A Cambro-Ordovician submarine volcanic succession hosting massive sulfide mineralisation: Mount Windsor Subprovince, Queensland

by

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Statement

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Abstract

The Seventy Mile Range Group is a Cambro-Ordovician, dominantly submarine volcano-sedimentary succession that occurs within the Mount Windsor Subprovince of northern Queensland, Australia. Detailed facies analysis of the Mount Windsor Formation and Trooper Creek Formation, between Coronation homestead and Trooper Creek prospect (approximately 15 km strike length), has clarified the facies architecture in this part of the Seventy Mile Range Group. The rocks have been affected by regional greenschist facies metamorphism and deformation, and hydrothermal alteration is intense around VHMS deposits.

The Mount Windsor Formation comprises associations of rhyolitic to dacitic autoclastic breccia and coherent facies, 100-500 m thick, which form submarine lavas, domes and syn-volcanic intrusions. The Mount Windsor Formation (300-3500 m thick) is conformably overlain by the Trooper Creek Formation (500-4000 m thick). The Trooper Creek Formation is divided into two members, the Kitchenrock Hill Member and the overlying Highway Member. The new stratigraphic scheme is based on mappable compositional and lithological variations which reflect changing provenance. The Kitchenrock Hill Member (90-110 m thick) includes rhyolitic to dacitic syn- and post-eruptive volcaniclastic facies, minor syn-sedimentary intrusions and rare siltstone units. Some volcaniclastic units contain rounded clasts with geochemical and petrographic properties which suggest they were sourced from the Mount Windsor Formation. Rounded clasts were reworked prior to deposition and imply that the source areas were subaerial or shallow marine. The Highway Member comprises compositionally and texturally diverse volcano-sedimentary facies, including rhyolitic to basaltic lavas and intrusions, andesitic scoria- and bomb-rich breccia, dacitic to rhyolitic pumice breccia, and volcanic and non-volcanic sandstone and siltstone.

The presence of turbidites, hyaloclastite and fossils within the Kitchenrock Hill Member and Highway Member suggest that the depositional setting for the Trooper Creek Formation was largely submarine and below storm wave base. The exception is in the upper part of the Highway Member at Trooper Creek prospect, where microbialites, gypsum molds, and traction current structures indicative of wave activity, collectively suggest a depositional environment above storm wave base. The lithofacies exposed at Trooper Creek prospect suggest that shoaling of the succession occurred in response to construction of a small, submarine andesitic volcano which temporarily emerged above sea level. Growth of the edifice involved: (1) eruption of lava and intrusion of syn-sedimentary sills; (2) strombolian-style volcanism in a near-storm-wave-base environment; (3) hydrovolcanic interactions above storm wave base and possibly
subaerially; and (4) post-eruptive and possibly syn-eruptive degradation of the volcanic edifice. When subsidence due to compaction and/or tectonism outpaced accumulation, the depositional environment returned to below storm wave base.

Syn-eruptive pumiceous and crystal-rich sediment gravity flow deposits in the Trooper Creek Formation were sourced from rhyolitic to dacitic eruptions at volcanic centres which are either; located outside the study area, not exposed, or not preserved. The abundance of pyroclasts (principally pumice, shards and crystals) reflects the importance of explosive magmatic and/or phreatomagmatic eruptions, and suggests that the source vents were in shallow water or subaerial settings. Some key units may be traceable for over 30 km, and are an important framework for exploration within the Trooper Creek Formation.

The Highway Member includes the products of intrabasinal, non-explosive silicic to intermediate eruptions that formed lava- and intrusion-dominated volcanic centres. The volcanic centres are: (1) dominated by syn-sedimentary sills and cryptodomes; (2) comprise thick (> 5 km) lava complexes; or (3) form lava-lobe hyaloclastite domes, up to 500 m thick and at least 1.5 km long. The Cu-Au-rich Highway-Reward massive sulfide deposit is hosted by one small syn-sedimentary intrusion-dominated volcanic centre. Detailed mapping of contact relationships and phenocryst populations suggest the presence of more than thirteen distinct porphyritic units in a volume of 1 x 1 x 0.5 km. The peperitic upper margins of more than 75% of these units suggests that they were emplaced as syn-sedimentary sills and cryptodomes. Evidence for partial extrusion of lava is limited to one rhyolite. The shape, distribution and emplacement mechanisms of the units were influenced by: (1) the density of magma relative to wet sediment; (2) the location and shape of previously emplaced lavas and intrusions; and (3) possibly syn-volcanic faults which may have acted as conduits for rising magma. Deformation and disruption of bedding, resedimentation, dewatering and low-grade metamorphism of the enclosing sediment accompanied emplacement of the intrusions. The resulting patterns of permeability and porosity in the volcanic succession are inferred to have strongly influenced the location and evolution of the syn-genetic hydrothermal system which formed the Highway-Reward massive sulfide deposit. The pyrite-chalcopyrite pipes and marginal pyrite-sphalerite-galena-barite mineralisation are largely syn-genetic sub-seafloor replacements of the host sediment, cryptodomes and volcaniclastic deposits.

Hematite alteration and ironstones are regionally distributed and are not preferentially associated with mineralisation. Most ironstones have geochemical and lithofacies characteristics which suggest they are the deposits from low temperature fluids circulating around lavas and intrusions and within the proximal facies association of shallow submarine volcanoes. At Handcuff, some ironstone lenses have REE and trace element
signatures which suggest they may be associated with as yet undiscovered massive sulfide mineralisation. Many ironstones are sub-seafloor replacements of pumiceous breccia and sandstone beds.

The top of the Highway Member marked the end of intrabasinal volcanism and volcanic-dominated sedimentation. This was then followed by a phase of post-eruption erosion of the subaerial to submarine source areas that led to an influx of variably rounded volcanic- and basement-derived detritus into the submarine basin, during deposition of the overlying Rollston Range Formation.

This research provides insights into understanding comparable modern and ancient submarine volcanic successions and assessing prospective host sequences for massive sulfide mineralisation.
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Chapter 1

Introduction
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Introduction

1.1 Initial statement

The Cambro-Ordovician Mount Windsor Subprovince crops out discontinuously in an east-west belt about 160 km long, south of Charters Towers in northern Queensland, Australia (Fig. 1.1). Volcanic and sedimentary rocks within the subprovince form the Seventy Mile Range Group (Henderson, 1986). The Seventy Mile Range Group includes several volcanic-hosted massive sulfide prospects and two significant deposits (Thalanga and Highway-Reward). This thesis presents the results of an integrated and field oriented volcanological, sedimentological and ore deposit study of Seventy Mile Range Group in the area around the Highway-Reward deposit. Of the four regionally mappable formations, the Trooper Creek Formation is the principal host to the known massive sulfide mineralisation. The present study focuses on this formation and the underlying Mount Windsor Formation. They comprise dominantly felsic to intermediate submarine volcanic and volcanogenic sedimentary lithofacies.

Major differences exist between facies associations of subaerial and subaqueous volcanic centres, and much remains to be learned about the controls on submarine facies. The Seventy Mile Range Group provides a rare opportunity to characterise the volcanic facies associations generated by felsic to intermediate eruptions in a submarine environment. Most research on the modern seafloor has concentrated on the abundant basaltic and andesitic volcanoes. Felsic submarine volcanic centres have been less well studied, in part due to their relative scarcity on the modern seafloor. Modern submarine volcanoes are also largely inaccessible for detailed textural and facies research. Understanding the processes and products of subaqueous felsic volcanic activity is thus largely dependent on studies of ancient sequences, such as the Seventy Mile Range Group, that are now exposed on land.

Although significant advances have been made in understanding Australian volcanic-hosted massive sulfide deposits, detailed studies of the interrelationships between volcanic facies, alteration and mineralisation are rare (e.g. Allen, 1992; Waters and Wallace, 1992; Sainty, 1992). Studies of younger volcanic-hosted massive sulfide deposits show that structures within the volcanic pile, proximity to eruptive vents and the physical properties of the volcanic facies all strongly influence the character, size and location of mineralisation (e.g. Horikoshi, 1969). Careful reconstruction of the palaeovolcanological setting may allow recognition of those parts of the volcanic
Figure 1.1 Distribution of the Trooper Creek Formation and other major lithostratigraphic units in the central part of the Seventy Mile Range Group (Mount Windsor Subprovince), northern Queensland. The locations of the main mineral deposits and prospects are also shown. The inset map shows the full extent of the Seventy Mile Range Group. The study area is largely confined to the exploration licence area in the Mount Windsor Subprovince held by Aberfoyle (MLA 1560/A to P 3380M; dark outline). Modified from Berry et al. (1992).
succession that are most highly prospective for as yet undiscovered ore deposits. The potential for discovery of new massive sulfide deposits in the Seventy Mile Range Group has been an important motivation for the current study.

The primary outcomes of the research are an understanding of the temporal and spatial relationships between volcanism and mineralisation, an interpretation of the style of volcanic activity, detailed definition of the Cambrian depositional setting in the Highway-Reward area and a refined stratigraphy for the Seventy Mile Range Group.

1.2 Location and access

A comprehensive analysis of the Mount Windsor Formation and Trooper Creek Formation, involving drill core logging and field mapping, has been undertaken in the area between Coronation homestead in the west and Trooper Creek prospect to the east (Fig. 1.1). Drill core studies were based in Charters Towers at Aberfoyle Resources Limited's core store. Field mapping focussed on the Seventy Mile Range Group in the exploration licence area held by Aberfoyle (MLA 1560/A to P 3380M) and within the Highway mine lease (ML 1571). The abandoned Highway open cut is located approximately 32 km south of Charters Towers. Access from Charters Towers is via the Gregory Developmental Road. The Highway workings are located on the eastern side of the road, approximately 0.5 km south of the Policeman Creek bridge. Satellite prospects are accessible by four-wheel-drive vehicle. The Seventy Mile Range Group is generally poorly exposed, often covered by younger deposits (e.g. Tertiary Campaspe Formation) and Ordovician to Permian granitoids. The best outcrops are along ridges and in some creeks.

The physiography of the area is dominated by the east-trending Seventy Mile Range, 300 to 580 m above sea level. The range is bordered by extensive plains covered by semi-arid woodlands, and cut by shallow ephemeral creeks and streams. Rainfall is low, averaging 658 mm per year, and long periods of drought are common.

1.3 Thesis objectives

The principal objectives of this thesis are to:

(1) document the character and geometry of volcanic and sedimentary facies associations;
(2) determine the style and setting of the Cambro-Ordovician volcanic activity, including the identification of proximal facies associations;
(3) establish the temporal evolution in the volcanic succession;
(4) evaluate the genetic links between volcanic, alteration and mineralisation processes.

In addition, clarification of the facies relationships and differences in composition and lithology in the area necessitated revision of the stratigraphic framework for the Seventy Mile Range Group. A detailed study of the textural and morphological features of peperite from basaltic to andesitic Palaeozoic examples in eastern Australia was also undertaken (Appendix A).

1.4 Methods and thesis organisation

Systematic geological mapping and drill core logging of the Seventy Mile Range Group was undertaken over three field seasons. Particular emphasis was placed on identification and interpretation of volcanic textures and facies using a process-orientated approach. Reconnaissance geological mapping in order to establish a general outline of the volcanic facies distribution preceded detailed mapping of areas where outcrop is best. Geological mapping was carried out using 1:5000 and 1:2500 base maps acquired from Aberfoyle Resources Limited. Critical sections were mapped at 1:200 scale using tape and compass methods. Field data were compiled onto photo centred, 1:10000 base maps in order to make regional correlations. Eighty three diamond drill holes from Highway-Reward and surrounding prospects were logged at 1:200 scale using graphic lithological logs. The logs were used to construct cross-sections (±6.25 m spacing) through the Highway-Reward deposit (Appendix B). Important drill log summaries are presented in Appendix C. Petrographic and geochemical techniques were used to complement field interpretations.

With the exception of Chapter 7, grid co-ordinates referred to in this thesis are Australian Metric Grid (AMG) co-ordinates. Chapter 7 is based on data from diamond drill holes collared along successive E-W cross-sections within the mine grid, and the mine grid co-ordinates are used. Mine grid north is 55.5° to the east of AMG north.

The thesis objectives have been addressed in eight chapters. Chapters 1 and 2 are introductory chapters and provide reviews of previous work, the geological setting, characteristics of the massive sulfide deposits and tectonic history. Chapter 3 reinterprets the position and nature of contacts between some formations in the Seventy Mile Range Group and clarifies the character of the lithofacies. A new formal subdivision of the Trooper Creek Formation is proposed. Chapter 4 documents the internal facies characteristics of a shoaling andesitic volcanic centre in the upper part of the Trooper
Creek Formation, at Trooper Creek prospect. The facies architecture of this part of the Trooper Creek Formation contrasts with the deeper water setting implied for the remainder of the succession. Chapter 5 describes a silicic, submarine syn-sedimentary intrusion-hyaloclastite host sequence to the Highway-Reward deposit. The character of the Highway-Reward volcanic centre is distinct from some other intrusion- and lava-dominated volcanic centres in the Trooper Creek Formation. Chapter 6 evaluates volcanic influences on the formation of iron oxide-silica units in the Trooper Creek Formation. Chapter 7 critically assesses the roles of sub-seafloor replacement versus seafloor accumulation in emplacement of the Highway-Reward deposit. Many of the ores formed by infiltration and replacement of sub-seafloor strata rather than by exhalation onto the seafloor. In Chapter 8, the results of preceding chapters are used to reconstruct the Cambro-Ordovician palaeogeography and volcanic history during accumulation of the Seventy Mile Range Group.
Chapter 2

Regional Geology
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Regional Geology

2.1 Regional geological setting

The Mount Windsor Subprovince (Henderson, 1980) in northern Queensland, Australia is part of the Tasman Fold Belt. The Tasman Fold Belt was an active orogenic region from Cambrian to Cretaceous time. In Queensland, the Tasman Fold Belt comprises the Hodgkinson-Broken River Fold Belt in the north, the Thomson Fold Belt in the central part, and the New England Fold Belt to the south (Fig. 2.1). Murray (1986, 1990) and Wellman (1995) consider the Mount Windsor Subprovince to be part of the Thomson Fold Belt.

The Mount Windsor Subprovince comprises the remnants of a formerly extensive Cambrian to Early Ordovician volcano-sedimentary terrain. The subprovince forms an approximate east-west belt extending for 165 km from the Leichhardt Range in the east, to near Pentland in the west (Fig. 2.2). The subprovince has been extensively intruded along its northern margin by the Ordovician to Permian Ravenswood Batholith and Lolworth Igneous Complex, and is overlain by the Carboniferous Drummond Basin succession to the south (Henderson, 1980, 1986). A discontinuous cover of Tertiary alluvial and ferricrete deposits obscures much of the subprovince.

Together the Mount Windsor Subprovince, Lolworth-Ravenswood Subprovince and the Burdekin Subprovince form the Charters Towers Province (Henderson, 1980, 1986). The Charters Towers Province is separated from the Ordovician Broken River Province to the north by the Clarke River Fault (Fig. 2.1; Murray, 1986). The Mount Windsor Subprovince was first recognised by Wyatt et al. (1971) who incorporated it into the Cape River Beds. Henderson (1980) demonstrated that the Cape River Beds have a different metamorphic and structural history from rocks of the Mount Windsor Subprovince and grouped them with other Proterozoic amphibolite grade metamorphic rocks of the Lolworth-Ravenswood Subprovince. Murray (1990) following earlier interpretations (Wyatt et al., 1971; Murray, 1986) suggested that the Cape River Beds are equivalent in age to, and gradational with, the volcanic rocks of the Mount Windsor Subprovince.

The Lolworth-Ravenswood Subprovince occupies a central position in the Charters Towers Province and includes the Running River and Argentine Metamorphics to the north, the Cape River Beds in the west of the province, and the Charters Towers Metamorphics in its central segment. Devonian to Early Carboniferous marine and
Figure 2.1 Simplified geological map of northern Queensland showing the distribution of elements in the Tasman Fold Belt and major Cambrian-Devonian structural blocks and provinces. Based on Murray (1990), Stolz (1995) and Withnall et al. (1987).
Figure 2.2 Simplified geological map of the Seventy Mile Range Group between Waddy's Mill and Sunrise Spur. The distribution of major geological formations, significant mineral deposits, and locations discussed in the text are illustrated. Modified after Berry et al. (1992).
continentally-derived sedimentary successions of the Burdekin Subprovince on-lap the northern margin of the Lolworth-Ravenswood Subprovince (Henderson, 1980).

The Balcooma Metamorphics, 250 km north of the Mount Windsor Subprovince, comprise volcano-sedimentary units of similar age to the Mount Windsor Subprovince and also host massive sulfide mineralisation (Fig. 2.1; Huston et al., 1992). The two belts have been correlated although the former is metamorphosed to a higher (amphibolite) grade (Henderson, 1986; Withnall et al., 1991). Volcano-sedimentary sequences at the southern end of the Tasman Fold Belt in western Tasmania (Mount Read Volcanics) and in the western part of the Lachlan Fold Belt also show similarities with the Mount Windsor Subprovince and host many world class massive sulfide deposits (Henderson, 1986; Withnall et al., 1991).

2.2 The Seventy Mile Range Group

The volcanic and sedimentary rocks of the Mount Windsor Subprovince are assigned to the Seventy Mile Range Group (Henderson, 1986). As presently understood, the Seventy Mile Range Group can be subdivided into four conformable formations: the Puddler Creek Formation, Mount Windsor Volcanics, the Trooper Creek Formation and the Rollston Range Formation (Figs. 2.2–2.3; Henderson, 1986; Berry et al., 1992). Type sections and fossil localities for each of the formations were presented by Henderson (1983, 1986) and Berry et al. (1992) reassessed the distribution and character of the component lithofacies. The word 'Mount Windsor Volcanics' has often been used as an informal name for the Seventy Mile Range Group (e.g. Stolz, 1995), and is now unworkable as a formation name. In order to avoid confusion, the 'Mount Windsor Volcanics' of Henderson (1986) is referred to in this thesis as the Mount Windsor Formation.

The Seventy Mile Range Group comprises a relatively thick volcano-sedimentary succession (possibly in excess of 12 km; Henderson, 1986). The true thickness of the succession is poorly constrained, due to structural complexities in the Puddler Creek Formation. However, as noted by Stolz (1995), such a thickness may not be unrealistic as more than 10 km of sediments and volcanics have accumulated in the northern Okinawa Trough since the Miocene (Letouzey and Kimura, 1985).
2.2.1 Puddler Creek Formation

The oldest exposed part of the Seventy Mile Range Group is the Puddler Creek Formation, a mixed volcanic and sedimentary succession dominated by poorly sorted, quartz- and lithic-rich sandstone, greywacke and siltstone (Fig. 2.3; Henderson, 1986; Berry et al., 1992). Sandstone beds are generally massive, locally graded, and range from 5 cm to 2.5 m thick (Henderson, 1986). The largely continentally-derived sedimentary component includes quartz, feldspar, phyllite grains, polycrystalline quartz and detrital mica (Henderson, 1986). Andesitic lavas frequently dominate the upper 500 m of the formation (Berry et al., 1992). The lavas record intermittent volcanism prior to eruption of the overlying Mount Windsor Formation (Stolz, 1991; Berry et al., 1992). The lithofacies associations are consistent with a submarine depositional setting, most probably below storm wave base.

Figure 2.3 Generalised stratigraphic column for the Seventy Mile Range Group between Waddys Mill and Sunrise Spur. The location of the main massive sulfide deposits is also shown. Modified after Large (1991).
Mafic dykes and sills become abundant towards the top of the Puddler Creek Formation but do not intrude younger formations. The intrusions are dolerite (Henderson, 1986) and microdiorite (Hartley et al., 1989) and have compositions similar to those of mafic lavas at the top of the formation (Berry et al., 1992). Quartz- and quartz-feldspar-phyric rhyolitic dykes and sills cut across the mafic intrusions (Henderson, 1986) and are interpreted to be feeders to volcanic units in the overlying Mount Windsor Formation and Trooper Creek Formation (Berry et al., 1992). Hartley et al. (1989) considered the mafic and silicic dykes to be similar to dykes that intrude the Charters Towers Metamorphics in the Lolworth-Ravenswood Subprovince.

The base of the formation is not exposed due to widespread granite intrusions of the Ravenswood Batholith. The maximum apparent thickness of the succession is 9 km, assuming no structural repetition (Henderson, 1986; Berry et al., 1992). Contacts between the Puddler Creek Formation and the Mount Windsor Formation are sharp and conformable (Berry et al., 1992) or faulted.

2.2.2 Mount Windsor Formation

The Mount Windsor Formation comprises a thick sequence of rhyolitic volcanic rocks with minor dacite and rare andesite (Henderson, 1986; Berry et al., 1992). Sedimentary rocks are absent except at the base of the formation. Massive coherent and flow-banded rhyolite and autoclastic breccia form lavas and domes which are mostly 100-150 m thick (Berry et al., 1992), but yield little unambiguous information on the depositional environment. In contrast to the remainder of the subprovince, the Mount Windsor Formation in the vicinity of Mount Windsor includes thick intervals of rhyolitic, mass-flow deposited breccia and sandstone (Stolz, 1991). Bedforms in the volcaniclastic facies provide evidence for emplacement in a subaqueous setting (Berry et al., 1992).

The Mount Windsor Formation forms the core to the Seventy Mile Range and is exposed almost continuously between Sunrise Spur in the east and Waddys Mill in the west. The rhyolitic succession is absent in the area north of Highway-Reward where the Puddler Creek Formation is conformably overlain by the Trooper Creek Formation. The Mount Windsor Formation has a maximum thickness of around 3500 m at Sunrise Spur (Fig. 2.2) in the eastern part of the subprovince, decreases in thickness to around 1000 m west of Highway-Reward, and is 300-400 m thick at Waddys Mill in the west (Henderson, 1986; Berry et al., 1992). Henderson (1986) argued that thickening of the Mount Windsor Formation between Highway-Reward and Sunrise Spur indicated an eastern source for the formation.
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The Mount Windsor Formation and the younger Trooper Creek Formation are conformable. The contact was defined by Henderson (1986) as the first appearance of dacite or andesite and associated volcaniclastic and sedimentary facies. This change from dominantly effusive rhyolitic volcanism to deposition of silicic to intermediate lavas and volcaniclastic units also coincides with a decrease in volcanic activity and/or an increase in the rate of sedimentation in the subprovince (Stolz, 1991).

2.2.3 Trooper Creek Formation

The Trooper Creek Formation mainly comprises basaltic, andesitic, dacitic and rhyolitic lavas, intrusions and volcaniclastic rocks, laminated siltstone and mudstone that locally contain graptolite fossils, calcareous metasediment and thin ironstone lenses (Henderson, 1983, 1986; Berry et al., 1992). The bedforms, graptolites, local intercalated pillow lavas, and peperite and hyaloclastite associated with many lavas and intrusions, collectively suggest a submarine setting (Berry et al., 1992; Doyle, 1994). The Thalanga and Highway-Reward massive sulfide deposits and several other prospects are thought to be part of the Trooper Creek Formation. Andesitic dykes and sills in the Trooper Creek Formation and underlying Mount Windsor Formation may represent feeders for units higher in the succession (Berry et al., 1992).

There are regional variations in the proportions of lavas, intrusions, volcaniclastic facies and sedimentary facies in the Trooper Creek Formation and in its thickness. The formation ranges from 4000 m at Mount Windsor in the central part of the subprovince, to a minimum of 500 m along the Thalanga Range to the west. Abundant lavas, syn-sedimentary intrusions and autoclastic breccia, and thickening of the volcanic sequence at Trafalgar Bore and between Mount Windsor and Highway-Reward suggest proximity to major volcanic centres in these parts of the Trooper Creek Formation (Stolz, 1991; Berry et al., 1992; Doyle, 1994). Apparent thickening of the formation in the Sunrise Spur area (Henderson, 1986) and south of Mount Windsor is due to structural repetition (Berry et al., 1992).

The Trooper Creek Formation is conformably and gradationally overlain by the Rollston Range Formation. The transition was defined by Henderson (1986) as the uppermost substantial volcaniclastic unit dominated by pyroclasts.
2.2.4 Rollston Range Formation

The Rollston Range Formation is the youngest exposed part of the Mount Windsor Subprovince and is composed of a thick (at least 1 km) sequence of sandstone and siltstone units and rare dacitic lavas (Henderson, 1986; Berry et al., 1992). Sandstone and siltstone beds are dominated by volcanic components but also contain basement-derived muscovite and phyllite grains. Graptolites and trilobites are known from several locations within the formation and suggest that the base of the formation is older in the west than in the east (Henderson, 1983, 1986). The bedforms and fossils suggest a deep marine depositional environment (Henderson, 1986). The top of the Rollston Range Formation is not exposed, being covered by Tertiary sedimentary formations.

2.2.5 Age relationships

An early Ordovician age for the Trooper Creek Formation is well constrained by radiometric dating on zircons (468±5.4 Ma, 479.7±5.6 Ma; Perkins et al., 1993) and by fossiliferous units within the Trooper Creek Formation and at the base of the Rollston Range Formation (Henderson, 1983). Graptolites and trilobites in the Rollston Range Formation are from the Lancefieldian, Bendigonian and Chewtonian stages of the Early Ordovician epoch. Late Lancefieldian graptolites are the oldest known from the Rollston Range Formation and indicate that the base of the formation is younger in the west than the east. The underlying Mount Windsor Formation and Puddler Creek Formation were interpreted as Late Cambrian by Henderson (1986) on the basis of conformable contacts with lower Ordovician (Lancefieldian) fossiliferous units at the base of the Trooper Creek Formation. Stolz (1995) suggest that the Puddler Creek Formation and Mount Windsor Formation could be of similar age to the Trooper Creek Formation if they were deposited in rapidly subsiding basins. Pb-U dates (474.6±5.1 Ma and 479.1±4.6 Ma) on zircons from the Mount Windsor Formation (Perkins et al., 1993) are consistent with an early Ordovician age. Wyatt et al. (1971) report a whole-rock isochron Rb-Sr age of 528±100 Ma for the Mount Windsor Formation.

2.3 Metamorphic grade

The Seventy Mile Range Group has been affected by low to medium grade regional metamorphism (Hutton et al., 1993; Henderson, 1986; Berry et al., 1992). The metamorphic grade is prehnite-pumpellyite facies around Trooper Creek prospect in the east and increases to greenschist facies in the central part of the subprovince near Highway-Reward (Berry et al., 1992). To the west around Trafalgar Bore and Thalanga,
2.9 Regional Geology

Metamorphic grade increases through actinolite, hornblende, biotite and andalusite isograds to amphibolite facies at Waddys Mill. The regional metamorphic assemblage is correlated with deformation event D3 and regional S3 cleavage development (Berry, 1991). In contrast to rocks in the east, the metamorphic assemblage at Waddys Mill is characterised by higher strain, possibly reflecting recrystallisation in response to simultaneous deformation and granitoid emplacement.

In the east, the syn-deformational early regional metamorphic assemblage has been overprinted by hornblende hornfels assemblages which form contact metamorphic aureoles around post-kinematic granitoids of the Lolworth-Ravenswood Batholith (Berry, 1991; Berry et al., 1992). Contact metamorphic aureoles are less prominent in the western part of the subprovince (including the study area) even though post-orogenic granites are common (Fig. 2.2). Because the regional metamorphic grade was higher in the west, the effects of contact metamorphism were diminished and growth of prograde minerals was restricted to granite margins (Berry, 1991).

2.4 Regional deformation and structure

A complex deformational history is recorded in Seventy Mile Range Group (Henderson, 1986; Laing, 1988; Gregory et al., 1990; Berry, 1989; Berry, 1991; Berry et al., 1992). Many pumiceous rocks preserve relics of an early bedding-parallel foliation (S1) which is defined by compacted, phyllosilicate-altered pumice. S1 is interpreted as a diagenetic compaction fabric or early tectonic fabric, and has not previously been recognised in the subprovince. The foliation is similar to that documented in other massive sulfide districts (e.g. Allen and Cas, 1990; McPhie et al., 1993; Allen et al., 1996b). The structural nomenclature of Berry et al. (1992) has been modified to include the S1 foliation. Three main deformations are recognised in the Seventy Mile Range Group between Waddys Mill and Sunrise Spur (Fig. 2.2; Berry et al., 1992). The first (D2) comprises thrusts (now wrench faults) and associated F2 folds and cleavage (S2). These structures are overprinted by the regional S3 cleavage and upright east-west F3 folds associated with D3. A second strong cleavage (S4) has been recognised in the subprovince but is only locally developed (Berry et al., 1992). S4 often grades with increasing intensity into D4 faults. F4 folds are locally developed.

Berry (1991) and Berry et al. (1992) interpreted thickening of units in the Mount Windsor Formation and Trooper Creek Formation to reflect syn-depositional extensional faulting. They suggested that a listric growth fault forms the boundary between the Puddler Creek Formation and Trooper Creek Formation in the area between Mount Farrenden and Truncheon. The fault south of Mount Farrenden is complicated by short-wavelength open
folding. Small growth faults were interpreted to occur at Oakvale prospect, Leopardtown prospect, Mt Windsor, Highway East prospect and Trooper Creek (Berry, 1991). Mapping of the current study suggests there is little evidence for a growth fault at Highway East prospect.

Variations in bedding and S3 cleavage intersections and F3 fold plunges throughout the Seventy Mile Range Group imply that beds were steeply dipping and folded prior to the regional S3 cleavage (Berry et al., 1992). The F2 folds are variable in form and many are closely associated with moderate to steeply south-dipping D2 faults which are bedding-parallel. D2 faults are folded about F3 so that although they were formed as thrusts the faults now have a wrench movement. The S2 cleavage is restricted to narrow zones along the faults. Berry (1989) suggested that the early thrust-related fold event correlates with Ordovician folding and thrusting in provinces to the north.

Volcano-sedimentary rocks of the Seventy Mile Range Group form the subvertical, south-facing limb of an east-west trending F3 fold (D3). The north-facing limb is exposed only at Waddys Mill (Berry et al., 1992). Short wavelength (100-500 m) parasitic F3 folds are common in some parts of the subprovince. The dominant regional cleavage (S3) is axial planar to F3 folds (Henderson, 1986; Gregory et al., 1990; Berry et al., 1992) and overprints S1. Bedding-S3 cleavage intersections are highly variable in orientation but mostly have moderate to shallow plunges (Berry et al., 1992). At Waddys Mill, the strongly foliated hinge of a major F3 syncline plunges moderately east. The age of the D3 event is equivocal. The orientation of F3 folds is similar to folds of Devonian age in the Hodgkinson and Broken River Provinces to the north (Bell, 1980; Berry, 1989, 1991; Berry et al., 1992). East-west trending faults within the Ravenswood Batholith are interpreted as equivalent to S3 in the Seventy Mile Range Group (Berry et al., 1992). Henderson (1986) related the D3 fold phase to intrusion of the Ravenswood Batholith.

The S3 slaty cleavage is locally overprinted by a weaker S4 spaced cleavage spatially associated with steep south-side up faults striking 040° to 080° (Berry et al., 1992). In many cases the intersection of S3 and S4 forms a moderately-steeply plunging lineation and a “diamond” pattern in outcrop. The S4 cleavage at Thalanga is associated with F4 folds. The F4 folds are spatially, and probably genetically related, to unfoliated granitoids of the Ravenswood Batholith (Berry, 1989). At Highway-Reward, S4 forms a northeast trending zone 2 km wide associated with the Truncheon Fault (syn-D4).

The D4 cleavage and faults have orientations similar to regional lineaments including the Mount Leyshon Corridor (Berry, 1991; Berry et al., 1992). The Mount Leyshon Corridor is interpreted as a major structure that controlled Carboniferous to Permian magmatism in the Mount Leyshon area, and is thought to be Carboniferous (Morrison, 1988; Hartley et
al., 1993). Carboniferous magmatism in the corridor generated the Mount Leyshon gold deposit (Wormald et al., 1993). Other lineaments were interpreted by Hartley et al. (1993) as pre-Ordovician to Ordovician in age.

The youngest structures in the Seventy Mile Range Group disrupt earlier deformation patterns (Berry et al., 1992). Minor late wrench faults occur throughout the Seventy Mile Range Group. Faults are at a high angle to stratigraphy and are associated with narrow cleaved domains. In addition there are north-striking dextral faults, steep west-plunging folds related to late kinking, and long wavelength (10 km) south-plunging folds, which post-date S4. A major fault extending west from Trooper Creek cuts off the Truncheon Fault (syn-D4) near Truncheon (Berry et al., 1992).

Structure of the Highway-Reward area

In the eastern part of the study area, bedding dips and faces moderately (30–60°) to the SSW and is consistent with regional trends. The regional S3 spaced cleavage is weak in this region and overprinted by S4. In many cases the intersection of S3 and S4 forms a moderately-steeply plunging lineation and a “diamond” pattern in outcrop. S4 (040–080°) has a brittle-ductile style with narrow zones of strong cleavage development separated by wider weakly cleaved zones (cf. Berry et al., 1992). At Highway East, variations in bedding orientation mark the position of an F4 anticline which plunges moderately to the southwest (Map 1; cf. Berry et al., 1992). The expected matching syncline is cut out by the northeast striking Truncheon Fault. The Truncheon and North Truncheon Faults (syn-D4) offset the contact between the Trooper Creek Formation and underlying Mount Windsor Formation, by at least 1 km. Movement on the fault is sinistral. S4 is strongly developed in a 2 km wide zone parallel to the Truncheon fault (syn-D4). This is the Policemen Creek Fault Zone of Henderson (1986). The Highway-Reward massive sulfide deposit is located within this zone, west of the Truncheon Fault. At Highway-Reward, the massive sulfide bodies have been fractured and diluted by D4, whereas the alteration halo has sheared parallel to the subvertical S4 cleavage. Structural readings from drill core suggest that beds dip (around 10°) and face south east in the subsurface volcanics (cf. Laing, 1988). Flow banding in rhyolite from the Highway open pit has steeper dips (18–55°) but may be at a high angle to bedding. The host succession is cut by steep reverse faults which parallel the Truncheon Fault. North of Highway-Reward (around Handcuff), the pattern of bedding is complex and consistent with a small syncline which plunges moderately to the southeast. The western limb of the syncline is cut off by the Handcuff Fault, which is in turn offset by the Truncheon and North Truncheon Faults (Berry et al., 1992). West of Highway-Reward, there is insufficient mapping to confidently outline the structure. Where exposed, beds dip and face northwest or dip and face south (Map 1).
2.5 Tectonic setting of the Mount Windsor Subprovince

Models for the tectonic evolution of the Seventy Mile Range Group were presented by Henderson (1986) and Stolz (1994, 1995). Henderson (1986) interpreted the Seventy Mile Range Group as the fill of a Cambro-Ordovician back-arc basin which developed on stretched Precambrian continental lithosphere west of a continental margin volcanic arc. Nd isotopic studies by Stolz (1995) supported this interpretation and suggested that the volcanic rocks were derived by partial melting of the older crust. The thickness of the calc-alkaline volcanic and sedimentary succession (possibly greater than 12 km; Henderson, 1986) suggests syn-depositional basin subsidence accompanied accumulation (Stolz, 1995). Quartz-rich micaceous sedimentary rocks of the Puddler Creek Formation are interpreted to comprise sediment derived from erosion of a Precambrian source on the attenuated Late Proterozoic to Lower Precambrian margin of northeastern Australia (Stolz, 1995).

During the early Cambrian, westerly dipping subduction of oceanic crust was initiated beneath the passive margin (Stolz, 1995). Modification of the mantle wedge by dehydration of the subducting slab initiated continental margin volcanism (Stolz, 1995). Extension of this arc, possibly due to trench retreat, subsequently initiated the development of a back-arc basin. Rapid accumulation of Puddler Creek Formation sediments within the subsiding basin preceded the eruption of a small volume of basalt and andesite (Stolz, 1995). The increased thermal gradient is interpreted to have generated voluminous crustal melts which were subsequently erupted as the Mount Windsor Formation rhyolite and dacite (Henderson, 1986; Stolz, 1995). This phase of magmatism was complete prior to initiation of Trooper Creek Formation volcanism from mantle sources variably modified by subduction.

Volcanism subsided during the early Ordovician and deposition of the Rollston Range Formation proceeded mostly by reworking of older Trooper Creek Formation volcanioclastic material. The formation may also contain detritus sourced from an active arc (Henderson, 1986; Stolz, 1995).

The orientation of the arc and adjacent back arc basin is equivocal. Henderson (1986) argued that the arc and back-arc basin were oriented north-south between the palaeosubduction zone to the west and a continental province to the east. Within the Puddler Creek Formation, an increase in volcanic components compared with continentally derived sediment moving from west to east supports Henderson’s (1986) model. The current east-west orientation of the Seventy Mile Range Group was regarded by Henderson (1986) as the result of disruption and folding during emplacement of the Ravenswood Batholith. However, observations from the Puddler Creek Formation are
also consistent with a northeast-southwest-oriented arc that was aligned parallel to the Precambrian cratonic margin (Stolz, 1995).

2.6 VHMS deposits in the Seventy Mile Range Group

The Seventy Mile Range Group contains a number of significant volcanic-hosted massive sulfide (VHMS) deposits. After Thalanga, the Highway-Reward deposit is the largest known deposit within the succession. There have been few studies (Gregory and Hartley, 1982; Gregory et al., 1987, 1990; Kay, 1987; Beams et al., 1989; Beams et al., 1990; Hill, 1991, 1993, 1996; Hutton et al., 1993; Doyle, 1994; Doyle and McPhie, 1994; Herrmann, 1995; Huston et al., 1995) of the physical, mineralogical and geochemical properties of mineralisation, alteration and volcanic host rocks in this VHMS district. Berry et al. (1992) provide a synthesis of the current understanding of deposits in the subprovince (Table 2.1). Mineralisation at Thalanga, Liontown, Handcuff and many smaller occurrences is mostly as stratiform sulfide lenses which are typically zinc-rich but contain little copper. At Highway-Reward, small stratiform Zn-Pb-Cu massive sulfide lenses occur marginal to sub-vertical Cu-Au-rich pyrite pipes.

There is a strong stratigraphic control on mineralisation in the Seventy Mile Range Group. The Thalanga deposit occurs within the Trooper Creek Formation at the contact with the underlying Mount Windsor Formation. Other VHMS-style deposits and prospects, including Highway-Reward and Liontown, are confined to the Trooper Creek Formation (Table 2.1).

2.7 Ravenswood Batholith

The Ravenswood Batholith (Ravenswood Granodiorite of Clarke and Paine, 1970) intrudes the early Palaeozoic (Cambrian?) Charters Towers Metamorphics and the Seventy Mile Range Group (Wyatt et al., 1970; Hutton et al., 1993). Rhyolitic, dacitic and andesitic units in the Mount Windsor Formation and Trooper Creek Formation may be comagmatic with a Late Cambrian to Early Ordovician phase of the batholith (Table 2.2; Hutton et al., 1990; Hutton and Crouch, 1993). Leucocratic granitoid stocks that intrude the Puddler Creek Formation along contacts with the Ravenswood Batholith are considered components of the latter (Levingston, 1981). Henderson (1986) interpreted the stocks to be comagmatic with quartz-feldspar porphyritic dykes in the Puddler Creek Formation and both as hypabyssal correlatives of the Mount Windsor Formation.
<table>
<thead>
<tr>
<th>Deposit</th>
<th>Grade and tonnage</th>
<th>Geometry</th>
<th>Ore mineralogy</th>
<th>Host lithologies</th>
<th>Stratigraphic position</th>
</tr>
</thead>
<tbody>
<tr>
<td>Thalanga</td>
<td>primary 7.5 m.t. @ 1.6% Cu, 9.3% Zn, 3% Pb, 77 g/t Ag, 0.4 g/t Au</td>
<td>sheetforming</td>
<td>sphalerite, pyrite, galena, chalcopyrite, minor molybdenite, zincoenstatite</td>
<td>Hanging wall dacitic and andesitic lavas, minor andesite</td>
<td>Within the Trooper Creek Formation</td>
</tr>
<tr>
<td></td>
<td>supergene 0.067 m.t. @ 5.8% Cu, 8.5% Zn, 2.1% Pb, 83 g/t Ag, 1.7 g/t Au paragenesis</td>
<td>pyrite, (Cu-Au) pipe with marginal underground Pb-Zn lenses</td>
<td>sphalerite, pyrite, galena; minor chalcopyrite, tennantite; trace electrum</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>ozide 0.184 m.t. @ 9% Ag, 0.8 g/t Au (proven or probable)</td>
<td>pyrite, (Cu-Au) pipe</td>
<td>pyrite, chalcopyrite, sphalerite, galena; minor pyrite, tennantite; trace electrum</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Reward</td>
<td>primary 0.2 m.t. @ 3.5% Cu, 13% Ag, 1 g/t Au</td>
<td>pyrite (Cu-Au) pipe</td>
<td>pyrite, chalcopyrite, sphalerite, galena; minor chalcopyrite, tennantite</td>
<td>dacytic to rhyolitic lavas, syn-sedimentary intergrowths, pseudobrecia-sandstone, siltstone, polymeric breccia</td>
<td>Upper Trooper Creek Formation</td>
</tr>
<tr>
<td></td>
<td>supergene 0.3 m.t. @ 11.0% Cu, 21 g/t Ag, 1.8 g/t Au</td>
<td>pyrite (Cu-Au) pipe with marginal underground Pb-Zn lenses</td>
<td>pyrite, chalcopyrite, sphalerite, galena; minor chalcopyrite, tennantite</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>oxide 0.1 m.t. @ 33 g/t Ag, 6.9 g/t Au</td>
<td>pyrite (Cu-Au) pipe</td>
<td>pyrite, chalcopyrite, sphalerite, galena; minor chalcopyrite, tennantite</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Hayford</td>
<td>primary 1.2 m.t. @ 5.5% Cu, 6.5% Ag, 1.2 g/t Au</td>
<td>pyrite (Cu-Au) pipe</td>
<td>pyrite, chalcopyrite, sphalerite, galena; minor chalcopyrite, tennantite</td>
<td>dacytic to rhyolitic lavas, syn-sedimentary intergrowths, pseudobrecia-sandstone, siltstone, polymeric breccia</td>
<td>Upper Trooper Creek Formation</td>
</tr>
<tr>
<td></td>
<td>oxide 0.07 m.t. @ 6.0 g/t Au</td>
<td>pyrite (Cu-Au) pipe</td>
<td>pyrite, chalcopyrite, sphalerite, galena; minor chalcopyrite, tennantite</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Lisburne</td>
<td>primary 1.0 m.t. @ 0.4% Cu, 0.2% Pb, 7.4% Zn, 8.8 g/t Ag, 0.3 g/t Au (inferred)</td>
<td>veins, lenses</td>
<td>sphalerite, pyrite, minor chalcopyrite, galena</td>
<td>Hanging wall dacitic pumice breccia, syn-sedimentary intergrowths</td>
<td></td>
</tr>
<tr>
<td></td>
<td>oxide 0.07 m.t. @ 6.0 g/t Au</td>
<td>sphalerite, pyrite, minor chalcopyrite, galena</td>
<td>Hanging wall dacitic pumice breccia, syn-sedimentary intergrowths</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Waterloo</td>
<td>primary 0.373 m.t. @ 3.8% Cu, 19.7% Zn, 2.8% Pb, 59 g/t Ag, 2.0 g/t Au (inferred)</td>
<td>lenses</td>
<td>sphalerite, pyrite, chalcopyrite, minor galena, tennantite; disseminated zinc sulfides</td>
<td>Hanging wall dacitic-dacitic volcaniclastics rocks and lavas; minor andesite; sandstone-siltstone</td>
<td>Central Trooper Creek Formation</td>
</tr>
<tr>
<td></td>
<td>oxide 0.07 m.t. @ 6.0 g/t Au</td>
<td>sphalerite, pyrite, chalcopyrite, minor galena, tennantite; disseminated zinc sulfides</td>
<td>Hanging wall dacitic-dacitic volcaniclastics rocks and lavas; minor andesite; sandstone-siltstone</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Agincourt</td>
<td>primary 2.0 m.t. @ 0.5% Cu, 6.6% Zn, 2.3% Pb, 59 g/t Ag, 0.0 g/t Au (inferred)</td>
<td>disseminated; breccia lenses</td>
<td>sphalerite, pyrite, galena, chalcopyrite, minor tennantite</td>
<td>Hanging wall dacitic-dacitic volcaniclastics rocks and lavas; minor andesite; sandstone-siltstone</td>
<td>Central Trooper Creek Formation</td>
</tr>
<tr>
<td></td>
<td>oxide 0.07 m.t. @ 6.0 g/t Au</td>
<td>sphalerite, pyrite, galena, chalcopyrite, minor tennantite</td>
<td>Hanging wall dacitic-dacitic volcaniclastics rocks and lavas; minor andesite; sandstone-siltstone</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Warramoe</td>
<td>primary 0.25 m.t. @ 2% Cu, 15% Zn, 2% Pb, 30 g/t Ag, 1 g/t Au (inferred)</td>
<td>multiple lenses</td>
<td>sphalerite, galena, chalcopyrite</td>
<td></td>
<td>Trooper Creek Formation</td>
</tr>
<tr>
<td></td>
<td>oxide 0.07 m.t. @ 6.0 g/t Au</td>
<td>multiple lenses</td>
<td>sphalerite, galena, chalcopyrite</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Magpie</td>
<td>primary 0.25 m.t. @ 2% Cu, 15% Zn, 2% Pb, 30 g/t Ag, 1 g/t Au (inferred)</td>
<td>stacked lenses</td>
<td>sphalerite, pyrite, chalcopyrite, galena; minor pyrite, tennantite, magnetite, marcasite</td>
<td>Hanging wall dacitic and andesitic lavas and volcanioclastic rocks</td>
<td>Central Trooper Creek Formation</td>
</tr>
<tr>
<td></td>
<td>oxide 0.07 m.t. @ 6.0 g/t Au</td>
<td>stacked lenses</td>
<td>sphalerite, pyrite, chalcopyrite, galena; minor pyrite, tennantite, magnetite, marcasite</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Age of intrusion</td>
<td>Composition</td>
<td>Associated volcanism</td>
<td>Deformation</td>
<td>Comments</td>
<td></td>
</tr>
<tr>
<td>-----------------</td>
<td>-------------</td>
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<td>-------------</td>
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<td></td>
</tr>
<tr>
<td>Late Cambrian to Early Ordovician (490±6 Ma)</td>
<td>Granodiorite</td>
<td>Co-magmatic with Mount Waddington Volcanics or Trooper Creek Formation</td>
<td>Recrystallised, deformed and cleaved (S3)</td>
<td>Coal mined in vicinity of the granodiorite.</td>
<td></td>
</tr>
<tr>
<td>Early to Mid Ordovician</td>
<td>Granite to granodiorite</td>
<td>Associated mafic intrusions</td>
<td>Recrystallised, foliated</td>
<td>Separate magma compared with other Middle Ordovician granitoids.</td>
<td></td>
</tr>
<tr>
<td>Middle Silurian to Early Devonian (426±4 to 406±4 Ma)</td>
<td>Calc-alkaline hornblende-biotite granodiorite to tonalite; possible gabbro-diorite</td>
<td>Co-magmatic with Mount Windsor Volcanics or Trooper Creek Formation</td>
<td>Unfoliаted; localized ductile deformation; foliation at margins</td>
<td>Lineament-controlled narrow component plutons.</td>
<td></td>
</tr>
<tr>
<td>Carboniferous to Permian</td>
<td>Carboniferous to Permian</td>
<td>(1) Gabbro to granite; (2) Granite to granodiorite</td>
<td>Brittle faults and fractures</td>
<td>Brittle faults and fractures.</td>
<td></td>
</tr>
<tr>
<td>(3) Rhyolite &amp; trachyte</td>
<td>(3) Rhyolite &amp; trachyte</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>
The batholith comprises numerous plutons with different geochemistry and ages, but is mostly adamellite and granodiorite (Wyatt et al., 1971; Levingston, 1981; Hutton and Crouch, 1993). Henderson (1986) identified two parts to the batholith: biotite tonalite west of Charters Towers, and hornblende granodiorite to the east.

Both pre- and post-orogenic granitoids are identifiable in the Ravenswood Batholith (Levingston, 1981; Hartley et al., 1989; Berry, 1989, 1991; Berry et al., 1992). Strongly foliated (S3) granodiorite and tonalite intruded prior to the regional, Mid to Late Ordovician deformation and are Ordovician in age. Undeformed post-orogenic granitoids are Late Devonian in age or younger. Recent mapping and isotopic dating of the Ravenswood Batholith (Hutton et al., 1990; Hutton and Crouch, 1993) indicate that intrusion occurred in four phases (Table 2.2). Ninety four percent of the batholith was emplaced between the Early Ordovician and Early Devonian. The remainder of the pluton intruded in Carboniferous to Early Permian time. Permian-Carboniferous granitoids occur near the eastern and southeastern margins of the batholith and are divisible into three groups (Table 2.2; Hutton et al., 1993; Hutton et al., 1994). The mineralised subvolcanic intrusive breccia complex at Mount Leyshon is associated with rhyolitic and quartz-trachyte plugs recording one of these intrusive phases (Hutton et al., 1993, 1994).

2.8 Lolworth Igneous Complex

The Precambrian basement and Ravenswood Batholith are intruded by the Lolworth Igneous Complex north and west of Thalanga (Paine et al., 1971). The complex is unfoliated and comprises adammellite, biotite granodiorite, muscovite granite and minor quartz diorite (Clarke and Paine, 1970; Paine et al., 1971; Vine and Paine, 1974; Levingston, 1981). A post-orogenic, Devonian age is suggested by K-Ar (390 Ma) and Rb-Sr (404 Ma) isochrons (Webb, 1970a,b, 1971) and by the absence of a penetrative foliation (Paine et al., 1971; Levingston, 1981).

2.9 Tertiary sedimentary sequences

Extensive areas of the Seventy Mile Range Group are covered by the Pliocene Campaspe Formation. Thicknesses of up to 120 m have been measured (Henderson and Nind, 1994) with values in the Highway-Handcuff area of 1 m adjacent to outcrop areas and up to 50 m in deep palaeodrainage channels (Beams and Jenkins, 1995).

The formation mostly consists of sandstone but occasional interbeds of siltstone, mudstone and rare conglomerate occur (Grimes, 1980; Henderson and Nind, 1994).
Grains are texturally immature, angular to subrounded and show little evidence of significant reworking. Cross-laminated sandstone beds have been interpreted as fluvial traction current deposits in a braided stream environment (Henderson and Nind, 1994). Medium to thick (0.5-3 m) sandstone beds are massive, poorly- to very-poorly sorted, silt/mud matrix-supported, and display evidence for substrate erosion, suggesting deposition from debris flows (Henderson and Nind, 1994). Ferruginous palaeosols are locally developed.

2.10 Summary

The Highway-Reward deposit and several other VHMS deposits including Thalanga are located within the Mount Windsor Subprovince of northern Queensland, Australia. The subprovince is a Late Cambrian to Early Ordovician volcano-sedimentary terrain in the Thomson Fold Belt, a poorly preserved segment of the Tasman Fold Belt. Volcanic and sedimentary rocks within the subprovince form the Seventy Mile Range Group (Henderson, 1986). As presently understood, the Seventy Mile Range Group comprises the following major lithostratigraphic units: the Puddler Creek Formation, Mount Windsor Formation, Trooper Creek Formation and the Rollston Range Formation. The study area contains the main lithostratigraphic units but is largely confined to the Trooper Creek Formation and Mount Windsor Formation. The Seventy Mile Range Group was deformed during three main events of equivocal age. Widespread faulting, folding and associated cleavage development occurred during these events. Evidence for D3 and D4 is recorded in the Highway-Reward to Trooper Creek area. The Seventy Mile Range Group has also been subject to diagenetic alteration and compaction, prehnite-pumpellyite to lower greenschist facies metamorphism and local hydrothermal alteration and mineralisation. The sequence has been intruded by Ordovician to Permian granitoids of the Ravenswood Batholith and Lolworth Igneous Complex. Contact metamorphic aureoles are locally developed along the margins of the granitoids, but not within the study area.
Chapter 3

Stratigraphy of the Seventy Mile Range Group: an evaluation of stratigraphic subdivisions and lithofacies
3.1 Introduction

Although the stratigraphic control on mineralisation in the Seventy Mile Range Group is strong, little is known about the volcanology and sedimentology of this succession. Systematic regional mapping concentrating on recognition of distinctive volcanic facies and facies associations as a means of reconstructing the Cambro-Ordovician facies architecture of the entire Seventy Mile Range Group has yet to be undertaken. The research reported here suggests that this approach will prove invaluable in regional correlation and mineral exploration.

A detailed analysis of the Mount Windsor Formation and Trooper Creek Formation, involving geological mapping and drill core logging, has been undertaken in the area between Coronation homestead and Trooper Creek prospect (Fig. 3.1; approx 15 km strike length). This research builds on earlier studies by Henderson (1986) and Berry et al. (1992) and reinterprets the position and nature of contacts between some of the formations and clarifies the character of the lithofacies. A new formal subdivision of the Trooper Creek Formation is proposed, based on mappable compositional and lithological variations, which primarily reflect changing provenance. Much of the discussion relates to a 1:25000 scale map of the area (Map 1) which accompanies the thesis.

In the study area, the Seventy Mile Range Group includes both syn-eruptive and post-eruptive volcaniclastic facies. Syn-eruptive facies comprise clasts that were initially created, transported and deposited by volcanic processes, but which were rapidly resedimented during (or shortly after) volcanic eruption but not significantly reworked (McPhie et al., 1993). In contrast, post-eruptive volcaniclastic facies consist of particles derived by weathering, erosion and/or resedimentation of pre-existing volcanic deposits, and involve significant sedimentary transport and reworking (McPhie et al., 1953). Post-eruptive deposits can be dominated by particles derived by surface weathering and erosion of volcanic rocks (epiclasts) or by pyroclasts and autoclasts that were resedimented long after eruption. Syn- and post-eruptive volcaniclastic facies can in reality be difficult to distinguish, especially for resedimented pyroclastic deposits. Resedimentation can conceivably be significantly post-eruptive and still generate pyroclast-rich deposits that appear syn-eruptive. Thick (several tens or hundreds of metres) pyroclast-rich mass-flow units provide the most distinctive examples of syn-eruptive resedimented volcaniclastic
Figure 3.1 Simplified geological map showing the distribution of formations and members within the Seventy Mile Range Group in the area between Coronation homestead in the west and Trooper Creek prospect in the east.
Figure 3.2 Measured sections along A-A', B-B' and C-C' in Figure 3.1 showing the three major lithofacies associations identified. True thickness shown. PCF = Puddler Creek Formation, MWF = Mount Windsor Formation, KHM = Kitchenrock Hill Member of the Trooper Creek Formation, HM = Highway Member of the Trooper Creek Formation, RRF = Rollston Range Formation.
deposits (e.g. McPhie and Allen, 1992). These units can resemble primary pyroclastic flow deposits, but lack evidence for hot gas-supported transport and emplacement and complete sedimentation units have different internal organisation (Chapter 4). The presence of epiclasts, textural evidence for reworking (e.g. particle rounding and good sorting) and general division into multiple, relatively thin sedimentation units are key criteria for distinguishing post- from syn-eruptive volcaniclastic deposits (e.g. White and McPhie, 1996). However, syn-eruptive volcaniclastic mass flows can incorporate epiclasts during resedimentation and in some environments, resedimentation processes may result in significant particle rounding, complicating the interpretation.

3.2 Puddler Creek Formation

Exposures of the Puddler Creek Formation are limited to one small (250 m) area to the northeast of Truncheon prospect (Fig. 3.1, Fig. 3.2 - section A) and have received only cursory attention. At this locality, the Puddler Creek Formation comprises interbedded non-volcanic siltstone and pale green to brown sandstone overlain by a thick (50 to 150 m) interval of coherent andesite. A poorly exposed monomictic, matrix-poor, clast-supported breccia occurs near the base of the andesite. The breccia comprises blocky to irregular, variably vesicular clasts. Clasts show some variation in phenocryst abundance suggesting mixing of clast types during emplacement of the breccia. Clast shapes reflect the importance of autobrecciation and cooling-contraction granulation during fragmentation, but clasts were resedimented following fragmentation. Upper contacts of the andesite are not exposed. However, the underlying breccia facies suggest that the andesite is a lava. Sandstone beds (centimetres to 2 m thick) are massive, locally graded, and composed of Precambrian derived detritus (principally subrounded to subangular quartz, minor feldspar and lithic fragments). Siltstone beds are strongly cleaved, micaceous and form horizons up to 4 m in thickness.

Rhyolite, dacite, andesite and dolerite dykes are abundant in this part of the Puddler Creek Formation. In this area, the Puddler Creek Formation is at least locally in faulted contact with the Mount Windsor Formation. The remaining contacts are not exposed.

3.3 Mount Windsor Formation

The Mount Windsor Formation was defined by Henderson (1986) as a thick rhyolitic volcanic succession, with minor dacite and rare andesite, devoid of intercalated sedimentary rocks except at its base. Within the study area, coherent massive and flow banded, quartz- and feldspar-phyric rhyolite dominate the Mount Windsor Formation and
dacite is a subordinate but significant component of the formation. Autoclastic breccia facies are relatively minor and primary pyroclastic rocks are not present.

3.3.1 Massive and flow banded rhyolite, rhyodacite and dacite

This facies is characterised by an even distribution of euhedral quartz and/or feldspar phenocrysts. The mineralogy and abundance of phenocrysts is uniform within a single unit. These properties have been used to map different units in the field and are a rough indication of their chemical composition. The rhyolites are characterised by 0.5-7% quartz phenocrysts (1 to 7 mm across) and subordinate alkali and plagioclase feldspar phenocrysts (7%), 1-3 mm long. The rhyodacites contain 7% feldspar phenocrysts, 1-1.5 mm across and subordinate quartz phenocrysts (3%, 0.5-1 mm). Dacitic lavas and intrusions are aphyric or contain 3-5% euhedral feldspar phenocrysts, 1-3 mm across. Geochemically, some rocks mapped as rhyolite plot as dacite and visa versa (Section 3.8.3). Quartz phenocrysts are round and embayed. Microphenocrysts of apatite and zircon are common accessory minerals. The groundmass was presumably originally glassy and has devitrified to a fine-grained mosaic of quartz and feldspar, or else has been altered to various assemblages of albite, chlorite, quartz or sericite. In some samples, the groundmass includes variably recrystallised spherical spherulites. Relic perlitic crack patterns suggest that parts of the groundmass in many units was formerly glassy (e.g. Allen, 1988; Ross and Smith, 1955; Friedman et al., 1966). Parts of some units are characterised by highly contorted flow banding (Fig. 3.3A). The flow foliations consist of pale siliceous bands alternating with darker, more phyllosilicate-rich bands or pinkish albite bands. The siliceous bands are composed of a quartzofeldspathic mosaic. Albite- or phyllosilicate-rich bands are probably an alteration of former glass.

Origin and significance of facies

Textural evidence and contact relationships favour interpretation of the rhyolite, rhyodacite and dacite intervals as coherent facies of lavas and/or shallow intrusions. Many of the units have been previously mapped as pyroclastic rocks (e.g. Johnson, 1991). Densely welded ignimbrite can also be perlitic, spherulitic or flow banded and could resemble the rhyolite, rhyodacite and dacite intervals. However, such an origin through dense welding of an ignimbrite can be discounted by evidence including: (1) the abundance of unbroken, evenly distributed phenocrysts, and absence of lithic clasts; and (2) absence of vertical or lateral variations in grain size or welding. Moreover, in some cases the flow banded rocks have been misinterpreted as bedded volcaniclastic rocks and apparent clastic textures have been generated by post-depositional processes including devitrification, hydration, hydrothermal alteration and regional greenschist metamorphism (cf. Allen, 1988; McPhie et al., 1993; Doyle et al., 1993).
Figure 3.3

Representative lithofacies from the Mount Windsor Formation (A-B), the Kitchenrock Hill Member of the Trooper Creek Formation (C-E), the overlying Highway Member (Trooper Creek Formation; F) and the Rollston Range Formation (G-H).

(A) Contorted flow banding in coherent rhyolite. 7747300 mN, 420500 mE.

(B) Monomictic rhyolitic breccia facies. This autobreccia is clast-supported and composed of blocky to slabby rhyolite clasts (arrow). Flow banding in the coherent facies is continuous into the autobreccia. 7747300 mN, 420500 mE.

(C) Massive to graded polymictic breccia and sandstone facies. The breccia is massive, clast-supported and comprises aphyric dacite clasts, 2 to 50 cm across. Some clasts are flow laminated. Clasts in the breccia vary from angular and blocky to well rounded. Intense silicification has modified groundmass textures within clasts and obscured clast margins. 7746550 mN, 421200 mE.

(D) Massive to graded polymictic breccia and sandstone facies. Large clasts of silicified, flow laminated aphyric dacite (d) up to 1 m across are the most conspicuous component in this breccia. These clasts are perlitic and spherulitic. They are supported in a strongly cleaved matrix of smaller clasts. Most of the smaller clasts are also dacite but some rhyolite clasts are present. Clasts vary from angular to subrounded in shape.

(E) Polymictic volcanic breccia facies. The unit is massive, matrix-poor and clast-supported. Clasts in the breccia include cherty siltstone, planar laminated siltstone and rhyodacite. These clasts have angular blocky shapes. The other constituent is angular to subrounded coarsely quartz- and feldspar-phyric rhyolite clasts (r). Rhyolite clasts range in size from 10 cm to 10 m across. The facies is interpreted as a sediment gravity flow deposit. 7746100 mN, 422000 mE.

(F) Rounded lithic-crystal sandstone facies. Discontinuous lenses of rounded to angular, lithic granules and pebbles (arrow) occur within the sandstone. Clasts are chert, phyllite and quartz-hematite fragments. 7741800 mN, 426700 mE.

(G) Rounded lithic-crystal sandstone facies. Thin beds of massive to weakly graded, crystal-lithic sandstone alternate with laminated siltstone beds. The sandstone beds are interpreted as turbidites. The siltstone beds were formed by suspension sedimentation. Hammer for scale (arrow). 7745000 mN, 420100 mE.

(H) Rounded lithic-crystal sandstone facies. The grain population in this sandstone is diverse but is dominated by quartz (q) and feldspar (f). The other components are detrital biotite, phyllite (p), apatite and well-rounded zircon and tourmaline grains. Stocksqaud, RD 813 - 95.9 m.
3.3.2 Monomictic breccia facies

These breccias are monomictic, poorly sorted, clast-supported and composed of evenly porphyritic (quartz and/or feldspar), non-vesicular dacite, rhyodacite or rhyolite clasts (Fig. 3.3B). Clasts have blocky to slabby shapes with planar and curvilinear to finely jagged margins. The groundmass of clasts can be perlitic, spherulitic, devitrified to an interlocking mosaic of quartz and feldspar, or altered to various assemblages of chlorite, sericite and quartz. The breccias contain small amounts of matrix, comprising cuneiform dacite, rhyodacite or rhyolite fragments and parts of crystals. In some cases, preferential quartz alteration of the matrix and margins of clasts has generated a more extensive apparent matrix domain.

Two different breccia types are identifiable. In the first, clasts in the breccia fit more or less together (jigsaw-fit texture). In the second, jigsaw-fit texture is variably modified suggesting that clasts have moved following fragmentation. Disruption varies from slight modification of jigsaw-fit texture to rotation and separation of clasts. Rotation and separation of clasts is most obvious in cases where flow banding in adjacent clasts has different orientations. The breccia facies is massive and non-stratified. In some cases, clast-rotated breccia grades through in situ jigsaw-fit breccia into coherent facies. Flow banding in the coherent facies may be continuous into the jigsaw-fit breccia (Fig. 3.3A). Other units consist entirely of clast-rotated breccia. Intervals of the monomictic breccia facies are a few to 50 m thick, and occur as discontinuous pods or lenses within coherent rhyolite to dacite.

Origin and significance of facies

Textural variations within the monomictic rhyolitic to dacitic breccia facies reflect varying roles for quench fragmentation and autobrecciation in fragmentation. Clasts with planar and curvilinear margins probably formed through the propagation of thermal contraction fractures, while the clast rotated breccia reflects the importance of autobrecciation (e.g. Pichler, 1965). Breccia comprising jigsaw-fit clasts is interpreted as hyaloclastite. The components and fabric in clast-rotated breccia suggest fragmentation by autobrecciation alone (e.g. Allen, 1988) or a combination of quench-fragmentation and autobrecciation. These breccias record fragmentation of parts of the lava that were cooler, more viscous, and/or subject to higher strain rates during extrusion than the associated coherent facies.

3.3.3 Monomictic pumice breccia facies

In the area northeast of Truncheon (around 77490200 mN, 420000 mE), the base of the Mount Windsor Formation is marked by a poorly exposed, thin (15 m) interval of
massive, monomictic, dacitic pumice breccia (Map 1). This unit is characterised by wispy, feldspar-phyric (5%, 0.4-4 mm) pumice. Formerly glassy vesicle walls have been replaced by sericite and chlorite, whereas feldspar crystals are relatively unaltered. Enclosed within the breccia are pods of coherent dacite up to 10 m across. The dacite is finely banded and contains a similar phenocryst assemblage to pumice in the surrounding breccia. The spatial association between the pumice breccia and coherent dacite suggests that the two facies are genetically related.

3.3.4 Associations of coherent and autoclastic facies

In the Mount Windsor Formation, associations of coherent rhyolite to dacite and monomictic breccia represent lavas, domes and syn-volcanic intrusions, 100 to 300 m thick. Criteria used to distinguish between intrusive and extrusive units in the drill core and outcrop are outlined in Chapter 5. Lava domes and flows in the Mount Windsor Formation are dominated by coherent facies. In some cases, in situ hyaloclastite and autobreccia are developed along contacts with the underlying or overlying units. Transitions between autoclastic and coherent facies are sharp but irregular.

At Trooper Creek prospect, a single aphyric dacitic unit, 300 m thick, is exposed along strike for at least 2.5 km. The lateral continuity and upper contact relationships suggest that the dacite was emplaced as a thick flow rather than a dome. The lower contact of the dacite is poorly exposed and in places is marked by a zone of cataclasite. The coherent interior of the dacite is massive and finely flow banded and overlain by a thin (30–60 m) carapace of non-stratified autoclastic breccia of the same composition. The breccia encloses domains of coherent dacite, up to 10 thick and 120 m long. Contacts between the coherent domains and the surrounding breccia are sharp. The coherent domains are interpreted as pods of dacite enclosed within coeval autoclastic breccia. Alternatively, the coherent domains could be lobes that intruded pre-existing autoclastic breccia.

Many rhyolite and dacite intervals remain as undifferentiated lavas and intrusions as their margins are not exposed. In these cases, changes in the phenocryst assemblage are the only indication that boundaries between units have been crossed. Some rhyolite and dacite intervals are thin (10–30 m) and/or laterally discontinuous with have sharp margins, suggesting that they are syn-volcanic intrusions.

Hyaloclastite breccia associated with the lavas suggests that the depositional setting for the Mount Windsor Formation was submarine. Regional context further constrains the depositional environment to submarine.
3.3.5 Distribution, thickness and contact relationships of the Mount Windsor Formation

The Mount Windsor Formation is continuous between Truncheon prospect in the west and Prisoner Creek in the East (Fig. 3.1). Complete sections through the formation were limited to one area northeast of Highway East prospect, where a true thickness of 760 m is indicated (Fig. 3.2 — section A). The formation is at least 700 m thick at Highway South prospect and 200 m of the formation has been mapped at Trooper Creek prospect in the east.

In areas of good exposure, the Mount Windsor Formation appears to be conformably overlain by volcaniclastic units of the Trooper Creek Formation (e.g. Trooper Creek prospect, Highway East prospect). The Mount Windsor Formation rhyolites and dacites along these contacts have autobrecciated tops, suggesting that the Trooper Creek Formation volcaniclastic units were deposited directly onto the upper surfaces of lavas or domes which were not subject to significant erosion in the interim period. However, mass-flow deposits in the Kitchenrock Hill Member contain rounded clasts of rhyolite, rhyodacite and dacite which are petrographically and geochemically similar to the Mount Windsor Formation. The clasts were reworked in a high-energy environment (above storm wave base) prior to redeposition, suggesting that the source areas were subaerial to shallow marine. These clasts indicate that parts of the Mount Windsor Formation were subject to significant erosion up until the initial stages of Trooper Creek Formation volcanism. It is likely that while some parts of the contact were being subject to erosion, others were sites of deposition.

3.4 Trooper Creek Formation

The recognition of compositional and lithological variations within the Trooper Creek Formation, which are mappable over at least 15 km strike length, has prompted a subdivision of the formation into two members, the Kitchenrock Hill Member and the overlying Highway Member. The Kitchenrock Hill Member comprises volcaniclastic sandstone and breccia units that are typically polymictic and include sub-rounded to well-rounded clasts. In contrast, the Highway Member is dominated by syn-eruptive volcanic breccia to sandstone units, syn-sedimentary intrusions, lavas, and volcanic siltstone.

The Kitchenrock Hill Member forms a discontinuous stratigraphic interval overlying rhyolitic and dacitic lavas and intrusions of the Mount Windsor Formation. Due to the variable nature of the member, a type area (Kitchenrock Hill area; 7746000 mN, 422000 mE – 7748500 mN, 419000 mE) is proposed for this stratigraphic unit, rather than a type section (Fig. 3.2 — section B). The Highway Member conformably overlies the
Kitchenrock Hill Member. The Highway Member is characterised by rapid lithofacies variations and contains varying proportions of coherent volcanic, volcaniclastic and sedimentary facies. Because of this heterogeneity three representative type sections have been constructed, rather than a single type section. These are located (from west to east) at Highway East prospect from 7746900 mN – 417700 mE (top) to 7749250 mN – 420250 mE (base), Highway South prospect from 7745170 mN – 419950 mE to 7747650 mN – 421200 mE and at Trooper Creek prospect from 7742650 mN – 426700 mE (top) to 7744250 mN – 427450 mE. These correspond to sections A–C in Figure 3.2.

The thickness of the Trooper Creek Formation varies regionally from approximately 1665 m in the Highway East area to 1115 m in the Kitchenrock Hill to Highway South prospect area (assuming minimal fault repetition and disruptions). At Trooper Creek prospect, the formation has an approximate thickness of 2140 m and includes two thick dolerite sills. The age of the sills is uncertain. They show the effects of ?Ordovician-Devonian regional metamorphism. If the sills are removed a thickness of 1835 m is indicated.

3.4.1 Kitchenrock Hill Member

The Kitchenrock Hill Member comprises four main facies: (i) normally graded pumice-crystal breccia and sandstone; (ii) massive to graded polymictic breccia and sandstone; (iii) massive polymictic volcanic breccia; (iv) coherent rhyodacite and dacite. Minor laminated siltstone beds occur within the member in some areas. Volcaniclastic facies within the member typically contain more feldspar than quartz or a greater proportion of dacite or rhyodacite clasts than rhyolite clasts. Some units are entirely dacitic in composition. The remaining units contain similar proportions of quartz and feldspar crystals, or clasts with similar proportions of quartz and feldspar phenocrysts. The presence of subrounded to well rounded clasts is characteristic of the member. Rounded clasts are not present in all volcaniclastic units within the Kitchenrock Hill Member and their abundance appears to vary within single units along strike. The composition of rounded clasts is not uniform throughout the Kitchenrock Hill Member and they can include rhyolite, rhyodacite, feldspar-phyric dacite or aphyric dacite. Some clasts have phenocryst assemblages and geochemical signatures (Section 3.8) which are similar to the underlying Mount Windsor Formation.

