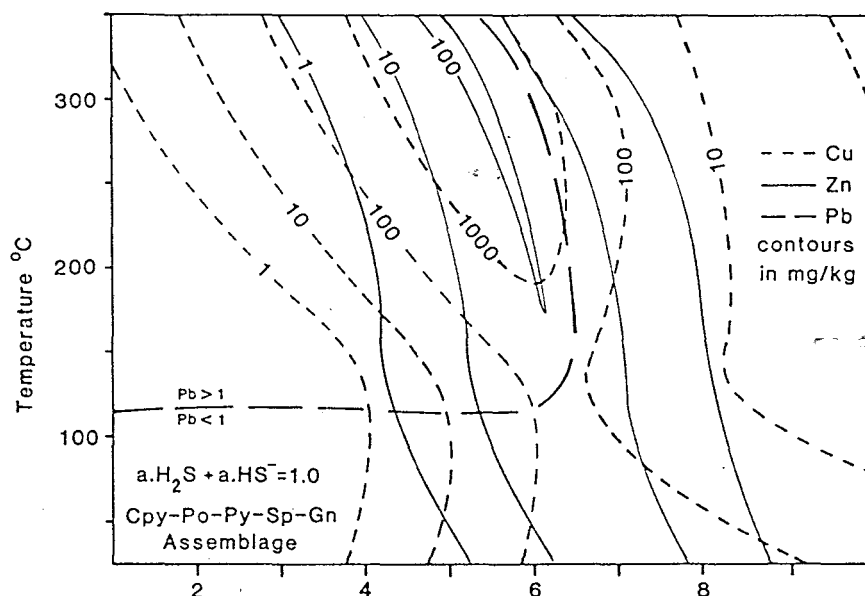


Fig. 39 Sulphide solubilities as bisulphide complexes. Calculated metal concentrations (in milligram/kg) in a solution in which the total sulphur concentration is 1 molal and which is in equilibrium with a chalcopyrite - pyrite - pyrrhotite - sphalerite - galena assemblage (After Lydon, 1988).

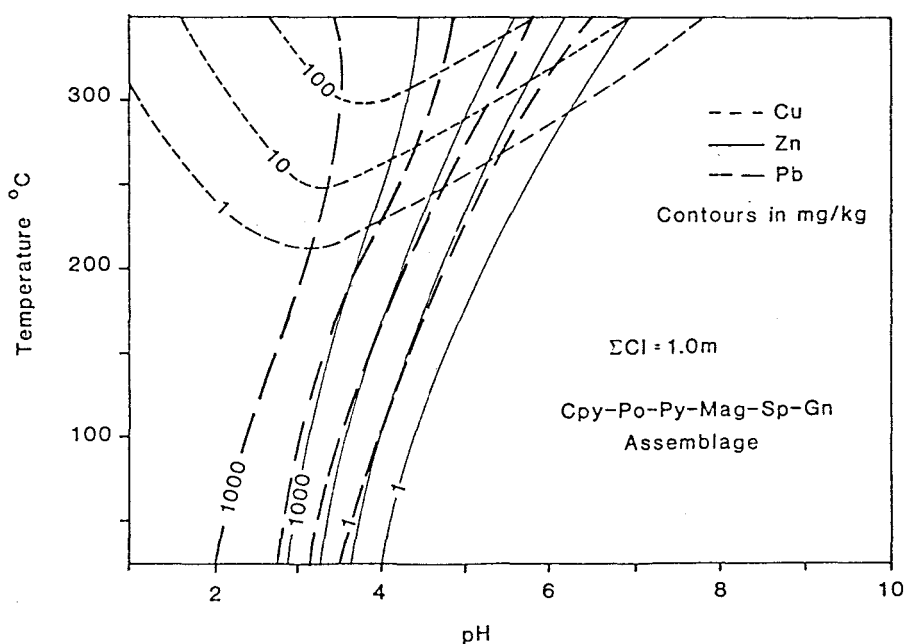


Fig. 40 Sulphide solubility patterns due to chloride complexing. Calculated total metal concentrations (in milligram/kg) in a 1 molar NaCl solution in equilibrium with a chalcopyrite - pyrite - pyrrhotite - sphalerite - galena assemblage. Note that the sulphur content of the solution ranges from 1 milligram/kg at low pH and low temperature to more than 10 000 milligram/kg at high pH and high temperature. The isothermal decrease in chalcopyrite solubility with decreasing pH at low pH is due to the decrease in activity of Cl^- as the relative proportion of chloride complexed with metal increases (After Lydon, 1988).

