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REGIONAL TECTONIC SETTING AND EVOLUTION
OF THE
LATE PROTEROZOIC MOUNT HARPER VOLCANIC COMPLEX,
OGILVIE MOUNTAINS, YUKON

by

Charles F. Roots, A.B., MSc.

A thesis submitted to the
Faculty of Graduate Studies and Research
in partial fulfillment of the requirements for
the degree of Doctor of Philosophy

Department of Earth Sciences
Carleton University
Ottawa, Ontario, Canada
December 4, 1987

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FRONTISPIECE. Mount Harper, looking east from the centre of the Volcanic Complex. The underlying conglomerate (Lower Harper Group) occupies the valley at left, with Pinguicula carbonate further left and beyond. The darker weathering volcanic rocks in ridges leading to Mount Harper constitute the most complete section of the Lower Suite (800 m thick), mostly basaltic pillows (lower spur) and scoriaceous flows and lapillistone (near summit). A remnant of Upper Harper Group sediments (Saddle outlier) lies on right skyline, with snowpatch.

The undersigned hereby recommend
to the Faculty of Graduate Studies and Research
acceptance of this thesis, submitted by Charles F. Roots,
in partial fulfillment of the requirements
for the degree of Doctor of Philosophy.

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ABSTRACT

The Ogilvie Mountains, near the Yukon-Alaska boundary, include inliers of Middle to Late Proterozoic strata within the Mackenzie Platform. Among these, the Harper Group preserves a record of crustal extension that correlates with the major unconformity that underlies Windermere Supergroup rocks along the eastern side of the North American Cordillera.

The Harper Group comprises three main units: a lower formation of fault-derived conglomerate-sandstone, the lenticular, bimodal Mount Harper Volcanic Complex (MHVC), and an upper clastic succession. Zircon from a felsic flow in the MHVC yield an U-Pb crystallization age of $751 \pm 26/-18$ Ma for extrusions near the beginning of Windermere deposition.

The MHVC is predominantly tholeiitic basalt, but includes minor felsic flows and dykes. It displays the results of repeated episodes of crustal movement and volcanism. The Complex has been divided into six lithostratigraphic members on the basis of field relations and chemical composition. Pillow flows (Member A), were overlapped by breccias and flows (Member B) that apparently constructed a volcanic island. Pyroclastic and epiclastic breccias (Member C) record the wasting and subsidence of the northern part of this early edifice. In the southern part, subaerially erupted rhyolite flows (Member D) and coeval andesitic floods (Member E) overlie eroded members A and B. Basaltic tuff and sills (Member F) are intercalated with overlying sediments.

The MHVC has been subjected to moderate tilting and was probably distorted by northward shortening during the late mesozoic Cordilleran orogeny. Normal fault movement both preceded and accompanied volcanic activity, and localized several dyke swarms. Many faults were reactivated during post-volcanic tectonic events.

Whole-rock chemical analyses reveal two magmatic suites. Members A and B are homogeneous, and relatively primitive, with low ($<1.0\%$) TiO_2 and $\text{Mg}/\text{Mg}+\text{Fe}$ ratios up to 0.54. Members D and E are differentiated, with 1.5-3 % TiO_2 and $\text{Mg}/\text{Mg}+\text{Fe}$ of 0.1-0.2; they have continental affinity. Trace element ratios are similar in the two suites, suggestive of a single magma source. Member F may be of hybrid origin. Contemporaneous igneous rocks related to the Windermere Supergroup are iron-rich tholeiitic or alkalic basalt, more enriched in incompatible elements than any member of the MHVC.

The MHVC erupted in a half-graben and appears to have resulted from penetration of the bounding normal fault to the top of the mantle. The Harper Group is a uniquely well-exposed succession that records the initial extension that led to crustal separation along the ancient western margin of North America.

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This thesis combines work on the mountains, in the laboratory and behind a desk; each setting has presented special challenges but given me the privilege of working with dedicated associates, and the resources of several institutions.

Foremost, I acknowledge Bob Thompson of the Geological Survey of Canada (GSC) for hours of stimulating discussion and unstinting encouragement of this study. He has championed the Ogilvie Mountains and Mount Harper in particular as an outstanding example of Cordilleran passive margin geology, and increased the exposure of my work. His leadership has been my inspiration.

My thesis supervisor, John Moore, has been sage and considerate. He brought a rare spell of fine weather on his visit to Mount Harper, and in recognizing the different fault systems helped elucidate the superimposed contraction and extension structures. I am most grateful for our regular discussions, his careful editing and considerable moral support. Professors Dick Brown and Ray Yole (Carleton University), Calvin Pride and Tony Fowler (University of Ottawa) advised me on points of structure, stratigraphic code, petrology and geochemistry.

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Early drafts were scrupulously edited by Fred Roots, Bob Thompson and Wayne Goodfellow. Gerry Carlson introduced me to the mysteries of the Ogilvie Mountains 'way back in 1978. The Mount Harper Complex first became that way during scintillating discussions between the floor shows at Diamond Tooth Gertie's Saloon in Dawson, Yukon.

....came out with a fortune last fall;
But somehow life's not what I thought it,
And somehow the gold isn't all.

Robert Service
"The Spell of the Yukon"

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Chapter One

INTRODUCTION

Most mountain belts are considered to have resulted from convergence of lithospheric plates. Their pre-collision geologic history has more recently become better understood; commonly a continental shelf assemblage - stable for long periods of geologic time - has been subjected to rifting that extended and weakened the basement rocks. Thus it appears that, in some cases, early rifting determined either the location or patterns of later contraction. Recognition of rifting as an early stage in the history of continental margins that are now exposed in mountain belts has provided better understanding of the behaviour of Earth's crust through time.

Numerous episodes of extension are recorded in the shelf deposits of the Cordilleran mountain belt. Late Proterozoic extension (ca. 750 million years ago) is examined here. This episodic or protracted extension event contributed to the development of a continental margin in latest Proterozoic or early Cambrian time (575 ± 25 million years ago; Bond *et al.*, 1984). Later, in Late Mesozoic time (ca. 170 million years ago) collision of continental fragments and island arcs with the continental margin forced the older sedimentary rocks upward and toward the craton (northeastward in Yukon). One of the best exposures of the Late Proterozoic rift succession lies in the Ogilvie Mountains of west-central Yukon.

