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Volcanology of the Archaean Lunnon Basalt and its relevance to nickel sulfide-bearing trough structures at Kambalda, Western Australia

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The Lunnon Basalt is an Archaean, subaqueous, tholeiitic metabasalt succession, with a minimum inferred thickness of 1750 m. It forms the oldest exposed stratigraphic unit at Kambalda in the nickel sulfide-rich Norseman-Wiluna Greenstone Belt, Western Australia, and is dominated by variable proportions of massive basalt, pillow basalt, and basalt breccia. These facies form intimately interlayered massive, pillow and complex lava flow units, with an average thickness of about 20 m. The stratigraphy of the basalt can be subdivided into a ‘lower’ MgO-rich member and an ‘upper’ less MgO-rich member, these being separated by a sedimentary horizon. Of the possible palaeovolcanic and tectonic settings (layer 2 of oceanic crust, a large shield volcano, or a tectonically-topographically ponded sea-floor lava field), the Lunnon Basalt appears to be the remains of a ponded (rift) lava succession. The presence of inherited zircon xenocrysts indicates that the subjacent crust at the time of formation was sialic, Archaean crust. Komatiite-associated nickel sulfide is normally localised in or over depressions in the upper surface of the Lunnon Basalt. The origin of these depressions or ‘troughs’ is uncertain. The four most-favoured models are: synvolcanic faulting; original volcanic topography; thermal erosion; and structural. Most depressions have faulted margins but there is very little field evidence to support synvolcanic faulting. Furthermore, physical evidence for thermal erosion is non-existent in the exposures seen. Palaeoflow indicators from the pillow basalts and basalt breccias indicate flow approximately parallel to the major depressions, suggesting a case for volcanic topography. Since significant relief occurs on the surfaces of all modern lavas, it is therefore clear that the surface of the Lunnon Basalt must have had depressions which may have controlled the behaviour of the succeeding komatiites, and therefore perhaps the location of the nickel sulfide mineralisation. However, the depressions confining the nickel sulfide mineralisation at Kambalda are extremely complex and variable in nature, and their present form is most likely a result of several processes.

Key words: Archaean, basalt, Kambalda, nickel sulfides, palaeovolcanology.

INTRODUCTION

The Archaean Lunnon Basalt is the oldest known stratigraphic unit in the Kambalda area of the Eastern Goldfields Province, Yilgarn Craton, Western Australia (Figure 1); it is the substrate upon which nickel sulfide-bearing komatiite lavas were emplaced. Despite the immense economic importance of the Kambalda succession and the importance of the Lunnon Basalt to models for accumulation of nickel sulfides (see below), surprisingly little work has been done on the palaeovolcanology. Previous studies of the Lunnon Basalt have focused on its stratigraphic or regional setting (Halleberg & Williams 1972; Gemuts & Theron 1975; Gresham & Loftus-Hills 1981; Redman & Keays 1985; Archibald 1987; Claoué-Long et al. 1988; Clout 1990) or aspects of its geochemistry and geochronology (Redman & Keays 1985; Claoué-Long 1986; Compston et al. 1986; Cowden & Archibald 1989; Morris 1990, 1993a; Lesher & Arndt 1995).

This paper documents the geological characteristics of the Lunnon Basalt and considers its importance for mineralisation. Specific aims are: (i) assessment of the palaeoenvironmental setting; (ii) documentation of the facies; (iii) assessment of the palaeovolcanology (i.e. eruption style and flow behaviour); and (iv) determination of the relationship between the Lunnon Basalt and nickel sulfide mineralisation. The study is based on mapping 11 underground exposures in the Otter-Juan and Poster mines south of Kambalda, and logging 13 associated diamond drillcores, and the longest drillhole in the Kambalda Dome, KD 1029, amounting in total to 3027 m of drillcore (Squire 1992).

REGIONAL GEOLOGICAL SETTING

Kambalda is located within the Archaean Norseman-Wiluna Greenstone Belt, of the Eastern Goldfields Province in the Yilgarn Craton, Western Australia (Gee 1979; Gee et
al. 1981) (Figure 1). The Kambalda stratigraphic succession is subdivided into the Kalgoorlie Group (oldest), the Black Flag Group and the Merougil beds (youngest) (Woodall 1965; Gresham & Loftus-Hills 1981; Cowden & Archibald 1989). This succession correlates in large part with the stratigraphic succession around Kalgoorlie, 80 km north (Gemuts & Theron 1975; Gresham & Loftus-Hills 1981; Morris 1983b). Detailed reviews of the Kambalda region stratigraphy have been presented by Gresham and Loftus-Hills (1981), Lesher (1983), Cowden and Archibald (1989), Clout (1990) and Cowden and Roberts (1990). The Kalgoorlie Group comprises mostly komatiitic and tholeiitic rocks (Gresham & Loftus-Hills 1981).

The Lunnon Basalt is the oldest known stratigraphic unit in the Kalgoorlie Group and is exposed in the core of the Kambalda Dome although its base is not exposed. The Lunnon Basalt is overlain successively up-sequence by the Kambalda Komatiite (with its nickel sulfide mineralisation), the Devon Consols Basalt, Kapai Slate, Paringa Basalt, and the Black Flag Group felsic volcanic and sedimentary succession (Brauns 1991; Ong 1994; Hand & Cas 1996) which represents the end of greenstone volcanism. The overlying Merougil beds comprise polymictic conglomerate and arenite, and the unconformable to faulted lower contact is believed to represent a major time gap, a significant tectonic event in the source region, and a major change in palaeogeography (McCall 1969; Brauns 1991; Bader 1994). The Kalgoorlie and Black Flag Groups are cut by a pre- to syn-peak metamorphic suite of dykes, stocks and sills ranging in composition from mafic calc-alkaline lamprophyre, dolerite, diorite, granodiorite and rhyolite (Archibald 1985; Cowden & Roberts 1990; Perring et al. 1990).

At least three deformational events have been identified at Kambalda and these probably occurred progressively during a single regional phase of shortening. Clout (1990) identified three main deformation events at Kambalda, with peak metamorphism approximately synchronous with D5. However, Archibald (1985), Cowden and Archibald (1987) and Cowden and Roberts (1990) identified four main deformation events, in which peak metamorphism was approximately synchronous with D5 at about 2660 Ma. Swager and Griffin (1990) proposed that regional thrust stacking related to D5 caused repetition of the regional stratigraphy in the Kambalda–Kalgoorlie region. The succession at Kambalda has been metamorphosed to uppermost greenschist and lower amphibolite grade (Binnis et al. 1976) that has significantly overprinted the original mineralogy and textures of the volcanic and associated sedimentary rocks.

The age of the Lunnon Basalt has been debated for some time. Compston et al. (1986), conducted SHRIMP dating of xenocrystic zircons from the Lunnon Basalt and derived ages of 3427 ± 10 and 2681 ± 23 Ma. Clauwé-Long et al. (1988) reported an age of 2692 ± 4 Ma for zircon xenocrysts from the younger Kapai Slate and a 2702 ± 4 Ma age for zircons from a sedimentary horizon above the second lowest komatiite lava flow. More recently Foster et al. (1996) have defined a Re–Os isochron of 2706 ± 36 Ma for the komatiite succession using mineralised and unmineralised komatiite samples from Kambalda, Mt Keith and Perseverance. Since
the Lunnon Basalt is separated from the overlying komati-ite by a thin (~2 m) sedimentary layer it is probably about 2710 Ma. The older xenocrystic zircon date from the Lunnon Basalt therefore represents an older crustal source whereas the younger must represent an overgrowth or alteration event.

DEPOSITIONAL SETTING AND VOLCANOLOGY

The aim of this section is to describe the volcanic and sedimentary facies and interpret them in terms of the processes and conditions that operated during their emplacement.

Stratigraphy and geochemistry

The Lunnon Basalt has a known minimum extent of 500 km$^2$ from Kambalda to the Tramways and Bluebush areas to the south (Figure 1b). However, stratigraphically equivalent basalts are also known for almost 500 km along strike in the Norseman-Wiluna belt and may represent a volume of as much as 1.5 million km$^3$ (Lesher & Arndt 1995). In the absence of detailed geochronological correlations, and because of the existence of numerous structural blocks, this paper will focus on the Lunnon Basalt in the Kambalda Dome. The deepest diamond drillhole in the Kambalda Dome, KD 1029 (Figure 2), was terminated while still in the basalt at a depth of approximately 1750 m. From the limited data at depth, the Lunnon Basalt appears to be comprised of randomly distributed coherent and fragmental volcanic rocks with intercalated sedimentary horizons.