Normally graded pumice-crystal breccia and sandstone

Pumice- and crystal-rich sandstone and breccia beds are the most common facies in the Kitchenrock Hill Member. Beds (10's cm to 60 m thick) are normally graded with tuffaceous sandstone tops and, in some cases, polymictic lithic-rich bases. The crystal
composition suggests dacitic to rhyodacitic volcanic provenance. The principal components are finely fibrous feldspar-quartz-phryic tube pumice, crystals and crystal fragments. Pumice clasts are variably oriented, uncompacted and altered to assemblages of chlorite, sericite and/or feldspar. Lithic clast populations vary between beds and comprise various assemblages of rhyolite, rhyodacite and dacite. Clasts can be perlitic, finely flow banded or spherulitic and, although mostly less than 40 cm across, some beds contain clasts (rhyolite) up to 2 m across. The majority of clasts have angular shapes but subrounded rhyodacite or dacite clasts are present in many beds. Rhyodacite clasts in a few beds are intensely silicified, whereas other clasts are weakly sericite-chlorite-altered. This implies that the rhyodacite clasts were altered at source prior to incorporation into the breccia.

**Origin and significance of facies**

The strongly pumiceous character of this facies suggests that pyroclasts were sourced from explosive silicic magmatic and/or phreatomagmatic eruptions. Sustained eruptions of this style are largely limited to subaerial settings or water shallower than about 1 km. In deeper water, volatile expansion is suppressed by the hydrostatic pressure exerted by the overlying water column (e.g. McBirney, 1963). Although pumiceous units within this facies have similarities to subaerial ignimbrites they show no evidence of hot emplacement. Their internal organisation is consistent with deposition from cold, water-supported, sediment gravity flows (cf. Lowe, 1982). Rounded clasts within the coarse lithic-rich bases of some of the deposits suggests reworking in a subaerial or shallow submarine environment prior to final deposition (below storm wave base). These clasts imply that the volcaniclastic mass flows may have transgressed a shallow-water environment. Alternatively, the mass flows may have collected clasts from the substrate during transport in a below storm wave base environment.

**Massive to graded polymictic breccia and sandstone**

Massive to weakly normally graded, polymictic breccia units (generally 1–20 m thick), intercalated with crystal-lithic sandstone beds (0.5–2 m thick), form a major lithofacies within the Kitchenrock Hill Member. Breccias are clast- to matrix-supported and lower contacts are sometimes erosion surfaces. Most beds are dominated by clasts 2 to 7 cm across, but the coarse base of some beds include clasts up to 80 cm across. Clast compositions vary between beds. Some beds are essentially monomictic and comprise aphyric- or feldspar-phyric dacite clasts and rare siltstone intraclasts (Fig. 3.3C). Finely flow banded clasts that superficially resemble tube pumice are a significant component of many beds. Other beds are polymictic but dominated by rhyolite clasts and feldspar-phyric dacite clasts (Fig. 3.3D). Clasts vary from angular and blocky to well rounded (Fig. 3.3C). Rounded clasts can be dacite or rhyolite. In some intervals of the facies,
intense silicification has modified groundmass textures within clasts and obscured clast margins. In these areas, dacite clasts have a finely granular texture and resemble fine-grained sandstone.

Intercalated sandstone beds contain abundant quartz and feldspar crystal fragments (15%) and angular felsic lithic fragments, with minor leucoxene. Beds comprise a lower division of massive to weakly graded sandstone (Bouma Ta) which passes up into a thin diffusely planar laminated division (Tb) followed by finely laminated siltstone (Te). Some beds display loading structures (e.g. flames).

**Origin and significance of facies**

The massive to graded polymictic breccia beds are interpreted to have been deposited from high-density turbidity currents and possibly debris flows in a submarine below-wave-base environment (cf. Lowe, 1982). The association with sandy turbidites and intercalated finely laminated siltstone beds support this interpretation. Rounded clasts were reworked in an above-wave-base environment prior to incorporation into the mass flows.

**Massive polymictic volcanic breccia**

Exposures of this facies are limited to an unnamed creek between Kitchenrock Hill and Highway South prospect (around 7746100 mN, 422000 mE). The top contact of the breccia is not exposed but the unit is at least 20 m thick. The bed is massive, matrix-poor and clast-supported. Clasts in the breccia include cherty siltstone, planar laminated siltstone and rhyodacite. These clasts have angular blocky shapes and vary from 1-30 cm across. The remaining clasts are angular to subrounded, coarsely quartz- and feldsparphyric rhyolite. Rhyolite clasts are mostly 10 cm to 1 m in size, but one clast 10 m across is exposed 2-3 m above the base (Fig. 3.3E). Rhyolite clast margins are mostly sharp but segments of a few clasts have broken in situ into small angular fragments. Pieces of the in situ fragmented clast margins have progressively separated and mixed with other constituents in the surrounding framework. In situ fragmented domains separate and partially enclose coherent rhyolite, the margins of which appear rounded. Clasts are separated by an unidentifiable matrix which has either altered to quartz or is cleaved and weathered to clay.

Rhyolite clasts are absent in the upper few metres of the exposure so that the breccia may be normally graded, although the top is not exposed. Contacts between the breccia and underlying Mount Windsor Formation rhyolite are poorly exposed but appear to be irregular. Rhyolite clasts in the breccia have phenocryst assemblages, sizes and abundances similar to the Mount Windsor Formation.
Origin and significance of facies

The components and organisation suggest that this facies was deposited from a mass flow. In situ disintegration of some rhyolite clasts margins, possible weak grading and mixing of clasts with different compositions and erosional histories, are evidence for mass-flow transport. For such a poorly sorted aggregate, different particle sizes will have had different support/transport mechanisms. Large clasts were probably transported as bed load, whereas smaller fragments were supported by traction and suspension. The role of the interstitial fluid in particle support during transport cannot be determined as the preserved matrix is altered, weathered and cleaved. Minor granular matrix may have been derived from attrition of larger clasts.

The moderate thickness and restricted distribution of the polymictic volcanic breccia facies suggest a localised source. Rhyolite clasts have phenocystal populations which suggest they were probably derived from the Mount Windsor Formation. Rounded clasts were reworked in a high-energy environment (above storm wave base) prior to redeposition, and suggest that the source areas were subaerial or shallow subaqueous. However, rounding of some large clasts is attributable to abrasion during transport within the flow. The evidence for reworking, heterogeneous clast population, small volume and rounded clasts, collectively suggest that the polymictic breccia is a post-eruptive deposit. The breccia may have accumulated on the unstable slopes of the underlying Mount Windsor Formation rhyolite. Discontinuous outcrop means that the depositional surface cannot be reconstructed.

Coherent rhyodacite and dacite and associated autoclastic breccia facies

The Kitchenrock Hill Member also contains intervals of rhyodacite and dacite but these are relatively minor facies. The rhyodacite intervals are characterised by 7% feldspar phenocrysts, 1-1.5 mm across, and subordinate quartz phenocrysts (3%, 0.5-1 mm). Dacite intervals contain 3% feldspar phenocrysts, 1-2 mm long. Upper contacts, critical in evaluating intrusive versus extrusive emplacement, are not always exposed. The most revealing exposure of rhyodacite occurs in the Highway East area (around 7747800 mN, 419900 mE). In this case, the top contact is marked by intricate interpenetration of rhyodacite and the overlying lithic-crystal sandstone. Downward from the contact, siltstone seams cut across the rhyodacite and locally merge forming siltstone-rich breccia. In some of these domains, ragged clasts of rhyodacite are entirely enclosed by siltstone. The degree of mixing between the rhyodacite and siltstone at the contact, suggests that the magma invaded wet unconsolidated sediment as a syn-sedimentary sill, and is consistent with other well described examples of peperite (e.g. Busby-Spera and White, 1987; Hanson, 1991).
Stratigraphy

3.14.

The Kitchenrock Hill Member is present in all areas studied and ranges from 60 to 110 m thick (Fig. 3.2). Poor exposure and faulting make it difficult to trace single units within the member. In some areas, the thickness of single units and the overall thickness of the member appear to increase towards fault contacts with the Mount Windsor Formation rhyolite (Fig. 3.1, Map 1). Bedding orientations in the Kitchenrock Hill Member are locally discordant to contacts with the underlying Mount Windsor Formation rhyolite, suggesting that there was topography on the palaeodepositional surface. Variations in thickness and bedding orientation within the Kitchenrock Hill Member probably reflect palaeotopography on the depositional surface of the underlying Mount Windsor Formation that may have been generated by the lavas and/or syn-depositional growth faults.

Depositional setting

The underlying rhyolite of the Mount Windsor Formation provides little unambiguous information on the depositional setting of the Kitchenrock Hill Member. Volcaniclastic facies within the Kitchenrock Hill Member support the interpretation that the depositional setting was subaqueous, and regional context further constrains the depositional setting to submarine. The widespread occurrence of turbidites is good evidence for deposition below storm wave base. Rounded clasts and the abundance of pyroclasts in some lithofacies suggest that the source eruptive centres were at least in part subaerial or shallow subaqueous.

Upper contact of the Kitchenrock Hill Member

The lithic-rich volcaniclastic units of the Kitchenrock Hill Member grade upward into the volcaniclastic units of the Highway Member. The contact between these members is gradational and interfingering, and taken as the top of the uppermost unit with rounded clasts. As the proportion of rounded clasts within single units varies along strike, the contact is sometimes poorly defined. The Kitchenrock Hill Member contains a much lower portion of siltstone relative to the Highway Member. In some areas, the uppermost units with rounded clasts are followed by thick intervals of siltstone (Highway Member).

3.4.2 Highway Member

The Highway Member comprises coherent lithofacies and compositionally and texturally diverse volcaniclastic lithofacies which are intercalated with volcanic and non-volcanic
sedimentary facies. Summary descriptions and interpretations of the twenty principal volcanic and sedimentary facies are given in Table 3.1. Coherent lithofacies and clasts in the volcanioclastic lithofacies are mostly rhyolitic, dacitic or andesitic in composition but rare basaltic-andesitic examples are present.

The lithofacies can be grouped within three principal lithofacies associations.

1. Lithofacies of the primary volcanic facies association include coherent rhyolite to andesite and associated monomictic non-stratified breccia facies (autobreccia, hyaloclastite; Table 3.1) and sediment-matrix breccia facies (peperite; Table 3.1).

2. The resedimented volcanioclastic facies association mostly comprises dacitic to rhyolitic pumice breccia, andesitic breccia to sandstone, and stratified monomictic rhyolitic to andesitic breccia.

3. The sedimentary facies association is dominated by massive to laminated siltstone but include minor intervals of sandstone and stromatolitic ironstone. Siltstone and sandstone units are dominantly volcanic. Non-volcanic detritus is more prominent near the top of the Highway Member.

There are regional variations in the proportion of the three principal lithofacies associations and in the composition and texture of the constituent lithofacies within each association. In the area around Coronation homestead, the primary volcanic facies association is dominant and comprises massive coherent and flow-banded rhyolite and dacite (Map 1). The resedimented volcanioclastic lithofacies comprise rare exposures of monomictic dacitic breccia. Sedimentary facies are absent. In the Highway-Reward to Highway East area, the primary volcanic facies association is most abundant and represented by rhyolitic, dacitic and andesitic lavas, syn-sedimentary intrusions, autoclastic breccia and peperite (Map 1; Fig. 3.2A). The resedimented volcanioclastic association and sedimentary facies association are more prominent here than further to the west. The resedimented volcanioclastic facies association is dominated by monomictic volcanic breccia (resedimented autoclastic breccia), pumice-crystal-lithic breccia and sandstone and graded andesitic scoria breccia (Table 3.1). To the east, between Highway South prospect and Trooper Creek prospect, the sedimentary facies association is more common and intervals of interbedded siltstone and sandstone range up to 160 m in thickness (Figs. 3.2B-C). The resedimented volcanioclastic facies association is dominated by stratified and laminated pumice breccia and sandstone, graded andesitic scoria breccia, block-rich andesitic breccia and vitric-crystal sandstone.

Fossils (stromatolites, trilobites) and regional context suggest that the depositional setting for the Highway Member was submarine. The widespread occurrence of turbidites suggests that most of the volcanic succession accumulated below storm wave base (generally greater than 150-250 m). The exception occurs in the southern segment of
<table>
<thead>
<tr>
<th>Lithofacies</th>
<th>Characteristics</th>
<th>Interpretation</th>
</tr>
</thead>
<tbody>
<tr>
<td>Massive rhyolite to dacite</td>
<td>Aphyric or evenly porphyritic; columnar and tortoise shell jointing; massive or flow banded</td>
<td>Coherent facies of lavas and domes, cryptodomes and syn-sedimentary intrusions</td>
</tr>
<tr>
<td>Massive andesite</td>
<td>Aphyric or evenly porphyritic; vesicular (1-15%); massive or flow banded; platy joints</td>
<td>Coherent facies of lava flows and syn-sedimentary intrusions</td>
</tr>
<tr>
<td>Basaltic andesitic lobes</td>
<td>Close-packed lobes, 5-11 cm in diameter; strongly vesicular cores, glassy margins; inter lobe material is jigsaw-fit, formerly glassy, fragments</td>
<td>Quench fragmented lava. Result of incomplete quenching of lava leaving unfragmented pods of magma which cooled slowly and vesiculated</td>
</tr>
<tr>
<td>Non-stratified rhyolite to dacite breccia</td>
<td>Monomictic; poorly sorted; blocky to ragged clasts; clast- to matrix-supported; gradational into coherent facies and/or peperite</td>
<td>Autobreccia and in situ hyaloclastite</td>
</tr>
<tr>
<td>Non-stratified andesitic breccia facies</td>
<td>Monomictic; poorly sorted; blocky clasts; some clasts have tiny-normal joints (cf. Yamagishi, 1979); clast supported; gradational into coherent facies</td>
<td>Autobreccia and in situ hyaloclastite</td>
</tr>
<tr>
<td>Non-stratified sediment-matrix breccia facies</td>
<td>Rhyolitic to andesitic; blocky, ragged and globular shaped clasts; clast- to matrix-supported; jigsaw-fit texture in matrix-poor breccia; matrix may be siltstone, sandstone or pumice breccia; present along the upper or lower contacts of massive andesite to rhyolite facies; 0.1-1 m thick</td>
<td>Peperite (cf. Busby-Spera and White, 1987; Brooks, 1995)</td>
</tr>
<tr>
<td>Stratified, monomictic rhyolite to dacite breccia facies</td>
<td>Thick (0.5-11 m), internally massive or graded beds; clast-supported; blocky to elongate Jagged clasts; often associated with hyaloclastite, peperite and coherent lava</td>
<td>Gravity-driven collapse and resedimentation of unstable hyaloclastite (cf. Dimroth et al., 1978). Deposits from high-concentration sediment gravity flows</td>
</tr>
<tr>
<td>Siltstone-matrix rhyolite to dacite breccia</td>
<td>Stratified; polymictic, matrix- to clast-supported, thick (&lt; 7 m); internally massive or normally graded; blocky to ragged clasts locally with jigsaw-fit fabric; siltstone matrix and intraclasts</td>
<td>Gravity-driven collapse of unstable hyaloclastite or peperite from the margins of subaqueous lavas or cryptodomes; deposition from high-concentration sediment gravity flows (? debris flows)</td>
</tr>
<tr>
<td>Indurated siltstone-matrix rhyolite breccia facies</td>
<td>Stratified; polymictic; very poorly sorted; matrix-supported; &gt; 20 m thick; blocky to ragged rhyolite clasts with indurated siltstone rinds; other clasts are indurated siltstone and siltstone intraclasts; sediment matrix</td>
<td>Collapse of unstable peperite from the margins of subaqueous cryptodomes; deposition from debris flows (cf. Hanson and Wilson, 1993)</td>
</tr>
<tr>
<td>Graded dacitic to rhyolitic pumice breccia and sandstone</td>
<td>Essentially monomictic; normally graded; non-welded; 1-80 m thick; equant to ragged tube pumice, crystal fragments, shards and sparse angular lithic clasts</td>
<td>Resedimentation of subaerial or shallow submarine pyroclastic pumice into a deeper submarine setting; syn-eruptive; downslope transport by high-concentration turbidity currents</td>
</tr>
<tr>
<td>Lithofacies</td>
<td>Characteristics</td>
<td>Interpretation</td>
</tr>
<tr>
<td>----------------------------------</td>
<td>-------------------------------------------------------------------------------</td>
<td>--------------------------------------------------------------------------------</td>
</tr>
<tr>
<td>Stratified crystal-rich</td>
<td>Essentially monomictic; massive or weakly normally graded; rich in crystal</td>
<td>Syn-eruptive; crystal concentration during eruption and transportation;</td>
</tr>
<tr>
<td>volcanic sandstone</td>
<td>fragments and pumice with lesser shards and lithic clasts; 1-50 m thick</td>
<td>deposition from high-concentration, granular turbidity currents</td>
</tr>
<tr>
<td>Planar laminated dacitic</td>
<td>Monomictic; clast-supported; thinly planar laminated; &lt; 5 m thick; probable</td>
<td>Water-settled fall in a shallow submarine environment; source of pyroclasts was</td>
</tr>
<tr>
<td>pumice breccia</td>
<td>mantle bedding; non-welded; tube pumice</td>
<td>subaerial or shallow subaqueous</td>
</tr>
<tr>
<td>Polymictic limestone-pumice</td>
<td>Poorly sorted; matrix-supported; weak normal grading; ~12 m thick; clasts (3 cm-2 m) of siltstone, dacite, ironstone, dacite clasts with indurated siltstone rings; pumiceous matrix</td>
<td>Gravity-driven collapse of pre-existing unstable pepsite, pumice breccia and ironstone; down-slope transport by high-concentration sediment gravity flows (? detritus flow); deposited and sourced subaqueously</td>
</tr>
<tr>
<td>breccia facies</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Graded andesitic scoria breccia</td>
<td>Essentially monomictic; stratified; thin to thick (0.1-40 m), normally graded</td>
<td>Syn-eruptive resedimentation of pyroclasts from a shallow submarine scoria cone (cf. Lonsdale and Batiza, 1980); deposits from high-particle concentration sediment gravity flows and suspension sedimentation</td>
</tr>
<tr>
<td>Cross-stratified andesitic</td>
<td>Monomictic, moderately well sorted; high-angle trough cross-bedding; poorly</td>
<td>Resedimentation of pyroclasts into submarine setting from subaqueous to subaerial scoria cone; deposition from traction currents in an above-storm-wave-base environment</td>
</tr>
<tr>
<td>breccia and sandstone</td>
<td>vesicular andesite fragments, and subordinate (5-7 %) scoria fragments;</td>
<td></td>
</tr>
<tr>
<td></td>
<td>intervals &lt;20 m thick</td>
<td></td>
</tr>
<tr>
<td>Globular clast-rich andesitic</td>
<td>Monomictic, moderately-poorly sorted; massive to normally graded; fragments</td>
<td>Near vent pyroclastic deposit; subaqueous to subaerial strombolian eruption; minor down-slope transport</td>
</tr>
<tr>
<td>breccia</td>
<td>of bombs supported in a framework of poorly vesicular clasts and minor (10 %)</td>
<td></td>
</tr>
<tr>
<td>Vitric-crystal-lithic sandstone</td>
<td>Planar, laterally continuous beds; thin (15-70 cm) massive to normally graded</td>
<td>Sandstone beds are deposits from low concentration turbidity currents (Bouma Ta-Ta); siltstone predominantly from suspension sedimentation</td>
</tr>
<tr>
<td></td>
<td>crystal-vitric-lithic sandstone and interbedded siltstone; dominantly volcanic;</td>
<td></td>
</tr>
<tr>
<td></td>
<td>some beds contain non-volcanic detritus</td>
<td></td>
</tr>
<tr>
<td>Massive to laminated siltstone</td>
<td>Laminated or thinly bedded intervals up to 160 m thick; planar, even,</td>
<td>Predominantly suspension sedimentation; in part water-settled volcanic ash; deposited below storm wave base</td>
</tr>
<tr>
<td></td>
<td>continuous beds in some cases laminae drape small irregularities such as large pumice clasts</td>
<td></td>
</tr>
<tr>
<td>Ironstone</td>
<td>Quartz-hematite=magnetite-chlorite-sericite-feldspar-calcite; locally pumiceous; discontinuous lenses (10-160 m long) and pods</td>
<td>Hydrothermal precipitates</td>
</tr>
<tr>
<td>Microbialitic ironstone</td>
<td>Microbialites composed of stromatolites, oncolites, pyroclasts and skeletal</td>
<td>In situ stromatolites as thin films, domed biostromes and subspherical bioherms; above-storm-wave-base structures</td>
</tr>
<tr>
<td></td>
<td>fragments</td>
<td></td>
</tr>
</tbody>
</table>
Trooper Creek prospect, where the depositional setting for the upper part of the Highway Member was shallow submarine. Stromatolites, traction current structures indicative of wave activity, and evaporitic minerals collectively suggest that this succession accumulated above storm wave base and may have even been temporarily emergent.

Genetically-related lithofacies associations are further described and discussed in chapters 4 and 5. Chapter 4 documents the internal facies characteristics of a shoaling andesitic to dacitic volcanic succession in the upper part of the Highway Member, at Trooper Creek prospect. Chapter 5 examines the lithofacies associations characteristic of submarine (below storm wave base), non-explosive, lava- and intrusion-dominated volcanic centres in the area around Coronation homestead, Highway-Reward and Highway East prospect.

**Thickness of the Highway Member**

The thickness of the Highway Member varies regionally from approximately 1615 m in the Highway East prospect to Kitchenrock Hill area to 1035 m at Highway South prospect (assuming minimal fault repetition and disruptions). At Trooper Creek prospect, the member has an approximate thickness of 1920 m which includes two thick dolerite sills.

**Upper contact of the Highway Member**

The transition to the overlying Rollston Range Formation is poorly exposed throughout the study area. The most continuous sections are exposed at Trooper Creek prospect (around 7741500 mN, 427000 mE), near Highway South (around 7745000 mN, 420000 mE) and occur in diamond drill core from two prospects (Stocksquad and Rustler) to the south and east respectively of Highway mine. In the Rustler drill hole (RD 812, 132 m), the transition from the Highway Member to the Rollston Range Formation is gradational and conformable (Fig. 3.4). At Stocksquad (RD 813, 78-114.5 m), the dacitic pumice breccia facies of the Highway Member is in faulted contact with the Rollston Range Formation.

At Trooper Creek and Highway South prospects, rare polymictic sandstone and lensoidal pebbly sandstone beds (0.1-2 m thick) occur in places near the top of the Highway Member (Fig. 3.3F). The sandstone beds are massive or normally graded, sometimes show minor low-angle cross-stratification and are interbedded with siltstone units (0.3 to 2 m thick). The fragment population in the sandstone units is diverse, comprising dominantly volcanic quartz and feldspar crystal fragments, with subordinate tourmaline and zircon crystals, white mica, quartz-hematite clasts and basement-derived lithic clasts (phyllite, metachert). The sandstone units are interpreted to have been deposited from
Figure 3.4 Simplified graphic lithological logs showing important textural variations and contact relationships of units within diamond drill core from (a) Stocksquad prospect, DDH RD 813 and (b) Rustler prospect, DDH RD 812. Downhole thickness shown.
high-density turbidity currents (Chapter 4). The components and organisation of the sandstone units are similar to those in the overlying Rollston Range Formation (Section 3.5.1). Along contacts with the Rollston Range Formation, intervals of the polymictic sandstone facies are intercalated with dacite, andesite and thick pumiceous mass-flow deposits, and increase in abundance towards the top of the Highway Member. This mixed interval is transitional and implies that the contact is conformable.

3.5 Rollston Range Formation

Although the Rollston Range Formation is poorly exposed throughout the study area, the range of lithofacies observed is similar in the east and west. The formation consists of volcanogenic sandstone and siltstone and rare pumiceous units.

3.5.1 Rounded lithic-crystal sandstone facies

The sandstone beds are poorly sorted, immature and although typically massive are sometimes weakly normally graded (Fig. 3.3G). Some beds have erosional bases and contain ragged siltstone intraclasts. The grain population is diverse and dominated by quartz and subordinate feldspar (Fig. 3.3H). Quartz grains display undulose or straight extinction. Feldspars are plagioclase and alkali feldspar, including microcline and rare altered perthite fragments. Other components are, from most to least abundant, detrital biotite, white mica, zircon, tourmaline, apatite and rare rounded oxide grains. Grains in the sandstone are separated by small amounts of fine-grained sericite, chlorite and cryptocrystalline quartz. Some sandstone beds contain a small percentage of lithic fragments which can be sand, granule or pebble size. Fragments can include biotite phyllite, metachert, polycrystalline quartz fragments and sandstone. The metachert fragments comprise interlocking equant to sub-equant microcrystals of quartz with no visible relic texture. A few grains contain rare angular volcanic quartz fragments and flecks of sericite. Polycrystalline quartz fragments comprise semi-equigranular interlocking euhedral to anhedral quartz. Phyllite grains typically consist of fine-grained interlayered quartz, sericite and biotite.

Some sandstone beds are dominated by subrounded to rounded fragments, whereas others contain a higher percentage of angular fragments. Quartz and feldspar grains vary from angular to well rounded. Plagioclase crystal fragments are mostly angular but some grains are subangular. Zircon, apatite and tourmaline grains in many samples have subrounded to rounded shapes, but grains can be euhedral and little modified. White mica
and biotite fragments have elongate platy shapes and only rare fragments show any evidence of rounding. The remaining lithic fragments are angular to subrounded.

**Origin and significance of facies**

The massive to weakly graded character of the sandstone beds suggests that they are the deposits of sandy high-concentration turbidity currents. Complete sedimentation units from sandy high-density turbidity currents include (Lowe, 1982): a lower division (S1) of coarse sand to gravel showing traction structures; an overlying inversely graded, planar laminated sand division (S2); and an upper division of grain-supported massive or normally graded sand, commonly showing dewatering structures, and deposited directly from suspension (S3) (Lowe, 1982). The sandstone units are similar to the S3 division of Lowe (1982). However, there is no evidence for the S1 or S2 divisions. Siltstone beds include sediment deposited from the dilute sediment clouds trailing the turbidity currents and sediment which settled through the water column.

The sandstone units are dominated by volcanic quartz and feldspar, suggesting a predominantly felsic source. As a result of regional deformation, volcanic quartz commonly exhibits undulose extinction, especially in sandstone beds which are grain-supported. Zircon and apatite may also be volcanic in origin. The remaining fragment population is clearly non-volcanic and implies input from granitic and deformed basement sources (cf. Henderson, 1986).

Many grains are subrounded to well rounded, indicating reworking in an above-wave-base environment prior to deposition. Feldspar crystal fragments are relatively unaltered and angular suggesting they have been little reworked and may be locally derived. Quartz, zircon and apatite have suffered varying but generally higher degrees of reworking prior to final deposition, suggesting a separate source. The remaining fragment population is dominantly subrounded to rounded suggesting significant reworking prior to deposition.

3.5.2 Siltstone and mudstone facies

In the areas studied, the Rollston Range Formation includes thick intervals of siltstone to mudstone. These fine-grained deposits occur as horizons up to 50 m thick or interbedded with crystal-lithic sandstone beds. Siltstone/mudstone beds (centimetres to 2 m thick) are laterally continuous, generally massive and locally show diffuse planar lamination. They vary from light grey to light brown in colour. Light grey siltstone/mudstone beds are composed of cryptocrystalline quartz. Pale brown coloured siltstone/mudstone comprises a fine-grained mosaic of quartz and sericite.
Origin and significance of facies
The widespread occurrence of turbidites implies that the siltstone beds accumulated below storm wave base. The siltstone beds incorporate non-volcanic and volcanic components that settled from suspension. Some are probably deposits of pelagic or hemi-pelagic sediment.

3.5.3 Crystal-pumice sandstone facies

In one area (around 7743500 mN, 421100 mE), the Rollston Range Formation contains minor crystal-pumice sandstone. The sandstone is very poorly exposed (<1 m), massive, and composed of pumice, shards and crystal fragments (feldspar and lesser quartz). The crystal composition suggests a dacitic to rhyodacitic provenance. The sandstone is intercalated with pale grey cherty siltstone units.

Origin and significance of facies
The dominance of juvenile pyroclasts suggests that this facies may have been sourced from explosive magmatic eruptions in a subaerial or shallow marine environment. The association with laminated siltstone/mudstone units suggests that the beds were deposited below storm wave base, most probably as a sediment gravity flow.

3.5.4 Depositional setting

Most of the sandstone units in the Rollston Range Formation are interpreted as turbidites, indicating deposition below storm wave base. Thick intervals of siltstone are consistent with deposition in a relatively deep and/or quiet water setting. The succession contains fossiliferous (graptolites, trilobites) horizons which suggest a submarine depositional environment (Henderson, 1986).

3.5.5 Lower contact of the Rollston Range Formation

Henderson (1986) defined the lower boundary of the Rollston Range Formation as “the top of the uppermost substantial unit dominated by pyroclasts”. However, mapping of the current study and previous workers (Berry, 1991; Berry et al., 1992) suggests that pumiceous sandstone units and coherent dacite occur within the Rollston Range Formation up to 3 km above contacts with the Trooper Creek Formation. The lavas and volcaniclastic deposits imply that episodic volcanism occurred during deposition of the Rollston Range Formation. Further work is required to determine the character,
distribution and abundance of volcanic units within the Rollston Range Formation. The research will help to clarify the nature and position of contacts between the two formations. As defined by Henderson (1986), the boundary between the Trooper Creek Formation and Rollston Range Formation is obscure. The boundary between the two formations is redefined as the top of the uppermost stratigraphic interval dominated by syn-eruptive volcanioclastic units and/or rhyolitic to basaltic lavas and syn-sedimentary intrusions. The overlying succession is dominated by volcanogenic and non-volcanic sedimentary units and forms the Rollston Range Formation.

3.6 Intrusions within the Seventy Mile Range Group

3.6.1 Diorite dykes

Diorite dykes are conspicuous at Trooper Creek prospect (around 7743500 mN, 427000 mE). The dikes are medium grained with subophitic textures (e.g. 95-196). Plagioclase laths (50-600 μm long) are partially enclosed in hornblende which shows varying degrees of alteration to chlorite. Other components are alkali feldspar (1%), subhedral epidote crystals (5-7%, 100 μm across) and interstitial chlorite. The dykes are 1-3 m wide and are exposed discontinuously for up to 400 m. They are cross cut by coarsely quartz- and feldspar-phyric rhyolite dykes.

3.6.2 Dolerite

Two large dolerite bodies intrude the Trooper Creek Formation at Trooper Creek prospect. The dolerite has a subophitic texture in which plagioclase laths are partially enclosed by incompletely chloritised clinopyroxene. Other components are interstitial chlorite, prehnite, quartz, carbonate and granular sphene. The bodies are clearly intrusive, and vary in thickness along their exposed length (1.5 km) from 100 to 250 m wide. The age of the dolerite is uncertain.

3.6.3 Granodiorite

Granodiorite intrudes the Trooper Creek Formation at Highway South prospect. The granodiorite is equigranular and consists of approximately equal proportions of quartz and feldspar crystals (200-250 μm long). Plagioclase is the dominant feldspar but minor microcline and microperthite are also present. Feldspar is largely unaltered or is only weakly altered to sericite. Rare tabular-shaped domains of very fine-grained quartz may
be pseudomorphs of feldspar. Other components are rare zircon micropherocrysts, 100 μm long, and occasional white mica crystals (muscovite ?). Quartz and feldspar crystals are broken into jigsaw-fit fragments which are separated by sericite.

3.6.4 Intrusive polymictic breccia facies

Exposures of this facies are restricted to Highway East prospect in an unnamed creek at 7747200 mN, 419750 mE. In map view, the breccia cuts across andesitic turbidites and is roughly circular (4 m in diameter). The breccia is polymictic, non-stratified and clast-supported. The clast population is diverse and includes rhyolite, dacite, andesite and fragments which have altered to epidote, quartz or chlorite. Rhyolite clasts have varying abundances of quartz and feldspar suggesting they sample different primary sources. Clasts vary from angular to subrounded and rounded. Although clasts are mostly 3 mm to 4 cm in size some are up to 12 cm across. The matrix consists of sand- to granule-sized finely comminuted lithic and crystal fragments with similar compositions and shapes to the clasts. Epidote alteration has partly obscured the character of sand-sized components in the matrix and altered some larger fragments.

*Origin and significance of facies*

Cross-cutting relationships suggest that the breccia is a pipe-like intrusion. The components suggest that the breccia is composed of pre-existing lithologies. The breccia is correlated with similar bodies which occur throughout the Mount Windsor Subprovince and interpreted as Permo-Carboniferous in age (e.g. Morrison, 1988; Wormald et al., 1991; Wormald, 1992). These breccia pipes are up to 2 km in diameter, and are composed of both pre-existing lithologies and material sourced from associated porphyry intrusions. The breccias have been interpreted to form during rapid magmatic gas release from porphyry intrusions. They are subvolcanic in origin. Rounding of clasts is attributed to abrasion and attrition of fragments during fluidisation by the upward streaming of volatiles (Wormald et al., 1991; Wormald, 1992).

3.7 Regional correlations

The new stratigraphic scheme is based on mapping of only a 15 km segment in the central part of the Seventy Mile Range Group. Previous studies in other parts of the belt (e.g. Berry et al., 1992; Van Eck, 1994) describe lithological units that are similar in composition, components and position in the stratigraphy to those recognised in the study area. Mapping by Van Eck (1994) suggests that the Kitchenrock Hill Member is also represented at Mount Farrenden, 5 km to the west. Van Eck (1994) describes clast-
matrix-supported breccia units comprising angular to subrounded, finely flow banded, dacite clasts (1-3 cm), possible basalt clasts and highly silicified clasts. Near Thalanga, deposits with rounded clasts have also been recognised along the contact between the Mount Windsor Formation and Trooper Creek Formation (Anthea Hill pers. com., 1996).

Correlations based on tracing mass-flow units with distinct provenance characteristics suggest that some key facies are traceable over large distances within the Highway Member. The best facies for correlation are the thick mass-flow-emplaced syn-eruptive pumiceous units. These were erupted infrequently, emplaced rapidly, are widespread and characterised by distinct phenocryst assemblages and abundances (cf. McPhie and Allen, 1992).

Distinctive units of this type occur at Stocksqaud and Rustler prospects near the contact between the Trooper Creek Formation and the overlying Rollston Range Formation. At Stocksqaud and Rustler prospects, the upper part of the Trooper Creek Formation comprises normally graded, thin to thick beds (to 30 m) of non-welded crystal-lithic-pumice breccia (Fig. 3.4). At both locations, the lower units are feldspar > quartz bearing and may correlate with similar breccias intersected in drill core at Highway-Reward (e.g. REW 805), to the north at Policeman Creek prospect (e.g. HDD 004), and in outcrop at Highway East prospect. Whether or not this correlation is direct, their mineralogy contrasts markedly with the overlying units which consist entirely of feldspar-only vitric-crystal-pumice breccia and sandstone. A syn-sedimentary intrusion occurs between the feldspar-only units and feldspar > quartz pumice breccia beds at Rustler prospect. At Stocksqaud prospect, massive coherent dacite, in situ hyaloclastite and normally graded resedimented hyaloclastite units indicate the presence of a lava flow or dome at an equivalent stratigraphic position.

The thickness of the feldspar-bearing pumice breccia facies in REW 813 (106 m) and exclusion of other particle types, point to it as a potential marker horizon. Although dacitic pumice breccia units occur lower in the stratigraphy (e.g. south of Hanćuff) these are intercalated with thicker intervals of feldspar > quartz-bearing pumice breccia. The key association of facies at Rustler and Stocksqaud prospects allows correlation of the drill hole sections to Trooper Creek prospect, 15 km to the east. At Trooper Creek prospect, the upper part of the Highway Member includes thick intervals of dacitic pumice breccia (Fig. 3.2C). The principal units are as follows: at the base stratified andesitic scoria breccia units; overlain by a thick sequence comprising intercalated dacitic (feldspar-only) pumice breccia beds, microbialites, siltstone and minor sandstone; followed by lithic-crystal sandstone and siltstone beds intercalated with feldspar-bearing pumiceous mass-flow units up to 80 m thick. Differences in the key facies association between the different localities reflect different depositional settings and distance from the source. It is not
possible to directly correlate single pumice breccia beds at Trooper Creek prospect with those further to the west.

The potential for the key facies association to be regionally extensive incited a review of the available descriptions of the stratigraphy elsewhere in the Seventy Mile Range Group. The association could extend at least to Liontown, 15 km to the west, where similar lithofacies to those at Stocksquad and Rustler are exposed (e.g. Berry et al., 1992; Miller, 1996). A thick interval of feldspar pumice breccia is overlain by sericitic siltstone and sandstone beds (Liontown Horizon) followed by a thick succession of black shale, rhyolitic volcaniclastic sandstone, cherty siltstone and feldspar-quartz-phyric rhyolite and dacite. The Liontown Horizon is host to barite-carbonate-base metal sulfide lenses interpreted as seafloor exhalative VHMS deposits. Sub-seafloor replacement style sphalerite-galena-pyrite mineralisation occurs within the footwall dacitic pumice beds (e.g. Berry et al., 1992; Miller, 1996). The Liontown deposit has been interpreted to occur at or near the top of the Trooper Creek Formation (Berry et al., 1992). The correlations proposed here support this interpretation.

The recognition of a regionally extensive key facies association provides an important framework for mineral exploration within the Trooper Creek Formation (Chapter 8). In particular, the correlations suggest that the Liontown and Highway-Reward deposits occupy a similar stratigraphic position near the top of the Highway Member. Previous interpretations have placed the Highway-Reward deposit within the central part of the Trooper Creek Formation. The correlations also suggest that the host succession to the Liontown mineralisation could be continuous to the east for over 30 km. The key facies association thus provides a powerful guide for mineral exploration. Stocksquad and Rustler prospects occur at the same stratigraphic position and represent clear targets for further exploration.

3.8 Regional geochemistry

Regional lithogeochemical studies of coherent volcanic rocks within the Seventy Mile Range Group were undertaken by Berry et al. (1992) and Stolz (1995). Stolz (1995) subdivided the Seventy Mile Range Group into four igneous suites. These correspond to discrete stratigraphic units within the three major volcanic-bearing formations (Puddler Creek Formation, Mount Windsor Formation, Trooper Creek Formation) of the Seventy Mile Range. In this section, Stolz’s (1995) geochemical subdivisions are followed closely and only slightly modified from his paper. The major difference here is an interpretation of the geochemical data which draws on a more detailed understanding of the facies
architecture of the Trooper Creek Formation and Mount Windsor Formation in the study area.

3.8.1 Sampling and analytical techniques

The new trace- and major-element data are presented in Appendix D and relate to a suite of least altered, coherent lavas and intrusions in the Mount Windsor Formation and Trooper Creek Formation, sampled from outcrop between Coronation homestead and Trooper Creek prospect. An additional 18 samples were collected from drill core in the host sequence to the Highway-Reward deposit. Analyses from Highway-Reward have been interpreted with caution, in the absence of less altered examples.

Rocks were first crushed in a jaw crusher, and a hand picked separate of chips free of oxidised or weathered rinds, veins or amygdales was powdered in a tungsten carbide disc mill. Major element and trace element concentrations were determined on a Philips automated XRF spectrometer at the University of Tasmania using standard fused disc and pressed pellet techniques (Norrish and Chappell, 1977). The major element analyses have been recalculated to 100% anhydrous to remove variations caused by differing loss on ignition values.

3.8.2 Element mobility

In the study area, the Seventy Mile Range Group has been affected by regional deformation and prehnite-pumpellyite to greenschist facies metamorphism, and hydrothermal alteration is locally intense. Therefore all samples selected for this study have undergone some degree of mineralogical readjustment. The variable mobility of elements during low-grade metamorphism and hydrothermal alteration is relatively well documented (e.g. MacLean and Kranidiotis, 1987; Whitford et al., 1989; Rollinson, 1993). Elements considered to be essentially immobile during these styles of alteration include the high field strength elements such as Ti, Zr and Nb. Also generally reliable are P, Sc, Y and Th. The strong correlation between elemental pairs (e.g. Zr/Nb, r=0.945) in rocks from the Seventy Mile Range Group, confirms that the high field strength elements have remained relatively immobile (cf. MacLean and Kranidiotis, 1987; MacLean and Barrett, 1993; Stolz, 1995). The concentrations of the large ion lithophile elements (including K, Rb, Ba and Sr) are unlikely to reflect original magmatic concentrations.

Despite undoubted silica mobility, there is a general negative correlation between Ti/Zr and SiO₂ (Fig. 3.5A), suggesting that the SiO₂ abundances in these rocks are often within
a few percent of their primary concentrations. SiO$_2$ is a useful chemical discriminator and fractionation indicator, especially when considered together with immobile elements. Particular emphasis has been placed on interpretation of these patterns, for which Stolz (1995) based his geochemical subdivisions of the Seventy Mile Range Group.

3.8.3 Compositions

Previous studies (Stolz, 1989, 1991; Berry et al., 1992; Stolz, 1995) demonstrated a predominance of rocks with low- to medium-K calc-alkaline compositions in the Seventy Mile Range Group. In a plot of SiO$_2$ versus K$_2$O (Fig. 3.5B) the new data conform to the regional pattern. In samples with SiO$_2$ contents greater than 77 wt% neither SiO$_2$ nor K$_2$O are likely to be pristine. Samples from the Trooper Creek Formation display a broad range of K$_2$O values at various silica concentrations, but mostly have low- to medium-K signatures. Samples from Highway-Reward display a large scatter of K$_2$O concentrations which is attributable to the effects of hydrothermal alteration.

Mount Windsor Formation

Lavas and intrusions from the Mount Windsor Formation are mostly rhyolite but some rhyodacite and dacite is present (Fig. 3.5C). The rhyolites, rhyodacites and dacites have SiO$_2$ concentrations in the range 74-80 wt% and are characterised by relatively low abundances of TiO$_2$, P$_2$O$_5$, MgO (Fig 3.5D-F), CaO and Fe$_2$O$_3$ (Fig. 3.6A-B). In addition, they display a broad range of Na$_2$O, K$_2$O, Sr and Ba values (Appendix D). Much of the variation in the concentrations of these elements is interpreted to reflect post depositional alteration and metamorphism.

Trooper Creek Formation

The new data for the Trooper Creek Formation are comparable with trends identified within the formation throughout the remainder of the Seventy Mile Range Group (e.g. Stolz, 1995). On a plot of Zr/TiO$_2$ versus Nb/Y the samples range in composition from high-silica dacite to basalt (Fig. 3.5C). Some lavas and intrusions classified as rhyolite petrographically (because they contain 5-7 modal percent quartz phenocrysts, 1-3 mm across) plot in the high-silica dacite field. These samples are characterised by having greater than 70 wt% SiO$_2$ but higher Ti/Zr ratios than rhyolitic lavas from the Mount Windsor Formation (Fig. 3.6C; Stolz, 1995). Other lavas and intrusions in the Trooper Creek Formation that generally contain less than 1 modal percent quartz phenocrysts (mapped as rhyodacite and dacite) also plot in the high-silica dacite field (cf. Stolz, 1995).
Figure 3.5. Geochemistry of coherent lavas and intrusions from the Mount Windsor Formation (MWF) and Trooper Creek Formation (TCF) in the study area. (A) Ti/Zr vs. SiO$_2$, (B) K$_2$O vs. SiO$_2$ showing the low, medium and high K calc-alkaline series (after Rollinson, 1993); (C) Zr/TiO$_2$ vs. Nb/Y (symbols as in A; after Winchester and Floyd, 1977); (D) TiO$_2$ vs. SiO$_2$; (E) P$_2$O$_5$ vs. SiO$_2$; (F) Ti/Zr vs. MgO.
Figure 3.6. Plots of (A) Ti/Zr vs CaO, (B) Fe₂O₃ vs SiO₂, (C) P/Zr vs Ti/Zr, (D) TiO₂ vs Zr, and (E) Th vs TiO₂ for coherent lavas and intrusions in the study area. Plots C and E effectively discriminate the three major suites identified in the area. The rhyolite clast in the Kitchenrock Hill Member of the Trooper Creek Formation (TCF) falls within the field of the Mount Windsor Formation (MWF). Fields in D and E are based on data from Berry et al. (1992), Stolz (1995) and this study.
Overall the range of compositions from high-silica dacite to andesite fall along linear trends which are consistent with the wider sample set reported by Stolz (1995) (e.g. Fig. 3.5D-F). The compositions of coherent rocks from the Highway Member of the Trooper Creek Formation are similar throughout the field area and do not appear to vary stratigraphically. Coherent volcanic facies are minor in the Kitchenrock Hill Member and samples suitable for analysis have not been identified. Berry et al. (1992) identified two suites of andesitic rocks within the Trooper Creek Formation: a low Ti-Zr group, and a relatively high Ti-Zr group. In the study area, the low Ti-Zr andesite (Highway Member) typically has Zr concentrations of 26 to 144 ppm and TiO$_2$ abundances < 1.1 wt% (Fig. 3.6D). The high Ti-Zr suite (Zr > 100 ppm and TiO$_2$ > 1 wt%; Berry et al., 1992) is not represented.

The silicic coherent rocks of the Trooper Creek Formation are distinguishable from rhyolites, rhyodacites and dacites in the Mount Windsor Formation by their higher P/Zr and Ti/Zr values (Fig. 3.6C). In a plot of TiO$_2$ vs. Th the former are also generally lower in Th and higher in TiO$_2$ (Fig. 3.6E; Stolz, 1995). In the study area, the Mount Windsor Formation rhyolite, rhyodacite and dacite intervals also typically have lower Fe$_2$O$_3$ concentrations (Fig. 3.6B) and higher Y and Rb abundances.

Analyses of volcaniclastic rocks in the Trooper Creek Formation were limited to a single rhyolite clast (94-50) from the Kitchenrock Hill Member. The sample comes from a normally graded pumice-crystal breccia and sandstone unit near the contact with the Mount Windsor Formation. In plots of P/Zr vs. Ti/Zr (Fig. 3.6C) and Th vs. TiO$_2$ (Fig. 3.6E) the rhyolite clast falls within the field of the Mount Windsor Formation.

**Andesite dykes**

Andesite dykes are abundant in the Mount Windsor Formation and Trooper Creek Formation. In the Trooper Creek Formation, some of the dykes are comagmatic with lavas (Stolz, 1995). The dykes have SiO$_2$ concentrations in the range of 50.8 to 58.6 wt% and relatively high contents of TiO$_2$ (1.4-2.3 wt%) and moderate P$_2$O$_5$ (0.19-0.35 wt%) (Fig. 3.5D-E; Stolz, 1995). Their high TiO$_2$ contents and lower Th contents clearly distinguish them from a suite of unaltered andesite dykes which cross-cut massive sulfide mineralisation at Highway-Reward (Fig. 3.6E). The dykes at Highway-Reward plot along a similar trend to data for the Mount Windsor Formation and Trooper Creek Formation (Fig. 3.5D-E, 3.6B) suggesting they are probably cogenetic with other lavas and intrusions in the Seventy Mile Range Group.
3.9 Summary

Detailed facies analysis of a central part of the Seventy Mile Range Group, between Coronation homestead and Trooper Creek prospect, has led to a better understanding of lithofacies characteristics of the four component formations. The Mount Windsor Formation is a thick sequence of massive rhyolite, rhyodacite and subordinate dacite. Volcaniclastic rocks form a minor component of the formations and pyroclastic rocks are not present. Associations of coherent and autoclastic facies form lavas, domes and syn-volcanic sills and dykes but provide little unambiguous information on the depositional setting. Minor hyaloclastite is the only evidence for eruption in a subaqueous environment. The Trooper Creek Formation is subdivided into two members, the Kitchenrock Hill Member and the overlying Highway Member. The stratigraphic subdivision is based largely on lithological variations, which reflect different provenance characteristics. Volcaniclastic deposits in the Kitchenrock Hill Member are characterised by abundant rounded clasts that were probably sourced from the Mount Windsor Formation. Rounding occurred in a high-energy environment prior to redeposition, suggesting that the source areas were subaerial to shallow marine. In contrast, the Highway Member is a complex association of syn-eruptive volcaniclastic deposits, syn-sedimentary intrusions, lavas and siltstone. Both the Kitchenrock Hill Member and the Highway Member were deposited in a submarine environment. At Highway South prospect and Trooper Creek prospect, the Trooper Creek Formation is overlain by the Rollston Range Formation. At these locations, angular to well-rounded metamorphic- and granitic-basement derived fragments are increasingly abundant near the top of the Highway Member. This mixed interval is transitional and suggests that the contact is conformable. The overlying Rollston Range Formation comprises sandstone and siltstone beds that are dominated by volcanic quartz and feldspar but contain significant basement derived detritus. Rare pyroclast-rich units and dacitic lavas occur within the Rollston Range Formation (e.g. Berry et al., 1992; Henderson, 1986).

Coherent volcanic rocks in the Trooper Creek Formation and Mount Windsor Formation are geochemically similar to those from the equivalent formations in the remainder of the Seventy Mile Range Group (e.g. Berry et al., 1992; Stolz, 1995). The geochemical data support the stratigraphic correlations proposed in the current study. In particular, the position of the contact between the Trooper Creek Formation and Mount Windsor Formation in some parts of the study area (e.g. Highway East, Trooper Creek) has been modified from that of previous authors.

The recognition of regionally extensive key facies associations within the Trooper Creek Formation has important implications for mineral exploration within the Seventy Mile Range Group.
Chapter 4

A shoaling felsic to intermediate volcanic succession in the Highway Member of the Trooper Creek Formation
Chapter 4

A shoaling felsic to intermediate volcanic succession in the Highway Member of the Trooper Creek Formation

4.1 Introduction

This chapter focuses on the physical volcanology of the volcanic and sedimentary sequence exposed in the southern part of Trooper Creek prospect (Figs. 4.1 & 4.2). This area provides some of the best exposures of the upper part of the Highway Member (Trooper Creek Formation) and the transition to the overlying Rollston Range Formation. Strata in this area strike east-west and dip at angles predominantly between 44-60° S. The sequence is cut by several S to SSE striking minor faults and is locally folded on a small scale, but no major structural complexities have been detected. A discontinuous cover of Tertiary alluvium and Recent surficial deposits obscures large parts of the sequence.

The lithofacies exposed at Trooper Creek prospect provide an opportunity to document the complex history of a shoaling submarine andesitic scoria cone. The volcanic and post-volcanic history, changes in the depositional and eruptive setting, and eruption styles are considered. The volcanic centre provides an analogue for other shoaling andesitic scoria cones constructed by strombolian eruptions in a relatively shallow marine environment. The model for this style of volcanism contrasts with that of surtseyan eruptions which are relatively well understood (e.g. Walker and Croasdale, 1972; Kokelaar and Durant, 1983a,b; Cas et al., 1989) and important in the construction of many shoaling volcanic centres.

Basaltic volcaniclastic units on deep marine (1000 m) seamounts have been interpreted as the deposits from submarine lava fountaining (e.g. Batiza et al., 1984; Smith and Batiza, 1989; Batiza et al., 1989). In ancient subaqueous volcanic successions, similar volcanic deposits have been documented (e.g. Carlisle, 1963; Yamagishi, 1982, 1987; Staudigel and Schmincke, 1984; Schmincke and Sunkel, 1987). However, relatively little is known about the eruptive style and products of fire fountain and strombolian eruptions in these environments.
A shoaling volcanic succession

Figure 4.1 Simplified geological map showing the distribution of volcanic and sedimentary units within the three principal lithofacies associations. Strata dip and young to the south.
4.2 Lithofacies associations

In this part of the Seventy Mile Range Group, the Highway Member is characterised by rapid vertical and lateral changes in volcanic and sedimentary facies. Andesites dominate over dacites and both rock types are represented by lavas, intrusions and a variety of volcaniclastic facies. The 14 principal lithofacies can be grouped into three compositionally distinct lithofacies associations. These associations have genetic significance with respect to eruption style, proximity to source, volcano type, provenance and depositional setting.

(1) The sedimentary facies association mostly comprises siltstone and sandstone units that can be volcanic or non-volcanic. Intercalated dacitic to rhyodacitic volcaniclastic breccia and sandstone beds record occasional influxes of volcanic debris into the depositional environment. These beds record explosive silicic eruptions at subaerial or shallow subaqueous volcanic centres, and were emplaced by cold, water-supported, high-concentration turbidity currents.
(2) The andesitic facies association includes both primary volcanic and resedimented volcaniclastic lithofacies. The primary volcanic facies includes lavas and volcaniclastic facies for which eruption, transportation and deposition were directly controlled by volcanic processes. The resedimented lithofacies are composed of clasts that were initially formed and deposited by volcanic processes and subsequently redeposited although not significantly reworked (McPhie et al., 1993). This facies is dominated by graded andesitic scoria breccia facies and globular clast-rich breccia facies.

(3) Lithofacies of the dacitic volcano-sedimentary facies association comprise a complex association of resedimented facies, siltstone and microbialites. Exposures of this facies association are largely limited to one creek section within the central part of the area (around 7741900 mN, 426800 mE; Fig. 4.3, 4.4).

All three associations occur within a 1300 m thick section through the upper part of the Highway Member (Fig. 4.2). The sedimentary facies association dominates the base of the section and includes minor dacitic and rhyodacitic units. The andesitic volcanic facies association overlies the sedimentary facies association. The dacitic volcano-sedimentary facies association largely overlies a disconformity at the top of the andesitic facies association.

4.2.1 Sedimentary facies association

*Massive to laminated siltstone facies*

Siltstone in the southern part of the area is massive and thickly bedded, or is planar laminated. The siltstone beds are laterally continuous, lack mud cracks, soil horizons and evidence of tractional reworking. They are mostly pale to dark grey but on weathered surfaces are yellow brown. The texture in thin-section consists of a cryptocrystalline mosaic of sericite and quartz. Around 7742800 mN, 426700 mE very thinly bedded (cm's thick) and laminated siltstone contains bedding-parallel gossanous Fe-Mn oxide bands. No primary sulfide minerals remain. Intervals of the siltstone facies range from <1 m up to 160 m in thickness.

*Origin and significance of facies*

Recrystallisation hampers textural interpretation of the siltstone. They may record settling of fines from the dilute currents trailing turbidity currents, water-settled fallout, pelagic or hemi-pelagic sediment or deposits from weak bottom currents. Although the composition/provenance is not clear, they record suspension-settled sedimentation in a relatively deep and/or quiet water environment, below storm wave base.
Figure 4.3 Detailed outcrop map showing the distribution of the principal lithofacies within the lower part of the dacitic volcano-sedimentary facies association. The highly irregular contact with the underlying andesitic facies association is interpreted as an unconformity.
Figure 4.4 Measured sections presented as graphic logs for major lithologies in the Trooper Creek Formation in the area of the Trooper Creek Prospect. Locations of sections A-E are given in Figures 4.1 and 4.3.
4.2.2 Andesitic facies association

**Massive andesite**

Coherent andesite at Trooper Creek prospect is plagioclase- and pyroxene-phyric. Phenocrysts (2%, 1 mm long) are euhedral and distributed evenly throughout a groundmass which comprises a network of small (160 μm) aligned feldspar laths, pyroxene crystals and interstitial chlorite (probably after glass). Pyroxene (clinopyroxene) is unaltered or has been variably pseudomorphed by chlorite. The andesites typically contain 3-30% (mostly 3-5 wt%) round to ellipsoidal vesicles or amygdales, 0.5 mm to 5 cm across. Amygdales are filled with quartz, chlorite, carbonate, feldspar or zones of chlorite-feldspar or chlorite-quartz.

Intervals of andesite range between 10 m and 70 m in thickness. Coherent andesite sometimes grades into associated peperite and autobreccia facies along contacts with underlying or overlying volcano-sedimentary rocks.

**Non-stratified andesitic breccia facies**

At Trooper Creek prospect (around 7742100 mN, 426800 mE), coherent and moderately fractured andesite passes outward through a zone of in situ breccia into a framework-supported breccia of the same composition, 1-2 m thick (Fig. 4.5A). The contact between brecciated and coherent andesite is highly irregular. Coherent andesite near the contact has more abundant but smaller vesicles than towards the interior (3-5% vesicles, 1-5 cm across). Trains of aligned ellipsoidal vesicles are tightly contorted. Coherent andesite within 15-30 cm of the breccia facies displays platy joints which mirror the vesicle trains and are spaced 1-2 cm apart. Platy joints and vesicle trains are parallel to contacts with the breccia or are cut by the contact at a high angle. Clasts in the breccia have blocky shapes bound by planar to curvilinear margins along which are joints that penetrate short distances towards clast interiors ("tiny normal joints"; Yamagishi, 1979). Vesicle trains in adjacent clasts have different orientations and can be traced into the coherent facies only within an intervening discontinuous, 10-15 cm wide zone of in situ breccia.

**Origin and significance of facies**

Internal variations within the andesitic breccia facies reflect varying roles for quench fragmentation and autobrecciation in the generation of the breccia (cf. Pichler, 1965). The in situ breccia is regarded as hyaloclastite formed through the propagation of networks of thermal contraction fractures into the cooling andesite (cf. Yamagishi, 1979). Tiny-normal joints and gradations between coherent facies and in situ breccia facies, are characteristic of hyaloclastite (e.g. Allen, 1988). Although jigsaw-fit texture is lost,
fragments in clast-rotated breccia retain shapes characteristic of quench fragmentation. The hyaloclastite has undergone rotation and separation of clasts and probably further granulation due to continued movement of the lava, indicating a role for autobrecciation.

**Siltstone pod-bearing andesite facies**

Exposures of this facies are limited to one segment of a small unnamed creek in the southern segment of the area at the contact between weakly vesicular andesite and the overlying planar laminated siltstone. Most of the observed contacts are broadly conformable with bedding although locally discordant. The contact zone is marked by intricate interpenetration of andesite and pale yellow-brown siltstone and fine-grained sandstone. Downward from the contact, tongues of sediment penetrate the andesite and sedimentary inclusions up to 40 cm across occur within coherent andesite, up to several metres away from contacts with the sediment. The macroscopic texture of the andesite is unchanged from that of the coherent facies described above. However, the siltstone is cherty and pale green at the contact suggesting that it is indurated.

**Origin and significance of facies**

The interfingering of the siltstone and magmatic component, and induration of the siltstone at contacts, favour interpretation of this facies as peperite. Peperite provides evidence for the mixing of magma or lava and wet unconsolidated sediment and has been described in many subaqueous volcanic successions (e.g. Fisher, 1960; Schmincke, 1967; Williams and Mc Birney, 1979; Brooks et al., 1982; Kokelaar, 1982; Busby-Spera and White, 1987; Hanson and Wilson, 1993; Brooks, 1995). Fluidal contacts between the igneous component and sediment are thought to form where a vapour film is established and maintained at the magma-sediment interface. The vapour film insulates the magma from the wet sediment and suppresses both quench fragmentation and steam explosions. Sediment is displaced along the contact zone until cooling below a critical temperature causes the steam to condense and sediment to be deposited (Chapter 5, Appendix A).

**Stratified andesitic breccia and sandstone facies association**

This association comprises graded andesitic scoria breccia, cross-stratified andesitic breccia and sandstone, and globular clast-rich andesitic breccia facies. The cross-stratified andesitic breccia and sandstone facies overlies the graded andesitic scoria breccia facies, and is overlain by globular clast-rich breccia facies. These facies are essentially monomictic, but chlorite-altered clasts, silicified clasts and hematite-altered clasts can occur in the one bed. Clasts are typically elongate, platy or equant in shape. The degree of
Figure 4.5

Andesitic facies association.

(A) Non-stratified andesitic breccia facies. Coherent and mildly fractured andesite passes outward into a framework supported autoclastic breccia (b) of the same composition. Coherent andesite within 15-30 cm of the breccia displays platy joints (arrow). 7742050 mN, 426800 mE.

(B) Graded andesitic scoria breccia facies. The breccia has a coarse-grained base (bottom of page) and fine, sandstone top. Note pale silicified clasts. 7741900 mN, 426800 mE.

(C) Abundant uncompacted scoria and subordinate trachytic clasts in the graded andesitic breccia facies. Vesicles are filled with calcite, chlorite or zones of hematite-chlorite. Formerly glassy vesicle walls have altered to calcite and quartz. Plane polarised light. Sample 95-318.

(D) Graded andesitic scoria breccia facies. Classical perlitic defined by arcuate chlorite-filled fractures (arrow) surrounding kernels of non-fractured, calcite-altered andesite. Sample 95-132.

(E) Cross-stratified andesitic breccia facies. This breccia includes both planar and cross-stratified intervals. 7741800 mN, 427000 mE.

(F) Globular clast-rich andesitic breccia facies. This massive to diffusely bedded breccia is composed of fragments of bombs (arrow) which are supported in a framework of blocky, weakly vesicular clasts. 7741750 mN, 427000 mE.

(G) Globular clast-rich andesitic breccia facies. One bomb has a partially intact, poorly vesicular rind (arrow) surrounding a more vesicular interior. 7741750 mN, 427000 mE.

(H) Photomicrograph of the globular clast-rich andesitic breccia facies. Within the clasts, flow aligned microlites (arrow) are separated by chlorite, quartz and hematite (probably after glass). Patches of calcite occur between some clasts. Plane polarised light. Sample 95-132; 7741750 mN, 427000 mE.
vesicularity varies and two principal clast populations are present. Poorly vesicular clasts with 0-5% (visual estimate) amygdales are bound by planar to cuspate margins. Breakage across microlites in the groundmass generated clast margins which are planar but finely jagged. Scoriaceous clasts have 25-50% (visual estimate) round-ellipsoidal amygdales and have vesicle-controlled bounding faces which, although finely cuspatate, are broadly planar. Vesicles are filled with either calcite, chlorite, sericite, hematite or zones of quartz-chlorite or hematite-chlorite.

In thin-section, clasts are aphyric or contain 1-2% feldspar microphenocrysts less than 1 mm long and rare pyroxene phenocrysts. The trachytic groundmass comprises unaltered feldspar (sanidine and plagioclase) microlites (150-500 μm long) and minor interstitial sericite, chlorite and fine-grained hematite. In many clasts, the original groundmass textures have been overprinted by a quartzfeldspathic mosaic or by calcite, chlorite, epidote and/or sericite alteration. Colloform-like bands of hematite nucleate along some clast margins.

Amygdales in some clasts are round, whereas in others there are varying proportions of round and ellipsoidal amygdales. Small amygdales tend to be round, whereas large amygdales are more ellipsoidal. Some round vesicles must have formed late because groundmass microlites project into them and elsewhere feldspar microlites are displaced by them. In some cases, intense silicification, chloritisation and/or carbonate alteration have obscured or destroyed vesicular textures and clast margins, generating a breccia which is dominated by apparently poorly vesicular fragments (e.g. 95-12, 95-318). Relic chlorite- or quartz-filled vesicles are the only indication of former scoriaceous fragments in these domains. Some clasts, and the vesicles within them, are outlined by fine-grained hematite.

Graded andesitic scoria breccia facies

Medium to thick, normally graded breccia beds are dominated by clasts 2-5 cm across although rare clasts up to 20 cm occur at their base (Fig. 4.5B). The breccia and sandstone beds are framework supported, moderately well sorted, and lower contacts are sometimes erosion surfaces. The grains are separated by calcite, chlorite and hematite that may have replaced a fine-grained matrix. Thicker beds are overlain by a series of thinner, normally graded beds, 1-15 cm in thickness. Coarse sandstone at the base of these beds comprises fragments 3 mm across and is sometimes diffusely planar laminated. Tops of beds are mostly finer sandstone (0.2 to 1 mm) although some are siltstone. Planar and cross-lamination characterise the sandstone–siltstone tops of some beds. Intercalated planar laminated siltstone units, a few centimetres to several metres thick, separate some breccia beds.
Poorly vesicular clasts account for 1% of most beds but comprise up to 10% of some units (95-132). The remainder of the fragment population consists of scoriaceous clasts with 30-50% round-ellipsoidal vesicles (Fig. 4.5C). Rare tube vesicle scoria with hematite-lined vesicle walls are present. Some clasts have a trachytic groundmass texture, whereas others may have been more glassy with sparse feldspar microlites. In these samples (e.g. 95-132), the groundmass shows classical perlite (e.g. Ross and Smith, 1955) comprising arcuate, overlapping and intersecting chlorite-filled fractures (Fig. 4.5D). Rare clasts have concentric fractures which define margins, cut microlites in the groundmass, and generate rounded clast shapes. Perlitic fractures suggest that the groundmass of these fragments was in part formerly glassy. In many scoriaceous clasts, recrystallisation or alteration has obscured groundmass textures. Large clasts (10-20 cm) are mostly elongate with irregular shapes but some fluidal spindle-shaped clasts occur. The vesicularity of these clasts varies but is generally low, and some clasts have dense margins with a few ellipsoidal vesicles concentrically arranged around a more vesicular interior.

Intervals of this facies range from 15 m to greater than 120 m in thickness. Thinner intervals comprise a number of graded units (centimetres to metres thick); thicker intervals are made up of a small number of graded units each up to 40 m thick.

*Cross-stratified andesitic breccia and sandstone facies*

This facies consists of high-angle trough cross-bedded sandstone and planar laminated sandstone, interbedded with normally graded breccia to sandstone beds (Fig. 4.5E). Cross-beds occupy broad shallow channels and dip to the south. Stratifications are 0.5 to 3 cm thick and characterised by a coarser, normally graded, lower interval comprising fragments 2 to 7 mm in size, sharply overlain by a thinner sandy top with fragments 0.6 to 1 mm across.

The coarse sandstone to breccia is monomictic, moderately to well sorted and devoid of fine matrix. Five to seven percent of the framework consists of scoriaceous clasts (e.g. 95-317A, 95-272). The remaining fragment population consists of poorly vesicular (1-5% vesicles) fragments. Formerly glassy vesicle walls are now chlorite, sericite and calcite. Chlorite occurs as finely crystalline patches or as radial aggregates nucleating on clast margins and extending out into patches of calcite. A thin (10 μm) rind of recrystallised quartz bounds many fragments. Quartz, chlorite and calcite alteration of clast margins has resulted in smoother, rounder and more bulbous outlines than for weakly altered clasts, creating a more extensive apparent matrix domain.
Globular clast-rich andesitic breccia facies

The globular clast-rich andesitic breccia facies is monomictic, moderately to poorly sorted and characterised by up to 10% globular shaped clasts, 5-50 cm long (Fig. 4.5F-G). Units are massive or else show coarse-tail normal grading. One thin bed (<1 m) is normally graded. Contacts between beds are indistinct and defined chiefly by the abundance of blocks. Single units are mostly between 35 and 40 m in thickness, but because outcrop is poor, the boundaries between some units may not be exposed.

The globular clasts are elongate to ellipsoidal or sub-spherical in shape with smooth, hackly, planar or curviplanar margins (Fig. 4.6). Many clasts have poorly vesicular rinds 5 to 10 mm thick, surrounding a more vesicular interior (20-30% vesicles) in which the size and abundance of vesicles increases inward (Fig. 4.6B-C). Other clasts have surfaces

Figure 4.6 Field sketches of representative examples of bombs from the globular clast-rich andesitic breccia facies. (A) intensely hematite-altered poorly vesicular clast; (B-C) these bombs have relic poorly vesicular rinds surrounding a more vesicular interior; (D) irregular block with margins that are in part the former walls of vesicles. This clast type is interpreted to be a fragment of the vesicular interior of bombs that have disintegrated.
that lack rinds, and some are entirely coarsely vesicular with margins which are in part the walls of vesicles (Fig. 4.6D). In many clasts, vesicles in the centre are spherical and up to a few centimetres in diameter, and those in the outer portion are ellipsoidal and their long axes are aligned parallel to clast margins. The poorly vesicular interior and non-vesicular rind of a few clasts are separated by a 1-2 cm wide zone with 20 % aligned ellipsoidal vesicles up to 5 cm long. In thin-section, vesicles, phenocrysts and feldspar microlites in the pilotaxitic groundmass are truncated at clast margins and no glassy rinds are preserved. Sericite, epidote or quartz occur in the groundmass and have also replaced some microlites.

The globular clasts are supported by a matrix composed of elongate to equant fragments with irregular planar margins (Fig. 4.5H) and ranging from 300 μm to 5 mm across. Up to 90% of these fragments are poorly vesicular with 0-3% round to ellipsoidal vesicles, 50-900 μm across. The remaining fragment population has 10-25% vesicles, some defining cuspat e inflections along planar clast margins. Some smaller clasts are zoned with respect to vesicularity and have groundmass textures and vesicle abundances which suggest that they are fragments of blocks.

Intervals of this facies range from 40 m to greater than 70 m in thickness. Thinner intervals appear to comprise one unit; thicker intervals are made up of a few units each possibly up to 40 m thick.

Interpretation

Fragmentation processes
In volcaniclastic deposits fragment shape may be an indicator of the style of eruption, fragmentation mechanisms and eruption environment (e.g. Kokelaar, 1986; Heiken and Wohletz, 1985). However, particle shapes may be modified during transport and reworking prior to final deposition (e.g. Kokelaar and Romagnoli, 1995). The shapes of fragments in the andesitic breccia facies are dependent largely on their vesicularity. Poorly vesicular fragments have blocky, angular shapes with rare embayments where the clast margins cross the walls of vesicles. Vesiculated fragments that formed by breakage across vesicles have more irregular shapes, although planar surfaces are common.

Scoria — The scoraceous character of some fragments, particularly within the graded scoria breccia, suggests they are pyroclasts, whereby magma fragmentation was driven primarily by exsolution of magmatic volatiles. However, if vesiculated magma interacts with water, non-explosive quenching or phreatomagmatic fragmentation of the magma can also generate clasts with vesicle-controlled surfaces (Kokelaar and Durant, 1983a,b; Houghton and Wilson, 1989). Quenching may initiate magmatic explosions by breaking
vesicle walls allowing rapid decompression and expansion of the magmatic volatiles (Cas et al., 1989).

In the graded andesitic scoria breccia facies, scoriaceous clasts with concentric fractures and perlite imply a role for quench fragmentation in disintegration of the melt. The concentric fractures bound some clasts and are similar to fractures in subaqueous lavas interpreted to record brittle fracturing by thermal contraction (e.g. Yamagishi, 1987, 1991; Goto and McPhie, 1996). Increased rounding of a few clasts appears to record spalling of concentrically fractured margins from non-fractured cores during transport (cf. Cas et al., 1994). Perlitic cracks develop in response to hydration of glass (Ross and Smith, 1955; Friedman and Smith, 1958; Friedman et al., 1966). Although perlitic fractures are not solely the result of quenching, the release of residual stress acquired during rapid cooling is probably important in forming perlite (e.g. Allen, 1988; Yamagishi and Goto, 1992; Cas et al., 1994; Davis and McPhie, 1996).

Globular clasts — Fluidal-shaped clasts in the graded andesitic scoria breccia facies and globular clast-rich andesitic breccia facies have shapes similar to bombs. Bombs are not restricted to subaerial settings. A variety of bomb-like pyroclasts variably termed water-chilled bombs, pyroclastic pillow breccia (Yamagishi, 1982, 1987), scoria pillows (Staudigel and Schmincke, 1984; Schmincke and Sunkel, 1987) clots, spheroids and globules (Carlisle, 1963) have been described from submarine volcaniclastic deposits (Table 4.1). These occur in shallow subaqueous deposits as a result of direct fallout from explosive eruptions (Yamagishi, 1982, 1987) or shallow subaqueous fire fountaining associated with high magma discharge and eruption through a restricted conduit (e.g. Staudigel and Schmincke, 1984). Downslope resedimentation of primary deposits from littoral or shallow subaqueous settings can transport such pyroclasts into deep water (Dolozi and Ayres, 1991). In situ bombs, shards and spatter also occur in modern, deep submarine (> 1000 m) settings on seamounts near the East Pacific Rise, and have been interpreted as proximal “hyaloclastite” deposits from submarine, mildly explosive lava fountaining associated with high effusion rates (Batiza et al., 1984; Smith and Batiza, 1989; Batiza et al., 1989).