This thesis is concerned with volcanic rocks which were erupted during that rift event. The Mount Harper Volcanic Complex (used in the sense of the North American Stratigraphic Code, 1983) comprises a record of volcanic eruptions and structures, which in addition to the compositional evolution preserved in lava flows, elucidate the relationship between faulting, volcanism and sedimentation of a rift-related sedimentary succession. Furthermore, an isotopic date from zircons of a

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rhyolite lava flow indicates the age of enclosing sedimentary strata, which can then be correlated with contemporaneous, rift-related successions in the Cordillera lacking isotopically dateable rocks.

LOCATION AND PHYSIOGRAPHY

Mount Harper Volcanic Complex comprises volcanic rocks in the southern Ogilvie Mountains, about 80 km northwest of Dawson, Yukon Territory (Fig. 1). Mount Harper ($64^{\circ}49'N$, $135^{\circ}35'W$) is one of two peaks similarly named (the other lies in adjacent Charley River quadrangle, Alaska) after Mr. Arthur Harper, a pioneer miner who outfitted prospectors in central Yukon during the 1880's, before gold deposits of the Klondike were discovered (Ogilvie, 1913).

The southern Ogilvie Mountains, bounded to the south by the physiographic Tintina Trench (occupied by major faults), have two general rock successions that produce contrasting landforms. Most prominent is a rugged central area (80 by 35 km) of treeless ridges of Proterozoic carbonate rock. To the north and south it is flanked by more subdued, vegetated ranges composed of fine-grained, Proterozoic and Paleozoic clastic rocks. Mount Harper (1865 m; 6120'), on the southwest boundary of the carbonates, forms one of the higher peaks in that part of the Ogilvie Mountains, and is frequently subjected to moisture-laden weather systems from the central Alaskan lowlands.

Parts of western Yukon and Alaska (called the Beringia refugium; Hopkins *et al.*, 1982) were not covered by the Pleistocene ice sheets, although major valleys of the central Ogilvie Mountains contained alpine glaciers. Coal Creek, near Mount Harper, occupies a glacier-carved, flat-bottomed valley, but most of its tributaries are narrow gorges. These mountains were shaped by erosion (mostly rockfall) which tends to emphasize rock types resistant to weathering, such as the ancient basaltic flows.

Fig. 1.1. Location of Dawson map-area, Yukon Territory.

Fig. 1.2. Major drainage, topography and cultural features in Dawson map-area.

In middle Tertiary time, the Ogilvie Mountains were headwaters of a major southeast-flowing drainage across central Yukon. Isostatic rebound of the lower watershed and uplift in the Coast Mountains resulted in a major drainage reversal 10-15 million years ago (Tempelman-Kluit, 1980) and Coal Creek now feeds the Yukon River which flows swiftly westward through entrenched meanders toward central Alaska and the Bering Sea.

PREVIOUS INVESTIGATIONS

The Ogilvie Mountains are remote and contain no established settlements. Exploration for base and precious metals has been sporadic, often prompted by federal government-sponsored regional surveys. The earliest recorded expedition, in 1887, surveyed the Alaska-Yukon boundary (Ogilvie, 1917), preceding discovery of the Klondike placer goldfields. A five-mile-wide transect along the Alaska-Yukon boundary (Cairnes, 1914) established the geologic framework used today. Systematic (1:250,000 scale) mapping of the Ogilvie and Wernecke Mountains began in 1959, and regional reports (Green and Roddick, 1962; Green, 1972) have been augmented by studies of Paleozoic stratigraphy further north (Norris *et al.*, 1963; other references in Norris and Yorath, 1981). The best studied part of the Ogilvie Mountains includes Paleozoic and younger units near Tombstone Mountain (Tempelman-Kluit, 1970). Recent mapping has clarified Middle and Late Proterozoic stratigraphy and structural relationships (Thompson and Roots, 1982; Anderson, 1987). Related studies, more local in scope, include: stratigraphy of the central region (Blaise and Mercier, 1984; Mercier, 1985); volcanic rocks (Roots, 1982, 1983; Roots and Moore, 1983); sedimentology of the Harper Group (P.S. Mustard, Ph.D. thesis in prep.), paleomagnetism of the Mount Harper volcanic rocks (Brunet, 1986); Proterozoic intrusive breccias (R. Lane, M.Sc. thesis in prep.) and the syenite plutons (Olade and Goodfellow, 1978; Anderson, 1987).

No mines are currently operating in the Ogilvie Mountains, although numerous mineral showings have been tested. Tertiary lignite

was extracted from small operations in the Tintina Trench near Coal Creek at the turn of the century. Zinc-lead-silver prospects in Proterozoic carbonate rocks, uranium in Cretaceous granitic intrusions, adjacent copper skarns and breccia pipes with uranium, gold and rare-metal potential are the principal mineral prospects in the Dawson map-area (see: Tempelman-Kluit, 1981 and the annual Yukon Mineral Industry reports; Department of Northern Affairs, Whitehorse, Yukon).

PRESENT WORK

This thesis is a contribution to the Ogilvie Mountains Project of the Geological Survey of Canada. The project will provide comprehensive maps, as well as structural and stratigraphic analyses for Dawson map-area.

Portions of two 1:50,000 map sheets (116 B/12 and 116 C/9) were investigated in 1981, and the volcanic rocks near Mount Harper were mapped and sampled, and stratigraphic sections measured in 1982. Field work by Bob Thompson and co-workers in adjacent regions during 1983 and 1985 led to recognition of rock successions representing periodic extensional tectonism in the Ogilvie Mountains, and the relationship of the volcanic rocks to the sedimentary units therefore became an important element of this thesis. The basic data consist of maps (1:25,000 scale) showing the distribution of sedimentary (Fig. 3.1, in pocket) and volcanic rocks (Fig. 4.1, in pocket), and structures (Fig. 5.1, in pocket); petrography of rock samples using megascopic and microscopic methods (140 thin sections) and whole-rock analyses (106 representative volcanic and related intrusive rocks).