The Lunnon Basalt appears to have been structurally emplaced over younger sequences in the St Ives and Tramways areas to the south (Cowden & Archibald 1989; Swager & Griffin 1990). However, logging of drillhole KD 1029 from the Kambalda Dome showed no evidence for thrust stacking and stratigraphic repetition. One significant shear zone occurs at a depth of 745 m and several smaller less intense shears occur elsewhere in the hole; however, the significance and sense of displacement are uncertain. In addition, the basalt is cut by a series of intrusions which may have intruded shear zones. Therefore, although there is a possibility of fault stacking and thickening, there is no definitive evidence that it has occurred in the Lunnon Basalt.

Gresham and Loftus-Hills (1981) and Redman and Keays

Figure 2 Simplified graphic log of diamond drillhole KD 1029 drilled into the Lunnon Basalt in the Kambalda Dome.
(1985) divided the Lunnon Basalt into a 'lower' high-
magnesian basalt member and an 'upper' low-magnesian
basalt member. These members are separated by a sedi-
mentary horizon. Most underground exposures studied
here are in the upper member; whereas the majority of drill-
hole KD 1028 penetrated the lower member. Geochemical
analyses by Morris (1993a) indicate similar Ti/Zr and Zr/Y ratio characteristics in the upper and lower
members, consistent with derivation from a single mantle
source. Plots of Ti/Sc versus Zr for the lower member
showed profiles with almost zero slope, which is consistent
with olivine fractionation, whereas higher Ti/Sc at a given
Zr for the upper member is consistent with clinopyroxene
fractionation (Morris 1993a). Morris (1993a) suggested that
although a uniform mantle source was likely for the
Lunnon Basalt, the lower and upper parts may not be geneti-

cally related, which is plausible because a time break is
indicated by the sedimentary horizon separating the two
members. Lesher and Arndt (1985) proposed that the
Lunnon Basalt originated from uncontaminated, slightly
fractionated melts resulting from low degrees of partial
melting of a mixture of depleted and undepleted mantle.

Whole-rock analysis of 13 samples collected from a
variety of locations and stratigraphic levels in the Lunnon
Basalt during this study revealed some minor variations in
major element abundance. In the least altered specimens
(Table 1), all from the 'lower' member of Redman and Keays
(1985), SiO2 varied from 49.6 to 52.6%, MgO from 6.8 to 7.9%,
Fe2O3 from 10.2 to 12.5%, CaO from 10.1 to 13.8%, and Al2O3
from 13.6 to 14.6%. These data are consistent with the data
of Redman and Keays (1985) who analysed 25 specimens
from the Lunnon Basalt, 12 from their high-magnesian
basalt member. These members are separated by a sedi-
mentary horizon. Most underground exposures studied
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fractionated melts resulting from low degrees of partial
melting of a mixture of depleted and undepleted mantle.

Table 1 Whole-rock XRF analyses of least-altered specimens of
Lunnon Basalt from drillhole KD 1028.

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Facies

Four main facies have been distinguished: (i) fine-grained
sulfidic sedimentary rocks; (ii) massive basalts; (iii) pillow
basalts; and (iv) basalt breccias. Several subfacies were also
defined.

FINE-GRAINED SULFIDIC SEDIMENTARY ROCKS

Description

A number of sedimentary horizons are known within the
Lunnon Basalt (called interflow sediment) and immediately
above it at the contact with the komatitites (contact sedi-
ment). They are important in assessing the palaeovol-
canological and palaeoenvironmental setting of the
Lunnon Basalt. Bavinton (1979, 1981) identified three sedi-
mentary facies in the contact sediment and interflow sedi-
mentary horizons in the komatite succession: (i) pale
siliceous sedimentary units; (ii) carbonaceous sedimentary
units; and (iii) dark-green chlorite-rich sedimentary
units. This subdivision will be used here also although no

Claoué-Long (1986) proposed that differences between the
upper and lower members were also reflected in their
mineralogy and textures. He identified tiny olivine crystals
or their pseudomorphs only in the lower member, consist-
ent with Morris' (1993a) interpretation of olivine frac-
tionation. Claoué-Long (1986) also suggested that the upper
member was marked by a subophitic texture, whereas the
lower member had widespread radite textures developed.
However, in this study subophitic textures have been
observed in both the upper and lower members. The
radite textures described by Claoué-Long (1986) were
found near pillow and flow margins, and most likely
represent varioles.
chloritic facies were encountered in this study. Bavinton (1979, 1981) provided a detailed petrological documentation of the fine-grained sulfidic sedimentary facies whereas Beckingham (1991) documented their geochemistry. Sedimentary facies are lenticular, of limited lateral persistence and usually less than 6 m thick (Bavinton 1979). There is apparently a general antithetic relationship between the occurrence of nickel ores and contact sedimentary units at Kambalda (Gresham & Loftus-Hills 1981), although recent studies (Hollamby 1995; Marchiori 1995) indicate that this is not universal.

Pale siliceous sedimentary units are hard, very fine grained, with regular banding or thin laminae, approximately 2 mm thick (Figure 3a), which in some cases are defined by thin horizons of pyrite and pyrrhotite, variations in sulfide content, microgranular grain size variations and colour. Siliceous sedimentary units occur in the 24 drill drive of the Otter-Juan mine in the 2 m thick contact sedimentary interval, and also in a thin 50 cm thick interflow horizon in the 5 footwall drive. The non-sulfide bands comprise more than 50% of the facies, they range in colour from pale to medium grey and have a distinctive cherty appearance, though the chemical composition is not that of a true chert (Beckingham 1991). Rare low-angle cross-laminations, small-scale scourings and microfaulting were observed.

Carbonaceous units are best exposed in the 2 m thick contact sedimentary horizon in the 1210 incline of the Otter-Juan Mine (Figure 3b, c). They are very fine-grained (microcrystalline to silt), dull-black pelites with sulfide banding, planar lamination and thin bedding which is defined by alternating carbon-rich black layers and carbon-poor, very dark grey layers, highlighted in places by secondary carbonate along lamination planes (Figure 3b, c). The thickness of facies intervals ranges from centimetres up to metres, and they are generally of limited lateral persistence. Undulating laminae were observed mantling surface irregularities in the underlying basalts (Figure 3c) and rare low-angle cross-laminations and scour truncations were also observed (Figure 3b).

Interpretation

The uppermost greenschist–lower amphibolite facies metamorphism (Binns et al. 1976) hinders interpretation of the original sedimentary rock types. Bavinton (1979) pointed out that the mineralogies and chemistries are unlike any modern rock types. Petrologic and geochemical interpretation by Bavinton (1979, 1981) and Beckingham (1991) suggested the sediments were a mixture of chemical (exhalative and non-exhalative) and very fine detrital material, in varying ratios. The chloritic facies was thought to include significant komatiitic and/or basaltic detritus, whereas the siliceous and carbonaceous facies were considered to be dominated by distantly derived fine terrigenous detritus (Bavinton 1979, 1981).

The very fine grainsize, thin laminae, and rare cross-laminae are indicative of a very low energy, subaqueous environment, indicating either deep- or quiet, shallow-water conditions. Although there is no specific indication of water depth, the absence of both coarse terrigenous clastics and abundant tractional structures suggest water depths at least below the influence of storm waves (several tens to >200 m), tidal currents and bottom currents, and away from the influence of coarse terrigenous sedimentation of land masses. The mantling nature of bedded carbonaceous sedimentary facies on the irregular surface of the underlying lava indicates that pelagic and hemipelagic fallout, as well as dilute fine-grained suspension density currents were the most likely sedimentation processes. Rare low-angle discordances may indicate local weak bottom currents or turbulence (possibly associated with hydrothermal plumes). Similar fine-grained
pyroxenes are difficult to discern (Claoué-Long 1986). The discrete plagioclase laths are set in an irregular aggre-

gate of actinolite, in which grain boundaries of the original

fine-grained basalts and lack swallow-tail terminations.

Claoué-Long (1986) also observed rare, tiny olivine crystals and their serpentine pseudomorphs in the lower member

of the Lunnon Basalt. Despite the metamorphic effects, igneous textures are well preserved in places. Claoué-Long (1986) identified numerous, very small plagioclase laths. These commonly preserve swallow-tail terminations and impinge upon one another to form a network with secondary interstitial amphiboles producing an intergranular or subophitic texture. The same textures and mineralogy are also present in the fine-grained pillow basalts and basalt breccias. The chilled margins of the basalts rarely have igneous textures preserved.