Bombs from breccia beds in the Trooper Creek Formation are either ellipsoidal to spindle shaped, with smooth surfaces, and similar to water-chilled bombs, or are elongate with irregular margins and more similar to “clots” described by Carlisle (1963). Fluidal-shaped bombs have been shaped by surface tension during flight, whereas the more irregular clasts are probably the result of the tearing apart of the vesiculating magma. Planar and curvilinear fractures which form the broken margins of some globular-shaped fragments may be quench fractures that formed as the pyroclasts fell into or through water. Bombs from subaerial eruptions are also often fractured and sometimes disintegrate on impact or
Table 4.1 Characteristics of fluidally-shaped clasts and associated deposits from subaqueous and subaerial volcanic eruptions.

<table>
<thead>
<tr>
<th>Water-chilled bombs</th>
<th>Water-chilled scoria</th>
<th>Pyroclastic pillow breccia</th>
<th>Clots in broken-pillow breccia</th>
<th>Globules in isolated-pillow breccia</th>
<th>Scoria pillows &amp; lobes</th>
<th>Subaerial bombs</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Environment</strong></td>
<td></td>
<td></td>
<td></td>
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<tr>
<td>Shallow subaqueous</td>
<td>shallow subaqueous</td>
<td>subaqueous</td>
<td>subaqueous</td>
<td>subaqueous</td>
<td>subaqueous</td>
<td>subaerial</td>
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<td>Shapes</td>
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<tr>
<td>Elongate ellipsoidal; wrinkled margins</td>
<td>ellipsoidal; wrinkled margins</td>
<td>irregular nodular-amoeboidal clots</td>
<td></td>
<td>branching, irregular, ellipsoidal-amoeboidal agglomerate; stringer &amp; bomb-shaped clots</td>
<td></td>
<td></td>
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<tr>
<td>Size</td>
<td></td>
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<tr>
<td>5-10 cm</td>
<td>lapilli-sized (2.64 mm)</td>
<td>1-30 cm</td>
<td>0.25 mm-10 cm</td>
<td>mostly 10-50 cm across; pillows up to 2 m diameter</td>
<td>64 mm to 6 m</td>
<td></td>
</tr>
<tr>
<td>Chilled margins</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
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<tr>
<td>Black chilled glossy rim with palagonite envelope; tiny cracks perpendicular to margin; colour zones</td>
<td>black chilled glossy rim with palagonite envelope; tiny cracks perpendicular to margin; colour zones</td>
<td>bleached chloritized former palagonite rims</td>
<td></td>
<td>devitrified chloritized sideromelane-tachylyte rims</td>
<td>partially detached quenched rinds</td>
<td>bombs generally large, glassy; commonly cracked</td>
</tr>
<tr>
<td>Vessicles</td>
<td></td>
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<tr>
<td>Concentric zones; core: spherical to ellipsoidal, few cm across. Outer core &amp; chilled margin: elongate, sun diameter</td>
<td>concentric zones; core more vesicular than interior</td>
<td>vesicular to scoriaceous ellipsoidal vesicles</td>
<td></td>
<td>zones of concentric vesicles</td>
<td>40 to &gt;80 % vesicles; vesicularity increases towards interior</td>
<td>variably vesicular; concentric layers; a vesicles more abundant &amp; larger in core than rim; rarely hollow</td>
</tr>
<tr>
<td>Matrix</td>
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<tr>
<td>Lapilli-sized scoria and/or ash with associated accretionary lapilli</td>
<td>no interstitial ash</td>
<td>highly vesicular scoriaceous lapilli and ash; more abundant than &quot;pillows&quot;</td>
<td></td>
<td>n effusive matrix of ellipsoidal globules, glassy, shards; teardrop- &amp; spindle-shaped; little in situ cracking</td>
<td>detached vesicular rinds and lapilli-sized, very vesicular stringer-shaped sideromelane fragments; minor ash</td>
<td>irregular vesicular cinder &amp; scoria</td>
</tr>
<tr>
<td>Bedforms</td>
<td></td>
<td></td>
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<td></td>
<td></td>
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<tr>
<td>Doubly-grained upper portion; massive lower portion</td>
<td>well sorted; graded; units a few 10's cm thick</td>
<td>massive normal to reverse graded lower part &amp; fine-grained, bedded top</td>
<td></td>
<td>non-mottled, unsorted</td>
<td>non-mottled, unsorted</td>
<td>beds 1-5 m thick</td>
</tr>
<tr>
<td>Eruption style</td>
<td></td>
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<tr>
<td>Water-sentided fall &amp; mass flow deposits from shallow subaqueous strombolian eruption</td>
<td>water-sentided fall; deposits from shallow subaqueous strombolian eruption</td>
<td>hydrovolcanic explosives followed by turbulent flow</td>
<td></td>
<td>rapid effusion</td>
<td>rapid effusion</td>
<td>vesicleted amoeboidal submarine lava flow (pillows) or submarine lava fountaining (clasts)</td>
</tr>
<tr>
<td>Reference</td>
<td></td>
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</table>
explode in flight (Macdonald, 1972; Self et al., 1974). The size and shape of the bombs in the globular clast-rich andesitic breccia facies strongly suggests that they formed by rupture of the melt due to volatile expansion or, more likely, high eruption rates or eruption through a restricted conduit (cf. Staudigel and Schmincke, 1984; Von der Borch, 1971).

Some submarine volcanlastic deposits include fluidal-shaped clasts similar to subaerial and water-chilled bombs, but are thought to have formed through budding from vesicular pillow-like lava lobes. This type of breccia is associated with the shoaling stage of La Palma seamount and is composed of lobes (“scoria pillows”) and “scoria lapilli/bombs” (Table 4.1; Staudigel and Schmincke, 1984). The scoria pillows were formed by effusion, and the lapilli and bombs which occur in upper part of the unit were thought to result from submarine lava fountains (Staudigel and Schmincke, 1984).

It is unlikely that bomb-shaped fragments from the Trooper Creek prospect formed by budding from lava lobes because: (1) the globular clast-rich breccia facies is not spatially associated with lavas; (2) gradational relationships between intact lava lobes and globular clast-rich andesitic breccia facies have not been observed; (3) the bombs have higher vesicularities than many compositionally similar lavas from the same stratigraphic section; and (4) the bombs differ in size and shape from the scoria pillow clasts.

Poorly vesicular clasts — Magmatic volatile-driven explosivity does not appear to have played a role in the generation of these clasts. Clasts which are segments of former bombs imply that quenching postdated the initial pyroclastic fragmentation mechanism. Other clasts have blocky to elongate shapes and planar margins consistent with brittle fragmentation. An origin involving quench fragmentation of a vesiculating magma or lava is inconsistent with the association with fluidal-shaped bomb fragments, and with the absence of gradational transitions into in situ brecciated and coherent facies. Rather, these components are more likely to be pyroclastic in origin.

Their shapes are similar to phreatomagmatic pyroclasts, which are typically blocky, and also sparsely to moderately vesicular (e.g. Fergusson et al., 1994). However, if vesiculation of the magma is advanced at the time of magma-water interaction, strongly vesicular, irregularly-shaped clasts can be abundant (Houghton and Wilson, 1989). The deposits of subaerial phreatomagmatic explosions are typically fine grained, “at most 8 mm in median diameter” (Wohletz, 1983). Some coarse phreatomagmatic deposits are poor in fine ash and show similar grainsize characteristics to Strombolian scoria deposits (e.g. Self et al., 1980; Kano et al., 1994). Low ash contents in phreatomagmatic deposits are thought to reflect inefficient magma-water interaction, for example, at high confining pressure and/or low magma/water ratios (e.g. Wohletz, 1983). In the Trooper Creek
prospect deposits, the coarse grain size and absence of fine ash suggests that phreatomagmatic explosions were suppressed, or that fines have been lost during transport to the depositional environment. Suppression of phreatomagmatic explosions is consistent with a wholly submerged vent and a low magma-water ratio in the vent, inhibiting the expansion of steam and resulting in incomplete fragmentation (cf. Cas et al., 1989). Both quenching and phreatomagmatic fragmentation were probably important.

Altered clasts — Pale silicified clasts are a subordinate but ubiquitous component in many beds. Clasts have groundmass textures and feldspar phenocryst abundances and sizes which are similar to other clasts in the breccia. Clasts are interpreted to represent either; (1) andesite clasts which were altered prior to redeposition (cf. Kokelaar and Romagnoli, 1995); or (2) accessory or accidental clasts. Patchy and mottled silicic domains which cut across clast boundaries suggests that many clasts are the product of post-depositional hydrothermal alteration.

Transport and depositional processes
The graded andesitic scoria breccia facies and cross-stratified andesitic breccia and sandstone facies are composed of clasts which were initially formed and deposited by pyroclastic processes, but were subsequently redeposited but not substantially reworked. Redeposition was more or less syn-eruptive; hence the deposits are essentially monomictic and primary clast shapes are largely unmodified.

The graded andesitic scoria breccia facies is interpreted to comprise the deposits of high-concentration gravelly turbidity currents. Breccia at the base of beds forms the S3 turbidite division (Lowe, 1982), whereas planar and cross-stratified sandstone and siltstone tops are the Td-e divisions. In the mass flows, clasts were probably supported by more than one mechanism. A combination of dispersive pressure and interstitial fluid turbulence is likely. Siltstone intervals between breccia beds are interpreted as suspension-settled sediment deposited from the dilute sediment clouds trailing mass flows and/or during ambient sedimentation.

The overlying cross-stratified andesitic breccia and sandstone facies is interpreted as a transition zone where traction currents and turbidity currents were both important in sediment transport and deposition. It is possible that the fragments deposited from turbidity currents, but with traction current modification during accumulation to form the cross-stratified intervals. Alternatively, the sandstone units may be genuine traction current deposits emplaced below wave base (Section 4.5) but within a channel setting. Both cross-stratification and planar lamination are identifiable. In relatively well sorted cohesionless sediments, these bedforms imply a transition from a low energy, lower-flow
regime toward high energy, high-flow regime. Channels within the facies may have been cut by turbidity currents or traction currents.

The restricted distribution of the globular clast-rich facies, its massive character and abundance of bombs suggests emplacement proximal to the source vent. The setting of similar bomb-rich deposits from other volcanic successions (e.g. Yamagishi, 1982, 1987; Staudigel and Schmincke, 1984) supports this interpretation. The overall upward fining in some units in the facies suggests a primary eruptive control or else sorting during transport to the depositional environment.

The globular clast-rich facies may be the deposit of water-settled fallout and/or sediment gravity flows. The bedforms, sorting and coarse grain size of the facies favour deposition as near-to-vent fallout. However, the unit probably also includes deposits emplaced by avalanching and rolling of pyroclasts down slope and from sediment gravity flows. Impact sags beneath coarse clasts were not observed, probably because the clasts landed on an accumulating pile of coarse, loosely packed pyroclasts (cf. Cas and Wright, 1987). Settling through water and/or transport from the initial depositional site may also explain the absence of bomb sags. Feeder dykes or spatter, characteristic of vent positions (e.g. Houghton and Landis, 1989), are not exposed in the study area suggesting that the facies flanks source vents. The diffuseness of bed boundaries, absence of any internal discordances and thickness of the globular clast-rich andesitic breccia facies suggest sustained eruptive activity or high eruption rates during deposition of the facies.

4.2.3 Dacitic volcano-sedimentary facies association

The dacitic volcano-sedimentary facies association is best exposed in one small unnamed creek in the central part of the area (Fig. 4.4 - section B).

*Planar laminated dacitic pumice breccia facies*

This breccia is monomictic and comprises a clast-supported framework of elongate irregular pumice clasts. Pumice fragments are feldspar-phric suggesting a dacitic composition. Although outcrop is poor, the pumice breccia appears to mantle the underlying stratified andesitic breccia and sandstone facies association and polymictic dacitic lithic-pumice breccia beds. The pumice breccia underlies and is partially replaced by a discontinuous lens of massive quartz-hematite ironstone (Chapter 6). Within a 45 cm wide zone beneath the ironstone lens, the pumice breccia contains abundant hematite nodules, 1-5 mm in diameter.
The pumice breccia consists of upper and lower planar laminated divisions separated by a more massive, diffusely laminated interval (2.9 m thick; Fig. 4.4 - section B; Fig. 4.7A). The lower division is 0.3 m thick but poorly exposed. The upper laminated division is 1 m thick and a thin (0.8-2 cm) pumiceous quartz-hematite horizon occurs 60 cm from the top. Laminae in the upper and lower divisions vary from 5 mm to 1 cm thick, and are defined by thin discontinuous concentrations of unaltered feldspar crystals, and generally also by colour. The breccia is pervasively sericite-hematite-altered but includes pale yellow, bedding-parallel, sericite-rich bands less than a millimetre thick. The breccia appears well sorted. However, most pumice clasts are compacted and strongly altered so that grain size variations are obscured.

Pumice clasts have elongate and ragged shapes. Surfaces normal to the tube vesicles are jagged, whereas the remaining surfaces are smooth and more planar. Formerly glassy pumice have been replaced by sericite and are strongly compacted. The pumice clasts and former vesicles within them are outlined by thin discontinuous trails of fine hematite. Some pumice clasts have been partially or entirely replaced by hematite. In these clasts, vesicle textures have been destroyed but the clast margins are preserved. In more quartz-rich-altered domains, relic uncompacted tube- and round-vesicle pumice clasts are preserved. Quartz and sometimes sericite and hematite have infilled vesicles and replaced former glassy walls. Some vesicle fills are zoned with an outer hematite zone passing into quartz and/or sericite zones. The phyllosilicate-altered pumice clasts are commonly deformed around the more competent quartz-hematite-altered pumice and feldspar crystals, and define a bedding-parallel compaction foliation.

Origin and significance of facies
The abundance of highly vesicular pumice within the laminated pumice breccia suggests this facies was sourced from explosive magmatic eruptions from a vent in a shallow subaqueous or subaerial environment. The angularity of clasts and absence of other particle types suggests the pumice clasts were deposited without having experienced significant reworking or abrasion and suggests that the facies is syn-eruptive. Although rich in juvenile pumice, there is no textural or lithofacies evidence for hot emplacement of the facies. Randomly oriented relic tube pumice clasts are preserved and the bedding-parallel foliation is interpreted as a diagenetic compaction foliation (cf. Niem, 1977; Allen and Cas, 1990; Branney and Sparks, 1990; Chapter 3).

The enclosing facies (siltstone, turbidites, microbialite; Section 4.4.1) suggest that deposition occurred in a subaqueous environment. Mantle bedding, good sorting and well developed planar lamination within the pumice breccia bed suggest that the pumice settled through the water column. The pumice may have settled through the water from fallout generated by subaerial eruptions, or from eruption columns above totally or partially
Figure 4.7

Dacitic volcano-sedimentary facies association (7741900 mN, 426800 mE).

(A) Planar laminated dacitic pumice breccia facies. The breccia is non-welded and pervasively hematite-sericite-altered. (B-H) Stromatolitic-oncolitic ironstone facies.

(B) The upper surface of this domed biostrome comprises a series of close linked hemispheroids (arrow). The intercolumn material and substrate (s) is tuffaceous sandstone or breccia.

(C) The structures consist of successive algal cappings or laminae over a nucleus such as an oncolite or irregularity within the substratum (s). At the base, laminae forming compound hemispheroids are not connected and neighbouring columns are separated by small amounts (<1 mm to 5 mm) of tuffaceous sandstone. The columns merge upward forming a continuous mat. Sample 95-216.

(D) Branching columnar stromatolites composed of fingers of gently- to moderately-convex hemispheroids. The intercolumn material comprises crystal fragments and shards. Sample 95-200.

(E) Oncolites comprising alternating quartz- and hematite-rich laminae concentrically arranged around a nucleus. Nuclei are either small ovoid quartz grains, single pumice fragments, crystal, or stromatolite fragments. Plane polarised light. Sample 95-200.

(F) Branching networks of filamentous structures (arrow) preserved within grains. Filament walls are made up of very fine-grained hematite. Cryptocrystalline quartz fills the space between the filaments. Plane polarised light. Sample 95-198.

(G) Fragments of trilobites (arrow) with characteristic cross-sectional morphologies and double thickness walls. The fragments have been infilled with hematite and walls are now composed of finely crystalline quartz and in some cases hematite. Plane polarised light. Sample 95-200.

(H) Brachiopod replaced by hematite (arrow). Plane polarised light. Sample 95-200.
submerged vents. Vitric ash was probably transported elsewhere in the eruption column or remained in suspension to be dispersed by currents. Siltstone beds which overlie the pumice breccia interval may include a component of this vitric ash.

The internal stratification in the pumice breccia may have been caused by changes in eruption column height, fragmentation processes or dispersal directions during the eruption. Shallow water currents may have enhanced stratification in the laminated divisions. The more massive breccia interval is interpreted to record a sustained period of rapid deposition and/or a sustained period of steady eruptive activity.

Distal water-settled fall deposits are composed of glass shards and crystal fragments and can be normally graded with coarse, crystal-rich bases and finer shard-rich tops. They typically occur in thin (centimetres to tens of centimetres) but widespread intervals interbedded with deep marine sedimentary deposits (e.g. Ninkovich et al., 1978; Lød debuted and Sparks, 1979; Sparks and Huang, 1980). Proximal water-settled fall deposits consist of coarse pumice clasts occurring together with finer but denser lithic clasts and crystal fragments. Deposits are typically internally planar stratified, massive or graded, very well sorted, and can be several metres thick (Dimroth and Yamagishi, 1987; Cashman and Fiske, 1991). The grainsize and thickness of the planar laminated pumice breccia facies suggests that it is a relatively proximal deposit.

**Stromatolitic and oncolitic ironstone facies**

"Microbialite" is a general term used for organosedimentary deposits composed of benthic microbial communities and sediments (Burne and Moore, 1987). Microbialites in ironstone at Trooper Creek prospect are composed of oncolites, stromatolites, pyroclastic components and fossil fragments. The microbialites occur in two layers (upper and lower microbialites) separated by a thick interval of polymictic lithic-pumice breccia (Fig. 4.4 - section B).

**Oncolites and stromatolites**

Oncolites are ellipsoidal to spherical, concentrically laminated microbialites. The term stromatolite is used here to describe microbialites with fine, relatively flat, internal laminations (e.g. Burne and Moore, 1987). Laminae in columnar and non-columnar stromatolites are typically very thin (8-20 μm) and are quartz-rich or hematite-rich. In some thick (70-80 μm) quartz-rich laminae, relic tube-vesicle pumice clasts, shards and crystal fragments are preserved. No attempt has been made to apply rigorous taxonomy to the microbialites studied. There is debate over the application of formal nomenclature to structures that are only partly biogenic and were probably constructed by a variety of different micro-organisms (e.g. Monty, 1977; Burne, 1994). Instead the structures have
been described using simple morphological terms following the scheme of Preiss (1976) and Walter et al. (1982) (Fig. 4.8). Both columnar and non-columnar stromatolites are present and these show a range of different internal forms. Non-columnar stromatolites have flat laminated, undulatory, pseudo-columnar, cumulate or columnar layered forms (Figs. 4.7B-C, 4.8). Columnar stromatolites consist of non-linked, vertically stacked thin hemispherical laminae. Columns have uniform variability, vary from upright to inclined and have stubby shapes (Fig. 4.8). Laminae are typically smooth but are sometimes wrinkled. Column walls are absent. Single non-branched columns are thin at the base, become wider towards the top, resembling turbinate and bulbous forms. Branching columnar varieties have multifurcate, coalesced or anastomosed forms. Laminae forming turbinate columns vary from steeply to moderately curved. Coalesced and anastomosed columnar varieties have steeply convex laminae or, in laterally linked segments, gently- to moderately-convex laminae. A number of the fingers branch upward, others terminate, and some coalesce upward by development of lateral linkages (Fig. 4.7D). Fingers have increasingly slender shapes as the proportion of pumice, crystals and shards increase.

Oncolites have ellipsoidal to spherical shapes and vary from 0.5 to 1.5 cm in diameter (Fig. 4.7E). The component laminae are concentrically arranged around a nucleus and can be continuous or comprise discontinuous overlapping shells. Nuclei can be: (1) small ovoid quartz grains; (2) quartz-hematite mosaic; (3) stromatolite fragments with interstitial tuffaceous sandstone; (4) single oncolites with ovoid quartz cores; or (5) a single pumice or tuffaceous sandstone fragment. Large oncolites typically have a single pumice or fragment of stromatolitic mat as the nucleus.

Branching networks of filamentous structures are preserved in the cores of oncolites or between grains in sample 95-198 (Fig 4.7F). The filaments form dense networks of randomly oriented fibres with cylindrical cross-sections, 5-8 \( \mu \)m in diameter. Filaments vary from 10 to 200 \( \mu \)m long. Filament walls are made up of very fine-grained hematite. Cryptocrystalline quartz fills the space between the filaments. The filaments are similar to those described and interpreted by Duhig et al. (1992) as microaerophilic chemolithotrophic bacteria or fungi. In most samples, the structure of the laminae forming stromatolites and oncolites has been destroyed during replacement of the components by quartz and hematite. Quartz is very fine grained (5-20 \( \mu \)m) but has locally recrystallised to coarser grains with 120° triple junctions. Cracks which follow or cut across laminae in some stromatolites and oncolites are filled with fine-grained (15 \( \mu \)m) quartz.

Matrix
The matrix between columns, clasts and oncolites consists of crystal fragments, pumice clasts and lithic fragments. Formerly glassy particles have been replaced by finely crystalline (20-50 \( \mu \)m) quartz and subordinate hematite. Some pumice fragments are now
Figure 4.8 Morphological terms for the description of microbialites with stromatolitic and oncolitic forms. Modified from Preiss (1976) and Walter et al. (1982).
entirely hematite. Relic pumice clasts, shards and the vesicles within them are outlined by thin films of hematite. Pumice clasts have uncompacted round- and tube-vesicles and are feldspar-phyric. Feldspar crystal fragments are largely unaltered but some have been pseudomorphed by polycrystalline quartz.

Fossils and fossil fragments

Fragments of trilobites display characteristic cross-sectional morphologies and, in the best preserved examples, double thickness walls (Fig. 4.7G). The fragments have been infilled with hematite and walls are now composed of either finely crystalline quartz or hematite. Other biogenic components are possible sponge spicules, gastropods and a brachiopod (Fig. 4.7H). Quartz and hematite alteration has destroyed or significantly modified the structure of these grains so that precise identification is difficult.

Mode of occurrence

In situ stromatolites occur as thin laminae 2-3 millimetres thick separated by pumice- and shard-rich sandstone and breccia (lower microbialite); as domed biostromes 1-3 cm thick in which there is an intimate association of stromatolites and associated trapped and bound detrital tuffaceous sediment (lower microbialite); and as subspherical bioherms built upon, and intergrown with, tuffaceous sandstone (upper microbialite). The in situ stromatolites are built on sandstone and pebble conglomerate that contain oncolites and stromatolitic clasts as well as pyroclastic components.

Domed biostromes — The upper surfaces of some mats are wrinkled into a series of small domes or hemispheroids, 0.5-1 cm in diameter and 1-4 mm high. The hemispheroids are linked laterally to neighbouring structures forming the mat. The distance between the hemispheroids is less than 2-3 mm, less than the diameter of the structures, so that they can be classified as close-linked hemispheroids. The structures consist of successive algal cappings or laminae over a nucleus such as a crystal or pumice fragment, an oncolite, or an irregularity within the substratum. In the simplest case, laminae are continuous between neighbouring hemispheroids (simple hemispheroids; Logan et al., 1964; Fig. 4.7B-C). At the base of the structures, compound hemispheroids are not continuous and neighbouring columns are separated by small amounts (<1 mm to 5 mm) of oncolite-bearing tuffaceous sandstone. The turbinate columns widen upward and merge where lateral linkages form a continuous mat covering the substratum. The non-columnar component of the structure can be pseudo-columnar or cumulate in form. Simple laterally linked forms occur within millimetres of compound forms in the same biostrome.

Subspherical bioherms — These are small dome-like structures 6-7 cm high and around 8 cm in diameter (Fig. 4.9A). The microstructural elements of the bioherms are compounded forms composed of concentrically laminated, columnar and non-columnar
elements. The pedestal of the structure comprises a series of oncolites from which extend radially arranged branching and turbinate columnar stromatolites (Fig. 4.9B). At the margins of the pedestal, initially discrete turbinate columns pass into a zone of pseudo-columnar stromatolites and planar laminated stromatolite. The upper part of the bioherm is a complex mixture of stromatolite forms (Fig. 4.9C). Turbinate columns pass into laterally linked hemispheroids forming pseudo-columnar structures. Branching and columnar layered stromatolites have grown upward from these surfaces.

Stromatolite clast-bearing sandstone and breccia — Sandstone beds are dominated by volcanic detritus or consist of grain-supported aggregates of sand- to gravel-sized oncolites, stromatolite fragments and volcanic fragments (principally crystals, pumice and shards). Breccias vary from matrix- to clast-supported. Stromatolite clasts have elongate slabby shapes, vary from 0.3-12 cm long, and are up to 4 cm thick. Clasts are separated by a matrix of volcanic fragments, oncolites and small stromatolite fragments.

Stromatolite clasts within sandstone and breccia beds are fragments of one stromatolite type or are compounded forms comprising several different morphologies. The former include non-columnar, undulatory and flat laminated forms. Others are branching columnar stromatolites. The morphological variability of compounded forms provides insights into the character of the bioherms or biostromes from which they were derived. Many clasts are composed of branching forms which locally develop lateral linkages and pass into pseudo-columnar varieties. Others are zoned with lower and upper zones of pseudo-columnar stromatolite separated by branching divisions. Many stromatolite clasts are encrusted by stromatolite laminae which grew out from the initial clast boundaries. Some encrustations completely surround clasts and comprise thin, smooth, concentric laminae. Other encrustations are restricted to the tops and margins of clasts and are compound forms involving turbinate columnar varieties, non-columnar undulatory forms and concentrically laminated stromatolite.

Origin and significance of facies
Microbialites at Trooper Creek prospect have grown by both trapping and binding of detrital volcanic particles on the organic film (cf. Monty, 1977; Burne and Moore, 1987). Compound structures that exhibit a change vertically from one stromatolite growth form to another are evidence for minor changes in the physical environment during deposition. Although it is not possible to attribute any one factor as responsible for a specific stromatolite form, the influx of volcanic detritus appears to have been a major influence in the Trooper Creek examples. For example, digitate stromatolites have more slender branches in parts of biostromes that are richer in volcanic detritus, and stubby shaped branches where there is less intercolumnar sediment. The unlinked branched portions reflect periods of rapid sedimentation when growth was limited to small patches which attempted
Figure 4.9

Dacitic volcano-sedimentary facies association (7741900 mN, 426800 mE). (A-C) Stromatolitic-oncolitic facies.

(A) Simplified sketch showing the distribution of microstructural elements forming a subspherical bioherm. Sample 95-218.

(B) Subspherical bioherm. The pedestal of the structure comprises a series of oncolites (o) from which extend radially arranged turbinate columnar stromatolites (c). Sample 95-218.

(C) The upper part of the bioherm is a complex mixture of turbinate columnar, pseudo-columnar and columnar-layered stromatolite forms. Sample 95-218.

(D) Polymictic lithic-pumice breccia facies. This breccia is a poorly sorted, matrix-supported and weakly normally graded. The clast population includes siltstone (s), ironstone (arrow), dacite (d) and composite clasts of dacite and indurated siltstone (i). (E-F) Laminated siltstone and vitric sandstone facies.

(E) Thinly bedded, normally graded, shard-crystal sandstone beds, some with thin cross-stratified tops. Flame structures (arrow) are locally developed.

(F) Gypsum molds on a parting within siltstone. Isolated euhedra and clusters of intergrown crystals have discoidal or lenticular morphologies similar to those characteristic of displacive gypsum. Sample 95-291.

(G) Trace fossils (*Planolites*) within siltstone (arrow). Sample not oriented. Sample 95-271.
to maintain a surface position. During periods of low sediment influx lateral linkages were
able to develop and in some cases non-columnar mat formed. The mechanism of
branching and factors controlling differential growth responses are poorly understood
(Grey and Thorne, 1985). Retardation of growth and restriction of growth to column
margins and sides due to excessive wetting, mechanical erosion, rapid sedimentation and
desiccation-induced cracking have been suggested as important (e.g. Haslett, 1976;
Logan et al., 1964). At Trooper Creek prospect, pumice clasts, crystal fragments and
vitric particles which washed on to the mat interrupted growth until the microbes re­
established themselves on the new substrate surface. There is increasing evidence to
suggest that the gross morphology and fabric of stromatolites is strongly dependent on the
kind of organisms forming the microbial community (e.g. Grey, 1984; Park, 1977).

Preservation of stromatolites and oncolites requires penecontemporaneous lithification.
Syn-depositional lithification of Proterozoic stromatolites mostly involves carbonates, but
silica (Knoll and Simonson, 1981) and iron oxides (Walter and Hoffman, 1983) have
been recorded. In examples of stromatolitic and oncolitic ironstone from Trooper Creek
prospect, former microbial laminae were preserved by hematite, while pumice and shards
were replaced by quartz and minor hematite. It is not known whether quartz and hematite
were the first minerals to precipitate or if they are replaced precursor minerals such as
carbonate or oxyhydroxide (e.g. Cloud and Semikhatov, 1969; Kah and Grotzinger,
1992; Hofmann and Jackson, 1987). There is no textural evidence of former carbonate
minerals in the stromatolitic samples. Penecontemporaneous lithification by silica has
been recorded in modern (e.g. Walter et al., 1976) and ancient (Knoll and Simonson,
stromatolites in the Early Proterozoic Sokoman Iron Formation are preserved as very fine­
grained hematite, whereas quartz replaces the remainder of the structure. Conversely,
Oehler (1976a) noted that quartz crystallites forming spherules commonly nucleated on
the surfaces of organic matter.

Polymictic lithic-pumice breccia facies

This breccia overlies the laminated pumice breccia facies and is poorly sorted, matrix-
supported and weakly normally graded. The breccia is 12 m thick and may comprise two
units. However, the upper, lower and internal contacts are unexposed. The size of lithic
clasts decreases from up to 2 metres at the base to 3 cm towards the top. The matrix is
intensely sericite-hematite-altered, but contains an even distribution of euhedral feldspar
crystals suggesting the presence of compacted pumice fragments (e.g. McPhie et al.,
1993). The clast population is diverse and includes laminated siltstone, quartz-hematite
ironstone and vesicular (20 %) aphyric dacite fragments (Fig. 4.9D). Other clasts
comprise vesicular dacite with indurated siltstone-filled fractures and rinds. These clasts
are similar to those which characterise the indurated siltstone-matrix breccia facies (Chapter 5.3.2). The sediment immediately surrounding these clasts is different from the matrix of the breccia. Dacite clasts have elongate to blocky shapes with planar and curviplanar to highly irregular margins. Siltstone clasts have elongate blocky shapes, are up to 2 m long and some contain laminae of feldspar crystal-vitric sandstone which are partially replaced by quartz and hematite.

**Origin and significance of facies**

Weak normal grading and large (1-2 m) matrix-supported clasts are consistent with deposition as a subaqueous mass flow (e.g. Smith, 1986; Lowe, 1982). Clasts in the breccia were either incorporated at source or collected from the substrate during transport. The underlying planar laminated pumice breccia facies suggests that the pumice formed by explosive magmatic eruptions but may not be juvenile. Dacite clasts with indurated siltstone rinds may record the mixing of magma and wet-sediment prior to, or during, their incorporation into the flow. The clasts may have been derived from a partially extrusive cryptodome in the source area of the breccia. One possibility is that the mass flow was initiated by partial extrusion of a cryptodome through a succession of pumice breccia, siltstone and ironstone beds.

**Laminated siltstone and vitric sandstone facies**

Planar, medium to thick (15-70 cm) beds of fine-grained sandstone interbedded with laminated siltstone characterise this facies. The siltstone and sandstone units are increasingly abundant towards the top of the section and are intercalated with thin intervals of microbialitic ironstone and pumice breccia. The principal components of the siltstone and sandstone beds are devitrified ash and shards, with subordinate crystal fragments (quartz and/or feldspar), suggesting that the beds are rhyolitic to dacitic in composition. Vitric components are now composed of fine-grained (5-10 µm) quartz and the grain boundaries are commonly not preserved. In many beds, devitrification and silicification have combined to destroy most primary textures, hampering their interpretation. Some non-volcanic material is likely to be intermixed with volcanic components and some siltstone intervals may be entirely non-volcanic.

The sandstone beds are massive or normally graded with crystal-rich bases and finer vitric-rich tops (Fig. 4.9E). Some beds display basal scours and flame structures. Cross-stratification is commonly developed within the fine sandstone to siltstone tops of beds. Siltstone units are massive or are planar laminated. Lamiae are typically massive but in a few samples (e.g. 95-270) are graded from shard-rich bases to very fine-grained relic ash-rich tops. Although mostly pale to dark grey, some siltstone beds are purple due to intense hematite alteration. One siltstone bed contains sericite pseudomorphs of gypsum.
crystals, best observed on weathered surfaces. The crystals have lenticular shapes and occur separately or in clusters (Fig. 4.9F). Although most crystals occur on bedding-parallel partings in the siltstone, some single crystals occur between the partings. The gypsum crystal-rich bands are 1-2 mm wide and separated by 3-4 cm of massive siltstone.

Evidence for bioturbation within the facies is limited to rare trace fossils within siltstone float from the same area (Fig. 4.9G). The trace fossils are temporary feeding structures (fodinichnia) and comprise mostly Planolites. The trace fossils belong to the Cruziana ichnofacies (Frey and Pemberton, 1984).

**Origin and significance of facies**

The crystal composition suggests that that some of the components in this facies may have been sourced from the same eruptions which deposited the underlying facies. Crystal fragments have unmodified angular shapes suggesting that they were not significantly reworked prior to deposition. Sandstone units display bedforms which suggest they are deposits from low-concentration turbidity currents (cf. Lowe, 1982). Crystals and large vitric particles settled from suspension forming a weakly graded division (Bouma Ta). Subsequent deposition of finer particles involved both traction (Bouma Tc ± Tb) and suspension (Bouma Te) generating the cross-laminated division and fine siltstone top. Shard-rich siltstone could have originated by settling from suspension in dilute currents trailing volcaniclastic mass flows, and/or water-settled fallout from subaqueous or subaerial explosive eruptions.

Gypsum is traditionally interpreted as evaporitic in origin, being deposited in brine pools at or above the sediment-water interface. However, gypsum can also grow displacively from saline pore fluids within permeable sediment. Displacive growth produces both isolated euhedra and clusters of intergrown crystals which commonly have “discoidal” or “lenticular morphologies” (Demicco and Hardie, 1994) similar to those at Trooper Creek prospect. Intra-sediment growth of gypsum crystals provides evidence for post-depositional crystallisation within an evaporitic environment.

**Graded dacitic pumice breccia facies**

This facies is dominated by feldspar-phyric pumice clasts and feldspar crystal fragments suggesting a dacitic composition. Units are non-welded, up to around 80 m thick, and are normally graded with massive or diffusely bedded tuffaceous sandstone tops, and in some instances, polymictic lithic-rich bases. A few beds have thin reverse-graded sandstone bases overlain by normally graded breccia to sandstone. Sandstone tops consist
mostly of relic pumice and shards. The base of one very thick (> 80 m) unit includes siltstone clasts up to 40 cm long. Lamination in siltstone underlying the breccia is contorted and invaded by tongues of the pumice breccia up to 0.5 m in length. Siltstone clasts are blocky or have wispy contorted shapes and sharp margins. Other clasts have irregular shapes and merge with the matrix, suggesting they were not fully lithified at the time of incorporation into the breccia and are intraclasts.

Thin-sections reveal relic uncompacted round- and tube-vesicle pumice clasts. Vesicles and former glassy vesicle walls are now optically continuous K-feldspar or have altered to albite. A mosaic of fine-grained (5 μm) albite, quartz, sericite and chlorite replaces K-feldspar throughout much of the samples (eg. 95-308), destroying or obscuring pumice textures. In some single pumice clasts, feldspar fills the vesicles and former glassy walls have altered to chlorite. Pumice clasts which have altered to chlorite and/or sericite are compacted and define a bedding-parallel foliation (S1) in outcrop. Most units contain varying proportions of albite-quartz-altered pumice and phyllosilicate-altered pumice. A few units consist entirely of sericite-chlorite-altered compacted pumice. These are characterised by an even distribution of crystals that resembles the porphyritic texture usually found in coherent lavas or intrusions. Margins of pumice clasts and former tube vesicles are defined by thin (2-5 μm) discontinuous trails of chlorite or mica.

Pumice breccia beds vary from a few metres up to more than 80 m in thickness and are interbedded with siltstone and fine-grained vitric sandstone.

**Origin and significance of facies**

The dominance of juvenile pyroclasts with angular, unmodified shapes and great thickness of some beds suggests that this facies is syn-eruptive and was sourced from explosive magmatic eruptions at a subaerial or shallow marine vent (Chapter 5). Clearly vesiculation was not inhibited by high confining pressures that would be imposed by a deep water column (e.g. McBirney, 1963; Fisher and Schmincke, 1984). Although rich in pumice that is most likely pyroclastic in origin, there is no textural evidence for hot emplacement preserved in this facies. Eutaxitic texture, columnar jointing, vapour phase minerals, gas-escape pipes, syn-depositional thermal oxidation, high temperature crystallisation textures (e.g. spherulites) and plastically deformed shards are absent (e.g. Niem, 1977). Some compacted pumice clasts in the breccia beds resemble fiamme in welded pyroclastic deposits. However, tube pumice clasts with randomly oriented uncompacted vesicles are also preserved and the bedding-parallel foliation is interpreted as a diagenetic compaction foliation (cf. Niem, 1977; Allen and Cas, 1990; Branney and Sparks, 1990). Furthermore, the lithofacies characteristics of the pumice breccia beds are consistent with deposition from water-supported subaqueous mass flows, most probably
high-concentration turbidity currents, some of which scoured the poorly consolidated sediment substrate.

4.3 Lateral and vertical lithofacies variations

The oldest parts of the succession are dominated by the sedimentary facies association. This association displays little internal heterogeneity. In contrast, the overlying andesitic and dacitic facies associations are characterised by rapid lateral and vertical changes in the proportion of resedimented volcaniclastic and sedimentary facies. Because there is no evidence for disruption of the section by post-depositional faults, the facies variations are interpreted as primary and syn-depositional. These facies associations incorporate variations controlled by shifts in relative proximity to source, as well as complexities caused by changes in water depth during emplacement, the rate of sediment supply, and volume of volcanic and non-volcanic material entering the basin.

The initial phase of andesitic volcanism is recorded by intervals of the graded andesitic scoria breccia facies and massive andesite facies. Graded scoria breccia beds dominate the western part of the section, with a clear shift to massive andesite and associated autoclastic breccia in the eastern half of the area (Fig. 4.1). The graded andesitic scoria breccia facies encloses a thin interval of coherent andesite and autoclastic breccia. The overlying cross-stratified breccia facies is thin in the west but thickens to greater than 30 m in the east (Fig. 4.4). Globular clast-rich breccia occurs only in the eastern part of the area and overlies the cross-stratified breccia facies (Fig. 4.4 - section E). At this position, a high angle surface juxtaposes lithofacies of the andesitic facies association (globular clast-rich facies) with those of the overlying dacitic facies association (Fig. 4.1). In the west, the surface is less irregular and marks the top of the graded andesitic scoria breccia facies, or the cross-stratified andesitic breccia and sandstone facies. The contact is draped by overlying dacitic water-settled fall beds. This relationship is inconsistent with post-depositional faulting but does not preclude the presence of syn-depositional growth faults. There are several possible interpretations for the surface: (1) topography generated through growth faulting; (2) erosion; or (3) primary constructional topography reflecting proximity to the vents within the andesitic facies association. The globular clast-rich facies is absent to the west and is interpreted as a near-vent deposit. This is consistent with the eastern sections being closer to source (and so thicker) than the central and western sections. Over 100 m of palaeorelief is suggested (Fig. 4.1). However, the whole surface is highly irregular which suggests that post depositional erosion and/or growth faults were probably also important.

In the central part of the area, the base of the overlying dacitic volcano-sedimentary facies association comprises massive to planar laminated pumice breccia (water-settled fall)
which passes up through a thin interval of microbialitic ironstone (lower microbialite) into polymictic lithic-pumice breccia (mass-flow deposit). The microbialite is around 10 cm thick and consists of thin (1-5 cm) units of tuffaceous sandstone, oncocolite-rich sandstone and in situ domed biostromes. Siltstone is increasingly abundant in the upper part of the succession and encloses thin intervals of microbialitic ironstone (upper microbialite), mud-matrix dacitic breccia and pumice breccia (Fig 4.1 — section B). The upper microbialite consists of stromatolite-rich breccia and subspherical bioherms. The relationship between the breccia and bioherms is not clear due to poor exposure. To the east, lithofacies at the base of the volcano-sedimentary facies association are poorly exposed along a steep contact with the globular clast-rich andesitic breccia facies. To the west, the contact between the andesitic facies association and dacitic volcano-sedimentary facies association is marked by a thin (10’s of centimetres to a few metres) discontinuous ironstone lens (Chapter 6). The overlying deposits are very poorly exposed and comprise intercalated siltstone, vitric-crystal sandstone and dacitic pumice breccia.

The top of the succession comprises siltstone, rounded lithic-crystal sandstone (Chapter 3) and a thick interval (> 80 m) of dacitic pumice breccia, the top of which is obscured by the Tertiary Campaspe Formation (Fig. 4.4 - section B). To the east, the equivalent dacitic pumice breccia is thinner (around 60 m) and overlain by thick intervals of siltstone intercalated with thin tuffaceous sandstone and breccia beds (Fig. 4.4 - section E). Thickening of the pumice breccia to the west is interpreted to reflect ponding against a palaeotopographic high to the east. The palaeotopographic high is marked by thickening of the underlying andesitic facies association. In the west, the dacitic pumice breccia is absent and a large body of coherent dacite occurs at the same stratigraphic position.

4.4 Evolution of the Trooper Creek centre

At Trooper Creek prospect, the Highway Member provides an excellent example of the impact of changing environment on eruption style and products. The depositional setting of the three principal lithofacies associations is interpreted to have been submarine on the basis of fossil evidence and regional context. The preserved succession records a transition from a relatively deep to shallow water (above fairweather wave base) depositional environment. Three evolutionary stages are identified (Fig. 4.10), each of which corresponds to one phase in the shoaling of an andesitic volcanic centre.
4.4.1 Deep water phase

Conditions in the depositional environment prior to the onset of andesitic volcanism are recorded by the sedimentary facies association. The lithofacies characteristics of this association are consistent with deposition below storm wave base. Furthermore, the thick accumulation of siltstone and fine sandstone (565 m) imply a relatively deep and/or quiet water setting.

The initial phase of andesitic magmatism is recorded by vesicular andesite in the eastern part of the area. Peperitic upper contacts suggest that some intervals of the andesite were syn-sedimentary sills emplaced into poorly consolidated siltstone and sandstone (Fig. 4.10A). In many other cases upper contacts are not exposed, so that interpretation of intrusive versus extrusive emplacement is not possible. The andesites were constructional and are interpreted to have played an important role in initial shoaling of the volcanic edifice. They formed a pedestal in the eastern part of the area upon which subsequent eruptive centres were constructed. Coeval environments in the western part of the area remained in relatively deep water.

4.4.2 Transitional phase

In the west, sedimentation was dominated by scoriaceous turbidity currents together with minor lavas, and ambient deposition of sands and silts (sedimentary facies association) was terminated. The graded andesitic scoria breccia facies also accumulated in water depths below storm wave base as indicated by several lines of evidence: (1) the units are interpreted as turbidites which are, in general, diagnostic of subaqueous below storm-wave-base depositional settings; (2) hyaloclastite which is associated with a single intercalated lava suggests a subaqueous setting; (3) tractional structures which are abundant in subaerial environments and fluvial and shoreline settings are absent. Isolated occurrences of planar and cross-stratification reflect traction and shearing on the depositional boundary layer of the sediment gravity flows.

The high vesicularity of fragments in the graded andesitic scoria breccia facies (30-50%) suggests that eruptions took place in a relatively shallow water or subaerial setting (probably to the east) where confining pressure allowed vesiculation to proceed uninhibited (e.g. McBirney, 1963; Fisher and Schmincke, 1984). Depths shallower than 500 m are implied (McBirney, 1963) and were probably less than 200 m or even subaerial, although no subaerial facies are preserved. The bombs within the graded andesitic scoria breccia facies and globular clast-rich andesitic breccia facies, may indicate a subaerial explosive origin for the pyroclasts. However, bombs can also form by
Figure 4.10 Cartoon of the principal stages in the development of the Trooper Creek prospect andesitic volcano. See following page for discussion.
Figure 4.10 continued. Cartoon of the principal stages in the development of the Trooper Creek prospect andesitic volcano. (A-E) and emplacement of the overlying dacitic volcano-sedimentary facies association (F-J). (A) Subaqueous (below storm wave base) lava effusion and syn-sedimentary intrusion of andesite into wet unconsolidated sediments of the sedimentary facies association; (B) Strombolian-style volcanism in a storm-wave-base (SWB) environment; (C) Post-eruptive erosion and resedimentation of pyroclasts and construction of a platform near fairweather wave base (FWB); (D) Eruption of magma by fire fountaining and hydrotelluric interactions, above fairweather wave base (FWB). Subsidence concurrent with volcanism (arrow); (E) Slumping and post-eruptive degradation of volcanic edifice; (F) Mass-flows and water-settled fall of pyroclasts sourced from distal dacitic eruptions; (G) Colonisation of platform by microbialites. Ironstone deposition from circulating, low-temperature hydrothermal fluids (arrows); (H) Partial extrusion of a cryptodome generating a polymictic debris flow; (I) Deposition of volcanogenic sediments and recolonisation of platform by microbialites. Minor evaporite minerals. Ironstone emplacement; (J) Subsidence and transition to siltstone-dominated sedimentation.
eruption in both deep water (e.g. Batiza et al., 1984) and shallow water (e.g. Yamagishi, 1982, 1987; Staudigel and Schmincke, 1984) and pyroclasts ejected during subaerial eruptions can fall into water after following ballistic trajectories (e.g. Kokelaar and Romagnoli, 1995). During subaqueous eruptions, bombs and agglutinate can form within an insulating vapour plume (Batiza et al., 1984; Kokelaar, 1986; Smith and Batiza, 1989). Hot fluidal clasts are isolated from the surrounding seawater by a steam envelope or cupola, so that hydromagmatic fragmentation is inhibited (Kokelaar and Durant, 1983a, Smith and Batiza, 1989). Magmatic volatiles may also contribute to the vapour plume (Kokelaar and Durant, 1983a; Kokelaar, 1986; Mueller and White, 1992).

Clasts in the graded andesitic scoria breccia facies are interpreted to record strombolian explosive eruptions which built ephemeral scoria cones subject to collapse and resedimentation into deeper water environments that existed in the western part of the area (Fig. 4.10B). The mean grainsize of scoria in the graded andesitic scoria breccia facies is coarse sand to granule size. This is typical of strombolian scoria deposits. Scoria fall deposits from the 1973 eruption of Eldfell on Heimaey, Iceland (Self et al., 1974) are representative of those of strombolian activity (e.g. Walker, 1973a; Walker and Croasdale, 1972). These deposits contain few pyroclasts finer than 1 mm, and consist of scoria mostly 2-8 mm across and bombs. They contrast with deposits of surtseyan eruptions by being coarser grained, better sorted and containing fragments with more fluidal shapes (Walker and Croasdale, 1972; Kokelaar and Durant, 1983a).

The degree to which the eruption column extended above seawater is uncertain. Vesiculation may have been enhanced by the eruption column reaching shallow water and so lower hydrostatic pressure. Lackschewitz et al. (1994) suggest that convection cells within subaqueous eruption columns are capable of lifting particles several centimetres across to shallower water depths and rapidly decreasing confining pressure. Decreasing confining pressures may promote vesiculation of incandescent fragments generated during initial low-energy fragmentation of the magma by, for example, fountaining driven by high effusion rates. Volatile expansion within vesicles already formed in deep water may cause secondary fragmentation of particles (Lackschewitz et al., 1994).

4.4.3 Shallow subaqueous phase

Following the first phase of magmatic explosivity (stage 1) there was a change in eruption style (stage 2) and depositional setting which is initially recorded by a thick interval of cross-stratified andesitic breccia and sandstone beds in the eastern part of the area. The cross-stratified andesitic breccia and sandstone facies records the transition to an above-storm-wave-base depositional environment. High-angle cross-bedding implies deposition
above fairweather wave base and suggests that the intercalated turbidity current deposits were also deposited above storm wave base (Fig. 4.10C). The depth of fairweather wave base varies but is typically in the 5 to 15 m range (Walker, 1984). The thickness (>70 m) of single mass-flow deposits in the upper part of the graded breccia and sandstone facies suggests very rapid aggradation to an above fairweather wave base depositional environment. The transition to above fairweather wave base probably included removal and redeposition of some deposits from the first phase, especially near vent pyroclastic deposits (Fig. 4.10C).

The globular clast-rich andesitic breccia facies overlies the cross-stratified interval and is also thought to have been deposited partially, if not entirely, above storm wave base for the following reasons: (1) the facies overlies above-storm-wave-base deposits; and (2) the overlying dacitic volcano-sedimentary facies association contains shallow marine fossils. However, an unconformity separates the andesitic facies association and dacitic facies association so the globular clast-rich andesitic breccia facies may have been partially emergent. If deposited subaqueously, the thickness of the globular clast-rich andesitic breccia facies (>70 m) requires that subsidence occurred prior to its emplacement and/or that subsidence kept pace with accumulation. If subsidence did not occur the deposit would have rapidly become emergent during emplacement in the shallow water depositional setting. Base surge deposits which might be expected on the emergent part of a marine cone (cf. Surtsey, Kokelaar, 1986) are absent. The subaerial facies may have been eroded prior to deposition of the dacitic volcano-sedimentary facies association.

There is an overall decrease in the vesicularity of clasts from scoriaceous in the graded andesitic scoria breccia facies to dominantly poorly vesicular in the overlying cross-stratified andesitic breccia facies and globular clast-rich andesitic breccia facies. The clasts within each of the three facies are mineralogically similar suggesting that they were probably sourced from the same eruptive centre although probably not from the same vent. The absence of rounded clasts, intervening silicic volcanic deposits and microbialites (cf. overlying succession) suggests that volcanic activity of stage 1 and 2 was relatively continuous. Vesiculation of a magma is dependent on external confining pressure, primary magmatic volatile content and cooling history (e.g. McBirney, 1963; Kokelaar, 1986). In deeper water, higher confining pressures might suppress volatile exsolution and restrict the increase in magma viscosity that usually accompanies cooling (e.g. McBirney, 1963). Eruption and deposition of the globular clast-rich andesitic breccia facies occurred in water depths shallower than 15 m (above fairweather wave base) where confining pressures would have been insufficient to suppress vesiculation. However, quenching of the magma prior to peak vesiculation may have been important in generating poorly vesicular clasts. Collectively, the evidence suggests a cooling and/or
primary magmatic volatile control on the vesicularity of clasts and an overall upward decrease in vesicularity within the andesitic facies association.

Pyroclasts in the globular clast-rich andesitic breccia facies and cross-stratified breccia facies appear to have been sourced from eruptions driven mostly by the high magma pressure, eruption through a restricted conduit and hydrovolcanic interactions (Fig. 4.10D). The expansion of exsolving magmatic volatiles does not appear to have been important as scoria is a minor component of this facies. The bombs show varying stages of disintegration and are accompanied by blocky clasts which implies brittle fragmentation, perhaps involving hydrovolcanic processes, and also suggests that water gained access to the column and/or that the pyroclasts fell into water. Reddish oxidation of a few clasts in the andesitic breccia facies suggests these may have suffered subaerial weathering and oxidation prior to final deposition in a marine setting. However, hematite alteration of single clasts and domains in the breccias is locally intense and oxidation is more likely to reflect post depositional hydrothermal alteration.

In the initial phases of eruption the influx of water into the eruption column was high resulting in effective fragmentation of pyroclasts by quenching and possibly phreatomagmatic explosivity, so that the proportion of bombs in the resulting deposit was low. As the eruption proceeded and the magma discharge rate increased, hydrovolcanic processes were suppressed and fluidly-shaped pyroclasts were favoured. As the eruption waned and hydrovolcanic processes became more important, the normally graded, bomb-poor top of the globular clast-rich andesitic breccia facies formed.

The absence of epiclasts and the monomictic character of the andesitic breccia suggests that the edifice was not emergent for more than short periods of time and/or that aggradation was rapid and there was insufficient time for reworking. However, explosive eruptions from shallow marine to temporarily emergent volcanoes may deposit poorly consolidated aggregates which are easily resedimented. In these cases, the resedimented pyroclasts may not be appreciably rounded and evidence for emergence might not be recorded in the resedimented equivalent. Lavas are important in preserving the emergent parts of a volcanic edifice (e.g. Surtsey, Kokelaar and Durant, 1983a,b), being less easily eroded.

Spatter, agglutinate and feeder dykes are typical of near-vent environments (e.g. Smith and Batiza, 1989; Houghton and Landis, 1989). The absence of these lithofacies supports the interpretation that the globular clast-rich andesitic breccia facies was generated by activity at a vent most probably located a few hundreds of metres to the east.
The stratified andesitic breccia and sandstone facies association at Trooper Creek prospect is interpreted as the submarine apron of a shallow-water volcanic centre. The size of the volcano is not well constrained due to poor exposure. The facies associations are inconsistent with the former existence of a large volcano, the erosion of which would have substantially influenced sedimentation in the Trooper Creek Formation (cf. Romagnoli et al., 1993; Kokelaar and Romagnoli, 1995).

4.4.4 Post-eruptive degradation

The original volcanic cone was probably easily eroded, especially the deposits emplaced subaqueously above storm wave base. Some observed historical shoaling volcanoes are ephemeral and show repeated cycles of construction and degradation. During the 1964-1965 eruption of Surtsey, two satellite vents, Syrlingur and Jolnir, went through cycles of construction above sea level followed by erosion and submergence during storms (Kokelaar and Durant, 1983a,b). They were planed off to the level of storm wave base. Subaerial lavas were important in preserving the island of Surtsey against major erosion from storms (Kokelaar and Durant, 1983a,b). At Trooper Creek prospect, an unconformity at the top of the andesitic facies association records erosion and/or collapse of part of the primary volcanic edifice, possibly by slumping, in the post-eruptive degradational stage (Fig. 4.10E). Collapse may have been initiated by gravitational instability, earthquakes and/or magma intrusion (Siebert, 1984).

Following degradation of the andesitic centre, sedimentation at Trooper Creek prospect was dominated by deposits sourced from a silicic, shallow subaqueous or subaerial volcanic terrain. Thick pumiceous mass-flow and water-settled fall deposits were emplaced onto the relatively stable shallow water platform and maintained an above-storm-wave-base depositional setting (Fig. 4.10F). Pyroclast-rich sediment gravity flow deposits that occur interbedded with marine and lacustrine sedimentary rocks can be sourced intrabasinally or extrabasinally. Such deposits can record: (1) collapse and resedimentation of pumiceous hyaloclastite from the margins of lava domes and flows (e.g. Kurokawa, 1991,1992); (2) explosive tuff cone-forming eruptions accompanying subaqueous dome emplacement (e.g. Horikoshi, 1969; Cas et al., 1990; Allen et al., 1996b); (3) syn-eruptive resedimentation of unstable accumulations of pyroclasts temporarily deposited during subaerial or shallow subaqueous explosive eruptions (e.g. McPhie et al., 1993; White and McPhie, 1996); (4) post-eruptive slumping and sliding of coastal, non-welded, pyroclastic deposits into deep water (e.g. Wright and Mutti, 1981); (5) explosive eruptions from partly or wholly submerged calderas (e.g. Howells et al., 1986; Busby-Spera, 1986) or vents (e.g. Fiske, 1963; Fiske and Matsuda, 1964); (6) passage of subaerially erupted hot, gas-supported, pyroclastic flows into subaqueous
A shoaling volcanic succession

Environments and transformation of flows into water-supported volcaniclastic mass flows (e.g. Stanton, 1960; Lock, 1972; Niem, 1977; Carey and Sigurdsson, 1980; Wright and Mutti, 1981; Cas and Wright, 1991; Jagodzinski and Cas, 1993); or (7) sedimentation of pyroclasts from the base of pyroclastic flows moving over water (e.g. Ui et al., 1983; Cas and Wright, 1991; Carey et al., 1996).

Parts of a few lavas and domes in the Trooper Creek Formation are pumiceous (e.g. REW 802, 170-242 m; Chapter 5) and could have supplied pumiceous clasts to volcaniclastic units. However, the small component of non-vesicular juvenile clasts, regional distribution (Chapter 3) and thickness of some intervals of the graded pumice breccia facies suggest that resedimentation from the pumiceous campace of nearby submarine lavas was not the major source of pumice. The mass flows were not sourced locally but from explosive eruptions at a volcanic centre which is located outside the study area.

A significant feature of the sequence was the colonisation of the platform by microbialites and a variety of other organisms (Fig. 4.10G). Only a few microbialite types are indicative of specific depositional settings (e.g. Grey and Thorne, 1985; Bauld et al., 1992). Conical columnar stromatolites occur only in deep to shallow subtidal environments, and have been identified as the sole or dominant stromatolite type in basin and slope facies, drowned platform sequences, and probable deep ramp facies. Microdigitate columnar stromatolites are almost exclusively restricted to intertidal or supratidal settings (Grey and Thorne, 1985; Bauld et al., 1992). The remaining stromatolite types have a less restricted distribution and occur in deeper subtidal to intertidal reef, open shelf, lagoonal and tidal flat settings (Bauld et al., 1992). The morphologies of stromatolites at Trooper Creek prospect fall within this group and so only provide evidence for deposition in a quiet water environment, above fairweather wave base. Wave activity was of sufficient intensity to generate oncolites (Logan et al., 1964) and disrupt the growing mats. Intercalated siltstone and pumiceous water-settled fall deposits suggest that quiet water conditions prevailed during the deposition of these intervals. The appearance of microbialites was coeval with deposition of a thin horizon of quartz-hematite rock (ironstone). The ironstone partially replaced intervals of the pumice breccia substrata to the microbialites (Fig. 4.10G). The ironstones are interpreted to have been deposited from short lived, low temperature, local hydrothermal systems (Chapter 6).

The polymictic lithic-pumice breccia facies contains clasts which were probably sourced from a dacitic cryptodome extruding through previously deposited pumiceous facies, ironstone and siltstone (Fig. 4.10H). Volcanic siltstone and subordinate fine-grained vitric sandstone become increasingly abundant towards the top of the sequence, reflecting
a return to relatively quiet water conditions following deposition of the upper microbialite horizon (Fig. 4.10I). Gypsum molds within siltstone immediately overlying the uppermost microbialite horizon are interpreted to have grown displacively within the sediment. Displacive gypsum has been reported in many marginal marine and marginal lacustrine sediments in modern evaporites (Demicco and Hardie, 1994). Siltstone units contain trace fossils which can be assigned to the *Cruziana* ichnofacies (Frey and Pemberton, 1984). The *Cruziana* ichnofacies is most characteristic of subtidal environments (Frey and Pemberton, 1984). The thickness of siltstone and abundance of pumiceous turbidity current deposits in the overlying section suggests a return to relatively deep water conditions, probably in response to subsidence by compaction and/or tectonism, in the subsequent history of the depositional environment (Fig. 4.10J).

4.5 Summary

The Highway Member (Trooper Creek Formation) in the southern part of the Trooper Creek prospect comprises a complex association of lavas, intrusions, volcaniclastic units and sedimentary facies. The oldest parts of the succession are dominated by the sedimentary facies association which records a relatively deep water depositional setting prior to the onset of andesitic magmatism. The andesitic facies association represents the submerged slopes of a small, submarine, strombolian-style volcano which temporarily emerged above sea level. Internal facies variations suggest that the eastern exposures are more proximal than central and western exposures. The four principal lithofacies record progressive stages in the shoaling of the volcanic centre including: (1) subaqueous (below storm wave base) lava effusion and syn-sedimentary intrusion into wet unconsolidated sediments of the sedimentary facies association; (2) strombolian-style volcanism in a near-storm-wave-base environment; (3) hydrovolcanic interactions, above storm wave base and possibly subaerially. The volcanism played a major role in physically modifying the depositional environment. Lavas, syn-sedimentary intrusions and mass flows of resedimented pyroclasts were important in rapid shoaling of the depositional environment.

The post-eruptive history began with partial erosion/collapse of the edifice creating a stable plateau for deposition of the overlying dacitic volcano-sedimentary facies association. Pumiceous mass-flow and water-settled fall deposits record influxes of pyroclasts into the depositional environment from a distal shallow subaqueous or subaerial environment. These units were important in maintaining a shallow water depositional setting during emplacement of the dacitic volcano-sedimentary facies association. When subsidence outpaced accumulation, the depositional setting returned to below storm wave base.
Chapter 5

A silicic syn-sedimentary intrusion-dominated host succession to massive sulfide mineralisation, Trooper Creek Formation, Highway-Reward
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A silicic syn-sedimentary intrusion-dominated host succession to massive sulfide mineralisation, Trooper Creek Formation, Highway-Reward

5.1 Introduction

Although significant advances have been made in understanding the chemistry of the processes involved in the formation of volcanic-hosted massive sulfide deposits, the nature of host-rock successions, their palaeovolcanological setting, and the interplay between volcanism and mineralisation have received much less attention. Modern submarine volcanoes hosting massive sulfide mineralisation are largely inaccessible for detailed textural and facies research (cf. Binns et al., 1992; Binns and Scott, 1993). Understanding the eruptive and sub-volcanic processes at submarine volcanic centres is therefore dependent on studies of ancient sequences that are now exposed on land, such as the Seventy Mile Range Group. The quality and abundance of drill core intersections available in the Highway-Reward area enabled precise volcanic facies definition and evaluation of the spatial and temporal relationships between volcanism and mineralisation.

The research presented here suggests that the Cu-Au-rich Highway-Reward massive sulfide deposit is hosted by a silicic syn-sedimentary intrusion-dominated submarine (below storm wave base) volcanic centre. This setting differs from that proposed in many early studies of silicic volcanic-hosted massive sulfide deposits (e.g. Ohmoto, 1978; Green et al., 1981; Ohmoto and Takahashi, 1983) but both ancient (Allen, 1992; Allen et al., 1996b) and Tertiary (e.g. Horikoshi, 1969) equivalents have been recognised.

The Highway-Reward volcanic centre is compared with other effusive silicic to intermediate volcanic centres in the Highway Member, with the aims of characterising the style of eruptive and intrusive activity, establishing the emplacement mechanisms of the diverse lithofacies, and tracing the spatial and temporal evolution. In particular, the evidence for sub-seafloor emplacement of magma and the role of the intrusions in influencing the character of the host succession are considered. Chapter 7 examines the inter-relationships between volcanism and mineralisation at Highway-Reward.
5.2 The enclosing facies

In the Trooper Creek Formation, rhyolitic to dacitic lavas, intrusions, and associated in situ and reworked auto-clastic facies are intercalated with and over lain by volcaniclastic and sedimentary facies associations comprising siltstone, sandstone turbidites, and non-welded rhyolitic to dacitic pumiceous and crystal-rich breccia and sandstone. The host succession to the intermediate, lava- and intrusion-dominated volcanic centres also includes graded beds of andesitic scoria breccia and globular clast-rich breccia facies (Chapter 4).

5.2.1 Siltstone facies

The siltstone facies is generally host to the other main facies associations in the Trooper Creek Formation and therefore defines the ambient depositional conditions. In some sections, siltstone beds contain graptolite fossils (Henderson, 1983, 1986). The siltstone units are massive, thickly bedded or finely planar laminated (Fig. 5.1A). Although mostly pale to dark grey, some are green, yellow brown or purple, reflecting differing alteration mineral assemblages and possibly also variations in the original composition and/or oxidation state during deposition.

Pale to dark grey siltstone sometimes contains quartz and/or feldspar crystal fragments and devitrified shards (50 μm), suggesting a rhyolitic to dacitic composition. Vitric components are now composed of fine-grained (5-10 μm) quartz and the grain margins are rarely preserved. Quartz and feldspar crystal fragments are mostly angular but a few grains are subrounded. In many beds, devitrification and alteration have destroyed all primary texture, hampering their interpretation. These beds now comprise a mosaic of quartz and feldspar (5-10 μm). Pale green siltstone has altered to fine (<10 μm) sericite, chlorite and quartz. Some beds contain discontinuous, quartz and feldspar crystal-rich laminae (50-200 μm). The crystals are angular to subangular. In a few samples, rare white mica (20-50 μm) and subrounded polycrystalline quartz fragments (40 μm) are also present. Yellow-brown siltstone comprises quartz and/or feldspar crystal fragments (<60 μm), and in some samples, white mica (100 μm), opaque grains, euhedral apatite and rutile. (Fig. 5.1B). Crystals (20-100 μm) are mostly angular but some are subrounded. Grains in the siltstone are separated by fine-grained (5 μm) sericite, chlorite, epidote and quartz. White mica is either aligned parallel to lamination or the cleavages (S3 and S4). Siltstone which is purple contains similar components to grey siltstone, as well as fine-grained disseminated hematite. At Handcuff prospect, peperite at the top of an intrusion provides evidence for mixing of dacite with hematite-rich siltstone. Hematite is restricted
Figure 5.1

(A) Planar laminated siltstone. Light grey (white weathering) silicified siltstone. Highway East prospect, 7748450N, 418500E.

(B) Photomicrograph of a siltstone containing feldspar crystal fragments (arrow) and minor non-volcanic detritus. Plane polarised light. 94-145, Highway East prospect, 7748450N, 418500E.

(C) Thinly laminated siltstone enclosing dark green, evenly porphyritic, blocky and ragged, chloritic (formerly glassy) clasts (c). The clasts truncate or are mantled by laminations in the siltstone. Although vesicular microstructures are rarely preserved in the clasts, they are interpreted as relic pumice clasts which settled from suspension. Drill core is 5 cm wide.REW 805 282-290 m, Highway-Reward.

(D) Graded lithic-crystal-pumice breccia and sandstone facies. The prominent dark green lenses (I) in this outcrop are relic fiamme and are aligned along a spaced stylolitic- and compaction-foliation (S1) which is crenulated by a steep regional cleavage (S3). Trooper Creek prospect, 7742800N, 427600E.

(E) Graded lithic-crystal-pumice breccia and sandstone facies. This photomicrograph shows an altered tube pumice clast in non-welded pumice breccia. The pumice clast contains large round vesicles (v) adjacent to a euhedral feldspar phenocryst (f). Away from the phenocryst there is a transition to tube vesicles. Pumice preservation is due to replacement of former glassy vesicle walls and infilling of vesicles by albite prior to compaction. Vesicle walls are outlined by sericite (arrow). Plane polarised light. REW 805 (192 m), Highway-Reward.

(F) Graded lithic-crystal-pumice breccia and sandstone facies. This photomicrograph shows the character of the sandstone in figure 5.2C. Dominant components are angular to subangular feldspar (dominantly plagioclase) and quartz crystals and crystal fragments, pumice and quartzofeldspathic matrix (presumably altered formerly glassy particles). Carbonate patches in the sample contain ovoid structures (arrow) which are interpreted as vesicles in former pumice fragments. REW 801 (192.7 m), Highway-Reward.

(G) Quartz- and feldspar-phyric rhyolite. The groundmass contains spherical spherulites (arrow) which have variably recrystallised to interlocking quartz and feldspar. Areas between the spherulites were presumably originally glassy but have altered to hematite. REM 600, 245.5 m, Highway-Reward. Plane polarised light.

(H) Macro-perlite in coherent dacite. The fractures (arrow) are accentuated by weathering out of sericite. Areas between fractures have moderate to strong phyllosilicate-quartz alteration. Trooper Creek area; 7744300N, 424500E.
to the siltstone, suggesting that it is not an alteration product, but deposited from suspension as the siltstone accumulated.

Siltstone beds overlying normally graded pumiceous mass-flow units locally contain outsized (3-10 cm) chloritic, evenly quartz- and/or feldspar-phyric clasts with wispy shapes. The clasts have similar phenocryst assemblages to the associated pumiceous units and either truncate, or are mantled by, laminae in the enclosing siltstone. Some laminae are highly contorted (Fig. 5.1C). Although vesicular microstructures are rarely preserved, clast shapes and lithofacies characteristics are consistent with interpretation of the lenses as compacted pumice clasts.

**Origin and significance of facies**

Least recrystallised siltstone horizons contain volcanic fragments (principally glass shards and crystal fragments) which are often mixed with minor non-volcanic material (polycrystalline quartz, some white mica). Most volcanic detritus is relatively unaltered and angular suggesting the grains have not been significantly reworked. However, some crystal fragments are rounded and have undergone higher degrees of reworking prior to final deposition, suggesting separate sources. The remaining fragment population has been variably reworked. The lithofacies characteristics suggest an origin through the settling of fines entrained in the dilute currents trailing sediment gravity flows, pelagic or hemi-pelagic sedimentation and water-settled fallout. Pumice clasts enclosed by siltstone are interpreted to have settled from suspension (e.g. Reynolds and Best, 1976; Reynolds et al., 1980; Whitham and Sparks, 1986).

5.2.2 Graded lithic-crystal-pumice breccia and sandstone facies

Lithic-crystal-pumice breccia units in drill core and outcrop throughout the Trooper Creek Formation are texturally similar (Chapter 4; Fig. 5.1D). The breccia and sandstone beds are dominated by quartz- and feldspar-phyric pumice and crystal fragments or by feldspar-phyric pumice and feldspar crystal fragments (Fig. 5.1E). The crystal and phenocryst populations suggest rhyolitic, rhyodacitic and dacitic compositions. Units are non-welded, up to 80 m thick (e.g.REW 814, 236-305 m; Chapter 3), and are normally graded with massive or diffusely bedded tuffaceous sandstone tops, and in some instances, polymictic lithic-rich bases (Fig. 5.2). A few beds have thin (< 1m) reverse-graded sandstone bases overlain by normally graded breccia and sandstone. Sandstone tops consist mostly of relic pumice and shards or are crystal-rich. Up to half of the thickness of a few very thick (> 80 m) beds comprises crystal-rich breccia to sandstone (e.g. REW 805, 188-276 m; Fig. 5.2B). Single crystal-rich breccia to sandstone beds (1-50 m) are massive, ungraded or weakly normally-graded. Medium to thick sandstone
beds have fine-grained tuffaceous sandstone tops which are vitric-rich and contain smaller and less abundant crystals than the rest of the bed. Thicker beds are sometimes overlain by a series of normally graded beds, 40-70 cm thick (Fig. 5.2C). The thin beds contain 5% crystals and crystal fragments and were presumably vitric-rich, but are now composed of sericite, quartz and feldspar.

The crystal-rich breccia to sandstone consists of 15-25% coarse sand to granule, whole and broken volcanic quartz and/or feldspar (dominantly plagioclase) crystal fragments. The nature of the matrix cannot be completely resolved due to alteration that comprises various assemblages of quartz, feldspar, chlorite, sericite, epidote and carbonate. At least some of the matrix was composed of pumice, shards and vitric ash but their proportions cannot be resolved. Carbonate patches (1.2 mm) at the margins of feldspar crystals contain circular structures (180 μm) which are interpreted as relic vesicles in pumice (Fig. 5.1F). Vesicles are outlined by epidote and filled with carbonate which is optically continuous with that replacing vesicle walls.