Chapters 2 and 3 outline the regional setting and sedimentary rocks of the Harper Group. Chapters 4 through 7 document the lithostratigraphy, structure, age and chemical composition of the Mount Harper Volcanic Complex. The tectonic setting and geologic history of the volcanic rocks are summarized in the final chapters.

Chapter Two

REGIONAL GEOLOGY AND TECTONICS

GENERAL FRAMEWORK

The Cordilleran orogen (Fig. 2.1) is generally considered to have resulted from the accretion of crustal fragments and oceanic arcs to the western margin of North America, between Middle Triassic and Miocene time (e.g. Monger and Price, 1979; Coney *et al.*, 1980). The easternmost of its five belts, called the Foreland, consists of sedimentary rocks thrust eastward (in the south) and northeastward (in Yukon). The Mackenzie, Wernecke and Ogilvie Mountains are part of the Foreland Belt. They expose laterally shortened strata that earlier formed the broad continental shelf and basin (sometimes called the miogeocline) off the western margin of ancient North America during most of the Paleozoic era. This study concerns the Late Proterozoic rocks related to the inception of the miogeocline.

The Ogilvie Mountains comprise two Paleozoic tectonic elements called the Mackenzie Platform and Selwyn Basin (Fig. 2.2). They are separated near Mount Harper by a steeply dipping thrust fault (Dawson-Fault) that coincides with the Paleozoic carbonate-to-shale facies transition. The boundary appears to overlie an older structural break, possibly a normal fault that was a precursor to formation of the Selwyn Basin.

The Selwyn Basin consists principally of black shale (Road River Formation), but south of Mount Harper the underlying grit unit, here called the Hyland Group, is extensively exposed. The Mackenzie Platform consists of thickly bedded Cambrian to Devonian dolostone, called the CD₀ formation near Mount Harper, beneath which are Middle and Late Proterozoic strata exposed by uplift in an erosional inlier (the

Fig. 2.1. Tectonic units of the Canadian Cordillera, modified from Tempelman-Kluit (1980) and Eisbacher (1982). Area of Figure 2.2 is outlined, and circled letters show location of sections in Fig. 2.7. Mount Harper lies in the Ogilvie Mountains, near B.

Fig. 2.2. Distribution of Proterozoic and Paleozoic rocks in central Yukon. Grid outlines 1:250,000 scale map areas (numbered); regional geologic reports are listed in Appendix A. Line A-A' is location of cross-section in Figure 2.6.a. Outlined area, the southern Ogilvie Mountains, is shown in Figure 2.3.

Coal Creek Dome of Green, 1972). Along the southern edge of the Coal Creek inlier (alternative new name) is a succession of mostly clastic rocks called the Harper Group, which enclose the Mount Harper Volcanic Complex (Fig. 2.3).

The Harper Group is the youngest Proterozoic unit of the Proterozoic shelf succession, which comprises thick carbonates separated by thinner or wedge-shaped clastic intervals (Fig. 2.4). The shelf succession is interpreted to represent a stable platform depositional environment, punctuated by episodes of tectonic instability.

In contrast, rocks beneath the Selwyn Basin are poorly exposed. Fine-grained clastic and carbonate rocks of Late Proterozoic to Cambrian age are here called the off-shelf succession, or Hyland Group, in regional correlations.

The appearance of the Mackenzie Platform and Selwyn Basin as tectonic elements approximately coincided with deposition of clastic units which represent an extensional episode, correlated along the length of the Cordillera and including the Harper Group.

This chapter summarizes the stratigraphy of the shelf and off-shelf successions. The Proterozoic strata of the Southern Ogilvie Mountains have been inadequately described, and the repeated extension events have only recently been recognized (R.I. Thompson, pers. comm., 1985; Thompson *et al.*, 1987). Furthermore the shelf succession, encompassing strata from about 1500 Ma to less than 500 Ma old, is among the most complete Proterozoic records of sedimentation in the Cordillera and worthy of comparison with better-known stratigraphy of southeastern British Columbia. The Proterozoic structural elements are also significant because they have influenced patterns of later deformation.

Fig. 2.3. General geology of the southern Ogilvie Mountains. Map is from Green (1972) with modifications from Thompson and Roots (1982) and additional field data. Line B-B' is schematic section (Fig. 2.6,b). Outlined area is shown on 1:25,000 scale maps of the stratigraphy and structure of the Mount Harper Volcanic Complex (Figs. 3.1, 4.1 and 5.1; in pocket).

Fig. 2.4. Schematic stratigraphic sections of the shelf and off-shelf successions near Mount Harper. All units vary in thickness, and lithology along strike; for carbonates average local thicknesses are shown, and the Harper Group is entirely schematic. Sections are modified from Thompson and Roots (1982) with additional unpublished data from R.I. Thompson and P.S. Mustard.

STRATIGRAPHY

The **off-shelf succession** is divided into the Hyland Group, Cambro-Ordovician volcanics, and Road River Formation (Table 2.1).

The Hyland Group, named after a lithologically similar succession in Nahanni map area (105 I on Fig. 2.2; S. Gordey, in prep.) contains three units, and its base is unknown. The oldest rocks are chlorite-muscovite phyllite and sandstone, extensively exposed in the eastern part of the Southern Ogilvie Mountains and also present in thin strips 11 km south of Mount Harper. Locally the sandstone coarsens to grit which contains distinctive blue quartz granules typical of metamorphic source rocks. The middle limestone unit is discontinuous in the eastern Ogilvie Mountains, but forms a prominent stratigraphic and structural marker south of Mount Harper. The upper division of the Hyland Group mainly comprises maroon and green argillite and siltstone. It locally hosts Oldhamia, a trace fossil of deep water depositional environments and of Late Hædrynian to Middle Cambrian age (Hofmann and Cecile, 1981; Nowlan et al., 1985). Scour and tool marks are commonly preserved in mudstone beneath sandstone layers and suggest sediment transport was from the south or west. These indications of a source region beyond the carbonate shelf succession suggest that the clastic basin had two sides, and uplifted continental crust lay to the west (present coordinates).