Vesicles and amygdales are conspicuously absent in the upper member, but occur in the lower member of the Lunnon Basalt. Amygdales ranging from millimetres up to centimetres in diameter were observed in drillhole KD 1029 at depths greater than 357 m. Amygdales are rarely spherical. A 1 m thick zone of irregular-shaped cavities at the top of a massive basalt is well exposed at the 5 footwall drive of the Foster Mine. These cavities may be up to several tens of centimetres in their maximum dimension and are associated in some exposures with a zone of brecciation. This zone is then overlain by a thin, pale siliceous sedimentary unit separating the upper and lower members of the Lunnon Basalt.

Dolerites are rare and were only observed in drillhole KD 1029 (Figure 2), where they are represented by two distinctly different bodies with an aggregate thickness of 140 m and have irregular contacts with associated basaltic intrusions. Contacts with the Lunnon Basalt stratigraphy are obscured by missing core samples and one highly altered faulted contact. The dolerites are also dark greenish-grey in colour, they have been metamorphosed to the same grade as the rest of the Lunnon Basalt, and show no evidence of a penetrative foliation. The mineralogy is almost identical, but coarser grained than the fine-grained massive basalts, but no olivine was observed. The most distinguishing feature of the doleritic basalts is the larger grain size. Plagioclase laths are up to 1.5 mm long, but are not as abundant as in the fine-grained basalts and lack swallow-tail terminations. The discrete plagioclase laths are set in an irregular aggregate of actinolite, in which grain boundaries of the original pyroxenes are difficult to discern (Claoué-Long 1986). Whole-rock analysis of the dolerites is similar to the fine-grained massive basalt, and the immobile-trace-element characteristics (Ti, Zr, Y) are similar to the Lunnon basalts, but different to analyses from other basalts and dolerites in the Kambalda-Kalgoorlie stratigraphy (Morris 1993a).

**Interpretation**

Massive basalts are intimately associated with pillow basalts and basalt breccias, confirming a subaqueous depositional environment for both the volcanics and intercalated sulfidic sedimentary horizons. Similar associations have also been identified from studies of the present ocean floor and other ancient successions (Dimroth et al. 1978; Hargreaves & Ayres 1979; Schmincke & Sunkel 1987). Bonatti and Harrison (1988) proposed that submarine massive basalts represent lavas erupted at high magma discharge rates and relatively low viscosities. Ballard et al. (1979) suggested submarine massive lavas erupted at high discharge rates are most likely surface fed, although the basis for this is not clear. The dolerites in hole KD1029 are interpreted as being cogenetic sills (i.e. not part of the regionally younger dolerite intrusive event) based on metamorphic grade, major element and immobile-trace-element chemistry similar to the Lunnon Basalt, and in the absence of contrary data.

Gresham (1986) and Claoué-Long (1986) suggested the absence of vesicles in the upper member is indicative of the basalts being extruded under high confining pressures, at water depths possibly exceeding 5000 m. Vesiculation in subaqueous lavas occurs where the vapour pressure exerted by a volatile phase(s) in a magma exceeds the magmastatic pressure, the inertial force of the magma, the surface-tension force, the viscous force (Sparks 1978), and the hydrostatic pressure of the water mass (McBirney 1963; Cas & Wright 1987; Cas 1992). Therefore if these forces, especially hydrostatic pressure, are high enough, and the volatile content is relatively low, no vesicles may form. Although the vesiculosity of submarine basalts has been correlated with water depth for specific compositional suites of lavas (e.g. Jones 1969a; Moore & Schilling 1973), generalisations are not possible because variations in volatile and other elemental contents influence magma viscosity, and vesiculosity. The absence of vesicles is consistent with eruption and emplacement in considerable water depths (>1000 m).

**Pillow basalts**

**Description**

Pillow basalts comprise approximately 45% of the stratigraphy of the Lunnon Basalt, and facies intervals vary from 3 to at least 15 m thick. Two subfacies, mesopillow and megapillow subfacies, have been identified in underground exposures. Pillows are fine-grained, non-vesiculated, closely packed and vary from tens of centimetres to several metres in maximum dimension. Mesopillow basalts consist of elongate pillows, approximately 30–150 cm long in their maximum exposed dimension, and are the most common pillow basalts observed. Megapillows are commonly associated with mesopillows, but are much larger, ranging from 150 cm up to 5 m in maximum dimension.

Pillow margins are defined by very fine-grained chilled rims which are 10–30 mm thick and commonly exhibit...
radial and subconcentric perlitic fracturing. The cores of the pillows consist of uniformly fine-grained basalt. Varioles defined by radiating clusters of amphibole needles, occur in a 5 cm concentric zone adjacent to the chilled margins of some pillows.

In cross-sectional exposure (Figure 4a) pillows are circular or exhibit a convex upper margin, and a lower margin which moulds into the shape of subjacent pillows. In longitudinal exposure (Figure 4b, c) pillows are interconnected and may contain a series of paired re-entrant embayments at the pillow margins, representing successive propagation stages (Yamagishi 1985). Hargreaves and Ayres (1979) identified similar features in Archaean metabasalts in Canada. Small budded pillows occur on the ends of some pillows (Figure 4b) and have been used as a palaeoflow indicator (Table 2).

The upper parts of some pillows contain a small intrapillow cavity with an arched roof and flat floor. The cavities are filled by carbonate with minor quartz, and may occur singularly or multiply, stacked one on top of another, with two or more on the same horizon (Figure 4d). The cavity shape varies from flat, cigar-shaped openings (exposure through the long axis), to smaller, less elongate openings (exposure through the short axis. Cavities may be up to 60 mm thick, and range from 5 to 30 cm in their longest dimension.

Pillow interstices generally comprise 10% or less of the rock and are commonly filled by secondary carbonate with minor quartz. Elongate and splintery clast breccias may also fill the interstices (see below). The secondary carbonate and quartz in places also appears to have replaced pillows and/or interstitial breccias. Nickel sulfides also occur, but rarely, and generally less than 15 m from the contact with a massive ore zone (Harley 1980).

Interpretation

Pillow basalts indicate a subaqueous depositional environment. They may form in shallow- or deep-water settings and are therefore not effective palaeodepth indicators. Each pillow represents a small, individual flow lobe (Furnes &

Figure 4 (a) Cross-sectional exposure of mesopillows at the 437 level of the Otter-Juan mine. (b) Oblique longitudinal section through mesopillows at the 450 level of the Otter-Juan mine. Note the interconnected and the elongate tube form of the basalt compared with (a). Note also the megapillow with large internal cavity, top left hand corner (field of view 4 m). (c) Re-entrants at margins of pillows creating bud-like shapes at the 437 level of the Otter-Juan mine (field of view 1 m). (d) Multiple intrapillow cavities in pillow basalts at the 510 ramp of the Foster mine (field of view 2 m).
Fridleifsson (1978). Pillows are interconnected and form by budding and digital advance from cracks on the surface of propagating, branching and curving subaqueous lava tubes (Moore 1975; Hargreaves & Ayres 1979; Cas & Wright 1987). Re-entrants on pillow margins (Figure 4b) represent the contact between successive budding pillows, each re-entrant pair representing a separate budding event along a transverse spreading crack (Yamagishi 1985).

The pillow basalts are most likely extrusive in origin, as intrusive pillows tend to be less regular in shape and size, frequently have quench-shattered peperitic margins and have preexisting sediments filling the interstices (Cas & Wright 1987; Cas 1992), none of which is the case for the Lunnon Basalt. The chilled margins result from rapid chilling when hot lava comes into contact with cold water. Radial and concentric fracturing in the margins may also develop as a result of rapid cooling. Varioles, which are most commonly found at the margins of mafic massive and pillow basalts. The transition from mesopillows to megapillows may therefore be attributed to locally higher magma supply rates, or vice versa. Perhaps the megapillows represent significant and longer lived feeder tubes in the pillow lava system. Longitudinal exposures through pillows (Figure 4b) were interpreted as resulting from flattening due to deformation (Cowden & Roberts 1990; Morris 1993a). However, the absence of penetrative deformation suggests that most pillows were originally elongate and even now are relatively undeformed.

### BASALT BRECCIAS

**Description**

Basalt breccias are disorganised, may be matrix or clast supported and are composed of ash- to block-sized non-vesicular fragments. They comprise approximately 8% of the stratigraphy and are generally of limited lateral persistence with rapidly fluctuating thicknesses. Two subfacies have been defined: irregular lobe-shaped and angular clast breccias, and elongate splinter clast breccias; both are monolithic.