4.1.5 Types of fluids

Sato (1972) postulated that a hydrothermal solution discharged into seawater would behave in one of three ways, depending on its initial temperature and density and its subsequent degree of mixing with seawater (Fig. 41): Type I is a highly saline hydrothermal solution whose density is greater than cold seawater at all degrees of mixing. Type II is a solution which is initially less dense than seawater, but at some stage in the mixing process pass through a maximum density which is greater than that of cold water. Type III is a solution that is initially less dense than cold seawater and which remain buoyant at all degrees of mixing.

Before the discovery of the mound-chimney sulphide deposits on the East Pacific Rise, mechanisms by which sulphides accumulated were most frequently pictured as analogous to those for the Red Sea brine pool metalliferous sediments or, were based on the buoyancy models of Sato (1972), whereby sulphides were precipitated from the hydrothermal fluids after discharge from the vent and accumulated as heavy gelatinous muds. In contrast, studies of modern mound-chimney sulphide deposits indicated that sulphides accumulate as rigid edifices.

As seen in previous sections, the temperature of the hydrothermal fluids measured at the recent sea-floor deposits ranges from 400 degrees Celsius to just a few degrees above ambient seawater. The highest temperature fluids are considered to be pristine hydrothermal fluids, namely fluids which have equilibrated with reservoir rocks at pressure-temperature conditions close to the maximum values of the hydrothermal system, whereas those at lower temperatures are usually mixtures of pristine hydrothermal fluid and pore water (mainly seawater) that had become entrained into the hydrothermal conduit in the subsurface (Lydon, 1988).

These hydrothermal fluids (Table 6) have salinities up to twice that of modern seawater (3.2 weight percentage NaCl), and as predicted by Sato (1972) for his Type III solutions, form buoyant plumes on discharge.