The Hyland Group is correlated with the Windermere Supergroup, a term now used to refer to predominantly clastic successions that underlie the lowest Paleozoic carbonate and quartzite (Eisbacher, 1981). The tectonic implications of the Windermere Supergroup are addressed in the second part of Chapter 3.

Age

Unit

Lithology

SHELF SUCCESSION

Early Ordovician to Devonian	Road River Formation (165 m)	black shale, chert
transgressive contact		
Early Cambrian to Devonian	"CD_b" formation (>250 m)	thick-bedded dolostone
disconformity; local angular unconformity, thrust contact		
Late Proterozoic	HARPER GROUP	
700 Ma →	Upper Harper Group (0-450 m)	paraconglomerate, shale, siltstone, sandstone
	Mount Harper Volcanic Complex (0-1500 m)	basaltic flows, breccia minor rhyolite flows
	Lower Harper Group (0-1100 m)	orthoconglomerate, pebbly sandstone, mudstone
disconformity, local angular unconformity		
Middle to Late Proterozoic	PINGUICULA GROUP	
~1200 Ma →	Cryptalgal member (0-300 m)	thick-bedded dolostone
	Fifteenmile Form- ation (2500 m)	
	Rackla unit (0-1500 m)	limestone, siltstone, conglomerate, olistoliths
angular unconformity; locally conformable		
Middle Proterozoic	WERNECKE SUPERGROUP	
	Gillespie Lake Group (300 m)	limestone, dolostone
	Quartet Group (>1500 m)	phyllite, siltstone.

OFF-SHELF SUCCESSION

Early Ordovician to Devonian	Road River Formation (200 m)	black shale, chert
conformable, locally disconformable or thrust contact		
Mid. Cambrian to Mid. Ordovician	C-O volcanic rocks (0-600 m)	basaltic breccia, vesicular flows
disconformable, locally conformable or thrust contact		
Late Proterozoic to Early Cambrian	HYLAND GROUP (>2000 m)	maroon and green shale, white dolo-limestone quartz grit, sandstone

Table 2.1. Table of Formations, Mount Harper area.

The Early Paleozoic volcanic rocks comprise vesicular meta-basalt flows, mostly pillowed, and flow-related breccias that outcrop in a band along Dawson Fault and in synclinal keels 15 km south of Mount Harper (Fig. 2.3). Numerous, widely separated meta-diorite dykes may have been feeders which imply that the volcanics were once more extensive and have been removed by erosion. Least deformed exposures are oblique cross-sections of overlapping subaqueous shield volcanoes. Isolated limestone outcrops were probably reefs. Brachiopod fragments (R.I. Thompson, unpubl. data) and a conodont fragment of Early Ordovician age (M.P. Cecile, pers. comm., 1987) have been recovered from the limestone.

These volcanics are part of an early to middle Paleozoic basaltic province discontinuously exposed along the margin of the Selwyn Basin. Most occurrences are only briefly described (references in Cecile, 1982) but are probably alkalic basalts, subaqueously deposited. Analyses of flows sampled near the Dempster Highway in the Ogilvie Mountains (Roots, in press) are compositionally similar to those of the Middle Ordovician to Devonian volcanic Marmot Formation (Goodfellow *et al.*, 1980; Cecile, 1982) in the Misty Creek Embayment, 500 km east of Mount Harper (Fig. 2.2).

The Road River Formation overlies the Hyland Group and the volcanics. It is poorly exposed in a narrow strip 10 km south of Mount Harper. A basal black chert layer 5 to 25 m thick is succeeded by about 170 m of black argillite and siltstone, but the unit is structurally thinned and thickened by thrusts and isoclinal folds. Graptolites, lingulids and trilobite fragments of Late Ordovician and Late Silurian age were collected (Green, 1972, p.49; R. I. Thompson, unpubl. data) southeast of Mount Harper.

The **shelf succession** comprises three major sedimentary divisions: the Wernecke Supergroup, the Pinguicula Group (provisional name) and Harper Group (new name). All three are, at various places along the edge of the Coal Creek Inlier, unconformably overlain by CD₆ dolostone.

The oldest exposed rocks are correlated with the Quartet Group of the Wernecke Supergroup (Delaney, 1981, in map-areas 105 C and D on Fig. 2.2). In the centre and along the northern margin of Coal Creek Inlier it is mostly meta-sandstone, but southeast and east of Mount Harper, micaceous, thinly bedded siltstone and chloritic phyllite predominate. It is unconformably overlain by a laminated to thin-bedded orange- and white-weathering dolostone correlated with the Gillespie Lake Group of the Wernecke Supergroup (Delaney, 1981).

Within Wernecke units are intrusive breccia bodies up to 1 km across that are similar to those in the Wernecke Mountains described by Archer and Schmidt (1977), and Bell and Delaney (1977). Most occur in the northeast part of the Coal Creek Inlier, where their stratigraphic relations suggest emplacement at about 1200 Ma (Bell, 1986) during the tectonic event that preceded deposition of the Pinguicula Group.

Unconformably overlying the Wernecke Supergroup is the Pinguicula Group, provisionally correlated with a stratigraphically analagous succession (Eisbacher, 1981) near Pinguicula Lake (map-area 106 C on Fig. 2.2). At its base is an assemblage of fine- and medium-grained siliciclastic sediments with carbonate reefs and breccia that vary laterally and are several kilometres thick across the northern part of the Coal Creek Inlier. It is informally known as the **Rackla unit** (R.I. Thompson, pers. comm., 1985) after the 1200 Ma tectonic event that it may represent. North of Mount Harper it consists of 350 m of red mudstone with metre-thick beds of jasper pebble conglomerate, overlain by thinly bedded, locally stromatolitic, dolomitic limestone.