Irregular lobe-shaped and angular clast breccias are comprised of highly irregular, fluidal lobe-shaped and ash-to block-sized angular clasts of basalt (Figure 5a). Jigsaw-fit textures and perlitic cracks within clasts are commonly

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**Table 2: Palaeoflow indicator measurements taken in the Lunnon Basalt.**

<table>
<thead>
<tr>
<th>Location of palaeoflow indicator (level, coordinates)</th>
<th>Palaeoflow direction (plunge/plunge direction*)</th>
<th>Style of palaeoflow indicator</th>
</tr>
</thead>
<tbody>
<tr>
<td>Otter-Juan Mine 2542 (552275N 370715E, -502.5mRL)</td>
<td>40°/250-250°</td>
<td>Apophysys in breccia</td>
</tr>
<tr>
<td>1832 (551411N 371085E, -306mRL)</td>
<td>35°/245-255°</td>
<td>Budding direction from re-entrant margin in mesopillow</td>
</tr>
<tr>
<td>1832 (551455N 371106E, -302mRL)</td>
<td>40°/255-259°</td>
<td>Bud on elongate pillow</td>
</tr>
<tr>
<td>1732 (551405N 371110E, -293mRL)</td>
<td>30°/265-295°</td>
<td>Bud on elongate pillow</td>
</tr>
<tr>
<td>1732 (551413N371098E, -296mRL)</td>
<td>34°/260-340°</td>
<td>Apophysys in breccia</td>
</tr>
<tr>
<td>1732 (551300N 371104E, -276mRL)</td>
<td>40°/260-300°</td>
<td>Axis of intrapillow cavity</td>
</tr>
<tr>
<td>1210 (551408N 371215E, -101mRL)</td>
<td>43°/330°</td>
<td>Pillow shape</td>
</tr>
<tr>
<td>1210 (551402N 371215E, -102mRL)</td>
<td>30°/260-340°</td>
<td>Axis of intrapillow cavity</td>
</tr>
<tr>
<td>437 (551082N 371782E, 216mRL)</td>
<td>35°/330°</td>
<td>Drooping elongate pillow</td>
</tr>
<tr>
<td>437 (551080N 371703E, 217mRL)</td>
<td>30°/260-340°</td>
<td>Budding formed by re-entrant margins in mesopillow</td>
</tr>
<tr>
<td>450 (551205N 371640E, 205mRL)</td>
<td>30°/260-340°</td>
<td>Apophysys in breccia</td>
</tr>
<tr>
<td>Foster Mine 885 FWD (530455N 382155E, -14mRL)</td>
<td>43°/220-290°</td>
<td>Drooping elongate pillow</td>
</tr>
<tr>
<td>885 FWD (530455N 382155E, -14mRL)</td>
<td>75°/220-280°</td>
<td>Apophysys in breccia</td>
</tr>
</tbody>
</table>

*A sector range rather than specific directions are listed in most cases because plunge directions were estimated from largely irregular, two dimensional exposures.*

---

Fridleifsson (1978). Pillows are interconnected and form by budding and digital advance from cracks on the surface of propagating, branching and curving subaqueous lava tubes (Moore 1975; Hargreaves & Ayres 1979; Cas & Wright 1987). Re-entrants on pillow margins (Figure 4b) represent the contact between successive budding pillows, each re-entrant pair representing a separate budding event along a transverse spreading crack (Yamagishi 1985).

The pillow basalts are most likely extrusive in origin, as intrusive pillows tend to be less regular in shape and size, frequently have quench-shattered peperitic margins and have preexisting sediments filling the interstices (Cas & Wright 1987; Cas 1992), none of which is the case for the Lunnon Basalt. The chilled margins result from rapid chilling when hot lava comes into contact with cold water. Radial and concentric fracturing in the margins may also develop as a result of rapid cooling. Varioles, which are most commonly found at the margins of mafic massive and pillow basalts. The transition from mesopillows to megapillows may therefore be attributed to locally higher magma supply rates, or vice versa. Perhaps the megapillows represent significant and longer lived feeder tubes in the pillow lava system. Longitudinal exposures through pillows (Figure 4b) were interpreted as resulting from flattening due to deformation (Cowden & Roberts 1990; Morris 1993a). However, the absence of penetrative deformation suggests that most pillows were originally elongate and even now are relatively undeformed.
preserved. These breccias form discontinuous, tabular, generally concordant units which pinch and swell rapidly and may be up to 2.5 m thick. The upper and lower contacts between the breccias and adjacent pillow or massive basalts are sharp but irregular, but lower contacts may also be gradational. At the base of thick breccia units (>1 m), the clasts may be closely packed and are aligned approximately parallel to the bedding. At the top of thick breccia units the packing is more open and clasts may be embedded at angles between 30° and 70° with the stratigraphy. Irregular lobe-shaped clasts are diagnostic of this facies, and may comprise 20–70% of the breccia. They are generally elongate (up to 80 cm) with high length/width ratios (up to 8:1) and commonly show radial and subconcentric fracturing, though chilled margins are rarely as well defined as in the pillow basalts. A single subcircular cavity, similar to those associated with mesopillows, occurs in the centre of many of the larger lobe-shaped clasts. Varioles are rare, though where present, are found in a 1–3 cm concentric zone adjacent to the chilled margins. The margins of the clasts are occasionally moulded upon each other, creating a deformed appearance. Associated angular fragments may range from rare pie-slice-shaped, to very angular ash- to block-sized fragments (<1 mm to tens of centimetres). They may be elongate or blockly in shape and frequently have sharp edges and corners, and curviplanar margins. Perlitic cracks are common in fragments or on one or more of the fragment margins.

Elongate splinter clast breccias consist of very fine-grained, elongate, subrounded to angular basalt fragments, commonly with perlitic fractures (Figure 5b). The fragments vary from millimetres to several centimetres in maximum dimension and fracturing generally occurs sub-parallel to both the long and short axis of the fragments. They are most commonly observed in the interstices of pillow basalts.

Approximately 20–80% of the breccias consist of carbonate with minor quartz filling interstitial spaces between clasts, though minor nickel sulfides may also be present, most commonly where breccias occur in proximity to a massive nickel sulfide ore body.

**Interpretation**

Lateral and vertical transitions between massive basalts, pillow basalts and basalt breccias indicate emplacement entirely underwater. The transitions are unlike those observed from littoral environments, in which massive basalts (or pahoehoe) pass abruptly into foreset-bedded pillow breccias or pillow basalts (Jones & Nelson 1970; Moore et al. 1973; Furnes & Sturt 1976). The pillow basalts developed on a slope of less than 10° (see above) and they are not foreset-bedded. The absence of conglomerates, resedimented hyaloclastite and tuffs also preclude deposition in a littoral lava delta (Cousineau & Dimroth 1983).

**In situ** jigsaw fit textures are common in the basalt breccias of the Lunnon Basalt and are indicative of a non-explosive, in situ mode of fragmentation. Fragmentation therefore appears to have developed due to **in situ** brittle shattering of the lava, which in subaqueous environments readily occurs through the cooling-quenching effects of cold water on hot magma, producing an aggregate of angular glassy debris called hyaloclastite (Kokelaar 1986; Cas & Wright 1987; Cas 1992).

Moore (1973) noted that when basaltic pillows propagate under water, a film of steam formed (called film boiling or the Leidenfrost Effect: Mills 1964) around the surface of the pillows where red-hot incandescent magma is exposed to seawater. This film insulates them from direct contact with water for the few (micro)seconds necessary for conductive and radiative cooling to form a crust that then further insulates the still-hot pillow interior magma, so preventing quench fragmentation or more dynamic explosive fragmentation. However, if that film collapses before a significant crust forms, due to boundary layer instabilities (Buchanan 1974; Board et al. 1974; Corradini 1981; Wohletz 1983), including turbulence caused by thermal convection, direct contact between water and magma may occur. The magma at the interface with the water chills too rapidly to glass, so that thermal stresses develop, tensional, contractional fractures propagate inwards and lead to cooling-contraction granulation (Kokelaar 1986). **In situ**, jigsaw-fit textures between clasts are a diagnostic feature for recognising hyaloclastites (Yamagishi 1987); however, they are not always preserved. Jigsaw-fit relationships may be destroyed by clast rotation due to further magma intrusion into the hyaloclastitic pile, or continued flowage of the host lava, or resedimentation of hyaloclastite debris down the steep sides of lavas or domes (Cas 1992).

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**Figure 5** Breccia types of the Lunnon Basalt. (a) Lobe-shaped and angular clast breccia at the base of a massive lava. 1832 level, Otter–Juuan Mine. (b) Pillow margin splintery clast breccia in interstice of pillow lava. Jigsaw-fit texture is evident. 1832 level, Otter–Juuan Mine.
According to Bonatti and Harrison (1988) the formation of basaltic hyaloclastite is favoured by subliquidus eruption temperatures, relatively high viscosities, and relatively low eruption rates. The pseudomorphed and relict crystals in the Lunnon Basalt indicate subliquidus eruption temperatures, and the occurrence of numerous pillow basalt horizons indicate moderate to low magma supply rates. Carlisle (1983) and Staudigel and Schmincke (1984) proposed that local slope, such as that associated with pillow mounds or seamounts, would enhance hyaloclastite formation. Since the palaeoslope on which the Lunnon Basalt was emplaced was <1°, it did not have a major influence on the formation of the hyaloclastite.

