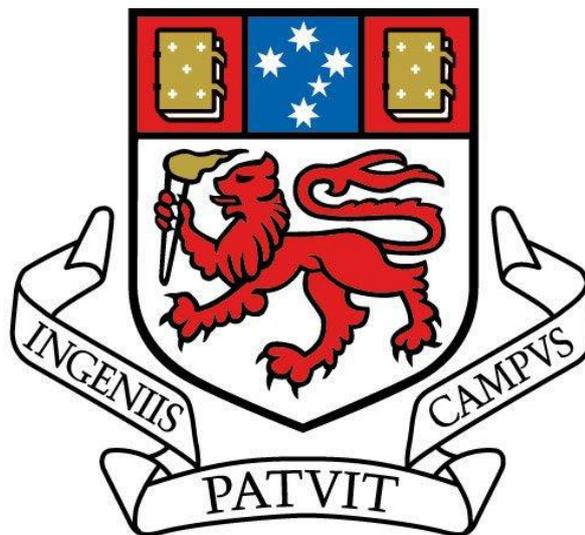


# Correlations and Exploration Significance of the Natone Volcanics, Rosebery Group, Western Tasmania

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# UNIVERSITY OF TASMANIA

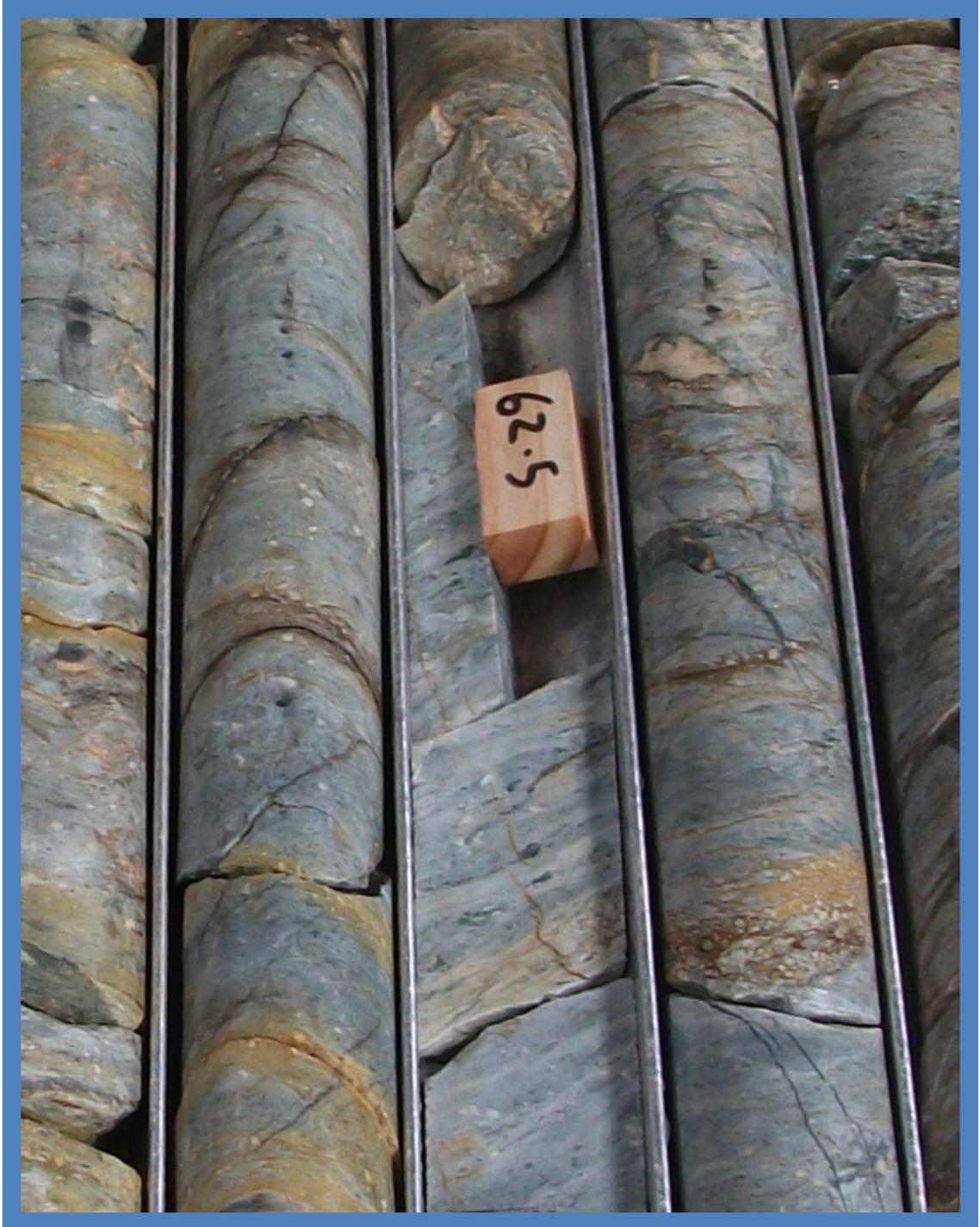
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**School of Earth Sciences**

*A research thesis submitted in partial fulfilment of a Graduate Diploma  
Science (Geology) August 2013*



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## **Declaration**

This thesis contains no material which has been accepted for the award of any other degree or diploma in any institution and, to the best of my knowledge and belief, contains no copy or paraphrase material previously written or published by another person, except where due reference is made in the text of the thesis.

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## Abstract

The significance of this study is to determine the likely correlations of the Natone Volcanics, and therefore its prospectivity in terms of Rosebery-Hercules type VHMS.

Logging of 2 drill holes indicates the Natone Volcanics are a largely homogenous unit comprising weakly chlorite-sericite altered, massive ungraded, volcanic lithic, fiamme-bearing, quartz-feldspar-phyric sandstone and siltstone. The volcanic components are interpreted to have been deposited from a single eruptive event, and the presence of minor lithics indicates emplacement as a mass flow.

The Natone Volcanics and enclosing sequences include two volcanogenic-sedimentary facies associations, one volcanic facies association, one coherent facies, and one sedimentary facies: Interbedded Fine-Grained Sandstone and Mudstone; Fuchsitic Conglomerate and Sandstone (Salisbury Conglomerate correlate); Volcanic-lithic Sandstone and Siltstone (Natone Volcanics); Coherent Dacite; and Quartz-Rich Sandstone (Stitt Quartzite correlate). The Natone Volcanics conformably overlie the Salisbury Conglomerate correlate, and a fault contact is inferred from changes in younging direction with the stratigraphically younger Stitt Quartzite correlate. A thin (2 to 4 m) dacite intrusion separates the Stitt Quartzite correlate from the Natone Volcanics, the timing of which is unknown. The majority of the Natone Volcanics are overturned, dipping west and facing east; however, at least one fold is present below the sequence. The internal structure within the Natone Volcanics is obscured by alteration and local cleavage development.

Whole rock compositional data has been used to classify the Natone Volcanics as suite I (Crawford *et al*, 1992). Major element relationships show that the Natone Volcanics closely match upper White Spur Formation and Rosebery hangingwall data; however do not match samples taken from below the Rosebery Fault. Petrographically, the Natone Volcanics are most similar to the middle sub unit of the White Spur Formation described by Lees (1987). The dacite intrusion is also inferred to be suite I (Crawford *et al*, 1992), and plots within range of Rosebery hangingwall data.

The provenance of detrital chromite grains found in units enclosing the Natone Volcanics was constrained by comparing element ratios with samples from in and around the Mount Read Volcanics. Chromite found in the Fine-Grained Sandstone and Mudstone is similar to chromite in Mafic-Ultramafic Complexes (MUC). Chromite from the Salisbury Conglomerate correlate and Stitt Quartzite correlate are also similar to chromite in MUC; and overlap slightly with chromite in Hellyer Basalt, and even less so with samples of chromite in the Crimson Creek Formation (CCF). The mixture of chromite types supports the Rosebery Group having a local intrabasinal provenance.

U-Pb dating of a sample from the Natone Volcanics indicates that deposition occurred  $498.26 \pm 0.78$  Ma (Mortensen *et al*, unpublished). This date, together with petrographic information and compositional analyses, confirms that the Natone Volcanics are a correlate of the White Spur Formation and Rosebery hangingwall volcanoclastic facies. This is an important finding in terms of exploration beneath and west of the Rosebery Fault.

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## Acknowledgements

This work represents a milestone achievement in my life. A decade ago I would never have imagined myself to be a University graduate, let alone submitting a dissertation such as this. For that I would like to thank myself, for I am proud of my achievement.

It occurs to me now that behind every author are myriad unwitting assistants, in my case these include the underappreciated Wikipedia, the overrated Google, Bing, and the UTAS library database.

On a more serious note and for their unwavering help and assistance I would like to thank my parents. If it were not for their encouragement, financial assistance, and in particular my mother's editorial skills, then this would not have been possible. Next I would like to thank my primary supervisor Dr. Andrew McNeill for his patience and faith in me. I have always found him to be approachable and helpful with his constructive feedback. Furthermore I extend gratitude to my secondary supervisor Dr. Jocelyn McPhie. Her feedback has been instrumental in connecting the dots, crossing the t's and dotting the i's. Other members of faculty I would like to thank include Dr. Garry Davison, Dr. Rebecca Carey, Dr. Ron Berry, Dr. Dima Kamenetsky and Dr. Sebastien Meffre for their advice, enthusiasm, encouragement and guidance through the project.

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## 1.0 Introduction to the Natone Volcanics

### 1.1 Introduction

The Cambrian Mount Read Volcanics (MRV) of western Tasmania has been described as one of the most mineralized areas in the world (Corbett, 1992). The sequence forms a belt along the western margin of the Neoproterozoic Tyennan Region, extending from Elliott Bay in the south to Deloraine in the northeast (Figure 1.1).

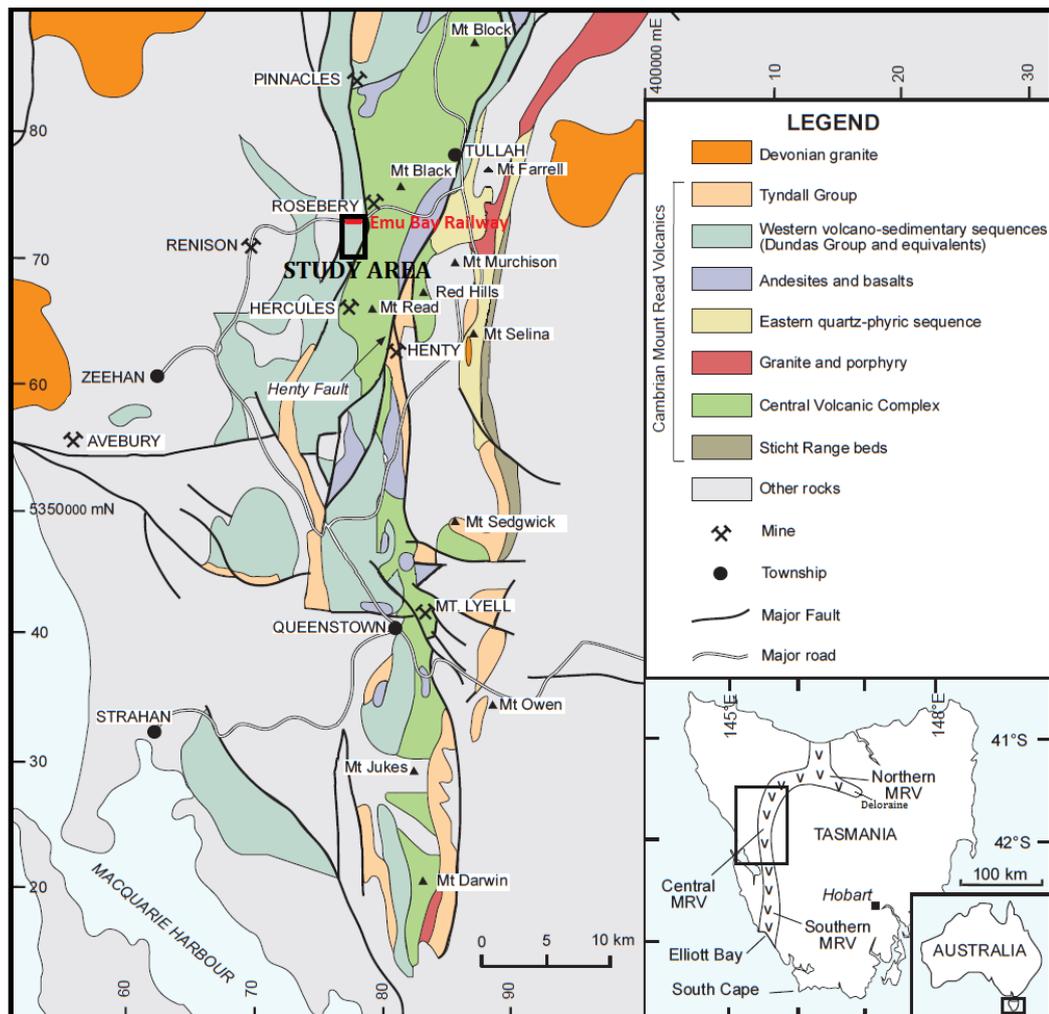


Figure 1.1: The Mount Read Volcanics and Dundas Group with the location of the study area shown (modified from Gifkins *et al.*, 2005).

Due to the economic significance of the MRV, the structure, stratigraphy, volcanology and geochemistry has been extensively studied (e.g. Corbett and Lees, 1987; Berry, 1992; Crawford *et al.*, 1992; McPhie and Allen, 1992). However, the geology of some areas, largely on the periphery of the main volcanic belt, remain poorly understood. One of these areas is west of the Rosebery Fault, between Rosebery and Colebrook Hill, where the Rosebery Group of Green (1983) outcrops.

The Rosebery Group has historically been a subject of confusion (Green, 1983). The sequence includes correlates of the MRV, the Chamberlain Shale, and other units of uncertain affinities, such as the Westcott Argillite and the Salisbury Conglomerate. In the western part of this area, trending north to south, there is an approximately 170 m-wide

band of volcanoclastic rocks called the Natone Volcanics by Campana and King (1963). The overall distribution of the Natone Volcanics is well defined; however their internal structure, contacts with enclosing units, provenance and potential correlations are poorly understood. In this study, new information on the geology of the Natone Volcanics is provided, with the aim of establishing the regional correlation of this unit.

## 1.2 Aims and Significance

The key aim of this project is to complete a combined study of the volcanology, sedimentology and provenance of the Natone Volcanics, in order to determine the likely correlations of the sequence, and therefore their prospectivity in terms of Rosebery-Hercules type Volcanic Hosted Massive Sulphide (VHMS) deposits. The Pb-Zn/Cu rich Rosebery deposit is an important example of its eponymous sub-class, the Rosebery-Hercules type (Large *et al.*, 2001; Green *et al.*, 2006). The Rosebery-Hercules host unit is in fault contact with the Mount Black and Sterling Valley Volcanics to the east (Mount Black fault) and the volcanoclastic units of the Dundas Group to the west (Rosebery Fault). The distribution of the Rosebery-Hercules host unit has been an important focus of regional exploration in the MRV (e.g. Gifkins and Allen, 2001). Overlying the host sequence are feldspar-quartz-phyric volcanoclastic units which are correlated with the White Spur Formation of the Dundas Group (Large *et al.*, 2001). Green (1983), Lees (1987) and Parfrey (1993) have described the Natone Volcanics as lithologically similar to the White Spur Formation, largely on the basis of the presence of abundant quartz phenocrysts, and thus hangingwall to the Rosebery ore position.

## 1.3 Methods

This study is based on detailed descriptions of two drill holes that intersect the Natone Volcanics (DDH MD-1 and BH-2) (Green, 1983) and geological mapping of outcrops near the Murchison Highway. These data are used to establish contact relationships, internal stratigraphy and facies characteristics of the Natone Volcanics. Eleven representative samples from MD-1 and six samples from BH-2 were thin sectioned for petrographic description. Thin section descriptions and drill log results are included in Appendix 1.

Detrital chromite was identified in four samples and these were analysed using electron microprobe, to determine their composition. The whole rock composition of three samples from the Natone Volcanics in BH-2 was determined by a combination of X-ray Fluorescence (XRF) and Inductively Coupled Mass Spectrometry (ICP-MS) analysis; all analyses results are discussed in Chapter 4, and analytical conditions are included as Appendix 2.

## 1.4 Study Area

The study area is located from the Murchison Highway approximately 2.5 km west of the town of Rosebery, extending south along Natone Creek to the Rosebery Fault in the region of the Jupiter prospect (Figure 1.1; Figure 1.2; Figure 1.3). The unit is well exposed over a width of about 120 m in a cutting along the Emu Bay railway.

## 1.5 Previous Studies

Units within the Rosebery Group, and their structural and stratigraphic relationships, have been notoriously difficult to define. Green (1983) wrote that subdivision of the group into formations has been a subject of much confusion, limited by a lack of consideration of facing evidence and contact relationships. Corbett and Lees (1987) summarized evolution of the naming of the units in the Rosebery Group (Figure 1.2) and asserted that

recognition of the Rosebery Fault and stratigraphic correlations with the Dundas Group have largely removed historical confusion. For the purposes of this study the stratigraphic nomenclature of Green *et al.* (1981) will be used (Table 1.1).

**Table 1.1: Stratigraphic nomenclature of the Rosebery Group (Green *et al.*, 1981)**

(i)	(ii) Central Units	(iii) Eastern Units
Munro Creek Formation	Natone Volcanics Salisbury Conglomerate Mudstone-dominated sequence	Stitt Quartzite Westcott Argillite Chamberlain Shale

Campana and King (1963) first identified and named the Natone Volcanics in their research on Palaeozoic tectonism, sedimentation and mineralization in western Tasmania. Green (1983) defined the overall distribution of the Natone Volcanics and provided a summary of units west of the Rosebery Fault. This work brought to light the presence of overturned units and facing changes, and the presence of a syncline at Pieman River gorge. Lees and Corbett (1987) modified previous descriptions of the Natone Volcanics, and studied the stratigraphic and structural relationships at the western margin of the MRV, concluding that a complex system of faults, overturned beds and folds is present in the sedimentary sequences west of the Rosebery Fault. Parfrey (1993) logged DDH BH-1 and provided the first whole-rock compositional data of the Natone Volcanics.

Corbett and McNeill (1986) mapped the rocks of the Natone Volcanics as part of a larger group of generally quartz-phyric felsic tuff and agglomerate. To the west of this map (Corbett and McNeill, 1986), the Natone Volcanics abut a sequence defined as a polymictic conglomerate with fuchsite clasts, and to the east is quartzwacke interbedded with black pyritic mudstone. Corbett (2002) provided the most recent regional map of the area showing rocks in the area of the Natone Volcanics as part of a marine sandstone-siltstone-conglomerate sequence (Early Ordovician–Late Cambrian Owen Conglomerate and correlates), fault bound to the west and in contact to the east with quartz-feldspar crystal-rich volcanoclastic sandstone, siltstone and conglomerate (Middle Cambrian Tyndall Group and correlates). This map (Corbett, 2002) suggests that units west of the Rosebery Fault are younging to the west.

## 1.6 Thesis Outline

Chapter 2 provides a discussion of the regional geology of the Mt Read Volcanics to set the context for a detailed discussion of the Rosebery Group, which hosts the Natone Volcanics. Chapter 3 is a description of facies observed within the Natone Volcanics and enclosing units, based on observations made of DDH MD-1 and BH-2 (Figure 1.3). Chapter 4 outlines the whole rock and detrital chromite compositions of the Natone Volcanics and enclosing sequences, and considers implications for correlations. Chapter 5 discusses the results of the present study.

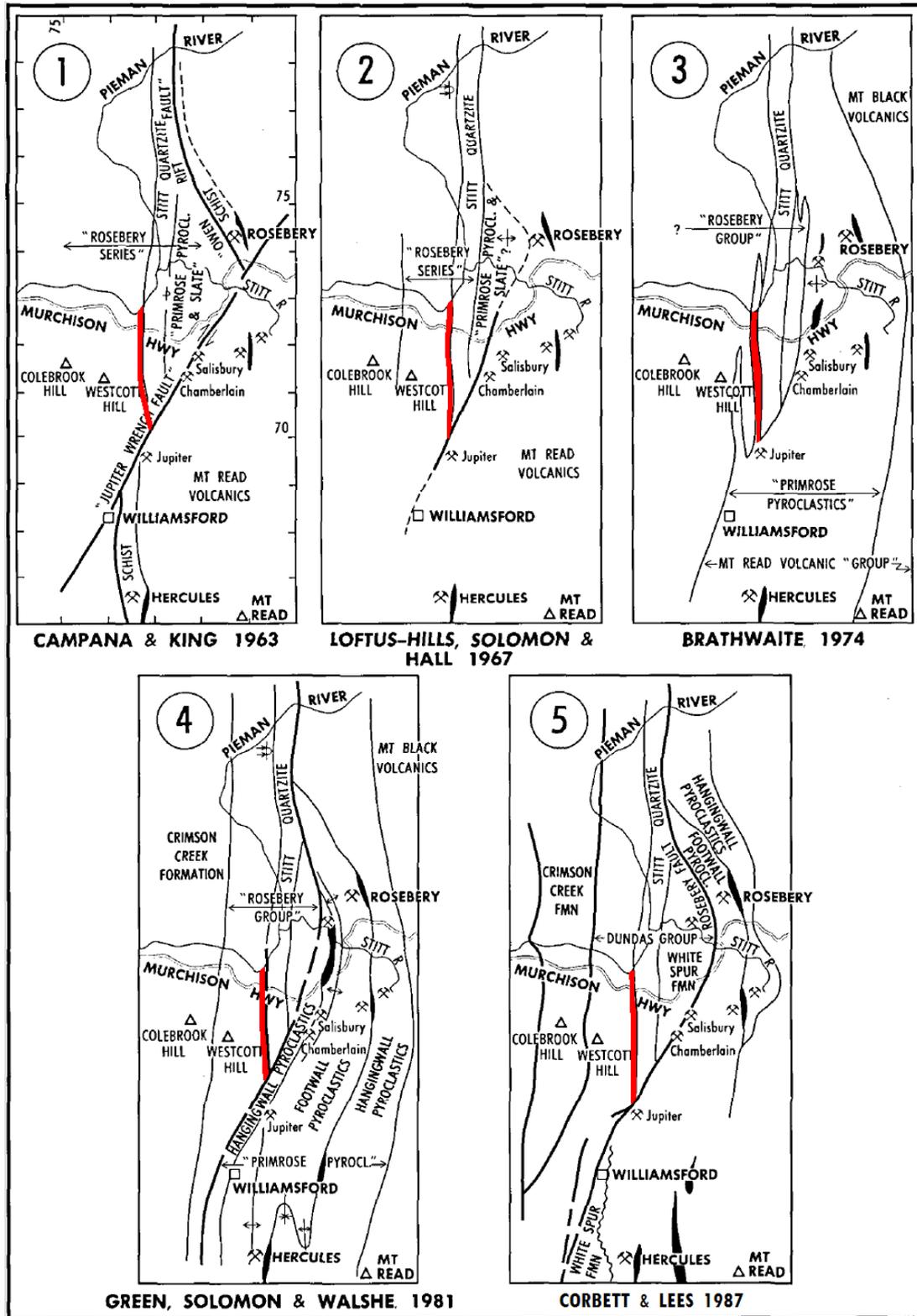


Figure 1.2: Historical summary of different subdivisions of the sequence west of the Rosebery Fault showing the distribution of the Natone Volcanics in red (modified from Corbett and Lees, 1987).