Sparse lithic clasts (0.5-3 cm) occur throughout many pumiceous and crystal-rich beds, but the size, abundance (generally < 1%) and sometimes the diversity of lithic clasts increase towards the base. Lithic clast populations differ among units and can include rhyolite, dacite, andesite and/or siltstone. Some apparently lithic clasts may in fact be juvenile as they have phenocryst assemblages similar to those in pumice clasts. Rhyolite and dacite clasts typically have irregular or blocky shapes and can be perlitic or flow banded.

**Origin and significance of facies**
This facies association is the submarine record of a distal volcanic terrain that is not exposed, not preserved, or located outside the study area (Chapter 4). Pumice, shards, and crystals within the facies are interpreted to be pyroclasts and reflect the importance of explosive magmatic and/or phreatomagmatic eruptions and suggest that the source vents were in shallow water, or else in basin margin or subaerial settings (Chapter 4). The abundance of quartz and/or feldspar in the facies suggests that the source magmas were dominantly rhyolitic to dacitic in composition. In the Highway-Reward area, intervals of lithic-crystal-pumice breccia and sandstone are thicker than in some sections to the east (e.g. Trooper Creek prospect). The apparent spatial association of the lavas, intrusions and pumiceous mass-flow deposits is similar to that documented by Horikoshi (1969) in the Kosaka Formation, Japan. Horikoshi (1969) interpreted the pyroclastic debris to be derived from intrabasinal tuff-cone forming eruptions preceding effusive eruptions. At Highway-Reward, the mass-flow deposits are widespread and not restricted to the areas dominated by lavas and intrusions. The pumiceous mass flows may have infilled seafloor
Figure 5.2 Representative graphic lithological logs of pumiceous and crystal-rich mass-flow deposits in the Trooper Creek Formation at Highway-Reward. (a) Normally graded pumice breccia with a fine sandstone top containing outsized pumice clasts. REW 814; (b) Two main divisions are evident in this drill core: (1) a thick interval graded from lithic-rich pumice breccia at the base to pumice breccia towards the top; (2) crystal-vitrific sandstone and siltstone. The contact between the divisions is sharp but gradational and the deposit is normally graded. The subtle change in grain size and components probably indicates an amalgamated sub-unit boundary. REW 805; (c) The graphic log shows a >28 m thick weakly normally graded, crystal-rich sandstone overlying laminated siltstone. The finer top of the interval comprises poorly defined, normally graded sandstone beds, 0.3-1.2 m thick. REW 801.
topography generated by lavas, syn-sedimentary intrusions and syn-volcanic faults, but there is no positive evidence that they were not sourced locally.

The lithofacies characteristics of the graded lithic-crystal-pumice breccia and sandstone units are consistent with deposition from syn-eruptive, cold (or cool), water-supported, high-concentration sediment gravity flows (Chapter 4). In some crystal-rich breccia and sandstone units, relic shards and ash are concentrated into the finer tops or within thin beds overlying thicker massive or graded intervals (e.g. REW 801, 177-205 m; Fig. 5.2C). This suggests that separation of ash from the crystals was at least in part due to sorting during subaqueous mass flow transport. However, it is likely that other processes also contributed to the observed crystal concentration in these units. For syn-eruptive subaqueous facies, there are a number of possible interpretations and the final deposit can reflect a multi-stage concentration process (e.g. Cas, 1983). Crystal concentration can begin during eruption as glassy ash is lost to the atmosphere or, if erupted subaqueously, air and/or water. Elutriation of glass shards from pyroclastic flows or cold-water supported mass flows may generate vitric enriched ash clouds (e.g. Walker, 1972; Sparks and Walker, 1977). Crystal enrichment can be further enhanced by loss of vitric fragments to ash plumes generated by secondary explosions where subaerial pyroclastic flows interact with water (e.g. Walker, 1979; Cas, 1983; Jagodzinski and Cas, 1993; White et al., 1993) and within ensuing subaqueous mass flows. Parts of pyroclastic flows that are less dense than water or that transgress shorelines with a low angle of incidence may continue to flow over water (e.g. Cas and Wright, 1991; Ui et al., 1983). Preferential sedimentation of denser crystals from the base of the subaerial flow may contribute to the crystal enrichment (Carey et al., 1996).

5.2.3 Stratified andesitic breccia and sandstone facies association

This association comprises graded andesitic scoria breccia and sandstone and globular clast-rich breccia facies. These facies have a spatial and compositional association with coherent andesite (Chapter 4). West of Trooper Creek, exposures of the globular clast-rich breccia facies are limited to one creek section at Highway East prospect (around 7747300 mN, 420000 mE). The breccia is at least 40 m thick and appears massive, but because outcrop is poor the boundaries between units may not be exposed. The breccia is underlain by sandstone turbidites. Coherent andesite and graded andesitic scoria breccia occur above these units but contacts are not exposed. Intense pervasive hematite-chlorite-feldspar (albite) alteration of the breccia obscures primary textures, clast margins and the matrix. The breccia is composed of around 20% blocks, 5 cm to 47 cm across (cf. Chapter 4). Globular clasts (bombs, Chapter 4) are supported by a matrix composed of hackly to blocky scoria clasts (50 to 900 μm across) with 15-25% vesicles (visual
estirate). The vesicularity of these clasts is higher than those forming the matrix in globular clast-rich breccia facies exposed at Trooper Creek prospect (Chapter 4). However, the phenocryst assemblage and groundmass textures of clasts from Trooper Creek prospect and Highway East prospect are similar.

An intrabasinal source is considered likely in view of the close association with compositionally similar coherent andesite, the localised distribution, and the proximal character of the block-rich facies (Chapter 4). The dominant clasts are interpreted to be juvenile pyroclasts generated by strombolian eruptions and deposited in ephemeral scoria cones subject to collapse and resedimentation (Chapter 4). The strongly scoriaceous character suggests pyroclasts were probably generated by eruptions in water shallower than about 500 m. At Highway East prospect, the enclosing lithofacies and theoretical constraints (McBirney, 1963) imply that the vent for the globular clast-rich facies (and probably the depositional environment) was below storm wave and in water probably less than 500 m deep. At Trooper Creek prospect, similar bomb-rich breccia units were deposited above storm wave base (Chapter 4). The graded scoria breccia facies may have been sourced locally or from other intrabasinal andesitic volcanic centres (e.g. Trooper Creek prospect, Chapter 4). Intervals of the graded scoria breccia facies are laterally discontinuous, suggesting that the mass-flows were sourced from a number of small eruptive centres which are not exposed or have been eroded, and/or that distribution was controlled by palaeo-seafloor topography. Because of poor exposure, the importance of each alternative is difficult to determine.

5.2.4 Depositional setting for the volcanism

Siltstone within the host succession contains graptolite fossils which suggest a relatively deep, submarine depositional setting (Henderson, 1986). Most of the volcaniclastic units that are associated with the lavas and intrusions are interpreted as resedimented submarine sediment gravity flow deposits (including turbidites) and also indicate below-storm-wave-base settings, especially in the absence of tractional sedimentary structures. In modern oceans, depth of storm wave base ranges between 10 m and 200 m (Johnson and Baldwin, 1996). The strongly scoriaceous character of the andesitic volcaniclastic facies association implies that clasts were generated by eruptions in water probably less than 500 m deep (e.g. McBirney, 1963). Interpretation of the globular-clast-rich breccia facies as intrabasinal and proximal (Chapter 4) also implies that the depositional environment was, at least locally (Highway East), less than about 500 m deep. Combined, the lithofacies associations, fossils and regional context imply that the volcanism was submarine, below storm wave base, and possibly not in extremely deep water, but rather, around 500 m.
5.3 Lithofacies associations

Lava- and intrusion-dominated volcanic centres in the Trooper Creek Formation comprise coherent lithofacies and texturally diverse volcanioclastic lithofacies. The phenocryst assemblage suggests that coherent lithofacies and clasts in the volcanioclastic lithofacies are largely rhyolitic to andesitic in composition but rare basaltic andesitic examples are present.

The lithofacies can be assigned to one of two principal lithofacies associations; the primary volcanic facies association and resedimented volcanioclastic facies association. Lithofacies of the primary volcanic facies association include coherent rhyolite to basaltic andesite and associated non-stratified monomictic breccia facies (autoclastic breccia and peperite). The resedimented volcanioclastic facies association contains clasts which initially formed and deposited by volcanic processes, but which were re-entrained into another transport system and redeposited (e.g. McPhie et al., 1993). Resedimentation was more or less syn-eruptive so that the deposits are essentially monomictic and clast shapes are unmodified. This facies includes monomictic rhyolitic to andesitic breccia (resedimented autoclastic breccia) and stratified siltstone-matrix rhyolitic to dacitic breccia. Within the primary volcanic facies association there are complex transitions between coherent facies, autoclastic breccia and peperite (Fig. 5.3). There are equally complex transitions between constituent facies of the resedimented volcanioclastic facies association.

5.3.1 Primary volcanic facies association

Massive and flow-banded rhyolite, rhyodacite and dacite

This facies is characterised by evenly distributed quartz and/or feldspar phenocrysts in a fine-grained groundmass. The mineralogy, abundance and distribution of phenocrysts are, in most cases, uniform within a single unit (Fig. 5.1G). These properties provide a means of mapping different units in the field and are a rough indication of their chemical composition. Dacites are aphyric or contain 3-15% feldspar phenocrysts. Rhyodacites contain 7% feldspar phenocrysts, 1-2 mm across and up to 3% quartz phenocrysts. Rhyolites have 5-25% quartz phenocrysts, 0.5 to 7 mm across and 3-15% feldspar phenocrysts ranging from 0.4 to 4 mm across. Apatite and zircon are common accessory phases in the rhyolites to dacites. Geochemically, some rocks mapped as rhyolites and rhyodacites plot as high-silica dacites (Chapter 3).

Feldspar phenocrysts are euhedral and although dominantly plagioclase sometimes include sanidine. Feldspar phenocrysts are unaltered or variably altered to sericite,
Figure 5.3 Graphic lithological logs for parts of diamond drill holes in the Trooper Creek Formation at Highway-Reward (REM 122, REM 125), Handcuff (HDD 008, HDD 012, HDD 022), Vicesquad (MVO 32), Rustler (RD 812) and Stock squad (RD 813). Important textures, structures and facies relationships are shown.
chlorite, carbonate, albite or polycrystalline quartz. Quartz phenocrysts have a partial bipyramidal habit or are embayed and rounded as a result of magmatic resorption. No fragments of quartz or feldspar are present in the groundmass between the phenocrysts. However, many phenocrysts are fractured and, in strongly cleaved rock, some quartz phenocrysts have broken in situ. All fragments derived from the one crystal fit together (jigsaw-fit) and few fragments have rotated or separated.

The groundmass has devitrified to a quartzofeldspathic mosaic or else altered to various assemblages of chlorite, sericite, carbonate, epidote, quartz and albite. In some samples, the groundmass includes relic spherical spherulites, microspherulites, lithophysae, and/or micropoikilitic quartz that encloses sericitised feldspar microlites. Partial recrystallisation of spherulites to interlocking anhedral quartz and feldspar has destroyed some microstructures. The areas between coalescing spherulites were presumably originally glassy but have altered to phyllosilicate, carbonate or very finely crystalline hematite (e.g. 95-147; HMO 41, 173.5 m; REM 600, 245.5 m). Relic classical- and banded-perlite suggest that parts of the groundmass in many units were formerly glassy. Perlitic fractures are delineated by sericite or chlorite. Near Trooper Creek prospect (around 7744300 mN, 424500 mE), feldspar-phyric dacite is dissected by macroperlitic fractures (cf. Yamagishi and Goto, 1992). The fractures are recessive, outlined by sericite, and separate spherical to sub-spherical kernels of non-fractured silicified dacite. On weathered surfaces the kernels resemble clasts in monomictic, clast-supported conglomerate (Fig. 5.1H). Contacts with non-fractured dacite are sharp and irregular.

Regular columnar joints characterise some thick intervals of massive coherent dacite (Fig. 5.4A). The pattern of jointing in parts of other units is characteristic of pseudo-pillows (Yamagishi, 1987, 1991). Each pseudo-pillow is outlined by relatively continuous, smoothly curved joints and is internally dissected by polyhedral joints (Fig. 5.4B).

Flow banding occurs in some units, particularly along flow or intrusion margins. Flow foliations are defined by alternating siliceous and phyllosilicate-rich bands. The siliceous bands are composed of a quartzofeldspathic mosaic with occasional relic spherulites. Phyllosilicate-rich bands are probably an alteration of former glass. Parts of some units are composed of relic unflattened tube- and round-vesicle pumice. The vesicles are outlined by chlorite and an unidentified opaque mineral. A mosaic of interlocking quartz and feldspar has infilled vesicles and replaced formerly glassy vesicle walls. In some drill cores (e.g. REW 802, 126-342 m), there are complex transitions between intervals of pumiceous, spherulitic and perlitic, massive and flow banded rhyolite or dacite.

This facies shows complex gradational relationships with monomictic rhyolitic to dacitic breccia facies and non-stratified sediment-matrix breccia facies, particularly along upper
and lower contacts (Fig. 5.3). Intervals of the massive or flow banded rhyolite or dacite can range up to several tens to hundreds of metres in thickness, or can be less than 1 m in cases where associated monomictic breccia facies dominate the outcrop or drill core.

**Monomictic rhyolitic, rhyodacitic or dacitic breccia facies**

Intervals of this facies are massive, non-stratified, clast-supported frameworks of dacite, rhyodacite or rhyolite clasts. Clasts are evenly porphyritic (quartz and/or feldspar), non-vesicular and have blocky to elongate shapes bound by planar to curviplanar or irregular finely jagged margins. The groundmass within the fragments can be perlitic, spherulitic, devitrified to an interlocking mosaic of quartz and feldspar, or altered to various assemblages of chlorite, sericite, albite, and quartz. Polyphase alteration sometimes results in an apparent polymictic clast assemblage. Perlitic suggests that many of the clasts were originally glassy. Clasts are separated by small amounts of millimetre- to sub-millimetre-sized matrix composed of juvenile fragments, crystals and crystal fragments. In some cases, preferential quartz, sericite and/or chlorite alteration of the matrix and margins of clasts has generated an apparent matrix-supported fabric.

Flow banded clasts are common in some intervals and highlight two styles of breccia. In the first, continuity of flow banding between fragments suggests that the clasts are in situ and moved little following fragmentation. Clasts display jigsaw-fit fabric and crystals and juvenile fragments at clast margins show arrested stages of in situ disintegration. Although in situ breccia is dominated by clasts 2-4 cm across, some intervals include clasts 20 cm across. In the second style of breccia, significant rotation and translation of clasts is implied by flow banding at different orientations. There is a variation from slight modification of the jigsaw-fit texture (clast-rotated breccia; Fig. 5.4C) to mixtures of clasts with different groundmass textures and fabrics (disrupted breccia; Fig. 5.4D). The clast-rotated and disrupted breccia facies are non-stratified and not graded (Fig. 5.3B-C). At Trooper Creek prospect and Vicesquad prospect, there are two textural types of disrupted breccia: one is matrix-poor and although dominated by clasts 1-3 cm across, contains clasts up to 7 cm across. The second is coarser (clasts 2-30 cm across) and is generally matrix-poor. In Vicesquad drill hole MVO 32, intervals of the two textural types are 7 to 30 m thick and alternate down hole (Fig. 5.3B). Contacts are sharp but gradational and the proportion of fine-grained breccia appears to increase down hole.

Contacts between the coherent facies and monomictic breccia facies vary from sharp to gradational. In some cases, coherent rhyolite to dacite passes out through a zone of in situ and clast-rotated breccia into disrupted breccia facies. The monomictic rhyolitic to dacitic breccia facies sometimes encloses massive or flow banded facies with distinctly different phenocryst assemblages. These units have sharp contacts. One interval of the monomictic
Figure 5.4

(A) Columnar joints in the interior of a feldspar-phyric dacite. Highway East; 7747800N, 419500E.

(B) Pseudo-pillows in feldspar-phyric dacite. Each pseudo-pillow is outlined by relatively continuous, smoothly curved joints (arrows) and is internally polyhedrally jointed.

(C) Monomictic dacitic breccia facies. Gradation from coherent, flow-banded dacite into matrix-poor jigsaw-fit hyaloclastite (i) that in turn grades into clast-rotated hyaloclastite (c). Clasts in the breccia, flow bands and the matrix have altered to various assemblages of sericite, chlorite and quartz. Trooper Creek prospect; 7743700N 427000E.

(D) Disrupted feldspar-phyric dacitic hyaloclastite. The clasts have been variably silicified, chloritised and sericite-altered. Clasts in the breccia have different groundmass textures and are variably oriented. The breccia is clast-supported and although non-stratified and massive, the textural variety among clasts suggests fragments moved following brecciation. Trooper Creek prospect; 7743400N 427100E.

(E) Non-stratified sediment-matrix breccia facies. This example of blocky, matrix-rich andesitic peperite occurs along the upper margin of a syn-sedimentary sill. Andesite clasts have curviplanar margins and locally display jigsaw-fit texture (arrow). The siltstone matrix (s) is strongly silicified.

(F) Non-stratified sediment-matrix breccia facies. Peperite with ragged quartz- and feldspar-phyric rhyolite clasts (r) and siltstone matrix (s). Clasts in the breccia display jigsaw-fit (arrow). Highway East; 7747250 mN, 418000 mE.

(G) Non-stratified sediment-matrix breccia facies. Dacite clasts (d) in this peperite have ragged and globular shapes and are separated by hematitic siltstone (s). Laminae in the siltstone (arrow) are contorted and absent in the peperite. Handcuff prospect; 7747400 mN, 417250 mE.

(H) Non-stratified andesitic breccia facies. This outcrop shows a gradation from relatively coherent andesite (a) into in situ hyaloclastite. Many of the clasts have curviplanar margins and groups of clasts locally show jigsaw-fit texture (arrow). The matrix and some small clasts have altered to quartz (pale). Highway South; 7745600 mN, 420000 mE.
breccia facies can be traced for 1.3 km along strike (Trooper Creek prospect; Map 1), but most are exposed for less than a few metres. Thicknesses of the monomictic rhyolitic to dacitic breccia facies vary from less than a metre to at least 100 m. Rhyolite, rhyodacite or dacite can be dominated by the monomictic breccia facies, but in most cases the coherent facies is more abundant. At Trooper Creek prospect, massive and flow banded dacite occur as pods and lobes, 1 to 30 m wide and 10 to 150 m long enclosed by monomictic breccia facies. Contacts between the lobes and the monomictic breccia facies are gradational over less than a metre.

**Origin and significance of the massive facies and monomictic breccia facies**

The textural characteristics and contact relationships of the rhyolites, rhyodacites, and dacites favour their interpretation as coherent igneous rocks. An origin as densely welded ignimbrite can be discounted by evidence including: (1) the absence of crystal fragments and significant lithic clasts (broken phenocrysts in strongly cleaved samples can be attributed to brittle fracturing during deformation); (2) preservation of uncompacted vesicles in the interior of units; (3) lack of vertical or lateral variations in grainsize or welding; and (4) peperite along the top and side contacts which is characteristic of sills and cryptodomes and inconsistent with ignimbrite; peperite can occur at the base of ignimbrites deposited on unconsolidated wet sediment (e.g. Francis and Howells, 1973) but has not been recorded along top contacts.

Gradational contacts between intervals of the monomictic rhyolitic to dacitic breccia facies and coherent facies suggests that the facies are genetically related. The shapes and textures of clasts, and local jigsaw-fit is consistent with quench fragmentation, as documented in hyaloclastite from similar volcanic successions (e.g. Pichler, 1965; Dimroth et al., 1978; De Rosen-Spence et al., 1980; Furnes et al., 1980; Fridleifsson et al., 1982; Yamagishi and Dimroth, 1985; Yamagishi, 1987, 1991; Bergh and Sigvaldason, 1991; Kano et al., 1991). The thickness and lateral extent of the facies are inconsistent with brecciation by hydraulic fracturing.

Clasts in the disrupted breccia facies have undergone rotation following fragmentation, implying that autobrecciation accompanied quenching. Clast rotation and additional fragmentation occurred in response to stresses imposed on the chilled parts of the lavas and intrusions by continued movement of the more ductile interior. In the disrupted breccia facies, mixing of texturally variable clast types suggests that some redistribution of clasts occurred following fragmentation. It is not clear if this was solely due to inflation of the breccia pile or if clasts tumbled, bounced, and rolled downslope under the influence of gravity. Transport distances were limited as clasts are bound by original fracture surfaces, clast margins are unaffected by rounding, and because the talus is non-stratified
and not graded. The disrupted breccia facies might be considered analogous to aprons of talus breccia that accumulate at the margins of subaerial lava domes (e.g. Novarupta lava dome, Alaska).

Yamagishi (1979, 1991) identified two textural types of hyaloclastite in submarine sequences of southwest Hokkaido, Japan; hyaloclastite (A) and (B). The textural differences between hyaloclastite (A) and hyaloclastite (B) were considered by Yamagishi (1979, 1991) to reflect differing magma viscosities at the time of fragmentation. Hyaloclastite (A) is associated with magma of relatively low viscosity (basalt, basaltic andesite) and hyaloclastite (B) is associated with relatively viscous magmas (silicic and some intermediate compositions). Rhyolitic to dacitic hyaloclastite from the Trooper Creek Formation is most similar to Yamagishi's hyaloclastite (B).

Hyaloclastite in the Trooper Creek Formation indicates emplacement of lava into subaqueous settings and/or intrusion of magma into wet unconsolidated sediment. However, the presence of hyaloclastite does not necessarily imply that the vents were also submarine, as subaerial volcanoes may erupt lava that flows into the sea and fragments (e.g. Moore et al., 1973).

Non-stratified sediment-matrix breccia facies

This facies is a texturally complex mixture of rhyolite to andesite and either, grey cherty siltstone, sandstone, non-welded pumice breccia and sandstone or crystal-rich sandstone. The relative proportions of each component varies considerably. In sediment matrix-poor breccia, the rhyolite to andesite is coherent, mildly fractured, or comprises tightly packed jigsaw-fit clusters of clasts separated by up to 1 cm of siltstone or pumice breccia and sandstone. Along some contacts with the host facies, inclusions, tongues and seams of siltstone are present in the rhyolite to andesite up to several metres away from contacts with the host. In sediment matrix-rich breccia, the igneous component often comprises less than 10 to 30% of the breccia. Clasts form jigsaw-fit aggregates or else are dispersed widely in the sediment matrix. Matrix-rich breccia often grades into matrix-poor breccia which in turn grades into coherent facies or in situ hyaloclastite of the same composition (Fig. 5.3D). In other cases, matrix-poor breccia forms a 2 to 5 m thick zone between the coherent facies and surrounding undisturbed host facies.

Clasts vary from blocky, cuneiform, ragged to globular in shape (Fig. 5.4E-G). Blocky and cuneiform clasts are angular with arcuate and subplanar or finely serrate margins. Some larger blocky clasts are enveloped by smaller cuneiform fragments and whole and broken crystals. Clasts occur in jigsaw-fit clusters and dispersed in the sediment matrix (Fig. 5.4E). Ragged clasts are elongate to lenticular and show progressive stages of in
situ disintegration into smaller fragments (Fig. 5.4F). Globular clasts have bulbous, fluidal shapes. The non-stratified sediment-matrix breccia facies may be dominated by one clast shape, or consist of a mixture of clasts with different shapes, or comprise distinct zones of clasts with different shapes (Fig. 5.4G; Appendix A). In andesitic breccia facies, only blocky clasts have been observed.

Clasts which are entirely perlitic were originally glassy. Some clasts that have altered to assemblages of chlorite, sericite, and albite may also have been glassy. The alteration is typically pervasive, without evidence of chilled rinds. In other cases, rhyolite or dacite adjacent to sediment-filled fractures is perlitic and the remainder of the rock comprises a mosaic of fine quartz and feldspar. The perlitic fractures are filled with quartz, sericite or chlorite. Formerly glassy kernels have devitrified or altered.

Lamination or bedding in the surrounding siltstone or sandstone rarely extends into the non-stratified sediment-matrix breccia facies. Where lamination or bedding in the host is preserved, it is generally contorted and extends to within a few centimetres of contacts with the non-stratified sediment-matrix breccia facies (Fig. 5.4G). Along many contacts, the sediment is silicified and has a different colour to the surrounding host, suggesting that it is indurated (Fig. 5.4E). Induration can be limited to sediment immediately adjacent to the clasts or else be more extensive.

Intervals of the non-stratified sediment-matrix breccia facies range in thickness from a few tens of centimetres to in excess of 30 m (rhyolite to dacite). Due to poor exposure, the facies can rarely be followed along strike for more than a few metres. Some rhyolite and dacite intervals consist entirely of non-stratified sediment-matrix breccia (e.g. REM 122, 152-161 m; Fig. 5.3E). In most cases, the non-stratified sediment-matrix breccia facies occurs at contacts between coherent rhyolite to andesite and the host sediments, and/or is interleaved with coherent facies and/or hyaloclastite (Fig. 5.3A). Gradations between the coherent facies, hyaloclastite and the non-stratified sediment-matrix breccia facies occur both laterally and vertically over distances of a few centimetres to several tens of metres.

Origin and significance of facies
These breccias, and particularly the matrix-rich intervals, superficially resemble deposits from small debris flows of igneous clasts and sediment. Emplacement as a debris flow is incompatible with the widespread preservation of jigsaw-fit fabric and gradational contacts with coherent facies (cf. Allen, 1992). An origin by the filtering of sediment down into hyaloclastite or autobrecchia does not account for: (1) matrix-support in parts of the breccia; (2) occurrences of breccia along the base of flows; (3) local induration of the sediment; (4) disrupted lamination and bedding within the breccia facies; and (5) occurrence within the host sediment of sediment gravity flow deposits.
The components and organisation favour interpretation of the breccia facies as peperite (e.g. Jones, 1969; Williams and Mc Birney, 1979). Peperite provides evidence for the mixing of magma or lava and wet unconsolidated sediment and has been described in many subaqueous, particularly submarine, volcanic successions (e.g. Fisher, 1960; Schmincke, 1967; Williams and Mc Birney, 1979; Brooks et al., 1982; Kokelaar, 1982; Busby-Spera and White, 1987; Hanson and Wilson, 1993; Brooks, 1995; Goto and Mc Phie, 1996). The examples described by these authors include magma compositions ranging from rhyolitic to basaltic. The shape of clasts in peperite is an indication of fragmentation processes (e.g. Busby-Spera and White, 1987). Globular clasts and fluidal surfaces are thought to form where a vapour film is established and maintained at the magma-sediment interface (e.g. Kokelaar, 1982). The vapour film insulates the magma from the wet sediment and suppresses both quench fragmentation and steam explosions. Sediment is displaced along the contact zone until cooling below a critical temperature causes the steam to condense and sediment to be deposited. Portions of the magma can be detached, sourcing globular clasts to the surrounding sediment. Magma-sediment interactions which involve quench fragmentation, dynamic stressing or steam explosions generate clasts with blocky and ragged shapes (e.g. Busby-Spera and White, 1987; Branney and Suthren, 1988; Appendix A).

**Massive andesite facies**

Massive andesite in the Trooper Creek Formation is aphyric, weakly feldspar-phyric or pyroxene- and plagioclase-phyric. Phenocrysts are euhedral and distributed evenly throughout a groundmass which is microcrystalline or formerly glassy. Groundmass microphenocrysts and microlites are plagioclase±pyroxene. Pyroxene (clinopyroxene) is unaltered or has been variably pseudomorphed by chlorite. The andesites typically contain 1-15% round to ellipsoidal vesicles or amygdales, 0.5 mm to 5 cm across. Amygdales are filled with quartz, chlorite, carbonate, feldspar, or zones of chlorite-feldspar or chlorite-quartz. Flow foliations in the andesites are defined by trains of aligned ellipsoidal vesicles and/or microlites in the groundmass.

Intervals of andesite range between 20 m and 170 m in thickness and can sometimes be traced for 1 km along strike. Massive andesite locally grades into non-stratified andesitic breccia facies or non-stratified sediment-matrix breccia facies along contacts with underlying or overlying volcano-sedimentary facies. Contacts between the different facies are generally sharp but gradational.
Non-stratified andesitic breccia facies

These breccias are monomictic, clast-supported frameworks of blocky and cuneiform andesite fragments, 5 mm to 30 cm across. Clasts have planar to curvilinear margins and are separated by minor millimetre- to centimetre-sized cuneiform andesite fragments, whole crystals and crystal fragments (feldspar and pyroxene). Flow banded clasts are not abundant. Breccia with jigsaw-fit clasts grades into clast-rotated breccia in which clasts have separated up to a few centimetres following fragmentation (Fig. 5.4H). Contacts with associated coherent andesite are sharp but gradational. Along some of these contacts, intersecting planar and curvilinear fractures dissect the andesite into polyhedral and cuneiform blocks which are separated by little or no matrix. Intervals of the non-stratified andesitic breccia facies are typically less than 1 to 2 m thick and rarely exposed for more than a few metres along strike.

Clasts in the breccia have altered to various assemblages of chlorite, sericite, carbonate and/or quartz. The fractures and/or matrix are accentuated by sericite or quartz. In some cases, these alteration minerals have encroached on the margins of andesite clasts partly obscuring original clast shapes and creating a more extensive apparent matrix domain.

Origin and significance of the massive and brecciated andesite facies

The groundmass textures, even distribution of phenocrysts, and contact relationships of the massive andesite facies are consistent with coherent igneous rock. The in situ breccia is interpreted as hyaloclastite formed through the propagation of networks of quench fractures in the cooling andesite. Characteristic of hyaloclastite are intimate and gradational relationships with associated coherent facies, in situ jigsaw-fit textures, and clasts with blocky to cuneiform shapes and curvilinear margins. Clasts in the clast-rotated hyaloclastite retain shapes characteristic of quench fragmentation but have undergone rotation, separation, and probably further granulation, due to flow concurrent with fragmentation (cf. Pichler, 1965). Andesitic hyaloclastite from the Trooper Creek Formation is similar to hyaloclastite (B) of Yamagishi (1979, 1991).

Basaltic andesite lobe facies

At Highway East prospect (around 7746800 mN, 418900 mE), outcrops of basaltic andesite have an apparent lobe-in-matrix fabric. The basaltic andesite lobe facies is exposed in one creek section for less than 10 m. The lobes are elliptical to tongue-shaped when viewed in two dimensions. They are 5 to 11 cm wide and vary from 2 to 20 cm long. In detail, many are interconnected and bifurcate (Fig. 5.5A–B). The lobes have recessive reddish, finely vesicular rinds (vesicles 400-800 μm across), 1-2 cm wide. The rinds surround a dark brown core in which the size (400 μm-2 mm) and abundance of
Figure 5.5

(A) Basaltic andesite with an apparent lobe-in-matrix fabric. The lobes have recessive reddish rinds which surround a dark brown core in which the size and abundance of vesicles increase inward. The inter-matrix material (m) is pervasively chloritised and has a similar abundance and distribution of vesicles and phenocrysts to lobe margins. Highway East prospect; 7746800N, 418900E.

(B) Sketch showing the principal textures in A. The inter-lobe matrix comprises chlorite (black) and domains of chlorite with minor crystallised basaltic andesite (stippled). Lobes (white) contain amygdales and vesicles. Diameter of lens cap is 6 cm.

(C) Photomicrograph of the interior of lobes shown in 5.5A. Feldspar microphenocrysts (arrow) are separated by a dense network of microlites. Vesicles (v) are filled with quartz±chlorite. Plane polarised light.

(D) The inter-lobe matrix consists of euhedral feldspar microphenocrysts (arrow) and chlorite containing abundant single oxide globules that probably replaces glass. Plane polarised light. 94-259; Highway East; 7746800N, 418900E.

(E) Clasts in this breccia were formed by quench fragmentation. The bed exhibits normal grading and sharply overlies coherent flow-banded rhyolite (c). The section of drill core is typical of resedimented hyaloclastite. Drill core is 5 cm wide. DDH REMM 121 (120-126m); Highway-Reward; Grid 10024.51N, 10900.07E.

(F) This polymictic breccia consists of blocky rhyolite clasts (r) which are supported in a matrix of pale brown, friable siltstone (s). The remaining components in the breccia are siltstone rafts and rhyolite clasts with rinds of indurated siltstone. The components and organisation are consistent with resedimented intrusive hydroclastic breccia (Section 5.3.2). Highway South prospect; 7745420 mN, 421400 mE.

(G) Part of the outcrop shown in 5.5F contains jigsaw-fit rhyolite clasts (arrow) with fractures and rinds of indurated siltstone which is different from the siltstone matrix to the breccia; together they define clasts which are similar to blocky pepyrite. Highway South prospect; 7745420 mN, 421400 mE.

(H) The other component in the breccia illustrated in 5.5F is rafts of friable siltstone to sandstone (s, arrows). Relic lamination is preserved in the largest rafts. Highway South prospect; 7745420 mN, 421400 mE.
vesicles increase inward. The inter-lobe matrix is pervasively chloritised and has a similar abundance and distribution of vesicles and phenocrysts to lobe margins. Contacts with the lobes are sharp. Euhedral feldspar phenocrysts and ellipsoidal vesicles cut across the boundaries.

In thin-section, groundmass textures in the lobes differ from the core to rim. Lobe interiors consist of feldspar microphenocrysts separated by a felted network of microlites. Vesicles are filled with quartz±chlorite (Fig. 5.5C). At lobe margins, the matrix is chloritised and the abundance of oxide granules decreases gradationally but sharply outward. There is an outer zone of chlorite-filled vesicles and an inner zone in which vesicles walls are lined by feldspar and vesicle centres are filled by chlorite. The inter-lobe matrix consists of euhedral feldspar microphenocrysts and chlorite containing oxide granules that probably replaces original glass (Fig. 5.5D). Arcuate chlorite-rich networks may outline former fractures. Vesicles are filled by chlorite and some vesicle walls are outlined by a thin oxide film. Feldspar microphenocrysts in the lobes and inter-lobe matrix are the same size (150-400 μm) and have similar distributions. There are no crystal fragments between euhedral feldspar phenocrysts in the inter-lobe matrix. The transition from the lobes to the inter-lobe matrix, although sharply defined in hand specimen, is gradational in thin-section.

**Origin and significance of facies**

The lobes superficially resemble pillows but they lack characteristic surface features (e.g. ropy wrinkles, tension cracks), mutually accommodating shapes and internal structures (e.g. tortoise-shell contraction cracks, radial columnar joints, pipe vesicles or spreading cracks; Yamagishi, 1985, 1991). There is no interstitial sediment between the lobes and the margins of single lobes are indistinct. They cannot be interpreted as fractured and dismembered lava lobes, extruded into and completely enclosed by related glassy hyaloclastite, as there are no fragments in the inter-lobe matrix.

The lobes are interpreted to have developed progressively as initial fractures allowed access of external water to the interior of the sheet (Fig. 5.6A) and caused rapid quenching to glass and further fracturing (Fig. 5.6B) Those parts of the sheet which did not quench to glass cooled at a slow enough rate to allow crystallisation (Macdonald, 1972; Fisher and Schmincke, 1984). Continued flow within the hot interior may have been important in generating fluidal-shaped lobes.

Vesicles in both the glassy and more crystallised domains are largely round suggesting that they formed after flow had ceased. However, vesicles in the inter-lobe material are smaller than those in the lobes suggesting they were trapped early. Bubbles are largest in the centre of lobes as these had more time to nucleate, coalesce and grow (Fig. 5.6C).
In the basaltic andesite lobe facies, alteration and regional greenschist metamorphism have converted the originally glassy domains to chlorite. Domains that consist largely of microlites and microphenocrysts are less altered, with only minor interstitial chlorite after glass.

5.3.2 Resedimented volcanioclastic facies association

**Stratified monomictic rhyolitic to dacitic breccia facies**

The stratified monomictic rhyolitic to dacitic breccia facies contains clasts with morphologies, textures and compositions similar to the massive monomictic rhyolitic to dacitic breccia facies. Breccia at the base of a few beds includes rare cherty siltstone clasts, volcanic lithic clasts and pumice (e.g. TA 024, 94 m).

Beds can be normally graded or massive, and range from 0.3 to 11 m in thickness (Fig. 5.3F-G; 5.5E). Massive breccia beds are close-packed frameworks of clasts up to 30 cm across, separated by sand and granules of the same composition. Normally graded beds
Subaqueous lavas and intrusions

typically have matrix-poor, clast-supported breccia at the base and fine sandstone tops (Fig. 5.3F). A few beds have thin reverse-graded sandstone bases (e.g. HDD 008, 58 m; Fig. 5.3G) and others are matrix-supported. Sandstone beds are normally graded or diffusely laminated and, in some cases, subtle variations in grain size mark the contacts between poorly defined amalgamated units.

Intervals of the stratified monomictic rhyolitic to dacitic breccia facies range from a few metres to more than 20 m thick and either comprise a number of graded units or else a single very thick breccia bed. Near Vicesquad prospect, resedimented hyaloclastite overlying the in situ facies contains clasts up to 0.5 m across along the contact. Both mean (base, 3 cm; top, 1 cm) and maximum (base, 0.5 m; top, 10 cm) clast size decrease towards the top of the resedimented hyaloclastite interval. Some intervals of the stratified monomictic rhyolitic to dacitic breccia facies overlie in situ hyaloclastite, peperite, sediment-matrix breccia or coherent rhyolite to dacite which have similar, or distinctly different, phenocryst assemblages, sizes and abundances. Basal contacts of these intervals are sharp or gradational. In a few cases, the stratified monomictic rhyolitic to dacitic breccia facies is enclosed by massive and finely-laminated siltstone (e.g. HDD 022, 220-230 m; Fig. 5.3H). Coherent facies and autoclastic breccia with similar phenocryst assemblages are sometimes exposed in the correlative sections. In drill core at Highway-Reward, intervals of the stratified monomictic rhyolitic to dacitic breccia facies are laterally discontinuous and cannot be traced for more than a few 10's of metres. Contacts with underlying siltstone units are sharp and a few are weakly erosive (e.g. REMM 124, 125 m).

Origin and significance of facies

Gradations between intervals of the stratified monomictic rhyolitic to dacitic breccia facies and in situ hyaloclastite suggest that the units are genetically related. Although jigsaw-fit fabric is lost in the stratified facies, the angular blocky nature of clasts is consistent with a quench fragmentation origin. These components, associated facies, and the stratification are consistent with resedimented hyaloclastite (e.g. Dimroth et al., 1978). Resedimentation of clasts from the quench fragmented margins of lavas, domes and partly-extrusive cryptodomes may have been syn-eruptive, in response to continued magma supply or oversteepening of dome margins, or post-eruptive and related to collapse of unstable dome of flow margins, or triggered by seismic activity.

The bedforms are consistent with deposition from high-concentration sediment gravity flows. Intergranular dispersive pressure and fluid turbulence may have been important particle support mechanisms (Lowe, 1982). Diffusely stratified intervals may reflect zones of high shear at the base of flows and fluctuations in the current (Kokelaar, 1993).
Dimroth et al. (1978) recorded a lateral decrease in grain size with distance from source in resedimented hyaloclastite associated with Archaean basalt lavas in the Rouyn-Noranda area, Quebec, Canada. Because it is not possible to trace single beds in drill core at Highway-Reward, lateral grainsize variations cannot be assessed.

**Indurated siltstone-matrix rhyolitic breccia facies**

Exposures of this facies are limited to Highway South prospect (around 7745420 mN, 421400 mE). The breccia is poorly sorted, matrix-supported and contains abundant rhyolite clasts up to 0.5 m across which have blocky shapes and planar to curviplanar margins, or are irregular with ragged margins (Fig. 5.5F). Some clasts form jigsaw-fit aggregates which are enclosed by thin rinds of indurated siltstone and are penetrated by thin siliceous sediment seams (Fig. 5.5G). The other clasts are siliceous (indurated?) siltstone clasts and friable laminated siltstone clasts which can be more than a metre in length (Fig. 5.5H). A matrix of brown, friable, non-stratified (andesitic?) siltstone surrounds the clasts. The deposit is massive to weakly normally graded and underlain and overlain by planar laminated siltstone. The base of the unit is not exposed. However, the minimum true thickness is 20 m. The lateral extent is poorly constrained (> 30 m).

**Origin and significance of facies**

Rhyolite clasts in this facies retain shapes which suggest they formed by quench fragmentation. Jigsaw-fit aggregates of clasts with fracture fillings and rinds of indurated siltstone are similar to blocky rhyolite-siltstone peperite. Although these clasts formed during the mixing of magma and wet sediment there is evidence for transport following fragmentation. This includes: (1) weak normal grading; (2) an absence of gradational contacts with coherent lava or hyaloclastite; (3) rhyolite clasts with fractures and rinds filled by siltstone which is different from the enclosing matrix; and (4) clasts of indurated siltstone which are isolated in non-indurated siltstone. Deposition from a subaqueous debris flow, in which larger clasts were supported by a watery silty interstitial fluid/continuous phase is suggested (e.g. Lowe, 1982; Smith, 1986).

Peperite incorporated into mass-flows during breaching of the seafloor by intrusions has previously been termed resedimented peperite or redeposited peperite (e.g. Hanson and Wilson, 1993). Resedimentation involves initial deposition of an aggregate followed by re-entrainment into another transport system and deposition at another locality. In the case of resedimented peperite, intrusion and mixing with the sediment may have been synchronous with eruption and entrainment into mass flows (e.g. White and Busby-Spera, 1987) so that use of the term is questionable. Peperite is a genetic interpretive term that is best restricted to in situ, or near in situ, breccia. In order to preserve the distinction and avoid confusion, the term resedimented intrusive hydroclastic breccia is adopted here.
This term also accommodates steam explosions as a potentially important fragmentation process accompanying shallow intrusion and extrusion.

As for peperite, the shape of igneous clasts in resedimented intrusive hydroclastic breccia reflect primary fragmentation processes (e.g. Busby-Spera and White, 1987; White and Busby-Spera, 1987). Blocky and ragged fragments reflect quench fragmentation and dynamic stressing (Kokelaar, 1986) respectively. Groups of jigsaw-fit clasts separated by sediment are thought to be characteristic of these fragmentation processes (e.g. Hanson and Wilson, 1993; Cas et al., 1990). However, similar textures can form if mass flows contain clasts with pre-existing fractures that progressively open during transport and become filled with finer sediment. Identification of resedimented hydroclastic breccia is therefore dependent on identifying igneous clasts with rings of indurated sediment or with fractures filled by sediment which is texturally or compositionally different from the enclosing matrix. It is important to note that the sedimentary component in peperite is not always indurated, especially in fluidal peperite, so that recognition of resedimented hydroclastic breccia can be difficult.

In the Trooper Creek Formation, the resedimented intrusive hydroclastic breccia facies provides evidence for contemporaneous shallow intrusion and extrusion of rhyolite. The debris flow is unlikely to have travelled far from the source, which is likely to be intrabasinal and may be a partially extrusive cryptodome.

*Stratified siltstone-matrix rhyolitic to dacitic breccia facies*

This facies is texturally similar to the indurated siltstone-matrix breccia facies. The igneous components vary between deposits and are either feldspar and quartz crystals and rhyolite clasts, or feldspar crystals and dacite clasts. Igneous clasts are elongate with ragged margins or are blocky and bound by planar to curviplanar surfaces. Groups of clasts locally display jigsaw-fit fabric (e.g. REMM 121, 268-275 m). Some clasts are cut by siltstone-filled fractures and/or seams (e.g. REMM 147, 236-250 m). The remaining components are siltstone clasts, siltstone matrix, sparse felsic volcanic clasts, and rare coarsely porphyritic (quartz and feldspar) formerly glassy lenses which are now chlorite and sericite. These lenses are interpreted to have originally been pumice clasts, although vesicular textures are rarely preserved. Some siltstone clasts are blocky with sharp margins. Others have irregular or contorted shapes, contain scattered crystal fragments near their margins, and merge with the matrix, suggesting that they were not fully lithified at the time of incorporation into the breccia and are intraclasts. The siltstone matrix includes sand-sized igneous fragments, crystals and crystal fragments.
Beds are mostly less than 10 m thick and either massive and ungraded or normally graded with siltstone to sandstone tops. Most beds are matrix-supported, but a few are clast-supported. Tops of beds are sharp if overlain by other graded units, and otherwise pass gradationally up into fine-grained ambient sedimentary facies (turbidites or mudstone). Bases of beds vary from sharp to diffuse and are locally erosional. Beds are intercalated with siltstone, turbidites and resedimented hyaloclastite or overlie in situ peperite, hyaloclastite or coherent facies of the same composition. In drill core, it is not possible to trace single beds for more than around 10 m.

**Origin and significance of facies**

Clasts in the siltstone-matrix rhyolitic to dacitic breccia facies retain shapes which suggest they formed by quench fragmentation and dynamic stressing (Kokelaar, 1986). Groups of jigsaw-fit clasts may have been hot when incorporated in the mass-flow and broken in situ by cooling contraction granulation during transport or following deposition, or else were entrained cold but with already prepared fractures that progressively opened during transport. In either case, a lava, lava dome or cryptodome probably contributed the igneous components, and gravitational collapse probably initiated the sediment gravity flows. The siltstone matrix implies that sediment gravity flows eroded the substrate during transport or that parts of the lava or dome were covered by a sediment carapace that was also dislodged by collapse. Sediment filling fractures within the igneous clasts is texturally and compositionally similar to the matrix and not indurated, so there is no positive evidence for interpretation of the breccia as resedimented hydroclastic breccia. However, it remains a viable alternative.

The components and organisation suggest that intervals of this facies were deposited from high-concentration sediment gravity flows, possibly debris flows, that incorporated a significant component of largely unconsolidated silt, either from the substrate during transport or at source. Particle support probably depended on the cohesive silt matrix. Breccia beds with weakly graded tops probably record increasing dilution of the flow by incorporation of water or watery silt (e.g. Smith, 1986).

5.4 The Highway-Reward volcanic centre

5.4.1 Interpretation of the lithofacies associations

Syn-sedimentary sills, cryptodomes, and partly extrusive cryptodomes have been identified in the host succession to the Highway-Reward massive sulfide deposit. Upper contact relationships and the distribution and arrangement of coherent facies, hyaloclastite, peperite and resedimented autoclastic breccia are the basis for determining
the mode of emplacement. Distinguishing between the intrusive and extrusive units is important because the stratiform mineralisation at Highway-Reward can only be correlated using units emplaced at the seafloor (i.e. lavas, sediments and some volcaniclastic deposits).

**Syn-sedimentary sills and cryptodomes**

Contacts of the syn-sedimentary intrusions in the Highway Member show varying effects of quenching and of interaction with the poorly consolidated, water-saturated host sequence. Mixed contacts occur along with unmixed contacts. Mixed contacts vary from smooth to irregular with tongues and apophyses of peperite and associated coherent facies invading the host. Along parts or all of some contacts, matrix-rich peperite passes through a zone of matrix-poor peperite with jigsaw-fit clasts into a coherent core. In situ hyaloclastite locally occurs between the coherent facies and peperite. Irregular seams and wisps of sandstone and/or siltstone that occur within coherent facies were enveloped by the intrusion and injected along cooling fractures, perhaps simultaneous with fluidisation of the host sediment. Along the margins of other intrusions (e.g. RD 812, 150-201 m), contacts between the coherent facies and matrix-rich peperite are sharp.

With a few exceptions which are predominantly massive, the intrusions consist of at least 10 percent by volume of in situ hyaloclastite and peperite. One thin (7-11 m thick) sill (rhyolite 5) consists entirely of matrix-rich and matrix-poor peperite. At Policeman Creek prospect (0.4 km to the northeast), a single pumiceous mass-flow unit is intruded by a rhyolitic sill with sharp planar contacts which show little interfingering with the host and intruded subparallel to bedding (e.g. HDD 004, 260.49-331.83 m). Flow banding at the margin of the sill mirrors the contact, but within the interior varies widely in orientation and can be strongly contorted. Pumice breccia within a 6.5 m zone above the rhyolite is bleached and silicified due to induration and alteration caused by the intrusion. Tube pumice textures have been destroyed at the contact, and there are no signs of secondary welding (cf. McPhie and Hunns, 1995). However, silicification decreases gradationally away from the contact and non-welded pumice is identifiable in both hand specimen and thin-section.

**Partly extrusive cryptodomes**

Partly extrusive cryptodomes are high-level domes that intrude the sediment pile and locally breach the sediment surface. They combine facies common to shallow syn-volcanic intrusions and extrusive domes. The example intersected in drill core at Highway-Reward (rhyolite 4) is similar to well described examples from similar subaqueous volcanic successions (e.g. Cas et al., 1990; Allen, 1992; Hanson and
Figure 5.7 Cartoon illustrating successive stages in partial extrusion of a cryptodome. (A) Magma advances into wet unconsolidated sediment. Mixing of the magma and sediment is accompanied by up-doming of the intruded section. (B) Intruding magma destabilises the section and slumping occurs. Mass flows of up-arched sediment, mixed sediment-igneous clasts and igneous clasts are deposit on the seafloor. (C) Quench fragmentation of extruded lava forms an unstable hyaloclastite pile, subject to collapse and resedimentation into flanking environments.
Wilson, 1991; Hanson and Wilson, 1993). The rhyolite has a squat dome shape and a relatively flat top. Contacts with the intruded host sequence comprise complex associations of peperite, hyaloclastite and coherent facies. Within the contact zone, bedding in the sediment has been destroyed or disrupted (Fig. 5.7A). Along extrusive margins a coherent core passes gradationally out into in situ hyaloclastite. In some cases, in situ hyaloclastite is overlain locally by resedimented hyaloclastite and stratified sediment matrix-rich breccia formed during local extrusion or in response to over-steepening of the dome surface following breaching of the seafloor (Fig. 5.7B–C). The cryptodome varies from 40 to 120 m in thickness. The diameter of the body (> 300 m) is uncertain due to insufficient drill core.

**Indeterminate units**

Poor exposure hampers the interpretation of emplacement mechanism of many rhyolites, dacites and andesites in the study area. In other cases, intense hydrothermal alteration and mineralisation, shearing and faulting have combined to destroy critical upper contact relationships required to interpret the mode of emplacement.

### 5.4.2 Volcano morphology and size

The Highway-Reward massive sulfide deposit is hosted by a silicic intrusion-dominated sequence, that includes sedimentary facies, turbiditic sandstone and pumiceous and crystal-rich sandstone. Contact relationships and phenocryst mineralogy, size and percentages indicate the presence of more than thirteen distinct porphyritic units in a volume of 1 x 1 x 0.5 km (Table 5.1; Figs. 5.8-5.9). More than 75% of the rhyolites to dacites studied in the immediate host sequence to the Highway-Reward deposit are entirely intrusive. Another 15% remain undifferentiated lavas and intrusions, as no diagnostic criteria of the emplacement mechanisms are preserved. Evidence for partial extrusion of dacitic to rhyolitic sills above the seafloor is limited to one example (rhyolite 4). Single porphyries vary from <10 to 350 m in thickness and some are less than 200 m in diameter. The intrusions have steep margins and are often separated from neighbouring porphyries by only a thin (0.2 to 30 m) interval of peperite, siltstone, crystal-lithic breccia or pumice breccia. Many of the smaller intrusions are sandwiched between much larger cryptodomes. The margins of the intrusions and lavas commonly conform to the surfaces of the neighbouring porphyries (Fig. 5.8).

The syn-sedimentary sills and cryptodomes have dissected the pre-existing host succession, so that correlation within the host volcanic succession is difficult. Many previous interpretations of the volcanic succession have attributed this complexity to
intense faulting. Pumiceous mass-flow units provide the best available framework for correlation within the host succession, as these were emplaced infrequently, are widespread and have distinctive phenocryst populations (cf. McPhie and Allen, 1992). Partly extrusive cryptodomes, resedimented hyaloclastite and resedimented hydroclastic breccia are important indicators of palaeo-seafloor positions at Highway-Reward. Sills and cryptodomes may have influenced seafloor topography and therefore sedimentation, but do not mark sea-floor positions. Removing the syn-sedimentary intrusions and cryptodomes leaves a relatively simple sedimentary stratigraphy, which is similar to the remainder of the “relatively deep-water” facies association in the Highway Member.

Table 5.1 Form, dimensions and phenocryst populations for sills, cryptodomes and partly extrusive cryptodomes at Highway-Reward.

<table>
<thead>
<tr>
<th>Unit</th>
<th>Form</th>
<th>Dimensions (m)</th>
<th>Quartz</th>
<th>Feldspar</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td>Length</td>
<td>Width</td>
<td>Thickness</td>
</tr>
<tr>
<td></td>
<td></td>
<td>(mm)</td>
<td>(mm)</td>
<td>(mm)</td>
</tr>
<tr>
<td>Dacite 1</td>
<td>cryptodome</td>
<td>250</td>
<td>300</td>
<td>&gt;300</td>
</tr>
<tr>
<td>Rhyodacite 1</td>
<td>sill</td>
<td>150</td>
<td>&gt;250</td>
<td>&gt;100</td>
</tr>
<tr>
<td>Rhyodacite 2</td>
<td>cryptodome</td>
<td>175</td>
<td>275</td>
<td>50-150</td>
</tr>
<tr>
<td>Rhyodacite 3</td>
<td>?</td>
<td>&gt;50</td>
<td>&gt;75</td>
<td>&gt;80</td>
</tr>
<tr>
<td>Rhyolite 1</td>
<td>sill</td>
<td>?</td>
<td>?</td>
<td>?</td>
</tr>
<tr>
<td>Rhyolite 2</td>
<td>cryptodome</td>
<td>&gt;100</td>
<td>&gt;125</td>
<td>120-170</td>
</tr>
<tr>
<td>Rhyolite 3</td>
<td>cryptodome</td>
<td>175</td>
<td>5-150</td>
<td>20-100</td>
</tr>
<tr>
<td>Rhyolite 4</td>
<td>partly ext.</td>
<td>312</td>
<td>&gt;300</td>
<td>40-120</td>
</tr>
<tr>
<td>Rhyolite 5</td>
<td>sill</td>
<td>38</td>
<td>25-140</td>
<td>5-10</td>
</tr>
<tr>
<td>Rhyolite 6</td>
<td>sill</td>
<td>150</td>
<td>&gt;300</td>
<td>25 to &gt;60</td>
</tr>
<tr>
<td>Rhyolite 7</td>
<td>?</td>
<td>&gt;37</td>
<td>&gt;60</td>
<td>&gt;30</td>
</tr>
<tr>
<td>Rhyolite 8</td>
<td>sill</td>
<td>75</td>
<td>&gt;300</td>
<td>35 to &gt;85</td>
</tr>
<tr>
<td>Rhyolite 9</td>
<td>cryptodome</td>
<td>?</td>
<td>&gt;350</td>
<td>&gt;225</td>
</tr>
</tbody>
</table>
Figure 5.8. Simplified geological cross-section of the Highway-Reward deposit at 10050N. The location of dacite D1, rhyodacite RD2 and the main rhyolite units (R3-7) are also shown. RL = relative position above sea level.
Figure 5.9 Simplified cross section of the Highway-Reward deposit at 10150N. The positions of rhyolites R3–6, rhyodacites RD1–3 and dacite D1 are shown. RL = relative position above sea level.
5.4.3 Evolution of the volcanic centre

Figure 5.10 is a schematic reconstruction showing successive stages in the development of the Highway-Reward volcanic centre. The character of the centre evolved over its extended history during which there were several successive seafloor positions. The most distinctive of these is marked by deposits of reworked hyaloclastite and pumice breccia which overlie a large, partly extrusive cryptodome (rhyolite 4; Fig. 5.10B). Thick, pumiceous mass-flow units which occur near the top of the succession may record another seafloor position (Fig. 5.10D). The rhyolites, rhyodacites and dacites were probably erupted/ intruded from a series of feeders that could have been located along syn-volcanic faults, rather than from a single vent which would result in a dome-cryptodome complex with a more marked palaeotopographic expression.

5.4.4 Other similar intrusion-dominated, submarine volcanic centres

The morphology of the submarine volcanic centre which hosts the Highway-Reward deposit (Fig. 5.11A) is similar to that described by Allen (1992) for the host sequence of the Silurian Currawong and Wilga massive sulfide deposits, southeastern Australia. Up to 50% of the rhyolites to basalts in the Wilga-Currawong area are syn-volcanic intrusions. Horikoshi (1969) also recorded a dominance of small dacitic to basaltic cryptodomes in Kuroko host sequences in the Miocene Kosaka Formation of NE Japan. The submarine dome-tuff cone volcano described by Horikoshi (1969) is similar to those documented by Cas et al. (1990) and Allen et al. (1996b). The explosive tuff cone-forming eruptions at these volcanic centres are not recorded in the stratigraphy at Highway-Reward, possibly reflecting differences in the volatile content of erupted magma and/or the external confining pressure. Pyroclast-rich mass-flow units within the host succession at Highway-Reward are interpreted to have been sourced from vents in a distal shallow submarine or subaerial setting. The thickness of these units suggests that they may fill topography created by the intrusion and lava complex and/or contemporaneous syn-volcanic faults.

At Highway-Reward, the small size of the lavas and intrusions contrasts with silicic lavas, domes and cryptodomes from many modern and ancient, subaerial (e.g. Clough et al., 1982; Cortese et al., 1986) and subaqueous environments (e.g. Pichler, 1965; Cas, 1978; Allen, 1992; Table 5.2). The size of the lavas and intrusions reflects a restricted supply of magma during eruption/ intrusion. The shapes and emplacement mechanisms of the porphyritic units are in part a reflection of their subaqueous depositional environment. Due to rapid quenching and mixing with unconsolidated sedimentary and volcanic facies, the sills and cryptodomes did not spread far from their conduits. The shape and
Subaqueous emplacement of feldspar ± quartz-bearing pumiceous mass-flow deposits and overlying siltstone. Intrusion of rhyolitic to dacitic syn-sedimentary sills and cryptodomes. The intrusions dismembered the pre-existing host succession and influenced seafloor topography.

Intrusion of a rhyolitic cryptodome (R3) between cryptodomes D1 & D02. Partial extrusion of a rhyolitic cryptodome (R4) through the seafloor forming unstable hyaloclastite subject to collapse and re-sedimentation. Intrusion or extrusion of RD3. Emplacement of distinctive rhyolitic to dacitic pumiceous mass-flow units above R4.

Deposition of quartz- and feldspar-bearing breccia and sandstone units and minor siltstone. Intrusion of syn-sedimentary sills (RS-R7) into re-sedimented hyaloclastite of rhyolite 4, and the overlying deposits.

The sills generated seafloor topography which was infilled by thick rhyolitic to dacitic pumiceous mass-flow deposits sourced from explosive eruptions in a distal sub-aerial or shallow marine environment.

Figure 5.10: Cartoon showing successive stages in the evolution of the syn-sedimentary intrusion-dominated host succession to the Highway-Reward deposit. Only the principal elements are illustrated.
Table 5.2 Dimensions of representative silicic to intermediate lavas and syn-sedimentary intrusions from ancient submarine volcanic successions.

<table>
<thead>
<tr>
<th>Locality</th>
<th>Age</th>
<th>Composition</th>
<th>Morphology</th>
<th>Diameter/Length</th>
<th>Thickness</th>
<th>Reference</th>
</tr>
</thead>
<tbody>
<tr>
<td>Wilga-Currawong, NSW, Australia</td>
<td>Silurian</td>
<td>rhyolite</td>
<td>flow</td>
<td>?</td>
<td>300 m</td>
<td>Allen, 1992</td>
</tr>
<tr>
<td></td>
<td></td>
<td>rhyolite</td>
<td>sill</td>
<td>3 km</td>
<td>500 m</td>
<td></td>
</tr>
<tr>
<td>Rouyn-Noranda, Quebec, Canada</td>
<td>Archaean</td>
<td>rhyolite</td>
<td>flows &amp; domes</td>
<td>up to 10 km</td>
<td>30-400 m; 200 m; De Rosen-Spence et al., 1980</td>
<td></td>
</tr>
<tr>
<td>Tadami, Northern Honshu, Japan</td>
<td>middle Miocene</td>
<td>rhyolite</td>
<td>dome</td>
<td>300 m</td>
<td>?</td>
<td>Kurokawa, 1991</td>
</tr>
<tr>
<td>Saskotan Peninsula, Honshu, Japan</td>
<td>Neogene</td>
<td>rhyolite</td>
<td>dome</td>
<td>50-100 m</td>
<td>50 m</td>
<td>Yamagishi &amp; Matsuda, 1991</td>
</tr>
<tr>
<td>Abishiri area, Hokkaido, Japan</td>
<td>late Miocene</td>
<td>rhyolite</td>
<td>lava lobes</td>
<td>40 m long, 16-24 m wide</td>
<td>5-7 m</td>
<td>Yamagishi &amp; Goso, 1992</td>
</tr>
<tr>
<td>Shimane Peninsula, Honshu, Japan</td>
<td>Miocene</td>
<td>rhyolite</td>
<td>block lavas</td>
<td>4 km</td>
<td>50-200 m</td>
<td>Kano et al., 1991</td>
</tr>
<tr>
<td>Northern Sierra Nevada, California, USA</td>
<td>upper Devonian</td>
<td>rhyolite</td>
<td>sill</td>
<td>?</td>
<td>600 m</td>
<td>Hauston &amp; Schweickcr, 1982</td>
</tr>
<tr>
<td>Ramsey Island, SW Wales</td>
<td>lower Ordovician</td>
<td>rhyolite</td>
<td>sill</td>
<td>?</td>
<td>35 m</td>
<td>Kokelaar et al., 1985</td>
</tr>
<tr>
<td>Moelwyn Hills, central Snowdonia</td>
<td>Ordovician</td>
<td>rhyolite</td>
<td>sill</td>
<td>?</td>
<td>55 m</td>
<td>Kokelaar, 1982</td>
</tr>
<tr>
<td>Japan &amp; Quebec, Canada</td>
<td>Miocene &amp; Archaean</td>
<td>rhyolite</td>
<td>flows</td>
<td>10 km</td>
<td>&lt; 400 m</td>
<td>Yamagishi &amp; Dimroth, 1985</td>
</tr>
<tr>
<td>Boyd Volcanic Complex</td>
<td>Devonian</td>
<td>rhyolite</td>
<td>cryptodome</td>
<td>1300 m</td>
<td>180 m</td>
<td>Cas &amp; Bull, 1999</td>
</tr>
<tr>
<td>Ultima Esperanza district, southern Chile</td>
<td>Jurassic</td>
<td>rhyolite</td>
<td>peperitic cryptodomes</td>
<td>areal extent of c. 3 km²</td>
<td>c. 300 m</td>
<td>Husson &amp; Wilson, 1993</td>
</tr>
<tr>
<td>Kosaka district, Honshu, Japan</td>
<td>Miocene</td>
<td>dacite</td>
<td>cryptodome (M4)</td>
<td>400x220 m</td>
<td>200 m</td>
<td>Horibashi, 1969</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>cryptodome (M6)</td>
<td>800x400 m</td>
<td>200 m</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>cryptodome (M8)</td>
<td>340 m</td>
<td>&gt; 50 m</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>cryptodome (M9)</td>
<td>200 m to &gt;600 m</td>
<td>150 m</td>
<td></td>
</tr>
<tr>
<td>Wilga-Currawong, NSW, Australia</td>
<td>Silurian</td>
<td>dacite-basalt</td>
<td>sills</td>
<td>?</td>
<td>150 m</td>
<td>Allen, 1992</td>
</tr>
<tr>
<td></td>
<td></td>
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<td></td>
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<td></td>
</tr>
<tr>
<td>Hill End Trough, NSW Australia</td>
<td>mid Paleozoic</td>
<td>dacite- andesite</td>
<td>flows</td>
<td>areal extent 1200-2000 km</td>
<td>150-380 m</td>
<td>Cas, 1978</td>
</tr>
<tr>
<td>Southern Sierra Nevada, California</td>
<td>Devonian</td>
<td>andesite</td>
<td>sill</td>
<td>?</td>
<td>av. 0.3-12 m thick</td>
<td>Brooks et al., 1982</td>
</tr>
</tbody>
</table>
distribution of the rhyolites and dacites were further influenced by the positions of previously or concurrently emplaced porphyries. Magma preferentially invaded the sediment, avoiding earlier porphyries or conforming to their margins. Large intrusions and their dewatered envelope may have formed a barrier to the lateral progression and ascent of subsequent batches of magma.

5.5 Other lava- and intrusion-dominated volcanic centres in the Highway Member

Five principal lava- and intrusion-dominated volcanic centres have been identified in the Highway Member (Trooper Creek Formation). They occur at Highway-Reward, Highway East prospect, Handcuff prospect, around Coronation homestead and at Trooper Creek prospect. The morphology, components and size of the volcanic centres vary. The Handcuff and Highway East volcanic centres are separate from, but similar to, the Highway-Reward centre (Fig. 5.11A). They differ from the volcanic centres at Coronation (Fig. 5.11B) and Trooper Creek prospect (Fig. 5.11C).

5.5.1 Highway East

In the area around Highway East, the Highway Member is dominated by silicic to intermediate lavas, cryptodomes, sills and partly-extrusive cryptodomes. Less than 10% of the dacitic to basaltic volcanics are extrusive in origin and these are invariably andesitic in composition. Another 70% remain as undifferentiated lavas and intrusions because critical upper contacts are not exposed. The remaining rhyolites to basalts were emplaced as syn-sedimentary sills and cryptodomes. Rhyolites and dacites display contact relationships and lithofacies associations which are similar to those at Highway-Reward. Andesites occur as both syn-sedimentary sills and lavas.

Andesitic syn-sedimentary intrusions

Many andesites display upper contact relationships which suggest that they were emplaced as syn-sedimentary sills into unconsolidated, wet sediment including volcaniclastic deposits. These include relatively smooth contacts with seams and wisps of siltstone invading the andesite and more complex zones of matrix-poor peperite and pods of coherent andesite. Contacts between peperite and the coherent facies vary from sharp to gradational.

The sills are lenticular, sheet-like, or more irregular in shape and although broadly concordant are locally discordant. Single sills can be 2-13 m thick and only 100-200 m
Figure 5.11 Schematic reconstruction of the three principal types of subaqueous lava- and intrusion-dominated volcanic centres identified in the Trooper Creek Formation. (A) Rhyolitic to dacitic syn-sedimentary intrusion-dominated volcanic centre, e.g. Highway-Reward; after Allen (1992). (B) Stacked sequence of lavas; e.g. Coronation. (C) Lava lobe-hyaloclastite complex, e.g. Trooper Creek prospect; after Cas (1992) and Yamagishi (1987).
long. The largest sill is up to 500 m thick and has an exposed (minimum) length of 1.5 km. This sill bifurcates parallel to bedding and includes intervals of pumice breccia and siltstone. The bifurcation may record a lateral transition from a central, mainly coherent facies, to a marginal facies of coherent andesite and peperite.

**Andesitic lavas**

Entirely coherent andesites have sheet-like morphologies and sharp concordant upper and lower contacts without any autoclastic facies. Lamination in the overlying siltstone beds is undisturbed right to the contact and displays no evidence of induration or fluidisation. These andesites probably represent thin (25-40 m thick) lava flows. However, their interpretation as sills cannot be entirely discounted due to poor exposure.

### 5.5.2 Handcuff volcanic centre

At Handcuff prospect, the arrangement and density of rhyolitic to dacitic lavas and intrusions is similar to that at Highway-Reward and Highway East. Syn-sedimentary sills, cryptodomes and lavas have been recognised. Less than 13% of the dacites and rhyolites are lavas, and another 60% remain as undifferentiated lavas and intrusions. The remaining 30% of the rhyolites to dacites are syn-sedimentary sills and cryptodomes; these have similar shapes and dimensions to those at Highway-Reward. Thick intervals of andesite exposed to the north near Truncheon prospect remain as undifferentiated lavas and intrusions. The lavas and intrusions are intercalated with thick intervals of suspension-settled siltstone, pumice breccia and sandstone to siltstone turbidites.

**Dacitic lava**

Part of the Highway Member intersected in Handcuff diamond drill holes (HDD 007, HDD 008, HDD 012 and HDD 022) provides an example of the facies and facies geometry associated with a small dacitic lava flow. In the northernmost hole (HDD 022), coherent massive and flow-banded dacite is interleaved with intervals of jigsaw-fit hyaloclastite, 1-13 m thick (Fig. 5.12A). The correlative section 150 m to the south, comprises coherent dacite overlain by jigsaw-fit hyaloclastite (Fig. 5.12B). The coherent interior includes some segments (1-10 m thick) of jigsaw-fit hyaloclastite. In drill hole HDD 012 (200 m south of HDD 007), thick mass-flow resedimented hyaloclastite units are intercalated with siltstone and are intruded by a rhyolitic syn-sedimentary intrusion (Fig. 5.12C). Lateral facies changes may thus be significant in silicic lavas (cf. Allen, 1992), involving in some cases a change from massive lava in the central facies to increasingly hyaloclastite-dominated lava, and at the distal margins, into resedimented
Subaqueous lavas and intrusions

5.38.

hyaloclastite. The Handcuff lava thins from around 110 m in the central facies to less than 70 m towards the margin.

Figure 5.12 Graphic lithological logs for parts of diamond drill holes HDD 007 (11575N), HDD 012 (11025N) and HDD 022 (11909.5N) at Handcuff prospect.

5.5.3 Coronation volcanic centre

Whereas the foregoing volcanic centres are dominated by volumetrically small lavas and intrusions, there are also situations where lavas form very thick complexes (Fig. 5.11B). Dacites in the Coronation area form a stacked sequence of lavas, with single flow unit boundaries being marked by variations in the abundance and size of feldspar phenocrysts and sometimes by a carapace of in situ and re sedimented hyaloclastite. Because of poor exposure, the thickness, lateral extent and geometry of single lavas are poorly constrained. Some appear to be at least 500 m thick. The coherent interior is overlain by less than 10 m of hyaloclastite dominated by disrupted breccia facies. Resedimented
hylolclastite overlying the in situ facies contains clasts up to 0.5 m across along the contact. Both mean and maximum clast size decrease towards the top of the resedimented hyaloclastite interval. Although not exposed, flow margins (and fronts) are most likely flanked by in situ and resedimented autoclastic breccia.

A 5 km thick section through the succession has been mapped (Map 1) and this represents a minimum thickness for the sequence. However, the succession includes several faults and abundant andesitic and rhyolitic syn-volcanic intrusions. The andesites and some rhyolites occur as dykes up to 450 m long and 100 m wide. More voluminous intervals of rhyolite have irregular, bulbous shapes and are up to 800 m across. The thickness of the succession suggests that the lava complex may have constructed significant seafloor topography. As a consequence, sedimentary rocks and volcaniclastic facies form a very minor component of the dacite complex. Mass-flow deposits probably infilled topography created by the lavas. Only the most voluminous and energetic mass flows are likely to have overridden the complex.

The thickness of both single lavas and of the succession is similar to that of Archaean, subaqueous felsic lavas in the Rouyn-Noranda area (e.g. De Rosen-Spence et al., 1980; Gibson, cited in Cas, 1992). The rhyolites documented by De Rosen-Spence et al. (1980) are up to 400 m thick and have steep flow fronts (up to 40°). Gibson (cited in Cas, 1992) demonstrated that the rhyolitic lavas commonly form shields of lava with slopes up to 15°. Single domes occur above feeder fissures and extend less than 2 km from the vent.