Irregular lobe-shaped and angular clast breccias in the Lunnon Basalt have commonly been referred to as 'flow-top' breccias (Gresham & Loftus-Hills 1981; Lesher 1983; Cowden & Roberts 1990). This is consistent with the gradational contacts into coherent lavas at the base of the breccia horizons. However, at least some breccias appear to have formed at the base of flow units. These breccias may develop due to the intrusion and brecciation of irregular apophyses of hot lava into an unconsolidated quench breccia developed at the front and/or base of an advancing lava flow. Brecciation of apophyses may have been caused partly by continued flow movement, but the major mechanism would have been further quenching of the apophyses by interstitial water in the hyaloclastite. The presence of perlitic fractures in many of the fragments supports the influence of water in their formation. An alternative explanation is that apophyses of hot lava intruded the quenched top of an underlying flow unit. The elongate splinter clast breccias are generally confined to pillow interstices. Their formation is believed to be the result of a mechanical unrimming process that may occur in pillow basalts, and in rare cases in sheet lava flows (Carlisle 1963). The mechanical contrast between the brittle glassy crust of tube or sheet lava flows may result in the exfoliation of the crust (Fisher & Schmincke 1984). This process may occur several times, and clast shapes are defined by the radial and subconcentric quench fractures on the pillow margins prior to unrimming. High carbonate abundances in these breccias indicate that the carbonate has either replaced a preexisting material (?sediment) or has wedged splinter fragments apart during (displacive) crystallisation in inter-pillow interstices. The lack of evidence for a preexisting sediment and the common jigsaw-fit arrangement of the fragments, even when separated by carbonate cement, in these breccias suggest the latter process.

Lava flow units
The Lunnon Basalt is comprised of a series of superimposed flow units and both compound (lobed, pillow) and simple (massive) types, as defined by Walker (1970), as well as complex lavas occur. Although pillow and massive basalt units are readily identifiable in drillcore, it is difficult to accurately distinguish flow-unit boundaries in drillcore.

Determination of the dimensions and extent of flows is difficult due to the restrictions of limited underground mining development and limited drillcore intersections. Flow-unit thicknesses, for both massive and pillow lavas, ranging from approximately 5 to 20 m were interpreted from underground plan mapping in the Otter-Juan and Foster mines (Figure 6). The most distinctive criterion for recognising flow-unit boundaries, whether massive or pillow, are hyaloclastite breccias which form tabular to lensoidal units of limited lateral persistence, and are up to 2.5 m thick (Figure 6). These breccias are comprised of both pillow clast derived breccias and irregular lobe-shaped and angular clast breccias, and appear to be developed at the base and/or tops of flow units, although some could represent the propagation of quench fragmentation zones within the interior of flows. Less common thin sedimentary horizons up to 3 m thick also define flow-unit boundaries (Figure 6a, b).

Flow-unit widths could not be determined with certainty. The 885 footwall drive of the Foster mine preserves one flow unit or flow lobe, the width being approximately 100 m (Figure 6a). In drillcore apparent massive and pillow lava flow units have similar dimensions, but there are also intervals of massive lava in particular that are many tens of metres thick (Figure 2). In drillcore in particular, it is unclear whether or not multiple flow units have been missed in the absence of intervening breccias or sedimentary horizons, and because of the effects of alteration and deformation. Although the foregoing dimensions provide some initial two-dimensional understanding of the flow-unit dimensions they cannot be considered representative of the entire stratigraphy. Many more measurements, and especially more plan-view and cross-sectional exposures are required before a better understanding of the flow units and their dimensions can be developed.

Although discrete massive and pillow lavas almost certainly occur in the Lunnon Basalt, there are also exposures where transitions from massive to mega- to mesopillow lava occur, and similar transitions have been noted by Morris (1993a) in basalt exposures on the shores of Lake Cowan to the south. Dimroth et al. (1978) and Hargreaves and Ayres (1979) first documented such transitions in Archaean metabasalts in Canada. These authors attributed lateral and upward variations from massive to pillow basalts to decreasing lava supply rate and increasing viscosity and cooling with distance from the eruption point. Lateral changes from massive facies close to vent, to pillow facies around the margins of flows result from decreasing magma supply rates to the margins of flows as the flows spread and their surface area increases (Dimroth et al. 1978), even though the magma eruption rate from the vent may remain constant.

Eruption style and lava flow propagation processes
Many of the features and textures of Archaean subaqueous lava flows are so similar to those of recent flows that interpretation by analogy is possible. Although only limited submarine eruptions have been filmed (e.g. shallow-marine basaltic lava forming events on Hawaii: Moore 1975), drilling of oceanic crust (e.g. Ballard et al. 1978; Anderson et al. 1982), submersible studies (e.g. Ballard & Moore 1977) and study of uplifted subaqueous basaltic extrusives, including Archaean flows (e.g. Dimroth et al. 1978; Hargreaves & Ayres 1979; Yamagishi 1985; Walker 1992),
have contributed to a better understanding. In addition, recent observations of the propagation mechanisms of subaerial basaltic lavas from the current, most recent eruption on Hawaii (e.g., Hon et al. 1994) have improved understanding of ancient lava flow processes and characteristics.

The association of massive basalts, pillow basalts and basalt breccias are indicative of a subaqueous, effusive eruption of basaltic lava. As discussed above, massive flows can be inferred to represent high magma discharge rate eruptions whereas pillow lavas represent either low magma discharge rate eruptions and/or the distal margins of massive lavas which develop a marginal pillow facies because the magma supply rate to the margins of the lavas decreased as the area of the lavas increased. The rarely exposed, thin sedimentary units within the stratigraphy of the Lunnon Basalt represent a period of quiescence between eruptions, but their general scarcity indicates that lavas were erupted in relatively rapid succession with few significant periods of eruptive repose.

Since no fire-fountain-type spatter deposits (Smith & Batiza 1989) or pyroclastic deposits are known from the Lunnon Basalt, the eruptions were relatively passive, probably tube-fed eruptions, with the erupting magma insulated from the water mass by a chilled lava crust surrounding interior lava tubes, and at vent, perhaps a lava pond. Observations on Hawaii have shown that pahoehoe lavas, the subaerial analogues of pillow lavas, inflate in time through the development of an internal feeder tube magma plumbing system that feeds magma internally to the margins of the lava (Hon et al. 1994). Once a chilled skin and crust form on the surface of the lava, which is almost instantaneous when hot lava is exposed to the atmosphere, this crust insulates the hot magma in the interior of the lava from heat loss. Hot lava then reaches the flow margins, which propagate forward through break-out zones. Through time the lava area increases, it becomes longer and slowly thickens through inflation from the interior feeder tube system, so causing the crust to rise (Hon et al. 1994; Self et al. 1996). Similar processes are thought to be possible with submarine lavas, which if insulated from direct contact with water through film boiling also develop a coherent crust or chilled skin which insulates the still hot, liquid interior of the lava. Features indicating internal inflation and the likely existence of internal feeder tube systems have been documented from lavas of the modern ocean floor; including channels or collapsed feeder tubes (Fornari 1996) and tumulus-like features (Applegate & Embley 1992).

The large volume of coherent basaltic lava flows in the Lunnon Basalt indicate that film boiling at the lava surfaces has effectively insulated the propagating lava flows from direct contact with water most of the time. However, the occurrence of locally developed hyaloclastite at the margins of some flows indicates that film collapse and quenching occurred periodically. There appears to be no depth constraint on quench fragmentation and so the hyaloclastites do not help constrain palaeo-water depths. Similarly, the associated sedimentary facies only suggest water depths below storm wave-base (? several tens to >200 m).

The absence of pyroclastic deposits provides an indication of minimum water depths. Explosive eruptions driven by exsolving magmatic volatiles occur only if magmatic gas pressure significantly exceeds ambient pressure (Cas & Wright 1987; Cas 1992). In this case ambient pressure would be hydrostatic, increasing at a rate of $10^5$ Pa/10 m of water depth. Since no in situ pyroclastic deposits have been found within the Lunnon Basalt the water depth was too great to allow explosive expansion of vesicles. This is consistent with the low vesicle content of the Lunnon Basalt. Either parental magmas had low volatile contents, so allowing passive eruptions in even relatively shallow water depths or, if the volatile content was high, very substantial water depths existed (? >1000 m) to suppress explosions.