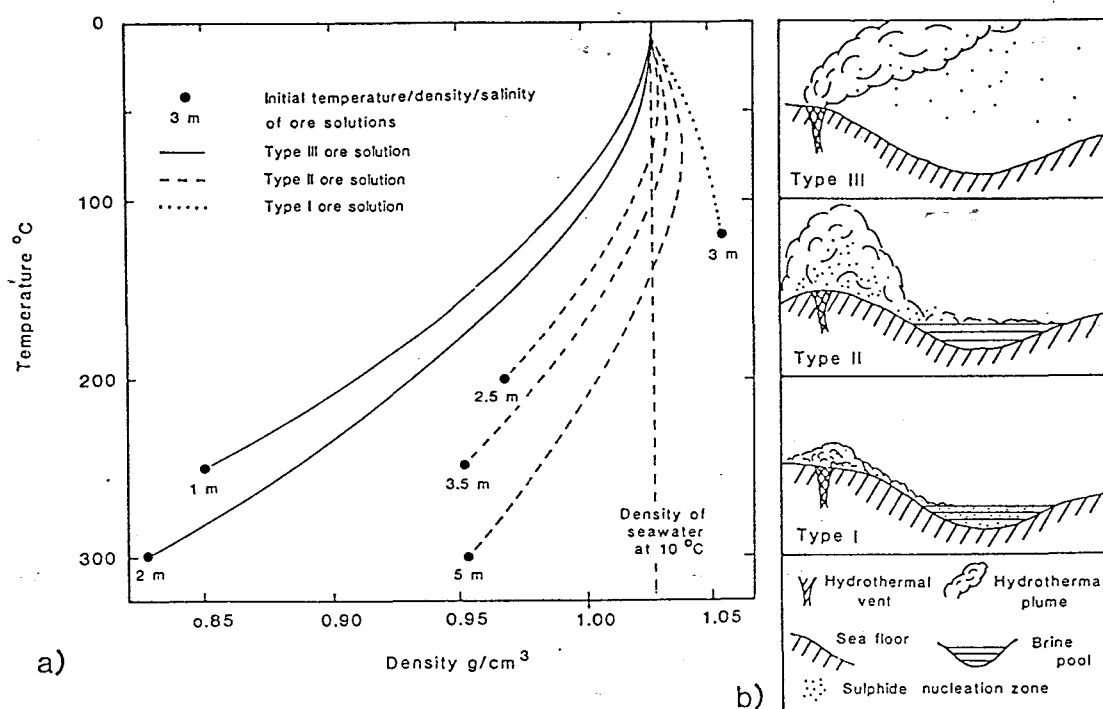


Fig. 41 (a) Density - temperature relationships of hot NaCl solutions mixed with seawater at 10 degrees Celsius illustrating the definition of Sato's (1972) three types of hydrothermal solutions.

(b) Schematic representation of the buoyancy behaviour of, and sulphide precipitation from Sato's (1972) three types of hydrothermal solutions (After Lydon, 1988).

Red Sea brines, with a salinity of seven times that of seawater, discharge at temperatures of 210 degrees Celsius (Pottorf and Barnes, 1983) and like the Type I solutions, collect as brine pools in the deepest parts of the axial valley. Turner and Campbell (1987) found that all high temperature (>300 degrees Celsius) fluids will be initially buoyant in seawater, but if their salinities are greater than about 7 weight percentage NaCl will show reversing buoyancy on mixing with seawater.

All the vent fluids in the Eastern Pacific, as discussed in earlier sections, are of Sato's Type III solutions. Several authors, from visual observations and fluid inclusions, suggest that fluids with

salinities high enough to form reversing plumes like Sato's Type II solutions do in fact exist in the modern oceanic crust (Lydon, 1988).

4.1.6 Sulphide accumulation

The literature places considerable emphasis on the stratiform and commonly sheet-like morphology of Besshi-type deposits and on their general lack of recognisable feeder zones. Although post-ore deformation has influenced the geometry of many deposits, some have preserved primary morphologic features.

The sheet-like shape of deposits (Matchless, Namibia; Gossan Lead, United States) suggests sulphide formation by exhalative or syngedimentary processes. Exhalative mineralisation requires precipitation of sulphide and most gangue minerals directly from venting hydrothermal fluids, either in the vicinity of black smoker chimneys or in a laterally extensive brine pool. Slack (1993) envisaged a black smoker model to be a problem since to date no sheet-like massive sulphide deposits have been discovered in modern settings near submarine black smoker fields probably because high-temperature, low-salinity black smokers disperse most of their metal into submarine plumes that are rapidly diluted by seawater. A buoyant or near-neutral density plume might produce sheet-like Besshi-type massive sulphides if the plume was sustained for tens or hundreds of years and was sufficiently large to yield widespread sulphide precipitation. Anoxic bottom waters would enhance this process and may have played a critical role in the formation of some Besshi-type deposits (Slack, 1993). A brine-pool for the sheet-like deposits is more favoured since it allows for relatively uniform precipitation of sulphides over a large area. This model is further supported by the modern analogues recognised in the Red Sea, where massive copper-zinc sulphides are currently precipitating from aerially extensive hot brines in an intracontinental rift setting similar to that postulated for many ancient Besshi-type deposits. The apparent lack of feeder zones in most sheet-like deposits, is consistent with a Red Sea analogue, where metalliferous saline brines are believed to discharge from scattered epigenetic vents

(Zierenberg and Shanks, 1983) rather than from a single pervasive discharge zone (Slack, 1993).