Massive, poorly bedded dolostone, with 1-10 m thick black shale intervals, overlies the Rackla unit. Its thickness increases westward across the Ogilvies; it is about 2500 m thick north of Mount Harper. The most prominent layer is a 600 m thick, medium grey, craggy weathering dolostone, here called the **Fifteenmile formation**. Along the north side of the Volcanic Complex is a white, smooth-weathering dolostone biostrome with large (40 cm diameter) domal stromatolites, the **Cryptalgal member** of the Fifteenmile Formation.

East of Mount Harper the Harper Group overlies the Pinguicula Group, and overlaps the Wernecke Supergroup. Two clastic sedimentary successions, here called the Lower Harper and Upper Harper groups, are described in Chapter 3. The Mount Harper Volcanic Complex lies between them.

The shelf succession is capped by thickly bedded, medium crystalline, light grey dolostone, called the 'CD_b' formation (an informal unit of Norris, 1976, north of Dawson map-area). It is at least 250 m thick where it unconformably overlies the Volcanic Complex and Upper Harper Group. Archaeocyathid fragments found 25 km west of Mount Harper (E. Mercier and R.I. Thompson, pers. comm., 1982) are probably of Early Cambrian age (e.g. Read, 1978).

Black shale of the Road River Formation conformably overlies CD_b Formation southwest of the Volcanic Complex. This unit indicates that the Selwyn Basin transgressed over the Mackenzie Platform in Devonian time.

REGIONAL STRATIGRAPHIC CORRELATIONS

Proterozoic shelf successions exposed in the Foreland and adjacent Omineca belts along the Cordillera are remarkably consistent in their lithologic progression, as indicated by generalized sections from the northern Cordillera and in southeastern British Columbia (Fig. 2.5). All sections contain two major unconformities that were formerly interpreted as representing orogenies and are now considered generally related to rifting. The older 'event', called the Rackla (Yukon) or East Kootenay (southeastern B.C.) occurred about 1200 Ma, and the younger, tentatively referred to as the Hayhook (Yukon) or Goat River (southeastern B.C.) is broadly 750 Ma. Evidence and correlations are summarized in Yeo *et al.* (1978), Young *et al.* (1979), Eisbacher (1981) and McMechan and Price (1982).

In southeastern British Columbia (section F on Fig. 2.5), the two unconformities are superimposed because sedimentary rocks that represent the 1200-750 Ma interval are absent (except locally, such as the Dutch Creek Fm.; Root, 1987). Uplift and erosion of the older rocks during the 750 Ma extension event is indicated by syndepositional extension faults (Lis and Price, 1976; Root, 1987). Faults related to the Harper Group are described in Chapter 3.

STRUCTURE OF THE COAL CREEK INLIER

Both shelf and off-shelf successions have been thrust northwards, and are detached from their original basement. The Tombstone thrust, recently recognized in the eastern part of Dawson map-area (R.I. Thompson, pers. comm. 1986), juxtaposes Late Paleozoic units whose facies relationships require at least 100 km relative northward displacement (Anderson, 1987). The minimum age of tectonic contraction is indicated by undeformed contacts of Late Cretaceous syenite. These plutons cut folded rocks of the off-shelf succession and some thrust faults, although they too may be translated northward in the largest thrust panels.

Fig. 2.5. Composite stratigraphic sections for the northern Cordillera, modified from Eisbacher (1981) and Young (1982) with additional information. Locations of sections are shown on Fig. 2.1, and references for lithologic units (circled numbers), fossil localities [F] and isotopic dates (system indicated) are listed in Appendix B. Lithologic symbols are those of Figure 2.4.

The off-shelf succession is more strongly deformed than the shelf succession. Older-over-younger stratigraphic relationships indicate that multiple thrust faults separate at least five panels, each containing a distinct lithologic package (commonly upper Hyland Group and volcanic rocks) that were previously tucked into upright, isoclinal folds. South of Mount Harper the Hyland Group carbonate outlines these early folds. Because large areas of deformed strata expose only a limited stratigraphic thickness, and because basement rocks are not exposed by folds or thrusts, this deformation is considered thin-skinned (Thompson and Eisbacher, 1984).

In contrast, the shelf succession is tilted southward in imbricate stratigraphic packages several kilometres thick, and overlain to west and north of the Coal Creek Inlier by gently dipping Paleozoic rocks. High-angle reverse faults with minor displacement occur in the Proterozoic rocks (Thompson and Roots, 1982), and low-angle thrusts are inferred at the base of the Paleozoic succession (Chapter 5). Together they are considered to represent a system of thrusts with ramps (through Proterozoic rocks) and flats (along incompetent horizons). The continuity of the regional stratigraphy and the lack of exposed crystalline basement imply that most Mesozoic faults flatten to a sole detachment, at least beneath the southern part of the Coal Creek inlier.

The deep structure of the Coal Creek inlier remains speculative; two models are shown in Figure 2.6. The first, from a structural interpretation of the Mackenzie Mountains by Gordey (1983), involves a basement step, possibly a relic of earlier extension faulting, that provided a ramp over which eastward-translated strata were uplifted (Fig. 2.6,a). The eastward displacement of the thick Rapitan Group has resulted in the Redstone Plateau. Steep faults along the eastern side of Redstone Plateau reveal little of the lateral shortening (estimated 30%

Fig. 2.6 (top) True scale cross-section for the Mackenzie Mountains (location on Fig. 2.2) from Gordy (1983). This section was balanced to conserve bed lengths. Bottom) Schematic cross-section of the Ogilvie Mountains (location on Fig. 2.3) with interpretation discussed in Chapter 2. North-dipping faults are Late Proterozoic listric normal faults (R.I. Thompson, pers. comm.).

by Gordey, 1983) within the Proterozoic shelf succession, which is obscured by Paleozoic rocks and shallow listric thrust faults.

In the Coal Creek inlier, no single unit is similarly elevated in the manner of the Rapitan Group of the Mackenzie Mountains. In contrast, the Wernecke Supergroup rocks, whose higher metamorphic grade may indicate substantial burial, were previously uplifted in several parts of the Inlier, because they are unconformably overlain by the Harper Group and CD₆ formation. In at least one place (5 km east of Mount Harper) these faults were reactivated during the Mesozoic contraction. An alternative model (Fig. 2.6,b) shows Mesozoic uplift as a result of stacking of strata on a series of splays from a major sole thrust.