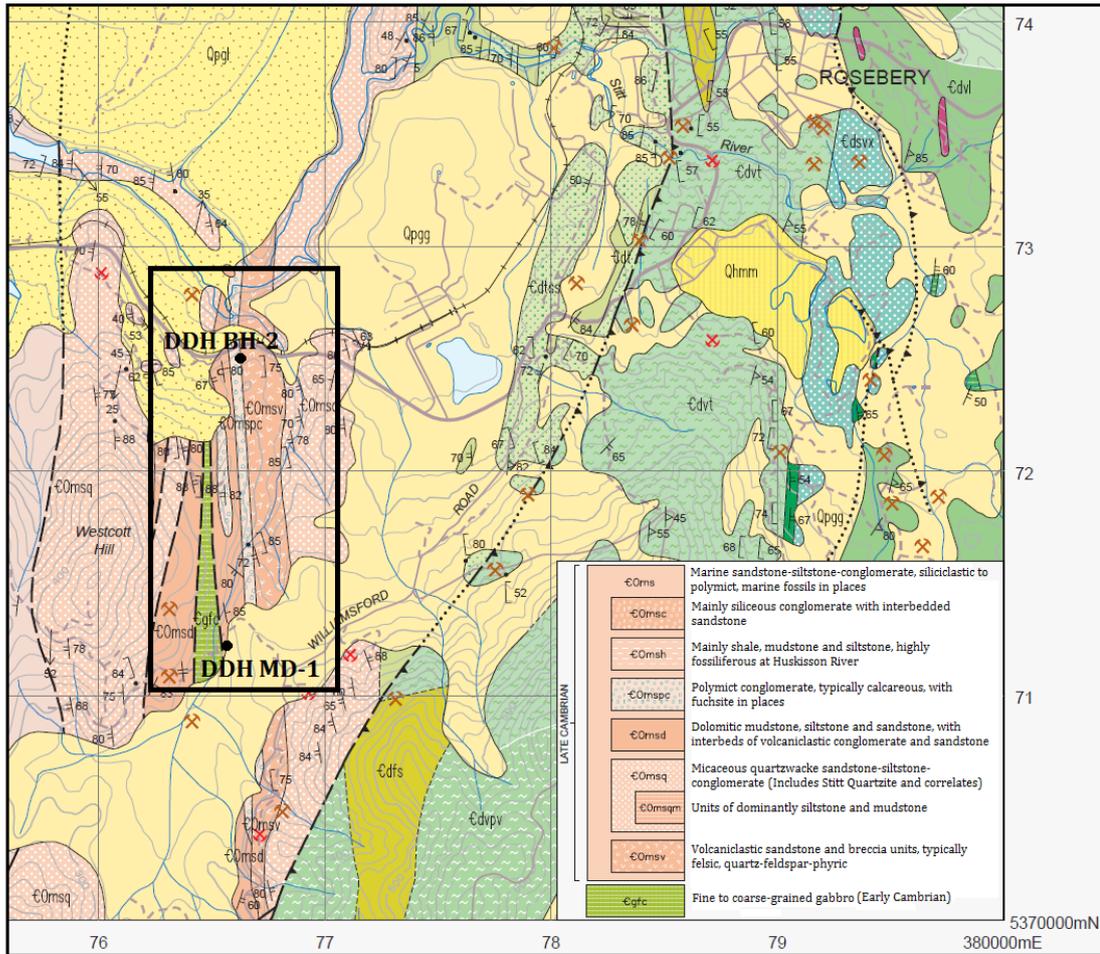


Figure 1.3: Geological map of the study area showing approximate location of drill holes (modified from Seymour and McMclenaghan, 2003).

## 2.0 Regional and Local Geology

### 2.1 Introduction

This chapter summarizes the geology of the MRV, western Tasmania, in order to set the context for a detailed discussion of the stratigraphy and setting of the Rosebery Group of Brathwaite (1974) and Green *et al.* (1981).

The MRV are a 20 x 200 km belt, extending from Elliot Bay in the southwest to Deloraine in the northeast (Figure 2.1), of volcanic and sedimentary rocks of Middle Cambrian age formed along the western margin of the Neoproterozoic Tyennan Region (Corbett, 1992). The MRV host six world class VHMS ore bodies: the Pb-Zn-rich polymetallic Que River; Hellyer; Rosebery, and Hercules deposits; the hybrid VHMS Au±Cu Henty deposits; and the Mount Lyell deposits (Corbett, 1992). As a result of their economic importance, the VHMS deposits of the MRV have been studied in some detail (Berry, 1992; Crawford, 1992; McPhie and Gemmell, 1992; McNeill, 2000; Corbett, 2002; Green *et al.*, 2006).

The MRV include silicic, intermediate and mafic lavas, syn-eruptive and re-worked volcanoclastic deposits, and syn-volcanic intrusions intercalated with non-volcanic sedimentary facies. McPhie and Allen (1992) suggested that the deposits formed below wave base in a marine environment. Although the MRV were emplaced over a period of at least 12.4 m.y., the majority of VHMS deposits formed at  $\sim 500 \pm 1$  Ma (Mortensen *et al.*, unpublished).

### 2.2 Allocthonous facies of the Arthur Lineament

The Arthur Lineament transects the northwest of Tasmania (Figure 2.1). Upper Neoproterozoic – Lower Cambrian allocthonous facies of the Togari Group dominate the 10 km wide tectonic feature. These rocks rest unconformably or disconformably on older successions (Seymour *et al.*, 2006).

#### 2.2.1 Crimson Creek Formation

The Crimson Creek Formation (CCF) represents a correlate of the Togari Group sedimentary and mafic volcanic succession. The group can be subdivided into four main phases: (1) a mid-Cryogenian lower dolomitic succession with basal siliceous conglomerate-sandstone; (2) a late Cryogenian to early Ediacaran phase of mafic rift volcanism and associated volcanoclastic sedimentation; (3) mid-late Ediacaran renewal of shallow-marine carbonate sedimentation; and, (4) at the top, a Cambrian phase of deep-water siliciclastic sedimentation (Seymour *et al.*, 2006).

Foden (1973) identified the olivine tholeiites of the CCF as 'abyssal tholeiites'. An early rifting or spreading phase is commonly suggested to explain the abundance of tholeiites and associated ultramafic facies in the lower sequences of the Dundas Trough (Burrett and Martin, 1989).

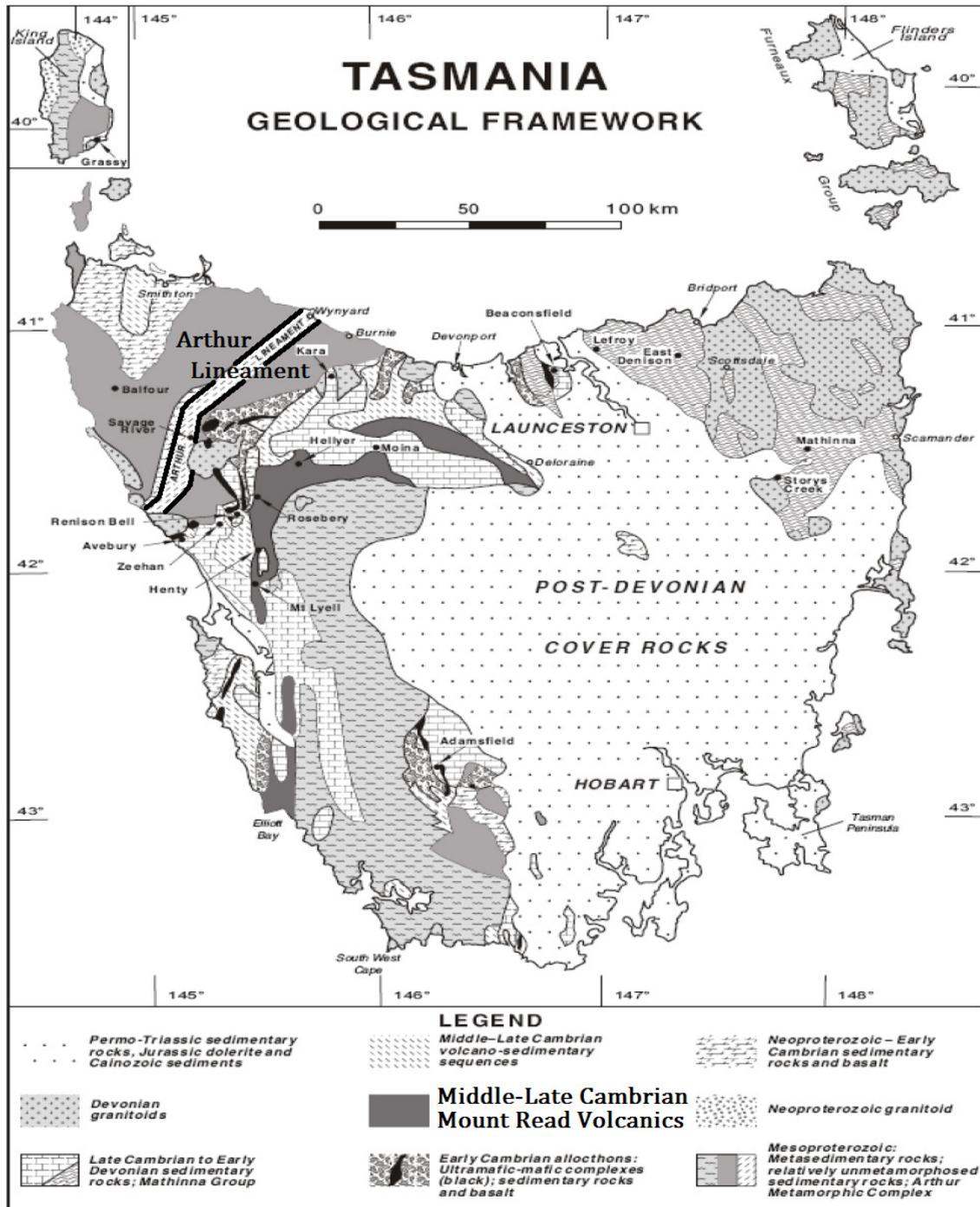


Figure 2.1: Distribution of the Mount Read Volcanics and the Arthur Lineament (modified from Heard, 2007).

### 2.2.2 Cleveland-Waratah association

Considered as emplacement products of the Early Cambrian Tyennan Orogeny, the Cleveland-Waratah association comprise lithicwacke, red mudstone, chert, mafic volcanic rocks with Ocean Floor Basalt characteristics, and rare carbonate rocks which host Devonian skarn mineralisation at the Cleveland mine (Seymour *et al.*, 2006). Brown (1988), utilizing rare earth elements (REE) and Sm-Nd isotopes, indicated that the tholeiites in the Cleveland-Waratah area are sub-alkaline with 'Ocean Floor Basalt'

affinities, whereas those in the lower Pieman River area are sub-alkaline to alkaline with affinities 'Within Plate Basalt'.

In the Pieman River Gorge, abutting the western side of the Munro Creek Formation, Green (1983) observed a green to maroon mudstone sequence interbedded with graded beds of lithicwacke and finer grained sandstone. The components are predominantly sericitized plagioclase, mafic lava fragments, clinopyroxene and chloritic fragments in a chlorite-rich matrix (Green, 1983).

### 2.3 Stratigraphy of the Mount Read Volcanics

The MRV have been divided into five major lithostratigraphic units (Figure 2.2): (1) the Sticht Range Beds; (2) an Eastern Quartz-Phyric Sequence; (3) the Central Volcanic Complex (CVC); (4) The Tyndall Group; and (5) the Western Volcano-Sedimentary Sequences (Dundas Group and equivalents with andesites and basalts) (Corbett, 1992).

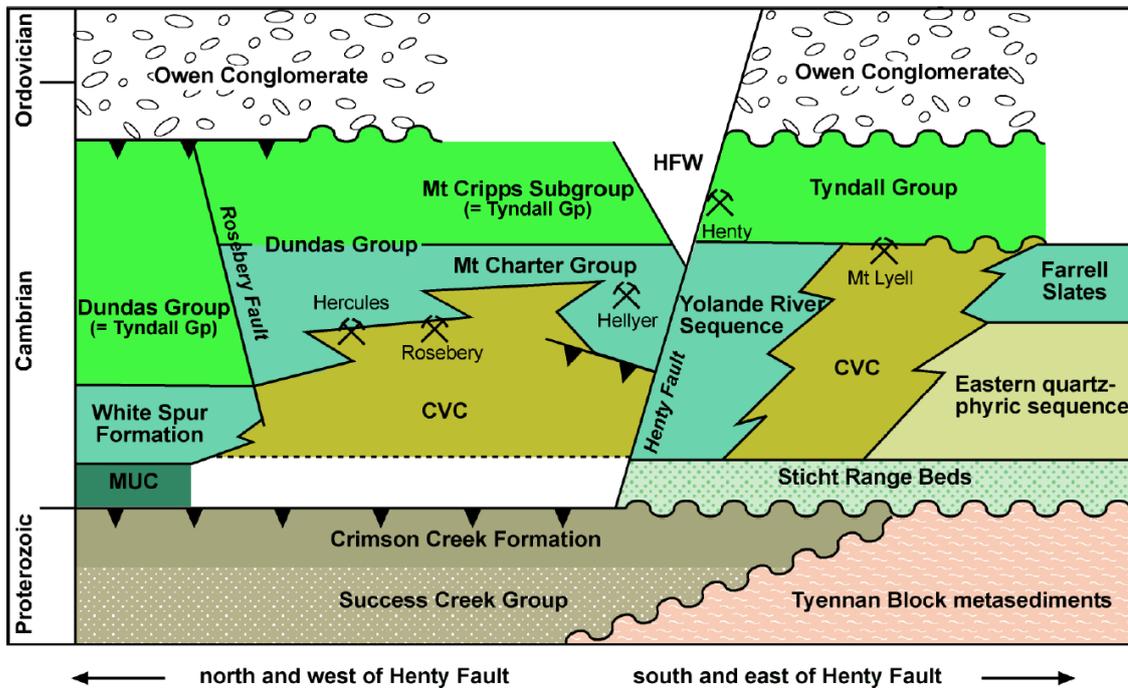


Figure 2.2: Lithostratigraphic units of the MRV (from Martin, 2004).

#### 2.3.1 The Sticht Range Beds

The predominantly west-dipping, early-Cambrian Sticht Range Beds comprise siliciclastic conglomerate, sandstone and minor siltstone. The beds overlie Neoproterozoic basement rocks at an angular unconformity, and are conformably overlain by the Eastern Quartz-Phyric Sequence and rocks of the Tyndall Group (Baillie, 1989). Corbett (1982) suggested that the succession was deposited prior to the MRV volcanism or during an extended pause in volcanism; however Baillie (1989) identified a quartz-feldspar porphyry within the lower sandstone unit. Baillie (1989) suggested that the lower conglomerate and sandstone were deposited as a series of fans, possibly close to a fault scarp, and that finer-grained upper sequences with bi-polar current patterns represent a transition from fluvial to shallow marine conditions.

### 2.3.2 The Eastern Quartz Phyrlic Sequence

The Eastern Quartz-Phyrlic Sequence (EQPS) of quartz-feldspar-phyric coherent rocks includes rhyolite, dacite and minor andesite and volcanoclastic facies. The EQPS overlies the Sticht Range Beds between Lake Dora and Mount Farrell (Figure 2.3) and are themselves overlain by the west-dipping Farrell Slates (Mount Charter Group). The sequence has an interfingering relationship with the CVC (Corbett, 1992).

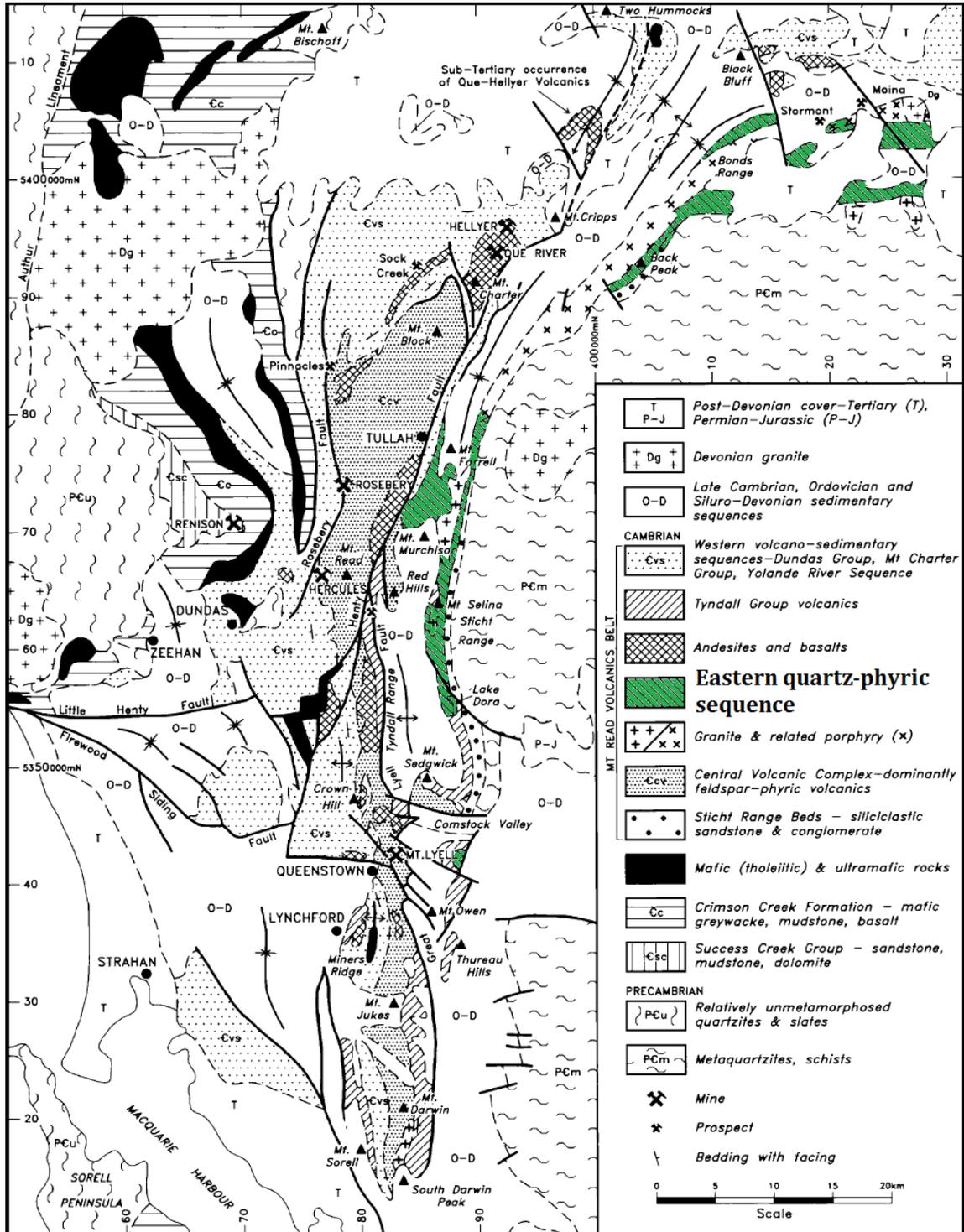


Figure 2.3: Distribution of the Eastern Quartz Phyrlic Sequence (modified from Corbett, 1992).

### 2.3.3 The Central Volcanic Complex

Most of the known sulphide mineralization in the MRV occurs within a central belt of volcanic facies, 9 to 12 km wide and roughly 90 km long, that is called the Central Volcanic Complex (CVC) (Corbett, 1981). The CVC is dominated by feldspar-phyric rhyolitic-dacitic-andesitic lavas and very thick, pumiceous volcanoclastic units (Corbett, 1981; McPhie and Gemmell, 1992). The complex includes the Rosebery and Hercules VHMS deposits.

The Southern CVC is separated from the Northern CVC by the Henty Fault. Gifkins and Allen (2001) subdivided the Northern CVC into three units, younging to the west (Figure 2.4); the Sterling Valley Volcanics (andesitic-basaltic) at the base, the Mount Black Volcanics (dacitic-rhyolitic), and the Kershaw Pumice Formation conformably at the top.

The Mount Black Volcanics (MBV) and Kershaw Pumice Formation formed through effusive and explosive eruptions respectively. These units were produced by eruptions in a large submarine felsic volcanic complex consistent with an intracaldera setting (Gifkins and Allen, 2002). The MBV cover the area from Mount Read to Hellyer west of the Henty Fault and are thick (>1.5 km) and laterally extensive, occupying large open folds. The predominantly feldspar-phyric sequence comprises interbedded pumice breccia, rhyolite breccia and rhyolitic lava intruded by hornblende  $\pm$  feldspar-phyric sills (Gifkins and Allen, 2002).

The Kershaw Pumice Formation is predominantly non-welded pumice breccia, pumice-rich sandstone and shard-rich siltstone with brecciated rhyolitic and dacitic lavas and intrusions. Hydrothermal alteration and mineralisation occurred during the late-stage intrusions. The Kershaw Pumice Formation forms part of the Rosebery-Hercules host sequence (Gifkins and Allen, 2002).

### 2.3.4 The Tyndall Group

The Middle Cambrian Tyndall Group overlies the CVC and the EQPS. The distinctive pink and green banded lower part of the group, known as the Comstock Formation (White, 1996), comprises crystal-rich mass-flow sandstone and breccia, minor limestone, mudstone, ignimbrite and rhyolitic intrusions. The upper part of the group, the Zig-Zag Hill Formation, consists of volcanoclastic conglomerate and sandstone which underlies the siliciclastic Owen Conglomerate at Mt Lyell and along much of the eastern side of the MRV. The group probably formed in a shallow marine environment close to subaerial volcanoes (White and McPhie, 1997). Correlates of the Tyndall Group overlie the Western Volcano Sedimentary Sequence (Yolande River Sequence) in some areas (Corbett, 1992).

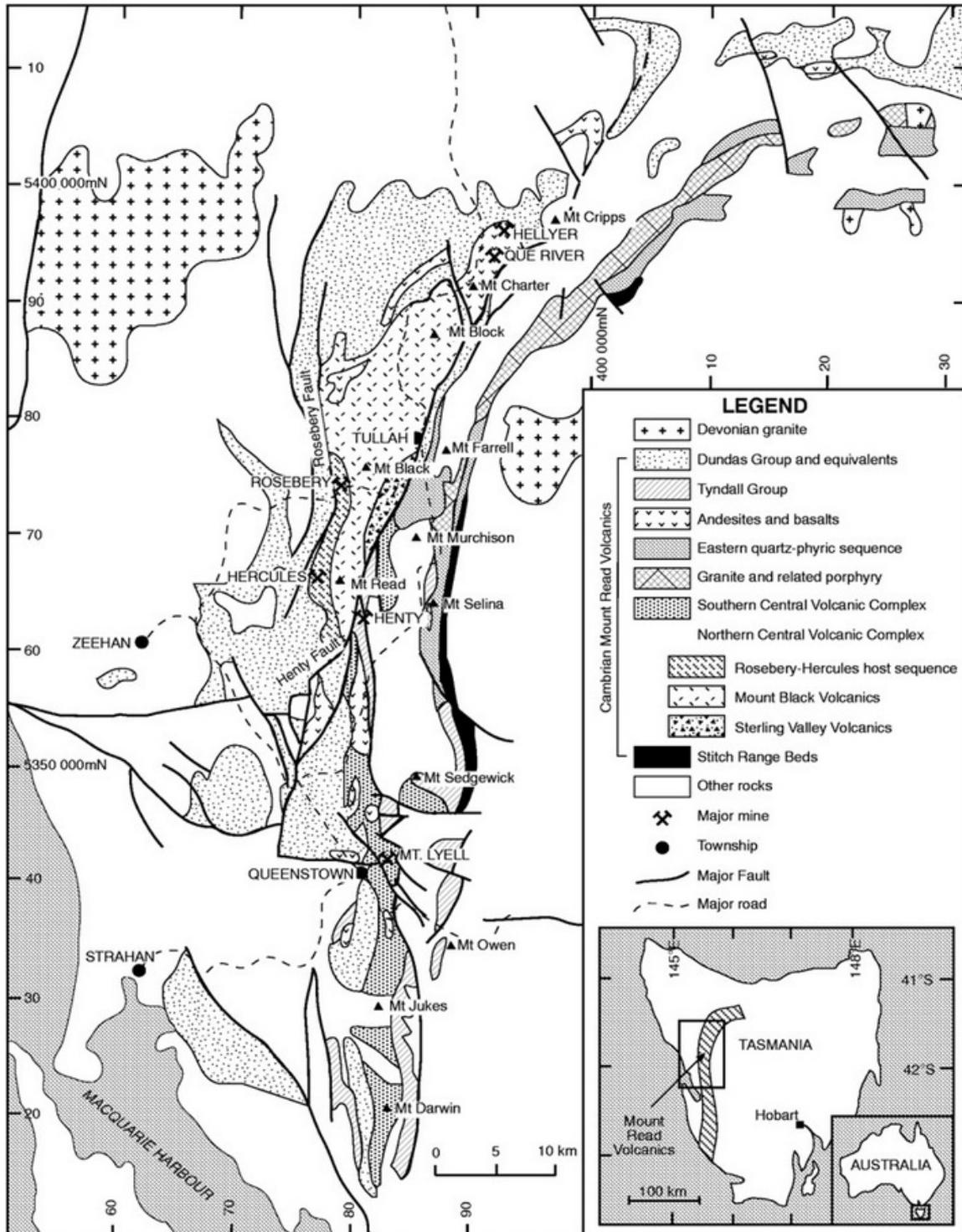


Figure 2.4: Distribution of the Sterling Valley Volcanics, Mount Black Volcanics, and the Rosebery-Hercules host sequence. The Kershaw Pumice Formation forms part of the Rosebery-Hercules host sequence (from Gifkins and Allen, 2002).