There are Besshi-type deposits which consist of lenses of massive sulphide which lack a sheet-like morphology. A few of these deposits also have well-developed footwall feeder zones, such as Windy Craggy, Canada (Peter and Scott, 1990). Fluid-inclusion studies of the Windy Craggy deposit indicate that the temperatures and salinities of the hydrothermal fluids are inconsistent with sulphide formation in a dense brine pool (Slack, 1993). These results suggest that the Windy Craggy deposit formed largely by exhalative processes or possibly by selective replacement of permeable sediments and/or volcanic rocks in the upper parts of a submarine hydrothermal system (Slack, 1993).

4.1.7 Palaeotectonic setting

Several authors have proposed a variety of paleotectonic settings for the origin of Besshi-type deposits. These include convergent plate settings on the flanks of island-arc volcanoes (Mitchell and Bell, 1973; Mitchell and Garson, 1976), fore-arc/trench environment (Hutchinson, 1980), extensional epicontinental or back-arc settings (Fox, 1984), rifted cratonic margins or fore arcs (Scott, 1985) and intracontinental rifts floored by oceanic crust (Gair and Slack, 1984; Breitskopf and Maiden, 1988; Misra and Lawson, 1988). These models are based on the regional geological settings of specific deposits, on the inferred protoliths of associated clastic metasedimentary rocks and on the geochemistry of mafic metavolcanic rocks/metabasalts (Slack, 1993).

The vast majority of these metabasalts have a geochemical signature of oceanic tholeiites or mid-oceanic-ridge basalts (Sugisaki *et al.*, 1972; Gair and Slack, 1984; Misra and Lawson, 1988). Some deposits have associated amphibolites with alkalic, within-plate basalts, such as some of the Besshi deposits in Japan (Sugisaki *et al.*, 1972; Banno and Sakai, 1989), Windy Craggy, Canada (Peter and Scott, 1990), the western Matchless deposits, Namibia (Breitskopf and Maiden, 1987, 1988) and Pirén Alto, Chile (Schira *et al.*, 1990).

Besshi-type deposits are not unique to any specific paleotectonic setting (Koski, 1990). Slack (1993) proposed that based on probable modern analogues such as Escanaba Trough, Guaymas Basin and the Red Sea, ancient Besshi-type deposits possibly formed in several different environments which include intracontinental rifts, rifted continental margins, oceanic spreading ridges near continental landmasses and back-arc basins. Koski (1990) suggested a setting involving a spreading ridge that is being consumed by a subduction zone.

Critical factors in each setting are the presence of spreading ridges as sources of basalts and heat needed to drive the hydrothermal systems together with the proximity of subaerial landmasses as a source of clastic detritus (Slack, 1993).

4.1.8 Isotopic characteristics

Isotope analyses provide data pertaining to the source of the metals and the conditions during mineralisation. Adequate information exists for sulphur, lead and boron isotopes, while limited data are available for stable (carbon and oxygen) and radiogenic isotopes (strontium).

Sulphur isotope analyses were conducted on several Besshi-type deposits in Japan (Yamamoto *et al.*, 1983, 1984a, b); Ducktown, Tennessee (LeHuray, 1984); Elizabeth, Vermont; the Gossan Lead, Virginia; Mount Molloy, Dianne and O.K., Australia (Gregory and Robinson, 1984); Kanmantoo, Australia (Seccombe *et al.*, 1985); Killingdal, Norway and the Beatson and related deposits in the Prince William Sound region, Alaska (Crowe *et al.*, 1992). A total range of $\delta^{34}\text{S}$ values from -2 to +12 per mil was obtained. Such values suggest derivation of sulphur from the following possible sources: basaltic sulphide with $\delta^{34}\text{S}$ values near 0 per mil; sedimentary bacteriogenic sulphide depleted of ^{34}S and ^{34}S -enriched sulphide originating from the reduction of seawater sulphate, by either inorganic or organic processes (Slack, 1993).