5.5.4 Trooper Creek dacite dome

At Trooper Creek prospect, a lava lobe-hyaloclastite dome occurs near the base of the Highway Member (Trooper Creek Formation). The dome is around 500 m thick and has an exposed (minimum) length of 1.4 km. Detailed mapping of phenocryst percentages and contact relationships suggests that two distinct phases of dacitic magmatism combined to form the dome. Massive coherent aphyric dacite intrudes an earlier feldspar-phyric lava lobe-hyaloclastite pile (Fig. 5.11C). Contacts between the different dacites are sharp but apophyses of aphyric dacite protrude into the earlier breccia facies. The irregular shape of the contact implies that the aphyric dacite was intruded when the host hyaloclastite was poorly consolidated. Flow banding in the aphyric dacite is often planar and laterally continuous and superficially resembles sedimentary bedding.

The feldspar-phyric dacite consist of at least 80% by volume of hyaloclastite and autobreccia. Relic coherent intervals occur as ellipsoidal to sheet-like lobes, 1-30 m wide and 10-150 m long, enclosed by autoclastic breccia. Lava lobes and pods are often
aligned, defining a crude bedding-like structure. The pods and lobes are zoned with crystallised coherent massive to flow banded cores and a brecciated border zone. The border zone is about 0.3-1 m thick and consists of brecciated dacite which grades into in situ jigsaw-fit hyaloclastite. In situ hyaloclastite passes out through clast-rotated facies into hyaloclastite consisting of disorganised blocks and comminuted fragments (disrupted breccia facies; cf. Yamagishi and Dimroth, 1985; Yamagishi, 1987). The dacite dome is overlain by graded beds of rhyodacitic crystal-vitric sandstone intercalated with intervals of feldspar-bearing resedimented hyaloclastite sourced from the oversteepened margins of the dome. Feldspar-phryic dacite lobes locally intrude the resedimented facies (e.g. DDH. TA 024).

The feldspar-bearing phase of the lava dome at Trooper Creek prospect is interpreted to have grown by propagation of feeder dykes intruding a cogenetic pile of hyaloclastite derived from quenching and autobrecciation of the outer parts of lobes (cf. Yamagishi, 1987). Large parts of the breccia were disrupted during inflation of the pile, in response to oversteepening of the dome margins and intrusion/effusion of lava lobes. Late in the emplacement of the dome, a second petrographically distinct magma intruded the hyaloclastite pile. The lava dome is interpreted to have formed directly above a feeder dyke(s). A single feldspar-phryic feeder dyke, 10 m wide, has been identified in the underlying Mount Windsor Formation.

Yamagishi and Dimroth (1985) proposed a similar model for small felsic domes in the Miocene Green Tuff belt of Japan and the Archaean, Abitibi belt in Canada (Table 6.2). Lava lobes, feeder dykes, and in situ and resedimented autoclastic breccia are all important components of their submarine lava dome model. The internal form and facies of these examples share similarities with other well studied subaqueous and subglacial lava-lobe hyaloclastite complexes (e.g. Pichler, 1965; De Rosen-Spence et al., 1980; Furnes et al., 1980). All have highlighted the importance of feeder dykes and lobes in growth of the effusive pile. De Rosen-Spence et al. (1980) documented proximal to distal facies variations in Archaean subaqueous felsic lavas of the Rouyn-Noranda area and Quaternary equivalents in Iceland. The proximal facies consists largely of coherent lavas and associated feeder dykes. The medial facies comprises lava lobes and relic sections of lobes enclosed by hyaloclastite. The distal facies consists largely of breccia composed of lava lobe fragments and hyaloclastite. In the Trooper Creek occurrence, lava lobes are more abundant and larger towards the base and centre the dome. Although a gradation into coherent facies has not been observed, the transition is consistent with the western exposures being proximal to the vent.
5.6 Controls on magma emplacement

In the Trooper Creek Formation, rising magma that encountered unconsolidated, water-saturated sediment commonly remained sub-surface and was emplaced as syn-sedimentary sills or cryptodomes rather than erupting as lavas or domes. Intrusion of magma is favoured where: (a) its density exceeds that of the surrounding host; and (b) the hydrostatic pressure of the magma is low (McBirney, 1963; Walker, 1989a). The density of volcano-sedimentary units into which the Trooper Creek Formation magmas intruded varied. Siltstone and sandstone probably had densities of between 1.6 and 2 g/cm³ (cf. Moore, 1962; McPhie, 1993), whereas pumiceous breccia units probably had densities of around 1.2 (cf. Kato, 1987). The densities of the rising magmas were greater, probably between 2.2 g/cm³ (rhyolite) and 2.5 g/cm³ (andesite) (Murase and McBirney, 1973). The water column contributed to the confining pressure (lithostatic and hydrostatic pressure) exerted on a magma, and probably promoted intrusion at shallow levels into wet, unconsolidated sediment (cf. Hanson, 1991; McPhie, 1993; Rawlings, 1993). Some syn-sedimentary intrusions were emplaced within a few tens of metres of the seafloor and were locally extrusive. Einsele (1986) identified the upper few hundred metres of sedimentary successions as the favoured site for the emplacement of syn-sedimentary sills. Sediments are wet and poorly consolidated within this zone but at greater depths become progressively denser, more compacted and less amenable to intrusion. At Highway-Reward, the position of previously or concurrently emplaced intrusions was equally as important as the host sediment properties in determining the position, shape and extent of the intrusions.

5.7 Lava- and intrusion-related syn-eruptive resedimented facies

Studies of comparable ancient volcanic successions (e.g. Fiske, 1963; Niem, 1977; Carey and Sigurdsson, 1980; Wright and Mutti, 1981; White and McPhie, 1996) have demonstrated that the shape and texture of fragments can be used to constrain the fragmentation mechanisms and eruption environment. Because particle shapes and populations may be modified during transport and reworking prior to final deposition (e.g. Kokelaar and Romagnoli, 1995), volcaniclastic facies vary widely in the degree to which their characteristics and geometry reflect the eruptive setting and depositional environment. Deposits can be syn- or post-eruptive (McPhie et al., 1993; Chapter 3). Syn- and post-eruptive volcaniclastic deposits can in reality be difficult to distinguish, especially for resedimented autoclastic deposits. Resedimentation of unstable autoclastic breccia at the margins of lavas can conceivably be significantly post-eruptive and still generate deposits that appear syn-eruptive.
Lava and intrusion-related syn-eruptive resedimented facies in the Trooper Creek Formation share common characteristics. (1) Deposits are dominated by formerly glassy juvenile clasts with angular unmodified shapes which suggest that fragments formed by quench fragmentation and were redeposited without significant abrasion. Lithofacies character indicates that subaqueous sediment gravity flows rapidly transported fragments from the source to the site of final deposition. (2) Autoclasts were transported and deposited during the eruptive cycle and so are often intercalated with compositionally equivalent coherent and/or autoclastic facies. (3) Some mass flows travelled a short distance from source (probably 10's to 100's of metres) and their deposits are interbedded with contemporaneous sedimentary facies. If unique, the phenocryst assemblage of clasts within the resedimented facies can be used to identify the source cogenetic lava. Mass flows that incorporate largely unconsolidated silt, either from the substrate during transport or at the source, are also syn-eruptive. The abundance of juvenile fragments, lack of evidence for sedimentary reworking, association with a cogenetic lava or partly extrusive cryptodome, and recognition of intraclasts, are key criteria in identifying these syn-eruptive deposits.

The post-eruptive facies in the Trooper Creek Formation are characterised by clasts with different compositions and erosional histories. Many clasts are subrounded to well-rounded indicating that they were reworked in a high-energy environment prior to redeposition (Chapter 3).

5.8 Influence on the facies architecture

The dominance of syn-sedimentary intrusions in the Trooper Creek Formation between Highway-Reward and Highway South prospect probably significantly modified the physical environment and pore-fluid properties in these parts of the basin. Deformation, disruption, dewatering, resedimentation and low-grade metamorphism of the enclosing sediment accompanied emplacement of the intrusions (cf. Einsele et al., 1980; Delaney, 1982; Kokelaar, 1982; Duffield et al., 1986; Hanson, 1991; McPhie, 1993; Davis and Becker, 1994; Brooks, 1995). Quenching of the cryptodomes generated abundant glassy, perlitic, fractured coherent and breccia facies, and slumping at the margins of some partly extrusive cryptodomes introduced small volumes of glassy detritus into flanking environments. These porous and permeable facies may have acted as conduits for connate and hydrothermal fluids circulating beneath the seafloor.

In the Trooper Creek Formation, emplacement of syn-sedimentary intrusions locally deformed bedding in the enclosing volcano-sedimentary sequence (cf. Pollard et al., 1975; Lorenz, 1984; Duffield et al., 1986; Krynauw et al., 1988) and updoming of
Subaqueous lavas and intrusions

sediment above the intrusions sometimes generated seafloor topography (cf. Yamamoto et al., 1991; Davis and Villinger, 1992). Young submarine sediment mounds associated with high level syn-sedimentary intrusions and massive sulfide mineralisation have been mapped in the Middle Valley rift of the northern Juan de Fuca Ridge using seismic reflection, SeaBeam bathymetry and SeaMARC side-scan imagery (Davis and Villinger, 1992; Goodfellow and Franklin, 1993). Sediment mounds are circular, commonly tens of metres high, and several hundreds of metres across. The largest, a polygenetic mound complex, is over 2 km wide and dissected by orthogonal fractures. Similar sediment domes occur in the Escanaba Trough (Davis and Becker, 1994; Denlinger and Holmes, 1994). Sediment is uplifted by as much as 150 m in these examples and although syn-sedimentary intrusions occur in the area, deformation is interpreted to reflect larger volume injections of magma near the base of the sediment pile. Faults generated during updoming of sediment above syn-sedimentary intrusions not only localise the sediment deformation but focus convecting hydrothermal fluids (e.g. Denlinger and Holmes, 1994). At Highway-Reward, the density of syn-sedimentary intrusions suggests that this was probably important.
circulating pore fluids may hydrothermally alter the cooling intrusion (e.g. Wilshire and Hobbs, 1962; Brooks et al., 1982; Pike, 1983; Hanson and Wilson, 1993) and leach Fe, Si and other elements from glassy parts of the succession (cf. Sigerdsson, 1977). In the Trooper Creek Formation, widespread hematite alteration and discontinuous ironstone lenses are interpreted to have deposited from similar fluids circulating around lavas and intrusions (cf. Einsele et al., 1980; Einsele, 1986; Boulter, 1993a; Chapter 6).

Stable isotope studies of hydrothermal vent fluids at the Escanaba Trough have suggested that the magma may also contribute to the hydrothermal fluid (Böhlke and Shaaks, 1994). In the Escanaba Trough, the hydrothermal fluids ascend growth faults above the intrusion and deposit massive sulfides at the seafloor (e.g. Davis and Becker, 1994). Considering the case of a hydraulically open intrusion, Delaney (1982) demonstrated that if magmatic pressure is high, magmatic fluids will flow out into the host rock and displace connate water around the intrusion. If the magmatic pressure is low, connate water will flow into the intrusion.

In the Trooper Creek Formation, sediments at the margins of many intrusions display evidence of induration and dewatering which is consistent with observations from other volcanic successions (e.g. Kokelaar, 1982; Einsele, 1986; Kano, 1989; Hanson, 1991). Einsele (1986) noted that sediment within several tens of metres of syn-sedimentary sills in the Guaymas Basin, Gulf of California, were indurated, metamorphosed to low grades, and less porous than the enclosing strata. The reduction in porosity accompanying intrusion is a mechanism for creating sufficient space for the magma without uplifting the host sequence. In areas of long-lived magmatism, progressive dewatering and induration of sediments by early intrusions may favour the emplacement of subsequent magma at even shallower levels in the sequence where sediments remain poorly consolidated (e.g. Einsele et al., 1980). Eruption of lavas and domes may result as interactions between magma and sediment are minimised and magma ascends through discrete conduits in intrusions and sedimentary rock.

By acting as a relatively impermeable barrier (cf. McPhie, 1993), the indurated sediment may have inhibited the development of broad convection cells during syn-genetic hydrothermal discharge. Circulating connate fluids and ascending hydrothermal fluids may have been focussed through more porous and permeable units, including the fractured glassy margins of lavas and intrusions, pumice breccia beds, resedimented hyaloclastite units and fractures and faults within the host sequence. An increase in the strength of the host sediment may have also enhanced the ability of fractures to form and remain open. The resulting hydrological configuration probably promoted very efficient, focused, hydrothermal fluid discharge. At Highway-Reward, this was important in localising sulfide accumulation and promoted the formation of the pipe-like sub-seafloor
massive sulfide deposits (Chapter 7).

**Eruption of magma-sediment slurries?**

The dominance of syn-sedimentary intrusions in the Highway-Reward, Handcuff, and Highway East areas suggests that significant volumes of host sediment and volcaniclastic deposits were displaced during magmatism. Venting to the seafloor of fluidised sediment-magma slurries may have been important (cf. Kokelaar, 1982).

Eruption of sediment-magma mixtures is envisaged as an initial stage in breaching of the seafloor by intrusions. Intrusions which breach the seafloor are likely to consume associated fluidisation pipes during ascent or initiate slumps, slides or mass flows destroying evidence of eruption mechanisms. Partial extrusion, seismic activity related to magma emplacement, oversteepening of accumulated debris, and water currents will combine to initiate reseidemation of primary erupted deposits, which accordingly are interpreted to have a low preservation potential.

Examples of eruptions of the style envisaged here are primarily restricted to subaerial and shallow subaqueous environments (e.g. Sanders and Johnston, 1989; Leat and Thompson, 1988). Sanders and Johnston (1989) described an extrusion of “fluidised peperite” in Precambrian shallow subaqueous (< 10m, Lawson, 1972) sediments, Stoer, Scotland. A slurry comprising igneous fragments and sediment is interpreted to have formed up to 400 m below the surface when magma interacted with ground water in porous, unconsolidated sediments.

5.9 Summary

The deeper water parts of the Highway Member (Trooper Creek Formation) comprise the juxtaposed and interleaved products of: (1) explosive volcanism from distal silicic to intermediate volcanoes; (2) scoriaceous deposits from intrabasinal strombolian eruptions; (3) siltstone from volcanic and non-volcanic sources; and (4) intrabasinal, silicic to intermediate, lava- and intrusion-dominated volcanism. Lithofacies character and fossils indicate a submarine (below storm wave base) depositional setting at Highway-Reward.

The Cu-Au-Pb-Zn Highway-Reward massive sulfide deposit is hosted by a silicic lava- and intrusion-dominated volcanic succession. Contact relationships and phenocryst mineralogy, size and percentages indicate the presence of more than thirteen distinct coherent porphyritic units in a volume of 1 x 1 x 0.5 km. The peperitic upper margins to many porphyries demonstrate their intrusion into wet unconsolidated sediment. Syn-
Sedimentary sills, cryptodomes, partly extrusive cryptodomes, and associated in situ and resedimented autoclastic deposits have been recognised. These are the principal facies in the environment of mineralisation and represent a proximal facies association from intrabasinal, intrusive/extrusive, non-explosive magmatism. The shape, distribution and emplacement mechanisms of porphyritic units were influenced by: (a) the relative density of magma to wet sediment; (b) the positions of previously or concurrently emplaced porphyries; and (c) possibly, by syn-volcanic faults which may have acted as conduits for magma. Partly extrusive cryptodomes and deposits of resedimented hyaloclastite are important indicators of palaeo-seafloor positions at Highway-Reward. Sills and cryptodomes may have influenced seafloor topography and therefore sedimentation, but do not mark sea-floor positions.
Chapter 6

Volcanic influences on the formation of iron oxide±silica units in a VHMS terrain
Chapter 6

Volcanic influences on the formation of iron oxide±silica units in a VHMS terrain

6.1 Introduction

Iron oxide±silica rocks form the ore equivalent horizon, or favourable stratigraphic position, within the host succession to many volcanic-hosted massive sulfide (VHMS) deposits and have long been considered a potential marker horizon in mineral exploration (e.g. Large, 1992). Such rocks are associated with many Australian VHMS deposits, including Highway-Reward (Fig. 6.1), Mount Morgan (Taube, 1986; Taube and Messenger, 1994, Messenger and Taube, 1994), Thalanga (Duhig et al., 1992a,b), Mount Chalmers (Large and Both, 1980; Hunns, 1994; Hunns et al., 1994), Scuddles and Gossan Hill (Ashley et al., 1988; Barley, 1992) and Captains Flat (Davis, 1975). The iron oxide±silica rocks may form the ore equivalent volcano-sedimentary units (e.g. Mt Morgan), occur as veins and massive replacements of volcanic host rocks (e.g. Highway-Reward), or occur within a 10 to 50 m thick interval which includes the ore horizon (e.g. Thalanga, Large, 1992). The iron oxide±silica rocks are sometimes laterally separated from mineralisation by a few tens to hundreds of metres (e.g. Duhig et al., 1992a,b) and not all iron oxide±silica units are associated with mineralisation. Similar iron oxide±silica rocks mark the ore equivalent horizon of some Canadian VHMS deposits (e.g. Ridler, 1971), and Japanese Kuroko deposits (Kalogeropoulos and Scott, 1983; Fig. 6.1F).

Recent observations of the seafloor confirm the presence of a variety of Fe-Si enriched hydrothermal precipitates. They may occur as chimneys, thin sediment layers/beds, irregularly shaped mounds, or fill fractures within lavas (Hekinian et al., 1993). Such units are forming around actively venting sulfide mounds at mid-ocean ridge spreading centres and have been mapped at the East Pacific Rise (Barrett et al., 1988; Juniper and Fouquet, 1988; Boyd et al., 1993; Hekinian and Fouquet, 1985; Hekinian et al., 1993; Alt et al., 1987; Janecky and Seyfried, 1984), the FAMOUS site on the Mid-Atlantic Ridge (Juniper and Fouquet, 1988), the Juan de Fuca Ridge (e.g. Normark et al., 1983; Tivey and Delaney, 1986; Hannington and Scott, 1988) and the Galapagos Spreading Centre (Herzig et al., 1988). Similar iron oxide±silica deposits have been recorded at the Vatu Fa Ridge, Lau Basin (Fouquet et al., 1993), the Okinawa Trough (Juniper and Fouquet, 1988) and submarine volcanoes of the Society Islands, South Pacific (Boyd et al., 1993; Hekinian et al., 1993). Others are associated with silicic to intermediate submarine volcanic settings including the PACMANUS site, western Woodlark basin (Binns et al., 1993; Boyd et al., 1993).
There is clearly a spectrum of iron oxide-silica deposits/rocks in host successions to VHMS deposits. A detailed study of iron oxide-silica rocks and alteration in the Trooper Creek Formation between Coronation homestead and Trooper Creek prospect, has led to a better understanding of the relationships of the iron oxide-silica rocks to mineralisation and the volcanic host successions. This research builds on earlier work by Duhig et al. (1992a,b) and demonstrates that many of the iron oxide-silica rocks are sub-seafloor replacements of rhyolite and dacite, peperite, stromatolitic and oncolitic units and pumiceous units (Fig. 6.1G-I). In this chapter, terminology is reviewed, the involvement of iron oxide secreting micro-organisms in iron oxide-silica deposition is assessed, and the role of volcanic facies and volcanism on iron oxide-silica precipitation discussed.

Figure 6.1 (A-E) Cartoon showing the distribution of ironstone lenses associated with several VHMS deposits. Data from Horikoshi (1969), Taube (1986), Large and Both (1990), Duhig et al. (1992b), Sainty (1992), Large (1992) and the present study. (F-H) Schematic representation of the lithofacies associations hosting iron oxide-silica units in the study area.
Iron oxide±silica rocks and alteration in volcanic successions have been variably termed exhalite, ironstone, jasper, jaspilite, chemical sediment, ochre, umber and ferruginous chert. The term “tuffaceous exhalite” has been used for iron oxide±silica rock which overlies some of the Noranda deposits in Canada (e.g. Kalogeropoulos and Scott, 1983) and “tetsusekiei” (literally “iron quartz”) which overlies some Kuroko deposits (e.g. Kalogeropoulos and Scott, 1989), in recognition of the volcanioclastic components in these deposits. Exhalite (Ridler, 1971) is a genetic interpretive term and is not suitable for description of iron oxide±silica rocks for which an exhalative origin cannot be clearly demonstrated. In sedimentary classifications, an ironstone is a rock containing > 15 wt% Fe (James, 1954). In the current study, the term ironstone has been used to refer to massive or laminated iron oxide±silica-rich rock with or without a sedimentary or volcanic component. The term tuffaceous ironstone is adopted for ironstone which contains a recognisable volcanioclastic component (e.g. shards, pumice, scoria, crystals) or alteration products of a former volcanioclastic component.

Intense pervasive hematite alteration occurs at the margins of the Highway-Reward massive sulfide deposit, as stratiform zones beneath some ironstones, as discontinuous pods in andesitic scoria breccia and at the margins of lava domes and cryptodomes. The term iron oxide±silica alteration/rock is adopted here as a non-genetic general descriptive term that can be applied to hematite-altered volcanic rock, ironstone and tuffaceous ironstone.

6.3 Stratigraphic distribution of ironstones in the Seventy Mile Range Group

Small (1-60 m) discontinuous lenses and pods of ironstone are common in the Seventy Mile Range Group (Duhig et al., 1992b; Berry et al., 1992). In the area west of Thalanga mine, ironstones occur in three stratigraphic positions (Duhig et al., 1992b): at the contact between the Puddler Creek Formation and rhyolites of the Mount Windsor Formation; within the Thalanga ore horizon at the contact between the Mount Windsor Formation and the overlying Trooper Creek Formation; and within the Trooper Creek Formation, 80-100 m above the Thalanga ore position.

Elsewhere in the Seventy Mile Range Group, ironstones are largely restricted to the Trooper Creek Formation or occur at the contact between the Trooper Creek Formation and the overlying Rollston Range Formation (Fig. 1.1; Duhig et al., 1992b; Berry et al., 1992; Doyle, 1996). In the study area, ironstones are restricted to the middle and upper
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Trooper Creek Formation, cropping out at Trooper Creek prospect and along strike 1.6 km to the west, north of Trooper Creek prospect, and at Highway East and Handcuff (Map 1). At Highway-Reward, iron oxide±silica alteration/rock occurs in drill core and at the base of the Highway pit.

6.4 Occurrence and volcanic facies

In the study area, iron oxide±silica alteration/rock is associated with different volcanic facies (Fig. 6.1). Four principal facies associations are important. (1) The stratified andesitic breccia and sandstone facies association comprises graded andesitic scoria breccia, cross-stratified andesitic breccia and sandstone, and globular clast-rich andesitic breccia. These units deposited from subaqueous mass flows and by fallout from strombolian eruptions (Chapter 4). (2) The second facies association was sourced from explosive rhyolitic to dacitic eruptions and comprises planar laminated dacitic pumice breccia (water-settled fallout) and graded lithic-crystal-pumice breccia and sandstone (sediment gravity flow deposits; Chapters 4 & 5). (3) Associations of rhyolitic to dacitic coherent facies, peperite and autoclastic breccia which form lavas, domes and cryptodomes. (4) Massive and laminated siltstone units.

Table 6.1 summarises the character and lithofacies associations of iron oxide±silica alteration/rock in the study area. Only the best exposed localities are described in the subsequent sections. The ironstone units are more often hosted by volcaniclastic and sedimentary facies rather than coherent facies, but do not appear to be preferentially associated with one lithofacies type. However, the thickest and longest ironstone lenses are hosted by rhyolitic to dacitic pumice breccia and sandstone units.

6.4.1 Ironstone lenses in rhyolitic to dacitic pumice breccia and sandstone

At Trooper Creek prospect, ironstone occurs at the contact between stratified andesitic breccia facies association and overlying planar laminated dacitic pumice breccia units, forming a discontinuous conformable horizon (horizon 1) approximately 500 m in length (Figs. 6.2–6.3). There is evidence for palaeotopography on the contact between the andesitic breccia and overlying dacitic pumice breccia (Chapter 4). The ironstone lenses occur on both the palaeotopographic highs and palaeotopographic lows. The lenses range in thickness from 1 cm to 10 m and although mostly 10 to 20 m in exposed (actual?) length, one lens has a strike length of approximately 116 m. Detailed mapping of the best exposed sections of horizon 1 ironstone (Figs. 6.3–6.5A) suggests that the lenses occur at
Table 6.1: Significant iron oxide-silica occurrences in the Trooper Creek Formation between Coronation homestead and Trooper Creek prospect.

<table>
<thead>
<tr>
<th>Prospect</th>
<th>Locality</th>
<th>AMG reference</th>
<th>Occurrence</th>
<th>Character</th>
<th>Samples</th>
</tr>
</thead>
<tbody>
<tr>
<td>Trooper Ck</td>
<td>Horizon 1</td>
<td>7741900 mN, 426800 mE</td>
<td>dacitic pumice breccia</td>
<td>massive</td>
<td>95-206, 95-207, 95-208, 95-210, 95-223, 95-276</td>
</tr>
<tr>
<td>eastern lenses</td>
<td></td>
<td></td>
<td>tuffaceous</td>
<td></td>
<td>95-209, 95-275</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>stromatolitic &amp; tuffaceous</td>
<td></td>
<td>95-202, 95-217</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>pumice breccia</td>
<td></td>
<td>95-274, 95-203</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>andesitic breccia</td>
<td></td>
<td>95-272</td>
</tr>
<tr>
<td></td>
<td>Horizon 2</td>
<td>7741900 mN, 426800 mE</td>
<td>dacitic pumice breccia</td>
<td>tuffaceous</td>
<td>95-269</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>microbialite, dacitic pumice breccia</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>Horizon 3</td>
<td>7741900 mN, 426800 mE</td>
<td>dacitic pumice breccia</td>
<td>hematite-altered pumice breccia</td>
<td>94-334</td>
</tr>
<tr>
<td></td>
<td>Horizon 4</td>
<td>7741700 mN, 427000 mE</td>
<td>stratified andesitic breccia</td>
<td>massive</td>
<td>95-316, 95-317B, 95-318</td>
</tr>
<tr>
<td>Trooper Ck</td>
<td>Horizon 1</td>
<td>7742000 mN, 426100 mE</td>
<td>unexposed; occurs above stratified andesitic breccia</td>
<td>massive</td>
<td>95-179, 95-180, 95-211, 95-213</td>
</tr>
<tr>
<td>western lenses</td>
<td>(527)</td>
<td></td>
<td>tuffaceous</td>
<td></td>
<td>95-212, 95-214, 94-327</td>
</tr>
<tr>
<td>Trooper Ck</td>
<td>cattle yard</td>
<td>911</td>
<td>unexposed; occurs above stratified andesitic breccia</td>
<td>massive lenses</td>
<td>95-138, 95-129, 95-136, 95-131</td>
</tr>
<tr>
<td>Trooper Ck</td>
<td>north</td>
<td>946</td>
<td>coherent rhyolite</td>
<td>massive pods</td>
<td>95-149, 95-150</td>
</tr>
<tr>
<td>Highway East</td>
<td>439</td>
<td>7746600 mN, 418250 mE</td>
<td>unexposed; adjacent to rhyolite</td>
<td>massive</td>
<td>94-197</td>
</tr>
<tr>
<td>Highway East</td>
<td>344</td>
<td>7747500 mN, 419900 mE</td>
<td>rhyolitic pumice breccia</td>
<td>massive pod</td>
<td>94-246</td>
</tr>
<tr>
<td>Highway</td>
<td>DDH REW 803</td>
<td>10150N, 10515E (mine grid)</td>
<td>peperitic dacite</td>
<td>veins</td>
<td>REW 803 (38.77 m &amp; 117 m)</td>
</tr>
<tr>
<td>Handcuff</td>
<td>DDH HDD 012</td>
<td>11385 mN, 10654 mE</td>
<td>coherent &amp; peperitic rhyolite</td>
<td>veinlets</td>
<td></td>
</tr>
<tr>
<td>Handcuff</td>
<td>Loc. 53, 69, 72</td>
<td>7743700 mN, 424600 mE</td>
<td>contact &lt;&lt; rhyolite &amp; siltstone</td>
<td>massive with disseminated pyrite</td>
<td>94-25, 94-18</td>
</tr>
<tr>
<td>Handcuff</td>
<td>DDH HDD 007; DDH HDD 022</td>
<td>11571N, 10740E (mine grid)</td>
<td>hyaloclastite &amp; coherent dacite</td>
<td>veins, alteration of matrix in hyaloclastite</td>
<td>HDD 007 (444.3 m)</td>
</tr>
<tr>
<td>Handcuff</td>
<td>Loc. 150</td>
<td>7749500 mN, 418950 mE</td>
<td>coherent dacite</td>
<td>veins</td>
<td>94-61</td>
</tr>
<tr>
<td>Handcuff</td>
<td>Loc. 98</td>
<td>7749520 mN, 418100 mE</td>
<td>massive/laminated siltstone</td>
<td>patches</td>
<td>94-401</td>
</tr>
<tr>
<td>Handcuff/Truncheon</td>
<td>Loc. 76</td>
<td>7748750 mN, 417750 mE</td>
<td>laminated siltstone</td>
<td>laminae</td>
<td></td>
</tr>
</tbody>
</table>
Figure 6.2 Simplified geological map showing the distribution of ironstone lenses and volcano-sedimentary units in the southern part of Trooper Creek prospect.
Figure 6.3 Detailed outcrop map showing the distribution of the principal lithofacies and iron oxide-silica horizons at Trooper Creek prospect (around 7741900 mN, 426800 mE). Locations of sections A-C are shown in Figure 6.2.
Figure 6.4 Measured sections presented as graphic logs for major lithologies in the Trooper Creek Formation in the area of the Trooper Creek Prospect. Locations of sections A–E are given in figures 6.2 and 6.3.

Section A
- Dacitic lithic-pumice breccia
- Globular clast-rich andesitic breccia
- Stratified andesitic sandstone
- Quartz-hematite ironstone
- Laminated siltstone (...)
- Polymictic, matrix-supported pumice-lithic breccia
- Dacite (○), Jasper (□), and laminated siltstone (△) clasts
- Dacitic breccia - hematitic patches and apparent matrix

Section B
- horizon 1
- horizon 2
- horizon 3

Section C
- horizon 1
- horizon 2

Section D
- horizon 1

Section E
- horizon 4

Legend:
- Sand-matrix dacitic breccia
- Globular clast-rich dacitic breccia
- Stratified andesitic sandstone
- Dacitic pumice breccia
- Dacitic lithic-pumice breccia
- Polymictic, matrix-supported pumice-lithic breccia
- Dacite (○), Jasper (□), and laminated siltstone (△) clasts
- Dacitic breccia - hematitic patches and apparent matrix
- Quartz-hematite ironstone
- Strong hematite alteration
- Cross laminations
- Microbialite
- Break in outcrop
Figure 6.5 Outcrop maps showing the distribution of ironstone lenses and their relationships to the enclosing volcanic facies. (A) Horizon 1 ironstone in the western part of the Trooper Creek prospect ("western lenses"). Locality shown in figure 6.2. (B) Ironstone lenses at "cattle yard" (around 7743700 mN, 424600 mE).
the base, top or are enclosed by the planar laminated dacitic pumice breccia facies. The pumice breccia is non-welded, diffusely laminated and comprises pumice clasts, shards, crystals and crystal fragments (plagioclase) (Chapter 4). The crystal composition suggests a dacitic composition. Top contacts of the ironstone lenses are sharp and sometimes overlain by hematite-altered siltstone. However, bottom contacts vary from sharp to gradational with massive ironstone (Fig. 6.6A) passing through tuffaceous ironstone (Fig. 6.6B) into sericite-hematite-altered, dacitic pumice breccia or stratified andesitic breccia (Fig. 6.6C). Laminae within the altered pumice breccia can be traced into tuffaceous ironstone. Within a 45 cm wide zone beneath the ironstone lens, the pumice breccia contains abundant hematite nodules, 1-5 mm in diameter.

Tuffaceous ironstone comprises non-volcanic quartz and hematite and relic pumice fragments, shards and crystals. Quartz-hematite-rich patches and bands alternate with quartz-dominant pyroclast-rich patches and bands, accentuating the primary laminated fabric (Fig. 6.6B). One segment of horizon 1 ironstone contains microbialites (stromatolites and oncolites; Figs. 6.3 & 6.4). The stromatolites occur as domed biostromes (1-3 cm thick) that contain oncolites as well as pyroclastic components (Chapter 4). Microbialites in the ironstone are now quartz-hematite, whereas pumice clasts and shards are pervasively silica-sericite-altered and outlined by fine-grained hematite (Fig. 6.6D).

A poorly exposed, 28.7 m thick sequence of planar laminated siltstone and polymictic lithic-pumice breccia (Chapter 4) separates horizon 1 ironstone from a second thin (10-30 cm), stromatolitic and oncolitic ironstone horizon (horizon 2, Fig. 6.3). The polymictic lithic-pumice breccia is poorly sorted, matrix-supported and weakly graded. The breccia contains clasts of dacite, laminated siltstone and ironstone. Strong hematite alteration of the pumiceous matrix obscures textures in the upper part of the unit. The internal organisation is consistent with deposition as a sediment gravity flow. Ironstone clasts in the breccia were incorporated at source or collected from the substrate during transport (Chapter 4).

Horizon 2 ironstone is very poorly exposed and comprises in situ stromatolites which are built on silicified sandstone and pebble conglomerate units containing oncolites, stromatolite clasts (0.3-12 cm) and pyroclastic components (Fig. 6.4 – section B). The in situ stromatolites occur as domed bioherms, 6-7 cm high and up to 8 cm in diameter (Chapter 4). The intercolumn matrix is a mixture of unaltered feldspar crystal fragments, lithic grains and pumice fragments. The bioherms are overlain by pervasively hematite-altered siltstone. Stratified dacitic pumice breccia and sandstone beds occurs above the
Figure 6.8

Photomicrographs of ironstone units from the Trooper Creek Formation.

(A) Coalescing type 1 spherules surrounding a cuspatate hematite patch. Fans of fibres (arrow) project out from the margin of spherules into the hematite patch suggesting that both the bundles of fibres and the hematite are space filling. Plane polarised light. 95-210; Trooper Creek prospect; 7741900 mN, 426800 mE.

(B) Botryoidal texture in tuffaceous ironstone. Alternating concentric quartz- and hematite-rich laminae nucleate around a hematite core. The remainder of the photomicrograph comprises hematite patches separated by finely crystalline quartz. Plane polarised light. 95-275; Trooper Creek prospect; 7741900 mN, 426800 mE.

(C) In this sample of massive ironstone, patches of spherules (s) and hematite are separated by an apparent matrix of fine-grained recrystallised quartz (r). Occasional relic domains of spherules are identifiable in some parts of the apparent matrix (arrow) and are separated by small cuspatate hematite patches. Plane polarised light. 95-210; Trooper Creek prospect; 7741900 mN, 426800 mE.

(D) Occasional pumice clasts and shards are preserved in this sample of massive ironstone. The pumice fragments (p) are now quartz and are outlined by hematite. Hematite has completely replaced pumice fragments in some parts of the sample. Plane polarised light. 95-273; Trooper Creek prospect; 7741900 mN, 426800 mE.

(E) Pumice fragments and shards in this sample are now fine-grained quartz and are delineated by hematite. The pyroclasts have compacted and deformed around a large quartz nodule (n) which grew within the pumice breccia during replacement by quartz and hematite. Plane polarised light. 95-212; Trooper Creek prospect; 7742000 mN, 426100 mE.

(F) Photomicrograph of hematite-sericite-altered pumice breccia. The pumice clasts and ovoid vesicles (arrow) within them are outlined by hematite. The vesicles and formerly glassy walls are now sericite. The breccia also contains hematite patches. Plane polarised light. 95-204; Trooper Creek prospect; 7741900 mN, 426800 mE.

(G) This sample comes from an ironstone pod in rhyolite. Relic coalescing spherules are separated by cuspatate hematite patches. Many spherules have recrystallised to fine-grained quartz. Plane polarised light. 95-150, north of Trooper Creek prospect; 7743500 mN, 428160 mE.

(H) Photomicrograph of filaments (arrow) in a quartz-hematite vein cutting coherent dacite. 94-61; Handcuff; 7749500 mN, 418950 mE.
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Silstone. Intense pervasive hematite-alteration of the lower 2 m of the pumice breccia has destroyed most primary textures in this part of the unit (horizon 3).

Horizon 4 ironstone is exposed 216 m to the west of horizons 1-3 (Fig. 6.4 - section E), and overlies stratified andesitic breccia containing bomb fragments (globular clast-rich andesitic breccia facies; Chapter 4). The ironstone has a granular sandy texture, with dark maroon hematite-rich patches (1 to 5 mm long) separated by a quartz-rich apparent matrix. Thin-sections reveal rare relict shards. Scoria breccia beneath the ironstone lens is intensely quartz-hematite-altered within a metre of the contact, and variably sericite-hematite-altered to 130 m below the contact. Hematite alteration is most intense within 26 m of the ironstone, obscuring clast margins and generating an apparent fine-grained sandstone.

6.4.2 Iron oxide±silica alteration/rock and stratified andesitic breccia

In a 1 m wide zone beneath horizon 4 ironstone, stratified andesitic breccia contains a network of quartz-hematite domains which replace the matrix of the breccia (Fig. 6.6E). The pod encloses (Fig. 6.6E). At Highway East prospect (around 7747350 mN, 420000 mE), stratified andesitic breccia includes zones of intense pervasive hematite alteration. Although ironstone is not exposed, the intensity of hematite alteration is similar to that associated with ironstone leases at Trooper Creek prospect, and the alteration may be a lateral or vertical equivalent of an unexposed or poorly developed ironstone.

6.4.3 Ironstone associated with lavas, domes and cryptodomes

Northwest of Trooper Creek prospect (around 7743700 mN, 424600 mE), pods of ironstone (1-3 m) are enclosed in massive coherent rhyolite. The ironstone pods have sharp margins and an apparent clast-in-matrix texture in which equant, irregular and ovoid hematite-rich “grains” are enclosed in a hematite-poor, quartz-rich, apparent matrix (Fig. 6.6F). Parts of the ironstone (2 to 10 cm across) are devoid of the apparent clastic texture and comprise a mosaic of quartz and hematite. Rhyolite surrounding the ironstones is purple in colour due to pervasive hematite alteration, but away from the pods is sericite-chlorite-altered. Quartz-hematite veins cut the rhyolite along contacts with the ironstone. The rhyolite is quartz- and feldspar-phyric and consists almost entirely of coalescing spherical spherulites (60-200 μm). Partial recrystallisation to interlocking anhedral quartz and feldspar has destroyed some microstructures in many of the spherulites. Cuspal areas (10-250 μm) of hematite occur between some coalescing spherulites and along the
margins of quartz and feldspar phenocrysts in the groundmass. The hematite patches are presumably altered glass.

Field relationships and drill hole sections in the Handcuff area (around 7748600 mN, 417850 mE) show that some ironstone lenses and rhyolite units are spatially, and possibly genetically, related. At one locality (around 7743700 mN, 424600 mE) the ironstone lenses occur within massive or laminated siltstone at the margin of a rhyolite syn-sedimentary intrusion. The ironstone is massive or comprises alternating hematite-rich and quartz-rich bands (laminae?). The rhyolite is mostly sericite-chlorite-quartz-altered, but along some contacts with the ironstone lenses both the rhyolite and siltstone are hematite-altered and contain disseminated pyrite (<1 mm). In diamond drill hole HDD 012 (25-100 m), quartz-hematite veinlets cut across chlorite-sericite-quartz-altered rhyolite and peperite. Rhyolite along the margins of the veins is intensely silicified. In Handcuff diamond drill holes HDD 007 (372-410 m) and HDD 022 (262-275 m), quartz-hematite veins occur in massive coherent dacite and monomictic dacitic breccia. The breccia is non-stratified and comprises angular blocky fragments. Clasts have mostly altered to sericite and chlorite, whereas the matrix has altered to quartz-hematite. The textural characteristics and contact relationships are consistent with interpretation of the breccia as hyaloclastite at the margins of a lava (Chapter 5). At Handcuff (around 7747400 mN, 417250 mE), peperite along the top contact of a partly extrusive cryptodome provides evidence for mixing of dacite and hematite-rich siltstone and sandstone. Dacite clasts in the peperite are sericite-altered and separated by iron oxide-rich siltstone (Fig. 6.6G). Laminae in the siltstone/sandstone are absent in the peperite but are preserved 0.5-1 cm away from the contact. The laminae are purple and hematite-rich or are pale and sericite-altered. Hematite occurs only in the siltstone suggesting that it is not an alteration phase but deposited as the siltstone accumulated. The dacite mixed with the siltstone while it was still poorly consolidated (Chapter 5), destroying bedding in the siltstone at the contact.

In Highway diamond drill hole REW 803 (31.55-130 m), dacite includes irregular bifurcating seams of siltstone and jigsaw-fit aggregates of dacite clasts that are separated by siltstone. The breccia facies is gradational into massive coherent dacite. Siltstone in the peperite is locally quartz-hematite-rich and veins of quartz-hematite-carbonate dissect the core. A silicified halo up to 4 cm wide surrounds some of the veins, suggesting that they are replacements of the dacite and are not sediment which mixed with the dacite during fragmentation.
6.4.4 Ironstone lenses in siltstone

At Handcuff prospect (around 7748750 mN, 417750 mE), massive and finely laminated, silicified, siltstone contains minor thin (2-5 mm), dark red, hematite-rich laminae. North of Handcuff prospect (around 7749320 mN, 418100 mE), massive and weakly planar laminated silicified siltstone includes lenses of ironstone up to 30 m in length. The lenses form a bedding-parallel horizon in the hinge of a steeply SSW plunging syncline. The ironstones are massive or patchy, hematite-rich and contain cubic pits after pyrite. Ironstone lenses are separated by finely laminated siltstone containing small (5 to 20 cm) hematite-rich patches with diffuse margins, alternating light grey and green–grey silicified bands and discontinuous, bedding-parallel, hematite-rich bands. The hematite-rich bands consist of irregular, round or ellipsoidal hematite-rich patches which are separated by quartz-rich domains. Thin (2-3 mm) semi-continuous hematite-rich bands (7 laminae; Fig. 6.6H) are locally present.

6.4.5 Indeterminate facies relationships

In some cases, contacts between ironstone lenses and volcanic facies are not exposed. To the west of Trooper Creek (around 7743700 mN, 414600 mE), a series of massive ironstone lenses are exposed discontinuously over a strike length of 175 m (Fig. 6.5B). At Highway East prospect (around 7746600 mN, 418250 mE), ironstone lenses occur near outcrops of rhyolite but contacts are not exposed.

6.5 Ironstone mineralogy and textures

The mineral assemblages associated with massive ironstone and tuffaceous ironstone are different. Petrography combined with X-ray diffraction (Appendix E1) show that quartz, hematite and locally fine-grained (5 μm) magnetite are the principal components of massive ironstone. In addition to quartz, hematite and magnetite, various assemblages of epidote, sericite, chlorite, albite, calcite, sanidine and plagioclase feldspar are present in tuffaceous ironstone and stromatolitic ironstone.

Ironstones have textures which can be subdivided into three main groups: (1) those reflecting a volcanic input or precursor; (2) textures recording biological activity in the depositional environment; (3) non-volcanic and non-biological textures, here defined as chemical textures. Many ironstones are characterised by textures from more than one group, and by different textures from the same group. Textures in ironstones associated
with each of the principal volcanic facies can be similar, but others are unique to a given facies association or ironstone outcrop. During metamorphism and tectonic deformation, earlier mineral assemblages were recrystallised or replaced by coarse metamorphic minerals, thus destroying or modifying primary (volcanic, biological, chemical) textures (cf. Duhig et al., 1992b).

6.5.1 Volcanic textures and their altered equivalents

**Pumice and shards:** Tube pumice in the ironstones is blocky with ragged ends and smooth planar margins. Most of the glass shards have cuspatel and microvesicular pumice shapes, but a few platy shards occur. In areas of strong sericite-chlorite alteration, the pumice fragments and former vesicles within them are delineated by fine hematite. Quartz-hematite has replaced former glassy vesicle walls and vesicles are filled with quartz, hematite, sericite, or zones of hematite-quartz-sericite (e.g. 95-203). Some pumice clasts have been partially replaced by hematite and have apparent blocky shapes. In these clasts, vesicle textures have been destroyed but the clast margins are preserved (e.g. 95-203). Many phyllosilicate-altered pumice clasts are compacted and deformed around quartz-hematite-altered pumice clasts and feldspar crystals. The compacted pumice clasts define bedding-parallel compaction foliation.

**Crystals and crystal fragments:** In ironstone, feldspar (sanidine and plagioclase) crystals and crystal fragments are largely unaltered, or only weakly altered. Sericite has partially replaced feldspar in a few samples and rarely (e.g. 95-200) polycrystalline quartz has replaced a tabular mineral, which may have been feldspar. Epidote has replaced fragments of an unidentifiable ferromagnesian mineral.

6.5.2 Biological textures

**Microbialites:** At Trooper Creek prospect, microbialites in ironstone comprise stromatolites and oncolites (Chapter 4). The oncolites are elliptical to spherical and vary from 0.5 to 1.5 cm across. The component laminae are arranged around a central nucleus which is often a pyroclast or lithic fragment. Stromatolites are characterised by fine (8-20 μm), relatively flat internal laminae. Non-columnar varieties have flat-laminated, undulatory, pseudo-columnar, cumulate or columnar layered forms (Chapter 4). Branching and non-branching columnar stromatolites are also present. The laminae comprising stromatolites and oncolites are typically very thin (8-20 μm) and quartz-rich or hematite-rich.
Filamentous structures: Branching networks of filaments are preserved in a few ironstone samples and in quartz-hematite veins in dacite (Chapter 4). The filaments are randomly oriented, 10 to 200 µm long, with cylindrical cross-sections, 5-8 µm in diameter. Filament walls are delineated by hematite granules, and quartz (5-500 µm) fills the interstitial space. The filaments are similar to those described and interpreted by Dahig et al. (1992a,b) as iron oxide-secreting bacteria and/or fungi.

Other biological structures: Fragments of trilobites are preserved in one sample of stromatolitic and oncotic breccia (Chapter 4). Other microfossils include possible sponge spicules, a single gastropod and a possible brachiopod (95-200; Chapter 4). The fossils are now cryptocrystalline quartz±hematite and delineated by hematite.

6.5.3 Chemical textures

Spherules: Spherules (50-200 µm diameter) in ironstone have been classified on the basis of the relative proportions and distribution of quartz to hematite, as well as on internal structure. Eight principal morphologies are recognised and their characteristics are summarised in Table 6.2 and Figure 6.7. Type 1A and 1B (Figs. 6.7A, 6.8A) spherules are composed only of quartz or albite, whereas the other spherule types are fine intergrowths of quartz and hematite. Type 2 and 3 spherules are characterised by a radial fibrous texture which is defined by variations in the abundance of radial hematite flecks. Similar radial fibrous textures characterise the core (type 4 spherules) or rim (types 5 and 6) of spherules which are concentrically zoned. Cores are opaque and hematite-rich, whereas rims are quartz-rich with hematite present as single granules and flecks, or else arranged in trails. Neither the hematite core nor the thin quartz rim of type 7 spherules are radially fibrous. In type 8 spherules, the cores comprise concentric quartz- and hematite-rich bands and rims are radially fibrous (Fig. 6.7A).

Isolated spherules are commonly spherical. Adjacent spherules may impinge on each other, producing elongate single or branching trains and coalescing patches. Domains (20 µm to 2 mm) of hematite occur between coalescing spherules and have cuspatelike shapes (Fig. 6.7B-C). Fan- to sheaf-shaped bundles of radial quartz and hematite fibres occur around the outer margin of some domains of coalescing spherules and project into large hematite patches (Figs. 6.7D & 6.8A). In some samples, similar bundles of fibres radiate out from a line, forming axiolite-like structures (e.g. 95-179).

Metamorphism has destroyed or obscured the primary textures in many spherules. Initially, the centres of spherules recrystallise to interlocking anhedral quartz and in some
### Table 6.2 Distinguishing characteristics of the eight spherule types identified in ironstone.

<table>
<thead>
<tr>
<th>Spherule</th>
<th>Size (μm)</th>
<th>Core</th>
<th>Rim</th>
</tr>
</thead>
<tbody>
<tr>
<td>Type 1A</td>
<td>50-200</td>
<td>Core &amp; rim absent. Radially fibrous quartz or albite crystals nucleate out from centre.</td>
<td>Increasing density of radially distributed hematite flecks &amp; granules towards hematite-poor margin.</td>
</tr>
<tr>
<td>Type 1B</td>
<td>50-60</td>
<td>Core &amp; rim absent. Finely crystalline quartz spherules.</td>
<td></td>
</tr>
<tr>
<td>Type 2</td>
<td>70-80</td>
<td>Absent. Sometimes centre is quartz dominant.</td>
<td></td>
</tr>
<tr>
<td>Type 3</td>
<td>50</td>
<td>Granules of hematite form a ring in from margin.</td>
<td>Trails of hematite extend in from jagged spherule margin.</td>
</tr>
<tr>
<td>Type 4</td>
<td>100</td>
<td>Radially fibrous hematite flecks &amp; granules. Core and rim have similar width.</td>
<td>Quartz rim with minor hematite granules.</td>
</tr>
<tr>
<td>Type 5A</td>
<td>50-60</td>
<td>Hematite nucleus thinner than rim.</td>
<td>Radial quartz and hematite flecks and granules.</td>
</tr>
<tr>
<td>Type 5B</td>
<td>50-60</td>
<td>Opaque hematite nucleus. Core wider than rim.</td>
<td>Radial hematite flecks and granules between hematite nucleus and thin quartz rim.</td>
</tr>
<tr>
<td>Type 6</td>
<td>50-100</td>
<td>Hematite nucleus wider than rim.</td>
<td>Radial extinction of component anhedral quartz.</td>
</tr>
<tr>
<td>Type 7</td>
<td>100-250</td>
<td>Spherical, ellipsoidal to cuspatate. Granular hematite ± minor quartz. Wider than rim (to 200 μm).</td>
<td>Quartz with disseminated hematite.</td>
</tr>
<tr>
<td>Type 8</td>
<td>50-100</td>
<td>Concentrically arranged hematite-poor bands and radially fibrous hematite-rich bands. Core narrower than rim.</td>
<td>Quartz with minor radial hematite flecks and granules.</td>
</tr>
</tbody>
</table>

---

**Figure 6.7 (A)** Cartoon of the principal spherule types (1-8) identified in ironstone. Not to scale. **(B-D)** Important textures in ironstone. (B) Coalescing, variably recrystallised type 1 spherules separated by cuspatate patches of hematite. (C) Coalescing type 7 spherules. (D) Fans of quartz-hematite fibres projecting out from the margins of type 1 spherules into a patch of hematite.
cases, radial segments of the spherules become coarser grained. Further recrystallisation produces a mosaic of interlocking coarse-grained, anhedral quartz without fibrous textures. The round margins of recrystallised spherules are preserved along contacts with cusptate hematite patches.

**Botryoidal texture:** Botryoidal texture comprises alternating concentric dark hematite-rich bands and light, hematite-poor, quartz-rich bands (Fig. 6.8B). Bands (<10 μm wide) are semi-continuous or discontinuous. In some samples, bands nucleate around single tube pumice clasts (e.g. 95-204), feldspar crystal fragments (e.g. 95-203) or hematite patches. The hematite patches are whole or form jigsaw-fit aggregates which are separated by cryptocrystalline quartz. Bands comprising botryoidal structures are smooth or have bulbous colloform-like shapes.

**Hematite patches and granules:** Hematite is mostly present as equant blocky to irregular patches surrounded by quartz (Fig. 6.8C). Wedge-shaped quartz-filled fractures extend in from the margins of the patches and are similar to cracks described in chert and attributed to the dewatering of silica gels (e.g. Schübel and Simonson, 1990). The hematite patches form jigsaw-fit aggregates or have elongate ragged shapes and are connected along mutual boundaries by thin stems. In some samples, the hematite patches contain small (5 μm), acicular, magnetite crystals. A few hematite patches are replacements of single pumice clasts. However, most iron oxide patches are not replacements of single pyroclasts and controls on their distribution are not obvious. Cusptate hematite patches between spherules appear to be filling pore space or replacing a precursor space-filling mineral.

Chlorite patches within the ironstone lenses contain small (5 μm) round hematite “globules” (e.g. 95-273). In detail, the globules comprise smaller aggregates of very fine-grained “granular” hematite. Granular hematite is also present as fine disseminations in many quartz and hematite patches and in hematite-rich bands of botryoidal structures. Hematite delineating the margins and vesicles of pyroclasts is also granular.

**Quartz textures:** Quartz is the principal component of the ironstones and is texturally diverse (Duhig et al., 1992b). The following quartz polymorphs are adopted from Duhig et al. (1992b). Megaquartz is clear, equant to tabular, and greater than 200 μm across. Chalcedony is optically fibrous quartz which forms radial fibres, mostly around 100 μm long, in spherules or fan- to sheaf-shaped bundles. Microcrystalline quartz is equant, 1 to 100 μm across, and displays undulatory extinction and pinpoint birefringence. Cryptocrystalline quartz appears isotropic under cross polars and is less than 1 μm in grain size. Microcrystalline quartz and cryptocrystalline quartz are often yellow-brown to pink in colour, possibly due to a very fine dusting of hematite.
6.6 Ironstone textures and volcano-sedimentary facies

6.6.1 Ironstone associated with dacitic pumice breccia

Cross-sections through ironstone associated with dacite pumice breccia show that ironstone textures vary passing from massive ironstone, through pyroclast-rich ironstone, down into hematite-altered pumice breccia. Massive ironstone is dominated by domains of coalescing spherules (type 1B, 2, 5B, 6, 7 and/or 8) and cuspatelike hematite patches (e.g. 95-179, 95-180). Recrystallisation of the spherules to interlocking medium grained (25 μm) or coarse-grained (400 μm) quartz has destroyed microstructures in many spherules and generates patches with only rare spherules (Fig. 6.8C). In a few samples (e.g. 95-179), quartz and hematite fibres are arranged in axiolite-like structures, separated by finely crystalline (5 μm) quartz (95-179).

In massive ironstone with relic volcanic particles (e.g. 95-273), pyroclasts are now composed of microcrystalline quartz, small type 1A spherules (e.g. 95-316) and/or delineated by hematite (Fig. 6.8D). In some samples, spherules cut across the margins of pumice clasts, shards and the vesicles within them. Microcrystalline quartz and megaquartz separate small irregular domains (200-700 μm) of coalescing spherules, hematite patches and rare botryoidal structures. The result in hand specimen is a fine granular texture. In one sample of horizon 1 ironstone (95-210), relic domains (500 μm) of calcite occur between patches of hematite and coalescing spherules. Light and dark bands (5-15 μm) in the calcite conform to contacts with the patches. Elemental maps from microprobe analysis show that zoning in the calcite is due to trains of very fine-grained (1-2 μm) anhedral quartz. Along the margins of the patches, recrystallisation to quartz-free calcite destroys original zoning in the calcite.

In tuffaceous ironstone, quartz-rich bands comprise coalescing type 2 spherules, fine (3-5 μm) quartz±hematite-altered pumice clasts, subordinate hematite patches and rare botryoidal structures. One sample (e.g. 95-214) contains branching networks of hematite filaments up to 200 μm long. The filaments project out from the margins of hematite patches and are enclosed by quartz. Single iron oxide globules at the ends of fibres may be cross-sections through filaments. Hematite-rich bands comprise single and interconnected hematite patches separated by finely crystalline (3-5 μm) quartz. Quartz nodules are also present (Fig 6.8E). The nodules have bulbous margins and comprise coarse quartz (25 μm), coalescing type 1 spherules and cuspatelike hematite patches (25 μm across). Pumice fragments and shards are deformed around the nodules, suggesting that the nodules formed prior to compaction or else, grew within the pumice breccia and displaced the pyroclasts.
Figure 6.8

Photomicrographs of ironstone units from the Trooper Creek Formation.

(A) Coalescing type 1 spherules surrounding a cuspatate hematite patch. Fans of fibres (arrow) project out from the margin of spherules into the hematite patch suggesting that both the bundles of fibres and the hematite are space filling. Plane polarised light. 95-210; Trooper Creek prospect; 7741900 mN, 426800 mE.

(B) Botryoidal texture in tuffaceous ironstone. Alternating concentric quartz- and hematite-rich laminae nucleate around a hematite core. The remainder of the photomicrograph comprises hematite patches separated by finely crystalline quartz. Plane polarised light. 95-275; Trooper Creek prospect; 7741900 mN, 426800 mE.

(C) In this sample of massive ironstone, patches of spherules (s) and hematite are separated by an apparent matrix of fine-grained recrystallised quartz (r). Occasional relic domains of spherules are identifiable in some parts of the apparent matrix (arrow) and are separated by small cuspatate hematite patches. Plane polarised light. 95-210; Trooper Creek prospect; 7741900 mN, 426800 mE.

(D) Occasional pumice clasts and shards are preserved in this sample of massive ironstone. The pumice fragments (p) are now quartz and are outlined by hematite. Hematite has completely replaced pumice fragments in some parts of the sample. Plane polarised light. 95-273; Trooper Creek prospect; 7741900 mN, 426800 mE.

(E) Pumice fragments and shards in this sample are now fine-grained quartz and are delineated by hematite. The pyroclasts have compacted and deformed around a large quartz nodule (n) which grew within the pumice breccia during replacement by quartz and hematite. Plane polarised light. 95-212; Trooper Creek prospect; 774200 mN, 426100 mE.

(F) Photomicrograph of hematite-sericite-altered pumice breccia. The pumice clasts and ovoid vesicles (arrow) within them are outlined by hematite. The vesicles and formerly glassy walls are now sericite. The breccia also contains hematite patches. Plane polarised light. 95-204; Trooper Creek prospect; 7741900 mN, 426800 mE.

(G) This sample comes from an ironstone pod in rhyolite. Relic coalescing spherules are separated by cuspatate hematite patches. Many spherules have recrystallised to fine-grained quartz. Plane polarised light. 95-150, north of Trooper Creek prospect; 7743500 mN, 428160 mE.

(H) Photomicrograph of filaments (arrow) in a quartz-hematite vein cutting coherent dacite. 94-61; Handcuff; 7749500 mN, 418950 mE.
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Ironstone breccia beneath the ironstone lens is pervasively sericite-hematite-altered and contains hematite nodules, 2-5 mm across (Fig. 6.8F). Some nodules have bulbous margins and are massive with hematite-rich cores and thin (50-200 μm) quartz-rich margins. A few nucleate around weakly sericite-altered feldspar crystals or tube pumice fragments. The sericite-altered tube pumice fragments are deformed around the hematite nodules and define a bedding-parallel compaction foliation (S1).

6.6.2 Stromatolitic-oncolitic ironstone

Stromatolites and oncolites in ironstones are red in hand specimen due to replacement by hematite or alternating hematite-rich and quartz-rich laminae (Chapter 4). Finely crystalline quartz and subordinate disseminated granular hematite has replaced pumice fragments and shards trapped and bound within the microbialites. Pumice fragments are delineated by a hematite film or are entirely altered to hematite. Feldspar crystal fragments and epidote are also present. The remainder of the ironstone lenses consists of medium grained (20-50 μm) anhedral quartz. In one sample (95-218), polycrystalline quartz has replaced a lozenge-shaped crystal which may have been gypsum. Spherule textures are absent in stromatolitic ironstone.

6.6.3 Iron oxide-silica pods in stratified andesitic breccia

Horizon 4 ironstone at Trooper Creek is underlain by a small pod of hematite-rich, stratified andesitic breccia. Hematite occurs between the clasts and as small patches within the clasts. The margins of some clasts and the vesicles within them are delineated by a thin (2-5 μm) film of hematite. Spherules, botryoidal structures and filaments are absent.

6.6.4 Ironstone pods in rhyolite

Ironstone pods in rhyolite have a clast-in-matrix texture with hematite-rich “grains” separated by a quartz-rich apparent matrix. Most “grains” comprise coalescing spherules (types 2A, 6 or 8) and cuspatate patches of hematite (Fig. 6.8G). Other grains comprise an outer zone of coalescing round and fan- to sheaf-shaped spherules, and a hematite-rich core containing single spherules or branching trains of spherules. The remaining grains comprise equant jigsaw-fit aggregates of hematite patches separated by quartz or spongy mosaics of hematite and quartz. The apparent matrix between grains comprises fine-grained (25 μm) quartz which has locally recrystallised to coarse-grained (300 μm)
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The apparent matrix locally contains relic spherules which are outlined by cuspathe patches of hematite.

6.6.5 Quartz-hematite±carbonate veins in dacite

In carbonate-hematite-quartz veins (e.g. REW 803, 117.2 m), hematite occurs as thin (20 µm) rinds around coarse-grained polycrystalline quartz patches (150-300 µm across). The quartz patches are separated by calcite. At the margins of the veins, sub-round hematite patches are intergrown with carbonate or quartz.

In quartz-hematite veins, hematite occurs as patches with bulbous margins or as fan- and sheaf-shaped bundles of fibres which are radially arranged around a hematite grain or patch. The hematite patches and fibres cut across boundaries between interlocking anhedral quartz crystals (20-200 µm across). The quartz is weakly peppered with granular hematite. Rare filamentous structures are preserved (Fig. 6.8H).

6.6.6 Ironstone in siltstone

Many ironstone lenses in silicified siltstone comprise hematite-rich patches (40-100 µm) separated by coarsely (100 µm) crystalline quartz (e.g. 94-401). Quartz occurs as interlocking crystals with 120° triple point junctions which suggest strong recrystallisation by metamorphism. Hematite-rich domains comprise euhedral fine-grained (5 µm) hematite and subordinate quartz.

6.7 Ironstone geochemistry

Detailed studies of the mineralogical and textural characteristics of the ironstone units provide a framework for geochemical studies using whole rock major, minor, trace and rare earth element (REE) analyses. The new data are presented in Appendix E2. Samples were first crushed in a jaw crushe and then powdered in a tungsten carbide disc mill. Major and trace element analyses were determined on a Philips automated XRF spectrometer at the University of Tasmania using standard fused disc and pressed pellets techniques (Norrish and Hatton, 1969; Norrish and Chappell, 1977). Ag and REE were determined by ICP-MS at ANALABS facilities in Perth. Samples (0.2 g) were digested in aqua regia/perchloric acid/hydrofluoric acid using ANALABS method 201. The major
Ironstone samples from the Trooper Creek Formation are principally composed of SiO₂ and Fe₂O₃ (Fig. 6.9A). All other oxides constitute less than 1 wt%, with the exception of Al₂O₃ which is high (4.6-8.7%) in some samples of massive, tuffaceous and stromatolitic ironstone. Single samples of tuffaceous ironstone from Trooper Creek prospect contain elevated K₂O (1.3%, 94-327) or MgO (2.3%, 95-275). Ratios of SiO₂/Fe₂O₃ are generally high (1.8-4.5) in massive ironstone, decrease (2.3-12) in tuffaceous and stromatolitic ironstone, and increase again in hematite-altered dacite pumice breccia (3.3) and a least altered equivalent (16.8; 95-308) (Appendix E2). Increasing concentrations of TiO₂, P₂O₅ and Al₂O₃ characterise the transition from massive ironstone to the least-altered equivalent (Fig. 6.9B–C). Ti/Zr ratios for the ironstone range between 16 and 40 but are mostly between 18 and 20, similar to the least-altered pumice breccia (95-308; Appendix E2). Variation in the ratios between samples may reflect initial compositional differences in the volcaniclastic precursor. Ironstone pods in rhyolite (95-150) have Ti/Zr ratios below detection limits and are significantly different to the rhyolite. This suggests that Ti and Zr have remained immobile and the ironstone pods contain little volcanic material.

Major element patterns for ironstone associated with pumice breccia, siltstone and rhyolite are generally similar (Fig. 6.10A). However, TiO₂ is low in ironstone associated with rhyolite, and ironstone lenses in pumice breccia contain elevated MgO. Trace element abundances vary between ironstone facies. Concentrations of Ba are higher in ironstone associated with siltstone (94-25, 94-401) than ironstone from other facies associations. Cu, Pb and Zn vary considerably between ironstone facies. One sample of ironstone associated with siltstone (94-401) contains anomalous Cu, Pb, Zn and Ag. Similar concentrations of Ba and Zn also occur in hematite-altered pumice breccia beneath horizon 1 ironstone (95-274) and in horizon 3 (94-334). V, Cr and Ni show a covariance with Fe (Fig. 6.11A–C). The highest values (V 1.3 wt%; Cr 26 ppm; Ni 18 ppm) occur in hematite-altered pumice breccia (94-334). Fe shows a negative correlation with Mn, and both Fe and Mn show a covariance with Ni (Fig. 6.11D). Sc, Sb, Bi and Ag are consistently low and mostly below detection limits (Appendix E2).

The composition of ironstone units from the study area is largely different from ironstone at Thalanga. Samples of ironstones associated with pumice breccia are enriched in MgO
Figure 6.9 Major element plots for ironstone lenses in pumice breccia units at Trooper Creek prospect. Elemental concentrations vary systematically passing from hematite-altered pumice breccia, through tuffaceous ironstone, into massive ironstone. (A) Fe₂O₃ vs. SiO₂; (B) TiO₂ vs. Al₂O₃; (C) TiO₂ vs. P₂O₅.
Figure 6.10 (A) Major element plot for ironstone associated with dacite pumice breccia at Trooper Creek Prospect, ironstone hosted by siltstone (Handcuff) and ironstone pods in a rhyolite lava (north of Trooper Creek prospect). (B) Major element patterns for the average composition of massive ironstone from Trooper Creek prospect and ironstone lenses at Thalanga. Data for Thalanga from Duhig et al. (1992b).
Figure 6.11 Selected trace element vs. major element plots for ironstones from the Trooper Creek Formation in the area between Coronation homestead and Trooper Creek prospect. (A) Fe$_2$O$_3$ vs. Ni; (B) Fe$_2$O$_3$ vs. V; (C) Fe$_2$O$_3$ vs Cr; (D) Ni vs MnO.
Figure 6.12 Major element concentrations of ironstone associated with dacitic pumice breccia at Trooper Creek Prospect and (A) Noranda chemical and clastic layers; (B) Tetsusekiei clastic and chemical layers; (C) modern seafloor deposits. Data from (1) Kalogeropoulos and Scott (1989); (2) Kalogeropoulos and Scott (1983); (3) Adachi et al. (1986); (4) Binns et al. (1993); (5) Barrett et al. (1988).
and K₂O and depleted in Na₂O, relative to ironstone from Thalanga (Fig. 6.10B). The low Na₂O content of ironstones associated with silstone distinguishes them from the ironstone lenses at Thalanga (Fig. 6.10A-B). Ironstone pods in rhyolite also have lower Na₂O values, in addition to higher CaO concentrations and lower K₂O abundances. Lithogeochemical studies of ironstone from the Kuroko and Noranda massive sulfide deposits were undertaken by Kalogeropoulos and Scott (1983, 1989). The major element patterns for ironstone from Trooper Creek prospect are similar to analyses from the “chemical layers” in ironstone from massive sulfide deposits in the Kuroko and Noranda districts (Fig. 6.12A-B). Variation in the concentration of MgO, CaO, Na₂O, K₂O and P₂O₅ in ironstone from the three districts, probably reflects differences in the composition of volcanic particles within the ironstone units. The major element patterns of ironstone from the Trooper Creek prospect and iron oxide-rich sediments from the modern seafloor are distinctly different (Fig. 6.12C). Adachi et al. (1986) demonstrated that hydrothermal Fe-Mn-Si units are readily distinguishable from non-hydrothermal precipitates by their low Al/ (Al+Fe+Mn) ratios. On a plot of Fe-Al-Mn, samples of ironstone from the Trooper Creek Formation plot within the hydrothermal field of Adachi et al. (1986) (Fig. 6.13).

![Figure 6.13 Al-Fe-Mn plot for ironstones from the Trooper Creek Formation in the study area. Samples (n=14) plot in the hydrothermal field for Fe-Mn-Si oxides defined by Adachi et al. (1986).]
6.7.2 Isocon analysis

Bulk chemical compositions of tuffaceous ironstone at Trooper Creek prospect cannot be equated with the composition of their precursor volcaniclastic units, because of major addition and/or depletion of Si, Fe and other elements. To determine the overall chemical changes with increasing alteration, a least-altered dacitic pumice breccia (95-308) from the stratigraphic section was compared with the average of each of its altered equivalents. The isocon method (Grant, 1986; Huston, 1988, 1993) provides a graphical method to determine compositional changes during alteration. In these calculations (Appendix E3), elements are ordered so that those usually considered immobile are evenly dispersed, and each element is assigned an integer (ni) in ascending order (e.g. SiO₂ = 1, Fe₂O₃ = 2). The scaled concentration of a particular element (Ci) can be calculated from:

\[ C_i = ni \cdot \frac{C_{ai}}{C_{ui}} \]  

where \( C_{ai} \) is the concentration of an element in the altered rock for the corresponding integer value \( ni \), and \( C_{ui} \) is the concentration of an element in the unaltered equivalent for the corresponding integer value \( ni \). The scaled values are plotted on the Y-axis against the corresponding integer which, for ease of interpretation, is replaced by the corresponding element symbol.

Once plotted the isocon can be determined by fitting a line through the immobile elements (Al, Ti, Zr, Nb and Y) and the isocon slope (m) calculated. The net mass change (\( M^\Delta \)) relative to the least-altered equivalent can be calculated using:

\[ M^\Delta (%) = 100 \left( \frac{1}{m} - 1 \right) \]  

Relative mass change for each element can be estimated from the isocon. Elements that gained mass through alteration plot above the isocon and those that lost mass plot below the isocon. The relative mass change for particular elements can be calculated by the relationship:

\[ C^\Delta_i (%) = 100 \left[ \frac{C_{ai}}{(mCUi) - 1} \right] \]  

where \( C^\Delta_i \) (%) is the relative mass change for the element corresponding to the integer \( ni \).

The absolute mass change has not been calculated because the present density of the least-altered equivalent is different from that of the unaltered and uncompacted equivalent. The
interpretation of the isocon has several limitations (Huston, 1988, 1993): (1) the relative mass changes are calculated using a least-altered equivalent which was sourced from a different eruption than that hosting the ironstone, and so may be geochemically different; (2) the altered and unaltered lithofacies are resedimented volcaniclastic units and so probably have compositional variations unrelated to alteration; (3) the primary composition of the least-altered pumice breccia was probably modified during alteration and compaction. Samples of ironstone and the least-altered equivalent have similar Ti/Zr ratios, suggesting that they were originally geochemically similar. To minimise the effect of the second limiting factor, the average of multiple samples has been used in the calculations where possible (e.g. Huston, 1988, 1993).

The relative mass change of elements in massive ironstone (95-210, 95-276), tuffaceous ironstone (95-275) and hematite-altered pumice breccia (95-274) have been calculated for horizon 1 (Fig. 6.4 - section B). Massive ironstone is enriched in Si, Fe, Cr, Cu, Pb and Mn relative to the least-altered equivalent (Fig. 6.14A). Zn, Mg, Ca, P and Ba are also elevated. Sr, Rb, Na and K are depleted, while the high field strength elements reflect original magmatic concentrations. The tuffaceous ironstone sample has a similar geochemical pattern to massive ironstone. However, the relative gains in Si, Cr, Cu, Pb and Mn are less substantial (Fig. 6.14B). Tuffaceous ironstone is enriched in Mg compared to massive ironstone. The concentration of Sr, Rb, Ca and K are less than those in the least-altered equivalent and similar to samples of massive ironstone. Hematite-altered pumice breccia is enriched in Fe, Zn and Pb but much less so than overlying ironstone facies (Fig. 6.15A). Cu and Ba are also depleted relative to the least-altered equivalent. The net mass changes (M^A) relative to the least-altered equivalent is 1044.2% for massive ironstone, 199% for tuffaceous ironstone and -15% for hematite-altered pumice breccia.