### 2.3.5 The Western Volcano Sedimentary Sequence

The Western Volcano-Sedimentary Sequence (WVSS) interfingers with and occurs along the western and northern flanks of the CVC. This together with lateral variations and fault disruptions make stratigraphic interpretation difficult (Corbett, 1992). The sequence consists of mass-flow deposits, turbiditic sandstone of mixed provenance, shard rich

mudstone, micaceous siltstone and black graphitic shale. Middle to Late Cambrian marine fossils are present in some units. Andesitic lavas and breccias, minor felsic lavas and large porphyry intrusions occur locally (Corbett, 1992).

Groups within the WVSS are defined geographically: rocks south of the South Henty Fault are called the Yolande River Sequence; north of the North Henty Fault is the Dundas Group; the sequence in the Hellyer area is the Mount Charter Group; and between the North and South Henty faults is the informally named Henty fault wedge sequence (Corbett, 1992).

#### *Yolande River Sequence*

The majority of the Yolande River Sequence dips and faces west; however the eastern part dips and faces east. The Miners Ridge Sandstone, occupying the core of the anticline at Miners Ridge, consists of siliciclastic sandstone underlain by tholeiitic Miners Ridge Basalt. The Lynch Creek Basalts occur at the top of the sequence at Lynchford and are dominated by andesitic-basaltic lavas, breccias and associated intrusions, and thick volcanoclastic units, turbidites and mudstones (Corbett, 1992).

#### *The Dundas Group*

The sedimentary sequence occupying the Dundas Trough west of Rosebery and flanking the MRV to the west and north of the North Henty Fault is known as the Dundas Group (Figure 2.5), the base of which is poorly defined.

The group is a complex mixture of immature and pebble-conglomerates, turbidites, shallow water epiclastic and volcanic sedimentary facies, mass-flow deposits, sandstone, mudstone, siltstone, shale, quartzwacke, and greywacke. Sequences show rapid facies changes and considerable structural disruption (Corbett, 2002). The pebble conglomerate contains volcanic fragments where it borders the MRV, indicating syn-volcanic deposition. At the top of the sequence are felsic to intermediate volcanoclastic sedimentary facies and minor lavas and intrusions. The volcanic components within the group are predominantly felsic (Burrett and Martin, 1989; Corbett, 1992; McNeill, 2000).

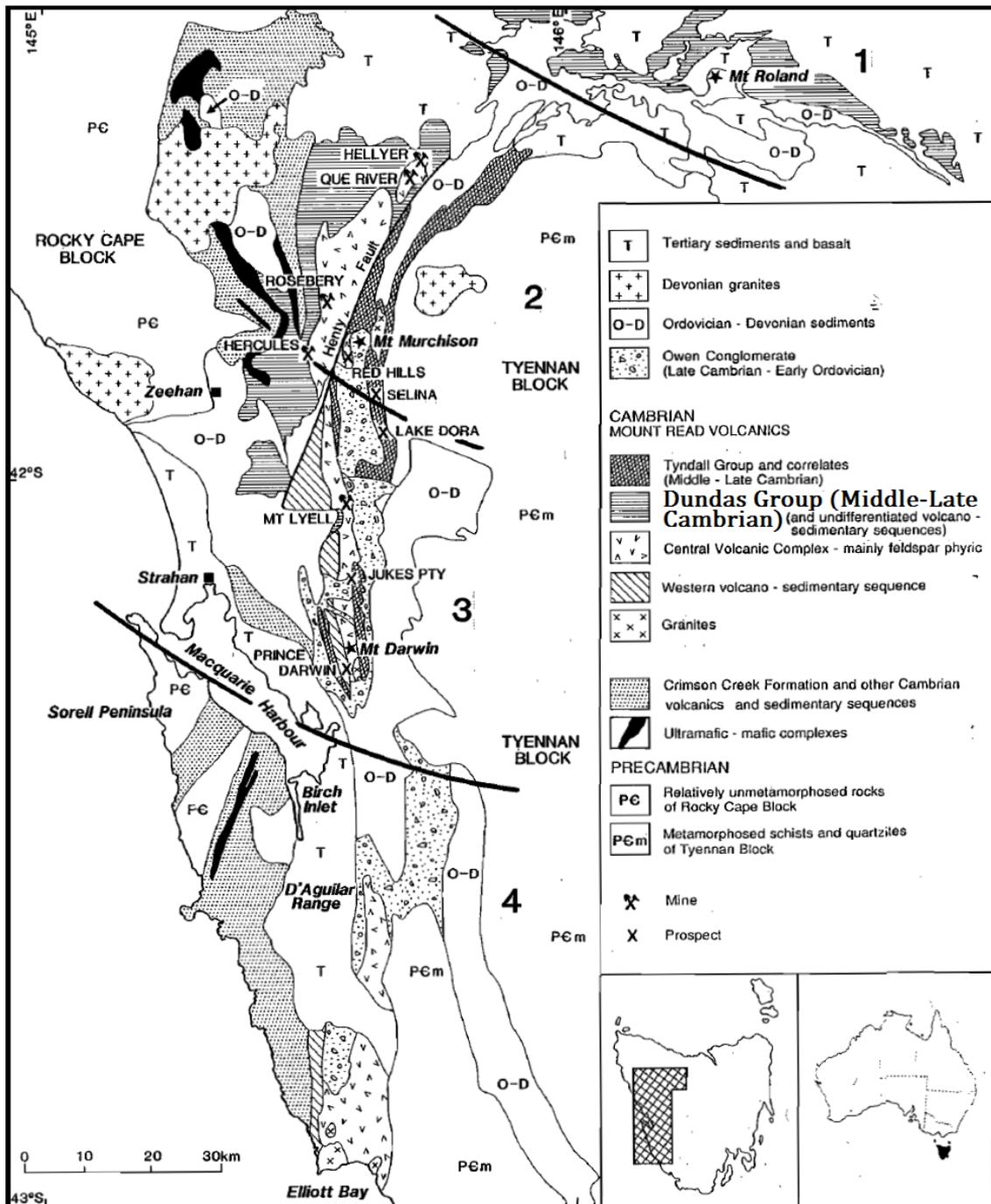


Figure 2.5: Distribution of the Dundas Group rocks (modified from Berry, 1992).

Fossil localities are perhaps the best available control on stratigraphy and indicate an age range from Middle Cambrian to Late Cambrian. The upper Dundas Group is considered to be an offshore equivalent of the Owen Conglomerate, and correlates of the Tyndall Group have been identified within the sequence (Corbett, 1992).

The Dundas Group has been subject to deformation in both the Cambrian and Devonian. Devonian deformation resulted in reactivation of the Rosebery Fault and propagation of a series of N-S trending thrust faults and associated strike-slip faults.

Much uncertainty has surrounded the interpretation and correlation of the Dundas Group lying adjacent to the MRV at Rosebery (Corbett and Lees, 1989). Finucane (1932)

assigned the name 'Rosebery Series'. Solomon (1965) placed the Rosebery Series below the MRV. Loftus-Hills *et al.* (1967) modified Solomon's view and incorporated the Rosebery Series with the lower part of the MRV (Lees, 1987). Corbett and Lees (1987) eventually redefined the Dundas Group west of Rosebery to include the White Spur Formation.

*The Mount Charter Group*

The Mount Charter Group contains the host andesite-basalt sequence of the Hellyer and Que River VHMS deposits. The group in this area overlies the CVC and underlies the Owen Conglomerate. This definition is problematic as it includes a correlate of the Tyndall Group (Mount Cripps Group). Its relationship with the EQPS is also complicated by the influence of the Henty Fault. The seven component units of the Mount Charter Group are well described by Corbett (1992) and shown in Figure 2.6. A brief summary of Corbett's description is given below.

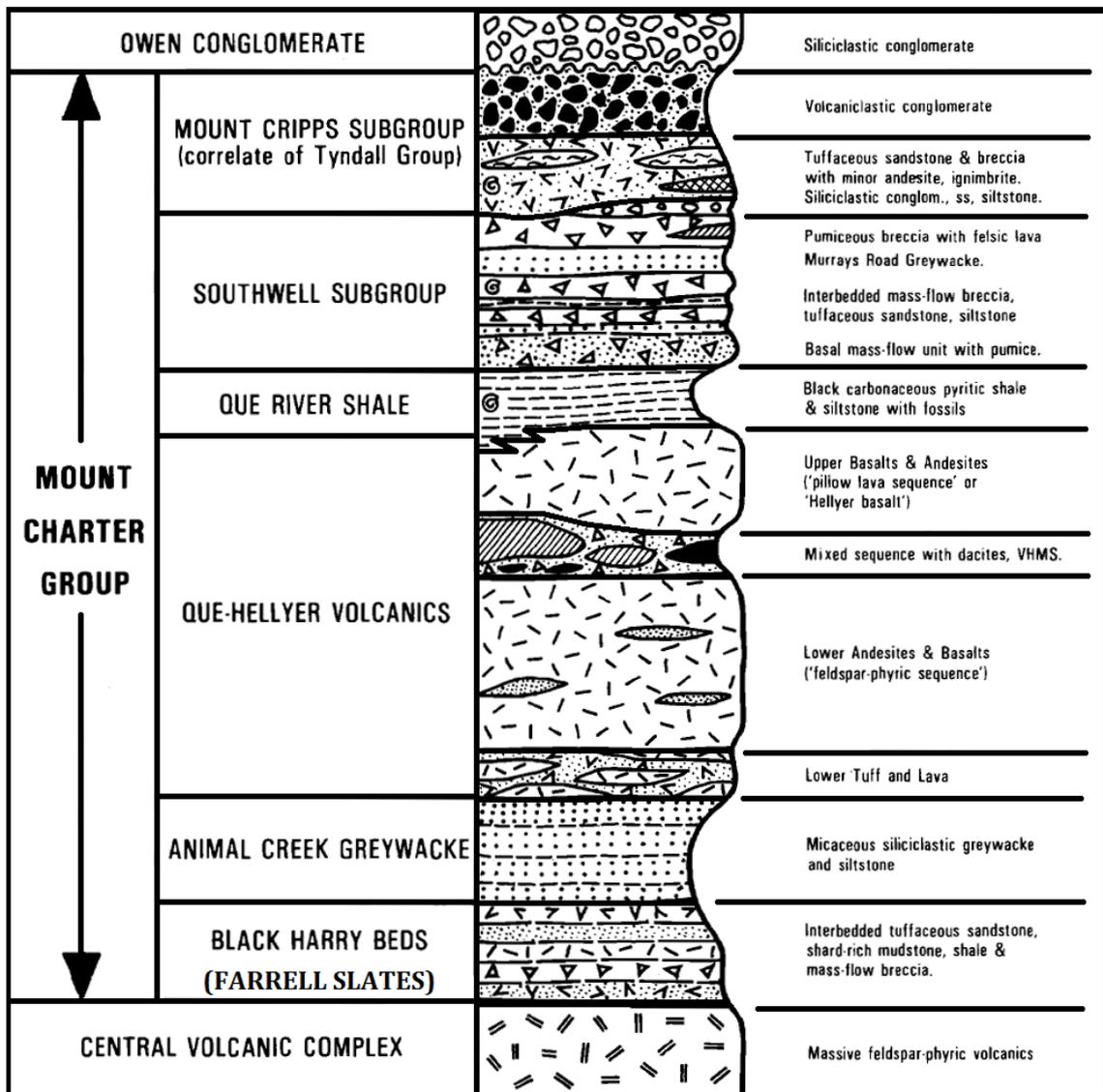


Figure 2.6: Components of the Mount Charter Group (modified from Corbett, 1992).

a). The Farrell Slates

A correlate of the lower Mount Charter Group, the Farrell Slates comprise black slate, volcanoclastic sandstone, mass-flow breccia and micaceous siliciclastic sandstone. Extending from the Henty Fault at Tullah to Mount Charter, the sequence overlies the EQPS to the northwest of Mount Murchison. The slates are sheared and steeply dipping to the west and appear to grade up into rocks of the Black Harry Beds and the Animal Creek Greywacke (Corbett, 1992).

b). The Black Harry Beds

The Black Harry Beds contain interbedded marine shard-rich mudstone, volcanoclastic sandstone, mass-flow breccia, and green to black shale. The sequence overlies the CVC and corresponds with the lower volcanoclastic part of the Animal Creek Greywacke (Corbett, 1992).

c). The Animal Creek Greywacke

The Animal Creek Greywacke is predominantly well-bedded grey micaceous siliciclastic sandstone interbedded with grey-black siltstone and shale. Volcanoclastic sandstone, shard-rich mudstone, and volcanoclastic breccia are minor components. Turbidite features are common (Corbett, 1992).

d). The Que-Hellyer Volcanics

The Que-Hellyer Volcanics comprise marine andesitic to basaltic lavas, breccias, volcanoclastic facies, minor felsic coherent facies and rare sedimentary units. Four informal subunits have been recognised and described by Corbett and Komyshan (1989). These include (1) the lower volcanoclastic facies and lavas; (2) the lower andesites and basalts; (3) the mixed sequence; and (4) the upper basalts and andesites.

The lower volcanoclastic facies and lava comprises interbedded felsic to andesitic volcanoclastic rocks, volcanoclastic sandstone and siltstone, and minor felsic to andesitic and basaltic lava (Corbett, 1992).

The lower andesites and basalts consist of breccias and lavas, with minor volcanoclastic and sedimentary interbeds and minor dacitic units (Corbett, 1992).

The mixed sequence, or mine sequence, comprise lavas and domelike structures of feldspar-porphyrific dacite intercalated with polymictic mass-flow breccias, thin andesite-basalt lavas, and minor shale and sandstone. The Que River and Hellyer massive sulphide lenses occur within the mixed sequence (Corbett, 1992).

The upper hangingwall basalts and andesites include pillow lavas and breccias of the Hellyer Basalt overlying the Hellyer orebody. The upper sequence is laterally variable; some regions consist almost entirely of andesite (Corbett, 1992).

e). The Que River Shale

The Middle Cambrian Que River Shale comprises distinctive black, carbonaceous, pyritic shale and siltstone. The shale contains trilobite fossils indicative of its age, and represents

a discontinuity in volcanism. It occurs between the underlying andesitic-basaltic facies of the Que-Hellyer Volcanics and the overlying felsic volcanoclastic facies of the Southwell Subgroup (Corbett, 1992).

f). The Southwell Subgroup

The Southwell Subgroup consists of quartz-feldspar-porphyrific pumiceous mass-flow breccia and sandstone interbedded with greywacke sandstone, siltstone, shale and minor felsic lava. The group was likely deposited in a marine basin close to active subaerial volcanism (Corbett, 1992).

g). The Mount Cripps Subgroup

The Mount Cripps Subgroup, a correlate of the Tyndall Group, has been divided into three informal units. The basal unit consists of lensoidal siliciclastic conglomerate, sandstone and siltstone, with boulders of Precambrian quartzite. Overlying this are intercalated crystal-rich volcanoclastic sandstone and fossiliferous siltstone (Middle Cambrian trilobite fauna), minor welded pink ignimbrite and minor andesitic lavas. The upper unit comprises purple-weathering volcanolithic conglomerate with porphyritic boulders and sandstone (Corbett, 1992).

### 2.3.6 The White Spur Formation

The thick (>1500 m) White Spur Formation (WSF) is present west of Rosebery in a wedge bounded to the east by the Rosebery Fault and to the west by the Stitt Quartzite (Lees, 1987). Previously, the WSF has been considered part of the 'Primrose Pyroclastics' of Brathwaite (1969). Corbett (1987) indicated a thickness of approximately 3500 m in the Howards Road area. Corbett and Lees (1987) assert that the upper part of the WSF is probably equivalent to the 'Judith Formation', a poorly defined basal unit of the Dundas Group. Green (1983), Lees (1987) and Parfrey (1993) identified lithological similarities between the WSF and the Natone Volcanics.

Quartzite and muscovite dominate the sequence; however felsic volcanic facies of MRV type and chert of intra-basinal origin also occur. Abundant quartz-feldspar crystals and porphyry clasts indicate that the source volcanoes were probably also quartz-feldspar-porphyrific (Corbett and Lees, 1987).

Lees (1987) divided the WSF into three sub units: (1) a coarse, lithic-rich, subrounded to angular matrix supported breccia containing shale, slate, felsic tuff, chert and pumice clasts; (2) a polycyclic central unit of quartz-feldspar tuff and bedded lithic wacke, volcanogenic sandstone and shale; and (3) an upper unit of laminated shale containing thin bands of massive pyrite, siltstone and minor sandstone, regarded by Corbett and Lees (1987) as the Chamberlain Shale. Contact relationships between the units are gradational.

### 2.4 The Owen Conglomerate

The Owen Conglomerate is a thick (up to 2000 m) sequence of siliciclastic, shallow marine to fluvial conglomerate and sandstone overlying the MRV south and east of the Henty Fault (Noll and Hall, 2005). The sequence is derived from Tyennan Precambrian rocks,

and ranges in age from middle Late Cambrian to Early Ordovician. This sequence was predominantly deposited in fault-controlled basins (Corbett and Lees, 1987).

### 2.5 Structural and Tectonic Evolution of the Mount Read Volcanics

The tectonic history of the MRV is largely interpreted from surface geology, and while some consensus has been reached, models are by no means complete.

The Neoproterozoic Rocky Cape and Tyennan Regions overlay the Mesoproterozoic basement, and form the major blocks of Western Tasmania (Figure 2.7). These formations record multiple deformation events. Angular unconformities and flat-lying Cambrian sedimentary facies succeeding folded Proterozoic units indicate that folding occurred during the Precambrian. Pelitic schist sequences found within the Tyennan Region contain garnets which record a rapid pressure increase, indicative of deep burial and subsequent exhumation prior to eruption of the MRV (Chmielowski, 2009).

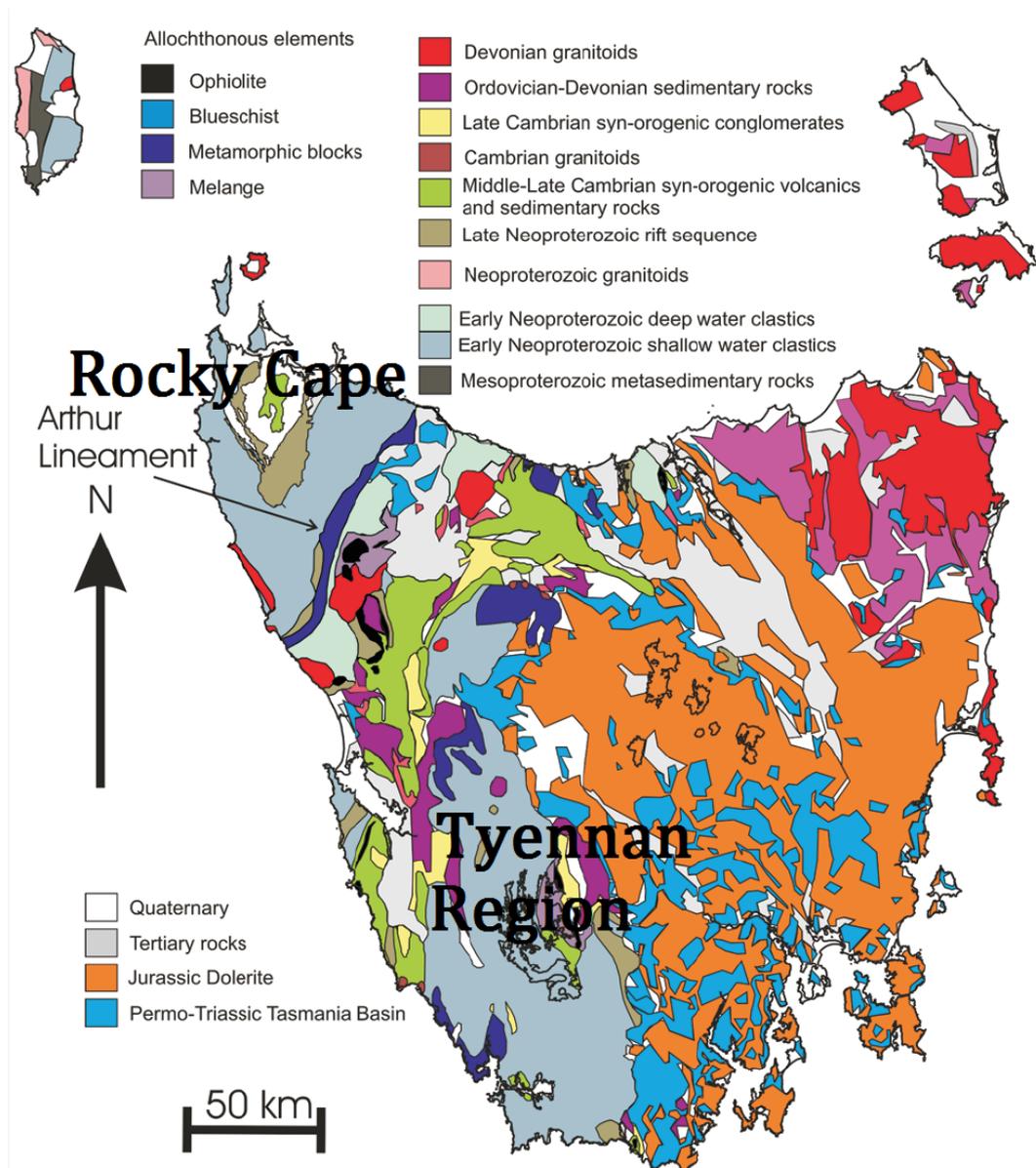


Figure 2.7: Showing Rocky Cape and Tyennan Region (modified from Berry and Brown, 2004).

Basement structures record evidence of a late Proterozoic rift phase. The position of Tasmania in Cambrian reconstructions of Rodinia remains controversial (Berry and Bull, 2012). The Western Tasmanian Terrane was being attenuated 600 – 570 Ma (Crawford and Berry, 1992). A magnetic high belt extends north of Tasmania across Bass Strait and into Victoria, wrapping around Broken Hill. Despite later structural reorganization this belt has geochemical, geometric and lithological similarities to Mesozoic Seaward Dipping Reflector Sequences (SDRS) along the North Atlantic and northwestern Australian volcanic passive margins. This suggests the formation of a volcanic passive margin in eastern Gondwanaland at the close of the Neoproterozoic (Crawford and Direen, 2003). Berry *et al.* (2007) concluded that the Western Tasmanian Terrane rifted from the East Antarctic margin at 580 Ma.

Rifting led to the formation of a thinned passive margin with small rift basins and associated tholeiites and MORB. Crawford and Berry (1992) suggested that eastward directed intra-oceanic subduction commenced to the east of the passive margin some time before the Middle Cambrian, forming an oceanic arc. Continued subduction between the arc and the passive margin resulted in an arc-continent collision at 520 Ma (Crawford and Berry, 1992). Major slices of fore-arc lithologies were thrust over the Proterozoic blocks. High temperature mylonites found in the Rocky Cape Region indicate that thrusting was predominantly toward the west (Berry and Stacey, 2004).

Most of the folding in the Neoproterozoic stratigraphy is correlated with the Late Cambrian Tyennan Orogeny, which began with ophiolite obduction 515-505 Ma. Major reverse faults and north trending open folds formed together with uplift of the Tyennan Block. Uplift is attributed to compressional stresses together with isostatic rebound due to crustal thickening following obduction. Compression and extension reactivated earlier extensional faults as reverse and thrust faults (Berry and Stacey, 2004). An intense compressional event recognised in the Delamerian and Ross Orogeny 500-495 Ma is correlated with a basin inversion in Tasmania (Berry *et al.*, 2007).

Subsequent erosion of the Tyennan Block and deposition by alluvial fans and braided streams into structural depressions formed the syn-orogenic Owen Conglomerate; lower sections record shallow marine MRV, higher sections record subaerial conglomerate. The identification of normal faults that host thickened accumulations of siliciclastic facies within the Owen Conglomerate support interpretation of syn-fill extension (Hall and Noll, 2005). The Great Lyell Fault extends along the western boundary of the Owen Conglomerate, and was active during this time (Banks and Baillie, 1989).

The Arthur Lineament (Figure 2.7) formed during the Tyennan Orogeny and marks the western limit to allochthonous facies, and thus the maximum extent of the collisional thrust complex (Berry and Stacey, 2004).