More importantly, the absence of phreatic pyroclastic deposits provides some limit to minimum water depths. When film collapse occurs at the interface between magma and seawater, the resulting direct contact between the two could lead to quench fragmentation and/or to phreatic explosions (Kokelaar 1986). Since quench fragmentation has occurred in the Lunnon Basalt direct magma–water interaction occurred. Steam explosions occur if the gas pressure of superheated seawater, which is flashed to steam, is significantly greater than the ambient hydrostatic pressure. Explosions involve instantaneous gas volume expansion, but changes in the specific volume of water at magmatic temperatures vary exponentially as a function of confining pressure (McBirney 1963). Only at pressures of about 10 MPa and less (≤1000 m water depth) are volume expansion rates high enough to become explosive. Peckover et al. (1973) calculated that at water depths greater than about 700 m, superheated steam cannot expand rapidly enough to be explosive and the explosive intensity is essentially suppressed. The emplacement depth of the Lunnon Basalt was therefore at least 700 m.

**PALAEOGEOGRAPHIC AND TECTONIC SIGNIFICANCE**

Because the Lunnon Basalt was a regionally extensive (>500 km$^2$), thick (>1750 m), basaltic succession of tholeiitic affinity, three possible palaeogeographic–tectonic settings are possible: (i) layer 2 of oceanic crust; (ii) topographically–tectonically ponded lava field; and (iii) submarine basaltic lava shield succession.

**Oceanic crust**

Macdonald (1982) suggested that mid-ocean ridge volcanism originated from fissures within a median rift valley. Mid-ocean ridge volcanoes are dominated by ridge-parallel fractures and faults, producing a very rough ridge and chasm terrain (Ballard & van Andel 1977). The stratigraphy of oceanic crust has been defined from a combination of seismic stratigraphic studies and drilling of the sea floor. Anderson et al. (1982) proposed that a complete section of oceanic crust at DSDP site 504B has a thickness of 5–7 km, and consists of: layer 1, pelagic sediments; layer 2a, a succession of permeable pillow lavas, breccias and minor massive lavas; layer 2b, predominantly pillow lavas, with minor breccias and massive lavas, and fractures filled with smectite; and a transition layer with decreased frequency of pillow lavas but dykes becoming predominant. The transition layer is about 600–700 m thick and includes a stock-
Figure 6  (a) Plan of part of 885 footwall drive, Foster mine, depicting the base of the komatite succession, the continuity of the contact sediment, and the thickness and apparent width of the uppermost pillow lava or one of its flow lobes. Note flow boundaries are marked by hyaloclastite breccia horizons and the contact sediment. (b) Plan of part of the 1210 incline, Otter-Juan mine, preserving the top few lava flows of the Lunnun Basalt, the contact sedimentary unit, and the base of the komatite which is cut irregularly by a young felsic dyke. (c) Plan of part of the 1832 level, Otter-Juan mine, showing the relationships between massive, pillow and breccia basalt facies. (d) Plan of part of the 1732 level, Otter-Juan mine, showing the relationships between massive, pillow and breccia basalt facies.
Volcanology of Lunnon Basalt, WA

Thin (5-30cm) breccia
005/40-W

Good exposure of pillowed basalts, occasional elongate & splintery fragment breccias

Pillows are very weakly defined by rare interstice spaces.

Longitudinal exposure through pillows in wall

Wall sketch of incipient pillow margins at transition between pillowed and massive basalt in complex pillowed flow

Mapping joins onto 1732 Level of the Otter-Juan Mine

Level of the Otter-Juan Mine

Scale: 1:500

1832 LEVEL, OTTER JUAN MINE

Contact between basalt and ore.
Strongly altered with no texture preserved in the basalt. The contact is very planar.

Very well preserved textures in breccia

Breccia dipping -45° towards the west

Longitudinal exposure through pillows in wall

Massive basalt

1732 LEVEL, OTTER JUAN MINE

work of Fe, Zn, Cu-sulfides with laumontite, chlorite, calcite and quartz veins in a fractured and brecciated pillow lava sequence. The transition layer passes downward into layer 3, a sheeted dyke complex. A similar stratigraphy is likely in backarc, marginal-sea oceanic crust.

Mid-oceanic ridge basalts or MORB are oceanic tholeiites (Basaltic Volcanism Study Project 1981; Wilson 1989), indicating derivation from primitive magmas. Campbell (1985) considered the Lunnon Basalt to have affinities with continental basalts, whereas our data and that of Redman and Keays (1985) suggest affinities with island-arc tholeiites when plotted on some tectonomagmatic discrimination diagrams for modern lavas. However, as pointed out above, the significance of this to the tectonic
context during the Archaean is unclear. The only significant physical evidence in ancient successions for oceanic crust and sea-floor spreading is a well developed sheeted dyke complex interfacing directly upwards with layer 2 basalts (Cas 1983; Cas & Wright 1987) and downwards into a sialic/mantle basement, unless they have been separated by faulting during tectonic emplacement (Moorees 1982).

The Lunnon Basalt, consisting of a thick succession of pillow basalts, massive basalts, hyaloclastite and sulfidic sediment, contains many of the features characteristic of layer 2a of the oceanic crust. Drillhole KD 1029 provided the only means of testing the stratigraphy of the Lunnon Basalt at depth, but did not intersect a sheeted dyke complex. There is, therefore, no definitive physical geological evidence that it represents ancient oceanic crust.

In addition, the presence of 3400 Ma xenocrystic zircons in the Lunnon Basalt indicates that the magmas passed through older sialic crust, indicating that the tectonic setting of the greenstones of the Kambalda region was ensialic not ensimatic (Compston et al. 1986). Therefore, unless the Lunnon Basalt represents transitional crust, associated with the rifting of a sialic crustal block during the early stages of opening of an ocean basin, it is unlikely to represent oceanic crust. This is consistent with the opinion of Bickle et al. (1994) that as yet no convincing sections of Archaean oceanic crust have been documented anywhere.

**Topographically-tectonically ponded lava field**

Alternatively, the Lunnon Basalt could represent a thick succession of basalt lavas erupted in an intraplate setting associated with mantle hot-spot activity and/or crustal extension/transtension. Intraplate basalt lava fields are well known, including from the modern ocean floor (e.g. Fornari et al. 1985, Lipman et al. 1988). Since the thickness of the Lunnon Basalt is about 1750 m or greater, it could have been ponded in a significant topographic or tectonic depression and/or experienced on-going subsidence during the overall eruptive history. The absence of a synvolcanic feeder dyke system in the studied areas of the Lunnon Basalt suggests that the eruption point was distant from the studied successions. If Lesher and Arndt (1995) are correct about the Lunnon Basalt being part of a regionally extensive, large-volume basalt lava succession, then a very large submarine lava field was involved and several eruption sites may have existed.

**Shield volcano**

The association of pillow lavas, hyaloclastites, massive lavas, and doleritic intrusions in the Lunnon Basalt is also consistent with the preservation of part of a submarine shield volcano (Jones 1989b; Moore & Fiske 1989). Basaltic shield volcanoes have gentle slopes between 2° and 10°, and are composed almost entirely of thin, extensive lava flows (Macdonald 1972); they have basal diameters from a few kilometres (e.g. Mauna Ulu on Hawaii (Searle 1983; Batiza & Vanko 1983)) to over 100 km (e.g. Mauna Loa in the Hawaiian Islands (Macdonald & Abbott 1970; Cas & Wright 1987)). Compositional variations from tholeiitic to more alkalic with time may occur during the growth of shield volcanoes (Feigensen & Spera 1981; Feigensen et al. 1983; Porter 1979). This observation is consistent with the Lunnon Basalt representing the early tholeiitic stages of development of a shield volcano. Palaeoflow indicators from pillow basalts and basal breccias in the Lunnon Basalt indicate the lavas flowed towards the west; ranging from southwest to north-northwest (Table 2). If the Lunnon Basalt is representative of an Archaean submarine shield volcano, the Kambalda district may represent the western flank. However, slide and debris-flow deposits are entirely lacking, and these would be expected on the flanks of a substantial volcanic edifice with its relief (Lipman et al. 1988). Furthermore the uniformly thick komatiitic succession indicates that the pre-komatiite relief was no more than several tens of metres.

**Summary**

In summary, the zircon xenocrysts in the Lunnon Basalt (Compston et al. 1986) indicate development in proximity to subjacent sialic or transitional crust, and together with the lack of the preserved stratigraphy of oceanic crust it is unlikely that the Lunnon Basalt originated as oceanic crust. The preserved characteristics are consistent with an origin as a tectonically-topographically (?rift) ponded, or shield volcano. However, the absence of significant resedimented facies, indicative of significant relief above basin-floor level, support the ponded lava field model. The apparent absence of a synvolcanic feeder system indicates that the preserved succession represents a distal succession away from the eruption point.