Samples of tuffaceous ironstone (horizon 1) from section B and localities further to the west (Fig. 6.5A) display similar relative mass changes. However, the western lenses are less enriched in Cu and Pb and depleted in Zn, Mn and Mg compared to samples from section B (Fig. 6.15B). Samples ofstromatolitic ironstone display elemental patterns which are similar to massive and tuffaceous ironstone (Fig. 6.16A). However, stromatolitic ironstone is depleted in Zn and Mg. The concentrations of Rb, Na and K are low compared to the least-altered equivalent. The net mass change during alteration of stromatolitic ironstone is equal to a 270% addition. Because the primary mineralogy of stromatolitic ironstone is different from the least-altered equivalent, direct comparison of analyses is not possible. Massive ironstone from horizon 4 is enriched in Mn and depleted in Nb and Zn, relative to both the least-altered equivalent (Fig. 6.16B) and massive ironstone from horizon 1. A net mass gain of 5570% is indicated.
Figure 6.14 Isocon diagrams illustrating scaled concentrations of particular elements (1) and relative mass change for each element (2) in (A) massive ironstone and (B) tuffaceous ironstone from Trooper Creek prospect.
Figure 6.15 Isocon diagrams illustrating scaled concentrations of particular elements (1) and relative mass change for each element (2) in (A) hematite-altered pumice breccia and (B) massive ironstone from the western lenses at Trooper Creek prospect.
Figure 6.16 Isocon diagrams illustrating scaled concentrations of particular elements (1) and relative mass change for each element (2) in (A) stromatolitic ironstone from horizon 2 at Trooper Creek prospect and (B) massive ironstone from horizon 4.
Interpretation

The low concentration of elements other than SiO₂ and Fe₂O₃ reflects the dominance of quartz and hematite in the ironstone lenses. The decrease in the concentration of TiO₂, P₂O₅, and Al₂O₃ in the transition from massive ironstone, through tuffaceous ironstone, into hematite-altered pumice breccia, records the depletion of these elements during replacement/infiltration of the volcanic component by addition of quartz and hematite. The covariance between Fe and Ni, and Cr and V is tentatively interpreted to reflect adsorption by iron oxides. Ni abundances may indicate a seawater input by adsorption (Davidson et al., 1996). Isocon analysis identifies those elements which have been added by the hydrothermal fluid (and seawater) during alteration of the pumice breccia. These include Si, Fe, Cr, Cu, Zn, Pb, Mn, Mg and P. Cr may have been sourced from the underlying stratified andesitic breccia facies. Alteration of pyroclasts in the pumice breccia resulted in loss of Sr, Rb, Na and K.

6.7.3 Rare earth elements

Chondrite normalised (Boynton, 1984) rare earth element concentrations for ironstone samples from the study area are generally high (Figs. 6.17 and 6.18). Samples of ironstone with incomplete REE patterns have high LREE concentrations and HREE concentrations below detection limits (Fig. 6.18B-C). These include analyses of ironstone hosted by pumice breccia at Highway East prospect (94-246) and massive ironstone from east of Trooper Creek (95-130).

Samples of ironstone associated with pumice breccia are characterised by light rare earth element (LREE) enrichment and negative Eu anomalies (Fig. 6.17A). Single samples of massive ironstone (95-316), tuffaceous ironstone (95-212) and hematite-altered pumice breccia also display negative Ce anomalies (Fig. 6.17B-C). The transition from hematite-altered pumice breccia into massive ironstone is marked by a decrease in the concentration of both LREE and HREE. The different ironstone types generally have similar REE patterns and slopes (Fig. 6.17A). Massive ironstone with the smallest component of volcanic detritus (95-316) has the lowest concentration of REE, greatest negative Ce anomaly and Eu is below detection limit. Overall the REE patterns of the ironstones that are associated with pumice breccia are similar to rhyolitic and dacitic lavas and intrusions from the Trooper Creek Formation (Fig. 6.18A). The rhyolitic lavas have relatively flat to slightly LREE enriched patterns with shallower slopes than ironstone associated with pumice breccia.
Figure 6.17 Rare earth element plots of ironstone from Trooper Creek prospect normalized to the chondritic values of Boynton (1984). (A) Massive, tuffaceous and stromatolitic ironstone and hematite-altered pumice breccia from section B and horizon 4. (B) Massive ironstone from the section B (horizon 1) and horizon 4. (C) Tuffaceous ironstone from the section B (horizon 1) and the western lenses.
Figure 6.18 REE patterns normalized to chondritic values of Boynton (1984). (A) Comparison of ironstones from Trooper Creek prospect and Thalanga ore horizon (data from Duhig et al., 1992) with rhyolite and dacite from Highway-Reward (data from Stolz, 1991). (B) Ironstone associated with siltstone (Handcuff), forming pods in rhyolite and lenses in dacitic pumice breccia (Trooper Creek prospect). Ironstone from "cattleyard" is also plotted. (C) Ironstone associated with siltstone.
Ironstone from pods enclosed in rhyolite is enriched in LREE, depleted in HREE, and display negative Ce and Eu anomalies (Fig. 6.18B). The REE pattern of this sample (95-150) mirrors those of massive and tuffaceous ironstone. Massive ironstone associated with siltstone displays a variety of different patterns. Sample 94-25 shows slight LREE enrichment and a negative Ce anomaly. Two samples (94-18, 94-401) have positive Eu anomalies, similar to ironstone from Thalanga ore horizon and distinct from other ironstones in the study area. Sample 94-18 displays a U-shaped pattern with slight LREE and HREE enrichment and a weak positive Eu anomaly (Fig. 6.18C). Ironstone dominated by quartz (94-401) is characterised by a relatively flat LREE pattern, positive Eu anomaly and HREE depletion.

Interpretation of rare earth elements

Oxyhydroxide deposits with light REE enriched patterns and positive Eu anomalies are currently forming at hydrothermal vents on the modern seafloor (e.g. Michard et al., 1983; Michard and Albarède, 1986). The REE patterns of these deposits contrast with the surrounding sea water which is characterised by HREE enrichment, negative Ce anomalies, and much lower REE abundances (e.g. Alt, 1988). Some ancient equivalents (e.g. Duhig et al., 1992b) are also characterised by positive Eu anomalies and are interpreted as hydrothermal in origin. Other ironstone units display REE patterns which reflect input from both seawater (negative Ce anomalies) and hydrothermal fluids (positive Eu anomalies) (e.g. Graf, 1977; German et al., 1990; Barrett et al., 1990). Barrett et al. (1990) documented proximal to distal geochemical variations in hydrothermal precipitates on the Southern Explorer Ridge. They noted that iron oxyhydroxide deposits at the vent have hydrothermal signatures, whereas distal precipitates have seawater-modified Eu patterns. Hydrothermal particles scavenge REE from seawater as they are dispersed in the plume and following deposition (e.g. German et al., 1990; Olivarez and Owen, 1991). Post-depositional scavenging is limited by burial and so is less important if sedimentation rates are high (Olivarez and Owen, 1991).

Hydrothermal precipitates with positive Eu anomalies indicate that the hydrothermal fluids were enriched in divalent Eu and either, hotter than 250 °C, or reduced (Sverjensky, 1984; Lottermoser, 1989; Michard, 1989). Eu is sourced to the hydrothermal fluids during alteration of feldspar in the source rock. Hornblende, sphene, calcite and pyroxene are also sources of Eu (Rollinson, 1993; Barrett et al., 1990).

In the study area, samples of ironstone associated with pumice breccia typically have pronounced negative Eu anomalies. In tuffaceous ironstone, feldspar crystals are
unaltered suggesting that the negative Eu anomalies are not related to the breakdown of feldspar. Feldspar crystals are rare in massive ironstone. However, samples of massive ironstone have weak negative Eu anomalies suggesting that Eu was relatively immobile (cf. McLennan and Taylor, 1979; Campbell et al., 1984; Schandl and Gorton, 1991). The negative Eu anomalies and hematite-rich composition of the ironstone lenses, suggest that the units deposited from oxidised and possibly low temperature (<100 °C) hydrothermal fluids in equilibrium with feldspar. Samples showing LREE enrichment and negative Eu and Ce anomalies reflect input from both the hydrothermal fluid and Cambrian seawater.

REE concentrations decrease passing from hematite-altered pumice breccia, through tuffaceous ironstone, into massive ironstone. However, the REE patterns for the different iron oxide±silica rocks are similar. The transition is interpreted to reflect the dilution of primary REE in a precursor pumice breccia by precipitation of quartz and other REE-poor minerals from the hydrothermal fluid. In hematite-altered pumice breccia, diageneric compaction may have resulted in a relative increase in the concentration of REE relative to massive ironstone and tuffaceous ironstone, which altered early and remained uncompacted. Post-depositional scavenging of seawater-derived REE may have been important.

Ironstone associated with siltstone shows a range of REE patterns. Sample 94-25 displays slight LREE enrichment and a negative Ce and Eu anomaly, suggesting both seawater and hydrothermal REE sources. The remaining samples (94-18, 94-401) display positive Eu anomalies suggesting deposition from relatively hot (>250 °C) and/or reduced hydrothermal fluids. The weak enrichment in HREE in sample 94-18 may reflect a seawater input.

6.8 Discussion

6.8.1 Mechanisms and conditions of quartz-hematite deposition

In most ironstone units from the study area, textures (e.g. syneresis cracks and spherules; Fournier, 1985) and negative Eu anomalies, suggest precipitation of amorphous silica and hematite (or oxyhydroxide) from oxidised and low temperature solutions (<100°C; cf, Davidson et al., 1996). The exception is two samples (94-18, 94-401) which have positive Eu anomalies and sometimes contain pyrite and high Ba (94-401). These ironstone units deposited from relatively “hot” and/or reduced hydrothermal fluids. Along the mid-ocean spreading centres, amorphous silica is precipitated from seafloor hydrothermal fluids at low temperatures (15-100°C). The fluids evolve through mixing of
relatively high temperature (175-350°C), silica-saturated hydrothermal fluids with cold seawater (Janecky and Seyfried, 1984; Tivey and Delaney, 1986; Alt et al., 1987; Hannington and Scott, 1988). At the Galapagos Rift (86°W), amorphous silica chimneys are forming at low temperatures (32-40°C), whereas amorphous silica intergrown with sulfides precipitates from fluids at 100°C (Herzig et al., 1988). Oxygen isotope studies suggest that tetsusekietie ores of Kuroko deposits also precipitated at less than 100°C (Tsutumi and Ohmoto, 1983). Most studies conclude that precipitation of amorphous silica proceeds by conductive cooling (e.g. Tivey and Delaney, 1986) or conductive cooling and mixing with seawater (e.g. Hannington and Scott, 1988; Herzig et al., 1988; Janecky and Seyfried, 1984). Hekinian and Fouquet (1985), examining massive sulfide fields on the East Pacific Rise near 13°N, suggested silica deposition during the waning stage of hydrothermal activity solely by mixing of low temperature (<100 °C), silica-rich solutions with seawater.

Silica solubility increases as solutions become more alkaline and in part, with increasing temperature and pressure (Williams and Crerar, 1985). In submarine environments, a sharp drop in pH or temperature can cause the hydrothermal fluid to become rapidly supersaturated with respect to silica (Fournier, 1985; Williams and Crerar, 1985). Theoretical modeling by Janecky and Seyfield (1984), suggests that quartz can deposit directly from high temperature (350°C) fluids by mixing with cool seawater. However, this is inconsistent with observations from the seafloor. The sluggish nucleation kinetics of quartz below 200°C inhibits precipitation from rapidly cooled fluids (Rimstidt and Barnes, 1980; Janecky and Seyfried, 1984). Alt et al. (1987) identified significant quartz in hydrothermal oxide deposits on seamounts near 21°N, East Pacific Rise. However, quartz there is a product of recrystallisation of amorphous silica in response to reheating within the growing sulfide deposit.

The formation of amorphous silica is facilitated by the presence of Fe–O–OH ions, which induce polymerisation of mutually repelling silica colloids or substrates in acid to neutral pH solutions (Williams and Crerar, 1985). Adsorption of silica by clays, iron oxides and other impurities can also remove silica from solution (Williams and Parks, 1985). In ironstones associated with pumice breccia, the high surface area of reactive glassy pumice fragments and shards may have provided nuclei for the precipitation of silica (and hematite) from the hydrothermal fluid (cf. Ohmoto et al., 1983). Early diagenetic minerals (e.g. clay, zeolite, hematite) were probably also present and promoted amorphous silica precipitation. In ironstone hosted by rhyolite and dacite, reaction of silica saturated fluids with glass may have promoted precipitation of hematite and silica. Ironstones with negative Ce anomalies suggest that precipitation of silica did not occur solely in response
to conductive cooling and that in addition, mixing with seawater or pore water was important.

Amorphous silica is relatively unstable and so readily transforms to more stable silica polymorphs (cristobalite, opal-CT, chalcedony or quartz). The time required for the transformations decreases as temperature increases and in the presence of solutions with high pH, high salinity or containing dissolved Mg (Fournier, 1985). In the Trooper Creek ironstones, preservation of pumice clasts and shards with delicate margins and vesicles implies replacement prior to significant compaction. This also suggests that the conversion of amorphous silica to quartz probably began early during the period of hydrothermal activity (cf. Renaut and Owen, 1988).

In many Trooper Creek Formation ironstones, the amorphous silica crystallised to form spherules (cf. Oehler 1976a,b). Coatings of iron minerals on the surface of some spherule types are interpreted to be due to surface adsorption effects or to exclusion of iron-rich impurities during spherule growth. Concentric bands of iron oxide were entrapped during growth. The bands provide evidence for the co-precipitation of iron oxide and silica or imply that hematite deposited first. In some cases, hematite acted as crystallisation nucleation centres. Similar textures have been recorded from some Precambrian ironstones (e.g. Oehler, 1976b). Oehler (1976b) demonstrated experimentally that chalcedonic spherules crystallise from a silica gel until the concentration of dissolved silica decreases below a critical value, following which, euhedral quartz crystallises from the remaining silica-depleted solution. In the Trooper Creek Formation ironstones, interlocking anhedral quartz which fills pore space in some ironstone samples may have deposited by the same mechanism. Botryoidal structures and nodules are interpreted to have grown at scattered nucleation sites (commonly feldspar crystals) within the amorphous silica gel.

Spherules originally comprised chalcedony (cf. Oehler, 1976b), but many progressively transformed to more stable quartz. Recrystallisation probably began early, but was most texturally destructive during metamorphism (cf. Duhig et al., 1992). Recrystallisation has destroyed or obscured radially fibrous textures of spherules, converting many to interlocking mosaics of anhedral quartz. In some cases, recrystallisation has not been pervasive and regions of quartz spherules grade into domains of interlocking anhedral quartz grains, and some spherules have only been partially converted to anhedral quartz.

At modern seafloor hydrothermal centres, Fe precipitates as iron oxyhydroxides (e.g. Hekinian and Fouquet, 1985; Alt et al., 1987; Holm, 1987). Hematite characterises some depositional settings (e.g. Soufrière volcano, Sigurdsson, 1977) and may have been the
primary oxide phase in the Trooper Creek ironstones. Alternatively, iron oxyhydroxide (e.g. goethite) may have formed an unstable precursor which later converted to more stable hematite. REE analysis and textures suggest that the Fe largely deposited at low temperatures and is consistent with studies of young deposits (e.g. Holm, 1987). At Santorini, Fe-oxidising bacteria in mud suggest that the iron-hydroxide deposited at 12-30°C (Holm, 1987). In the Trooper Creek Formation, ironstone lenses with filamentous algae/bacteria probably deposited at similar temperatures (cf. Dahig et al., 1992b). Syneresis cracks suggest that the oxide patches were more hydrated than the enclosing amorphous silica. In ironstone hosted by pumice breccia beds, fine-grained hematite delineates vesicle walls and pyroclast margin. As both vesicles and pumice/shard walls are now phyllosilicate or quartz, an early phase of hematite alteration is probably recorded. Cuspate patches of hematite between coalescing spherules probably deposited during and/or after growth of the spherules.

6.8.2 Evidence for sub-seafloor replacement

In the Trooper Creek Formation, relic pyroclasts suggest that some ironstone lenses formed by sub-seafloor replacement of pumice breccia units, or sedimentation was synchronous with hydrothermal activity and precipitation of iron and silica occurred at, above, and below the seafloor during ironstone deposition. Most of the iron and silica precipitated by sub-seafloor replacement and infiltration of pumice breccia because: (1) the ironstone lenses are hosted by rapidly or mass flow emplaced units (cf. Allen, 1994; chapter 7); (2) laminae within the pumice breccia units can be traced into the ironstone lenses; (3) there are replacement fronts passing from massive ironstone with rare pyroclasts, through tuffaceous ironstone, into hematite-altered pumice breccia with hematite-silica nodules; and (4) strong hematite alteration in siltstone beds which overlie some ironstone lenses, suggests that the hydrothermal activity continued during (and probably after) deposition of the siltstone units.

The distance below the seafloor at which infiltration and replacement occurred is difficult to interpret, but was not excessively deep because: (1) clasts of ironstone were incorporated into overlying subaqueous mass-flow deposits; and (2) the pumice breccia has been pervasively altered without producing veins. After lithification, alteration of pumiceous deposits is generally fracture controlled (McPhie et al., 1993).

The transition in mineralogy passing from massive ironstone (quartz-hematite), through tuffaceous ironstone (hematite>quartz), into pumice breccia (sericite>hematite), may record a decreasing involvement of hydrothermal fluids, in favour of diagenetic fluids,
during ironstone deposition. Alternatively, the ironstone lenses may mark the seawater mixing zone, whereas the hematite-sericite alteration may mark feeder zones within the sub-seafloor strata. Spherules are more abundant in massive ironstone compared to tuffaceous ironstone and hematite-altered pumice breccia. Massive ironstone may have deposited at, or near, the seafloor where the physical and chemical conditions promoted the growth of spherules. In tuffaceous ironstone and pumice breccia, early diagenetic compaction may have decreased the porosity and permeability and inhibited the growth of spherules in pumice breccia.

6.8.3 Role of micro-organisms in ironstone formation

Networks of filaments, similar to those described by Duhig et al. (1992a,b), are present in massive and tuffaceous ironstone, stromatolitic ironstone and in quartz-hematite veins in dacite. The filaments may be microaerophilic chemolithotrophic bacteria which were able to colonise sub-seafloor environments. The role of seafloor Fe-oxidising bacteria and algae in precipitation of iron oxide around hydrothermal vents is well documented (Trudinger and Mendelsohn, 1976; Holm 1987, 1989; Alt et al., 1987; Alt, 1988; Duhig et al., 1992b). Microaerophilic chemolithotrophic bacteria metabolise in oxygen-poor environments and are solely dependent on the immediate geochemical environment. The bacteria use energy liberated during the oxidation of Fe$^{2+}$ to Fe$^{3+}$ to secrete a mucus sheath composed of Fe-oxyhydroxides, which alter to more stable hematite during diagenesis. The recognition of filamentous structures in quartz-hematite veins supports the interpretation that microaerophilic chemolithotrophic organisms may be able to colonise sub-seafloor hydrothermal conduits (e.g. Haymon et al., 1993).

In the study area, microbial structures are rare, suggesting a limited role for micro-organisms in iron oxide deposition. Rather the microbial structures may have provided only a framework for iron oxide (or oxyhydroxide) minerals to deposit, and played no more direct role in mineral deposition than do crystal and vitric particles in the ironstone.

6.8.4 Palaeowater depths

In the study area, most ironstone lenses and pods are associated with volcanic and sedimentary facies which suggest a relatively deep (below storm wave base) submarine depositional setting (Chapters 3 and 5). The exception is at Trooper Creek prospect, where microbialites, gypsum molds, and traction current structures indicative of wave activity, imply that the ironstones accumulated above storm wave base (Chapter 4).
6.8.5 Models for ironstone emplacement

A spectrum of iron oxide deposit styles exist (Fig. 6.1), several of which depart from classic "blanket style" ironstone units associated with some massive sulfide deposits (e.g. Kalogeropolus and Scott, 1983,1989). These differences reflect variation in the chemistry of hydrothermal fluids, different volcanic facies, the facies architecture and the character of the seafloor. In host successions to VHMS deposits, several different models have been proposed for iron oxide±silica units. Ironstones can record: (1) diffuse venting of hydrothermal fluids from white smoker chimneys and during the waxing and waning stages of massive sulfide deposition (e.g. Tivey and Delaney, 1986; Alt et al., 1987; Lydon, 1988; Hannington et al., 1995); (2) precipitation of oxyhydroxides by oxidation of sulfide particles in hydrothermal plumes above black smoker chimneys (e.g. Hannington et al., 1995; Lilley et al., 1995); (3) deposition of iron oxyhydroxides and silica from brine pools (e.g. Large, 1977; Poitoff and Barnes, 1983); or (4) sub-seafloor replacement of wet unconsolidated sediment by mixing of hydrothermal fluids and sea water (Ohmoto et al., 1983).

Other iron oxide±silica units are unrelated to massive sulfide mineralisation. The hydrothermal systems responsible for these iron oxide±silica units often operate in periods of heightened volcanic activity (e.g. Goodwin, 1962; Sigurdsson, 1977). During the 1971-72 eruption of Soufrière volcano, hematite-rich mud was deposited in the crater lake. Iron in the mud was interpreted to have been derived from convecting fluids which leached Fe and other elements from a glassy lava which partially filled the crater (Sigurdsson, 1977). Heat released from syn-sedimentary intrusions can also generate local hydrothermal systems capable of depositing iron oxyhydroxides on the seafloor (e.g. Einsele et al., 1980; Einsele, 1986).

Model for the Trooper Creek Formation ironstones

In the study area, the REE patterns of the iron oxide±silica units are generally distinct from ironstone lenses which forms the ore equivalent horizon at Thalanga. The iron oxide±silica units are regionally distributed and not preferentially associated with known mineralisation. The exception is two samples (94-18, 94-401) from Handcuff. These units have REE patterns similar to ironstone from Thalanga, have anomalous Ba, Zn, Pb and Ag, and may be associated with as yet undiscovered massive sulfide mineralisation.

The evidence suggests that most of the ironstone units are unrelated to massive sulfide mineralisation. At Trooper Creek prospect, iron oxide±silica deposition consistently followed the emplacement of syn-eruptive pumiceous deposits. The iron oxide±silica
units may record the establishment of shallow water hydrothermal systems following explosive eruptions at a nearby (unexposed/eroded) subaerial or shallow marine volcanic centre. Alternatively, they may be related to circulation of fluids around lavas/intrusions which occur within the succession. Other ironstone units are hosted by lava- and intrusion-dominated volcanic centres. These iron oxide±silica units occur along contacts with the enclosing strata, form the matrix in peperite, or occur as pods in the lavas or intrusions. This suggests that the iron oxide±silica units deposited during, after and prior to, emplacement of the associated lavas and intrusions. The spatial and temporal relationships of the iron oxide±silica units with the volcanic centres suggests that the hydrothermal systems may have been genetically related to the magmatism associated with emplacement of the lavas and intrusions. Convecting pore waters probably leached Fe and Si from the enclosing glassy volcanic succession and reprecipitated the Fe and Si by conductive cooling and mixing with seawater in the enclosing volcanic succession. Experimental studies show that volcanic glass readily contributes silica and iron to circulating hydrothermal solutions and feldspar is a source of Eu (e.g. Fournier, 1985). The absence of positive Eu anomalies in most iron oxide±silica units suggests that feldspar destruction was not important during leaching of the volcanic succession and/or that the host rocks were feldspar-poor. Most units in the Trooper Creek Formation contain feldspar, suggesting that the former was more important. Samples of ironstone with positive Eu anomalies were sourced from relatively hot fluids which leached feldspar-bearing volcanic rocks.

Many ironstone units are sub-seafloor replacements of pumice breccia and sandstone. Diffuse fluids which passed through the pumice breccia units, deposited amorphous Fe-Si-O-OH gel in pore space within the units and replaced glassy pyroclasts. It is possible that initial precipitation of hydrothermal minerals occurred at or near the seafloor, where sediments were more porous and contained greater seawater. An advancing Fe-Si-O-OH front progressively moved downward and laterally through the less altered parts of the pumice breccia units. The transformation of the amorphous gel into more stable silica and oxide minerals, probably began during replacement of the pumice breccia and may have also proceeded down from the seafloor. The quartz and hematite minerals may have formed a barrier to the ascending hydrothermal fluids. The low-permeability cap might have restricted circulation of mineralising fluids within the developing lenses, caused an upward migration of isotherms in the lenses, and so promoted the transformation of amorphous silica into more stable polymorphs (e.g. Oehler, 1976a,b). Siltstone which occurs above some iron oxide-silica units also acted as an aquitard. Hematite alteration extends into the siltstone, suggesting that fluids were able to penetrate these units. The thickness of some ironstone lenses suggests that the hydrothermal systems which deposited the iron and silica were long lived, that precipitation of minerals was very
efficient, and/or that large volumes of hydrothermal fluids passed through the pumice breccia units. Intense silica-hematite alteration of units beneath the ironstone lenses is interpreted to have deposited from diffuse fluids which passed through these units. There is evidence for topography on the original upper surface of the pumice breccia hosting horizon 1 ironstone. Rather than occupying the topographic lows, the ironstone lenses often replaced pumice breccia that mantled palaeotopographic highs, suggesting that the lenses did not accumulate from brine pools.

Selective replacement of permeable horizons by hydrothermal fluids may continue uninterrupted as younger deposits are emplaced from concurrent volcanic activity. The preservation potential of ironstones formed in this way may be higher than for exhalative deposits. On the seafloor, currents can disrupt and disperse exhaling hydrothermal fluids, and sediment gravity flows may erode poorly consolidated gels. The depth at which infiltration and replacement take place is poorly constrained. The upper few tens of metres may be the favoured position for replacement as sediments are wet and unconsolidated in this zone, and at greater depths become progressively more compacted (e.g. Einsele, 1986). The shapes, dimensions and viability of hydrothermal circulation, important for iron oxide-silica deposition, will be strongly dependent on the volcanic facies and their properties. This study suggests that distinct or genetically related iron oxide-silica deposits that involve different facies may contrast markedly in texture, structure and geochemistry.

6.9 Significance of ironstones to mineral exploration

Iron oxide-silica units occur throughout the Seventy Mile Range Group. Type 1 ironstones are characterised by positive Eu anomalies and anomalous Zn, Pb, Ba, Ag and Au. They are sourced from relatively high temperature, acidic and/or reduced hydrothermal fluids (cf. Duhig et al., 1992). Type 2 ironstones are characterised by negative Eu and Ce anomalies and contain little Ba, Pb or Zn (cf. Davidson et al., 1996). These ironstones deposit from oxidised, low temperature fluids which are in equilibrium with feldspar (alkaline/neutral). Type 1 ironstones are spatially associated with massive sulfide mineralisation (Fig. 6.19A). Type 2 ironstones are interpreted to deposit from low temperature fluids circulating around lavas and intrusions and within proximity to subaqueous explosive volcanoes (Fig. 6.19B).

As some styles of massive sulfide mineralisation are hosted in the proximal facies association submarine volcanic centres, type 2 ironstones are likely to be represented in these environments. This is the case at Highway-Reward. The Highway-Reward VHMS
Figure 6.19 Models for the formation of ironstone units in the Seventy Mile Range Group. (A) Type 1 ironstones are spatially associated with massive sulfide mineralisation and are sourced from reduced, high temperature fluids. They are vectors to mineralisation (cf. Duhig et al., 1992). (B) Barren (type 2) ironstones are not related to massive sulfide mineralisation. Type 2 ironstones deposit from oxidised, low temperature fluids circulating around volcanic centres. Type 2 ironstones can occur in the immediate host succession to massive sulfide mineralisation.
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deposit is hosted by a syn-sedimentary intrusion-dominated volcanic centre. Iron oxide±silica units and alteration occur in the host succession to mineralisation and are interpreted to have deposited from type 2 fluids related to the emplacement of the lavas and intrusions. At Handcuff, both type 1 and type 2 ironstone units are present. Type 2 ironstones are interpreted to have been emplaced from fluids circulating around lavas and intrusions which occur in the host succession. Type 1 ironstone units represent clear targets for further exploration.

6.10 Summary

Careful elucidation of the volcanic facies and facies architecture has proven to be critical in unravelling the exploration significance and emplacement processes of the iron oxide±silica rocks in the Trooper Creek Formation. The iron oxide±silica units occur as pods in lavas, domes and cryptodomes, and as lenses in pumiceous units and silicified siltstone. Hematite-quartz±carbonate also occurs as veins in coherent rhyolite and dacite, and as an alteration of the matrix in hyaloclastite and peperite.

Many of the iron oxide±silica units are interpreted to have been deposited from oxidised, alkaline, low temperature hydrothermal fluids, circulating around lavas and intrusions and within proximity to shallow submarine volcanoes. Convecting hydrothermal fluids leached Fe, Si and other elements from glassy volcanic deposits in the host succession, and reprecipitated the Fe and Si by conductive cooling and mixing with seawater in the enclosing volcanic package. Many iron oxide±silica units are sub-seafloor replacements of pumice breccia. The iron oxide±silica units are barren, characterised by negative Eu anomalies and distinct from ironstones that flank the massive sulfide mineralisation at Thalanga. At Handcuff, ironstone lenses with positive Eu anomalies and anomalous Pb, Zn, Ba and Ag are interpreted to have been sourced from relatively high temperature, reduced and acid hydrothermal fluids. These ironstone units are untested for massive sulfide mineralisation and are targets for exploration.
Chapter 7

The sub-seafloor replacement origin of the Cambro-Ordovician Highway-Reward massive sulfide deposit, Mount Windsor Subprovince

7.1 Introduction

The literature on volcanic-hosted massive sulfide (VHMS) deposits emphasises formation on the seafloor by accumulation of sulfides precipitated from exhaling hydrothermal fluids (e.g. Large, 1992; Lydon, 1988a,b). Support for a seafloor exhalative origin initially came during studies of pristine ancient examples from uplifted and eroded subaqueous volcanic successions (Solomon, 1976; Franklin et al., 1981). The discovery of sulfide chimney mound deposits on the modern seafloor gave credence to the hydrothermal exhalative model for the genesis of VHMS deposits, but raised questions as to the viability of mineral precipitation by exhalation at the seawater-substrate interface. Sulfide accumulation from buoyant hydrothermal plumes above black smokers is a highly inefficient process. It has been estimated that more than 99% of the metal carried by hydrothermal fluids is dispersed in the water column by the plume and incorporated into distal sediments (Rona, 1984). Studies of the structure of modern sulfide chimneys and mounds (e.g. Goldfarb et al., 1983; Koski et al., 1984), and the texture and metal zonation in ancient deposits (e.g. Large, 1977), have highlighted the importance of sulfide accumulation by open space filling and replacement within the sulfide mound. There is increasing evidence to suggest that these processes may extend into the sub-seafloor environment and that some VHMS deposits form by replacement of the host volcanic and sedimentary rocks.

There are few detailed descriptions of sub-seafloor replacement deposits in modern and ancient volcanic successions. Young sub-seafloor deposits have been mapped at Middle Valley; a sediment-covered ridge located along the Juan de Fuca spreading centre (Goodfellow and Blaise, 1988; Goodfellow and Franklin, 1993). The deposit comprises a small sulfide mound that passes down into a discordant pipe-like body of massive sulfide, which extends more than 93 m into the underlying siltstone. The best studied examples of sub-seafloor replacement style VHMS deposits are confined to ancient volcanic successions. These include deposits hosted by silicic to intermediate sequences (Allen, 1988, 1992; Khin Zaw and Large, 1992; Doyle and McPhie, 1994; Bodon and Valenta, 1995; Galley et al., 1995; Allen et al., 1996b) and mafic volcanic successions (Zierenberg et al., 1988). The paucity of examples is probably a reflection of the difficulty of recognising these deposits, rather than their abundance.
This chapter focuses on the problems associated with distinguishing between exhalative and sub-seafloor replacement style VHMS deposits, with the aims of clarifying the terminology, constraining the diagnostic/critical evidence for sub-seafloor replacement, and assessing the role of host lithofacies in determining the character of the resulting deposit. On the basis of published descriptions of sub-seafloor replacement style VHMS deposits and a detailed study of the Highway-Reward deposit, speculative models for the interaction between hydrothermal fluids and host lithofacies in the sub-seafloor environment are presented.

7.2 Terminology

The term “volcanic-hosted massive sulfide deposit” is used for syn-genetic accumulations of massive sulfide which are hosted by submarine volcanic successions (e.g. Solomon, 1976; Franklin et al., 1981; Lydon, 1988a; Large, 1992). Most deposits comprise a stratiform and sometimes strata-bound sulfide zone overlying a discordant massive sulfide and/or stringer zone. Others (e.g. Mount Morgan, Mount Lyell, Highway-Reward) are dominated by massive sulfide pipes or stringer and disseminated styles of mineralisation.

In the context of massive sulfide deposits, the term “exhalation” has been used to indicate fluid emanations into the hydrosphere from the lithosphere (Franklin et al., 1981). Most primary textures in mound-style deposits are indicative of sulfide replacement and sulfide infilling of pore space, rather than precipitation from exhaling hydrothermal fluids at the seawater-sediment interface. Growth of sulfide chimneys and the mound is largely attributed to the deposition of anhydrite and barite on the outer surfaces of the mound, with sulfates being continually replaced by sulfides during mound growth (Lydon, 1988b). For this reason, the term “exhalative ore” is avoided here in favour of the term “seafloor massive sulfide deposit”. The position of the seafloor is variable throughout the life of the hydrothermal system due to mound growth, sedimentation, volcanism (strata building) and/or oxidation of sulfides and erosion (strata destruction). The term “seafloor massive sulfide accumulation” is used for the process of sulfide deposition at successive seafloor positions without significant contributions of volcanic detritus or sediment. “Seafloor massive sulfide deposit” is used for the products of this style of hydrothermal activity.

Replacement is defined as a “change in composition of a mineral or mineral aggregate, presumably accomplished by diffusion of new material in and old material out, without breakdown of the solid state” (Bates and Jackson, 1987). In the context of VHMS deposits, sub-seafloor replacement refers to the syn-genetic formation of sulfide minerals by replacement and infiltration of pre-existing volcanic or sedimentary deposits.
Mineralisation

7.3

restriction of depth beneath the seafloor is placed on the term. The term does not refer specifically to the process of zone refining (Large et al., 1989) within the developing deposit, although this is probably important during sub-seafloor replacement. Neither does the term incorporate mineralisation which is vein-hosted or not syn-genetic.

Separate lenses or segments of a single deposit can form by different processes and both seafloor massive sulfide accumulation and sub-seafloor replacement may be involved in constructing an orebody (e.g. Vivallo, 1985; Zierenberg et al., 1988; Galley et al., 1993; Humphris et al., 1995). Clearly there is a spectrum from wholly sub-seafloor replacement type deposits to seafloor massive sulfide deposits.

7.3 The role of sub-seafloor replacement vs. seafloor accumulation

Studies of ancient VHMS deposits suggest that there are interdependent criteria for evaluating the role of sub-seafloor replacement during syn-genetic sulfide (or other minerals) deposition: (1) presence of precursor particles or relic coherent facies within the massive sulfide; (2) facies characteristics indicating that mineralisation post-dates deposition of the host; (3) identification of syn-depositional replacement fronts; (4) discordance with the enclosing host lithofacies; and (5) the presence of hanging wall alteration zones with footwall affinities. In most cases, more than a one criterion is required to demonstrate a seafloor or sub-seafloor origin for the mineralisation.

1. The presence of precursor particles or coherent facies

Volcanic and sedimentary particles can sometimes be preserved within the massive sulfide mineralisation. The particles may be partially or completely replaced by sulfide minerals, or are altered and contain disseminated sulfides (e.g. Bodon and Valenta, 1995; Galley et al., 1995). Fragment types, shape, composition and texture help constrain the particle forming mechanism and, when combined with the lithofacies character (bedforms, geometry, internal organisation), can be used to establish the character (porosity, permeability) of the precursor lithofacies and mechanisms of sulfide accumulation in the sub-seafloor environment. Considered alone, precursor particles are consistent with either sub-seafloor replacement or synchronous sedimentation and sulfide precipitation.

In some VHMS deposits, the matrix of autoclastic breccia is replaced by sulfides (e.g. Galley et al., 1995), or the massive sulfide contains segments or patches of strongly altered, mineralised coherent igneous rock (e.g. Highway-Reward). Parts of the massive sulfide mineralisation contained within relic coherent facies or autoclastic breccia can only have formed by replacement and infiltration. Where emplacement as a lava or intrusion cannot be proven, it remains possible that the igneous component was resedimented onto
Mineralisation 7.3.

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Mineralisation

7.4.

seafloor massive sulfide or sourced from subsurface environment and transported with the hydrothermal fluid (e.g. Humphris et al., 1995).

2. Evidence for rapid emplacement of the host facies

This criterion involves a genetic interpretation of the host, with particular attention to transport and depositional mechanisms and the nature and position of contacts between the host rock and mineralisation. Massive sulfide mineralisation which is hosted by rapidly or mass-flow emplaced units can only have formed by replacement and impregnation (Allen, 1994; Allen et al., 1996b). Most often the host will be a lava or intrusion and associated autoclastic breccia (e.g. Zierenberg et al., 1988), or syn-eruptive pumiceous mass-flow deposit (e.g. Allen, 1994; Khin Zaw and Large, 1992). Parts of some sub-seafloor replacement style massive sulfide deposits are hosted by thinly bedded volcano-sedimentary units rather than a single thick depositional unit (e.g. Bodon and Valenta, 1995; Galley et al., 1995). In these, the possibility remains that sedimentation was synchronous with hydrothermal activity and that precipitation of sulfides occurred at, above, and below the seafloor over the life of the hydrothermal system.

The rate of sediment accumulation will determine the relative importance of seafloor and sub-seafloor processes in massive sulfide accumulation. If sedimentation surpasses sulfide precipitation at the seafloor, then sub-seafloor replacement will become important. Growth rates for sulfide mounds are poorly constrained. Chimneys on the mound can grow 8 to 30 cm per day (Goldfarb et al., 1983; Hekinian et al., 1983). Although a single typical black smoker chimney produces around 250 tonnes of sulfide per annum, much of this is dispersed by currents and incorporated into distal sediments (Scott, 1992). Sulfide precipitation at seafloor chimney mounds may be intermittent and separated by long periods dominated by ambient sedimentation (e.g. Rona et al., 1993). Regional sedimentation rates for oceanic spreading centres are generally low (e.g. 1.8 cm/kyr, TAG, Scott et al., 1978), whereas in back arc basin settings sedimentation can be rapid (e.g. Taylor et al., 1991). In the Sumisu Rift of the Izu-Bonin island arc, sediments accumulated at between 90 mm/kyr and 6 m/kyr from 0.1 to 1 Ma (Taylor et al., 1991). The basin fill includes hemipelagic sediment and voluminous volcaniclastic turbidites sourced from arc volcanoes (Nishimura and Murakami, 1988). Volcanism, especially explosive volcanism, has the potential to release large volumes volcaniclastic detritus into submarine settings (e.g. Cas and Wright, 1991; McPhie and Allen, 1992). Burial of chimney mound deposits by volcaniclastic mass flows or lavas can occur during the life of the hydrothermal system, interrupting or terminating seafloor sulfide deposition (e.g. Haymon et al., 1993). As the hydrothermal system attempts to re-establish a seafloor position, sub-seafloor replacement of the intervening lithofacies by sulfide minerals may become important (e.g. Haymon et al., 1993; Humphris et al., 1995).
3. Identification of syn-depositional replacement fronts
Bodon and Valenta (1995) documented primary and tectonic features in the Currawong Zn-Cu-Pb VHMS deposit, Benambra. At the margins of one massive sulfide lens there are replacement fronts passing from the various ores into sandstone beds. Intercalated siltstone laminae are not replaced. A thin quartz-sericite alteration halo extends out from the margins of the sulfide lenses into relic sandstone beds and provides critical evidence for replacement of a pre-existing lithology. Galley et al. (1995) identified similar replacement fronts in the Ansil VHMS deposit, Canada. Alteration intensity increases towards contacts between the massive sulfide lens and tuffaceous host rocks, and beds can be traced into the massive sulfide. Within the ore zone, bases of sandstone beds are mineralised and finer grained tops are silicified. Thin veins cut across silicified tops of beds and link the sphalerite-rich mineralisation. At South Hercules, massive sphalerite mineralisation often passes out through a zone of altered host rock with “spotty” sphalerite ore, into the enclosing pumiceous units (Khin Zaw and Large, 1992). Khin Zaw and Large (1992) interpreted the zonation to record the lateral and vertical migration of hydrothermal fluids within the pre-existing host lithofacies. The “spotty” ore records the nucleation of sulfides within the pumiceous host (Khin Zaw and Large, 1992) and can be considered one type of replacement front. Partially replaced beds and/or clasts are good evidence that bedforms in sulfide can be the product of selective replacement and that these parts of a deposit formed by sub-seafloor replacement of permeable sedimentary horizons.

4. Discordance within the enclosing lithofacies
Discordance within the enclosing volcano-sedimentary package can provide evidence for replacement of a precursor lithofacies (e.g. Khin Zaw and Large, 1992; Galley et al., 1993). However, progressive burial of the older parts of a growing mound/lens by sedimentation or volcanism concurrent with massive sulfide deposition might generate sharp discordant contacts. Accordingly, the criterion cannot be used alone and requires consideration of the transport and depositional mechanisms of the host lithofacies, and the recognition of the seafloor position(s) at the time of mineralisation.

5. Hanging wall alteration zones with footwall affinities
In seafloor VHMS deposits, the upward termination of intense hydrothermal alteration at the position of stratiform sulfides indicates that the ore-forming hydrothermal activity occurred after emplacement of the footwall rocks and before deposition of the hanging wall rocks. Zones of subtle hanging wall alteration have been reported above mineralised intersections in a number of Australian VHMS deposits (e.g. Mount Chalmers, Large and Both, 1980; Hellyer, Jack, 1989). These are interpreted to record weak hydrothermal activity during deposition of the hanging wall volcano-sedimentary package. In contrast, zones of intense hanging wall alteration suggest that ore-forming hydrothermal activity
was continuing during deposition of the hanging wall lithologies or that massive sulfide deposition was entirely sub-seafloor. Given that many volcanic facies are rapidly emplaced, strong hanging wall alteration is entirely plausible even for seafloor exhalative deposits. Burial of seafloor massive sulfide by lavas (e.g. Haymon et al., 1993) or volcaniclastic units can only be discounted by considering the lithofacies character of the host. Intense hanging wall alteration may be mineralogically similar to alteration in the footwall and can contain significant pipe- or vein-style mineralisation (e.g. Highway-Reward, Doyle, 1995; Scuddles, Ashley et al., 1988) and separate massive sulfide lenses/mounds/chimneys at successive seafloor positions (e.g. Haymon et al., 1993).

An assessment of other textures and structures in VHMS deposits

Chimneys and chimney fragments provide strong evidence for sulfide accumulation at the seafloor (Lydon, 1988b) and have been recognised in ancient deposits (e.g. Oudin and Constantinou, 1984). Fossil tube worms and bivalves in chemolithotrophic communities which colonise modern seafloor sulfide mounds can sometimes be preserved by sulfide minerals and are characteristic of seafloor deposits (Haymon et al., 1984; Oudin and Constantinou, 1984). Few other textures or structures in VHMS deposits have genetic significance in distinguishing seafloor deposits from sub-seafloor deposits. Many massive sulfide textures form through open space filling (e.g. colloform banding, network textures). However, this can include intramound porosity, primary lithological porosity, and/or space formed by fluid-rock-mineral interaction (e.g. dissolution of anhydrite). A critical line of evidence in support of a seafloor origin for VHMS deposits has been the recognition of sedimentary structures within some massive sulfide lenses. Graded bedding, cross bedding, and intercalations of laminated and fragmental ore have been identified in the Kuroko deposits (e.g. Ohmoto and Skinner, 1983) and Australian deposits (e.g. Mount Chalmers, Large and Both, 1980). The recognition of similar structures in sub-seafloor replacement style deposits suggests that selective replacement of clasts in pre-existing lithofacies can form similar structures (e.g. Bodon and Valenta, 1995; Galley et al., 1995). Apparent interbedding of sulfide and sediment can result if more permeable laminae are selectively replaced and fine-grained (impermeable) laminae remain unaffected (Bodon and Valenta, 1995). Breccia ore facies can also form in situ within sulfide mounds by dissolution of anhydrite in the matrix to sulfide patches (e.g. TAG, Humphris et al., 1995). Banded sulfides are a common feature of VHMS deposits. In most cases, it is difficult to determine whether the banding is primary or due to later deformation (e.g. Large et al., 1988).

Iron oxide-silica units ("ironstones", Chapter 6) that form the ore equivalent horizon to VHMS deposits can provide evidence for a seafloor position which includes the ore
horizon. However, very similar iron-oxide silica units can from by sub-seafloor replacements of pre-existing strata (Chapter 6). Microaerophilic chemolithotrophic organisms can colonise sub-seafloor sediments (Chapter 6) so are not characteristic of a seafloor position. In some cases, the favourable horizon is not represented by a chemical sediment but sulfide-clast bearing mass-flow deposits (e.g. Hercules, Allen and Huns, 1990; Hellyer, Waters and Wallace, 1992). Mass flows can erode several metres down into the substrate and successive mass flows up to 10's of metres. Sulfide clasts indicate only that massive sulfide accumulation occurred at or near the seafloor.

7.4 Highway-Reward massive sulfide deposit

The Highway Member (Trooper Creek Formation) at Highway-Reward hosts two spatially associated copper-gold-rich massive sulfide orebodies: Highway and Reward (Fig. 1.1). After Thalanga, the Highway-Reward deposit is the largest known VHMS deposit in the Mount Windsor Subprovince. The Highway orebody is located approximately 200 m NNW of the Reward orebody beneath the abandoned Highway open pit (Fig. 7.1). The massive sulfide deposit consist of sub-vertical pipe-like bodies of massive pyrite-chalcopyrite that are associated with minor marginal discordant and strata-bound zones of pyrite-sphalerite±galena±barite mineralisation. Supergene and oxide ores overlie the primary ores at both Highway and Reward. Most of the oxide resource at Highway was mined from the Highway open cut by North Queensland Resources in the period 1986 to 1988. A significant primary resource remains at Highway and the Reward orebody is presently undeveloped (Table 7.1).

<table>
<thead>
<tr>
<th></th>
<th>Highway</th>
<th>Reward</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Primary</strong></td>
<td>1.2 m.t. @ 5.5% Cu, 6.5 g/t Ag, 1.2 g/t Au</td>
<td>0.2 m.t. @ 3.5 Cu, 13 g/t Ag, 1 g/t Au</td>
</tr>
<tr>
<td><strong>Supergene</strong></td>
<td></td>
<td>0.3 m.t. @ 11.6% Cu, 21 g/t Ag, 1.8 g/t Au</td>
</tr>
<tr>
<td><strong>Oxide</strong></td>
<td>0.07 m.t. @ 6.04 g/t Au</td>
<td>0.1 m.t. @ 33 g/t Ag, 6.49 g/t Au</td>
</tr>
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</table>

Table 7.1 Grade and tonnage data for the primary, oxide and supergene ore zones at Highway-Reward. The Highway oxide resource is a pre-mining resource estimate. Data from Beams and Dronseika (1995).
Previous investigations have briefly addressed various aspects of the Highway-Reward deposit: geophysics and discovery (Kay, 1982; Beams et al., 1989, 1993; Beams et al., 1990; Beams and Dronseika, 1995), geology of the mineralisation and host rocks (Kay, 1987; Berry et al., 1992; McPhie and Large, 1992; Large, 1992; Doyle, 1994), sulfur and lead isotope geochemistry of the massive sulfide (Huston, 1992; Dean and Carr, 1992), deformation style and strain partitioning around the orebodies (Laing, 1988; Berry, 1989); and exploration geochemistry (Beams and Jenkins, 1995). The interpretations presented here are based on detailed diamond drill core logging (61 holes), 1:250 scale mapping of the Highway open pit and the few available surface outcrops, and detailed petrography of the host lithofacies.

Figure 7.1. Surface projection of the Highway and Reward orebodies from approximately 175 m RL (150 m below surface). Stippling highlights the distribution of Cu-rich massive sulfide. Modified from Aberfoyle Resources Limited.
7.4.1 Local geology

The lithofacies associations which host the Highway-Reward deposit are addressed in Chapter 5 and briefly reviewed here. The massive sulfide deposit is hosted by a silicic intrusive and volcanic succession intercalated with sedimentary facies that indicate a submarine, below-storm-wave-base environment of deposition (cf. Beams et al., 1989; Berry et al., 1992; McPhie and Large, 1992; Beams et al., 1993). Contact relationships and phenocryst mineralogy, size and percentages indicate the presence of more than thirteen distinct porphyritic units in a volume of 1 x 1 x 0.5 km. The peperitic upper margins to many porphyritic units demonstrate their intrusion into wet unconsolidated sediment (e.g. Allen, 1992; McPhie et al., 1993). Syn-sedimentary sills, cryptodomes, partly extrusive cryptodomes and associated in situ and resedimented autoclastic deposits have been recognised. These are the principal facies in the environment of mineralisation and represent a proximal facies association from intrabasinal, non-explosive, syn-sedimentary intrusion-dominated magmatism.

Porphyritic units are intercalated with a volcaniclastic and sedimentary facies association comprising suspension-settled siltstone, sandstone turbidites and thick, non-welded pumice- and crystal-rich sandstone and breccia units. Pumiceous and crystal-rich deposits record episodes of explosive silicic volcanism in a shallow marine, basin margin or extrabasinal environment, and were emplaced by cold, water-supported, high-concentration sediment gravity currents. Unaltered andesite dykes cut across the massive sulfide and altered host rocks.

Volcanic host rocks in the Highway-Reward area are weathered to depths in excess of 100 m. The weathered and oxidised zone is more extensive at Reward and extends upward to surface from near the top of the massive sulfide orebody on most sections. Prior to development the Highway orebody was marked by a gossan at surface. The Reward orebody occurs under 100 m combined thickness of Tertiary fluviatile sediments (Campaspe Formation) and weathered gossanous volcanics (e.g. Beams et al., 1989). Bedding measured in drill core suggests a shallow dip of around 10° south east in the subsurface volcanics (cf. Laing, 1988). Flow banding in rhyolite from the Highway open pit has steeper dips (18-55°), but may be at a high angle to bedding.

7.4.2 The massive sulfide orebodies

The mineralisation can be divided into five principal types on the basis of mineralogy, textures and the relationships to the host rocks (cf. Beams et al., 1989; McPhie and Large, 1992; Beams and Dronseika, 1995). These are: (1) pyrite-chalcopyrite pipes; (2)
pyrite-sphalerite±galena±barite; (3) footwall stringer; (4) hanging wall stringer; and (5) gossanous breccia.

**Pyrite-chalcopyrite pipes**

The Highway and Reward pyrite-chalcopyrite pipes are discordant to bedding, but parallel to S4, and have a plunge sub-parallel to a sub-vertical mineral lineation (cf. Berry et al., 1992). The Highway pipe is a NE trending, chalcopyrite-rich massive pyrite body that is approximately 175 m long. To the south, the pipe is less than 20 m wide and 200 m thick (section 10050N; Figs. 7.2, 7.3). On central and northern sections the deposit comprises two or three pipes spaced less than 3 to 10 m apart (section 10100N; Fig. 7.4) or a single “dome” or “tooth-shaped” massive sulfide body. Reward consists of one main and several smaller subvertical, NNE trending, massive pyrite pipes (Fig. 7.4). The largest pipe is saddle-shaped in plan, with a bulbous top in cross-section (Fig. 7.4). The pipe is 100 m x 150 m in plan and up to 250 m thick. It contains around 5 million tonnes of massive pyrite (Beams et al., 1989), which includes a small chalcopyrite-rich zone.

**Mineralogy, textures, and contact relationships**

The massive pyrite-chalcopyrite pipes have a relatively simple mineralogy of pyrite and chalcopyrite, with minor tennantite, sphalerite, quartz and sericite (Fig. 7.5A). Trace minerals include chlorite, aikinite, galena, barite, hematite, rutile and a carbonate (Huston, 1992). Pyrite is typically subhedral to euhedral and fine to medium grained (20-500 μm). Patches of coarse-grained (2-5 mm) pyrite are often associated with shear zones. Rarely pyrite has a lath-like habit (e.g. REM 150, 159.3 m) and may be replacing an earlier mineral such as hematite, barite, or anhydrite (cf. Large, 1992). Pyrite can also be spongy, shreddy, frambooidal, or exhibit snowflake texture (Huston, 1992). Chalcopyrite flecks, veinlets (< 1 mm wide) and patches cut across massive pyrite. Chalcopyrite fills the interstices between pyrite crystals. Barite, quartz-carbonate and anhydrite veins cut across the pyrite-chalcopyrite mineralisation. In some samples, barite occurs as tabular laths interstitial to pyrite and is replaced by quartz (e.g. REM 147, 140 m).

Matrix gangue is generally absent, other than along the margins of the massive sulfide pipes where quartz or sericite±quartz±pyrite-altered volcanic rock is abundant. The exception is a small segment of the Highway pipe (e.g. REM 551, 228-285), where interstitial sericite (now clay) comprises around 10% of the pipe. In most other holes, semi-massive and massive pyrite with 30-60% relic patches and segments of altered coherent rhyolite to dacite and peperite occur within a 1 to 10 m wide zone at the margins and tops of the pipes (Fig. 7.5B). Relic quartz phenocrysts are rarely preserved within the sulfide, but quartz and/or feldspar are common in altered volcanic intervals. The rhyolitic or dacitic component of the peperite is often more pyritic than siltstone in the matrix or
Figure 7.2. Simplified geological cross-section of the Highway-Reward deposit at 1002SN. The positions of the main rhyolite units (R1, R3-6) are also shown.
Figure 7.3. Simplified geological cross-section of the Highway-Reward deposit at 10050N. RL = relative position above sea level. The locations of dacite D1, rhyodacite RD2 and the main rhyolite units (R3-7) are also shown.
Figure 7.4. Simplified geological cross-section of the Highway-Reward deposit at 10100N. RL = relative position above sea level. Dacite D1, rhyodacite RD2 and rhyolites R2-4 and R6 occur on this section.
filling fractures in the breccias (e.g. REM 142, 170-177 m). Interstitial anhedral quartz constitutes up to 20 to 30% of the pipe along some of the contacts. Quartz is intergrown with pyrite, forms patches up to 2 cm across, or occurs in bands of pyrite-quartz±barite (Fig. 7.5A-2). Contacts vary from sharp to finely bulbous and many are sheared. A halo of disseminated and patchy pyrite extends cut into the surrounding altered rhyolite and dacite or peperite.

*Sphalerite-galena mineralisation*

The pyrite-chalcopyrite pipes are surrounded by a halo of Pb-Zn-Ba-rich mineralisation up to 500 m wide in an east-west direction and around 225 m long from north to south. Within this halo there are four styles of Zn±Pb±Ba mineralisation. They are: (1) veins and veinlets of sphalerite±galena-barite within altered volcanic rocks along the margins and tops of the pyrite pipes (Fig. 7.2); (2) disseminated, patchy and spotty sphalerite within sericite-quartz-chlorite-altered rocks at the tops of the pipes (Fig. 7.5C-1); (3) lenses of strata-bound massive pyrite-sphalerite-chalcopyrite-barite closely associated with volcaniclastic mass-flow units in the hanging wall to the Reward pipe (Fig. 7.2; 7.5C-2); (4) massive to semi-massive pyrite-sphalerite-chalcopyrite±barite at the margins of the pyrite pipes (Fig. 7.5C-3).

(1) The first style consists of low grade Pb-Zn and can be found throughout the host succession above and adjacent to the wall of the pipes. However, the strongest vein development occurs along the margins of the Reward pipe (Fig. 7.2). (2) Disseminated and patchy sphalerite and galena form a halo above the Highway pipe and extend around 5 m into the peperitic base of the hanging wall rhyolite (e.g. REM 155, 195-200 m). Above Reward, supergene clay, sphalerite, bornite and covellite occur as disseminations and patches near surface, partially oxidised samples. (3) The halo of Pb-Zn mineralisation extends up to 20 m into the hanging wall, 80 m laterally and includes a small strata-bound pyrite-sphalerite-chalcopyrite-barite lens around 60 m in length. The massive sulfide lens is spatially associated with, but physically separated from, the main Reward pipe (Figs. 7.2-7.4). Drilling to date suggests that southern part of the lens comprises massive and finely banded sphalerite-rich massive sulfide (Fig. 7.2). Banded sulfide textures are common towards the base of the interval, and comprise thin sulfide bands and siltstone bands which are deformed by small F4? folds. The northern part of the lens has a progressively thickening pyrite-rich base and grades into a discordant massive pyrite pod (Fig. 7.3, 7.6). The pod occurs around 20 m above the southern margin of the Reward pipe (Fig. 7.4). Sphalerite-barite-rich massive sulfide intersected at a similar stratigraphic position in drill holes further to the east is interpreted as continuation of the pyrite pod. (4) The fourth style of massive sulfide occurs within a thin (<1 m) zone at the margins of the pyrite-chalcopyrite pipes (e.g. REM 150, 224-228 m).
Figure 7.5

(A) 1. Massive pyrite-chalcopyrite mineralisation from the Highway pipe. REM 142, 194.4 m. 2. Banded pyrite-chalcopyrite-quartz mineralisation from the margin of the main Reward pipe. Chalcopyrite is aligned in the foliation (bar). REM 560 205.4 m.

(B) Semi-massive pyrite-chalcopyrite (p) with relic patches and segments of strongly altered coherent dacite (d) and peperite. Core is 5 cm wide. HMO 62, 157-162 m; Reward.

(C) Sphalerite-galena mineralisation. 1. Spotty sphalerite (arrow) within sericite-quartz-chlorite-altered sandstone at the top of the Highway pipe. REM 152, 160.8 m; 2. Sample of banded pyrite-sphalerite-chalcopyrite-barite mineralisation from a strata-bound lens above the main Reward pipe. REM 122, 157.7 m. 3. Sphalerite-pyrite mineralisation from the margin of the Highway pipe. REM 150, 225.4 m.

(D) Banded pyrite-quartz-barite-sphalerite mineralisation replacing rhyolitic peperite along the upper margin of the Reward pipe. The rhyolite (r) is strongly chloritised and pyritic, whereas the siltstone (s) is silicified. REM 116, 182-190 m; Reward.

(E) Photomicrograph of strata-bound pyrite-sphalerite-barite-rich mineralisation from the Reward orebody. Apparent grading in this quartz-sericite-altered unit is defined by euhedral pyrite (p). Plane polarised light. REM 122, 157 m.

(F) As for E, but in reflected light.

(G) Pyrite-quartz veins (v) in sericite-quartz-altered dacite below the Highway orebody. Fibre quartz forms a halo around the veins (1) or fills fractures within the pyrite mineralisation (2). REM 154, 314.4 m; REM 154, 257.5 m.

(H) In situ brecciated (tectonic) massive pyrite-chalcopyrite mineralisation from the Reward orebody. Core is 5 cm wide. HMO 60, 145-150 m; Reward orebody.
Sphalerite:pyrite.
CryStal-Yltric: Ihie
chalcopyrite-barik:
Breccia sulfide
Quartz-
reldspar -
Rhyolite
Qt<&F1d
 Bearing
Sediment-miltnx
Rhyolite-bearing
rhyolite breccia (peperite)
Banded sulfite
Rhyolite: figure-5
breccia (Hyphid)
Cry-ual-varsic-holc sulfide
Pyrite:chalcopyrite-barik: muscovite-chlorite
Banded sulfite
Breccia sulfide

Figure 7.6. (A) Selected drill logs showing JIP of plant internal textural variations and contact relationships of sphalerite-rich mineralisation and volcanic facies to the south of the Reward pipe. Alteration facies as illustrated in 7.8. Alteration codes outlined in appendix C. (B) Plan projection of diamond drill holes in "A".
It consists of massive to finely banded pyrite-chalcopyrite-sphalerite-barite replacing peperite. The bands are 1-2 mm wide, strongly contorted, and mimic contacts between the massive sulfide and altered volcanic rock. (Fig. 7.5D; REM 116, 182-190 m).

**Mineralogy and textures**

This mineralisation style consists of varying proportions of pyrite, sphalerite, chalcopyrite and galena, with a gangue dominated by quartz, sericite and barite. Minor and trace minerals include tennantite, carbonate, chlorite, bismuth minerals, marcasite, hematite and electrum (Huston, 1992). Minerals in the supergene zone are clay, bornite and covellite. Pyrite is euhedral or exhibits snowflake, colloform, or spongy textures (Huston, 1992). Sphalerite occurs as anhedral intergrowths with pyrite and chalcopyrite, and often exhibits chalcopyrite disease. Barite laths are commonly pseudomorphed by quartz.

Fine lamination in the strata-bound pyrite-sphalerite-chalcopyrite-barite mineralisation closely resembles bedded sulfide sediments (cf. Huston, 1992). A limited microscopic study (REM 122, 157.7 m) suggests that laminae consist of 1-2 mm microcrystalline bands of anhedral quartz intergrown with pyrite, alternating with 1-20 mm bands of pyrite-sphalerite-chalcopyrite with interstitial coarser-grained (0.05-0.2 mm) anhedral quartz and minor sericite. The grain size of pyrite grains decreases from the base to the top of the thicker bands and resembles sedimentary grading (Fig. 7.5E-F). Pyrite is euhedral to subhedral and generally associated with sphalerite. Chalcopyrite occurs as anhedral grains and as chalcopyrite disease in sphalerite. Banding in the strata-bound interval is cross-cut, and partially replaced by, veins and veinlets of pyrite and quartz with diffuse boundaries. Similar veins dissect massive sphalerite mineralisation at the margins of the pyrite pipes.

**Footwall and hanging wall stringer veins**

Pyrite±quartz±sericite stringer veins occur below the Highway and Reward pipes and adjacent to the walls of the pipes (Fig. 7.5G). On some sections, stringer pyrite±covellite±chalcopyrite veins also occur in sericite-quartz-altered volcanic rocks above the pyrite pipes (e.g. REM 147, 101-116; Fig. 7.4). Small veins are dismembered in the cleavage and larger veins are fractured internally. Fibre quartz fills the fractures or forms a selvedge surrounding a pyrite±sericite±quartz core. A few veins comprise discontinuous, finely bulbous bands of quartz and pyrite. In general, disseminated fine to coarse-grained, euhedral to subhedral pyrite accounts for 10-20% of the wall rock beneath the pyrite pipes and veins vary from 0.5 to 15 cm across, but are mostly around 1 cm wide. Towards the centre of the stringer zone, vein spacing decreases from 0.5-1 m to 15-30 cm (e.g. REM 142, 275-290 m). In the hanging wall, stringer veins are thinner
Mineralisation (0.1-1 cm wide) and spaced 30 to 40 cm apart (e.g. REM 560, 101-120; REM 147, 102-116 m). The wall rock typically contains only 6-10% disseminated pyrite.

**Gossanous breccia zones**

Gossanous zones occur within volcaniclastic units and coherent rhyolite overlying the pyrite pipes and exposed in the base of the Highway pit (Fig. 7.3). They comprise heavily silicified and/or clay weathered volcanic rock with iron oxide-filled veinlets and cubic hollows after disseminated pyrite. Relic quartz phenocrysts are preserved in intervals of gossanous rhyolite. Some quartz-bearing zones contain siltstone seams and patches and may be altered peperite. Veins and patches of tabular barite crystals (1-2 mm long) are abundant in a few cores (e.g. REM 132, 58-69 m).

**7.4.3 Effects of deformation and metamorphism on the massive sulfide orebodies**

The host lithofacies, mineralisation and associated alteration have the same tectonic fabric, although the intensity of the fabric varies. The massive pyrite pipes have been fractured and dilated by the deformation and include local domains of cataclasite (e.g. HMO 60, 133-151 m; Fig. 7.5H). The alteration halo has sheared parallel to the subvertical S4 cleavage. Shearing has been focussed along the margins of the pyrite pipes and porphyries. Spotty and disseminated sphalerite have been drawn out into the cleavage and bands in strata-bound pyrite-sphalerite lens folded. Fibre quartz has grown in the pressure shadows of disseminated euhedral pyrite grains. Stringer veins have been dismembered in the cleavage, whereas gypsum veins are crenulated by S4 (e.g. REM 149, 169.8 m). Chalcopyrite was mobilised into the cleavage which cuts across primary quartz-pyrite banding at the margins of the pipes.

**7.4.4 Alteration mineralogy and distribution**

Detailed drill core logging of alteration and volcanic facies allow for investigation of the interplay between primary volcanic texture, the texture and mineralogy of each alteration stage, and mineralisation. Graphic lithological logs have been used to record the nature and positions of contacts, volcanic textures and facies, alteration mineralogy and mineralisation (Appendix C). Textural relationships between each alteration stage and the pre-alteration texture, mineral abundances and associations, and chronology have been recorded using a code system which is detailed in Appendix C.
The Highway and Reward orebodies occur within a well-developed discordant alteration envelope. The envelope extends from at least 150 m below the orebodies to over 60 m above the Highway pipe (Fig. 7.7). At Reward, intense weathering and oxidation obscure the hanging wall alteration on most sections (Fig. 7.8). Footwall alteration occurs along the entire length of the deposit in a 500 m by 250 m post-deformational elliptical shape. Small zones of footwall alteration also occur peripheral to the orebodies at Gateway (200-250 m west of Highway) and 200 m east of Reward.

Hydrothermal alteration at Highway-Reward is complex, reflecting primary heterogeneity modified by subsequent faulting. Broadly the alteration envelope has a mineralogical zoning which is defined by assemblages of sericite, chlorite-sericite, quartz-chlorite-sericite, chlorite-anhydrite, quartz-sericite±pyrite, albite-sericite-chlorite-quartz and hematite-chlorite-sericite-quartz. Disseminated pyrite is a common accessory in all but the latter two assemblages. Although the stringer veining and sulfide orebodies at Highway and Reward are spatially distinct, the alteration zones associated with each overlap; and some alteration zones partially envelop both orebodies. A quartz-sericite±pyrite zone is centred beneath the pyrite pipes and on some sections extends into the hanging wall succession (Figs. 7.7-7.8). Small domains of intense chlorite-anhydrite alteration occur within the footwall quartz-sericite±pyrite zone and along some margins of the pipes (Fig. 7.7). Quartz-sericite±pyrite alteration gives way laterally and vertically to domains of sericite-chlorite-quartz and chlorite-sericite alteration. Beyond the hydrothermal alteration halo, rocks of rhyolitic to dacitic composition contain various assemblages of feldspar (albite), calcite, sericite, chlorite, quartz and hematite-chlorite-sericite-quartz (Fig. 7.8). Contacts between the alteration zones are typically gradational. Less intense alteration, as indicated by variable preservation of feldspar phenocrysts and volcanic textures, occurs more commonly marginal to the orebodies and in the hanging wall. The most intense alteration occurs in the stringer zone beneath the pyrite pipes, and on some sections, in quartz-sericite-altered hanging wall rocks.

**Quartz-sericite±pyrite zone**

Within this zone, the relative proportions of quartz and sericite varies considerably, and either can be the dominant phase. Quartz-dominant alteration is more common beneath the pipes. The quartz-sericite±pyrite alteration is pervasive or patchy and obscures or destroys original volcanic textures creating apparent clastic textures (Fig. 7.9A). Relic quartz phenocrysts are locally recognisable along the margins of the zone, particularly in sericitic apparent clasts enclosed by more intensely silicified wall rock. In some samples, disseminated pyrite preferentially replaced the sericitic component. The quartz-sericite±pyrite alteration encloses domains of intense pervasive quartz alteration which are associated with strong, quartz-pyrite stringer vein development (Fig. 7.8). In
Figure 7.7 Simplified geological cross section showing the distribution of volcanic facies, alteration and mineralisation at 10100N. RL = relative position above sea level.
Figure 7.8. Simplified geological cross section showing the distribution of volcanic facies, alteration and mineralisation at 10150N. RL = relative position above sea level.
the hanging wall, a discordant zone of strong quartz-sericite±pyrite alteration extends up to more than 60 m (10075N) above the Highway pipe into the overlying rhyolite and volcaniclastic deposits (Fig. 7.7). A similar, but poorly preserved zone occurs above the main Reward pipe. This style of alteration obscures primary textures, but only in the most silicified domains are the lithofacies unrecognisable.

**Chlorite±anhydrite zone**

Small zones of intense pervasive chlorite alteration are developed sporadically within the footwall quartz-sericite zone. In a few samples, chloritised wall rock contains lath- and ovoid-shaped quartz spots, 0.5-2 mm in diameter. The quartz may be pseudomorphing an earlier mineral such as barite or carbonate. At the margins of the Reward pipe, pods of very fine-grained (30-50 μm), weakly foliated chlorite contain abundant euhedral anhydrite crystals up to 1.5 cm (Fig. 7.9B). The foliation wraps around the anhydrite, which indicates a pre- or syn-kinematic timing for anhydrite crystallisation (Fig. 7.9C; cf. Huston, 1992). This observation contradicts the late-kinematic timing for anhydrite proposed by Beams et al. (1989, 1990). The chlorite-anhydrite alteration contains rare relic quartz-sericite patches.

**Sericite alteration**

This style of alteration is distributed throughout the alteration envelope, but is particularly common within shear zones at the margins of the Highway and Reward orebodies (Fig. 7.7). Sericitised wall rock also occurs as the gangue in semi-massive pyrite-chalcopyrite at the margins of the pyrite pipe. Subsequent tectonic foliation is strongly developed in the sericitic domains and has obscured or destroyed primary textures in the volcanic rock.

**Quartz-chlorite-sericite zone**

Outward from the quartz-sericite zone is a widespread quartz-chlorite-sericite zone. The zone comprises a complex alteration assemblage in which the dominant phase can be chlorite, sericite-quartz, or chlorite-sericite. Alteration within this zone is strongly controlled by primary volcanic textures, particularly fracture and matrix permeability. Pale to dark phyllosilicate-rich alteration is typically pervasive. Quartz-sericite±pyrite alteration is often strongly controlled by fractures and the matrix in autoclastic breccia, and has overprinted earlier phyllosilicate-rich alteration assemblages (Fig. 7.9D). Domains of chlorite-sericite and chlorite alteration are often preserved within more extensive quartz-sericite alteration supporting the paragenesis. However, chlorite stringer veins with diffuse margins sometimes cut across the quartz-sericite alteration, suggesting that there
Figure 7.9

Examples of typical alteration features at Highway-Reward.

(A) Quartz-sericite ± pyrite zone. Drill core showing quartz-sericite ± pyrite alteration in lithofacies from the footwall (1) and hanging wall (2) of the Reward and Highway orebodies. REW 807, 454.7 m; REM 137, 168.9 m.

(B) Chlorite-anhydrite zone. Intensely chlorite-altered volcanic rock containing abundant euhedral anhydrite crystals (a). Core is 5 cm wide. HMO 41, 283.6 m; Reward.

(C) Chlorite-anhydrite zone. The foliation (arrow) wraps around the anhydrite (a) suggesting a pre- or syn-kinematic timing for anhydrite. Plane polarised light. HMO 41, 283.6 m; Reward.

(D) Quartz-chlorite-sericite zone. This breccia is strictly monomictic, comprising blocky and cuneiform dacite clasts (c) that form jigsaw-fit aggregates. Brecciation was probably caused by quench fracturing of a formerly glassy cryptodome, producing in situ hyaloclastite. Quartz-sericite alteration (q) along and adjacent to the matrix overprints earlier, pervasive chlorite-sericite alteration. Core is 5 cm wide. REM 560, 305 m; Highway.