Post-collisional extension and magmatism produced the MRV. The Cambrian Dundas Trough foreland basin sequence of marine mudstones and conglomeratic turbidites formed during rifting, and contemporaneously with the MRV (Green, 1983). The Dundas Trough wraps around the western and northern margins of the Tyennan Region. The central region of the trough is occupied by the volcano-sedimentary sequences of the Dundas Group and equivalents (Green, 1983).

Sequences of the Dundas Trough interfinger with the MRV (Corbett, 1992). The volcanic facies occupy the eastern margin of the Dundas Trough and are divided by the Henty Fault Zone, which was active during the Cambrian and appears to have controlled the distribution of lithologies, such as the EQPS. Additionally, economic mineralisation to the north and west of the Henty Fault is Pb-Zn rich, whereas it is Cu-Au rich to the south and east (Berry, 1989; Berry and Keele, 1997).

The entire Western Tasmanian Terrain was folded in at least one major deformation event. The Early to Middle Devonian Tabberabberan Orogeny, dated in Tasmania at 390 - 420 Ma (Adams *et al.*, 1985), involved widespread thrusting, the propagation of N-S trending faults in the MRV and reactivation of earlier faults. The activation of the Rosebery Fault caused the MRV to be thrust over the Dundas Group. Subsequent emplacement of Devonian granitoids including the Heemskirk Granite imposed contact metamorphism, remobilization of elements and regional scale hydrothermal alteration. The Linda Zone, a major Devonian fault complex, intersects the MRV at Mt Lyell.

## 2.6 The geology of the Rosebery Group

The geology of the Rosebery area is strongly controlled by the Rosebery and Mt Black faults, which dip approximately 45° to the east and show reverse movement related to Devonian deformation (Green, 1983). The age and relationships of the volcano-sedimentary sequences west of Rosebery have been debated for many years (Burrett and Martin, 1989). Corbett and Lees' (1987) recognition of the east dipping Rosebery Fault helped to clarify the relationship between the MRV and sedimentary sequences to the west in this area, and mapping has confirmed that the sequence belongs to the Dundas Group (Corbett and McNeill, 1986). However, the historically debated nomenclature of the MRV, coupled with the complexity of the Dundas area continues to result in confusion.

The sedimentary rocks of the Rosebery Group dip steeply to the west (70°?), which complicates structural interpretation (Brathwaite, 1969). The rocks have undergone at least one period of deformation and cleavage development. This has traditionally been associated with the Devonian "Tabberabberan" Orogeny; however Green (1983) postulated that deformation style is different from that recognized to be Devonian. Tectonic disruption may in fact be associated with the Delamerian Orogeny 500-495 Ma (Green, 1983).

The volcano-sedimentary rocks of the Rosebery Group occupy the Dundas Trough west of Rosebery between the Marionoak Fault to the west and the Rosebery Fault to the east (McNeill and Simpson, 2000). A number of features indicate that the rocks were deposited well below wave base (70 to 100 m) (Green, 1983). The series include sandstone, mudstone, fuschitic conglomerate and altered volcanoclastic horizons. The sequence has been strongly disrupted, with many faulted boundaries, tectonic interleaving of lithologies and extensive overturning of bedding (Corbett and McNeill, 1986; Corbett and Lees, 1987). Fault bounded strips of differing lithologies, including gabbro near Westcott Hill, may be the result of the Devonian deformation; however, the Delamerian Orogeny is also likely. Carbonate-rich polymict conglomerates and interleaving lensoidal packages suggest erosion of local intra-basinal sources (Corbett, 2002).

Corbett (2002) considered the Rosebery Group to be largely an offshore equivalent to the Late Cambrian Owen Group. Possibly the oldest exposed unit of the group, the upper WSF (Chamberlain Shale) is bounded by the Rosebery Fault to the east and persists eastwards for at least 1.5 km beneath the fault (Burrett and Martin, 1989). This sequence of interbedded volcanoclastic sandstone, siltstone and mudstone is a probable correlate of the Middle Cambrian Tyndall Group (Corbett, 2002). The WSF is conformably overlain by the Stitt Quartzite, which shows considerable tectonic disruption and overturning (Burrett and Martin, 1989).

The Natone Volcanics appear as a local lens of weakly chlorite-sericite altered, volcanic lithic, fiamme-bearing, quartz-feldspar-rich mudstone separating the Stitt Quartzite from the Salisbury Conglomerate. The Salisbury Conglomerate reaches maximum development at Moores Pimple, where it contains abundant clasts of dolomite, felsic to mafic igneous rocks, quartzite, sandstone and fuchsite. A repetition of facies appears to occur prior to termination of the Rosebery Group by the Marionoak Fault (Burrett and Martin, 1989).

Green (1983) subdivided the group into three sequences: the 'Eastern units' comprising the Chamberlain Shale, Stitt Quartzite and Westcott Argillite; the 'Central units' including the Mudstone-dominated sequence, Salisbury Conglomerate and Natone Volcanics; and thirdly the 'Munro Creek Formation'.

The Eastern units are relatively well defined both structurally and stratigraphically. Faulting, deformation and hydrothermal alteration to lower-greenschist facies has resulted in some confusion regarding structural interpretation of the Central units.

Corbett (2004) redesignated most of the Rosebery Group to the Marine Owen Group, from fossil evidence and lithological correlations, including the Stitt Quartzite, Munro Creek Slate and Quartzite, Westcott Argillite and the Natone Volcanics. Some rocks of pre-Owen Group age such as the Tyndall Group or Western Volcano Sedimentary Sequence (WVSS) may also be present (Corbett, 2004).

### **2.6.1 Chamberlain Shale**

To the west of the MRV in the Rosebery area is the Chamberlain Shale (Braithwaite, 1972). The sequence overlies the WSF (McNeill, 2000). Lees (1987) identified a faulted contact with the MRV (Rosebery Fault). Green (1983) believed the west dipping rocks to be the right way up; however, few reliable facing indicators have been found. The sequence comprises tightly folded grey slate, mudstone and volcanoclastic quartz-feldspar sandstone. Green (1983) suggested the sequence may be a volcanic mass-flow deposit as it grades into the conformably overlying Stitt Quartzite (Parfrey, 1993).

### **2.6.2 Stitt Quartzite**

A prominent marker unit, the Stitt Quartzite comprises 350 m of quartzwacke, quartzarenite, siltstone, dark grey slate and conglomerate. Interbeds of sandstone, mudstone and laminated siltstone shale occur. Two distinct facies are present: thickly bedded and graded sandstone (4 to 100 m) and thinly bedded to laminated siltstone (1 to 20 cm) and slate. Graded beds, cross bedding, load structures and ball together with

pillow structures provide facing indicators and demonstrate that the west-dipping sequence youngs to the west (Green, 1983).

The discovery of fossils at Higgins Creek (Jell *et al.*, 1991) indicates that the Stitt Quartzite formed contemporaneously with the Late-Cambrian Owen Conglomerate. The group forms a narrow but continuous belt from the Hatfield River to the Dundas–Professor Range area, where it links to the conglomerate and shallow marine sandstone sequence at Mt Zeehan (Corbett, 2002).

### **2.6.3 Westcott Argillite**

In the Pieman River Gorge area, Green (1983) observed a gradational contact between the dark grey Stitt Quartzite and the pale green-blue Westcott Argillite. Beds gradually reduce in thickness and the argillite is finer grained. The sequence is at least 200 m thick and generally faces west, except for the presence of a major syncline in the Pieman River Gorge (Green, 1983).

Green (1983) described the Westcott Argillite as minor sandstone beds and laminae intercalated with massive mudstone laminae. A conglomerate horizon in the Pieman River Gorge area contains clasts up to 10 cm in diameter of chert, dolomite, quartz, quartzite and rare mafic volcanic facies. Sedimentary structures are common and include graded beds with flame structures at the base (Green, 1983).

Parfrey (1993) described the sequence as mixed dolomitic siltstone, sandstone and purple slate outcropping west of the Salisbury Conglomerate in the Pieman River Gorge and on Westcott Hill.

### **2.6.4 Central Section**

Green (1983) defined a Central Section of the Rosebery Group overlying the Westcott Argillite. The section comprises a Mudstone Dominated Sequence, the Salisbury Conglomerate and the Natone Volcanics. Green postulated that the Central Section was overlain by the Munro Creek Formation, which he noted commonly resembles the Stitt Quartzite. Lees (1987) later suggested that this was in fact a southern extension of the Stitt Quartzite however this is yet to be confirmed.

### **2.6.5 Mudstone Dominated Sequence**

The Central Section in the Pieman River Gorge area comprises siltstone, mudstone, dolomite, slate, lithicwacke and the fuschite-bearing Salisbury Conglomerate. The sequence is disrupted, and graded sandstone beds indicate an eastward facing. Interbedded dolomicrite and slate is succeeded by mauve-coloured graded dolomitic sandstone and mudstones with some conglomeritic beds. At the top of this section, a 6 m thick section of the Salisbury Conglomerate is truncated by a fault (Green, 1983).

Green (1983) considered the rocks to be a correlate of the Westcott Argillite, noting compositional and textural similarities. Brathwaite (1969) included the sequence in the Westcott Argillite. The sequence is most likely representative of a higher stratigraphic level of the Westcott Argillite in the eastern area of the Rosebery Group (Green, 1983).

### 2.6.6 Salisbury Conglomerate

At Rosebery, the Salisbury Conglomerate interfingers with the Natone Volcanics and is conformably overlain by the Westcott Argillite (Parfrey, 1993). Green (1983) observed a western fault contact with the Mudstone Dominated Sequence at Pieman Gorge and well exposed outcrops present in Natone Creek. Graded bedding occurs in sandy beds and sections found within the basal part of the Natone Volcanics indicate that the unit is overturned (Green, 1983).

The Salisbury Conglomerate section in Tasmanian Department of Mines drill hole, DDH MD-1 is about 90 m thick. Green (1983) and myself observed six fuchsitic conglomerate beds interfingering with a sandstone-mudstone sequence below the base of the 90 m section in DDH BH-1, which may also be included. Green (1983) observed a similar underlying sequence below the Salisbury Conglomerate in the Pieman Gorge section.

Green (1983) described a section of the conglomerate in the Pieman Gorge as closed framework conglomerate with clasts up to 30 cm well aligned parallel to cleavage. He observed clasts of rounded to subangular chert, quartzwacke, carbonate, siltstone, shale, phyllite, fuchsite and vein quartz in a carbonate bearing lithicwacke matrix. Green noted that the conglomerate in the drill hole is coarser grained than the section in the Pieman Gorge and contains felsic volcanic clasts.

Lees (1987) described the conglomerate as matrix to clast supported polymict conglomerate with rounded to sub angular clasts of chert, micaceous quartzite, siltstone, black shale, phyllite, vein quartz, fuchsite and mafic volcanics set in a dolomitic to cherty matrix. Campana and King (1963) noted sporadic grains of pyrite, chalcopyrite, galena and molybdenite (Parfrey, 1993).

### 2.6.7 Natone Volcanics

Numerous authors have written about the Natone Volcanics yet their significance remains uncertain (Corbett, 2002). Finucane (1932) and Taylor (1954) described the volcanics as a band of quartz porphyries separating the fuchsitic conglomerate (Salisbury Conglomerate) and the Lower Slates and Quartzites (Stitt Quartzite) west of Rosebery. Finucane suggested that the volcanic facies were intrusive, possibly a dyke; however, he noted numerous rounded and angular cherty inclusions which he believed were derived from the conglomerate (Finucane, 1932).

Campana and King (1963) named the Natone Volcanics after identifying the rocks in Natone Creek. Observing similarities with the sequence east of Primrose, the Stitt Quartzite, they considered the rocks to be sedimentary, not intrusive as Finucane (1932) and Taylor (1954) suggested. Dallwitz (1946) was the first to suppose that the Moores Pimple breccia-conglomerate may be the same as the one in Natone Creek (Salisbury Conglomerate). His inference was later substantiated by Campana and King's mapping work (1963). Loftus-Hills (1967) confirmed that in the Rosebery Cemetery area, the Stitt Quartzite is structurally overlain by the Natone Volcanics and the fuchsitic conglomerate.

Brathwaite (1969) observed that the Natone Volcanics pinch out to the north and converge southwards with the 'Primrose Pyroclastics' (Chamberlain Shale, WSF). He also

noted near vertical cleavage and that the mineral assemblages (quartz-sericite-muscovite  $\pm$  chlorite) would be stable at lower greenschist facies.

In 1976 a drill hole proposed by Geoffrey Green was drilled below the Natone Volcanics (DDH MD-1). The primary purpose of the hole was to obtain fresh samples for age dating; however strong alteration, cleavage development and the presence of black shale clasts rendered the unit unsuitable for Rb-Sr and U-Pb dating. The second purpose was to clarify facing relationship, which was successfully achieved. Graded bedding demonstrated that this part of the Rosebery Group youngs east (Green, 1983).

Corbett (1981) mapped a correlate of the Natone Volcanics at Moores Pimple, describing the unit as quartz-feldspar-phyric tuff and autobrecciated lava. He observed the volcanics overlying a carbonate-fuchsite-chert conglomerate (Salisbury Conglomerate correlate).

Green (1983) mapped a 120 m wide band of the Natone Volcanics trending north-south from the Pieman River Gorge to Westcott Hill. Green (1983) describes the Natone Volcanics as a strongly cleaved felsic-vitric-crystal-tuff with zones of quartz-sericite; quartz-chlorite and carbonate-rich-schist. Noticing similarities between the massive pyroclastics at Rosebery and the Natone Volcanics, Green (1983) suggested a mass-flow emplacement and that the unit might be classed as epiclastic rather than of pyroclastic origin; however, he warned that poor textural preservation prevents definitive judgement. Green (1983) also drew attention to the presence of relict cusped glass shards from one specimen, but noted that sericitization and strong cleavage development had generally obliterated much of the rocks original texture. Attempting to address the problem of contact relationships, Green (1983) inferred a fault contact between the Natone Volcanics and sedimentary units to the east, a projected southward continuation of the fault truncating the Salisbury Conglomerate in the Pieman River Gorge.

Lees (1987) identified horizons of the Natone Volcanics containing small lithics of shale, siltstone, chert, pumice, quartz and sericitized feldspar in a strongly sericitic matrix. He identified the rocks in outcrop in the Natone Creek (Pieman River bend area) and on the Williamsford road to near Jupiter, where they are truncated by the Rosebery Fault. Contact relationships between the Stitt Quartzite, Natone Volcanics and Salisbury Conglomerate have not been resolved (Parfrey, 1993).

### **2.6.8 Munro Creek Formation**

Green (1983) described the Munro Creek Formation as at least 250 m of sandstone, dark grey slate, a conglomerate horizon and minor fine-grained dolomite beds. Sandstone beds are graded and interbedded with dark grey and pyritic slate. The formation is bounded by faults at both its eastern and western margins, and shows lithological similarities to the Stitt Quartzite. Green (1983) mapped the formation from Williamsford in the South to the Pieman River Gorge in the North.

## **2.7 Summary**

The Middle to Late Cambrian MRV of western Tasmania were emplaced over a period of at least 12.6 m.y., from 506.8 to 496.0 Ma, and the majority of VHMS deposits formed at  $\sim 500 \pm 1$  Ma (Mortensen *et al.*, unpublished). The volcanic belt formed in a

predominantly subaqueous setting along the western margin of the Neoproterozoic Tyennan Region. Corbett (1992) divided the MRV into five major lithostratigraphic units: (1) the Sticht Range Beds; (2) the Eastern Quartz-phyric Sequence (EQPS); (3) the Central Volcanic Complex (CVC); (4) the Western Volcano-Sedimentary Sequence (WVSS); and (5) the Tyndall Group.

The study area lies within the WVSS of the northern MRV. The Natone Volcanics form part of the Dundas Group west of the Rosebery Fault, herein termed the Rosebery Group of Green, Solomon and Walshe (1981). The geology within this area has historically been a subject of confusion, limited by complex structural relationships (Green, 1983; Parfrey, 1993). The internal structure, contacts with enclosing units, provenance and potential correlations of the Natone Volcanics remain poorly understood.

### 3.0 Facies descriptions for the Natone Volcanics and enclosing sequences

#### 3.1 Introduction

This chapter provides detailed descriptions of volcanic and sedimentary facies that constitute the Natone Volcanics and the enclosing sequences, based on observations of hand samples, seventeen thin sections, and graphic logs of drill core from DDH MD-1 (Green, 1983) and BH-2. Although alteration and metamorphism have affected the succession, making identification of primary textures and mineralogy difficult in some cases, two volcanogenic-sedimentary facies associations, one volcanic facies association, one coherent facies, and one sedimentary facies have been identified; their characteristics are summarised in Table 3.1.

#### 3.2 Facies association A: Interbedded Fine-Grained Sandstone and Mudstone



Figure 3.1: Drill core of Interbedded Fine-Grained Sandstone and Mudstone (Facies association A) (DDH MD-1, 61.5 m).

Facies association A appears over six thinly (<10 cm) interbedded sequences in DDH MD-1. These sequences range in thickness from 2.5 to 20 m. The sequence shows up-hole grading at 17 m (west facing), and a possible fault or fold at 37 m is followed by down-hole facings at 50 – 53 m and deeper in the hole. Facies association A occurs immediately beneath and interfingers with Facies association B. Facies association A rocks are cleaved, weakly chlorite altered, and thinly interbedded. They consist of basement-dominated non-volcanic and volcanic fine-grained sand, and silt particles (Figure 3.2). Two red-brown chromium spinels (chromite) were identified; their chemistry will be described in chapter 4. Bedding is distinct to diffuse planar, with rare grading and cross bedding. The mudstone is cream coloured, and the sandstone is light to dark grey.

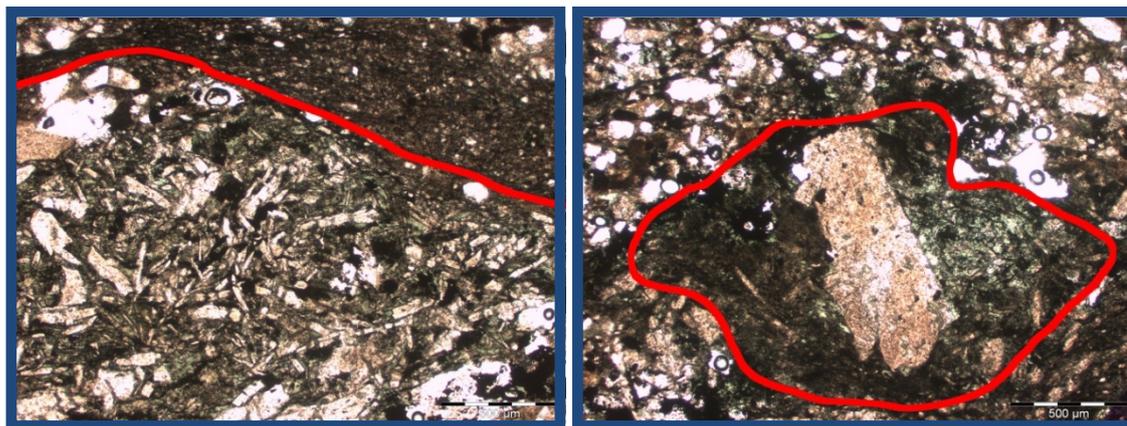


Figure 3.2: Photomicrographs of crystal dominated sandstone (Facies association A) showing grains of quartz, mica, and a lithic fragment below the red line (left); and a dacitic-andesitic? fragment (right) outlined red (DDH MD-1, 33.0 m PPL).

### 3.3 Facies association B: Fuchsitic Conglomerate and Sandstone



Figure 3.3: Drill core of Fuchsitic Conglomerate and Sandstone (Facies association B) (DDH MD-1, 65.4 m).

Facies association B interfingers with, and overlies Facies association A. Beds range in thickness from 1 to 98 m in DDH MD-1. Facies association B is a sericite-carbonate altered, fuchsitic (low abundance, typically <5 %), sandy matrix supported, poorly sorted, subangular conglomerate (Figure 3.4). The interval from 61.7 - 73.2 m in MD-1 is more clast supported. Components range from 1 mm to 6 cm, with an average of 1 cm; however, larger fragments (<10 cm) also occur. Lithic components include quartz, quartzite, chert (grey, white, red), micaceous clasts, fuchsite-altered clasts (green), rhyolite, dacite, dolerite, mafic volcanics (blue, purple), quartz-rich sandstone (dark grey, yellow), black shale, dolomite, carbonate, sandstone and mudstone (green, grey) in a sandstone matrix. Three chromite were identified in Facies association B, their chemistry will be examined and described in chapter 4. Numerous (>18) intercalated quartz-rich fine to medium-

grained sandstone lenses occur within the conglomerate, some of which are graded indicating a down-hole (i.e., east) facing.



Figure 3.4: Thin section (7.5 cm long) of Facies association B showing weak fabric and lithic fragments in quartz-mica sandstone matrix (DDH MD-1 65.5 m).

#### 3.4 Facies association C: Volcanic-lithic Sandstone and Siltstone (Natone Volcanics)



Figure 3.5: Drill core of Volcanic-lithic Sandstone and Siltstone (Facies association C) (DDH BH-2 62.5 m).

Fine-grained volcanic-lithic sandstone and siltstone occur within and above Facies association B, appearing in two intervals in DDH MD-1, 3.8 m and 1 m thick respectively, and the interval from 182.5 – 290.16 m (EOH) in MD-1; and from 0 – 146 m in DDH BH-2. Facies association C is grey-white with a pronounced cleavage, weakly carbonate and patchy chlorite-sericite altered, quartz-feldspar-crystal-rich, massive, fiamme-bearing sandstone and siltstone with rare well-preserved relic pumice and mica-rich domains of altered pumice preserved in carbonate (Figure 3.6). Quartz (volcanic and metamorphic)

and feldspar are subangular to subrounded. Yellow sandstone fragments incorporated into altered carbonate domains, and sedimentary clasts up to 3 cm (predominantly black slate >1 mm) suggest that the unit be classed as volcanic-lithic sandstone. Other components include equant and deformed volcanic and sedimentary clasts, quartzite and chert. A fine to medium grained sub-unit at 101.1 – 104.87 m in MD-1 shows grading with a decreasing abundance of lithic clasts down-hole, suggesting an east facing. Facies association C is quartz-feldspar-crystal-rich and compositionally homogenous.

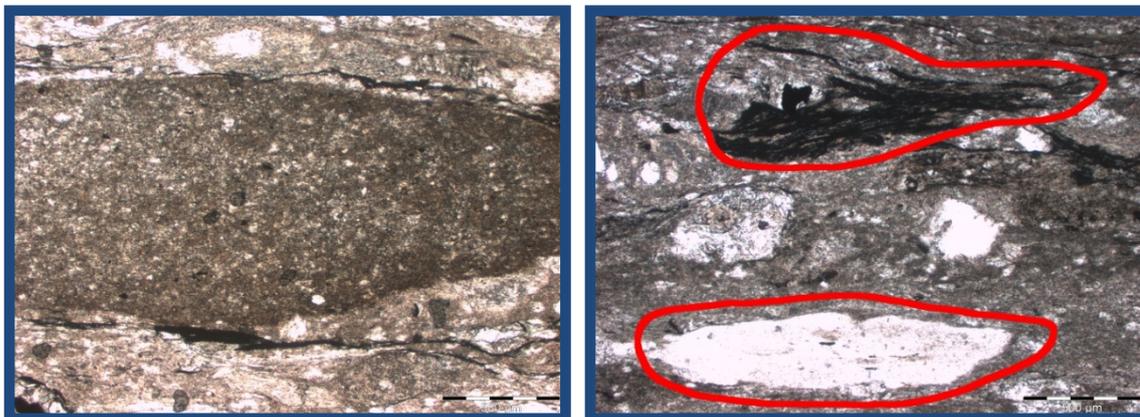


Figure 3.6: Photomicrographs of lithic fragments (left, centre; right, bottom) in carbonate altered, fiamme-bearing (right, top), volcanic-lithic sandstone and siltstone (DDH MD-1 164.35 m PPL).

### 3.5 Facies D: Coherent Dacite



Figure 3.7 Hand sample of Coherent Dacite (Facies D) taken from the Murchison Highway at 376780mE, 5372520mN (GDA94/Zone 55).

Facies D occurs immediately above Facies association C in DDH BH-2 for an interval of 2 m from 146 m to 148 m. The contact with Facies association C is obscured by gradational chlorite-carbonate alteration and the contact with Facies E is sharp. This coherent unit is feldspar>>quartz-phyric, massive, originally glassy with possible relic perlite domains, non-vesicular, microspherulitic dacite, with irregular and rare quartz microphenocrysts and a poorly developed snowflake (micropoikilitic) texture in the groundmass (Figure 3.8).