**Implications of the Volcanology for Mineralisation**

The principal nickel sulphide orebodies at Kambalda occur at or near the contact between the Kambalda Komatiite with the underlying Lunnon Basalt and are commonly referred to as contact ore (Gresham & Loftus-Hills 1981; Gresham 1986). The ore occurs as ribbon-like bodies or 'shoots' that are generally localised in and over depressions in the upper surface of the Lunnon Basalt (Lesher 1989). These depressions are generally narrow, elongate, complex structures and their lengths, widths and depths are highly variable (Gresham & Loftus-Hills 1981). Apart from the Durkin shoot at the northern end of the Kambalda Dome, the depressions generally trend between 310° and 345° (Gresham & Loftus-Hills 1981). An important feature of these depressions is the general antithetic relationship between the occurrence of sediments and the nickel sulhide mineralisation, although Holllambry (1995) and Marchiori (1995) have shown that this is by no means a constant relationship, at least at Tramways and Blair.

The origin of the depressions is uncertain, and interpretation is hampered by incomplete exposure and lack of marker horizons. A variety of models have been proposed including: (i) synvolcanic faulting (Ross & Hopkins 1975; Gresham & Loftus-Hills 1981; Gresham 1986); (ii) primary lava volcanic topography (Green & Naldrett 1981; Lesher 1983; Lesher et al. 1984; Lesher 1989); (iii) thermal erosion...
channels (Huppert et al., 1984; Huppert & Sparks 1985; Evans et al. 1988; Evans 1989); and (iv) post-volcanic structural controls (Cowden 1988; Cowden & Archibald 1989; Cowden & Roberts 1990).

**Synvolcanic faulting**

The recognition of a well-defined 'ore prism' (i.e. spatially related contact and hangingwall ores, a lack of sediment, an anomalous thickness of high-Mg komatiite with cumulative textures) led to the development of a model in which original linear depressions, bounded in part by a series of subvertical faults, were considered parts of eruptive fissure systems in a basaltic ocean floor (Ross 1974; Ross & Hopkins 1975; Gresham & Loftus-Hills 1981; Gresham 1986). The greatest problem with this model, however, is the lack of observed feeder dykes in trough positions in the Lunnon Basalt. Gresham (1986) suggested the fissures would have been no more than 1 m wide and may now be unrecognisable due to reactivation of associated faults or were obliterated by the extensive later intrusions in the area.

**Primary volcanic topography**

Lesher et al. (1984) proposed that the mineralisation depressions were linear topographic depressions in the surface of the Lunnon Basalt, such linear depressions or channel-like hollows bounded by the margins of non-overlapping basaltic lava flows or lobes on either side (Figure 7a, b) (Lesher 1989). The topographic depressions would focus and channelise the succeeding komatiite flows, and would thus control the position of major komatiite lava channels (Lesher et al. 1984). Lesher (1989) suggested the depressions were then variably modified by thermal erosion and superimposed structural deformation.

In Hawaii, modern subaerial pahoehoe lavas are lobate and inflate through the development of interior feeder tube systems generating relief of at least several metres (Hon et al. 1994). Submarine pillow lavas, which are the subaqueous equivalents of pahoehoe lavas, are also known to develop irregular relief and do not spread as sheets of even thickness with flat tops. In fact all modern lavas, whether they are basaltic or not, show significant surface relief. Many are lobate, and have steep leveed margins and channels. It is therefore certain that original relief of at least several metres existed on the surface of Lunnon Basalt lava flows. Hargreaves and Ayres (1979) and Dimroth et al. (1978) indicated that relief of up to 20 m existed between successive lava flows in subaqueous Archaean tholeiitic metabasalts in Canada, in successions very similar to the Lunnon Basalt.

Depressions may also develop on lava surfaces due to the collapse of the roof of large lava tubes (Figure 7c). Lava tubes are most commonly known in subaerial pahoehoe flows (Cas & Wright 1987; Hon et al. 1994; Self et al. 1996). However, subaqueous examples have also been documented for the Archaean by Cousineau and Dimroth (1982) and on the modern ocean floor (Fornari 1986). Hollow lava tubes form when lava supply to a tube ceases and magma drains from it (Cousineau & Dimroth 1982), in a similar way as for the development of intrapillow cavities, though on a much larger scale. Cousineau and Dimroth (1982) suggested that breccias may develop due to the collapse of the roof of these hollow tubes, although no evidence of this was observed in the Lunnon Basalt. Fornari (1986) suggested tube-fed flows may gradually deflate, so that breccias do not develop, but deflation canals do.

Limited exposure and deformation have unfortunately made it very difficult to evaluate how much original volcanic topography existed in the Lunnon Basalt. Although there is no evidence for collapsed or intact lava tubes in the Lunnon Basalt, it is likely that such features, as well as lava flow/lobe topography, influenced the localisation of initial flow paths or channels in the succeeding low-viscosity komatiite lava, the location of interior tubes, and perhaps the mineralisation. Palaeoflow indicators in the Lunnon Basalt, taken from a variety of locations at Kambalda (Table 2), indicate the lavas were flowing in a direction between southwest and north-northwest. This is comparable with the orientation of closely associated orebodies and is therefore consistent with a mineralisation model involving primary volcanic topography.

![Figure 7](image-url)
Thermal erosion channels

Huppert et al. (1984) and Huppert and Sparks (1985) proposed that the depressions confining the nickel sulfide mineralisation may have formed by thermal erosion beneath linear komatiite lava channels. The komatiite lavas were thought to have been emplaced as low viscosity turbulent flows in which mixing would maintain a uniformly high temperature and heat flux throughout the flow and to the substrate (Huppert et al. 1984). Huppert et al. (1984) suggested that up to 10% of the substrate could thus melt and become assimilated, resulting in the formation of deep channels and compositional contamination of the komatiites. Thin flows, such as at the flanking positions of the depressions, would be emplaced in a laminar flow regime, with heat flux to the substrate declining because heat loss through conduction from the base of a laminar flow would not be replenished and therefore substrate melting would be unlikely (Huppert et al. 1984). Whether or not a flow is thick and turbulent or thin and laminar at the time of flow is therefore an important question.

The recent observations of Hon et al. (1994) have shown that the final preserved thickness of Hawaiian pahoehoe lavas is due to inflation and does not represent the dynamically flowing thickness of the lavas at any time. Since komatiites apparently had viscosities two orders of magnitude less than basalts, it is likely that they would have spread as very thin fluidal sheets and inflated with time, not advanced as thick flows tens or more metres thick as the komatiites of the trough settings are preserved today. Also, if the flows were progressively thickening through inflation, fractional crystallisation was probably continuously leading to the deposition of olivine to the base of the flow, leading to the uprise of the interface between the developing crystal cumulate and the komatiitic liquid in the flow interior. The thickening basal cumulate zone would then have acted to insulate the substrate from thermal melting. If these factors are correct, then the heat budgets and the turbulence levels of komatiite flows at any time during their emplacement would have been less than expected, and perhaps significantly less than required and calculated by Huppert et al. (1984) to induce thermal erosion.

The relationship between the nickel sulfide ore and the discordant bases of some thick massive komatiite units to underlying stratigraphy is consistent with thermal erosion (Hill et al. 1995). However, there is considerable confusion from the isotope geochemistry of the komatiites about whether or not thermal erosion and compositional contamination have occurred (see below); detailed textures and contact relationships that should result from such thermal erosion are minimally documented and other alternatives (physical erosion, unconformities, relict submarine slide surfaces, syn- to post-emplacement faults) have not been adequately evaluated. At Kambalda, all mineralised depressions seen in this study show some evidence of structural modification.