(E) These three drill core samples come from a pumiceous, subaqueous mass-flow deposit. Cores 1 and 2 contain nodular domains (n) of albite alteration, separated by sericite-chlorite-altered pumice clasts (arrow). Alignment of compacted pumice clasts defines a bedding-parallel compaction foliation (S1). In core 3, nodules merge and form patches within the pumice breccia. Core is 5 cm wide. REW 805 (277.9-278.4 m).

(F) Samples of pumice breccia are composed of variably compacted pumice clasts and are non-welded. Sericite-chlorite-altered pumice clasts (c) are compacted around the more competent albite-altered pumice clasts (p) and define a diagenetic compaction foliation. The albite-altered pumice clasts have uncompacted tube vesicles (bar). Plane polarised light. REW 578, 203.7 m.

(G) The formerly glassy groundmass of this rhyolite contains relic classical perlitic fractures. The fractures are delineated by chlorite and sericite (c). The remainder of the glass has been replaced by albite (a) and quartz (q). Plane polarised light. REW 802, 183.1 m.

(H) These four samples are from a rhyolitic intrusion. In situ hyaloclastite (core 1) passes gradationally down into peperite comprising rhyolite clasts and a pumiceous matrix (core 2, 3). Clasts and the matrix in the breccia have been silicified, sericitised and hematite-altered (core 3). Pumice breccia away from the contact (core 4) is sericite-quartz-altered. Core is 5 cm wide. HMO 52, 146-160 m, Highway-Reward.
are several generations of chlorite alteration. The polyphase alteration has resulted in the widespread development of patchy, mottled, wispy and apparent clastic textures.

**Chlorite-sericite alteration**

Low intensity chlorite-sericite alteration with minor carbonate, characterises this alteration zone. The assemblage is largely restricted to domains within more widespread quartz-chlorite-sericite alteration and is common towards the northern end of the Highway orebody. Near the hanging wall contact of the Highway pipe, quartz-sericite-pyrite alteration passes gradationally out through a chlorite-sericite zone into weak pervasive chlorite alteration (e.g. REM 562, 100-197 m). Along the contacts, quartz-sericite alteration often occurs as a "wash" within the chlorite-sericite zone and overprints earlier chlorite and sericite alteration.

**Feldspar (albite & K-feldspar) ±carbonate-chlorite-sericite-quartz alteration**

Many formerly glassy rhyolitic to dacitic volcanic rocks in the Trooper Creek Formation have altered to various assemblages of feldspar, carbonate, chlorite, sericite and quartz. At Highway-Reward, these assemblages are restricted to hanging wall lithofacies greater than 200 m from the orebodies (Fig. 7.8). The alteration is regionally distributed but selectively replaces glassy parts of lavas and intrusions, pumiceous units and crystal-vitric breccia and sandstone beds. This style of alteration is very heterogeneous, forming pale feldspar-quartz-rich domains and green chlorite-sericite-rich areas. In pumiceous and shard-rich deposits, secondary feldspar or carbonate nucleated around feldspar phenocrysts and moved out as nodular zones (Fig. 7.9E), infilling vesicles and replacing formerly glassy vesicle walls (Fig. 7.9F). The feldspar alteration was incomplete, leaving weakly altered domains which were subsequently altered to more phyllosilicate-rich compositions. The phyllosilicate-altered domains are often strongly elongate and define a bedding-parallel diageneic compaction (or early tectonic) foliation (cf. Allen and Cas, 1990; Chapter 2). In the glassy parts of lavas and intrusions, early feldspar-carbonate-chlorite alteration was typically pervasive or fracture controlled. More advanced phyllosilicate-dominated alteration generally spread outward from fractures or the matrix, overprinting earlier feldspar alteration (Fig. 7.9G).

**Hematite±quartz-sericite-chlorite**

A second common alteration style in the Trooper Creek Formation, as represented in lithofacies 50-200 m from the Highway-Reward deposit, is widespread hematite±quartz-sericite-chlorite alteration. Occasionally, small zones of hematite-rich alteration are preserved along the margins of the orebodies (e.g. REM 122, 125-136 m). The alteration
is pervasive or domains of hematite, chlorite-sericite and chlorite-carbonate alteration, generate patchy, mottled or pseudoclastic textures. Formerly glassy, porous and permeable facies (e.g. pumice breccia and sandstone, fractured glassy lava) are particularly susceptible to hematite-quartz-sericite-chlorite alteration (Fig. 7.9H). In other cases, hematite overprints lamination and bedding within siltstone. The alteration is generally restricted to a single depositional unit or only a few beds.

Alteration paragenesis

The regional distribution of the feldspar-carbonate-chlorite-sericite-quartz alteration assemblage is consistent with an early diagenetic or regional metamorphic origin. The albitic alteration protected some pumice clasts from diagenetic compaction, suggesting alteration occurred prior to or during diagenesis. The diagenetic compaction foliation is crenulated by the regional cleavage indicating that the alteration was complete prior to regional deformation. The domainal feldspar-carbonate-chlorite-sericite-quartz alteration is similar to that documented by Allen and Cas (1990) in pumiceous breccia units at Rosebery and Hercules, in western Tasmania. Allen and Cas (1990) suggest that two phase albite-K-feldspar and chlorite alteration occurred prior to tectonic deformation during the onset of diagenetic compaction. Elements such as Na, K, Al and Si necessary to form the widespread albite-K-feldspar alteration were probably derived by dissolution of glassy pyroclasts and/or hydrothermal leaching of glass deeper in the volcanic pile. In the Trooper Creek Formation, the widespread development of a bedding parallel stylolitic dissolution foliation in the pumice breccia beds is good evidence for dissolution of glass in the volcanic pile (cf. Allen and Cas, 1990). It is uncertain whether albite and K-feldspar were the first alteration minerals formed or if they replaced earlier minerals such as zeolite.

Hematite-rich alteration and quartz-hematite veins are also regionally distributed and not preferentially associated with mineralisation. The hematite alteration is interpreted to have deposited from low temperature fluids circulating around lavas and intrusions and within the proximal facies association of shallow marine volcanoes (Chapter 6; cf. Einsele et al., 1980; Einsele, 1986; Boulter, 1993a). The convecting pore water leached Fe, Si, and other elements from the glassy volcanic rocks and reprecipitated the iron and silica by conductive cooling and mixing with seawater in the enclosing volcanic package (Chapter 6; cf. Sigurdsson, 1977). The same hydrothermal systems were probably important in contributing elements for regional diagenetic feldspar alteration.

The alteration paragenesis related to mineralisation appears complex and is beyond the scope of the present study. However, some preliminary comments can be made. From drill core, it appears that hydrothermal alteration associated with massive sulfide
deposition started with assemblages of sericite and chlorite which were subsequently overprinted by quartz-sericite alteration. Chlorite veinlets which cut across the quartz-sericite zone are probably late. The chlorite-anhydrite alteration is probably also late as these zones contain relic patches of earlier quartz-sericite alteration and anhydrite veins cut across the massive pyrite-chalcopyrite pipes and other alteration assemblages.

7.4.5 Interrelationships between lithofacies, mineralisation and alteration

The geometry and distribution of the Highway and Reward orebodies were strongly influenced by the position of cryptodomes in the host succession (Figs. 7.2–7.4; 7.8). The Highway pipe is bound by four porphyries: dacite 1 (western margin), rhyodacite 2 (footwall), and rhyolites 3 and 4 (hanging wall) (Figs. 7.3, 7.4). Rhyodacite 2 hosts most of the pyrite-chalcopyrite pipe and the associated stringer mineralisation. The upper contact of the pipe has partially replaced the peperitic lower margin of a rhyolitic partly extrusive cryptodome (rhyolite 4). The massive sulfide and peperite show no indications of mixing. Weak disseminated and semi-massive pyrite-sphalerite-barite mineralisation has replaced the sedimentary and rhyolitic components of the peperite. Intense quartz-sericite alteration and associated disseminated and stringer pyrite mineralisation extend for at least 60 m into the cryptodome. The Highway oxide resource occurs within this alteration zone, along the top of rhyolite 3 (cryptodome) and below an onlapping margin of rhyolite 4 (Fig. 7.3). The southern extent of the Highway pipe is limited by a substantial thickening of rhyolite 4, and termination of rhyodacite 2. Both rhyodacite 2 and significant pyrite-chalcopyrite mineralisation are absent north of 10200N.

The main Reward pyrite-chalcopyrite pipe straddles the boundary between two cryptodomes; rhyodacite 2 (western margin) and rhyolite 2 (eastern margin; Fig. 7.8). An underlying syn-sedimentary sill (rhyodacite 1) occurs below both rhyodacite 2 and rhyolite 2 implying that the cryptodomes are not in faulted contact. Stringer veins extend beneath the pyrite pipe into rhyodacite 2, the top of which is partially replaced by massive pyrite-chalcopyrite. As at Highway, rhyolite 4 forms the hanging wall to the main Reward pipe. On many sections the upper margin of the pipe corresponds to the extrusive top of rhyolite 4 (Fig. 7.4). Several smaller pyrite bodies cut across the overlying volcaniclastic units and intrusions (rhyolites 5–6; Figs. 7.3, 7.4). The southern extension of one small pyrite pipe comprises strata-bound sphalerite-pyrite-barite. The strata-bound mineralisation overlies resedimented hyaloclastite of rhyolite 4 and a distinctive syn-sedimentary sill (rhyolite 5) (Fig. 7.2). As at Highway, the northern extent of the Reward pyrite pipe is marked by the boundary between rhyodacite 2 and a northerly thickening cryptodome (rhyolite 9) (Fig. 7.2). The southern limit of the Reward pipe is poorly
constrained. Stringer vein-style mineralisation extends for at least 90 m north and 150 m south of the Highway and Reward pyrite pipes.

7.4.6 Evidence for a syn-volcanic origin

Laing (1988) and Beams et al. (1989, 1990) proposed two episodes of ore deposition at Highway-Reward: a Cambro-Ordovician event which deposited syn-genetic sphalerite-rich mineralisation, and a Siluro-Devonian syn-deformational episode which produced discordant pyrite-chalcopyrite pipes. Their interpretation of the pyrite-chalcopyrite pipes as syn-deformational was based on: (1) discordance with the host succession; (2) elongation of the pipes and associated stringer veins parallel to the S4 cleavage; and (3) the inferred late-kinematic timing for anhydrite. The textural and structural evidence presented by Laing (1988) and Beams et al. (1989, 1990) is inconclusive and not diagnostic of a syn-kinematic origin for the pyrite pipes. Rather the available evidence suggests that the pyrite-chalcopyrite pipes and strata-bound lenses formed together and are syn-genetic (cf. Large, 1992; McPhie and Large, 1992; Huston, 1992).

Discordance with the host succession does not rule out a syn-genetic origin for the pyrite pipes. Large parts of many ancient and currently forming VHMS deposits display cross-cutting relationships with the host succession (e.g., Bodon and Valenta, 1995; Galley et al., 1995; Humphris et al., 1995). At Highway-Reward, the gradation from Cu-rich massive pyrite mineralisation into marginal sphalerite-rich mineralisation is consistent with a single mineralising event and is similar to the mineralogical zonation seen in many VHMS deposits (e.g., Lydon, 1988a; Large, 1992).

Rare patches of sulfide that crosscut cleavage occur in parts of the Highway-Reward deposit. However, almost all of the mineralisation and alteration have the same tectonic fabric as their host rocks. S4 cleavage and faults are the most prominent structures suggesting most of the ore predated the Siluro-Devonian deformation. Studies of comparable volcanic successions suggest that superimposed structures are often localised within mineralised zones due to the relative incompetence of the altered host rock and some mineralisation types (e.g., Large, 1992). The intensity of the cleavage at the margins of the pyrite pipes, relative to the massive sulfide, reflects the different responses to deformation by the incompetent phyllosilicate-rich alteration zone and competent massive sulfide. It is possible that the pyrite-chalcopyrite pipes and stringer zones were rotated into the D4 structures during deformation (cf. Huston, 1992). Alternatively, the D4 faults reactivated pre-existing structures that controlled the initial deposition of the pyrite pipes (Berry et al., 1992; Huston, 1992) and acted as conduits for rising magma.
The late-kinematic timing of anhydrite proposed by Beams et al. (1989) is not consistent with relationships between the cleavage and the crystals documented in present study. Rather, textures observed within both the pyrite-chalcopyrite pipes and marginal sphalerite-rich mineralisation provide strong evidence for a syn-genetic origin (cf. Huston, 1992). Pyrite with colloform, framboidal, spongy, and snowflake textures is common within VHMS deposits (e.g. Large, 1992). In particular, colloform pyrite is characteristic of open space filling (e.g. Guilbert and Park, 1986), which is inconsistent with a syn-deformational model for mineralisation. Chalcopyrite disease in sphalerite is characteristic, but not diagnostic, of syn-genetic VHMS deposits (Huston, 1992).

The zonation of hydrothermal alteration associated with the massive sulfide bodies is similar to that associated with other Australian VHMS deposits characterised by well developed alteration pipes (e.g. Large, 1992). The distribution of alteration minerals and textures is closely related to inferred initial patterns of permeability and compositional contrasts in the volcanic succession. Alteration associated with syn-deformational mineralisation is more likely to be controlled by fracture and fault patterns than by original volcanic textures (e.g. Oliver, 1996).

Further evidence consistent with a syn-genetic origin for the Highway and Reward orebodies is their occurrence in a regionally altered submarine volcanic succession hosting other VHMS deposits (e.g. Liontown, Berry et al., 1992; Miller, 1996) and the strong stratigraphic control on mineralisation. The regional stratigraphy, depositional setting, sulfide mineralogy and alteration style are also similar to successions that contain relatively undeformed VHMS deposits. Sulfur isotope values for VHMS deposits in the Seventy Mile Range Group overlap with those of Permo-Carboniferous Mt Leyshon gold deposit, but are distinct from Devonian vein-hosted mineralisation (Huston, 1992).

7.4.7 Evidence for sub-seafloor replacement

With the exception of the strata-bound Pb-Zn ores, mineralisation and associated alteration at Highway-Reward are discordant to local bedding, and enclosed by intrusions and volcaniclastic deposits. Relics of coherent rhyolite, dacite and peperite are often preserved within the mineralisation. There is no evidence of mixing with sulfides at the margins of the intrusions and the evidence suggests that the massive pyrite pipes are syn-genetic. The massive sulfides must therefore be interpreted to postdate their host rock and have formed by syn-genetic sub-seafloor replacement and/or infiltration (cf. Large, 1992).
There are two possible interpretations for the strata-bound sphalerite-pyrite-barite mineralisation. Either the lens formed by sub-seafloor replacement of volcanioclastic deposits overlying rhyolite 4, or sedimentation was synchronous with hydrothermal activity and precipitation of sulfides occurred at, above, and below the seafloor during mineralisation. The former is more consistent with the available evidence and the criteria (1-5) for evaluating the role of sub-seafloor replacement in massive sulfide deposition. (1) The mineralisation contains siltstone laminae and significant interstitial sericite and quartz that is presumably altered volcanic (glassy ?) detritus. (2) The underlying and overlying lithic-crystal-pumice beds (1-7 m thick) display bedforms which suggest they were rapidly emplaced as sediment gravity flows. In some drill holes (e.g. REM 122, HMO 89 m), a peperitic syn-sedimentary sills (rhyolite 5) intruded siltstone and sandstone beds directly (0-30 cm) beneath the massive sulfide. The peperite and massive sulfide are not mixed suggesting that emplacement of the sill occurred after deposition of the overlying siltstone and sandstone beds, but prior to deposition of the sphalerite-rich mineralisation. The hanging wall rhyolite intrusion (rhyolite 6) is partially replaced by the massive sulfide lens and displays no evidence of mixing with massive sulfide mineralisation. It was also emplaced prior to the cessation of hydrothermal activity. Apparent graded bedding within some parts of the lens may result from replacement of primary grading in the volcanic precursor. The grading (normal) is defined by pyrite and unlikely to reflect reworking of sulfides on the seafloor because: (a) the pyrite is euhedral and displays no textural evidence of reworking; (b) spongy pyrite overgrows interstitial sericite, indicating that mineralising fluids passed through the sediments after they were deposited (cf. Huston, 1992); and (c) other sulfide minerals within the sample display no grain size trends.

(3-4) The strata-bound lens is zoned and grades from sphalerite-rich massive sulfide northward into progressively more pyrite-rich massive sulfide. The northern part of the lens is a discordant pyrite pod which partially replaces rhyolite 5 and the overlying syn-sedimentary intrusion (rhyolite 6). The pod clearly formed by replacement of the enclosing lithofacies. The gradation into progressively more sphalerite-rich mineralisation probably records a replacement front within the siltstone and sandstone beds which host the lens. Relic siltstone laminae within the sphalerite-rich part of the lens are also preserved within the pyrite-rich massive sulfide.

(5) Intense hanging wall quartz-sericite-pyrite alteration occurs above the pyrite pod and, although less intense, is also well developed above the strata-bound lens. At the southern end of the lens, domains of chlorite-sericite, feldspar-carbonate-chlorite-sericite-quartz and hematite-quartz-sericite-chlorite alteration are more abundant in the hanging wall intrusion (rhyolite 6). The alteration suggests that the ore-forming hydrothermal activity occurred after emplacement of the hanging wall syn-sedimentary intrusion. The zonation in alteration mineralogy is similar to that observed passing out from the margins of the
pyrite pipes into the host lithofacies. Quartz-sericite alteration is more intense above the pipes as these were probably zones of more focussed fluid flow.

The distance below the seafloor at which replacement occurred is difficult to interpret. At Highway, strong quartz-sericite alteration and pyrite veining extends more than 60 m into the hanging wall without any abrupt break in intensity. Consequently, the palaeoseafloor position at the time of mineralisation is not preserved. Prior to mining, barite-rich volcanic rocks were exposed at the surface (Kay, 1987). In VHMS systems, barite typically forms at the interface between ascending hydrothermal solutions and cold seawater and/or seawater saturated strata (e.g. Large, 1992). This suggests that the barite-rich mineralisation marks a near seafloor position.

At Reward, intense weathering and oxidation have obscured alteration and lithofacies above the main pipe hampering interpretation. On some sections, the top of the main pyrite-chalcopyrite pipe coincides with the extrusive top of rhyolite 4. It is possible that the top of the reworked hyaloclastite units overlying rhyolite 4 was the seafloor at the time of ore formation, and replacement occurred right up to (and/or down from) the seafloor. Above the main pyrite pipe, small discordant pyrite pipes cut across the hanging wall volcaniclastic units and intrusions. Consequently, replacement and infiltration may have taken place beneath a series of seafloor positions as the host succession accumulated. Alternatively, all of the host rocks may have been deposited prior to significant hydrothermal activity, and replacement occurred at a number of stratigraphic positions below a seafloor position which is not preserved.

7.4.8 Fluid chemistry

Fluid inclusion studies did not form part of the present research and there is little suitable material available for this type of research. Studies of VHMS deposits suggest that zinc-rich zones deposit at temperatures of 225-300°C, whereas copper-rich mineralisation deposits from fluids in excess of 300°C (e.g. Large, 1992). At Highway-Reward, the copper-rich mineralogy of the pipes suggests that they deposited at >300°C. The sphalerite-rich mineralisation deposited at less than 300°C (cf. Huston, 1992). The gangue mineralogy and iron-rich composition of sphalerite in the pyrite pipes suggests that this mineralisation precipitated from reduced and acid fluids (Huston, 1992). Conversely, the strata-bound mineralisation deposited from oxidised and slightly more alkaline fluids which evolved from those which formed the pyrite-chalcopyrite mineralisation (Huston, 1992). The range of δ^34S values for the pyrite and sphalerite
Mineralisation is consistent with a relatively low $\Sigma S0_4^{2-}/\Sigma H_2S$ ratio, and suggests that redox reactions were not important during ore genesis (Huston, 1992).

7.4.9 A genetic model for the formation of the Highway-Reward deposit

The syn-genetic Highway-Reward deposit formed within a submarine, syn-sedimentary intrusion-dominated volcanic centre. Figure 7.10 is a schematic reconstruction showing successive stages in the evolution of the volcanic centre. At the onset of significant hydrothermal activity, the centre had the configuration shown in frame B or D. Intensification of the hydrothermal system followed the main phase of intrusion-dominated volcanism. The magmatism may have acted as a heat engine during seawater convection causing metals to be leached from the volcanic pile and/or contributed to the ore forming fluid. Initially, hydrothermal fluids may have ascended growth faults beneath the Highway and Reward positions (cf. Berry et al., 1992). Within a few hundreds of metres of the sea floor, hydrothermal fluid flow was possibly complicated by the intersection of faults and stratigraphic zones of lower permeability. Hydrothermal fluids were focussed within, but close to, the steep margins of cryptodomes that intruded the host volcano-sedimentary deposits while they were wet and poorly consolidated. Emplacement of the syn-sedimentary intrusions had transformed the sediments into relatively impermeable rocks compared to the fractured glassy margins of the cryptodomes. The indurated sediments may have prevented the development of broad convection cells and focussed ascending hydrothermal fluids into fractures within the host sequence and along the glassy margins of the cryptodomes. The porosity and permeability of the cryptodomes allowed infiltration of seawater which mixed with the high temperature ($>300^\circ$C) hydrothermal fluid, promoting precipitation of pyrite-chalcopyrite in the developing pipes. Barite which is intergrown with the massive sulphide records the role of seawater in massive sulphide deposition (cf. Large, 1992). An advancing pyrite front gradually moved out through the host succession replacing rhyolite–dacite, volcanioclastic units and sediment. At Highway, a rhyolitic partly extrusive cryptodome (rhyolite 4) formed a barrier to ascending ore fluids and replacement occurred below and within it’s peperitic base. Although massive sulphide deposition was largely limited by the contact zone, fluid flow extended into the cryptodome, producing strong hanging wall quartz-sericite alteration enclosing pyritic stringer and disseminated styles of mineralisation. The outer alteration zones record the mixing of convecting seawater and ascending hydrothermal fluids along the margins of the pipes.

At Reward, it is possible that the hydrothermal system remained active during a small hiatus in volcanism, and that the pyrite pipes form a stacked system beneath two successive seafloor positions. If so, the extrusive top of rhyolite 4 was the seafloor
Mineralisation

Subaqueous emplacement of feldspar-quartz-bearing pyroclastic flow deposits and syn-sedimentary sills (R1 and R2). The intrusions influenced seafloor topography and partially dewatered the host succession.

Intrusion of a rhyolitic cryptodome (R3) and partial excision of a dacitic syn-sedimentary sill (R4) through the seafloor. Onset of significant hydrothermal activity. Hydrothermal fluids ascend growth faults and are focused along the margins of the overlying cryptodomes. Fluids mix with seawater circulating through the gassy, porous and permeable margins of the cryptodomes and volcaniclastic deposits. Sulfides replace the enclosing strata within a strongly altered and veined zone.

Intensification of the hydrothermal system to a maximum of >300°C. An advancing pyrite-chalcopyrite replacement front moves through the host strata overprinting earlier sulfide minerals. Ponding of hydrothermal fluids beneath rhyolite 4 pronounces sulfide deposition. Hydrothermal fluids were locally expelled in the water column.

Intrusion of syn-sedimentary sills (R5-R7) into resedimented hyaloclastite of rhyolite 4, and the overlying deposits. At Reward, the pyrite pipe eventually penetrated rhyolite 4. Continued hydrothermal activity deposits pipe- and vein-style mineralisation above the main pipe. Hydrothermal fluids moving out from the margins of the pipes. Minor stratabound Pb-Zn-Ba-rich lenses deposited by sub-seafloor replacement of permeable volcaniclastic deposits.

An advancing pyrite-chalcopyrite front progressively replaces earlier low-temperature (<300°C) sphalerite (Sp)- and barite (Ba)-rich mineralisation.

Figure 7.10. Schematic representation of successive stages in the genesis of the Highway and Reward orebodies.
position at the time the main pyrite pipe was forming, and replacement occurred up to near the seafloor (Fig. 7.10B–C). The Highway pipe may have formed beneath the same seafloor position, although it is not preserved above the orebody. Metals carried to the seafloor by the hydrothermal fluids were dispersed in the water column and incorporated into distal sediments. It remains possible that some strata-bound sphalerite-rich mineralisation deposited at this time. Volcaniclastic deposits and intrusions, which were emplaced above rhyolite 4 during renewed magmatism, were replaced by small discordant pyrite pipes below a new seafloor position (Fig. 7.10D). Alternatively, all of the mineralisation could have been emplaced within a single alteration pipe that extended through pre-existing volcanic units, beneath a palaeoseafloor position which is not preserved or is marked by pumiceous units at the top of the sequence (Fig. 7.10D). Poorly focussed fluid flow to the west and east of the Highway-Reward deposit produced zones of weak to moderate sericite-silica alteration.

Dispersed fluids which escaped from the margins of the pyrite pipes into the fractured and glassy host rock, mixed with seawater and deposited a broad halo of disseminated and patchy sphalerite-galena-barite mineralisation. Similar mineralisation may have deposited during initial phase of hydrothermal activity (cf. Large, 1992). A small strata-bound sphalerite-pyrite-barite-rich lens formed above the main pyrite pipe by lateral migration of low temperature (<300°C) hydrothermal fluids through permeable units (Fig. 7.10D). The northern extent of the lens became progressively more pyrite-rich, being deposited from higher temperature (>300°C) fluids. An advancing pyrite-chalcopyrite front gradually moved out through the sphalerite-barite-rich lens. The common development of chalcopyrite disease suggests that lead and zinc leached at the advancing copper front were reprecipitated within the volcanic precursor along a Pb-Zn-Ba front. Sphalerite-galena-barite veins and disseminations probably formed at the same time and continued to deposit in the waning stages of hydrothermal activity, as the fluids cooled from a peak of >300°C down to <300°C. Anhydrite was deposited in zones of intense chlorite alteration as hydrothermal fluids mixed with seawater along the margins of the pipes.

This model suggests that there were three principal controls on the location and formation of the Highway-Reward VHMS deposit: (1) a progressively evolving submarine intrusion- and lava-dominated volcanic centre. The high geothermal gradient associated with magmatism provided heat to drive the hydrothermal fluid flow, and the lavas and intrusions focussed hydrothermal fluids along the discrete mineralising pathways; (2) an impermeable barrier promoted sub-seafloor ponding of hydrothermal fluids and replacement; and (3) the hydrothermal system remained active during continued volcanism, dewatering of the sediment pile by synt-sedimentary intrusions, and sedimentation.
7.5 Discussion

7.5.1 The importance of lithofacies in sub-seafloor replacement

Sub-seafloor deposition of massive sulfides involves dissolution, replacement, infilling of pore space, and precipitation of minerals along fluid pathways. Consequently, the shapes, dimensions and distribution of hydrothermal circulation (and therefore the mineralisation) are closely related to the initial patterns of permeability and compositional contrasts in the volcanic host rock.

Many sub-seafloor replacement ores occur within rapidly emplaced units, particularly pumiceous facies (e.g. Khin Zaw and Large, 1992; Allen, 1994; Allen et al., 1996b). The originally highly porous, permeable, water-saturated, and glassy nature of these deposits make them favourable host rocks for sub-seafloor replacement deposits. Ascending hydrothermal fluids will be poorly focussed and permeate through the substrate to produce widespread strata-bound alteration and lens- or sheet-style massive sulfide mineralisation (cf. Large, 1992). Sulfide replacement of this type probably commences at the interface between the ascending hydrothermal fluid and overlying cold, seawater-saturated strata (e.g. Khin Zaw and Large, 1992; Allen et al., 1996b).

Within less permeable, syn-sedimentary intrusion- and lava-dominated volcanic piles fluids are likely to by focussed along faults, local autoclastic breccia intervals, or within the fractured glassy margins of lavas and intrusions. Under these circumstances, well focussed fluid flow gives rise to lens- or pipe-shaped massive sulfide mineralisation and well developed, zoned alteration pipes (e.g. Large, 1992). Massive sulfide deposition probably commences beneath a relatively impermeable barrier (e.g. massive crystalline/devitrified lava) and grows downward by mixing with seawater convecting through the fractured, glassy, porous and permeable parts of lavas and shallow intrusions and volcaniclastic deposits. In the absence of a barrier, ascending hydrothermal fluids are more likely to reach the seafloor, and may well form a seafloor massive sulfide deposit.

The distance below the seafloor at which infiltration and replacement take place is rarely well constrained. In some cases, mass-flow deposits directly overlying the host rocks to sub-seafloor replacement deposit contain clasts of the massive sulfide (e.g. Hercules, Allen and Hunns, 1990), suggesting that replacement and infiltration probably occurred within a few metres of the seafloor. The upper few tens of metres in the volcano-sedimentary pile are probably the favoured position for replacement, as sediments are wet, porous and poorly consolidated in this zone, and at greater depths become progressively more compacted (e.g. Einsele, 1986) and less amenable to replacement and infiltration by hydrothermal fluids. Ascending hydrothermal fluids will meet and mix with
cold seawater before reaching the seafloor, and can precipitate some of their metals subsurface.

7.5.2 Models for sub-seafloor replacement

In this section, the various circumstances under which sub-seafloor replacement and infiltration may develop and the character of the resulting mineralisation are considered. The scenarios depend on the sedimentation rate and whether or not the host succession is dominated by relatively poorly porous rocks (e.g. lavas and shallow intrusions) or by incompetent, very porous deposits (e.g. pumiceous units). Figure 7.11 summarises the main attributes of possible sub-seafloor replacement style deposits. It does not aim to present a comprehensive account of all possible scenarios for sub-seafloor replacement, but simply to highlight the spectrum of deposit styles which may form and their relationship to the enclosing strata.

7.5.3 Implications for mineral exploration

Pipes or plumes of strong hanging wall alteration are characteristic of sub-seafloor massive sulfide deposits, and can extend from tens to hundreds of metres into the hanging wall volcanic succession. At Highway-Reward, this style of alteration hosts significant disseminated and stringer vein-style mineralisation and several small massive sulfide lenses. Intense hydrothermal alteration and veining are more typical of footwall alteration and stringer zones beneath seafloor massive sulfide deposits. Hanging wall alteration with footwall affinities has major implications for mineral exploration. Exploration requires consideration of the possibility: (1) that massive sulfides occur beneath alteration and veining that would otherwise be regarded as the stringer zone to a seafloor deposit; (2) that massive sulfide lenses, pipes, veins and/or disseminations may form a stacked system; and (3) that the seafloor position at the time of hydrothermal activity may be located tens to hundreds of metres above the position of sub-seafloor mineralisation.

This analysis reveals that the seafloor hydrothermal systems responsible for the Highway-Reward massive sulfide mineralisation operated within a small non-explosive, syn-sedimentary intrusion-dominated volcanic centre. The Handcuff massive sulfide mineralisation is hosted by a similar but separate lava- and intrusion-dominated volcanic centre. The volcanic facies associations which characterise syn-sedimentary intrusion-dominated volcanic centres are as prospective as parts of the Trooper Creek Formation dominated by lavas, sediments and volcaniclastic deposits. The spatial and temporal relationship of the massive sulfide mineralisation with volcanic centres, suggests that the mineralising hydrothermal systems were intimately and possibly genetically related to the
mineralisation associated with emplacement of the volcanic centres. The magma may have acted as the heat engine driving convection and/or contributed to the hydrothermal fluid (cf. Large, 1992).

- Sub-seafloor replacement of rapidly emplaced mass-flow deposits
- Host commonly syn-eruptive & pumiceous deposits
  (e.g. Rosebery, Allen, 1994; South Hercules, Khin Zaw & Large, 1995; Litotown, Miller, 1996; Langdals & Langsete, Allen et al., 1997b)
- Sub-seafloor sulfide replacement front within rapidly emplaced units
- Synchronous sedimentation & sulfide precipitation both above & below the seafloor (e.g. Currawong, Bodon & Valenta, 1995; Ansil, Galley et al., 1995)
- Sub-seafloor replacement of a single rapidly emplaced unit (e.g. Renströma, Allen et al., 1997b)
- Synchronous sedimentation & massive sulfide deposition by replacement, infiltration & exhalation
- Seafloor massive sulfide deposits
- Burial: replacement of volcaniclastic deposits during ongoing hydrothermal activity (e.g. Que River, Large et al., 1988)

Figure 7.11. Schematic representation of the various scenarios in which ascending hydrothermal fluids can interact with the enclosing seawater-saturated strata and deposit mineralisation by replacement and infiltration of pre-existing volcaniclastic or sedimentary deposits.
Figure 7.11 continued. Cartoon showing various circumstances under which sub-seafloor replacement mineralisation may deposit. In these examples the host successions also includes lavas and intrusions.
7.6 Summary

The Highway-Reward deposit is hosted in the proximal facies association of a submarine (below storm wave base) silicic, syn-sedimentary intrusion-dominated volcanic centre. The characteristic facies associations of these volcanoes can be used as vectors to locate prospective parts of the volcanic succession. Pyrite-chalcopyrite pipes are surrounded by a zone of pyrite-sphalerite-galena-barite mineralisation. The pyrite-chalcopyrite pipes and sphalerite-rich mineralisation formed at the same time and are syn-volcanic, sub-seafloor replacements of the host sediment, syn-volcanic intrusions, partly extrusive cryptodomes, and volcaniclastic units. Alteration associated with mineralisation is zonally arranged around the pipes so that alteration in the footwall and hanging wall is similar. Intense quartz-sericite alteration and pyrite veins extend more than 60 m above the Highway orebody. The location, distribution, form and shape of massive sulfide mineralisation and alteration are closely related to inferred initial patterns of permeability in the host rocks. Pipe-shaped massive sulfide deposits are likely to form in host successions dominated by relatively poorly porous facies (e.g. lava- and intrusion-dominated volcanic centres).
Chapter 8

Synthesis: a palaeogeographic reconstruction of the Mount Windsor Formation and Trooper Creek Formation during the Cambro-Ordovician
Chapter 8

Synthesis: a palaeogeographic reconstruction of the Mount Windsor Formation and Trooper Creek Formation during the Cambro-Ordovician

8.1 Introduction

A detailed analysis of the Mount Windsor Formation and Trooper Creek Formation, in the area between Coronation homestead and Trooper Creek prospect (approximately 15 km strike length), has helped clarify the facies architecture of the Cambro-Ordovician Seventy Mile Range Group. Significant advances have been made in understanding the character and geometry of the submarine silicic to intermediate volcanic succession, and assessing genetic links between volcanic, alteration and mineralisation processes. The research addresses a number of specific problems which are relevant to understanding other ancient submarine volcanic successions. These include:

(1) the lithofacies and volcanic history of andesitic, shallow-marine strombolian volcanoes (Chapter 4);
(2) constraints on the lithofacies associations and volcanic (and sub-volcanic) history of submarine silicic lava- and intrusion-dominated volcanic centres (Chapter 5);
(3) assessment of the influence of sub-volcanic intrusions and domes on the genesis, localisation and evolution of syn-genetic submarine hydrothermal systems (Chapters 5-7);
(4) volcanic influences on the formation of iron oxide-silica units ("ironstones") and relationships between ironstones and massive sulfide mineralisation (Chapter 6);
(5) assessment of the diagnostic/critical evidence for sub-seafloor replacement style massive sulfide mineralisation, and the role of lithofacies in determining the character of the resulting deposit (Chapter 7).

The principal findings of this research are an understanding of the Cu-Au-rich Highway-Reward massive sulfide deposit and clarification of lithofacies and stratigraphy of the Seventy Mile Range Group.
8.2 Palaeogeographic reconstruction

In this section, the various elements of the preserved facies architecture are combined to reconstruct the Cambro-Ordovician palaeogeographic setting and volcanic history during accumulation of the Mount Windsor Formation and Trooper Creek Formation in the central part of the Seventy Mile Range Group.

8.2.1 Puddler Creek Formation

The oldest part of the Seventy Mile Range Group is the Puddler Creek Formation. The formation comprises a thick (9 km, Henderson, 1986) sedimentary succession and minor basaltic to andesitic volcanics. The siltstone and sandstone units are dominantly continentally-derived but contain minor volcanic detritus. The lithofacies associations suggest that the depositional setting for the Puddler Creek Formation was submarine (below storm wave base).

8.2.2 Mount Windsor Formation

The Mount Windsor Formation marks the onset of voluminous silicic volcanism in the Seventy Mile Range Group. The formation comprises a thick sequence of rhyolitic volcanic rocks with minor dacite and andesite. Associations of coherent facies and autoclastic breccia facies, 100-500 m thick, form lavas and intrusions. The feeder dykes for the extrusive units have not been identified. However, mapping of the unit boundaries suggests that many lavas probably extended less than a few kilometres from their vents. The paucity of intervening volcano-sedimentary units implies that the lavas were erupted rapidly from adjacent vents or from fissures, and probably constructed significant topography that strongly influenced sedimentation patterns. Henderson (1986) argued that thickening of the Mount Windsor Formation between Highway-Reward and Sunrise Spur indicated an eastern source for the formation. As the lavas and intrusions are proximal to vent, it is more likely that the rhyolitic and dacitic units were erupted from separate intrabasinal vents distributed along the length of the basin, and that volcanic centres in the east were more productive.

The lavas and intrusions yield little unambiguous information about the depositional setting. Hyaloclastite associated with the lavas provides evidence for emplacement in a subaqueous environment. West of the study area at Mount Windsor, the Mount Windsor Formation includes resedimented volcaniclastic units interpreted to have been deposited from sediment gravity flows in a submarine environment (Berry et al., 1992). At
8.3. Thalanga (Fig. 1.1), VHMS-style mineralisation is partly hosted by volcaniclastic rocks of the Mount Windsor Formation, suggesting that this part of the Mount Windsor Formation accumulated below storm wave base in a submarine setting (Hill, 1996).

8.2.3 Trooper Creek Formation

Water depth and depositional setting

The Trooper Creek Formation comprises coherent lithofacies and compositionally and texturally diverse volcaniclastic lithofacies which are intercalated with lavas, intrusions and non-volcanic sedimentary facies (Chapter 3). The sedimentary facies contain marine fossils. Voluminous hyaloclastite, syn-sedimentary sills and local pillow lavas are also consistent with a submarine setting (cf. Berry et al., 1992). In the study area, the widespread occurrence of turbidites and thick intervals of siltstone suggests that most of the succession accumulated below storm wave base. Depth of storm wave base varies in modern environments from 10 to 200 m (Johanson and Baldwin, 1996). At Highway East, the presence of proximal bomb-rich andesitic breccia facies implies that the depositional environment was, at least locally, less than about 500 m deep (Chapter 5). Combined, the lithofacies associations, fossils and regional context imply that depositional setting of the Trooper Creek Formation was submarine, below storm wave base, and possibly not in extremely deep water. The exception is at Trooper Creek prospect where the depositional setting for the upper part of the Trooper Creek Formation was shallow marine (Chapter 4). Stromatolites, traction current deposits indicative of wave activity, and evaporitic minerals collectively suggest that this part of the succession was deposited above storm wave base and may have been temporarily emergent. Further work is required to determine if the facies associations at Trooper Creek prospect record a local shoaling depositional environment or regional shallowing of the basin east of Trooper Creek.

Lithofacies and volcanic history

The large volume rhyolitic effusive eruptions of the Mount Windsor Formation were followed by a phase of rhyolitic to basaltic effusive and explosive volcanism, and sedimentation. This phase in the history of the Seventy Mile Range Group is recorded by the Trooper Creek Formation. The Trooper Creek Formation is subdivided into two members, the Kitchenrock Hill Member and the overlying Highway Member. The stratigraphic subdivision is based on lithological variations which reflect different provenance characteristics.
**Kitchenrock Hill Member**

The volcaniclastic facies in the Kitchenrock Hill Member contain abundant rounded rhyolitic and dacitic clasts with geochemical and petrographic characteristics that suggest they were sourced from the Mount Windsor Formation (Chapter 3). Rounding occurred in a high-energy environment (above storm wave base) prior to redeposition, suggesting that the source areas were subaerial to shallow marine and that the volcaniclastic mass flows transgressed a shallow-water environment (Fig. 8.1). Alternatively, the clasts may have been collected somewhere along the flow path from already deposited material in a below-storm-wave-base environment. The sources of the rounded clasts are located outside the study area.

![Palaeogeographic reconstruction of the Trooper Creek Formation and source terrain during deposition of the Kitchenrock Hill Member.](image)

Figure 8.1 Palaeogeographic reconstruction of the Trooper Creek Formation and source terrain during deposition of the Kitchenrock Hill Member. 1 - rhyolite and dacite domes; 2 - lava; 3 - eroded domes and intrusions; 4 - syn-volcanic fault related to basin subsidence; 5 - talus adjacent to fault scarp; 6 - explosive eruptions from subaerial to shallow subaqueous volcano; 7 - syn-eruptive sediment gravity flow deposit incorporating clasts derived from Mount Windsor Formation (MWF) rhyolite and dacite; 8 - seafloor topography created by the Mount Windsor Formation domes; 9 - syn-sedimentary intrusion. PCF - Puddler Creek Formation.
Volcaniclastic facies in the Kitchenrock Hill Member also contain juvenile components. The actual sources for the juvenile components have not been identified, are not exposed or have been eroded away. However, the high proportion of pumice, shards, crystals and crystal fragments within the volcaniclastic units implies that the sediment gravity flows were sourced from explosive magmatic or phreatomagmatic eruptions, and also suggests that the volcanic centres were shallow marine or subaerial (e.g. McBinney, 1963; Chapter 3). The crystal composition (feldspar±quartz) suggests a dacitic to rhyolitic provenance. Thickness variations within the Kitchenrock Hill Member reflect palaeotopography on the depositional surface that was generated by the underlying Mount Windsor Formation lavas and/or growth faults.

Highway Member
The Highway Member comprises compositionally and texturally diverse syn-eruptive volcaniclastic facies, syn-sedimentary intrusions and lavas which are intercalated with volcanic and non-volcanic sedimentary facies. Rounded clasts of Mount Windsor Formation provenance are absent in the Highway Member. This suggests that either: (1) the source areas for clasts of Mount Windsor Formation rhyolite and dacite were completely eroded prior to deposition of the Highway Member; or (2) that basin subsidence and/or volcanic accumulation caused a change from a mixed volcanic source (i.e. Kitchenrock Hill Member) to syn-eruptive sedimentation during deposition of the Highway Member (Trooper Creek Formation).

Syn-eruptive resedimented volcaniclastic facies in the Highway Member also provide a record of the character and setting of the volcanic activity in the source areas. The source areas probably included both intrabasinal and extrabasinal volcanic centres (Fig. 8.2; Chapters 3-5). The pumiceous and crystal-rich facies contain pyroclasts that originated from magmatic and/or phreatomagmatic eruptions at extrabasinal, basin margin or shallow subsaquesous intrabasinal volcanic centres. The crystal assemblage (feldspar±quartz) suggests a dacitic to rhyolitic magma composition for these facies. However, the actual source areas for the sediment gravity flows have not been identified. Intercalated siltstone and sandstone turbidites are dominated by volcanic detritus (principally shards, crystal fragments, devitrified ash) but may contain non-volcanic components (phyllite, meta- chert, tourmaline, white mica) from basement sources similar to those of the Puddler Creek Formation and Rollston Range Formation. This implies that deposition of non-volcanic detritus was the ambient sedimentation style and was interrupted during accumulation of the Mount Windsor Formation and Trooper Creek Formation.

The high proportion of scoria and bombs in the andesitic breccia facies association (Highway Member) suggests that pyroclasts were sourced from intrabasinal strombolian eruptions (Chapter 4). These eruptions built ephemeral scoria cones subject to collapse...
and resedimentation, delivering scoria and bombs into deeper water flanking environments. The bomb-rich breccia facies accumulated in close proximity to source vents (cf. Staudigel and Schmincke, 1984). In the western part of the study area, intercalated turbidites and siltstone imply that the vents and depositional environment were below storm wave base and in water probably less than 500 m deep (Chapter 5). The maintenance of a marine environment during volcanic aggradation implies that basin subsidence was probably important during accumulation of the Trooper Creek Formation.

Figure 8.2 Palaeogeographic reconstruction of the Trooper Creek Formation and subaerial source area during deposition of the Highway Member at the time of mineralisation. 1 - dome; 2 - partly extrusive cryptodome; 3 - resedimented intrusive hydroclastic breccia; 4 - resedimented hyaloclastite; 5 - lava; 6 - syn-sedimentary intrusion; 7 - submarine scoria cone; 8 - resedimented andesitic scoria breccia; 9 - microbialite; 10 - ponding of syn-eruptive sediment gravity flow deposit against lava dome; 11 - syn-eruptive resedimentation of pyroclastic debris by subaqueous mass flows; 12 - water-settled ash fall; 13 - subaerial explosive rhyolitic and dacitic volcanoes sourcing pyroclasts to the basin; 14 - massive sulfide body; 15 - syn-volcanic faults. KRHM - Kitchener Hill Member; MWF - Mount Windsor Formation.
To the east, around Trooper Creek prospect, the andesitic volcanoism constructed an edifice which shoaled to above fairweather wave base and may have been temporarily emergent (Chapter 4). The post-eruptive history of the andesitic volcanic centre began with partial collapse of the edifice, creating a stable shallow marine surface for deposition of the overlying dacitic volcano-sedimentary facies and microbialites. Volcanic siltstone and subordinate vitric-rich sandstone that are increasingly abundant in the overlying succession (Highway Member), reflect a return to relatively quiet, probably deeper water conditions in response to compaction and/or tectonic subsidence. In the west near Highway-Reward, Highway East and Coronation homestead there is no evidence for shallowing of the depositional environment towards the top of the Highway Member. This suggests that basin subsidence was regionally uneven, although largely kept pace with volcanic aggradation in the basin.

Syn-sedimentary sills, cryptodomes, partly extrusive cryptodomes and associated in situ and resedimented autoclastic facies are an important component of the Highway Member. These form the proximal facies association from intrabasinal, intrusive and extrusive, non-explosive volcanism (Chapter 5). In the Highway-Reward to Trooper Creek area, rising magma that encountered unconsolidated, water-saturated sediment commonly remained sub-surface and was emplaced as syn-sedimentary sills and cryptodomes rather than erupting as lavas and domes. The intrusions probably significantly modified the physical environment and pore fluid properties in these parts of the basin (cf. Einsele et al., 1980; Delaney, 1982; Duffield et al., 1986; Hanson, 1991; McPhie, 1993; Davis and Becker, 1994; Brooks, 1995). Dewatering, induration, disruption of bedding and low grade metamorphism commonly accompanied intrusion. The lavas, cryptodomes and sills influenced seafloor topography and therefore sedimentation, causing lateral facies and thickness variations (cf. Yamamoto et al., 1991; Davis and Villinger, 1992). The Highway-Reward VHMS deposit formed in the proximal facies association of one small, syn-sedimentary intrusion-dominated volcanic centre. At Highway-Reward, pyrite-chalcopyrite pipes and associated marginal sphalerite-galea-barite mineralisation are syn-volcanic replacements of the host sediment, syn-volcanic intrusions, cryptodomes and volcaniclastic deposits (Chapter 7). Syn-volcanic faults may have acted as conduits for ascending hydrothermal fluids and also influenced the position of the intrusion-dominated complex (cf. Berry et al., 1992; Huston, 1992).

In the study area, quartz-hematite pods and lenses ("ironstones") occur throughout Trooper Creek Formation (Chapter 6). Type 1 ironstones are characterised by positive Eu anomalies, anomalous Zn, Pb, Ag and Au, and are geochemically similar to ironstones which mark the ore equivalent of the Thalanga VHMS deposit (45 km to the west). These units are exposed at Handcuff prospect and are clear targets for further exploration. Type 2 ironstones are characterised by negative Eu and Ce anomalies, are regionally distributed
and are not associated with mineralisation. Type 2 ironstones are interpreted to have deposited from short-lived, low temperature, local hydrothermal systems in the proximal facies associations of intrusion- and lava-dominated volcanic centres and shallow marine volcanoes in the Highway Member (cf. Einsele et al., 1980; Einsele, 1986; Boulter, 1993a). Circulating fluids leached iron, silica and other elements from the glassy volcanic rocks and reprecipitated the iron and silica in the enclosing volcanic succession in response to conductive cooling and mixing with seawater (cf. Sigurdsson, 1977).

8.2.4 Rollston Range Formation

The top of the Highway Member corresponds to the end of intrabasinal volcanism and volcanic-dominated sedimentation. This was then followed by a phase of post-eruptive erosion and reworking that probably affected both basin margin and intrabasinal centres, accompanied by rare effusive and explosive eruptions, leading to deposition of the Rollston Range Formation. The lithofacies characteristics and fossils in the Rollston Range Formation imply a submarine depositional environment, below storm wave base (cf. Henderson, 1986; Chapter 3). The siltstone and sandstone units are dominated by volcanic quartz and feldspar suggesting a largely felsic volcanic source. The remaining fragment population is clearly non-volcanic (phyllite, polycrystalline quartz, detrital mica, tourmaline) and implies input from granitic and deformed basement sources. Feldspar grains are relatively unaltered and angular suggesting only minor transport and reworking. The other components are variably rounded indicating reworking in a high-energy environment prior to redeposition by sandy, high-concentration turbidity currents, and that the source areas were at least partly subaerial or shallow marine. In addition to the reduced volcanic input, tectonic uplift may have contributed to the change in the depositional style towards the top of the Trooper Creek Formation. This uplift might have exposed Precambrian basement, feeding significantly larger volumes of non-volcanic material into the basin during deposition of the Rollston Range Formation. Siltstone and sandstone units with similar provenance to those in the Rollston Range Formation also occur in the Trooper Creek Formation.

8.3 The Cambro-Ordovician tectonic setting of the Seventy Mile Range Group

Henderson (1986) and Stolz (1995) propose that the Seventy Mile Range Group is the fill of a back-arc basin developed on thinned Precambrian basement flanking a continental margin volcanic arc (Chapter 2). A reconstruction of the tectonic setting for the Mount Windsor Subprovince is beyond the scope of this study. However, the volcanic facies
analysis presented here provides constraints on the tectonic setting of the Trooper Creek Formation.

1. Local erosion of the Mount Windsor Formation occurred during deposition of the Kitchenrock Hill Member (Trooper Creek Formation). Erosion of the Mount Windsor Formation may record shoaling of the Mount Windsor Formation to above storm wave base in some parts of the basin and/or a period of differential uplift and subsidence prior to, or during deposition of the Kitchenrock Hill Member.

2. The palaeogeography of the Trooper Creek Formation comprises a marine setting (both above and below storm wave base), that flanked a subaerial or shallow water silicic volcanic terrain.

3. During deposition of the Trooper Creek Formation, the rapid accumulation of strata and continued subsidence to around or below storm wave base (e.g. Chapter 4) implies strong crustal extension (cf. Allen et al., 1996b). Evidence for local, temporary shoaling of the volcanic succession implies that basin subsidence was regionally heterogeneous but largely kept pace with volcanism and intrusion of magma.

4. Although palaeocurrent indicators are lacking in the Trooper Creek Formation, the shallowing of the depositional environment in the upper part of the Highway Member at Trooper Creek suggests a local westerly dipping palaeoslope in the upper part of the Trooper Creek Formation, between Coronation homestead and Trooper Creek prospect. The regional extent and significance of this shallowing-upward trend within the Trooper Creek Formation is unclear and requires further work.

5. Volcanism waned during the early Ordovician and deposition of the Rollston Range Formation occurred during post-eruptive erosion of a largely felsic volcanic source (Trooper Creek Formation ?, volcanic arc; cf. Henderson, 1986) and erosion of granitic and deformed basement sources. Uplift of the basement, possibly in response to a decrease in crustal extension, may have occurred at this time.

8.4 A modern analogue for the Trooper Creek Formation

The lithofacies, formative volcanic processes and volcano types represented in the Cambrian Mount Windsor Subprovince have analogues in modern successions. The offshore extension of the Taupo Volcanic Zone (TVZ) is proposed as an analogue of the Trooper Creek Formation and associated source terrain. However, it is important to note that the comparisons have several limitations.
(1) The interpretations presented here are based on mapping only a central segment of the Seventy Mile Range Group. Regional geological studies concentrating on identifying key facies associations as a means of reconstructing the facies architecture of the entire belt have yet to be undertaken. This approach will provide important insights into the evolution of the basin, source volcanoes and provenance.

(2) The combined effects of erosion and burial by younger deposits means that the understanding of the Trooper Creek Formation is incomplete, and biased towards the preserved submarine record. In the TVZ, the opposite is true, as few geological, geophysical or geochemical studies of the offshore extension of the TVZ have been undertaken. Neither is the subsurface stratigraphy of the TVZ well constrained for much of the TVZ.

(3) The original orientation and dimensions of the Seventy Mile Range Group depocentre are poorly constrained and complicated by faulting and folding (e.g. Henderson, 1986; Stoiz, 1995). The subprovince presently extends east-west for approximately 165 km, is oriented east-west, and may be more than 12 km thick (Henderson, 1986).

The Taupo Volcanic Zone (TVZ) in the North Island of New Zealand is a region of major Pliocene to Quaternary calc-alkaline volcanism and crustal extension resulting from subduction of the oceanic Pacific Plate beneath the continental Australian Plate (Cole, 1990; Wilson et al., 1995; Fig. 8.3). The TVZ is a NNE trending zone of vent and caldera structures extending for 200 km onshore from Ohakune in the south, northward through the central North Island, and offshore for around 150 km (e.g. Wright, 1992). The boundary between the TVZ and the Kermadec Ridge-Havre Trough system (its counterpart to the NNE) is marked by a NW-SE-trending structural discontinuity which is coincident with, but separate from, the Vening Meinesz Fracture Zone (Gamble et al., 1993). The Kermadec Ridge-Havre Trough system and the TVZ are offset sinistrally by approximately 50 km (Wright, 1992; Gamble et al., 1993; Fig. 8.3). The TVZ is about 20-60 km wide, and is bounded to the east and west by faults, and dominantly Mesozoic greywacke basement. Little is known of the basement beneath the TVZ as it is buried beneath 2-3 km of volcanic units (Rogan, 1982). A total volume of 15 000-20 000 km$^3$ of volcanic products is estimated to have erupted from the TVZ (Houghton et al., 1995). At present the TVZ is rifting at rates between 7 and 18 mm/a (Wilson et al., 1995) and subsiding at approximately 1 to 2 mm/a (Nairn and Beanland, 1989; Wright, 1992). Episodes of uplift and erosion affected the TVZ between c. 1 Ma and 0.32 Ma (Wilson et al., 1995).

The TVZ is divided into three segments (Wilson et al., 1995; Houghton et al., 1995). The northeast and southwest segments contain andesitic to dacitic composite volcanoes and lack calderas, whereas the central segment (125 km long) is dominantly rhyolitic (Wilson
et al., 1995; Houghton et al., 1995). The northeast segment extends offshore. The central TVZ includes at least eight caldera volcanoes. These account for 34 caldera-forming eruptions in the approximately 1.6 Ma history of the central TVZ (Wilson et al., 1995). Andesitic activity marking the onset of TVZ volcanism began at approximately 2 Ma. Cole (1990) divided the onshore and offshore segments of the TVZ into an eastern andesite-dacite continental arc, which is best developed in the northern and southern segments of the TVZ, and a marginal back arc basin characterised by bimodal basaltic-rhyolitic volcanism. However, this model is not universally accepted (e.g. Wilson et al., 1995).

Rhyolite is the dominant magma erupted in the whole TVZ (Wilson et al., 1995). Andesite is an order of magnitude less abundant, and dacite and andesite are volumetrically minor. Rhyolitic eruptions have generated ignimbrites, fall deposits and domes. The remaining compositions manifest in a variety of forms including tuff rings, scoria cones, fall deposits, lavas, domes, sills and pyroclastic flow deposits (Wilson et al., 1995).

Sedimentation in the offshore segment of the TVZ includes detritus from Upper Palaeozoic to Mesozoic rocks in the eastern part of the North Island and volcanic components from the TVZ (Lewis and Pantin, 1984). Terrigenous sedimentation is relatively continuous, whereas volcanic sedimentation is episodic and dominantly derived from onshore rhyolitic eruptions in the TVZ. Volcanic debris is delivered to the basin via rivers or directly from fallout (e.g. Lewis and Pantin, 1984) or by pyroclastic flows which reach the shoreline (e.g. Walker, 1979). The current fluvial input is minor as the onshore TVZ is being drained mainly by a river system which discharges on the west coast (e.g. Lewis and Pantin, 1984). Water-settled fallout and resedimented pyroclast-rich deposits, account for the greatest volume of sediment within the offshore TVZ (Lewis and Pantin, 1984). Volcanism within the offshore TVZ extends between the active andesitic White Island massif, northward along the Ngatoro Ridge, to the submarine Whakatane arc volcano (Gamble et al., 1993). These volcanic centres are an important syn- and post-eruptive source of basaltic, andesitic and minor rhyolitic and dacitic detritus to offshore TVZ (e.g. Lewis and Pantin, 1984). Whakatane volcano is inactive and comprises basaltic and andesitic lava flows and talus deposits (Gamble et al., 1993). The Ngatoro Ridge is studded with numerous vents, fissures and lava fields comprising sheet flows, pillows and talus fans (e.g. Wright, 1992; Gamble et al., 1993). In some areas, these deposits are blanketed by rhyolitic pumice clasts from an unknown source (Gamble et al., 1993). Submarine rhyolitic dome complexes and associated breccia facies flank White Island (Gamble et al., 1993). Andesitic detritus sourced from eruptions at White Island are largely restricted to <15 km from the volcano. However, volcaniclastic units derived from White Island occur in sediments up to 60 km from the source (Kohn and Glasby, 1978).
Shelf sediments in the offshore TVZ display varying provenance characteristics (Lewis and Pantin, 1984). Sediments in the western and central part of the shelf dominantly comprise volcanogenic sand sourced from the onshore TVZ and andesitic-dacitic eruptions from White Island. Some units contain epiclasts sourced from eroded intrabasinal rhyolitic and andesitic knolls (Lewis and Pantin, 1984; Kohn and Glasby, 1978). In the eastern part of the shelf, sediments are dominantly non-volcanic and delivered to the basin by rivers draining the adjacent greywacke ranges (Lewis and Pantin, 1984). On the shelf slope, late Pleistocene and Holocene sedimentation rates are around 0.1 to 0.2 m/ka (Kohn and Glasby, 1978). Water depths range from a few 10s of metres to > 3 km (e.g. Lewis and Pantin, 1984).

Figure 8.3 Structural setting of the Taupo Volcanic Zone (TVZ) of the North Island of New Zealand. Quaternary andesitic volcanoes are shown as solid triangles. The location of the central TVZ and position of major rhyolitic calderas is also shown. The position of the Vening Meinesz Fracture Zone (VMFZ) is from Wilson et al. (1995). Isobaths are in metres. Arrows show the direction of subduction of Pacific Plate beneath the Australian Plate. Modified from Davey et al. (1995).
The palaeogeographic setting of the Trooper Creek Formation and associated source terrain is envisaged to have been similar to that of the modern TVZ. Like the offshore TVZ, sedimentation in the Trooper Creek Formation included episodic influxes of pyroclasts from explosive eruptions at subaerial or shallow-submarine volcanic centres and from eruptions in relatively deep subaqueous environments. The style of eruptions represented by volcanlastic facies in the Trooper Creek Formation are interpreted to have been similar to some of the volcanism in the TVZ (e.g. Gamble et al., 1993; Wilson et al., 1995). Caldera-forming explosive eruptions in the source terrain may have been an important source of pyroclasts that were finally deposited in submarine environments. However, there is no positive evidence such as caldera-collapse megabreccias, very thickly ponded pumiceous mass-flow deposits or caldera-margin growth faults (e.g. Busby-Spera, 1986) to indicate that caldera volcanoes were present within the study area.

The interleaving and juxtaposition of volcanic facies observed in the Trooper Creek Formation is presently taking place in the offshore TVZ (e.g. Lewis and Pantin, 1984). Basin subsidence, syn-volcanic faulting and episodes of uplift are important elements of the TVZ and are likewise recorded in the Trooper Creek Formation. However, the TVZ differs from the Trooper Creek Formation in several ways. (1) Although the dimensions of the Trooper Creek Formation depocentre are poorly constrained the regional extent of the preserved succession (160 km) is significantly less than the dimensions of the TVZ; (2) In the Trooper Creek Formation, intrabasinal volcanoes contributed a greater volume of volcanic detritus to the depocentre than basin margin or subaerial volcanic centres. In contrast, the onshore extension of the TVZ is the focus of current magmatism, and is more extensive than the offshore segment; (3) Unlike the TVZ, the supply of terrigenous sediment to the Trooper Creek Formation depocentre was limited and overwhelmed by volcanic sedimentation; and (4) The Trooper Creek Formation contains several major VHMS deposits. Massive sulfide mineralisation has not been recorded in the offshore TVZ.

8.5 Implications for comparable volcanic successions

The research undertaken here has relevance to understanding comparable ancient submarine volcanic successions and assessing prospective host sequences for massive sulfide mineralisation.

(1) Studies of silicic submarine volcanoes are largely limited to ancient volcanic successions, such as the Seventy Mile Range Group, that are now exposed on land. The

The present study has contributed to the understanding of volcanic facies generated by felsic to intermediate eruptions in submarine environments. In particular, the importance of syn-sedimentary intrusions in submarine volcanic successions, and the lithofacies associations which characterise intrusion- and lava-dominated volcanic centres are documented (Chapter 5). The research constrains models for the growth of silicic lavas and intrusions, and suggests that they can play an important role in influencing the pore fluid properties of the volcanic succession and the location, geometry and chemistry of syn-volcanic submarine hydrothermal systems (Chapters 6-7).

(2) Although major advances have been made in understanding the chemistry and evolution of Australian VHMS deposits, few studies have evaluated the interrelationships between volcanism and mineralisation. Detailed definition of the lithofacies and palaeovolcanologic setting of the Seventy Mile Range Group in the area around the Highway-Reward deposit has allowed for recognition of those parts of the volcanic succession that are most prospective for VHMS deposits (Chapter 3). The results of the research can be applied to other parts of the Seventy Mile Range Group and comparable submarine volcanic successions elsewhere.

(3) The literature on VHMS deposits emphasises mineral deposition within caldera settings. This analysis suggests that silicic, syn-sedimentary intrusion-dominated submarine volcanic centres are also important settings for some massive sulfide deposits. The present study also highlights the importance of sub-seafloor replacement during massive sulfide accumulation. Based on a study of the Highway-Reward deposit and the few other published descriptions of sub-seafloor deposits, the various circumstances by which sub-seafloor deposits can develop have been summarised and the implications for mineral exploration assessed (Chapter 7).

(4) In the study area, type 1 ironstones have geochemical signatures which suggest they may be associated with as yet undiscovered massive sulfide mineralisation (cf. Duhig et al., 1992; Davidson, 1996), whereas type 2 ironstones are interpreted as deposits from hydrothermal fluids circulating around lavas, intrusions and explosive volcanic centres (Chapter 6). Ironstones occur within the host successions to many Australian VHMS deposits. In these successions, type 2 ironstones may also be targets for exploration.

8.6 Avenues for further research

A comprehensive analysis of the Seventy Mile Range Group concentrating on recognition of distinctive facies and facies associations with the aim of reconstructing the Cambro-Ordovician facies architecture has yet to be completed. The research presented here
suggests that this approach will provide important insights into lithofacies, composition, depositional environment and volcanic history of the Seventy Mile Range Group. The proposed study will also allow for recognition of other parts of the Seventy Mile Range Group that are prospective for as yet undiscovered VHMS deposits. There are several important topics which should be addressed. These include: (1) a determination of the regional extent of the Kitchenrock Hill Member in the Trooper Creek Formation (Chapter 3); (2) an evaluation of the regional significance of a shallowing upward trend in the Highway Member (Trooper Creek Formation) at Trooper Creek prospect (Chapter 4); (3) further assessment of the potential for some key facies associations within the Trooper Creek Formation to be traceable over several tens of kilometres (Chapter 3); and (4) clarification of the lithofacies comprising the Rollston Range Formation (Chapter 3).

The Mount Windsor Formation comprises a thick sequence of rhyolitic and dacitic lavas, domes, intrusions, subordinate volcaniclastic rocks and rare sedimentary units (e.g. Berry et al., 1992; Chapter 3). The formation extends for over 60 km within the Mount Windsor Subprovince. Extensive silicic submarine lava- and intrusion-dominated volcanic successions have been little studied (e.g. Gibson, cited in Cas, 1992), in part due to their relative scarcity in modern environments. The quality of outcrop available in large parts of the Mount Windsor Formation will allow precise volcanic facies definition, an interpretation of the style of volcanic activity and evaluation of the depositional setting. The research could be undertaken during systematic geological mapping of the Seventy Mile Range Group.

There are few detailed descriptions of sub-seafloor replacement style VHMS deposits. In particular, the geochemistry of hanging wall alteration associated with sub-seafloor replacement deposits remains poorly understood. Detailed definition of the lithofacies and volcanic history of the host succession to the Highway-Reward deposit provides a framework for geochemical studies of alteration associated with a sub-seafloor replacement style Cu-Au-rich VHMS deposit. Detailed mineralogical, paragenetic, fluid inclusion and isotope studies of the massive sulfide orebodies could form part of this research. The results of the proposed research will have applications for mineral exploration in the Seventy Mile Range Group and other ancient volcanic successions.

8.7 Summary

A detailed analysis of the Mount Windsor Formation and Trooper Creek Formation in the area between Coronation homestead and Trooper Creek prospect has led to a better understanding of the stratigraphy and palaeovolcanology of the central part of the Mount Windsor Subprovince. The preserved lithofacies were deposited in a submarine
environment. The first stage of Trooper Creek Formation volcanism and sedimentation (Kitchenrock Hill Member) reflects local post-eruptive erosion of the Mount Windsor Formation rhyolitic and dacitic units, and contemporaneous subaqueous (syn-eruptive) redeposition of rhyodacitic to dacitic, pyroclastic debris generated at an adjacent subaerial or shallow marine, explosive volcanic terrain. Minor siltstone units, turbidites, lavas and intrusions were also emplaced in the basin (below storm wave base) during this stage. The overlying Highway Member comprises compositionally and texturally diverse volcanioclastic facies which are intercalated with volcanic and non-volcanic sedimentary facies. The volcanioclastic facies reflect resedimentation of pyroclastic and autoclastic detritus, sourced intrabasinally (andesitic–rhyolitic volcanic centres) and from an adjacent subaerial or shallow marine, rhyolitic to dacitic, explosive volcanic terrain. The top of the Highway Member marks the end of volcanic-dominated sedimentation. This was then followed by a phase of post-emptive erosion and reworking of volcanic deposits in the source terrain with rare eruptions of lava and incursions of rhyodacitic sediment gravity flows, leading to deposition of the Rollston Range Formation. The Rollston Range Formation includes significant basement-derived detritus and was deposited in a marine environment (below storm wave base).

The palaeogeographic setting, lithofacies and formative volcanic processes represented by the Trooper Creek Formation are similar to those documented (e.g. Gamble et al., 1993; Wilson et al., 1995) in TVZ, New Zealand.
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Appendix A

Clast shape and textural associations in peperite as a guide to hydromagmatic interactions: Late Permian basaltic and basaltic andesite examples from Kiama, Australia
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Clast shape and textural associations in peperite as a guide to hydromagmatic interactions: Late Permian basaltic and basaltic andesite examples from Kiama, Australia

Introduction

Interaction between magma or lava and wet unconsolidated sediment is common in environments where sedimentation accompanies volcanism, especially in subaqueous settings where large volumes of magma are emplaced sub-seafloor as syn-sedimentary intrusions. A variety of processes and products attributable to magma-wet sediment interaction have been recorded, including intrusive pillows (Snyder and Fraser 1963a,b; Kano 1991), effusive magma-sediment slurries (Lawson 1972, Leat and Thompson 1988, Sanders and Johnston 1989), and peperite (Fisher 1960, Schmincke 1967, Williams and McBirney 1979, Brooks et al. 1982, Kokelaar 1982, Busby-Spera and White 1987, Brooks 1995). Peperite is a genetic term for a rock formed by the mixing of magma or lava with wet sediment. Peperite occurs at contacts between intrusions and the host sediment (Hanson and Schweickert 1982, Branney and Suthren 1988), along basal contacts of lavas (Schmincke 1967) or surrounds burrowing parts of lavas. Here I describe peperite and related structures in basaltic and basaltic andesite lavas and syn-sedimentary intrusions from the Late Permian Broughton Formation, Kiama, New South Wales. Because of continuous coastal exposure at this locality it has been possible to interpret from field observations the significance of textures and structures in peperite.

Peperite is useful for demonstrating contemporaneous volcanism and sedimentation, and because it preserves evidence of progressive stages in hydrovolcanic interactions (non-explosive mixing, steam explosions). Busby-Spera and White (1987) identified two textural types of peperite: in blocky peperite, clasts derived from the magma have polyhedral blocky shapes and commonly fit together like a jigsaw puzzle, whereas in globular peperite, juvenile clasts are bulbous. In this study variations in clast shapes and interrelationships are interpreted in terms of changing hydrovolcanic interactions during magma-sediment mixing. In particular, the role of host-sediment properties in determining peperite type is assessed and associations between peperitic, autoclastic and coherent facies are examined.

Terminology and description of peperite

Peperite can be identified, described and interpreted on the basis of (1) igneous clast shape; (2) fabric; and (3) location with respect to the margin of an igneous body. Clast
shapes described in this study are present in many other examples of peperite (e.g. Busby-Spera and White 1987, Branney and Suthren 1988, Hanson 1991, Hanson and Wilson 1993, McPhie 1993, Rawlings 1993, Brooks 1995). Important insights into hydromagmatism, and intrusive and mixing processes might be gained from the investigation of the complex relationships between different clast types and textural associations, so it is important that complexities are recorded. Peperite consisting of one clast type is termed blocky, globular, ragged or platy peperite following on from Busby-Spera and White (1987). Peperite containing a high proportion of clasts from more than one textural group is here classified as mixed peperite and the clast shapes indicated (e.g. mixed ragged-globular peperite). In peperite with a closely packed fabric (Hanson and Wilson 1993), sediment fills joints and fractures that define pseudo-pillows (Watanabe and Katsui 1976; Yamagishi 1987, 1991), and columns and polyhedral joint blocks (Brooks et al., 1982) in the coherent facies. Peperite with dispersed fabric (Hanson and Wilson 1993) is a sediment matrix-rich breccia with clasts and tongues of the igneous component. Peperite occurs at the margins of lavas and intrusions and is present as pods, sheets and dykes in massive coherent facies within the interior of the units.

**Geological Setting**

Peperite examined in coastal exposures at Kiama, New South Wales occurs in the upper part of the Late Permian Broughton Formation. The Broughton Formation and overlying coal-bearing Pheasants Nest Formation form part of a conformable regressive sedimentary succession within the Permo-Triassic Sydney Basin (Cas and Bull 1993).

The Broughton Formation and the lower part of the Pheasants Nest Formation include both sedimentary and volcanic facies associations (Raam 1969). The sedimentary facies association is dominated by thin to thickly bedded immature sandstone, pebble conglomerate and mudstone of volcanic provenance, and occurs as four intervening units between volcanic facies of the Broughton Formation. Units are interpreted as high-density turbidity current and tractional current deposits emplaced in a storm- and tide-dominated, shallow marine environment (Bull and Cas 1989). Dropstones within the lower part of the Broughton Formation suggest that periodic coastal sea ice and/or icebergs were present during deposition. Dips of bedding rarely exceed 2°. The volcanic facies association comprises nine shoshonitic basaltic to basaltic andesite lavas and syn-sedimentary intrusions, previously termed latites, and associated autochthonous breccia and peperite (Carr 1985). Three of the lowermost members of Broughton Formation are relevant to this study. They are, from oldest to youngest, the Blow Hole Latite Member, the Kiama Sandstone Member and the Bumbo Latite Member (Fig. 1). The Blow Hole Latite Member is holocrystalline and porphyritic, containing euhedral to subhedral
plagioclase and pyroxene phenocrysts, and chloritic pseudomorphs of olivine phenocrysts, in a fine-grained pilotaxitic groundmass. The groundmass consists of plagioclase microlites, pyroxene microlites, chlorite, an unidentified opaque phase (magnetite?), and interstitial potassium feldspar. The petrography of the Bumbo Latite Member is similar, although olivine phenocrysts are absent and the groundmass is finer grained. Volcanic and sedimentary facies associations are well exposed in coastal cliffs at Kiama. However, outcrop inland is restricted to quarries and road cuts.