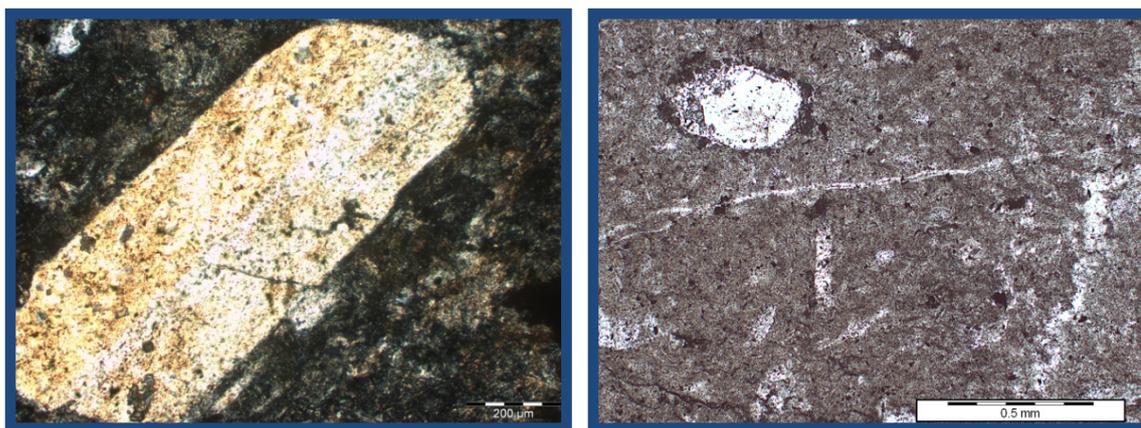


Figure 3.8: Photomicrographs of Facies D; feldspar phenocryst (left) in snowflake groundmass (DDH BH-2 147.0 m XPL); and rare quartz phenocryst (right) in feldspar-phyric dacite (DDH BH-2 147.0 m PPL).

### 3.6 Facies E: Quartz-Rich Sandstone



Figure 3.9: Hand sample of Quartz-Rich Sandstone (Facies E) taken from the Murchison Highway at 376790mE, 5372524mN (GDA94/Zone 55).

Facies E is down-hole of Facies D in DDH BH-2 from 148 m to 166.25 m (EOH). Facies E is massive, quartz-rich sandstone with minor feldspar and mica, a uniform distribution of lithic fragments, including sandstone and mudstone in a chlorite-sericite-carbonate altered groundmass (Figure 3.10). Facies E does not contain pumice or other volcanic components, quartz is predominantly metamorphic with undulose extinctions. Eleven chromite were identified, their chemistry is examined in chapter 4. Facies E is lithologically similar to the Stitt Quartzite described by Green, Solomon and Walshe (1981), Corbett and Lees (1987), and Parfrey (1993).

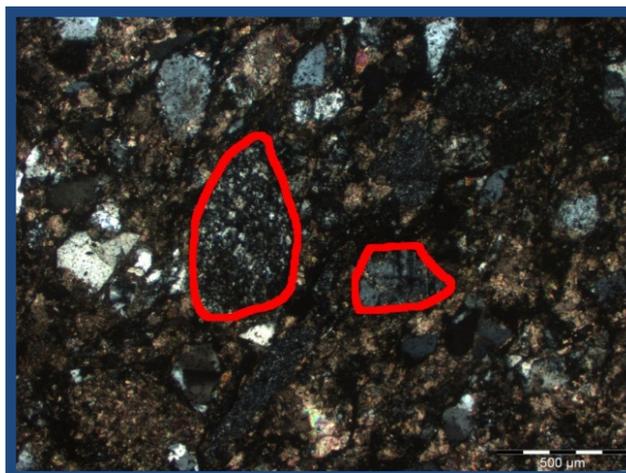


Figure 3.10: Photomicrograph of quartz-rich sandstone showing lithic fragment (left) and rare feldspar-crystal (right) in a mica-carbonate groundmass (BH-2 155.75 m XPL).

### 3.7 Discussion

Facies association A is of mixed sedimentary and igneous provenance. The presence of red-brown chromite implies that the source may have included exhumed and eroded ultramafic rocks, possibly derived from the Neoproterozoic basement. Facies association A interfingers with Facies association B, which conformably underlies Facies association C (Natone Volcanics).

Facies association B mostly consists of metamorphic detritus with a mixed population of quartz with straight to undulose extinction and very little plagioclase (<1%). Fuschite-altered clasts suggest the presence of mafic volcanic facies. Rapid changes in grainsize within conglomeritic intervals indicate amalgamation of beds. Graded sandy beds are lithologically similar to Facies association A; however, they contain a higher proportion of volcanic components (fiamme). Facies association B is the Salisbury Conglomerate described by Green, Solomon and Walshe (1981), Lees (1987) and Parfrey (1993).

Facies association C is strongly foliated and has undergone sericite alteration, making identification of primary textures and mineralogy difficult. Facies association C is a quartz-feldspar-phyric, volcanoclastic facies with a minor non-volcanic component lithologically similar to the White Spur Formation (WSF) described by Green, Solomon and Walshe (1981), and Corbett and Lees (1987). Green (1983), Lees (1987) and Parfrey (1993) compared the Natone Volcanics to the WSF; however, the volcanic-lithic Facies association C could equally be Eastern Quartz Phyric Sequence or Tyndall Group correlative, as these units are also volcanic-quartz-bearing.

Facies D displays a uniform distribution of phenocrysts and is not vesicular, nor associated with breccia. The lower contact is obscured by alteration and the upper contact with Facies E is sharp. The micropoikilitic texture in the groundmass is typical of the inner parts of lava domes or shallow intrusions (McPhie, pers. comm., 2012). Nothing definitive can be said about structural arrangement; however, the absence of breccia and the presence of snowflake texture in the groundmass may indicate that this is an intrusion.

Facies E is unrelated to Facies D, containing a very high proportion of quartz; the unit is lithologically similar to the Stitt Quartzite described by Green, Solomon and Walshe (1981).

### **3.8 Conclusion**

Facies association A is a sandy, volcanogenic-sedimentary facies association. Facies association B is a conglomeritic, volcanogenic-sedimentary facies association, and a correlate of the Salisbury Conglomerate. Facies association C is a volcanic facies association, and a possible correlate of the White Spur Formation. Facies D is a coherent facies, and also a possible correlate of the White Spur Formation. Facies E is a quartz-rich sedimentary facies and is a correlate of the Stitt Quartzite.

The Natone Volcanics and host sequences in the Natone Creek area comprise two volcanogenic-sedimentary facies associations (A and B), one volcanic facies association (C), one coherent facies (D), and one sedimentary facies (E). Facing indicators within sandy beds indicate that the majority of the western part of the sequence is overturned, dipping west and facing east. Contacts are gradational to sharp in the lower part of the sequence and obscured by chlorite-carbonate alteration in the upper part. The internal structure within the Natone Volcanics is also obscured by pervasive alteration and locally strong cleavage development.

Table 3.1 Major facies characteristics

Facies	Lithofacies Characteristics	Thickness	Clast Composition	Groundmass / Matrix Composition	Possible Correlate
association A: Interbedded Fine-Grained Sandstone and Mudstone	Diffuse planar lamination with rare cross bedding, grading and flute and flame structures, well sorted, crystal dominated (fragments), consistent fine grained sandstone and siltstone	2.5 - 20 m	Crystals: Quartz (15 %) [igneous + metamorphic, subangular, 3 mm * 3.4 mm], phyllosilicates, plagioclase ± orthoclase feldspar (5 %), muscovite ± biotite, tourmaline, zircon, chromite Lithics: Andesite-dacite fragment, microcrystalline quartz rich (chert?), fragment containing microcrystalline quartz and carbonate rhombs (900 µm wide, 1600 µm long)	Murky chloritic matrix ± carbonate	?
association B: Fuchsitic Conglomerate and Sandstone	Cleaved, fuchsitic, poorly sorted, matrix supported, subangular polymictic conglomerate	1 - 98 m	Crystals: Quartz (10 %) [large, polycrystalline + microcrystalline], plagioclase ± orthoclase feldspar (< 1 %), fuchsite, muscovite, biotite, minor chlorite, tourmaline, chromite, zircon, quartz-feldspar veins, carbonate (siderite?) veins Lithics: Chert, quartzite, dolomite, carbonate, sandstone, mudstone	Quartz-mica, disseminated quartz, sandstone and mudstone	Salisbury Conglomerate
association C: Volcanic-lithic Sandstone and Siltstone	Slaty? cleavage, weakly carbonate altered with patchy chlorite-sericite, massive, fiamme-bearing, quartz-feldspar-crystal-rich fine to medium grained sandstone and siltstone	> 146 m	Crystals: Quartz (15 %) [polycrystalline + microcrystalline], plagioclase ± orthoclase feldspar (3 %), albite, muscovite, biotite, sericite, chlorite, tourmaline, siderite, relic cuspsate Lithics: Fiamme (compacted pumice), mafic-intermediate porphyritic fragment, carbonate (vein), black slate - shale (1 mm - 3 cm), sandstone, muscovite-sericite-rich mudstone, siltstone	Very fine matrix, pumice, chlorite-sericite-carbonate altered, relic igneous texture	White Spur Formation
D: Coherent Dacite	Little to no cleavage, uniformly feldspar-phyric, cream-grey massive dacite	2 m	Crystals: Plagioclase ± K-feldspar (15 %), quartz (3 %) [micropoikilitic], biotite, muscovite, amphibole, pyroxene, tourmaline, chromite, zircon, trace disseminated pyrite	Feldspar, mica, quartz, patchy, carbonate altered, poorly developed snowflake texture (micropoikilitic groundmass)	White Spur Formation?
E: Quartz-Rich Sandstone	Weakly cleaved, cherty, well-sorted, uniform distribution of lithics	> 18.25 m	Crystals: Quartz (50 %) [predominantly microcrystalline (chert) ± strained polycrystalline ± metamorphic (subgrain nucleations)], plagioclase ± K-feldspar (3 %), biotite, muscovite, chromite, tourmaline, zircon, disseminated pyrite, no recognisable volcanics Lithics: Sandstone, mudstone, other lithic fragments	Chlorite-sericite-carbonate altered (mica-carbonate)	Stitt Quartzite

### 4.0 Whole-rock and chromite compositions of the Natone Volcanics and associated units

#### 4.1 Introduction

The main aim of this study is to determine the provenance and stratigraphic correlation of the Natone Volcanics. Studies of volcanic rock geochemical data can assist in constraining the composition, source area and regional correlation of lithofacies (Crawford *et al.*, 1992; Rollinson, 1993). In this chapter the compositional characteristics of the Natone Volcanics and enclosing sequences will be described and compared with possible correlates, including the White Spur Formation (WSF). The chapter focuses on whole rock and chromite compositions as potential indicators of correlations.

#### 4.2 Methods and Analytical Techniques

Two representative samples of the largely homogeneous Facies association C of the Natone Volcanics were selected from DDH BH-2 at 95 m and 126 m; and one sample of Facies D was taken from BH-2 at 147 m. The Natone Volcanics and associated units exhibit weak diagenetic alteration (Parfrey, 1993). Samples weighing approximately 200 g each were crushed to <3 cm fragments in a steel jaw crusher, and milled to powder in a tungsten carbide pulverizing ring mill. The powder was submitted for X-ray Fluorescence (XRF) and Inductively Coupled Plasma Mass-Spectrometry (ICP-MS).

Glass discs and powder pellets were prepared; whole rock major and selected trace element analyses were carried out by Jay Thompson (Analyst) using a PANalytical Axios Advanced X-Ray Spectrometer at the School of Earth Sciences, University of Tasmania. Elements analysed using XRF included the major elements as oxides and a suite of trace elements, including Zr. Major elements were determined from discs fused at 1100°C in 5% Au / 95% Pt crucibles, with a lithium tetraborate-metaborate flux. Trace elements were determined from pressed powder pills. A full set of XRF analytical conditions is included in Appendix 2. A more extensive suite of trace elements and Rare Earth Elements (REE) was determined at the University of Tasmania by Ian Little (Analyst) using solution ICP-MS methods. REE analysis utilized the methods of Robinson (2003). A full set of ICP-MS results and analytical procedures is included in Appendix 2.

XRF and ICP-MS data were screened by Dr Andrew McNeill and BSc Wayne Baker for contamination and other problems such as formulaic errors and excessive loss on ignition. ICP-MS results were used where accurate (including La, Ce, Nd, Pb and Ga) and XRF results were used for Sc, V, Ni, Cu, Zr and Y. Mobile elements were ignored (e.g. As). Results were normalized to 100 wt.% volatile free to account for loss on ignition (LOI) (Gifkins *et al.*, 2005; Rollinson, 1993), which in all cases was <3 wt.%. Because Zr was not corrected, the uncorrected Ti value was used for Ti/Zr ratios. Data were imported into and analysed in the software package ioGAS.

##### 4.2.1 Chromite Analyses

The compositions of chromite (Chr) in four samples from the Natone Volcanics, including one from Facies A (2 Chr), one from Facies B (3 Chr), and two from Facies E (11 Chr),

were determined by Electron Probe Micro-Analysis (EPMA) on the Cameca SX100 at the Central Science Laboratory, University of Tasmania. Samples were prepared as polished thin sections and then carbon coated. Analytical conditions used were an accelerating voltage of 20 keV, a current of 30 nA, and an effective beam diameter of approximately 2  $\mu\text{m}$ . Count times were (peak/background in seconds) 20/10 for Mg, Al, Si, Mn and Fe and 40/20 for Cr, Ca, Ti, V, Ni and Zn. All elements were calibrated using appropriate elemental and mineral standards and the standard Tiebghi chromite (USNM117075) was used to check for inter-session variations.

### 4.2.2 Use of the Alteration Box Plot

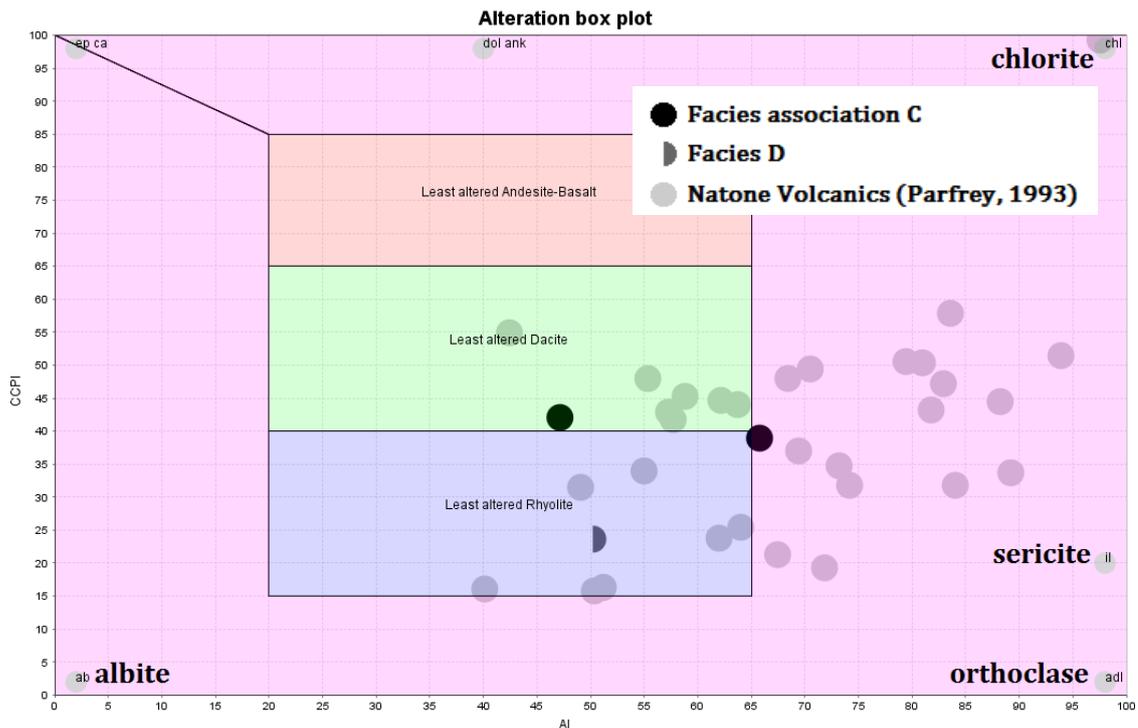
Ratios such as the Ishikawa alteration index (AI) and the chlorite-carbonate-pyrite index (CCPI) were developed to detect chemical and mineralogical changes in diagenetically and hydrothermally altered rocks (Ishikawa *et al.*, 1976; Large *et al.*, 2001a; Large *et al.*, 2001b).

The AI was developed to represent the principal components gained and lost in hydrolysis of Na and Ca bearing feldspars to form chlorite and/or sericite (Ishikawa *et al.*, 1976). The AI has been used extensively in VHMS exploration to provide estimates of the intensity of alteration; however it does not differentiate between chlorite- and sericite-altered rocks or account for carbonate alteration (Large *et al.*, 2001a).

The CCPI was developed to illustrate the prominence of chlorite, Fe-Mg carbonates and pyrite minerals commonly found in the alteration zones close to VHMS deposits (Large *et al.*, 2001a). CCPI values are dependent on the degree of fractionation in the precursor rock, for example basalt will have higher values than dacite due to higher levels of primary  $\text{Fe}_2\text{O}_3$  and MgO (Large *et al.*, 2001a).

The Alteration Box Plot (ABP) plots the Ishikawa AI against the CCPI to characterize alteration trends related to massive sulfide ores, discriminate between chlorite-carbonate or sericite-rich alteration, and to assist in the distinction of VHMS-related hydrothermal alteration from regional diagenetic alteration (Large *et al.*, 2001a). Least altered volcanics plot toward the centre of the diagram, and samples which plot outside the 'least altered box' have undergone some degree of either hydrothermal and/or diagenetic alteration (Large *et al.*, 2001a). Large *et al.* (2001a) found that the least altered samples of the MRV all displayed an AI range of 20 to 65 and a CCPI range of 15 to 85 (Figure 4.1). The mineral end members (albite, epidote, calcite, dolomite, ankerite, chlorite, sericite and adularia) plot along the boundaries of the box in the positions labelled on Figure 4.1 (Large *et al.*, 2001a).

Natone Volcanics Facies association C samples from this study plot within the least altered field and thus remain essentially unaltered (Figure 4.1). The dacite sample (Facies D) has undergone weak albite- and orthoclase-alteration, plotting outside the least altered dacite box. Seventeen of 31 samples of the Natone Volcanics taken by Parfrey (1993) plot outside the least altered squares, displaying weak to moderate chlorite-, sericite- and orthoclase-alteration; one sample plots at the chlorite end member boundary (Figure 4.1). As Facies association C samples from this study remain essentially unaltered and LOI values for the samples are low (<3 wt.%), major elements from these samples (e.g. Si, K and Na) are likely to reflect primary rock compositions.



**Figure 4.1:** ABP showing Natone Volcanics samples from Parfrey (1993) and this study. Facies association C samples taken in this study plot within the least altered field and thus remain essentially unaltered. The Facies D sample has undergone weak albite- and orthoclase-alteration and plots outside the least altered dacite box. Sixteen of 31 Natone Volcanics samples taken by Parfrey (1993) exhibit weak to moderate chlorite-, sericite- and orthoclase-alteration. One of the Natone Volcanics (Parfrey, 1993) samples is strongly altered, plotting in the top right corner at the chlorite end member.

#### 4.2.3 Immobile Elements

As indicated by the AI-CCPI boxplot Facies D of the Natone Volcanics are weakly altered. Consequently some of the elements originally present in the sample may have been mobilized, rendering the use of major element chemistry potentially unsuitable for rock classification (e.g. Figure 4.2). The mobility of major elements is controlled by the stability and composition of minerals in the unaltered parent rock, the stability and composition of the alteration product, and the temperature and volume of the fluid phase (Rollinson, 1993). Immobile elements are chemically resilient, and considered to remain relatively stable during metasomatism and metamorphism (MacLean and Kranidiotis, 1987), and are thus useful in determining the nature of primary igneous rock types (Rollinson, 1993).

Elements considered to be essentially immobile during the style of alteration that has affected the Natone Volcanics samples of Parfrey (1993) and Facies D of this study include V, Cr, Ni, Rb, Sr, Y, Ti, Zr, Nb and, particularly, the rare earth elements (REE) (Rollinson, 1993; MacLean and Barrett, 1993).

For the Facies association C samples selected for analysis, their position on the ABP and low LOIs indicate that elements such as SiO<sub>2</sub> may not have been strongly mobile; therefore SiO<sub>2</sub> remains a useful chemical discriminant, particularly when considered together with more immobile elements (e.g. Figure 4.2; Figure 4.3; Figure 4.4) (Crawford *et al.*, 1992; McNeill, pers. comm., 2013).

Total alkalis vs. silica (TAS) diagrams can be used to assign names to unaltered volcanic rocks (Le Maitre *et al.*, 2002). Figure 4.2 demonstrates that volcanic-lithic Facies association C has affinities with rhyolite. Weak alteration of the dacite sample (Facies D) results in the sample plotting outside of the dacite field.

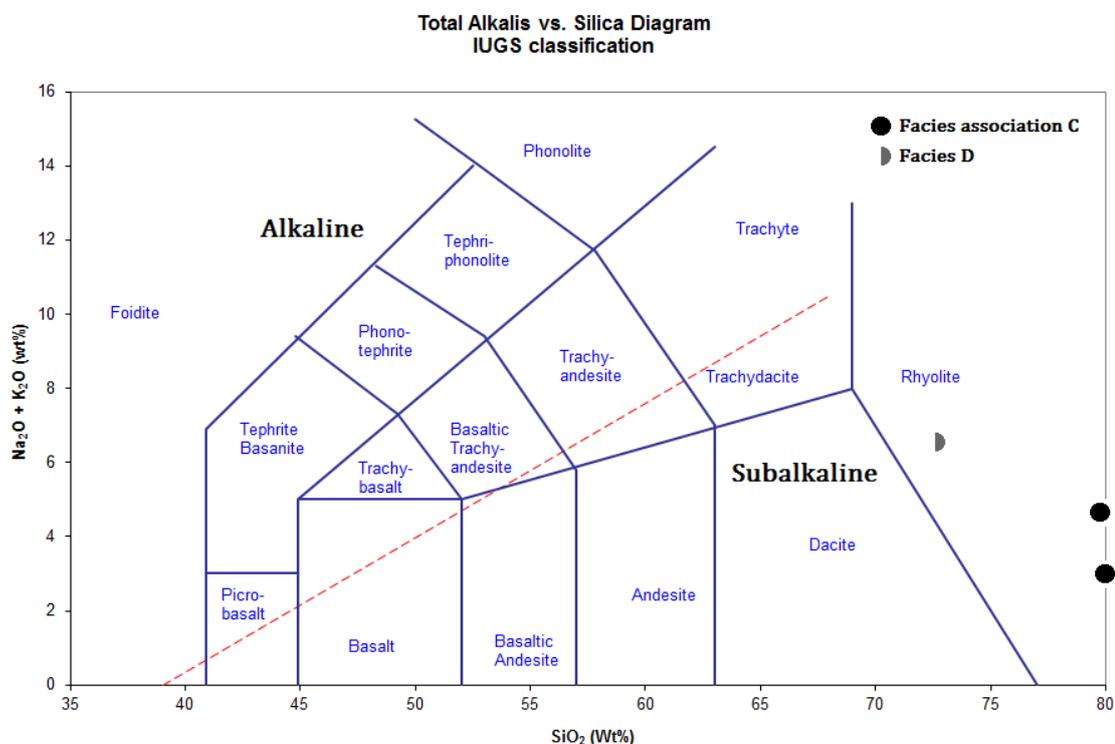


Figure 4.2: TAS diagram showing samples from this study. Volcanic-lithic Facies association C samples plot within the rhyolite field. Weak alteration of the Facies D dacite sample results in it plotting outside the dacite field.

### 4.3 Previous Studies

Parfrey (1993) analysed a suite of samples from both the Natone Volcanics and the WSF. Parfrey (1993) concluded that the Natone Volcanics show low levels of hydrothermal alteration when compared with the CVC at Rosebery, and suggested that alteration occurred with burial metamorphism to greenschist facies. Parfrey (1993) suggested that the Natone Volcanics may be correlative with the WSF.