Chapter Three

GEOLOGIC SETTING:

SEDIMENTARY ROCKS AND STRUCTURE OF THE HARPER GROUP

DEFINITION AND DISTRIBUTION

The Harper Group consists of clastic sedimentary and volcanic rocks that occupy the stratigraphic interval between the Pinguicula Group and CD_b formation (Fig. 2.4). The unit is exposed along the southern and western edges of the Coal Creek Inlier over an east-to-west strike length of 45 km (Fig. 2.3). The greatest thickness and lithological variety, however, occur in the southwest and coincide with the Mount Harper Volcanic Complex. In this area sedimentary successions that conformably underlie and overlie the Volcanic Complex are respectively called the Lower Harper and Upper Harper groups. Both units include thick conglomerate beds, but only those of the Upper Harper Group contain volcanic clasts.

A map and representative sections of the sedimentary rocks (Fig. 3.1, in pocket) indicate rapid changes in thickness and lithology near Mount Harper. The Lower Harper Group, up to 160 m thick where it underlies the volcanics, consists of thickly bedded coarse conglomerate with a pocket of well-stratified facies at the top (Plate 1,a) directly north of Mount Harper. The Upper Harper Group, at least 500 m thick northwest of the Volcanic Complex, contains volcanic and dolomitic pebbly mudstone, siltstone and argillite, with carbonate build-ups and some pyroclastic beds. Erosional outliers of sedimentary rocks that overlie the Volcanic Complex are informally referred to as Saddle (135 m thick), Baldie (200 m; Fig. 3.2) and Snowcap (140 m; Plate 1,b). They are lithologically correlated with the basal Upper Harper Group.

The emphasis in this chapter is on the sedimentary sequence and structural setting of the Harper Group. A more detailed analysis of the

Caption for Plate 1 (facing)

PLATE I - Lower Harper Group, Upper Harper Group

a) View eastward of well-stratified conglomerate and sandstone conformably overlain by massive basalt of Member A, at the base of the Mount Harper Volcanic Complex. White section of measuring stick, upper centre, is 0.5 m long. North face, 1.1 km NE of Mount Harper; 5000' at NTS 550 730.

b) Aerial view westward of Snowcap outlier (above dashed line). Light coloured resistant outcrops are dolostone of Upper Harper Group, enclosed by well-bedded volcanic-derived siltstone. Beneath the outlier, hydroclastic breccia (Member E) up to 60 m thick unconformably overlies pillow breccia of Member B.

c) Well-stratified facies of Lower Harper Group north of Mount Harper consists of normally graded, subrounded dolostone and quartzite pebbles interstratified with sandstone and grit. The densely packed pebble fabric suggests a braided stream deposit. White line scale is 10 cm long. From 5 m below the top of section 1 (Fig. 3.1); 5000' at NTS 550 730.

d) Poorly sorted, coarsely stratified Lower Harper Group at northwest end of the Complex contains subrounded dolostone cobbles. The 10 cm thick sandy layer in centre covers a pavement of tightly packed clasts (at bottom) and its top is deformed by large clasts of the overlying flow. The imbrication of cobbles suggests deposition from right to left. From 5 m below top of section 3 (Fig. 3.1); 7 km NW of Mount Harper, 4600' at NTS 485 768.

e) View eastward of the north-facing cliff of Baldie outlier, a superb exposure of Upper Harper Group. Massive flows of Member B are unconformably overlain by 1-3 m thick beds of conglomerate succeeded by well-stratified sandstone layers. Smooth ridge nose in centre is section 6 (Fig 3.1). Repetition of the thin bedded mid-section to the skyline at right indicates a reverse fault that is a splay from the southeast trending fault set. Plate 1(f) is taken from "x", near a sill of Member F.

f) Westward view of Baldie outlier from "x" in Plate 1 (e) shows cross-cutting relations at base of Member F (within dashed lines) interpreted as a sill, although its top is locally indistinct. Note interdigitated paraconglomerate at right, interpreted as lobate debris flows.

g) Laminated epiclastic siltstone, with basalt lapillistone and reworked tuff of the Upper Harper Group in Saddle outlier. Convolutions in dolomitic mudstone layer suggest slumping (direction unknown from this loose slab). Note delicate stacking of lapilli (top centre), indicating low energy deposition. From 1 km SSE of Mount Harper; 5200' at NTS 542 716.

h) Sorted basaltic gravel layer at the base of Upper Harper Group. The angular, non-vesicular clasts were probably derived from eroded flows of Members A or B. From east end of Baldie outlier; 5500', NTS 485 761.

* Location reference consists of distance and direction to geographic feature, station or sample numbers; elevation and National Topographic System coordinates (prefix 7W EB-) correspond to grid on 1:50,000 scale topographic maps 116 B/12 and 116 C/9.

Fig. 3.2. View looking south at the northern edge of the Volcanic Complex, with the crest of Baldie outlier at top centre. The westernmost exposed Lower Harper Group conglomerate is at the tarn left of centre; the ridge above the tarn is composed of volcanic members A and B, which were systematically sampled (section 118-). The North Harper Fault, exposed on the ridge crest at right, underlies the tarn and downfaults the Upper Harper Group (middle ground) and the northernmost exposure of the volcanic rocks (foreground). Field of view is 3.5 km; topographic relief above the tarns: 500 m (from 4200' to 5800'). Sketch from photographs.

Harper Group sedimentary rocks is contained in Mustard (Ph.D. thesis in prep., Carleton University). Stratigraphic relationships between the volcanic and sedimentary rocks are well exposed at Mount Harper and recognition of the large scale structures allows the history of the Complex to be interpreted in a tectonic, as well as stratigraphic, context. The sedimentary rocks also provide a link through regional correlations with other Late Proterozoic clastic and volcanic units. Thus the history of faulting, sedimentation and volcanism can be applied to less well-exposed occurrences in the Late Proterozoic strata in the Cordillera

THE LOWER HARPER GROUP

LITHOLOGY

— To the east and north of the Volcanic Complex, the Lower Harper Group is dominated by poorly sorted conglomerate (sections 1-3 on Fig. 3.1). North of Mount Harper a second facies of moderately to well-sorted, finer grained conglomerate with inter-stratified sandstone and mudstone occurs at the top of the unit.