Lead isotope data obtained span a relatively large range for $\delta^{206}\text{Pb}/\delta^{204}\text{Pb}$ 17.47 - 19.38; $\delta^{207}\text{Pb}/\delta^{204}\text{Pb}$ 15.39 - 15.64; $\delta^{208}\text{Pb}/\delta^{204}\text{Pb}$ 37.25 - 38.78 from analyses conducted by Sato and Sasaki (1976), LeHuray (1984), Fox *et al.* (1988) and Peter and Scott (1990). This range of data spans from below the mantle growth curve to above the orogene growth curve of Doe and Zartman (1979) and therefore the lead in the ores was derived from a mixture of radiogenic and unradiogenic sources. The dominant radiogenic source in most of the deposits is the surrounding clastic metasedimentary rocks, from which lead was leached during submarine hydrothermal alteration (Fox, 1984; LeHuray, 1984; Fox *et al.*, 1988). Based on the similarities in lead isotope composition between Besshi-type deposits and mid-oceanic-ridge basalts, several authors proposed that the source of unradiogenic lead was basalt and/or diabase (Sato and Sasaki, 1980), Fox *et al.*, 1988). Slack (1993) pointed out that not all the unradiogenic lead in Besshi-type deposits is necessarily from basalt. Granulite-facies rocks usually have low uranium/lead ratios and unradiogenic lead isotope compositions similar to those of mid-oceanic-ridge basalts (Zartman and Doe, 1981). Detrital feldspars within the footwall sedimentary sequences of Besshi-type deposits that originally were derived from continental granulite-facies terranes could have been the source for some of the lead in the ores, and may have contributed to the unradiogenic lead isotope composition observed for several deposits.

Tourmaline occurs as an accessory gangue mineral in many dominantly sediment-hosted, volcanogenic massive sulphide deposits (Slack, 1982; Taylor and Slack, 1984) together with the Shimokawa and Sazare deposits of Japan (Slack, 1993). Palmer and Slack (1989) conducted a worldwide survey of the boron isotope compositions of tourmaline from nine Besshi-type deposits. The $\delta^{11}\text{B}$ values were found to range from -15.4 per mil to -2.8 per mil which suggest that the boron in the tourmaline was derived from footwall siliclastics and related crustal rocks. In contrast, local tourmaline concentrations in the wall rocks of the Cherokee Mine, Ducktown district, have $\delta^{11}\text{B}$ values of +3.2 per mil which implies a component of marine-evaporite boron present in the tourmaline (Palmer and Slack, 1989).

4.2 Descriptive model

A descriptive model embodies those features of the geological settings, morphology, geochemistry, ore mineralogy, zoning (if any), judged to be unique or characteristic of the deposit type (Lydon, 1984). Table 8 contains descriptive features which are very similar to the models of Cox (1986), Koski (1990) and Maiden (1993a).

Key diagnostic features which should be considered essential in planning an exploration programme would include a general sheet-like morphology within clastic marine sediments and minor mafic volcanic rocks of Proterozoic to early Tertiary age; sulphide mineralisation that consists predominantly of pyrite and/or pyrrhotite with minor chalcopyrite, sphalerite and very little galena; relatively high cobalt contents and high cobalt/nickel ratios.

Table 8 Descriptive features of Besshi-type massive sulphide deposits (After Slack, 1993).