Crawford *et al.* (1992) subdivided the MRV into five geochemical suites (Table 4.1); three calc-alkaline suites (I, II and III) and two tholeiitic suites (IV and V). These five suites are useful in aiding correlations within the belt, but are not necessarily representative of discrete stratigraphic units (Crawford *et al.*, 1992). Crawford *et al.* (1992) almost exclusively sampled coherent facies. The most voluminous facies (suite I) in the MRV are volcanoclastic; however there is minimal data. Unless there is an attempt to control sampling, variations in volcanoclastic compositions in and around the MRV are very hard to interpret. The majority of samples taken from the Natone Volcanics and WSF are volcanoclastic. For the purposes of this study, basalt data from suite V will be omitted due to its limited stratigraphic occurrence (Table 4.1).

Table 4.1: The chemical characteristics of suite I to V of Crawford *et al.* (1992) (modified from MacIntyre, 2006).

Characteristics	Suite I	Suite II	Suite III	Suite IV	Suite V
<b>Stratigraphic occurrence</b>	Volumetrically abundant - Eastern Quartz Phyric Sequence - Tyndall Group - Central Volcanic Complex - Que Hellyer Volcanics footwall	- South east of Henty fault - CVC top - Yolande River Sequence	- Lynch Creek Basalts (Yolande River Sequence) - QHV hangingwall basalt & andesite - Sock Creek basalt	- Ophiolite sequence (HFW) - Henty dyke swarm	- Miners Ridge basalt
<b>Petrology</b>	- Plagioclase-phyric dacite - Augite+plagioclase-phyric andesite - Quartz+plagioclase-phyric rhyolite	- Hornblende-phyric andesite/dacite - Augite+plagioclase-phyric andesite with partly resorbed quartz phenocrysts	- Primitive crystal-rich olivine+chromite+cpx-phyric basalt - More evolved cpx+plagioclase-phyric	- Augite+plagioclase-phyric basalt - Aphyric-ophitic basalt	- Aphyric-ophitic basalt with rare altered olivine phenocrysts
<b>Range: (La/Yb)<sub>N</sub></b>	5 - 12 (avg 8.1)	10 - 26 (avg 16.7)	8 - 12	1.4 - 3.4	N/A
<b>Range: SiO<sub>2</sub> %</b>	58 - 78	58 - 68	48 - 57	50 - 54	48 - 52
<b>Range: P<sub>2</sub>O<sub>5</sub>/TiO<sub>2</sub></b>	< 0.4	0.3 - 0.7	0.2 - 1.2	< 0.2	< 0.2
<b>Range: Ti/Zr</b>	5 - 40	15 - 20	20 - 40	70 - 110	90 - 160
<b>FeO trend with differentiation</b>	Steady decrease	Steady decrease	Decrease	Increase	Increase
<b>TiO<sub>2</sub> trend with differentiation</b>	Steady decrease	Flat decrease	Flat	Increase	Increase
<b>REE enrichment</b>	Moderate LREE-enrichment (avg (La/Sm) <sub>N</sub> = 3.52)	Strong LREE-enrichment (avg (La/Sm) <sub>N</sub> = 5.47)	Variable (avg (La/Sm) <sub>N</sub> = 4.11)	Weak LREE-enrichment (avg (La/Sm) <sub>N</sub> = 1.52)	Weak LREE depletion (avg (La/Sm) <sub>N</sub> = 0.78)
<b>Affinity</b>	Medium - high K calc-alkaline	High K calc-alkaline	- Transitional medium-high K calc-alkaline - P-rich REE-rich shoshonites	Tholeiitic	Strongly magnesian tholeiitic
<b>Interpreted tectonic setting</b>	Initial post-collisional volcanism in back-arc basin setting	Intermediate post-collisional volcanics in back-arc basin setting	Evolved post-collisional volcanics in back-arc basin setting	Extensional rift tholeiites in opening back-arc basin setting	Allochthonous inlier of pre-existing basement

### 4.4 Results and Comparison with other Mt. Read Volcanics Geochemical Data

Geochemical data from this study were compared with MRV suite I – IV data from Crawford *et al.* (1992); WSF and Natone Volcanics analyses by Parfrey (1993); WSF (at Rosebery) data from Large *et al.* (2001a); unpublished data from the North Rosebery host rocks and North-end footwall (provided by Minerals and Metals Group or MMG); and data from units in the footwall of the Rosebery Fault (Winter, 2012).

#### 4.4.1 Major Elements

Diagrams of  $\text{TiO}_2$  versus  $\text{SiO}_2$  can be useful in distinguishing calc-alkaline suites I to III from tholeiitic suites IV and V of Crawford *et al.* (1992). Likewise, diagrams showing  $\text{P}_2\text{O}_5/\text{TiO}_2$  versus  $\text{SiO}_2$  clearly distinguish suites I, II and III (Crawford *et al.*, 1992). Figures 4.3 and 4.5 demonstrate that volcanic-lithic Facies association C samples from the Natone Volcanics plot within the trend of calc-alkaline suite I. Figure 4.5 shows that Facies D also plots within the range of suite I data (Crawford *et al.*, 1992).

Suite I type rocks are the most abundant of those sampled within the MRV and include the EQPS, the Tyndall Group dacites and rhyolites, intrusive quartz-feldspar porphyries mainly along the western side of the MRV, the Darwin and Murchison subvolcanic granitoids, and the andesites of the Que-Hellyer footwall sequence (Crawford *et al.*, 1992).

Figure 4.4 shows that samples of the WSF below the Rosebery Fault (Winter, 2012) have very low  $\text{P}_2\text{O}_5$  (<0.01 wt.%) and  $\text{TiO}_2$  (0.03 wt.%), making  $\text{P}_2\text{O}_5/\text{TiO}_2$  values appear broader than the Natone Volcanics. WSF samples generally have lower  $\text{SiO}_2$  values than the Natone Volcanics. Samples from Facies association C taken in this study and by Parfrey (1993) plot within the range of Parfrey's (1993) WSF data and follow the trend of Rosebery hangingwall WSF data (MMG).

Ti/Zr vs  $\text{P}_2\text{O}_5$  (Figure 4.6) was plotted to show that the Natone Volcanics samples from Parfrey (1993) and this study plot within the range of Parfrey's (1993) WSF data and Rosebery hangingwall WSF (MMG). The Facies D sample also plots within the range of Rosebery hangingwall data provided by MMG. Samples from the WSF below the Rosebery Fault (Winter, 2012) contain very low  $\text{P}_2\text{O}_5$ , and much lower Ti/Zr than samples from the Natone Volcanics Facies association C and are thus unlikely to be correlates.

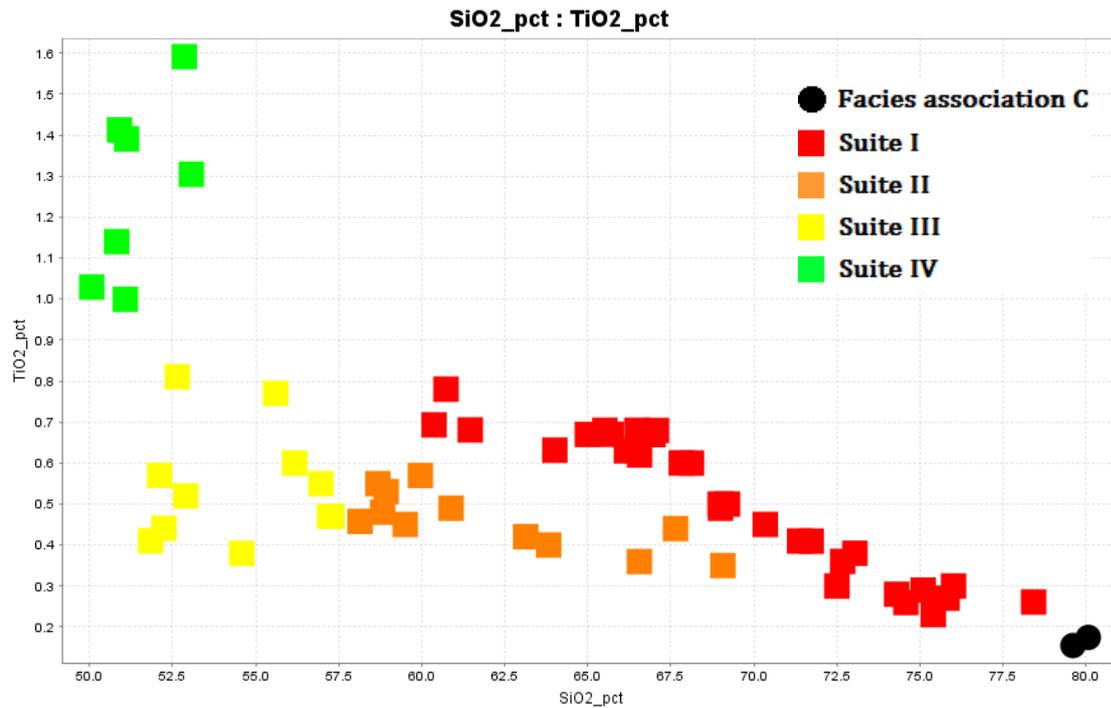


Figure 4.3:  $\text{SiO}_2$  vs  $\text{TiO}_2$  plot showing volcanic-lithic samples from this study compared with coherent facies data of the MRV (Crawford *et al*, 1992). Volcanic-lithic Facies association C samples plot within the trend of medium-high K calc-alkaline suite I samples (Crawford *et al*, 1992).

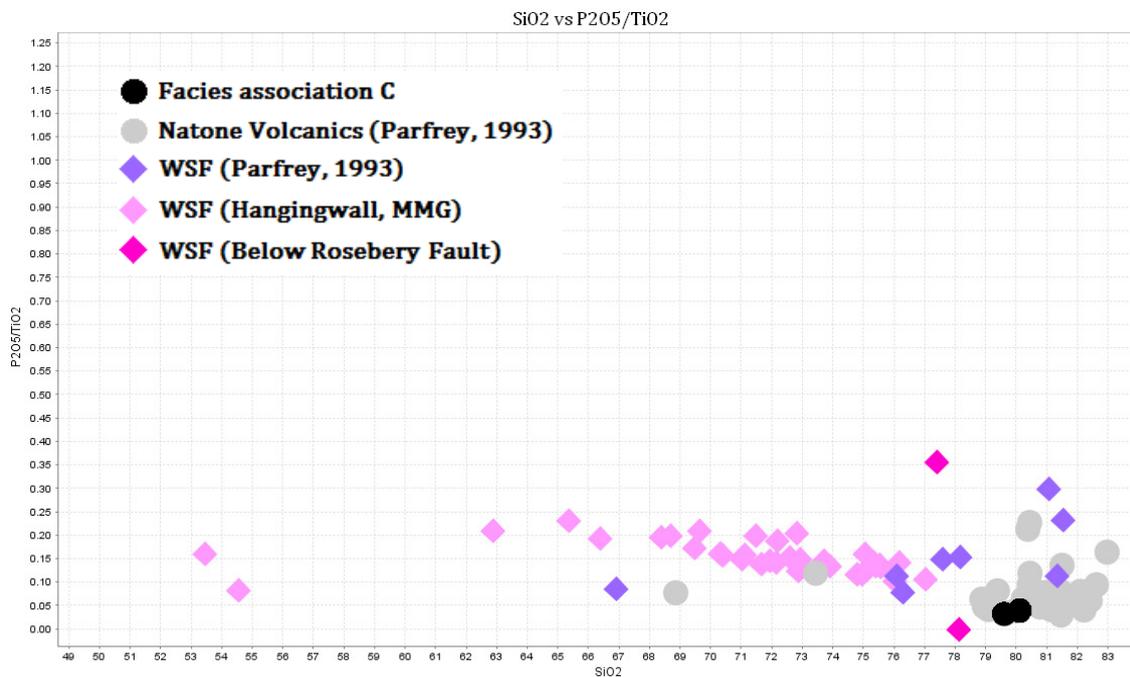


Figure 4.4:  $\text{SiO}_2$  vs  $\text{P}_2\text{O}_5/\text{TiO}_2$  plot showing data from the WSF (Parfrey, 1993; Large *et al*, 2001a; Winter, 2012; MMG) and Natone Volcanics. Samples from Facies association C taken in this study and by Parfrey (1993) plot within the range of Parfrey's (1993) WSF data and follow the trend of Rosebery hangingwall WSF data (Large *et al*, 2001a; MMG).

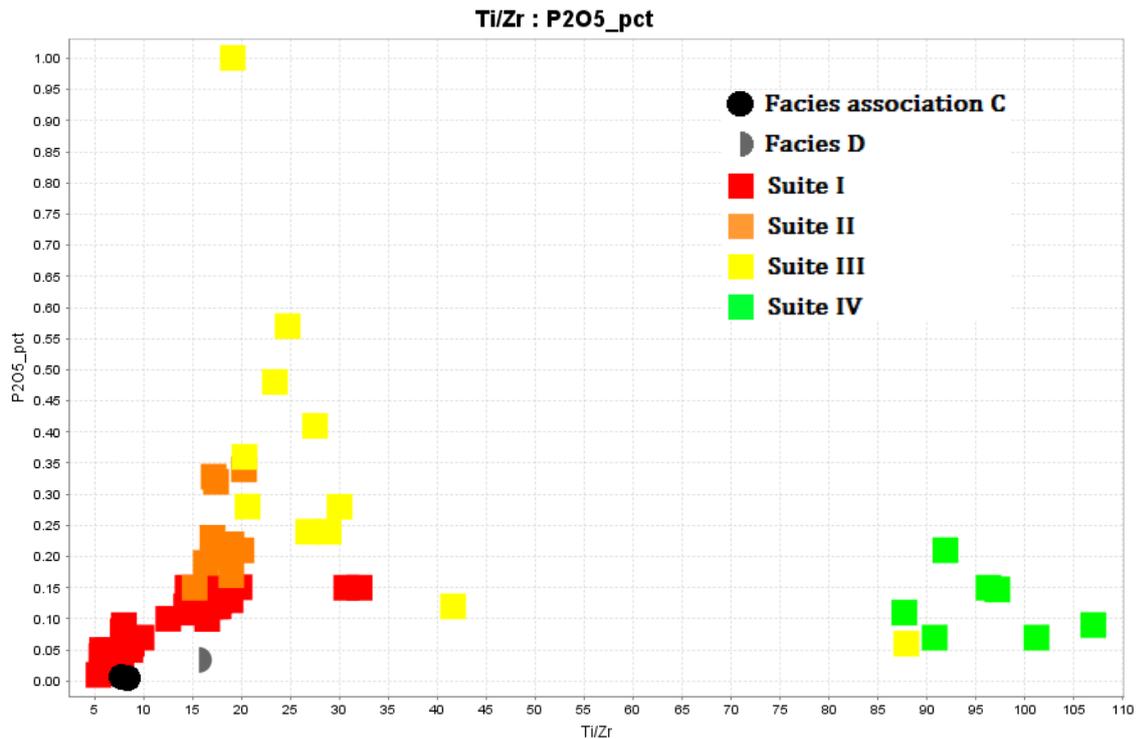


Figure 4.5: Ti/Zr vs P<sub>2</sub>O<sub>5</sub> plot for samples of the Natone Volcanics and associated Facies D compared with samples from the MRV. Volcanic-lithic Facies association C and coherent Facies D samples plot within the range of suite I data (Crawford *et al.*, 1992).

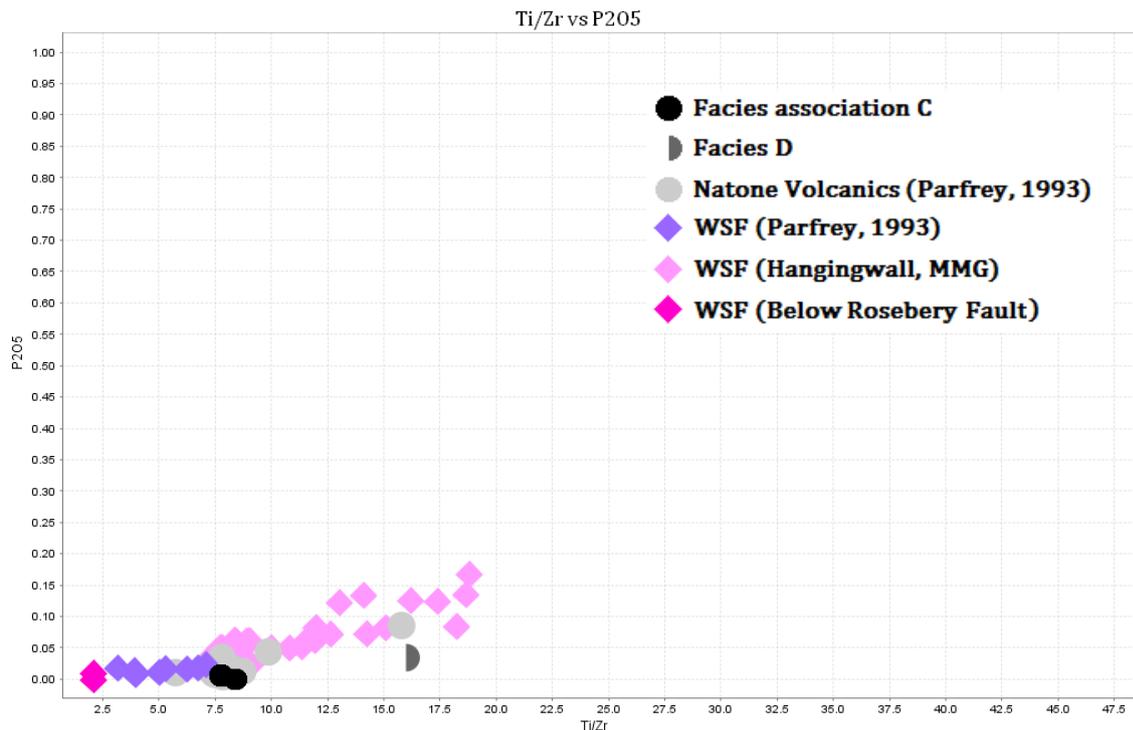


Figure 4.6: Ti/Zr vs P<sub>2</sub>O<sub>5</sub> plot showing Facies association C and Facies D samples of the Natone Volcanics compared with WSF data. The Natone Volcanics samples from this study and Parfrey (1993) plot within the range of Parfrey (1993) WSF data and Rosebery hangingwall WSF (Large *et al.*, 2001a; MMG). The Facies D sample plots close to the range of the Rosebery hangingwall. This diagram also shows that samples from below the Rosebery Fault (Winter, 2012) contain very low P<sub>2</sub>O<sub>5</sub> and much lower Ti/Zr than the Natone Volcanics.

### 4.4.2 Rare Earth Elements

Magma composition varies as minerals form during cooling. Fractional crystallization occurs when crystals are removed from the melt (Winter, 2001). Light REE or LREE (e.g. La) have a larger ionic radius and are more incompatible than heavy REE (HREE) (Winter, 2001). REE enrichment is dependent upon source concentration, the degree of partial melting and subsequent fractionation (Wilson, 1989). REE analysis thus allows for an interpretation of magma fractionation processes (Wilson, 1989; Winter, 2001).

$(La/Yb)_N$  values are a useful way of summarising and representing REE information. The REE end-members are chondrite normalized to represent primordial, undifferentiated, and thus primary rock compositions (Winter, 2001). REE diagrams illustrate the relative concentrations of elements in a given sample and can also be used to distinguish between the calc-alkaline suites I to III, and the tholeiitic suites IV and V of Crawford *et al.* (1992).

The Natone Volcanics samples (Figure 4.7) show LREE enrichment, a negative Eu anomaly and consistent low HREE enrichment. The dacite sample (Facies D) plots with a similar Eu anomaly, and higher LREE and HREE enrichment than the Natone Volcanics sample. Eu is divalent in contrast with other REEs which are trivalent, and consequently it behaves like Sr, which is highly reactive. The negative Eu anomaly is likely to be the result of a fractionating phenocryst or a residual solid in source; in this case it is an indicator of feldspar fractionation. The calculated Eu anomalies are 0.5 for both Natone Volcanics Facies association C samples and 0.7 (small) for the dacite (Facies D). Both the Natone Volcanics and dacite samples plot in the subalkaline range (High K calc-alkaline) (Figure 4.2; Figure 4.7).

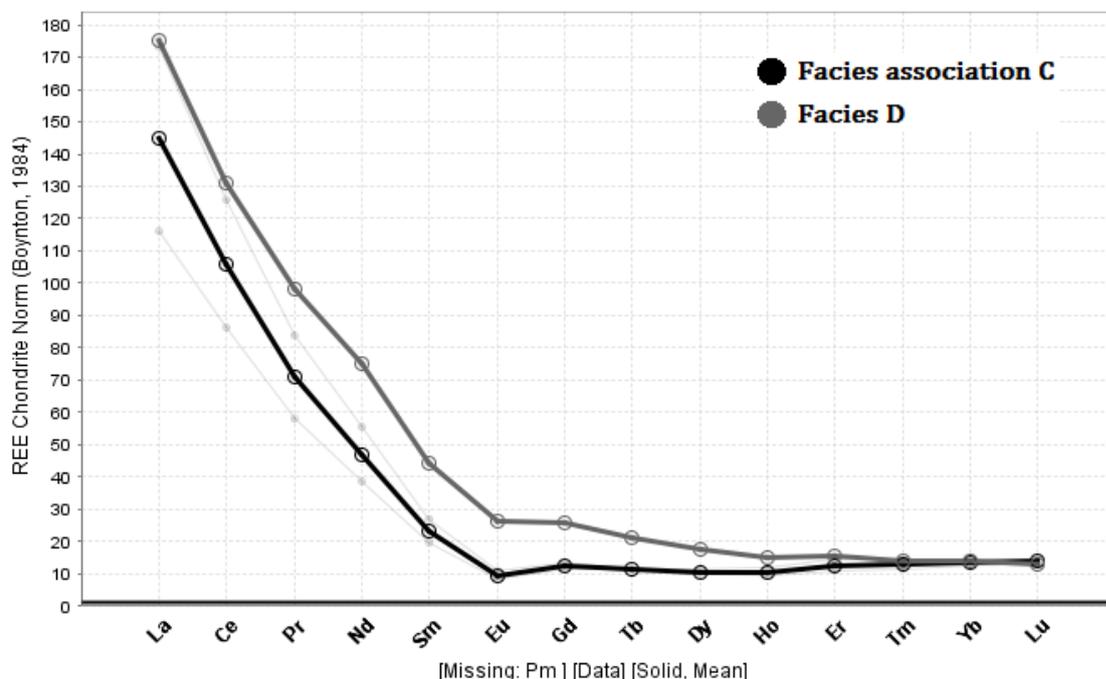


Figure 4.7: REE diagram for the Natone Volcanics Facies association C and Facies D samples. Samples from Facies association C of the Natone Volcanics are LREE-enriched, have a Eu anomaly, and consistent low HREE; Facies D are more LREE enriched than Facies association C, with a similar shaped Eu anomaly, and show HREE depletion.

REE plots have been used to compare the Natone Volcanics data with results from Crawford *et al.* (1992) and WSF data (Parfrey, 1993; Large *et al.*, 2001a; Winter, 2012; MMG). The Parfrey (1993) Natone Volcanics data were omitted as the REE analyses of these samples are incomplete. The Natone Volcanics have similar REE patterns to suite I (Figure 4.8), and have  $(La/Yb)_N$  values ranging from 8.1 to 13.8, consistent with Crawford *et al.* (1992) suite I dacites and rhyolites that have  $(La/Yb)_N$  values ranging from 5 to 12 (Table 4.1).

The Natone Volcanics Facies D (coherent dacite) REE patterns are similar to those of suite I, II and III of Crawford *et al.* (1992) (Figure 4.9). The dacite has a  $(La/Yb)_N$  value of 12.6, consistent with suite I values which range from 5 to 12, and also overlapping with suite II which ranges from 10 to 26 (avg 16.7) (Crawford *et al.*, 1992). However, suite II is more light REE and  $P_2O_5$  enriched than suite I, and has high-K calc-alkaline affinities (Crawford *et al.*, 1992). Figure 4.5 displaying  $Ti/Zr$  vs  $P_2O_5$  shows that Facies D is suite I correlative.

Figure 4.10 shows that Natone Volcanics samples taken from this study and by Parfrey (1993) are correlative with the WSF; however they do not match samples from below the Rosebery Fault (Winter, 2012). Natone Volcanics samples from Parfrey (1993) are almost identical to those that he took of the WSF. Missing data from the Rosebery hangingwall (MMG) prevents definitive judgement; however similar La, Ce and Nd can be observed. WSF below the Rosebery Fault have much lower LREE-enrichment, higher HREE enrichment and a lower Eu anomaly than the Natone Volcanics.