Arndt and Jenner (1986) claimed rare-earth element data supported crustal and sediment contamination of komatiite magmas at Kambalda. Their data for the Lunnon Basalt and hangingwall basalts are consistent with their interpretation (as are zircon xenocrysts). However, their data for the komatiites indicate minimal, if any, contamination. Lesher and Arndt (1985) by contrast state that the Lunnon Basalt is uncontaminated and the basal Silver Lake Member komatiites are minimally contaminated in the channel (trough) facies, but slightly contaminated in the flanking sheet facies. However, the geochemical data apparently can also be interpreted in terms of metamorphic-alteration effects associated with the long post-emplacement flux of fluids through the Kambalda succession (J. Foster pers. comm. 1996). Gresham (1986) conceded that sediment assimilation may have occurred locally however; he believed it was unlikely to have generated the nickel sulfide ores and the broadly antithetic sediment-ore relationships. His arguments were based on a variety of geochemical and stratigraphic observations against sediment assimilation, including mass-balance calculations. Recent Sm–Nd and Re–Os isotopic, and trace-element studies by Foster and Lambert (1993, 1996), Foster et al. (1995, 1996) and Lambert et al. (1998), also suggest that there is little or no evidence for significant sediment assimilation and contamination of the mineralised komatiites in the Kambalda and other komatiite successions. The geochemical data are therefore conflicting; the geochemists

Figure 8 Untectonised contact between the top of the Lunnon Basalt and the massive ore zone at the base of the komatiite succession, 1014/1, 2 cut-and-fill stope, Foster mine. Note the sharp, jagged contact, the whole pillow shape, and the apparent absence of segments of pillow, and even whole pillows, suggesting physical erosion (from Langworthy 1990).
cannot seem to agree, suggesting that convincing, unequivocal evidence for widespread thermal erosion and sediment contamination is still to be presented.

At the 1014/1.2 cut-and-fill stope at the Foster mine, an extremely rare unetonised contact between the Lunnon Basalt and massive sulfide ore was exposed (Figure 6) and was interpreted by Evans (1989) and Langworthy (1990) as evidence for thermal erosion. The sharp, jagged nature of the contact between the two units, preserving the shapes of underlying pillows and missing pillows and segments, however, poses problems for an origin by thermal erosion. If the Lunnon Basalt had been melted, the contact may be sharp and undulating with perhaps resorption-like embayments, and a gradational change from unmelted to partially melted substrate, but it is unlikely to be sharp and jagged and preserve the morphology of the underlying basalts. The very jagged nature of the contact, the missing pillows and segments, together with an absence of evidence for any melt textures in the basalt, are better attributed to physical erosion processes, rather than thermal erosion. Frost and Groves (1988) suggested that felsic spherules (ocellii) found at the top of the lowermost, mineralised komatite at Kambalda were the melt residues of melted assimilated sediment, based on chemical compositions of spherules and the enclosing groundmass. There is no other independent evidence that this is correct, and given their occurrence in altered, uppermost greenstitch-amphibolite facies rocks, further work seems justified to clarify if ocelli could be related to devitrification, alteration or metamorphism.

Although the general absence of sediment in depressions at Kambalda has made the concept of sediment assimilation attractive, Gresham (1986) observed the presence of sediment-free and sulfide-free zones in both contact and hangingwall depressions and Hollamby (1995) and Marchiori (1995) have noted that significant intervals of sediment do occur in ‘trough’ positions at the Tramways and Blair mines respectively at least. So clearly, komatitites have not nowhere thermally eroded sediment in trough settings. However, other possible mechanisms for removing large volumes of sediments, or other substrate lithologies in trough positions, such as physical erosion processes, appear not to have been considered. These require as much serious consideration as thermal erosion, simply because if effective they would have operated before thermal erosion could take effect.

The fronts of komatite lavas could have caused several types of physical erosion. First, dense turbulent komatiitic magmas would have significant erosive capacity, especially of unconsolidated sediments. Much of such eroded sediment would have been lofted as a suspension cloud, some falling back onto the flow surface, but most remaining suspended until after the cessation of turbulent komatite flow. Some could have been incorporated as intracretal material into the base, and perhaps even assimilated to varying degrees. Secondly, if komatite lavas were turbulent in their cores (= trough positions: Huppert et al. 1984) they should have generated an erosive turbulent bow wave in the water mass immediately ahead of the lava flow. Furthermore, the high temperature of the lava front should have produced significant film boiling (Mills 1984) at the lava–water interface. Both have the capacity to generate high enough turbulent circulation velocities to erode fine silt- to sand-sized sediment and lead to its suspension, as in the first case. Simple Hjulstrom diagram relationships (Sundborg 1956) for minimum threshold erosion velocities and grain size indicate minimum erosion velocities of 20 cm s–1 for unconsolidated silt–fine sand and up to 60–80 cm s–1 for finer cohesive sediment.

Thirdly, the very hot komatite lavas should have caused instantaneous heating and even boiling in the pore water of the sediments over which the flows were advancing. When the pore water in a wet sediment crosses the water–steam phase boundary due to heating by an igneous body there is an instantaneous expansion of water vapour, resulting in a momentary fluidising flow of vapour; and loss of strength in the sediment (Kokelaar 1982). During the resulting liquification, grains are suspended in the interstitial fluid, supported by high excess pore pressures, and the whole aggregate is in a ‘quick’, fluidal state (Cas & Wright 1987), which is highly erodible. Kokelaar (1986) suggested that hydrostatic pressure could suppress the expansion of steam, as it does in magma vesiculation. At depths greater than 3.1 km for seawater, Kokelaar (1982) suggested the critical pressures would be too great for steam-driven fluidisation to occur. However, significant heating of pore water alone may induce fluidisation through simple convection. Therefore thermally induced fluidisation of the sediment over which the fronts of the very hot (1500–1650°C; Lesher 1989) komatitites were flowing also seems to be a feasible mechanism for mobilising and eroding sediment.

In summary, although theory suggests that some komatitites may have had the capacity to cause thermal erosion of substrate, and almost certainly did in some instances, we have found no definite evidence of this at the examined contacts between the top of the Lunnon Basalt and the base of the mineralised komatite succession in the Otter-Juan and Foster mines. Other mechanisms for eroding sediments by physical erosion processes have surprisingly not been explored in any detail. Moreover, the occurrence of massive nickel sulfide ores directly on undisturbed pelitic sedimentary rocks in the Blair mine (Marchiori 1995) indicates that neither thermal erosion nor physical erosion have occurred in some cases. Komatite flow dynamics and the nature of the interaction between komatitites and their substrate require further consideration.

**Structural modifications**

Lesher et al. (1984) attributed the present re-entrant geometry of many of the mineralised depressions to deformation, although they maintained that 2–30 m deep pre-tectonic, volcanically related topography existed. Cowden (1988), however, suggested most of the depressions were formed almost entirely by deformation of a near-planar contact. In this study a number of underground exposures were seen where mineralised depressions had faulted margins (Figure 6d). However, it is unclear if the faulting is superimposed upon original topography.

**CONCLUSIONS**

The Lunnon Basalt comprises a thick succession of pillow basalts, massive basalts and basalt breccias. Flow units
comprise massive and complex pillow flow facies, and successive flows are separated in many cases by hyaloclastite breccia. Sediments are rare, indicating nearly continuous outpouring of lava. The morphological features of the Lunnon Basalt are similar to modern submarine basaltic flows in a range of tectonic and palaeogeographic settings. The excessive thickness of the basalts relative to modern and ancient oceanic crust, and the absence of an exposed or intersected subjacent sheeted dyke complex provides little support for an origin as oceanic crust. The great apparent thickness of the basaltic lava pile is consistent with a submarine shield volcano/sea mount, but the absence of mass-flow resedimented facies suggests an edifice did not exist. Alternatively, the thick lava succession formed as a tectonically-topographically ponded sea-floor lava field. The great apparent thickness (and inferred large volume), the presence of inherited zircons, suggestive of a preexisting subjacent sialic crust, and the close apparent temporal relationship with the overlying regionally extensive, voluminous and thick komatiite succession is suggestive of a tectonic setting conducive to the eruption of large volumes of magmas. Ponding of lavas in a submarine extensional (rift) basin may have occurred. The top of the Lunnon Basalt and the base of the komatiite are everywhere separated by only a thin, several metre thick, contact sedimentary succession, suggesting that the top of the Lunnon Basalt had relatively little local relief (cf. a shield volcano), and that the relatively flat sea-floor topography on top of the Lunnon Basalt was readily accessible to the flood of low viscosity komatiite lavas when they erupted.

The depressions hosting the nickel sulfide mineralisation are complex and varied, thus their present geometry is most likely the result of a variety of processes. Similarities between the Lunnon Basalt, Archaean basalts from Canada and modern basalt lavas indicate that significant relief would have existed at the surface of the Lunnon Basalt. This, together with similar palaeeoflow trends for pillow and massive basalts and the orientations of the linear nickel sulfide depressions, supports volcanic topography as an important factor in determining the localisation of komatiite channels or tubes and probably the ore. There is no unequivocal evidence for thermal erosion by komatiites at Kambalda, although it may have occurred in some instances. However, physical erosion of sediments by the fast flowing, very hot komatiite flows also explains the sediment-free zones in the areas of the depressions at Kambalda. Structural modification has also clearly occurred although the degree of modification remains uncertain.

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