Definition	Strata-bound cupriferous pyrite/pyrrhotite deposits hosted by mixtures of clastic sedimentary rocks and minor tholeiitic basalt
Synonym	Kieslager-type deposit
Commodities	Cu, S, Fe, Zn, Ag, Au
Important examples	Besshi, Shimokawa, Japan; Gossan Lead, USA; Windy Craggy, Canada; Matchless, Namibia; Gairloch, Scotland; Tverfjell, Norway; Kilembe, Uganda; Saladipura, India; Arintei, Spain; Liwu, China; Kanmantoo, Australia; Kizil-Dere, Georgia; Tisova, Czech Republic; Rapu-Rapu, Philippines; Hermioni, Greece; Altin-Tepe, Romania; Gjegjan, Albania; Walchen, Austria; Pirén Alto, Chile
Importance	Largest deposits (Ducktown and Windy Craggy) have resources of >1 Mt of contained copper
Grades and tonnages	Median deposit contains 1.3 Mt of 1.43 percent Cu, generally with significant zinc, silver and some gold
Age	Early Proterozoic to early Tertiary, with largest number late Proterozoic, early Palaeozoic or Mesozoic
Geological setting	Rift basins containing clastic sedimentary rocks and generally minor tholeiitic basalt as flows, tuffs and sills, generally with associated strata-bound alteration zones and/or chemical sediments
Host rocks	Pelitic schists, metagraywacke, quartzite, chlorite-rich schist, muscovite-rich schist; local metachert, magnetite iron-formation, coticule, albitite and tourmaline
Structure and texture	Laminated, banded or massive fine-grained sulphides; highly deformed deposits characterised by breccia locally; some deposits have footwall stringer zones
Ore mineralogy	Pyrite + pyrrhotite + chalcopyrite ± sphalerite ± magnetite ± galena ± arsenopyrite ± cobaltite ± tetrahedrite ± cubanite ± molybdenite
Gangue minerals	Quartz + muscovite ± carbonate ± albitite ± chlorite ± amphibole ± tourmaline
Palaeotectonic setting	Not unique; includes rifted continental margins, intracontinental rifts, back-arc basins and sediment-covered spreading ridges proximal to continental land masses
Related deposits	Transitional between Cyprus- and Kuroko-type massive sulphide deposits

5 CONCLUSIONS

Besshi-type deposits constitute an important class of stratiform ores in many orogenic belts of the world. These deposits are dominantly sediment-hosted and contain significant copper and in some cases zinc, silver, gold and/or cobalt. Important examples include Besshi and Shimokawa, Japan; Gossan Lead, USA; Windy Craggy, Canada; Matchless, Namibia; Gairloch, Scotland; Tverfjell, Norway; Kilembe, Uganda; Saladipura, India; Arintei, Spain; Liwu, China; Kanmantoo, Australia; Kizil-Dere, Georgia; Tisova, Czech Republic; Rapu-Rapu, Philippines; Hermioni, Greece; Altin-Tepe, Romania; Gjogjan, Albania; Walchen, Austria and Pirén Alto, Chile.

Besshi-type deposits range in age from early Proterozoic to early Tertiary, of which the largest number are late Proterozoic, early Palaeozoic or Mesozoic in age. No Archaean examples of Besshi-type deposits are known.

Besshi-type deposits are characterised by a blanket- or sheet-like morphology in which the dominant sulphide minerals are pyrite and/or pyrrhotite. Chalcopyrite is predominantly the base metal sulphide and sphalerite may be abundant locally while galena and arsenopyrite are rare. Besshi-type massive sulphide deposits contain important copper resources. The largest deposits are in the size range of 100 - 300 million tonnes with average grades of approximately 1 - 3 percent copper.

Besshi-type massive sulphide deposits are hosted within sequences of clastic metasedimentary rocks and intercalated tholeiitic metabasalt. Felsic metavolcanic rocks are generally absent or volumetrically minor. The amounts of metabasalt are highly variable and gradations exist from largely sediment-hosted (Ducktown) to deposits with abundant associated metabasalt (Windy Craggy). The sediment-dominated deposits may have formed within intracratonic rifts or along rifted continental margins where continental masses provided the source for abundant sedimentary detritus. The basalt-dominated deposits likely formed on or near oceanic spreading ridges either in continental-margin or back-arc basin settings. The

lack of associated felsic volcanic rocks in the deposits suggests settings which are distant from volcanic arcs.

It is suggested that Besshi-type deposits form by both exhalative and syngenetic replacement processes when considering geological features and comparisons with modern analogues in the Guaymas Basin, Middle Valley and Escanaba Trough. The currently forming metalliferous sediments in the Red Sea provide for a brine pool model explaining the lack of footwall feeder zones below sheet-like deposits. Where thick sulphide lenses are contained in some Besshi-type deposits, combinations of exhalative precipitation and sub-sea-floor replacement of permeable sediments and/or volcanic rocks, take place in the upper parts of submarine hydrothermal systems.

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