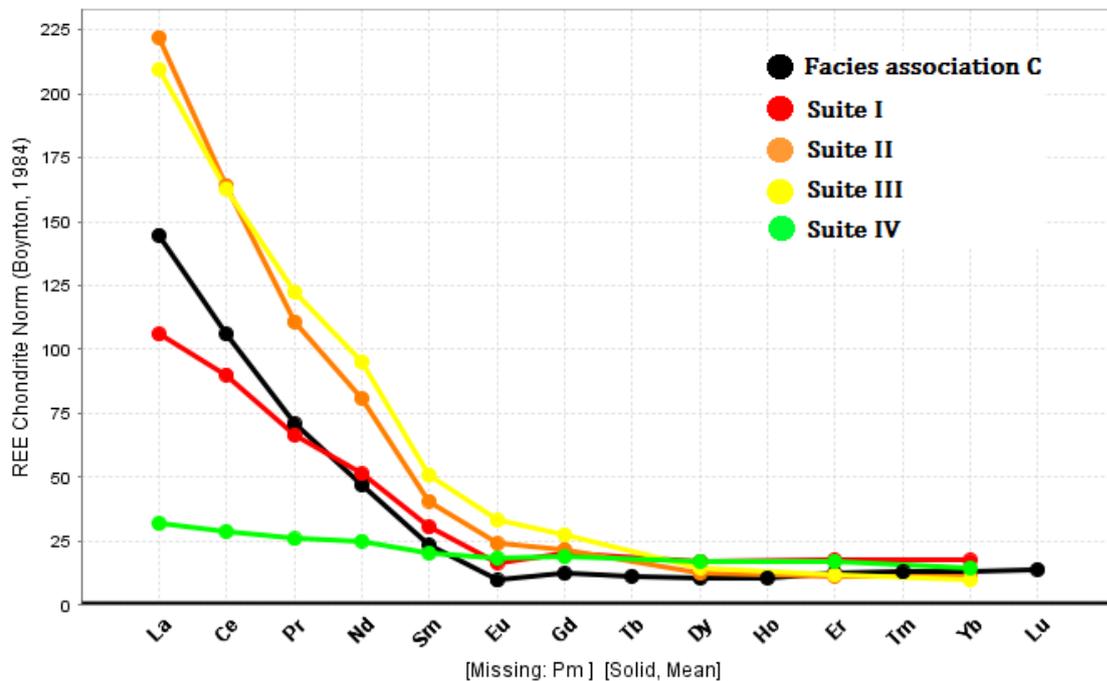


Figure 4.8: REE diagram comparing the Natone Volcanics with samples from the MRV. Facies association C has a similar REE pattern to suite I (Crawford *et al.*, 1992).

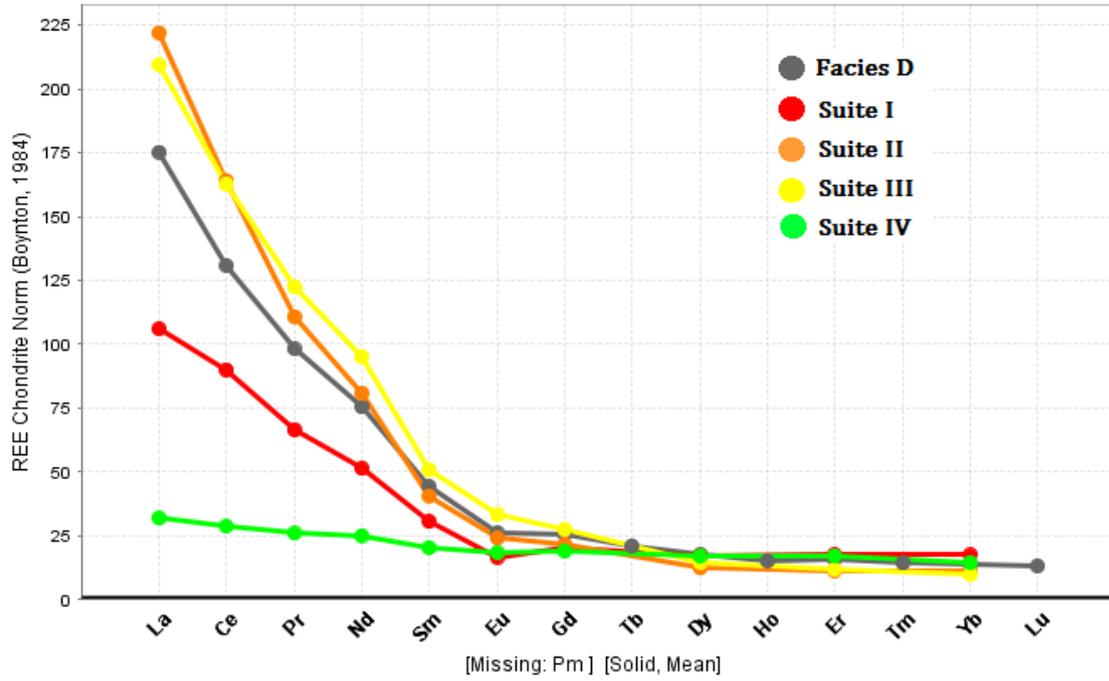


Figure 4.9: REE diagram showing Facies D compared to MRV data. The Facies D REE pattern appears similar to suite I, II and III (Crawford *et al.*, 1992).

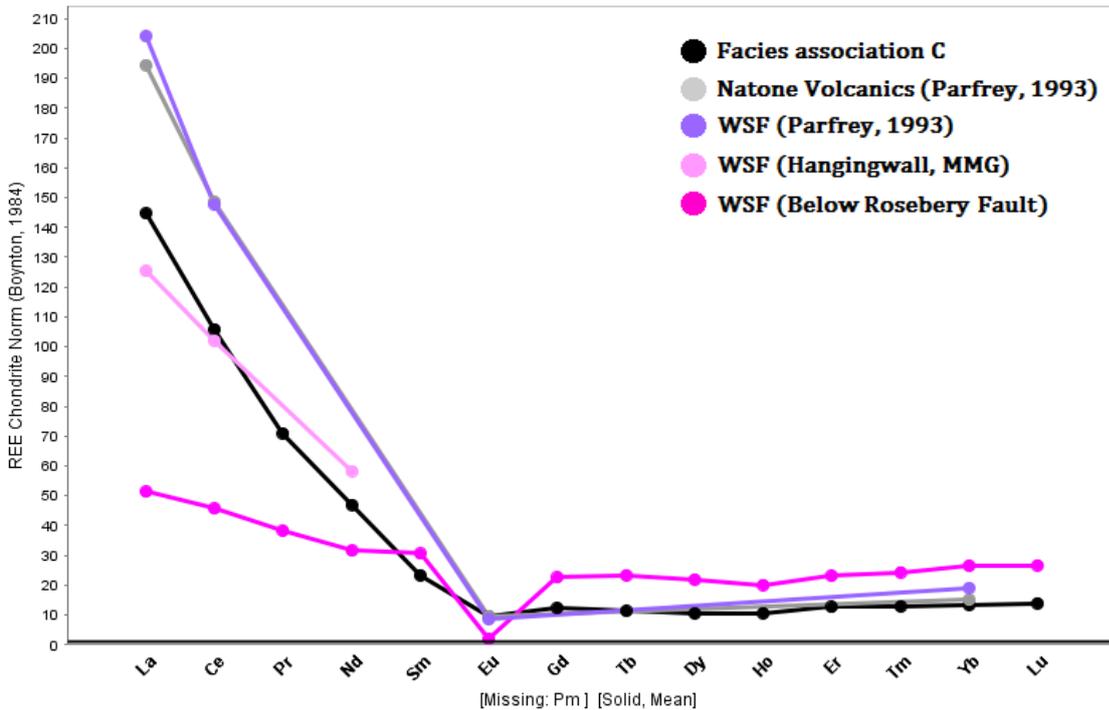


Figure 4.10: REE diagram showing samples from the Natone Volcanics compared with WSF data from Parfrey (1993), Rosebery hangingwall WSF (Large *et al.*, 2001a; MMG), and WSF below the Rosebery Fault (Winter, 2012).

### 4.4.3 Summary

Major element data were normalized to 100 wt.% volatile free in order to remove apparent variations in the proportions of major oxides due to differences in LOI (Rollinson, 1993).

Whole-rock compositional data have been used to show that the Natone Volcanics Facies association C and Facies D match suite I of Crawford *et al.* (1992). The Natone Volcanics composition is similar to that of the WSF (Parfrey, 1993; Large *et al.*, 2001a; MMG); however they do not match samples from below the Rosebery Fault (Winter, 2012).

### 4.4.4 Chromite Analyses

Primary minerals (e.g. olivine, pyroxene and plagioclase) are highly susceptible to weathering and alteration; consequently the use of these minerals as source indicators becomes problematic. Chromite is a high temperature mineral that crystallises over a wide range of conditions in mafic and ultramafic magmas. Its composition is a function of the magma from which it crystallised. The refractory nature of chromium spinel makes it a useful petrogenetic indicator relative to other high temperature phases (Barnes and Roeder, 2001). Chromite geochemistry has been successfully applied as a source indicator (Kamenetsky *et al.*, 2001).

Chromite has a specific gravity of 5.1, making it relatively dense (Deer *et al.*, 1966). It is therefore easily concentrated by settling out during sedimentary processes (Barnes and Roeder, 2001). Coupled with its chemical resilience, this makes chromite a useful provenance indicator in sedimentary rocks in western Tasmania. Detrital chromite composition has been successfully used to discriminate between sources of mafic components in Cambrian sedimentary sequences in Tasmania (e.g., Dower, 1991; Selley, 1997; van Eijndthoven, 2006; Stubbley, 2010).

At least 3 compositional sources of detrital chromite have previously been defined (Greenhill, 1995; Berry *et al.*, 1997; Selley, 1997):

1. Passive margin tholeiitic basalts (e.g. the Crimson Creek Formation).
2. Boninitic volcanic and peridotitic cumulates (Mafic Ultramafic Complexes; MUC).
3. Shoshonitic basalt in the MRV.

Direen and Crawford (2003) inferred that tholeiites of the Crimson Creek Formation (CCF) were emplaced during rifting of the Western Tasmanian Terrane from the East Antarctic Margin (Berry and Bull, 2012). These basalts have been dated to ~580 Ma (Meffre *et al.*, 2004). Low-Ti tholeiites (wt.% TiO<sub>2</sub> ~0.1 to 0.6) are associated with MUC in the Heazlewood, Serpentine Hill and McIvors Hill complexes (Crawford and Berry, 1992; Dower, 1991). Tonalite in the Heazlewood River Complex has been dated at ~515 Ma (Turner *et al.*, 1998), and ~515.9 Ma (Stubbley, 2010) and fossil evidence from strata overlying MUC at Serpentine Hill constrain emplacement of the MUC allochthon during the Tyennan Orogeny at ~505.2 – 504.5 Ma (Jago and Bentley, 2010). The MRV were emplaced from 506.8 to 496.0 Ma, and the shoshonitic units of the Hellyer area were emplaced between 502.5 and 499.6 Ma (Mortensen *et al.*, unpublished).

The compositions of detrital chromite found in Facies association A, Facies association B and Facies E of this study were compared with MUC data sourced from Brown (1986), Peck and Keays (1990); Hellyer Basalt data from Crawford *et al.* (1992) and Miedecke (2011); and CCF data provided by Meffre *et al.* (2004).

Mantle derived peridotite chromite is typically more reduced ( $Fe^{2+}/Fe^{3+} > 5$ ) than chromite from volcanic sources ( $Fe^{2+}/Fe^{3+} < 5$ ) (Kamenetsky *et al.*, 2001). Furthermore, the  $TiO_2$  wt.% value of chromite reflects the  $TiO_2$  values of the parental melt (Kamenetsky *et al.*, 2001).  $TiO_2$  can thus be plotted against  $Fe^{2+}/Fe^{3+}$  to discriminate between intraplate basalts with high  $TiO_2$  and peridotitic accumulates and subduction related volcanics including boninites with low  $TiO_2$  values (Kamenetsky *et al.*, 2001).

Figure 4.11 shows that Facies association A chromite are correlative with MUC data (Brown, 1986; Peck and Keays, 1990), and both Facies association B and Facies E contain chromite of mixed provenance, predominantly MUC. Relatively low  $TiO_2$  values observed within chromite from Facies association A (<0.07 wt.%), Facies association B (<0.15 wt.%), and Facies E (<0.1 wt.% ) suggest subduction related source rocks for these minerals (Figure 4.11).

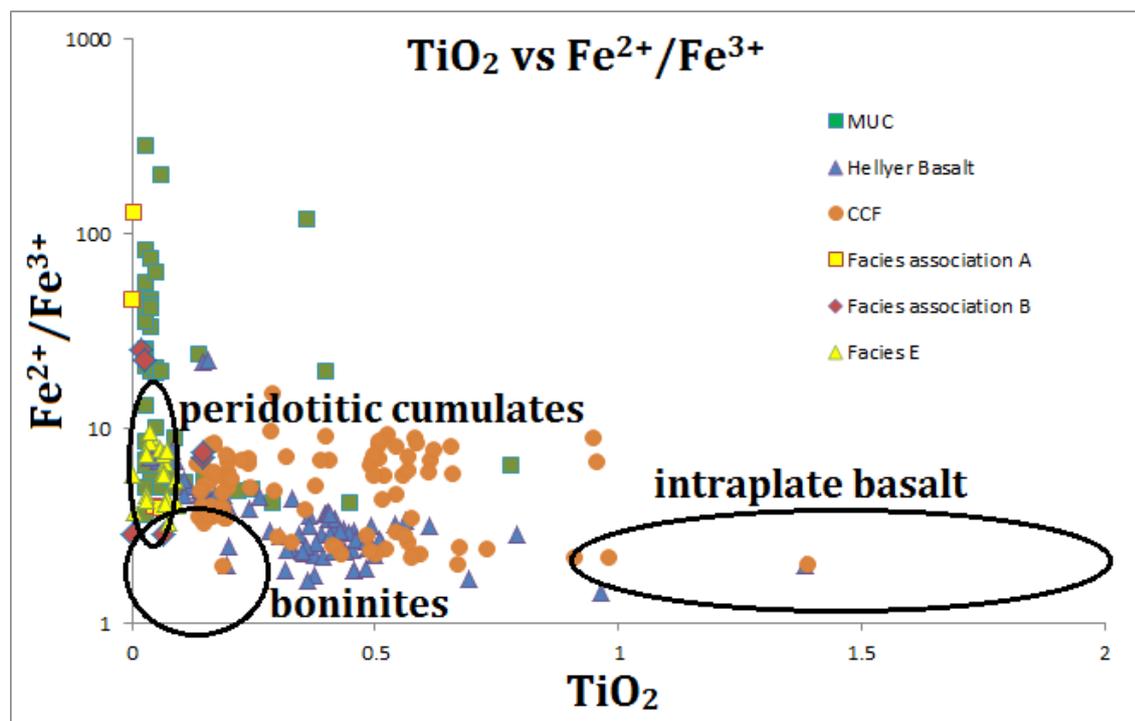


Figure 4.11:  $TiO_2$  vs  $Fe^{2+}/Fe^{3+}$  diagram comparing MUC (Brown, 1986; Peck and Keays, 1990), Hellyer Basalt (Crawford *et al.*, 1992; Miedecke, 2011) and CCF (Meffre *et al.*, 2004) chromite with Facies association A, Facies association B, and Facies E chromite. Facies E chromite plot within the field of peridotitic cumulates (after Kamenetsky *et al.*, 2001).

The chrome number (Cr#) abbreviates the relationship  $Cr/(Cr+Al)$  and can be useful in discriminating between basalt types (Barnes and Roeder, 2001). Boninites have higher Cr# numbers and mid-ocean ridge tholeiites have low numbers, whereas intermediate values indicate an ocean island basalt source (Roeder, 1994). This variability reflects depletion of a mantle peridotite source, i.e. Al availability (Roeder, 1994).

High Cr# values (>60) observed in chromite from Facies association A, Facies association B and Facies E of the Natone Volcanics (Figure 4.12) indicate increasing source depletion through generations of melting (Roeder, 1994). The overall range of chromite compositions indicates correlation with MUC.

Figure 4.13 and 4.14 demonstrate that Facies association A chromite match the composition of MUC chromite (Brown, 1986; Peck and Keays, 1990); Facies association B chromite overlap MUC and Hellyer Basalt (Crawford *et al.*, 1992; Miedecke, 2011); and Facies E chromite overlap MUC, Hellyer Basalt and CCF data (Meffre *et al.*, 2004).

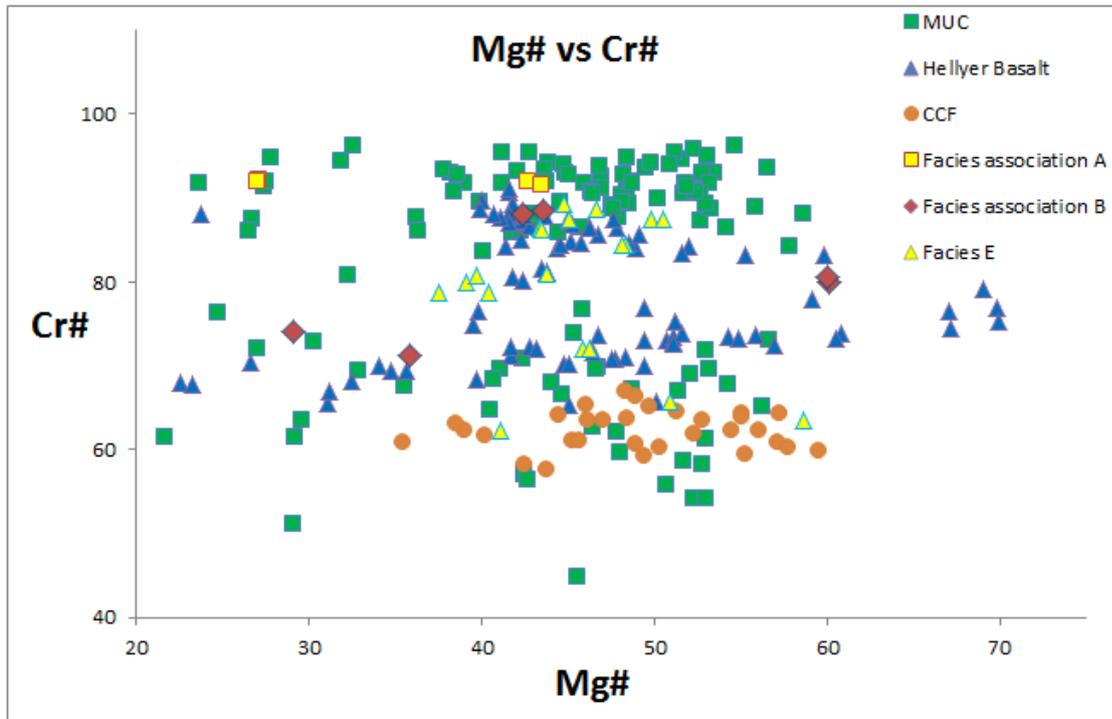


Figure 4.12: Mg# vs Cr# plot showing MUC (Brown, 1986; Peck and Keays, 1990), Hellyer Basalt (Crawford *et al.*, 1992; Miedecke, 2011) and CCF (Meffre *et al.*, 2004) data compared with the compositions of Facies association A, Facies association B, and Facies E chromite from this study.

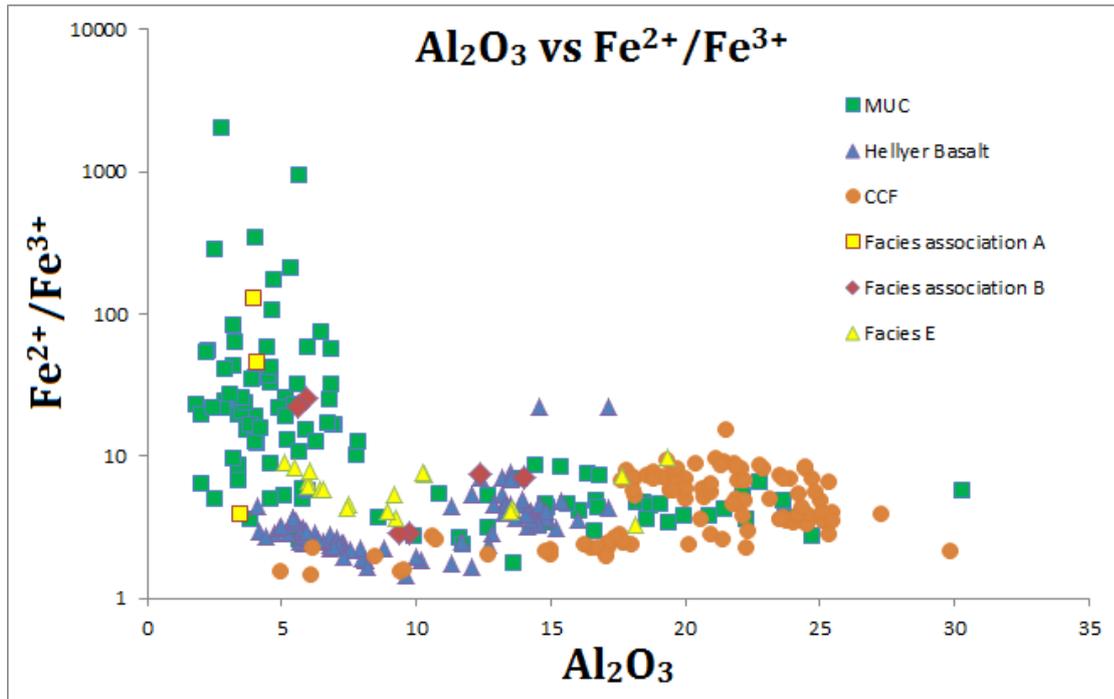


Figure 4.13: Plot of Al<sub>2</sub>O<sub>3</sub> vs Fe<sup>2+</sup>/Fe<sup>3+</sup> comparing samples from MUC (Brown, 1986; Peck and Keays, 1990), Hellyer Basalt (Crawford *et al.*, 1992; Miedecke, 2011) and CCF (Meffre *et al.*, 2004) with chromite from Facies association A, Facies association B, and Facies E of the Natone Volcanics.

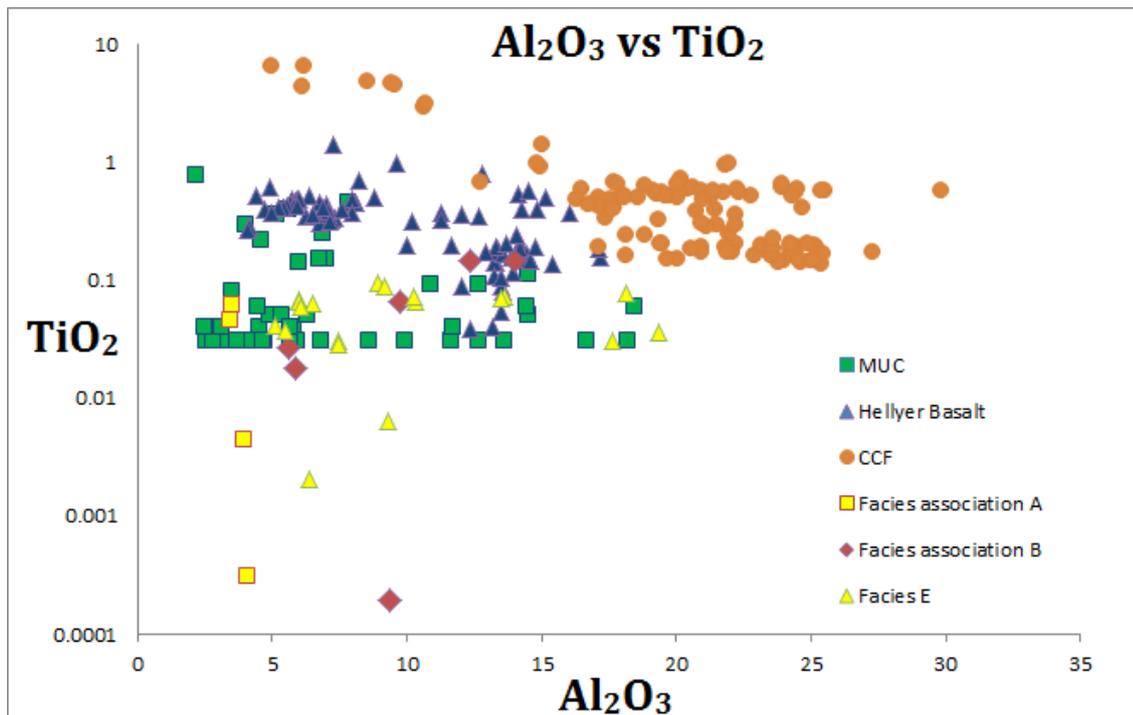


Figure 4.14: Diagram of Al<sub>2</sub>O<sub>3</sub> vs TiO<sub>2</sub> for MUC (Brown, 1986; Peck and Keays, 1990), Hellyer Basalt (Crawford *et al.*, 1992; Miedecke, 2011) and CCF (Meffre *et al.*, 2004) data compared with chromite from Facies association A, Facies association B, and Facies E of the Natone Volcanics (MUC TiO<sub>2</sub> samples below detection limit are shown at half the detection limit). Facies association A chromite overlap MUC data, and Facies association B and Facies E chromite are of mixed provenance, however the majority of samples plot within MUC range.