The conglomerate consists of 0.5-3.5 m thick beds which range from paraconglomerate to orthoconglomerate (Plate 1,c). Most are framework-supported and massive, except for crude normal-to-reverse grading of large clasts, in a poorly sorted matrix of gritty dolomitic wackestone (Plate 1,d). In many, a subtle internal stratification results from lenticular concentrations of boulders or the sub-parallel orientation of elongated clasts. Contacts between beds are generally conformable and can be traced hundreds of metres, indicating that the beds were emplaced in discrete events.

The conglomerate beds of the Lower Harper Group contain angular to well-rounded clasts of pebble to large cobble size, boulders up to 1.5

m in diameter, and isolated blocks up to 40 m in diameter (R.I. Thompson, pers. comm., 1985). All clasts were eroded from ~~rock units~~ whose uneroded equivalents are presently exposed nearby; no high grade metamorphic or plutonic debris has been found. Clast lithology represents a crude inverse stratigraphy: lowest beds are composed of Pinguicula fragments, while those from Wernecke Supergroup are increasingly abundant in some of the higher beds. This shift in provenance suggests progressive erosion of an uplifted block of the Proterozoic shelf succession. Some of the conglomerate beds are interpreted as debris flows because they are poorly sorted, internally unstratified, and do not have scoured bases.

The sorted, stratified facies includes pebble conglomerate (Plate 1,c), fine-grained sandstone and hematitic siltstone in alternating beds 10-50 cm thick (Plate 1,a). The well-rounded clasts, rare tabular cross-bedding and oversize lag pebbles (Fig. 3.3) suggest a submarine fan or fluvial-deltaic deposit. Uppermost layers contain angular volcanic chips and thin layers of microvesicular basaltic granules. Pebble-rich mudstone beds are disharmonically folded by overriding basaltic flows (Fig. 3.4) at the base of the Complex. Both features indicate volcanism concurrent with deposition.

DEPOSITIONAL ENVIRONMENT

The presence of well-rounded clasts, poorly sorted in very thick, massive layers implies at least a two-stage erosion cycle. The smooth clasts were shaped in a high energy stream environment. Abundant mud matrix and poor sorting of the deposits indicate mud flows that probably mobilized cobbles and boulders of the stream bed.

The textural immaturity of the matrix, enormous clasts, and rapid thickness changes (tapering from more than 1000 m to less than 150 m in under 5 km) imply rapid erosion of a topographically elevated, nearby

Fig. 3.3. Sandstone layer with angular dolostone clasts and siltstone (unpatterned), containing a well-rounded quartz pebble. From top of section 3 (Fig. 3.1); sketch from photograph.

Fig. 3.4. Upper contact of the Lower Harper Group. Laminated siltstone (lines) containing basalt and syn-sedimentary deformed clasts is disrupted and overlain by a basalt flow (dashes) with sedimentary clasts and isolated pillow breccia. From the ridge northeast of Mount Harper (NTS 545 731; elev. 5350').

source. At the southern edge of the Lower Harper Group, where the conglomerate is thickest, at least six blocks of Fifteenmile dolostone, up to 40 m high and 10 m long, lie along a single bedding plane. They are interpreted as a landslide deposit (P. Mustard, pers. comm., 1985). In addition, large angular dolostone blocks chaotically packed together along the southern edge of Lower Harper Group are interpreted as paleotalus (P. Mustard, pers. comm., 1986).

The Lower Harper Group was derived from uplifted carbonate shelf strata. Large clasts probably lay in high gradient valleys that were periodically flushed by debris torrents and mud slurries. The conglomerate layers extended from an eroding area possibly uplifted by normal faulting (Fig. 3.5,a). Landslide blocks were issued from the fault scarp, and the sorted, stratified facies indicates streams flowed over the debris fans. Although these sediments were deposited subaerially, they were submerged before or during initial volcanism. The absence of conglomerate interbedded with lava flows indicates rapid volcanic accumulation or abrupt shut-off of the sediment supply.

STRUCTURAL SETTING

The Lower Harper Group appears to be wedge-shaped because it is progressively thinner northwestward from Mount Harper. Its present form is broadly synclinal, indicated by the inward-dipping upper contact at both ends of the ridge that extends north and south from Mount Harper. It is not clear whether the syncline represents folding during Mesozoic regional contraction, or developed earlier in response to loading or extension.

The southern boundary of the Lower Harper Group, where thickest sections (up to 1050 m; P.S. Mustard, pers. comm., 1986) and coarsest clasts occur, is a west-trending fault zone. Two fault traces, representing sub-parallel faults of different ages, are exposed (Fig. 3.6a).

Fig. 3.5. Schematic depositional block diagrams. a) The Lower Harper Group consists of subaerial deposits from the uplifted south side of the Early South Harper Fault; b) the Upper Harper Group contains clastic derived from uplifted Pinguicula carbonate (top right) and the partly submerged Volcanic Complex (at left).

Fig. 3.6. a) View southeastward from Mount Harper, showing Late South Harper Fault and reactivated trace of Early South Harper Fault. An erosional remnant of the Volcanic Complex (outlier) indicates the top contact of the Lower Harper Group. Width of view: 1.5 km. Sketch from photo.

b) Sequence of fault movement in the South Harper Fault zone. View is looking east, as in the sketch above.

The southernmost, here called Late South Harper Fault, drops the south side of the Volcanic Complex. The Early South Harper Fault, whose reactivated trace is about 200 m north of the Late South Harper Fault, is the Late Proterozoic boundary fault implicated in deposition of the Lower Harper Group.