**Contact Relationships**

The Blow Hole Latite Member is a 50 m thick basaltic andesite sheet which was initially interpreted as a tripartite intrusion (Raam 1964). However, Bull and Cas (1989) considered that only the middle unit of the sheet was partly intrusive, and regarded it as a lava which locally burrowed into wet sediment. This study demonstrates that the Blow Hole Latite Member can be divided into two flow units with peperitic contacts suggesting their intrusion into wet unconsolidated sediments. A thin, poorly exposed horizon of bedded sandstone (Rifle Range Tuff Member, Raam 1964) exposed at Rifle Range Point (Fig. 1) separates the upper and lower flow units. The middle flow unit proposed by previous authors is a peperitic facies of the lower flow unit. The upper and lower units are interpreted as syn-sedimentary intrusions, due to the volume and extent of peperite development. However, critical facies relationships required to discount a burrowing flow are absent due to poor exposure inland.

The Bumbo Latite is a 150 m thick massive, columnar jointed basalt sheet above the Kiama Sandstone Member (Fig. 1). The base of the member is locally peperitic and the upper contact was not examined in this study. The Bumbo Latite also has been interpreted as a tri-composite extrusion (Bowman 1974).

At map scale the sheets are broadly concordant with bedding in the enclosing sedimentary rocks. However, at outcrop scale contacts vary from relatively planar to complex and highly irregular. Unmixed lower contacts vary from smooth to undulating with 10-20 cm amplitude lead casts of coherent basaltic andesite separated by flames of sandstone. Underlying sedimentary rocks are relatively undisturbed except for minor soft-sediment deformation attributable to the loading effect of the sheets.
Figure 1. Geology of the Permian Broughton Formation at Kiama, showing complex relationships between peperite, hyaloclastite and coherent facies in the Blow Hole and Bumbo Latite Members.
Facies of the Blow Hole and Bumbo Latite Members

Coherent Facies

Regular, well-developed, wide (to 1 m) columnar joints characterise the massive interiors of the Bumbo and Blow Hole Latite Members. In places (e.g. Kaleula Point) column faces are dissected by interconnected, broadly curved tortoise shell joints which, in three dimensions, define equant polyhedral blocks. More often columns are cut by less regular, curved and planar joints. Column axes are generally subvertical and perpendicular to sheet margins. However, along contacts with some dyke-like peperitic domains in the Blow Hole Latite Member, column axes are subhorizontal at contacts, but progressively steepen and become subvertical a few metres into massive basaltic andesite (Fig. 2, 3A). Along the top of peperite dykes, columns are subvertical, but are cut at right angles by concentric joints spaced a few 10’s of centimetres apart (Fig. 2). Concentric joints mirror the upper margin of the peperite domains, forming a wavy pattern where peperite dykes are closely spaced.

Near contacts with sedimentary facies or peperitic zones, columnar joints merge into a several metre wide interval of blocky jointing. Widely spaced, smoothly curved, intersecting joints outline polyhedral blocks, 2-6 metres in length (pseudo-pillows, Watanabe and Katsui 1976; Yamagishi 1987, 1991), many of which are internally jointed. Joints are progressively more closely spaced within a metre or two of contacts.
(cf. Brooks et al., 1982) dissecting the rock into small blocks, 5 to 30 cm across. Some blocks are defined by intersecting radial and concentric joints which diverge outward from small (20-30 cm) discontinuous apophysis-like tongues of peperite (Fig. 3B). Blocky jointed basalt or basaltic andesite is in direct contact with peperite along part or all of some contacts and elsewhere grades into hyaloclastite.

Locally in the Blow Hole Latite Member, subvertical platy joints form an intervening zone between columnar jointed and blocky jointed coherent facies. Platy joints are laterally continuous, spaced up to 1.5 metres apart, dissected by crude blocky jointing, and conform to contacts with peperitic and blocky jointed domains. Subhorizontal joints up to 10' s of metres in length form bifurcating networks in both platy- and blocky-jointed domains.

**Hyaloclastite Facies**

Exposures of hyaloclastite are monomictic and characterised by jigsaw-fit of polyhedral blocky and cuneiform clasts separated by minor amounts of finely comminuted magmatic rock. In the Blow Hole Latite Member, in situ hyaloclastite may be the brecciated equivalent of large parts of the coherent facies or form a narrow selvedge between blocky jointed coherent facies and peperite. Often, clasts decrease in size approaching peperitic contacts and some fractures have been invaded by sediment, forming peperite.

At Blow Hole Point, small pods of hyaloclastite are enclosed by massive columnar and blocky jointed basaltic andesite. Almost continuous outcrop between Blow Hole Point, Black Beach and Pheasant Point (Fig. 1) provides a section through the outer interior to the margin of the upper Blow Hole Latite Member, and suggests that it is a sill. The hyaloclastite facies can be regarded as an intermediate facies between the massive columnar- and blocky-jointed coherent facies and marginal peperite. Features which characterise this transition are, from the margin inward, a rapid decrease in peperite to hyaloclastite, reduction in the degree of brecciation, and replacement of blocky jointing by columnar jointing as the major joint style.

**Closely-packed peperite**

Peperite with closely-packed fabric occurs only within the interior of the Blow Hole Latite Member. Blocky jointed coherent facies merge into domains of closely-packed peperite where sediment is present between widely spaced, smoothly curved, intersecting joints which define polyhedrally jointed blocks (Fig. 2). More continuous sediment-filled subhorizontal joints, up to 30 m in length, outline pseudo-pillows (Fig. 3C). Pseudo-pillows are dissected by internal joints, which are free of sediment, or else separated by a thin or thick infill of sediment (cf. Yamagishi et al. 1989). Basaltic andesite in the interior and margins of pseudo-pillows is texturally equivalent to that of the massive facies.
Figure 3.

Outcrop features of the Blow Hole Latite Member (A-D, F) and Bumbo Latite Member (E).

(A) Transition from blocky jointing (b) to columnar jointing (c) passing out from the margin of a dyke-like body of dispersed peperite within the interior of the intrusion (p). Columns are sub-horizontal at the contact with the dyke but progressively steepen and become subvertical. Pack for scale. Marsden Head.

(B) Lobate incursions (arrow) of peperite (p) into blocky jointed coherent facies (b). Within the coherent facies, trails of ellipsoidal vesicles conform to the shape of some parts of the contact. Scale 10 cm long. Kendalls Point.

(C) Closely-packed peperite showing progressive dismembering of coherent basalt into pseudo-pillows (p). Sediment fills fractures between subhorizontal fractures (arrow) and fractures in pseudo-pillows. Kaleula Head.

(D) Cross section through lobes (l) dissected by incipient columnar and blocky jointing and partially enclosed in altered dispersed peperite (p). Clasts in the breccia and adjacent to lobe margins display jigsaw-fit texture demonstrating that the lobe and breccia are cogenetic. Marsden Head.

(E) Detailed drawing from photograph. Type D lobes (l) enclosed in cogenetic peperite have altered margins (a) and unaltered jointed (j) and cores (u). Parts of some lobe margins are strongly vesicular (v). Peperite with vesicular clasts (vp) contrasts with peperite-dominated by poorly vesicular polyhedral blocky clasts (bp). Scale 10 cm long. Bombo Point.

(F) Lamination (arrow) and concentration of lithic clasts (l) on the ?lee side of a juvenile clast (j) derived from the walls of the enclosing sheet fracture in closely-packed peperite. Juvenile clast is 2.5 cm long. Kaleula Head.
However, along some contacts with sediment less than a millimetre of the groundmass is black in colour and charged with a fine unidentified opaque phase.

Subhorizontal fractures in closely-packed peperite are filled with up to 10 cm of siltstone to sandstone. However, thicknesses of sediment vary considerably along their length. Towards fracture terminations, infills decrease to a sub-millimetre film which is present along the whole length of the fracture, or else fractures are sediment free. In some cases, segments or the terminations of subhorizontal fractures comprise stacked sets of interconnected, sediment-filled, en-echelon fractures. Similar, but subvertical en-echelon fractures characterise some outcrops of the polyhedrally jointed coherent facies. En-echelon fractures are interpreted as tensile fractures formed by non-rotational, dilational strain during the invasion of overpressurised sediment (cf. Beach 1975, Francis 1982). The surfaces of subhorizontal fractures are sharp, but have an irregular form which reflects small-scale steps in the direction of fracture propagation and incomplete exfoliation of incipient clasts from some walls. Platy clasts (cf. Brooks 1995) liberated from fracture surfaces form jigsaw-fit aggregates separated by minor amounts of sediment matrix. Apophyses of sediment extend a few centimetres in from some sheet fracture walls and locally have formed peperite comprising globular-shaped clasts.

Close to domains of dispersed peperite, outlines of pseudo-pillows are masked as the proportion of sediment-filled fractures increases. Remnants of large pseudo-pillows enclose multiple smaller pseudo-pillows which, with increasing brecciation, disintegrate into aggregates of blocky to ellipsoidal clasts separated by sediment matrix. Wedge-shaped, sediment filled fractures penetrate the pseudo-pillows. The largest fractures are over 1 m in length and, where closely spaced, generate complex serrated margins to pseudo-pillows. Thinner wedges extending in from the surfaces of larger fractures locally merge, outlining platy clasts surrounded by sediment.

At Marsden Head, well developed, subvertical columnar joints, cut at right angles by subhorizontal joints, extend upward from a subhorizontal sheet-like body of dispersed peperite in the interior of the sheet. An irregular, roughly ellipsoidal section of columnar jointing, 10 m wide and 5 m high, that occurs 1m above the peperite is dissected by blocky joints and sediment-filled fractures. Ghosts of former columnar joints are visible towards the centre of the zone, but are best observed along gradational contacts with intact columnar jointed basaltic andesite. Domains of blocky jointed basaltic andesite are dissected by fine sediment-filled fractures that are connected to the underlying peperite by a network of sediment veins (cf. Brooks et al., 1982). Some veins follow the margins of column faces, but most form bifurcating networks within the blocky jointed interiors of remnant columns. Farther to the south, sediment fills the space between some column
faces. Relationships at these two localities suggest that columnar jointing was initiated synchronous with peperite formation.

**Dispersed Peperite**

Peperite with dispersed fabric passes into massive blocky jointed coherent facies, or grades through an intervening zone of closely-packed peperite as the proportion of sedimentary matrix between clasts decreases. Contacts with the enclosing facies are highly irregular.

Dispersed peperite occurs from the base to top of the Blow Hole Latite and does not appear to be restricted to a specific level. In map view, this facies forms elliptical pods and interconnected peperite tongues, a few metres wide and up to 10 m long, isolated in blocky jointed coherent facies. Tongues separate lobe-like, blocky jointed, coherent domains which extend in from the surrounding coherent facies. In cross-section, dyke-like bodies, irregular branching networks, and sheets of peperite are surrounded by coherent facies or extend up from the base of the sheets to more than 10 m into coherent facies. Pods and tongues of peperite apparently isolated within coherent facies are interpreted as cross-sections through dykes (cf. Brooks et al., 1982). However, others are evidently rootless and direct connections to the enclosing sedimentary package are not apparent. Elliptical domains of coherent basalt or basaltic andesite partially or completely enclosed in peperite resemble cross-sections through lava-lobes (Figs. 3D, 4, 5A-B).

Most peperitic domains include poorly- and strongly-vesicular parts, resulting in apparent polymictic breccias in which pods and fingers of contrasting vesicularity are juxtaposed. Clasts contain a uniform to heterogeneous distribution of vesicles ranging from 0.1 to 3.5 cm in diameter, and vary from non-vesicular to containing around 15% vesicles; some are nearly scoriaceous. At the margins of some poorly vesicular coherent facies, a coherent vesicular rind passes out into peperite comprising vesicular clasts (Fig. 4), demonstrating that the facies are cogenetic. Along some contacts within the Blow Hole Latite Member, lobate apophyses of peperite (10-20 cm across) comprising vesicular clasts are enclosed in weakly-vesicular coherent facies (Fig. 3B). Aligned ellipsoidal vesicles in the weakly-vesicular coherent basalt-andesite mirror the broad shape of some of these contacts. In many apophyses, sediment is concentrated at the top of the structure, possibly trapped there as expanded pore water cooled, preventing further advance into the still plastic basaltic andesite. Clasts associated with vesicular domains have fluidal and globular shapes although some clasts in poorly vesicular domains also have these shapes. In some outcrops (e.g. Kendalls Point, Marsden Head), in situ hyaloclastite at the margins of the coherent facies passes into dispersed peperite containing jigsaw-fit aggregates of polyhedral blocky clasts. Within the peperite, groups of poorly vesicular clasts with jigsaw-fit texture are enclosed by areas where clast rotation and separation are
Appendix A

10.

evident. In some exposures (e.g. Kendalls Point), wide (5-40 cm) subhorizontal sediment-filled fractures can be traced through the breccia. Fracture walls are irregular and stepped.

Occurrences of dispersed peperite at the margins of the Blow Hole and Bumbo Latite Members consistently have a dispersed fabric. This is best illustrated along the contact between the Bumbo Latite Member and the underlying Kiama Sandstone Member at Bombo Point. Vesicular domains occur as small pods in coherent poorly vesicular basalt and as peperite which encloses small lobe-like bodies of poorly vesicular basalt up to 0.8 m in length (Figs. 3E, 5D). Away from contacts, there is a transition from tube-vesicles to round and ellipsoidal vesicles in coherent vesicular basalt. Margins of large lobes and all of the smallest lobes are light green in colour and altered, whereas lobe interiors are black and unaltered. Lobe-like bodies show progressive disintegration into jigsaw-fit aggregates of blocky clasts. Jigsaw-fit texture is poorly preserved in peperite containing vesicular clasts. Contacts between poorly- and strongly-vesicular domains are mostly sharp. However, mixing of vesicular and non-vesicular clast types has locally generated texturally complex peperite. Sandstone containing juvenile vesicular clasts fills some fractures in the poorly vesicular lobe-like bodies, so that the lobes appear to intrude earlier, texturally distinct peperite.

The upper contact of the upper Blow Hole Latite Member is extensively exposed on the shore platform at Pheasants Point. Pods, tongues and sheets of massive to blocky jointed basaltic andesite up to 5 m in length are enclosed in cogenetic peperite (Fig. 5C). Parts of some tongues are cut by wide to narrow sediment-filled fractures which dissect them into smaller bodies and irregular blocks with jigsaw-fit geometry. Small digitate apophyses of basaltic andesite up to 5 cm in length extend out from lobe margins. In detail, much of the peperite consists of interconnected, bulbous, entral-like domains of basaltic andesite which are separated by sediment, but which can be traced back to coherent facies of the lobes. Peperite at the margins of some lobes encloses pods comprising clasts which are more vesicular and/or have different shapes, and are separated by greater amounts of sediment. Bedding in sandstone above the contact zone is undisturbed, in contrast to the near complete destruction of bedding in the peperitic facies.

Lobes

Lobe-like bodies of coherent basalt and basaltic andesite are isolated in the peperite or connected to coherent facies by wide stems of the same composition. On the basis of size, shape and relationships with associated peperite, lobes are divided into four types; A to D (Fig. 5). Peperite in the interior of the sheets incorporates types A-D, whereas peperite at
contacts with then enclosing sediments contains only types C and D. In peperitic facies of the Bumbo Latite Member, only type D lobes have been recognised.

Type A lobes — are elliptical- to pendant-shaped when viewed in cross-section (Figs. 3D, 5A), and tongue-shaped to elliptical in map view. They are up to 25 m in length and 20 m wide. Lobe interiors are unaltered and dissected by intersecting polyhedral joints, or polyhedral-jointed basaltic andesite encloses an inner zone of incipient radial columnar jointing. Pale green, in situ hyaloclastite (± peperite) forms a selvedge along segments of some lobe margins. Parts of some margins are vesicular and grade out into peperite comprising vesicular clasts. Rarely, vesicular pods to 15 cm wide occur in the lobes. Lobe interiors are penetrated by sediment-filled fractures. Fractures are planar along contacts with poorly vesicular domains, but have more irregular shapes when cutting numerous vesicles.

Type B lobes — Fractures at the margins of the type B lobes are penetrated by sediment, whereas lobe interiors are sediment-free (Fig. 5B). Sediment-filled fractures cut across some larger lobes producing trains of progressively smaller remnant coherent domains, which become more widely spaced as larger segments of the lobes are brecciated. Jigsaw-fit aggregates of clasts separated by sediment outline former large lobes which have undergone complete brecciation. Clasts become smaller and separated by greater amounts of sediment forming a matrix between the lobes. Slight modification of jigsaw-fit textures by rotation and separation of clasts, to complete loss of jigsaw-fit texture is widespread in the matrix.
Figure 5. Field sketches of lobes formed by incomplete brecciation in peperite facies of the Blow Hole Latite (A-C) and dispersed peperite facies of the Bumbo Latite (D). A— Cross section of a type A lobe; Kaleula Head. B— Plan view of a type B lobe in peperite displaying in situ and clast-rotated textures; Marsden Head. C— Type C lobe gradational into peperite containing clasts varying from poorly to strongly vesicular and from blocky to globular in shape; Pheasant Point. D— Type D lobes enveloped by an altered margin and enclosed in peperite containing domains of poorly and strongly vesicular clasts. Coherent facies show an equivalent range in vesicularity to clasts in peperite. Bumbo Point.

Type C lobes — Type C lobes characterise the peperitic upper margin of the upper Blow Hole Latite Member. Sheets of relatively coherent jointed basaltic andesite enclose pods and large domains of peperite (e.g. Marsden Head). Outlines of lobes become distinct as the proportion of peperite increases, enclosing relic pods of polyhedrally jointed basaltic andesite to 1 metre in size (Fig. 5C). Sediment-filled fractures dissect large lobes into groups of blocky clasts and small lobes which are separated by sediment matrix-rich domains. Clasts fit together along some margins but others have moved following fragmentation. Variation in clast shapes and vesicularity produces texturally complex peperite.

Type D lobes — Within some peperitic domains, poorly vesicular coherent basalt or basaltic andesite is interleaved with strongly vesicular intervals to 1m across (Fig. 5D). In
Appendix A

strongly vesicular domains, there is a gradation between coherent basalt or basaltic andesite, hyaloclastite and sediment matrix-rich and sediment matrix-poor peperite. All facies contain isolated pods and finger-like protrusions of poorly vesicular coherent or polyhedrally jointed basaltic andesite (Figs. 3E, 5D). Those pods and fingers in peperitic domains resemble concentric pillows (cf. Yamagishi 1987) and small pillow lobes. Some lobes are enveloped by a hyaloclastite (± peperite) sheath comprising poorly vesicular blocky clasts. Similar clasts are isolated in the surrounding peperite which is dominated by vesicular clasts.

Clast types and shapes

Peperite contains igneous clasts that can be divided into six main textural types on the basis of clast shape and relationships between clasts (Fig. 6).

Globular clasts — Globular clasts have bulbous, globular shapes ("entrail globular" clasts) or are roughly equant but are bound by finely digitate, fluidal margins ("equant globular" clasts). There is a progression in clast shapes between entrail- and equant-globular. In detail, most "clasts" are connected by fluidally-shaped stems a few millimetres to several centimetres wide; they are incipient clasts formed through fragmentation mechanisms which did not go to completion.

Entrail globular
Interconnected incipient clasts with rounded globular shapes form complex branching entrail-like interdigitations with sediment (Fig. 6A). Digits terminate in the surrounding sediment or connect small rounded patches of relatively coherent igneous component. The patches are up to several tens of centimetres across and many contain small, centimetre-sized blebs of sediment. Pinching off of branches along the bifurcating digits has delivered discrete clasts to the surrounding sediment. Only a thin film of homogenised sediment separates some clasts from their parent digit, whereas others are surrounded by large amounts of sediment.

Equant globular
In peperite comprising equant globular clasts there is less disruption of the igneous component as incipient clasts are larger and interpenetration with sediment is largely restricted to their margins (Fig. 6B). Incipient clasts are cut by bifurcating sinuous seams of sediment which propagate in from clast margins or outward from the interior. Other clast margins are planar and have sharp or finely serrated margins which imply that they are quench fractures.
Mesoblocky clasts — Mesoblocky clasts are an important but relatively minor component of some vesicular and poorly vesicular closely-packed and dispersed peperite facies. Along margins of mesoblocky domains, jagged sediment-filled fractures dissect the igneous component, defining progressively smaller fragments. Remnant finger-like projections of coherent and in situ fragmented igneous component extend out from margins of the coherent facies into clouds of mesoblocky fragments (Fig. 6C). Fragments are angular with finely serrate margins, and are mostly 1-5 mm across. Adjacent to fingers, many fragments display jigsaw-fit texture and are separated by only small amounts of sediment. Jigsaw-fit texture is absent in sediment matrix-rich breccia only a small distance into the breccia. Large clasts with shapes similar to mesoblocky clasts are an important component of incompletely fragmented domains.

Polyhedral blocky clasts — Polyhedral blocky clasts have angular, blocky and cuneiform shapes bounded by curviplanar margins (Fig. 6D). In some outcrops, broadly curved first-order fractures outline large blocky clasts which are dissected by second-order fractures into jigsaw-fit aggregates of progressively smaller polyhedral blocky clasts. Jigsaw-fit textures are disturbed in some parts of the breccia. Disturbance produces results which range from the slight modification of jigsaw-fit, by rotation and translation of fragments, to large scale separation of clasts.

Irregular blocky clasts — Strongly vesicular domains of dispersed peperite are characterised by a high proportion of clasts with irregular blocky shapes. Clasts are equant in shape, but bound by irregular to feathered margins which are in part the former walls of vesicles (Fig. 6E). Strongly vesicular clasts are bound mostly by vesicle walls and have feathered terminations. Highly irregular clast margins reflect rapid changes in the direction of fractures as they cut vesicles. Along contacts with coherent vesicular domains, clasts commonly display jigsaw-fit texture. Jigsaw-fit texture is lost as more sediment separates clasts.

Platy clasts — Platy clasts (Brooks 1995) are common in both closely-packed and dispersed peperite facies but are the principal clast type of closely-packed peperite. Platy clasts are several times longer than they are wide and show planar or irregular margins. They reflect the propagation of planar sediment-filled fractures (e.g. sheet, en-echelon) within relatively coherent facies.

Some clasts in peperite are bound by both globular to spongy margins and sharp planar-curviplanar margins, so that they do not fall into any one of the main textural groups (Fig. 6F).
Figure 6.

Clast types in peperite associated with the Blow Hole and Bumbo Latite Members.

(A) Discrete and interconnected incipient clasts with entrail globular shapes (light) enclosing and enclosed by sandstone (s).

(B) Incipient equant globular clasts with bulbous digitate margins invaded by thin fluidally-shaped sediment seams (arrow).

(C) Finger-like projection of basaltic andesite (f) showing progressive disintegration into mesoblocky fragments with finely serrate margins. Jigsaw-fit between fragments (arrow) is lost as sediment (s) penetrates fractures.

(D) In this example of polyhedral blocky peperite, clasts are separated by small amounts of sandstone matrix (s). Groups of clasts with jigsaw-fit contrast with domains where clasts have rotated and moved (arrow).

(E) Irregular blocky clasts bound by margins which are in part the former walls of vesicles (arrow) and enclosed in sandstone (s).

(F) In this domain of dispersed peperite, margins of clasts vary from planar-curvilinear to delicately fluidal (skeletal/spongy). These clasts imply a change in fragmentation mechanism during magma-sediment interaction.
Textural associations

The foregoing discussion highlights the wide variation in clast types in peperite. The distribution of clast types is not random. Textural zones are defined here as a domain of one clast type in hyaloclastite or peperite. Peperite may consist entirely of one textural zone or of multiple textural zones, arranged geometrically in recurrent textural associations. Variation in vesicularity is a principal determinant of clast types and textural associations. In closely-packed peperite, the magmatic component is consistently poorly vesicular, observed clast types are restricted to platy, globular and mesoblocky types, and textural associations are less diverse. Only short segments of a few fractures have mesoblocky and globular textures. In dispersed peperite, four principal associations have been recognised: (1) blocky jointed - equant globular; (2) blocky jointed - mesoblocky - entrail globular; (3) polyhedral blocky - irregular blocky; and (4) hyaloclastite - polyhedral blocky (Fig. 7).
Sediment matrix

Sediment forms the matrix to clasts, partially surrounds incipient clasts, and fills fractures and joints. The three principal sediment types, from most to least abundant, are: reddish-brown sandstone and minor siltstone, yellow-brown sandstone and granular to pebbly sandstone. Wisps and laminae of one grain size are enclosed by sediment of another grain size. Discontinuous planar- and rare cross-lamination are common to all peperitic facies, but best developed and most continuous in sediment-filled subhorizontal fractures in closely-packed peperite facies. Within the fractures, lamination is broadly concordant to walls but locally terminates against steps in the fractures. At one locality, laminae partially mantle a clast-supported lens of well-rounded granules which are concentrated on the lee side of a juvenile clast derived from the walls of the sheet fracture (Fig. 3F). Concentration of lithic clasts and fines depletion are interpreted to reflect local turbulence as fluids (water and steam) and sediment streamed through the fracture. Similarly, elutriation of fine sediment from some parts of the peperite is suggested by their sediment matrix-poor, clast-supported, but disrupted character. In some of these cases, wide subhorizontal fractures in blocky jointed coherent facies have sediment-poor, juvenile clast-supported breccia at their bases and sediment-rich upper parts which support large juvenile clasts. The distribution of sediment and juvenile clasts is similar to reverse coarse-tail grading.

Discussion

Emplacement and cooling

Contraction that accompanied cooling of the Bumbo and Blow Hole Latite sheets produced a variety of joint styles which are zonally arranged relative to peperitic and sedimentary facies, and record unequal rates of cooling. There is a transition from columnar jointed facies, through blocky jointed facies, into hyaloclastite along contacts with the enclosing sediments and/or peperite.

Columnar joints developed as intersecting contraction cracks nucleated within the blocky jointed zone and migrated towards the interior of the sheets, perpendicular to surfaces of equal tensile stress (Spry 1962, Long and Wood 1986). The pattern of columnar jointing suggests that, in most domains, surfaces of equal stress were parallel to isothermal surfaces at the contacts of the sheets, and columns formed perpendicular to both. Cooling of the igneous component along contacts with some dyke-like peperitic domains produced a distinctive style of columnar jointing. Initially, columns formed perpendicular to subvertical isothermal surfaces at the dyke margin but progressively steepened away from the dykes under a greater influence of isothermal surfaces parallel to sheet margins.
Sediment fills the space between some columns and other columns are dissected by blocky joints filled with sediment. These relationships suggest that columnar joints acted as pathways for the infiltration of wet sediment into the interior of the sheets. In blocky jointed zones, similar fractures may have provided access for fluids (± sediment) to move in and fragment the margins of the sheets (cf. Watanabe and Katsui 1976, Yamagishi 1987, 1991, Yamagishi and Goto 1992). The inward progression from blocky jointing to pseudo-pillow structure reflects a decrease in the degree of fragmentation and decrease in the cooling rate. In places, blocky jointed coherent facies developed along peperitic contacts, but more often, blocky jointing formed in a distinct zone inward from the hyaloclastite zone. In the hyaloclastite zone, quench fractures dissected joint blocks into jigsaw-fit aggregates of polyhedral blocky clasts (cf. Dimroth et al. 1978, Yamagishi 1979).

Vesiculation
Vesicle distributions in the Bumbo and Blow Hole Latite sheets are interpreted to reflect both primary magmatic vesiculation and vesiculation due to injection of steam from external water prior to complete solidification (cf. Fuller 1931, Waters 1960, Macdonald 1972, Walker 1987). Vesicles in poorly vesicular, coherent and peperitic facies probably reflect degassing of primary magmatic volatiles. Strongly vesicular zones are sparse, invariably associated with peperite and are localised and discontinuous. Isolated strongly vesicular pods in otherwise dense, massive, poorly vesicular basalt and basaltic andesite have not been observed (cf. Dimroth et al. 1978, Sahagian et al. 1989, McMillan et al. 1987, 1989). The association of peperite and domains of strong vesicularity suggest that the lava incorporated limited amounts of steam from the wet sediment in the initial stages of peperite formation (cf. Smedes 1956). Vesicular domains are interpreted as a form of vesicle cylinder. Wet sediment was heated and pore water vaporised as it moved into the magmatic component in dispersed peperite. A vesicular front may have propagated out into the magmatic component as sediment entered peperitic domains. Vesiculation was complete prior to brecciation, as sediment-filled fractures cut across vesicles and no clasts are zoned with respect to vesicularity. Vesiculation of fracture walls in closely-packed peperite did not occur, as the sediment was partially dewatered or the fluid was not vaporised, or the magmatic component had cooled sufficiently to resist vesiculation, or the lava had already degassed. Fraser (1976) attributes vesicle cylinders (2-20 cm across) in high-alumina basalts of the Cascade Mountains and Modoc Plateau to segregation of bubbles and residual melt into regularly spaced vertical cylinders. Although this mechanism cannot be discounted, the association of peperite and strong vesicularity in the Bumbo and Blow Hole Latite sheets favours the interpretation of vesiculation by steam.

Stress waves generated by high-pressure vaporisation of pore water at the melt-sediment interface can induce vesiculation of the melt (Wohletz 1983). Steam explosions are
interpreted to have played a minor role in generating peperite in the Blow Hole and Bumbo Latite Members, suggesting that stress wave induced vesiculation was insignificant.

Relatively few vesicles are filled with sediment, even in nearly scoriaceous peperite facies of the Bumbo and Blow Hole Latite Members (cf. Branney and Suthren 1988, Brooks et al. 1982). This may reflect a lack of interconnection between vesicles or that particles were too large to move through interconnections.

*Lobes*

Lobe types A-D lobes may simply be isolated coherent patches within otherwise strongly brecciated material. Alternatively, they could be interpreted as fractured and dismembered lava lobes, extruded into and partially or completely enclosed by their own or earlier peperite and hyaloclastite. Along some contacts, coherent facies pass through peperite containing jigsaw-fit clasts into lobes, demonstrating that type A-D lobes have formed through incomplete brecciation of coherent facies. Along contacts and in peperite where jigsaw-fit textures are not preserved, formation of lobes through extrusion/intrusion cannot be discounted. Type D lobes formed as vesicular pods in the sheet fragmented and mixed with sediment, leaving poorly vesicular domains. Complete loss of jigsaw-fit texture is widespread in the breccia surrounding type D lobes, so that they appear to invade earlier peperite. However, poorly vesicular coherent facies along the margins of peperitic facies enclose strongly vesicular pods which are coherent analogues of the matrix to type D lobes in peperite.

*Fluidisation of the host sediment*

The ability of sediment to penetrate even the finest fractures and large spaces in the interior of the basalt-andesite sheets to distances of tens of metres from the base, indicates that the sediment was highly mobile during peperite formation. Kokelaar (1982) ascribed similar features in peperitic facies of Ordovician andesitic and rhyolitic sills from Scotland and Wales to fluidisation of sediment by heating of pore water at sediment-magma contacts. In the present case, water at contacts was vaporised and some sediment injected up into the sheets, forming domains of peperite. Injection was driven by the relatively low density of the fluid-sediment mix compared with the magma and undisrupted sediment, and possibly by fluid over-pressure. The density inversion requires a disturbance to initiate flow of the low density layer, so that vapour expansion driven by the transition of water to steam may be more important, at least initially. The fluid-sediment slurries may have moved along fractures formed by contraction and/or quenching, or as propagating sediment dykes. Vesiculation of the magma by steam preceded the formation of peperite by mixing with the fluidised sediment. Some parts of
the surrounding magma remained sufficiently plastic to deform around mushroom-shaped tongues of sediment which penetrated up from contacts with peperitic domains.

Irregularities, fractures or peperitic domains at the margins of the sheets may have been preferred sites for the injection of fluid-sediment slurries (cf. Brooks 1995). Invasion of the sediment was probably vigorous but was not obviously explosive as jigsaw-fit textures between clasts and incipient clast are widely preserved, and contacts between vesicular and non-vesicular peperite are sharp with little mixing of clast types. Also, igneous clasts in the peperite commonly have bulbous, feathered or irregular outlines, rather than the angular blocky shapes typical of phreatomagmatic brecciation.

Remnant sedimentary lamination in sediment filling space between clasts in peperite has been described by many authors (e.g. Hanson and Wilson 1993, Kokelaar 1982, Branney and Suthren 1988, Hanson 1991, Brooks 1995). In the present case, wisps, seams or planar and cross laminae of one grain size are enclosed in, or alternate with, sediment of another grain size, producing extremely complex relationships in some cases. Lamination could be interpreted as: (i) relic primary bedding rotated and disrupted during intrusion; (ii) laminated sediment which infiltrated from above; or (iii) non-primary lamination. Structures are often subhorizontal, consistent with regional bedding, but are interpreted as non-primary sedimentary lamination because: (1) lamination is well developed within peperite facies completely enclosed by massive coherent lava; (2) lamination filling fractures in closely-packed peperite is parallel to fracture walls and could only be introduced along the length of the fractures (up to 30 m) through fluidisation; (3) structures in the sediment (e.g. cross lamination and lithic lenses in closely-packed peperite; reverse coarse-tail grading) are not consistent with washing-in processes. Layering reflects the repeated streaming of highly mobile sediment through fractures, and the intrusion of initial fracture- or space-filling sediment by coarser grain sizes. Vapour pressure was building, equilibrating and waning rapidly and unevenly in the invading sediment as it streamed to fill propagating fractures and open spaces. Rapid changes in sediment paths, superposition of sediments with different grains during the merging of fractures, and propagation of fractures at different rates all may have all been important in affecting vapour pressure and generating layering.

Relative timing

Figure 8 illustrates the relative timing of development of textures and structures in the Blow Hole and Bumbo Latite Members. Degassing of the sheets occurred both during emplacement, as evidenced by elongate vesicles, and after flow ceased, as indicated by spherical vesicles. Formation of vesicle cylinders clearly must have occurred while the sheets were still ductile, but probably after emplacement. Mixing of the lava and fluidised sediment formed domains of dispersed peperite. The general restriction of hyaloclastite
and blocky jointed facies to the margins of peperitic domains suggests that fractures developed concurrent with peperite in these domains. Columnar joints developed over a large part of the cooling history. Incipient columns dissected by blocky joints formed early concurrent with peperite. Long, well developed columnar joints in the massive interior of the sheets reflect slow cooling, largely following fragmentation and peperite formation. Sediment penetrating columnar joints at the base of the Blow Hole Latite Member, and filling brittle (en-echelon) fractures, suggest that sediment was moving through the sheet even in the late part of the cooling history.

*Mechanisms of brecciation*

The shape of clasts and contacts between sediment and the igneous component in peperite is a guide to fragmentation processes. Experimental and theoretical studies of magma-water interaction (e.g. Sheridan and Wohletz 1983, Wohletz 1986, Kokelaar 1986) have produced textures, structures and clasts with shapes which are similar to those observed in peperite, suggesting the mechanisms of magma-water interaction and magma-water-sediment interaction may be similar. Four primary clast forming processes are currently recognised to occur during magma-water interaction: magmatic explosivity, steam explosivity, cooling-contraction granulation, and dynamic stressing (e.g. Wohletz 1983, Kokelaar 1986). Steam explosivity is divisible into contact-surface interaction and bulk interaction (Kokelaar 1986).

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![Figure 8. Relative timing of development of textures and structures in the Blow Hole and Bumbo Latite Members. Exsolution of magmatic volatiles (vesiculation 1) was probably initiated in the vent and continued through vesiculation by heating of pore water during interaction between magma and wet sediment (vesiculation 2).](image-url)
Peperite comprising globular clasts indicates that non-explosive, contact-surface interaction and bulk interaction are probably important in the formation of peperite. Good evidence for contact-surface interaction is seen where tongues and apophyses of the igneous component transect undisturbed laminated or bedded host sediment, implying the passive removal of sediment during emplacement (cf. Branney and Suthren 1988). This was achieved by film boiling of pore water (Leidenfrost effect; Mills 1984), causing fluidisation of sediment at the magma-sediment interface. Sediment is displaced along and away from the contact zone until cooling below a critical temperature (Leidenfrost temperature) causes steam to condense and the sediment to be deposited. Oscillations in the vapour film can distort the magma surface into delicate bulbous fluidal shapes which detach, generating small fluidally-shaped fragments (Sheridan and Wohletz 1983, Wohletz 1986). Vapour films insulated the magma from direct contact with sediment and suppressing both steam explosions and quench fragmentation.

A case for bulk interaction in peperite formation is suggested where pods and seams of sediment are enclosed in the igneous component or occur between incipient clasts (cf. Kokelaar 1986, Branney and Suthren 1988, Brooks 1995). The main clast-forming process is the tearing-apart of the igneous component around invading and expanding steam-sediment slurries. Propagation of sediment seams promotes the disintegration of relatively coherent igneous material into progressively smaller clasts. Initially only a thin film of sediment, a few millimetres or centimetres wide, fills the seams. Walls of clasts are progressively wedged apart as sediment penetrates the seams. Vaporisation of pore water may have generated pressure waves causing disintegration of the magma. Kokelaar (1986) suggests that heat exchange between the magma and sediment through convective heat transfer may be more important than by direct contact mixing during bulk interaction. However, fluidally-shaped margins to incipient clasts with entrail and equant globular shapes suggest that direct contact mixing is in some cases important, and implies that bulk interaction and contact-surface interaction have combined to fragment the magma. Conductive heat transfer, a function of surface area and time of heat transfer, may increase as margins are "roughened" and the melt fragmented by contact-surface interaction, but will be limited by the insulating effects of a continuous vapour film. Concurrent bulk- and contact-surface-interaction combined to fragment the greatest percentage of the Blow Hole Latite Member.

In examples of peperite comprising ragged clasts, higher yield strengths at the strain rates which accompanied fragmentation are suggested by finely serrated, ragged clast margins. Again, bulk interaction during magma-sediment interaction may be indicated by textures in these domains. However, clasts with ragged shapes formed during bulk interaction (e.g. Branney and Suthren 1988) are similar to those produced by dynamic stressing.
Dynamic stress fragmentation is ascribed to brecciation of the chilled parts of lavas or intrusions by the continued movement of fluid magma in the interior.

In peperite comprising polyhedral blocky clasts, fractures define equant blocks, whereas platy clasts form by intersecting subparallel planar fractures and more widely spaced short cross fractures (cf. Brooks 1995). Clast shapes reflect different local stress fields, and may represent end members of a spectrum of clast shapes formed by quenching. Small scale changes in the direction of propagation of quench fractures in response to internal heterogeneities in the igneous component (e.g. phenocrysts) form jagged blocky/platy clasts bounded by serrated margins rather than sharp planar and curviplanar margins characteristic of polyhedral blocky clasts and some platy clasts (cf. Brooks 1995).

It remains unclear what the mechanism of formation of mesoblocky clasts was. Brittle failure may have resulted from propagation of stress waves through the melt in response to the collapse or explosive expansion of vapour films (cf. Wohletz 1983), or through cooling-contraction granulation. Turbulent mixing following quenching of the resulting fragments promoted the movement of fragments out of the zone of interaction and loss of jigsaw-fit texture.

Vesicles strongly influence the character of peperite formed when magma or lava invades wet, unconsolidated sediment. Fractures which cut across vesicles generate irregular blocky clasts with margins which are in part the former walls of vesicles. Vesiculation which occurs concurrent with fragmentation is likely to play a more active role in determining clast shape, but will be limited because bubbles will be entrapped as cooling proceeds and viscosity increases. An insulating sheath of vapour which forms at the contact between the magma and enclosing wet sediment may allow some bubbles to reach the magma-sediment interface (Mills 1984). Vapour bubbles which reach, form at, or penetrate the melt-film interface will probably interact with it, creating local pressure gradients which will influence vapour flow and hence also the shape of the contact surface and clasts.

Textural associations: evidence for controls on peperite formation

Textural associations of more than two clast types, and individual clasts with both bulbous and planar margins, imply a change in fragmentation mechanism. In many cases, initial magma fragmentation and mixing with sediment is thought to have resulted mainly from the tearing apart of the magma (bulk interaction) and shaping of the magma-sediment interface into fluidal globular shapes by contact-surface interaction. In other cases, globular surfaces and clasts developed first. Planar fractures reflect fragmentation by cooling-contraction granulation and/or by propagating stress waves. Planar fractures which cut across and displace fluidal globular surfaces in the igneous component formed later (cf. Goto and McPhie 1996). The relationship between some planar fractures and
globular surfaces is ambiguous and both may have formed simultaneously with viscosity
and/or temperature being the control.

Bulk physical properties, such as the density and viscosity of the magma and sediment
will in part control their behaviour during interaction. Difficulties in determining the
physical properties driving transitions in fragmentation mechanism result from the
complex and rapidly changing states of the components. For example, the magmatic
component will become more viscous with time, and steam together with volatiles can
promote multi-stage vesiculation of the melt. The sediment may be progressively
dewatered during interaction, with intergranular fluids ranging in temperature from cold
to boiling or superheated steam. Also, the host sediment is itself a many-phase system.

Busby-Spera and White (1987) concluded that host sediment properties strongly
influence magma-sediment interaction, and hence the shapes of clasts. They suggest that
fluidal globular peperite is more likely to develop in fine-grained, well sorted, loosely
packed sediment, as it is more easily fluidised and vapour films can be maintained at the
melt-sediment interface. Coarser, poorly sorted sediment is associated with blocky-
shaped clasts (blocky peperite) at Punta China, Baja, California. In these, greater
permeability was interpreted to inhibit the development of vapour films, and only a small
percentage of the sediment grain size is amenable to fluidisation. In the absence of
insulating vapour films, quench fragmentation and steam explosions are the main
fragmentation processes. At Kiama, different clast types occur within sediment of
constant grain size (Fig. 6F). Similarly, clasts with the same shape occur in sediment
with different grain sizes. These examples suggest that factors other than sediment grain
size are also important in determining fragment shape (cf. Goto and McPhie 1996).
However, sediment surrounding clasts in peperite represents the final grain size
distribution at the time of fragmentation and not necessarily that which was present at the
time of fragmentation.

Fragmentation processes are complexly dependent on external confining pressure. In
cases where the lithostatic and hydrostatic pressure exceed the critical pressure (about
31.2 Mpa for seawater; Kokelaar 1982), the degree of expansion of heated pore water is
impeded, steam explosions are suppressed and fluidisation may be inhibited. At lower
confining pressures steam may expand explosively. The character of peperite examined in
this study suggests that confining pressures were insufficient to suppress fluidisation of
the host sediment along magma-sediment contacts or to prevent vesiculation of the
magma, but large enough to inhibit steam explosivity.

Experimental and theoretical studies (Sheridan and Wohletz 1981, 1983; Wohletz 1983,
1986) suggest that changes in the water/magma ratio may lead to changes in eruption
style. In peperite, it is possible that both short and long term variations in water (and sediment)-melt ratios may be responsible for the changing fragmentation mechanisms, and so clast shapes. Direct application of results from experimental and theoretical studies of magma-water interaction to magma-slurry systems involving peperite is probably not possible. Also, changes in the water/melt ratio may occur due to varying volume rate of magma or sediment supply and fluxing of sediment with varying pore water contents during fragmentation.

Viscosity reduces growth rates of instabilities at the magma-sediment interface (Wohletz 1986), so that high viscosity magmas may mix more slowly with sediment than would low viscosity magmas. One might expect clasts with fluidally-shaped margins to be more common in peperite involving magma of mafic rather than silicic composition. The spectrum of clast shapes recognised in peperite span magma compositions ranging from basaltic to rhyolitic, suggesting that this may not be the case. However, changes in the rheological behaviour of a given magma from ductile to brittle, most likely in response decreasing viscosity, are clearly important in cases where peperite contains single clasts bound by both globular and planar surfaces. Planar fractures displace fluidal globular surfaces suggesting that they formed later. During the globular clast-forming stage, the magma had a relatively low viscosity and sediment was displaced by fluidisation. Planar and curvilinear fractures formed as the magma became more viscous, most likely in response to decreasing temperature and/or the breakdown of insulating vapour films at the magma-sediment interface (cf. Goto and McPhie 1996)

Viscosity profiles in some lavas and intrusions are likely to be complex, varying in response to, for example, pulsatory flow or intrusion (cf. Goto and McPhie 1996), and differing volatile contents, crystallinity and temperature. If magma rheology fluctuates then different parts of an intrusion or lava may be associated with peperite with different clast types and/or textural associations. Fluidal contacts and clasts will be generated early or in domains where the magma temperature is highest and viscosity is at a minimum. Continued flow will stress those parts that have already begun to cool and solidify, promoting brittle disintegration along contraction fractures, and clasts with blocky or ragged shapes are more likely to form. Also, if wet sediment injects the magma in pulses, then magma rheology at the time or site of interaction might fluctuate and different clasts form.

Conclusions

Peperites associated with basaltic to basaltic andesite lavas and intrusions in the Late Permian Broughton Formation, Kiama, New South Wales have been described on the
basis of (1) igneous clast shape; (2) fabric; and (3) location with respect to the margins of
the lava or intrusion. The complexities of peperite, in terms of clast types and their
relative abundances and distribution, as well as textures and structures in the host
sediment, indicate that a spectrum of fragmentation and mixing processes may occur
together and thus interact.

Examples of peperite with more than one clast type, involving magma of the same
composition and sediment of constant grain size, are common. In many examples,
globular surfaces formed during an early, low viscosity phase of magma emplacement
into wet sediment. Planar and curvilinear fractures truncate some fluidal surfaces
suggesting that these, at least in part, formed slightly later as the magma became more
viscous (cooler) and/or vapour films at the magma-sediment interface broke down (cf.

The intimate mixing of magma and wet sediment recorded by peperite is commonly a
precursory step towards explosive hydromagmatism. At Kiama, peperite has developed
by one or a combination of (1) non-explosive oscillation of vapour films at the magma-
sediment interface (contact-surface interaction); (2) non-explosive expansion of pore
water following enclosure of sediment in the magma or entrapment of sediment at the
magma-sediment contacts (bulk interaction), (3) cooling-contraction granulation; and (4)
brecciation of the chilled parts of an intrusion-extrusion by flow of the hotter interior
(dynamic stressing).

Fluidisation of the host sediment during mixing with the melt is common to peperite
involving clasts from all of the textural groups. Lamination in sediment within peperite
can include remnants of original stratification (e.g. Kokelaar 1982) and layering formed
by the streaming of fluid-sediment slurries through fractures and between clasts.
References


Appendix B

Geological cross-sections for the Highway-Reward deposit
Andesite intrusion
Rhyolite
Rhyodacite
Dacite
Non-stratified monomictic breccia (hyaloclastite)
Non-stratified sediment-matrix breccia facies (pse-vMA)
Stratified monomictic pyroclastic breccia (sandstone)
Stratified polymictic breccia - sandstone
Dacitic pumice breccia - sandstone
Feldspar-quartz pumice-crystal breccia - sandstone
Quartz & feldspar pumice breccia - sandstone
Quartz & feldspar crystal-phyric sandstone
Siltstone
Gossanous
Massive pyrite-chalcopyrite
Stringer veins
Fault
Shear
Appendix B

Highway-Reward

X-Sect 10275N ± 6.25 m

Mappe<1:M.Doyle

Andesite intrusion
Rhyolite
Rhyodacite
Dacite
Non-stratified monomictic breccia (hyaloclastite)
Non-stratified sediment-matrix breccia (tuffite)
Stratified monomictic rhyolite breccia - sandstone (resedimented hyaloclastite)
Stratified polymictic breccia - sandstone
Dacitic pumice breccia - sandstone
Feldspar-quartz pumice-crystal breccia - sandstone
Quartz & feldspar pumice breccia - sandstone
Quartz & feldspar crystal-lime sandstones
Silica
Gossanous
Massive pyrite-chalcopyrite
Sulfur veins
Fault
Shear

Highway

Not condol

R 4

R 8

RD 1

Andesite intrusion
Rhyolite
Rhyodacite
Dacite
Non-stratified monomictic breccia (hyaloclastite)
Non-stratified sediment-matrix breccia (tuffite)
Stratified monomictic rhyolite breccia - sandstone (resedimented hyaloclastite)
Stratified polymictic breccia - sandstone
Dacitic pumice breccia - sandstone
Feldspar-quartz pumice-crystal breccia - sandstone
Quartz & feldspar pumice breccia - sandstone
Quartz & feldspar crystal-lime sandstones
Silica
Gossanous
Massive pyrite-chalcopyrite
Sulfur veins
Fault
Shear
### Appendix C

**Summary graphic lithological logs**

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### Lithology

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<td>Rhyodacite</td>
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<td>Perlite</td>
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<td>Non-stratified monomictic breccia (hyaloclastite)</td>
<td>Massive/banded pyrite-sphalerite-taseite</td>
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<td>Stratified polymictic breccia-sandstone</td>
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<td>Crystal-vitric sandstone</td>
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### Alteration

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**Legend:**
- **F**: Feldspar-bearing
- **Q**: quartz
- **P**: feldspar
Facies codes for alteration in volcanic rocks

(a) Phase(s)

- mineralogical and textural changes accompany hydrothermal alteration. Each alteration mineral can be referred to as a phase.
- each alteration domain comprises an area of rock that is characterised by a particular alteration mineral assemblage or by different proportions of similar minerals (phases) in similar mineral assemblages.

| C  | chlorite       |
| SI | quartz         |
| H  | hematite       |
| PY | pyrite         |
| S  | sericite       |
| K  | albite/K-feldspar |

e.g. SI-S quartz-sericite (alteration domain comprising quartz and sericite)

(b) Relative abundance (phases - domains)

- the least abundant mineral within an alteration domain is presented on the right hand side (RHS) and the most abundant mineral on the left hand side (LHS).

  e.g. S-SI (sericite-quartz) dominant phase - subordinate phase
- in a rock comprising two or more alteration domains, the phase(s) comprising the dominant domain are presented on the LHS and those of the remaining domains on the RHS in order of relative abundance

  e.g. C / S-SI (chlorite & sericite-quartz domains) dominant - subordinate

(c) Intensity

- allocation of a number to describe the intensity of alteration within each domain

  Weak (1-2) Moderate (3-4) Strong to intense (5-6)

  * e.g. C\(^5\) (strong chlorite alteration)

  S-SI\(^3\) (moderate sericite-quartz alteration)

(d) Controls/textures

The distribution of alteration minerals and domains can be controlled by the pre-alteration texture or superimposed structures. Alternatively, the alteration phases/domains can generate a range of new textures and patterns in the rock.

| x  | crystal      |
| mx | matrix       |
| c  | clasts       |
| fr | fracture (perlite, quench) |
| hf | hydraulic fracture |
| fb | flow banding |
| sh | shear        |
| v  | vein         |
| d  | disseminated |
| am | apparent matrix |
| ac | apparent clast |
| mo | mottled      |
| w  | wash         |
| fi | fiamme       |
| k  | fleck        |
| s  | spotty       |
| pt | patchy       |

- e.g. Cp\(^5\) (strong pervasive chlorite alteration)
- e.g. Cp\(^5\) / SIF\(^3\) (strong pervasive chlorite alteration and moderate, fracture-controlled quartz alteration)
Appendix D

Geochemical analyses of lavas and intrusions

Appendix D1 Mount Windsor Formation
Appendix D2 Trooper Creek Formation
Appendix D3 Trooper Creek Formation
### Appendix D1: Major (wt%) and trace element (ppm) analyses of volcanics from the Mount Windsor Formation and Ti-rich dikes

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**Notes:**
- **MWF** - Mount Windsor Formation
- Total Fe as Fe₂O₃
- Analyses recalculated to 100% anhydrous
Appendix D2: Major (wt%) and trace element (ppm) analyses of volcanics from the Trooper Creek Formation

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TCF - Trooper Creek Formation

1 Total Fe as Fe₂O₃
2 Analyses recalculated to 100% anhydrous
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TCF - Trooper Creek Formation
\(^1\) Total Fe as Fe₂O₃
\(^2\) Analyses recalculated to 100% anhydrous
Appendix 03: Major (wt%) and trace element (ppm) analyses of volcanics from the Trooper Creek Formation at Highway-Reward

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TCF - Trooper Creek Formation

1 Total Fe as Fe₂O₃
2 Analyses recalculated to 100% anhydrous
### Appendix D3 continued

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**TCF** = Trooper Creek Formation

¹ Total Fe as Fe₂O₃
² Analyses recalculated to 100% anhydrous
Appendix E

Geochemical analyses of ironstones

Appendix E1 XRD analyses for massive ironstone
Appendix E2 Major, trace and REE analyses
Appendix E3 Calculations for isocon plots
Appendix E1: XRD analyses for massive ironstone

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Appendix E.2: Major (wt%), trace element (ppm) and rare earth element (ppm) analyses of iron oxide-illiticates rocks from the study area

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Abbreviations: i01=ironstone; tuff=tuffaceous ironstone; bx=breccia; strom=stromatolithic ironstone.
Appendix E2 continued

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Abbreviations: im=ironstone; tuff=silicic tuff; bx=breccia; stsm=sulfuric ironstone.
# Appendix E3: Calculations for isocon plots

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Appendix E3: Calculations for isocon plots

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Appendix E3: Calculations for isocon plots

Trooper Creek prospect - western lenses (95-212, 214, 275)

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Appendix F

Publications


TEXTURAL EFFECTS OF DEVITRIFICATION AND HYDROTHERMAL ALTERATION IN SILICIC LAVAS AND SHALLOW INTRUSIONS, MOUNT READ VOLCANICS (MRV), CAMBRIAN, TASMANIA


Submarine silicic lava flows, domes and shallow intrusions in the MRV comprise coherent, massive and flow banded lava, hyaloclastite and autobreccia. Margins of lavas and intrusions were formerly glassy whereas interiors varied from glassy to crystalline. Perlitic fracturing, devitrification, and hydrothermal and diagenetic alteration acted on primary volcanic textures to generate diverse alteration textures, including false volcaniclastic textures, in the originally glassy parts of the silicic lavas and intrusions.

Perlitic fracturing of glass commenced during cooling of the silicic lavas and intrusions, generating pathways for migrating fluids. Devitrification refers to the nucleation and growth of crystalline minerals in glasses at subsolidus temperatures. “High” temperature devitrification of glass accompanied emplacement, and generated spherulites, lithophysae, and micropoikilitic texture. “Low” temperature devitrification of silicic glass to an assemblage of sericite, chlorite, quartz and feldspar is attributed to interaction with syn-volcanic hydrothermal fluids and early to late diagenetic fluids, and can be referred to as hydrothermal and diagenetic alteration. The textural effects of these alteration processes were strongly influenced by the pre-existing texture which was created by eruption and primary fragmentation, “high” temperature devitrification, and hydration. Textures were either enhanced, modified or destroyed during “low” temperature devitrification.

During lower greenschist facies metamorphism earlier mineral assemblages were recrystallised or replaced by coarse metamorphic minerals, overprinting or mimicking primary and alteration textures.

The outcome of this textural progression is that both coherent and autoclastic facies of silicic lavas and shallow intrusions in the MWV resemble matrix supported, monomict and polymict, welded and non-welded volcaniclastic deposits.
Evaluating the prospectivity of ancient volcanic sequences for volcanic-hosted massive sulfide (VHMS) deposits can be greatly enhanced by identifying original lithologies and emplacement processes (McPhie et al., 1993). In particular, distinguishing between syn-volcanic intrusions, lava flows, domes and cryptodomes and between autoclastic, resedimented volcanioclastic and epiclastic facies is critical in recognizing palaeo-sea floor positions which are important sites for exhalative and shallow sub-surface base metal sulfide accumulation in many VHMS systems. Detailed core logging and petrography of host rocks to the Cu-Au-Pb-Zn Highway and Reward deposits have revealed the nature of volcanic processes in a near vent, subaqueous (submarine), below-wave-base depositional environment.

The volcanic facies architecture at Highway and Reward includes the products of both intrabasinal and basin margin or subaerial eruptions. Rhyolitic, rhyodacitic and dacitic lava domes, partly extrusive cryptodomes, syn-sedimentary intrusions and associated in situ and resedimented autoclastic deposits are from an intrabasinal source. Contact relationships and phenocryst mineralogy, size and percentages indicate the presence of up to nine distinct porphyritic units within an area of approximately 1 x 1 x 0.5 km at Highway-Reward. Massive coherent and flow banded lava, hyaloclastite, autobreccia and peperite are the main component facies of the porphyritic units. Peperites vary from sediment-matrix-supported breccias in which porphyry clasts are sparse (dispersed peperite), through sediment-poor jigsaw-fit aggregates of porphyry clasts (compact peperite), to relatively coherent porphyry enclosing isolated stringers and/or globules of sediment. Porphyry clasts vary from blocky with curviplanar margins (blocky peperite) to lenticular with ragged margins (ragged peperite), which may reflect, respectively, the relative importance of cooling contraction granulation and dynamic stressing of chilled lavas surfaces during emplacement. The peperitic upper margins to many porphyry sheets demonstrate their intrusion into wet unconsolidated sediments. The high relative density of magma to wet sediment favoured emplacement as syn-sedimentary intrusions rather than extrusions (cf. McBirney, 1963; Walker, 1989). Dewatering and induration of the sediment pile by early syn-sedimentary intrusions may have favoured the subsequent eruption of lava domes and partly emergent cryptodomes at Highway-Reward. The shape and distribution of lava domes and cryptodomes was further influenced by the positions of previously or concurrently emplaced porphyritic units, and possibly by syn-volcanic faults which may have acted as conduits for magma. Because they are constructional, lava domes and cryptodomes influenced subsequent volcanioclastic sedimentation. Lava domes, cryptodomes and deposits of resedimented hyaloclastite sourced from over-steepened dome margins are an important indicator of palaeo-sea floor positions.

Porphyries intruded or were overlain by volcanioclastic and sedimentary facies association comprising suspension-settled siltstone, graded turbiditic sandstone and thick mass-flow-emplaced pumiceous- and crystal-rich sandstone-breccia. Pumiceous mass-flow deposits are emplaced rapidly in large volumes, erupted infrequently and are widely distributed (McPhie & Allen, 1992), and so provide an important framework for correlation within the Trooper Creek Formation at Highway-Reward. Quartz-feldspar and feldspar only, pumiceous and crystal-rich sandstone-breccia units are non-welded, up to 65 m thick, and normally graded with fine grained tops, and in some instances, polymict lithic-rich bases. Deposition from high-concentration turbidity currents sourced from explosive eruptions at a subaerial or shallow subaqueous basin margin centre is suggested.

Perlitic fracturing, devitrification, hydrothermal and diagenetic alteration have acted on originally glassy parts of lavas and intrusions, and pumiceous breccias to generate diverse alteration textures, including false volcanioclastic and welding textures. Alteration of lavas commenced during cooling from magmatic
temperatures (high temperature devitrification) generating spherulites, micropoikilitic texture and lithophysae. Hydration of residual glass to form perlitic fractures supplemented fracture and matrix permeability generated by autoclastic processes, both of which were important for migration of fluids during hydrothermal and diagenetic alteration. Hydrothermal and diagenetic alteration were also influenced by textural and compositional domains generated during high temperature devitrification. Apparent polymict and monomict volcaniclastic textures formed during this textural progression further evolved during greenschist facies metamorphism and tectonic deformation. Pumiceous breccias show the textural effects of early polyphase diagenetic and syn-volcanic hydrothermal alteration. Initial heterogeneous quartz-feldspar alteration replaced glassy vesicle walls of individual pumice shreds and domains within breccias, thereby largely preserving non-welded tube-vesicle textures. Remaining pumice clasts were phyllosilicate-altered and flattened by diagenetic compaction, resulting in false welding textures. Intensely silicified pumice shreds isolated in chloritic domains resemble felsic volcanic lithic fragments.

The density and complexity of non-explosive, coherent, intrusive-extrusive units at Highway-Reward is similar to that described by Horikoshi (1969) for Kuroko host sequences in the Miocene Kosaka Formation of NE-Japan. Analogues of the initial, explosive, tuff cone forming eruptions at the "Kosaka volcano" are not recorded in the stratigraphy at Highway-Reward, possibly reflecting differences in the volatile content of erupted magma, and/or the external confining pressure (lithostatic and hydrostatic pressure).

References:


A SILICIC SUBMARINE SYN-SEDIMENTARY INTRUSIVE - DOME - HYALOCLASTITE HOST SEQUENCE TO MASSIVE SULFIDE MINERALISATION: MOUNT WINDSOR VOLCANICS, CAMBRO-ORDOVICIAN, AUSTRALIA

DOYLE, M.G., and McPHIE, J., C.O.D.E.S., University of Tasmania, Hobart, Tasmania 7001, Australia.

The Cu-Au-Pb-Zn Highway and Reward massive sulfide deposits are hosted by a silicic intrusive and volcanic sequence intercalated with sedimentary facies that indicate a submarine, below-storm-wave-base environment of deposition. Contact relationships and phenocryst mineralogy, size and percentages indicate the presence of up to nine distinct porphyritic units in an area of 1 x 1 x 0.5 km. The peperitic upper margins to many porphyries demonstrate their intrusion into wet unconsolidated-sediment. Syn-sedimentary intrusions, partly emergent cryptodomes, lava domes, and associated in situ and reworked autoclastic deposits have been recognised. These are the principal facies in the environment of mineralisation and represent a proximal facies association from intrabasinal, intrusive/extrusive, non-explosive magmatism. The shape, distribution and emplacement mechanisms of porphyritic units were influenced by: (a) the relative density of magma to wet sediment; (b) the positions of previously or concurrently emplaced porphyries; and (c) possibly by syn-volcanic faults which may have acted as conduits for magma. Lava domes, partly emergent cryptodomes, and deposits of reworked hyaloclastite and peperite are important indicators of palaeo-sea-floor positions at Highway-Reward. Sills and cryptodomes may have influenced sea-floor topography and therefore sedimentation, but do not mark sea-floor positions. Massive sulfide ores are primarily subsea-floor syn-volcanic replacements of the host sedimentary rocks, syn-sedimentary intrusions, lava domes, and autoclastic breccia.

Porphyries intruded or were overlain by a volcanoclastic and sedimentary facies association comprising suspension-settled siltstone, graded turbiditic sandstone and thick, non-welded pumice- and crystal-rich sandstone-breccia. Pumiceous and crystal-rich deposits record episodes of explosive silicic volcanism in an extrabasinal or marginal basin environment, and were emplaced by cold, water-supported, high-concentration turbidity currents.
EVALUATION OF THE ROLE OF CAMBRIAN GRANITES IN THE GENESIS OF WORLD CLASS VOLCANOGENIC-HOSTED MASSIVE SULPHIDE DEPOSITS IN TASMANIA

Ross R. Large\textsuperscript{1}, Mark Doyle\textsuperscript{1}, David Cooke\textsuperscript{1} and Ollie Raymond\textsuperscript{2}
\textsuperscript{1}CODES Key Centre, Geology Dept., University of Tasmania, Hobart TAS 7005
\textsuperscript{2}AGSO, GPO Box 378, Canberra ACT 2601

Summary - New data on the distribution, composition and alteration zonation of Cambrian granites in the Mt. Read Volcanics provide evidence that there may have been a direct input of magmatic fluids during the genesis of the copper-gold volcanogenic-hosted massive sulphide (VHMS) mineralisation in the Mt. Lyell district.

INTRODUCTION

There has been considerable debate on the role of granitic magmas during the generation of volcanic hosted massive sulphide deposits; are they simply heat engines driving seawater (e.g. Ohmoto & Rye 1974, and Solomon 1976) or do they directly supply magmatic components to ore-forming solutions (e.g. Henley & Thornley 1979, Stanton 1985)? Pioneering research by Solomon and his students in the Mount Read Volcanics (e.g. Solomon 1976, Solomon 1981, Polya et al. 1986 and Eastoe et. al. 1987) clearly demonstrated a relationship between hydrothermal alteration and sulphur isotope zonation around the granites, indicating that the granites acted as heaters for the ore-forming convective fluid. In this paper we provide evidence to suggest that the Cambrian granites may have also provided important metal contributions to the ore-forming fluid, especially Fe, Cu, Au, P, F ± Ti and Zr.

FACTORs LINKING THE CAMBRIAN GRANITES TO MINERALISATION

Distribution: Two narrow bodies of Cambrian granite (Murchison Granite and Darwin Granite) intrude the eastern margin of the Central Volcanic Complex (CVC) in the Mt. Read Volcanics. Interpretations based on magnetic and gravity data indicate that the two granite bodies form a semi-continuous narrow vertical sheet of granite 65 km long and about 2 km wide. A series of copper-gold and base metal prospects occur along the margins of the granite sheet (e.g. Prince Darwin, Jukes Pty., Lake Selina). The Mt. Lyell Cu-Au VHMS deposits are located immediately west of the projected continuation of the subsurface granite.

Timing: Previous mapping by Corbett (1989) suggested that the Murchison granite intruded the Tyndall Group volcanics (which unconformably overlie the CVC) and is therefore younger than the VHMS deposits. However, later work (e.g. Corbett, 1992) has revised this interpretation, and recent dating by Perkins and Walsh (1993) has confirmed that the Murchison granite has an age of 501 ± 5.7 Ma (Ar/Ar), the same age as the host rocks to the massive sulphide deposits.

Composition: Both the Murchison and Darwin granites are high-K, magnetite series granites which show anomalous enrichment in barium and potassium. The Murchison granite varies in composition from granodiorite to granite (58 to 78% SiO\textsubscript{2}; Abbott, 1992), while the Darwin granite is composed of two highly fractionated granite phases (74-78% SiO\textsubscript{2}; Jones, 1993). K\textsubscript{2}O varies up to 8.5% and Ba up to 3000 ppm; however, some of this enrichment is related to alteration.

Alteration: Well developed zones of hydrothermal alteration have been mapped around the margins of the granites (e.g. Polya et al. 1986, Eastoe et. al. 1987, Hunns 1987, Doyle 1990). An extensive zone (Z\textsubscript{1}) of pink K-feldspar alteration extends from the outer part of the granites into the surrounding volcanics. An overlapping shell (Z\textsubscript{2}) of chlorite ± pyrite ± magnetite alteration overprints and extends outwards from the K-feldspar zone. Sericite-chlorite ± pyrite forms a distal alteration zone (Z\textsubscript{3}). At both Jukes Pty. and Lake Selina, Cu ± Au mineralisation occurs in the chlorite ± pyrite ± magnetite zone (Z\textsubscript{2}).
Magnetite-apatite association: The strongest link between the granites and VHMS Cu-Au mineralisation is provided by the common occurrence of magnetite-apatite-Cu ± Au vein style and disseminated mineralisation both within the Z2 alteration halo of the granites and within the centre of the Prince Lyell ore deposit in the Mt. Lyell VHMS district. A good linear correlation exists between Cu and P₂O₅, and Fe and P₂O₅ both within the mineralised alteration halo of the granites and in the Prince Lyell ores. Oxygen isotopes indicate that the magnetite veins within the granite halo and the Prince Lyell deposit have δ¹⁸O values that are consistent with a magmatic source (Doyle 1990, Raymond 1993). Apatite, which is commonly intergrown with magnetite, pyrite and chalcopyrite, has consistently high F/Cl ratios, with a mean of about 6 wt% F.

RELATIONSHIP OF COPPER-GOLD TO LEAD-ZINC-COPPER VHMS DEPOSITS

The Mt. Lyell field contains both stringer-style copper-gold deposits such as Prince Lyell and separate stratiform lead-zinc-copper deposits such as Comstock and Tasman & Crown Lyell Extended. Most previous workers (e.g. Solomon 1976, and Walshe & Solomon 1981) consider that the Cu-Au and Pb-Zn-Cu deposits formed as part of the same hydrothermal system; the Cu-Au stringer-style forming by subsurface replacement and the Pb-Zn-Cu massive sulphides by contemporaneous seafloor exhalation. Although our work suggests a source for Cu and Au from the Cambrian granites, the source for Pb, Zn, Ag and S remains unresolved and may be either magmatic or related to seawater leaching.

CONCLUSIONS

Cambrian granites in the Mt. Read Volcanics form a thin linear discontinuous sheet 65 km long which is spatially related to Cu-Au mineralisation, including the VHMS deposits at Mt. Lyell. The highly fractionated, oxidised, magnetite series granites have overlapping alteration shells of K-feldspar, chlorite-magnetite and sericite. Preliminary evidence suggests that the VHMS copper-gold mineralisation at Mt. Lyell may be associated with fluids enriched in Fe-Cu-Au-P₂O₅-F-Zr-Ti released directly from the granite magma.

REFERENCES


Acknowledgements: Permission to use geochemical data for Cambrian granites from the Mineral Resources Tasmania data-base, and from Steve Hunns for unpublished analyses on the Lake Selina prospect is gratefully acknowledged.
Evaluation of the role of Cambrian granites in the genesis of world class VHMS deposits in Tasmania

Ross Large, Mark Doyle, Ollie Raymond, David Cooke, Andrew Jones, Lachlan Heasman

Abstract

An analysis of the distribution, composition and alteration zonation of Cambrian granites which intrude the Mt Read Volcanics of western Tasmania provides evidence that there may have been a direct input of magmatic fluids containing Fe, Cu, Au and P to form the copper-gold volcanic-hosted massive sulphide (VHMS) mineralisation in the Mt Lyell district.

Interpretation of regional gravity and magnetic data indicates that a narrow discontinuous body of Cambrian granite (2-4 km wide) extends along the eastern margin of the Mt Read Volcanic belt for over 60 km. The Cambrian granites are altered magnetite series types which show enrichment in barium and potassium, and contrast markedly with the fractionated ilmenite series Devonian granites related to tin mineralisation elsewhere in the Dundas Trough.

Copper mineralisation occurs in a linear zone above the apex of the buried Cambrian granite body at the southern end of the belt, from Mt Darwin to the Mt Lyell district over a strike length of 25 km. Gold and zinc mineralisation are concentrated higher in the volcanic stratigraphy more distant from the granite. Overlapping zones of alteration extend from the granite into the surrounding volcanic rocks. An inner zone of K-feldspar alteration is overprinted by chlorite alteration, which passes outwards into sericite alteration. Magnetite ± pyrite ± chalcopyrite ± apatite mineralisation is concentrated in the chlorite alteration zone as veins and low grade disseminations. The Mt Lyell copper-gold stringer and disseminated mineralisation is hosted in felsic volcanic rocks 1 to 2 km west of the interpreted buried granite position. Magnetite-apatite ± pyrite veins in the Prince Lyell deposit at Mt Lyell are very similar to the veins in the halo of the granite. Further south, and provide evidence for magmatic fluid input during the formation of the copper-gold VHMS deposits.

A model involving deeply penetrating convective seawater, mixing with a magmatic fluid released from the Cambrian granites, best explains the features of VHMS mineralisation in the Mt Lyell district.

1. Introduction

There has been considerable debate over the past 25 years on the role of granitic magmas during the generation of volcanic-hosted massive sulphide (VHMS) deposits. Some workers (e.g. Uribe and Sato, 1978; Henley and Thornley, 1979; Sawkins and Kewalik, 1981; and Stanton, 1985; Stanton, 1990) have argued for a direct input of volatiles and metals from the magma to form the ore solutions, while others (e.g. Kajiwara, 1973; Spooner and Fyfe, 1973; Ohnoe and Rye, 1974; Solomon, 1976; Large, 1983).