### 4.4.5 Summary

Detrital chromite found within Facies association A match the composition of MUC chromite (Brown, 1986; Peck and Keays, 1990). The majority of Facies association B and Facies E samples plot within MUC range, displaying low-Ti and high  $\text{Fe}^{2+}/\text{Fe}^{3+}$ ; however they also overlap slightly with the Hellyer Basalt (Crawford *et al.*, 1992; Miedecke, 2011) and even less so with CCF data (Meffre *et al.*, 2004). Facies E chromite plot within the field of peridotitic cumulates (Kamenetsky *et al.*, 2001). High Cr# values ( $>60$ ) represent increasing depletion through generations of melting (Roeder, 1994).

### 4.5 Summary of Whole-rock and Chromite Compositions

Whole-rock compositional data has shown that the Natone Volcanics Facies association C and Facies D match suite I of Crawford *et al.* (1992). Major element relationships show that the Natone Volcanics samples closely match samples of the WSF and Rosebery hangingwall sequence (Parfrey, 1993; Large *et al.*, 2001a; MMG); however they are different to samples of the WSF taken from below the Rosebery Fault (Winter, 2012).

Facies association A chromite match composition of chromite from MUC (Brown, 1986; Peck and Keays, 1990). The majority of Facies association B and Facies E samples plot within MUC range, however they also overlap with Hellyer Basalt (Crawford *et al.*, 1992; Miedecke, 2011) and even less so with CCF data (Meffre *et al.*, 2004), indicating a mixed provenance. The range of chromite provenance for Facies association A, Facies association B and Facies E samples supports the theory of erosion and deposition of the sedimentary rocks of the Rosebery Group from intra-basinal sources (Corbett, 2002).

5.0 Discussion and Conclusion for the Correlations and Exploration Significance of the Natone Volcanics

5.1 Introduction

The overall aim of this project is to clarify the position of the Natone Volcanics (NV) with respect to the Rosebery-Hercules host sequence. A linked study of the volcanology, sedimentology and provenance of the NV and enclosing units was undertaken in order to determine the likely stratigraphic correlations of the sequence, and thus its relationship with the Mount Read Volcanics (MRV). In this chapter the results of the study will be discussed.

5.2 Stratigraphy

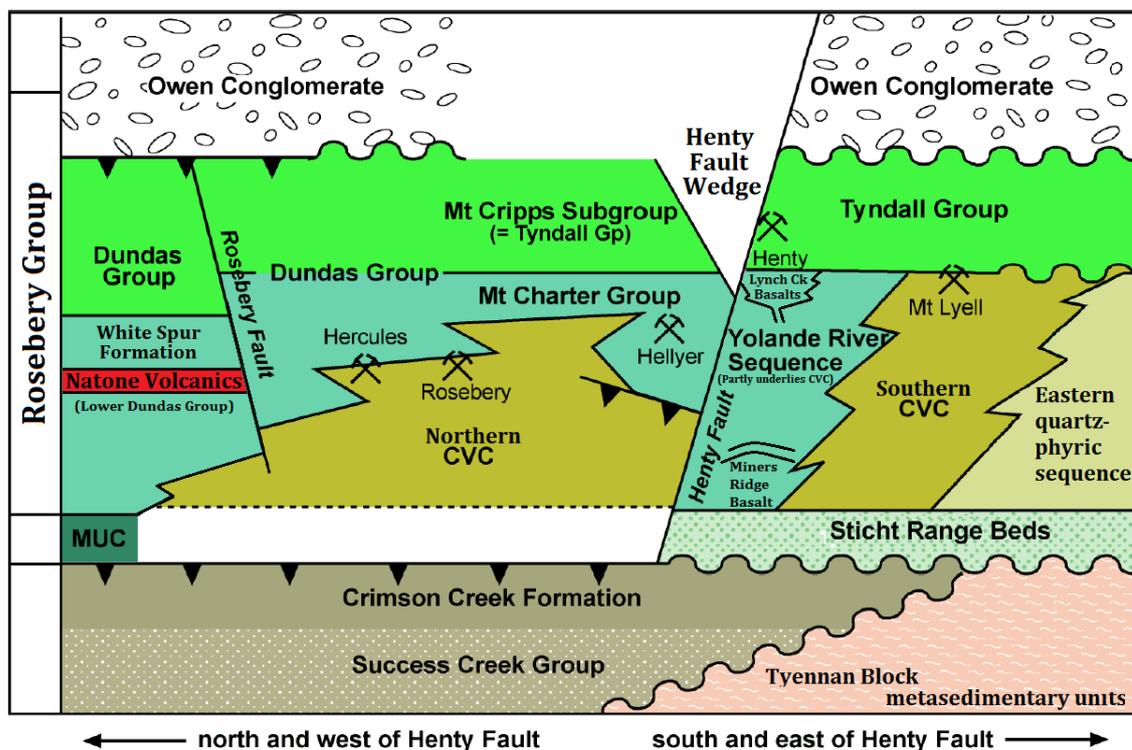


Figure 5.1: Lithostratigraphic units of the MRV showing the Rosebery Group and approximate position of the NV (modified from Martin, 2004).

The Central Volcanic Complex (CVC) of the MRV hosts the Pb-Zn rich polymetallic Rosebery and Hercules VHMS deposits (Figure 5.1). The Rosebery-Hercules host unit is separated from the Mount Black Volcanics and the Sterling Valley Volcanics by the Mount Black Fault, and is bound to the west by the Rosebery Fault and volcano-sedimentary rocks of the Rosebery Group. The unit above the Rosebery-Hercules host unit is called the black slate and hangingwall volcanoclastics by mine geologists (Corbett, 1992). Gifkins and Allen (2001) showed that the unit overlying the Rosebery-Hercules host unit is a correlate of the White Spur Formation (WSF).

The WSF occurs west of Rosebery and is a probable correlate of the Rosebery hangingwall sequence (Corbett, 1992). The WSF is bound to the west by the Stitt Quartzite (Lees, 1987) and to the east by the Rosebery Fault, and persists eastwards for at least 1.5 km

beneath the fault (Corbett and Lees, 1987). Green (1983), Lees (1987) and Parfrey (1993) identified lithological similarities between the NV and the WSF. The NV are thus potentially a correlate of the hangingwall to the Rosebery-Hercules host unit.

### 5.2.1 Geology

The NV is part of the Rosebery Group (Green, 1983) which occurs between the Rosebery Fault to the east and the Marionoak Fault to the west (McNeill and Simpson, 2000). The Rosebery Group includes sandstone, mudstone, fuchsitic conglomerate and altered volcanoclastic horizons. The sequence has been subject to tectonic deformation, with numerous fault bounded strips of differing lithologies, and extensive overturning of beds (Corbett and McNeill, 1986; Corbett and Lees, 1987; Corbett, 2002).

The NV appear as a local lens of weakly chlorite-sericite altered, massive ungraded, volcanic lithic, fiamme-bearing, quartz-feldspar rich fine grained sandstone and siltstone conformably underlain by the Salisbury Conglomerate (SC) (Figure 5.2).

<b>Rosebery Group (Green <i>et al.</i>, 1981)</b>			<b>Rosebery-Hercules (Gifkins and Allen, 2001)</b>
<b>(Central Units)</b>	<b>(Eastern Units)</b>		
<b>Stitt Quartzite</b>	<b>Mudstone Dominated Sequence</b>	<b>Dundas Group</b>	<b>Tyndall Group</b>
	<b>Westcott Argillite</b>		
?	<b>Chamberlain Shale</b>		
<small>Dacite</small>	<b>White Spur Formation</b>		
<b>Natone Volcanics</b>			
<small>SC</small>		<b>Hangingwall v'c'c</b>	
?		<b>Black Slate</b>	
<b>Northern Central Volcanic Complex</b>			<b>Rosebery-Hercules Host</b>
			<b>Footwall v'c'c</b>

**Figure 5.2:** Lithostratigraphic units of the Rosebery Group (Green *et al.*, 1981) and the Rosebery-Hercules mine sequence (Gifkins and Allen, 2001) showing the position of the NV inferred from this study.

Five facies were identified in DDH BH-1 (Green, 1983) and BH-2. Facies association A (interbedded fine-grained sandstone and mudstone) is of uncertain correlation, however Green (1983) observed a similar sequence below the Salisbury Conglomerate (SC) in the Pieman River Gorge section. Facies association B (fuchsitic conglomerate and sandstone) interfingers with, and stratigraphically overlies Facies association A. Facies association B is a correlate of the SC described by Green *et al.* (1981), Green (1983), Lees (1987) and Parfrey (1993). Facies association C, or the Natone Volcanics of Green (1983), Parfrey (1993) and this study, occurs within and conformably above Facies association B. While the contact between Facies association C and Facies D (dacite) is obscured by alteration, it is probable that the dacite is an intrusion. Facies E (quartz-rich sandstone) is unrelated to Facies D, and is likely a correlate of the Stitt Quartzite described by Green *et al.* (1981).

Facies association C (NV) appears to young east and conformably overlies the SC (Facies association B), however up-hole grading in Facies association A at 17 m (west younging) in DDH MD-1 (Green, 1983), followed by down-hole younging at 50 – 53 m and deeper, and east younging grading observed in sandy horizons of the SC on the Emu Bay Railway (Figure 5.4) indicate that at least one fold is present below the sequence. The NV and SC along the Emu Bay Railway are overturned, facing east but dipping steeply west. Facies association C (NV) structurally overlies Facies D (dacite), which is inferred to be a younger intrusion at the top of the NV. Facies D structurally overlies the west younging Stitt Quartzite correlate (Facies E). A fault contact is inferred from changes in younging direction (Figure 5.4).

Facies association C is largely homogeneous in composition, allowing for regional correlation, while making it difficult to correlate internal stratigraphy between drill holes. The volcanic quartz was most likely deposited from a single eruptive event, however the presence of minor lithics implies emplacement as a mass flow as suggested by Green (1983).

### 5.2.2 Discussion

The NV are volcanic quartz-phyric, suggesting that they are not a correlate of the CVC, which is predominantly feldspar-phyric. Overlying the host sequence at Rosebery are feldspar-quartz-phyric volcanoclastic units which are correlated with the WSF of the Dundas Group (Large *et al.*, 2001a; Corbett, 2002). Parfrey (1993) described the NV as lithologically similar to the WSF on the basis of abundant volcanic quartz phenocrysts. The Eastern Quartz Phyric Sequence (EQPS) and other quartz-phyric sequences of the MRV can largely be ruled out on the basis of stratigraphic occurrence (Figure 5.1).

### 5.3 Geochemistry

Whole-rock compositional data have been used to classify NV Facies association C (volcanic-lithic sandstone and siltstone) into suite I of Crawford *et al.* (1992). Suite I of the MRV is volumetrically abundant and includes Tyndall Group quartz-plagioclase-phyric rocks and CVC plagioclase-phyric rocks (Crawford *et al.*, 1992).

Major element relationships (e.g. SiO<sub>2</sub> vs P<sub>2</sub>O<sub>5</sub>) show that samples of Facies association C taken in this study and by Parfrey (1993) closely match Parfrey's (1993) upper WSF samples and Rosebery hangingwall data (Large *et al.*, 2001a; MMG). Rare Earth Element (REE) data from WSF samples taken below the Rosebery Fault (Winter, 2012) show significant variation to NV samples taken by Parfrey (1993) and this study, suggesting that the sequence below the Rosebery Fault is different.

The Facies D (dacite) sample from this study plots within range of Rosebery hangingwall data (Large *et al.*, 2001a; MMG). Facies D is also inferred to be suite I (Crawford *et al.*, 1992). The Rosebery-Hercules host sequence of the CVC contains similar dacitic intrusions (Crawford *et al.*, 1992).

The provenance of detrital chromite found within units surrounding the NV was constrained by comparing immobile elements ratios with data from chromite found in areas of the MRV. Chromite grains in Facies association A (fine-grained sandstone and

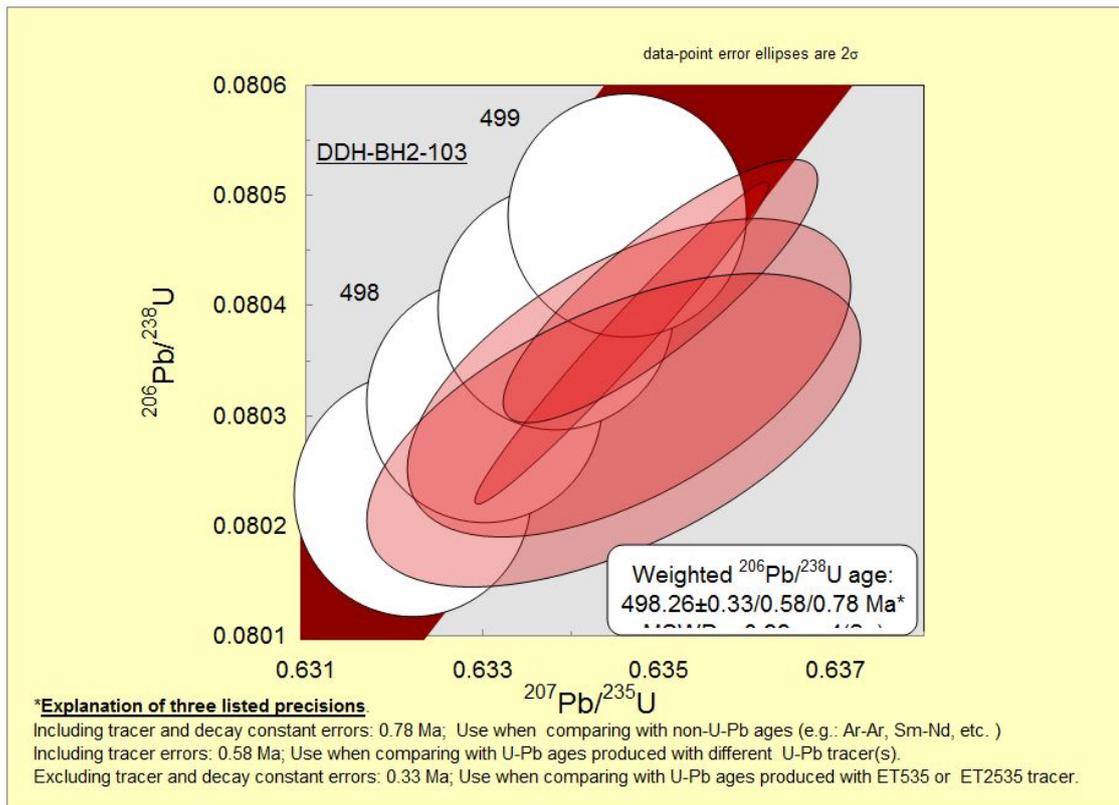
mudstone) are similar to chromite in the Mafic-Ultramafic Complexes (MUC) reported by Brown (1986) and Peck and Keays (1990). Fossil evidence (trilobite) from strata at Serpentine Hill constrain emplacement of the MUC allochthon at ~505.2 – 504.5 Ma (Jago and Bentley, 2010). Deposition of Facies association A is likely to have occurred after emplacement of the MUC.

Facies association B (SC) and Facies E (Stitt Quartzite) chromite grains are also similar to chromite in MUC (Brown, 1986; Peck and Keays, 1990) and overlap slightly with Hellyer Basalt chromite samples (Crawford *et al.*, 1992; Miedecke, 2011), and even less so with Crimson Creek Formation (CCF) chromite (Meffre *et al.*, 2004), indicating a mixed provenance. Shoshonitic basalt units of the Hellyer area were emplaced between 502.5 and 499.6 Ma (Mortensen *et al.*, unpublished). Basalts of the CCF have been dated to ~580 Ma (Meffre *et al.*, 2004).

### 5.4 U-Pb Dating

The footwall member of the Hercules Pumice Formation (upper CVC) at Rosebery and Hercules has been dated at ~503 Ma (Mortensen *et al.*, unpublished). The age of the “host rock” sequence (Rosebery-Hercules host sequence) at Rosebery is inferred to be ~499 Ma, 2-4 m.y. younger than the footwall. Mineralization in the Rosebery-Hercules area appears to be confined to this host sequence of post-CVC and pre-Tyndall Group age (Mortensen *et al.*, unpublished). Dating of syn-volcanic intrusions at the base of the WSF indicates that deposition occurred at ~499-500 Ma. Rocks of the Lower Tyndall Group (Mt Cripps Subgroup) have been dated to ~496 Ma. The majority of VHMS deposits within the MRV formed within a narrow time window at  $500 \pm 1$  Ma (Mortensen *et al.*, unpublished).

U-Pb dating of zircons from Facies association C of the NV (Figure 5.3) indicates that the unit was emplaced at  $498.26 \pm 0.78$  Ma (Mortensen *et al.*, unpublished). The NV are therefore younger than the CVC and older than the Tyndall Group. This date places the NV within the lower echelons of correlation with the WSF and inside the time window for mineralization. However, nowhere is the WSF host to economic mineralization. Locally the WSF contains clasts of sulphides most likely derived from the Rosebery host and therefore must be younger than the Rosebery-Hercules mineralization event. At Rosebery-Hercules, mineralization is thought to have occurred in a shallow sub-seafloor environment (Corbett, 1992), younger than the units it occurs within (host unit). The WSF and NV are thus not prospective. The critical part of the stratigraphy underlies the WSF. It can then be inferred that the NV are hangingwall to the Rosebery-Hercules host sequence.



**Figure 5.3: U-Pb dating of a sample from the Natone Volcanics indicates that deposition occurred  $498.26 \pm 0.78$  Ma (DDH BH-2 103 m; Mortensen *et al.*, unpublished).**

## 5.5 Conclusion

The geology west of the Rosebery Fault has long been a subject of confusion (Green, 1983; Parfrey, 1993). Complex geometries such as folding and overturned beds, disappearance of the SC along strike, alteration and poor outcrop have led to indefinite conclusions. Corbett (2002) considered the Rosebery Group to be largely a submarine equivalent to the Late Cambrian Owen Group.

This study has confirmed that the Natone Volcanics are a correlate of the WSF and Rosebery hangingwall sequence. Petrographically, the NV most resembles the middle sub unit of the WSF described by Lees (1987). This polycyclic central unit comprises quartz-feldspar and lithic-rich volcanic sandstone, volcanogenic sandstone and shale. The composition of NV samples taken in this study closely match the composition of samples from the WSF and Rosebery hangingwall reported by Parfrey (1993), Large *et al.* (2001a), and MMG (unpublished data). U-Pb dating of zircons found within Facies association C (NV) has been used in conjunction with petrographic and geochemical information to correlate the unit with the WSF and the Rosebery hangingwall (Mortensen *et al.*, unpublished).

The mixed provenance of chromite in the SC covers a broad time period (~580 Ma to 499.6 Ma). The latter date suggests that deposition began later than Facies association A (<504.5 Ma). The Stitt Quartzite is interpreted to have been deposited contemporaneously with the Late Cambrian to Early Ordovician Owen Conglomerate (Jell *et al.*, 1991; Corbett, 2002), and thus younger than the Salisbury Conglomerate.

The mixture of chromite types in Facies association A, Facies association B and Facies E samples supports the Rosebery Group having a local intra-basinal provenance as suggested by Corbett (2002).

### 5.6 Recommendations

The thickness of the WSF (<1500 m) (Lees, 1987) may vary laterally and thus the position of the WSF in relation to the Rosebery-Hercules host sequence remains unclear. Further drilling below the Rosebery Fault and mapping of the NV will provide insight into the lateral variation of the WSF and the distribution of the NV. The WSF dips west at approximately 80°, and a syncline can be inferred from the repetition of units to the west. Drilling beneath the Rosebery Fault might encounter either Facies association C or Facies D of the NV within the WSF, allowing for confirmation of the postulated syncline.

U-Pb dating of the dacite is recommended as it is feldspar-phyric and its composition plots in the suite I field of Crawford *et al.* (1992). Although an intrusion has been inferred, its relationship with the NV is obscured by alteration in DDH BH-1, and the contact with the Stitt Quartzite unclear.

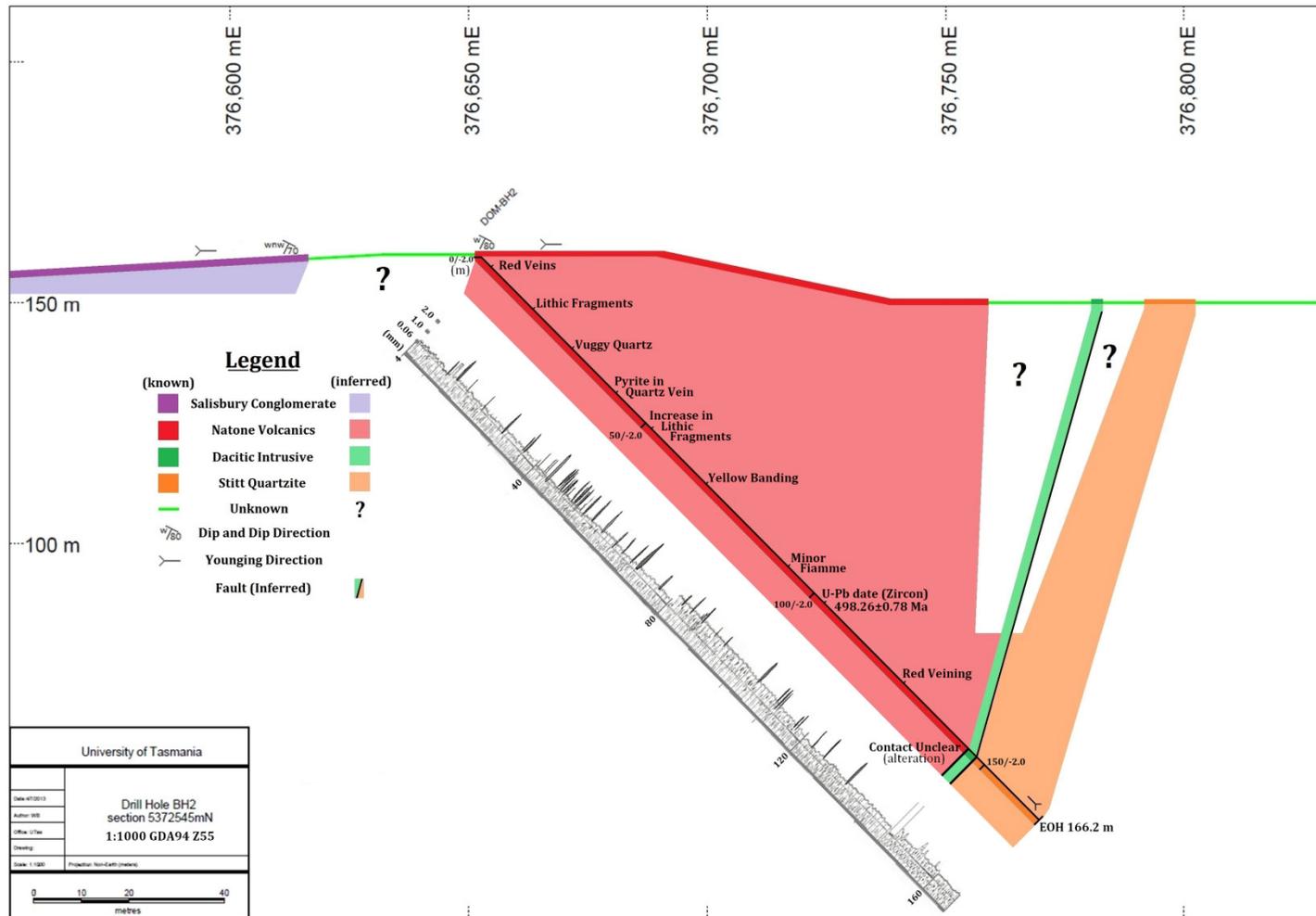


Figure 5.4: East-West cross section of DDH BH-2 at 5372545mN (GDA94/Zone 55). Facing evidence indicates that the NV and SC are overturned (Facing east but dipping steeply west).

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