The Early South Harper Fault separates areas with different stratigraphic successions (Fig. 3.6,b). Against its northern side is the greatest known thickness of Lower Harper Group, overlying southwest-tilted strata of the Pinguicula Group, which in turn unconformably covers the Wernecke Supergroup. On the south side Wernecke Supergroup rocks are exposed on the ridge crest, and further south, across the Late South Harper Fault, CD_b dolostone directly overlies the Wernecke Supergroup. This stratigraphic discontinuity, with mid-Proterozoic to early Paleozoic strata missing, is important; only thin, isolated remnants of Pinguicula strata remain south of the fault zone, and most of this unit must have been removed before early Paleozoic time. The uplifted southern area (bounded by the Early South Harper Fault to the north) is hereafter called the **South Block**.

The age of the South Harper faults is constrained between that of truncated older rock units, and the overlap of younger formations. Initial movement on the Early South Harper Fault must have postdated deposition of the Pinguicula Group, and pre-dated volcanic extrusions, which straddle the fault. It was therefore active during deposition of the Lower Harper Group. The Late South Harper Fault cuts CD_b dolostone and is therefore younger.

No piercing points are known that can be used to determine lateral movement on the Early South Harper Fault, although its vertical displacement can be estimated by the apparent dislocation of the Wernecke Supergroup. On the north side of the fault, this unit is overlain by

Pinguicula strata about 4 km thick, but this succession has been removed from the South Block. The Early South Harper Fault thus exhibits substantial north-side-down displacement.

Sedimentological evidence demonstrates that the scarp formed by the Early South Harper Fault resulted in erosion of Pinguicula from the south side, and northward deposition of the Lower Harper Group to the north. South-to-north sediment transport is indicated by imbricated clasts in the conglomerate (P. Mustard, pers. comm., 1986) as well as crossbeds and pebble trains within the sorted, stratified facies. Landslide blocks, large clasts and paleo-talus indicate nearby topographic relief, and the aggregate thickness of the conglomerate suggests incremental uplift that could have resulted from protracted faulting. Clasts of the Pinguicula Group are unlikely to have been derived from nearby to the north, because the Pinguicula appears intact throughout the Coal Creek Inlier and beneath the Lower Harper Group. The only potential source area is south of Early South Harper Fault, from where Pinguicula was mostly removed before Mount Harper volcanism. The wedge shape of the Lower Harper Group therefore resulted from infilling the half-graben created by the north-side-down Early South Harper Fault and extensive erosion of the adjacent South Block.

THE UPPER HARPER GROUP

LITHOLOGY

Outliers within the Mount Harper Complex (sections 4-11, Fig. 3.1) consist of paraconglomerate and fine-grained clastics with intercalated carbonate and volcanoclastic beds. Locally the paraconglomerate unconformably overlies lava flows of the Volcanic Complex, but commonly the contact is gradational from volcanic breccia. In more complete sections of the Upper Harper Group that overlie the western and northern extremities of the Complex, the clastic succession is overlain by maroon and beige thin-bedded dolostone and black siltstone (sections 12 and 13, Fig. 3.1).

Paraconglomerate comprises poorly sorted pebble, sand and mud mixtures (also called diamictite). In contrast to conglomerate of the Lower Harper Group, this rock type has higher matrix content, as well as numerous subangular and volcanic clasts. Basaltic clasts predominate in most beds adjacent to and overlying the Complex; rhyolitic cobbles and pebbles are prominent in the Saddle outlier. The matrix is locally green or maroon, apparently derived from breakdown of chloritized and hematized volcanic rocks.

Beds are 0.2-2.0 m thick, with sharply defined lower contacts and fine-grained, slightly reworked tops. Most are internally massive; some have concentrations of boulders near the base. Disconformable contacts between beds are locally useful marker horizons, and some thin pyroclastic and carbonate interbeds, typically deformed by load casts and slump folds (Fig. 3.7,a), assist in correlating the outliers.

The paraconglomerate beds are interpreted as subaqueous mudflows because they have not eroded their bases and are separated by fine-laminated interbeds.

Siltstone-argillite couplets, in sequences 0.5 to about 5 m thick, consist of 0.2-3 cm beds of laminated green siltstone and brown mudstone. Outsize pebbles, mostly basaltic, but also of dolostone, are irregularly distributed in the thinly laminated beds. Although some clasts resemble volcanic bombs or glacial dropstones, underlying laminae are not truncated (Fig. 3.7,c) and deformation surrounding them is considered to result from differential compaction. These deposits are attributed to winnowing of gravel-sized clasts from gravity flows (P. Mustard, pers. comm., 1985). The beds are interpreted as distal turbidites because of their erosional bases and planar laminated tops.

Fig. 3.7. Sedimentary structures in the Upper Harper Group:

- a) Syn-sedimentary deformation of siltstone and mudstone, at the base of Snowcap outlier.
- b) Three beds of reversely graded sharpclast mudstone. Black clasts are basalts; others are greenish dolostone. Beds are 10 cm thick at the base of the Upper Harper Group conformably overlying hematized vesicular flows. From Mount Harper summit.
- c) Oversize basalt and dolostone clasts in laminated mudstone and graded siltstone. Note that deformation around pebbles resulted from diagenetic compaction; layers are not disrupted as beneath dropstones or pyroclasts. From the middle of section 6 (Fig. 3.1) on Baldie outlier (NTS 469 761).
- d) Soft-sediment deformation and scouring of laminated mudstone by overlying dolomitic sandstone; paleo-current from the left. Black clasts are basalt. From the southwest edge of Baldie outlier (NTS 462 754, elevation 4950').

Volcaniclastic rocks including lapillistone (Plate 1,g), lapillituff, fine-grained basaltic breccia (Plate 1,h) and tuffaceous siltstone constitute 1-2 m thick beds in the outliers atop the Volcanic Complex and in the basal unit to the west. They are subaqueous deposits from pyroclastic eruptions and erosion of unconsolidated volcanic slopes.

Precipitated carbonate constitutes two rock units. Near the middle of the Upper Harper Group in West Trib laterally continuous, 40 m thick, orange- and red-weathering silty dolostone, comprises 10-40 cm thick beds with thin black shale separations. The other unit consists of 5-15 m thick lenses of beige and grey limestone that occur at several stratigraphic levels. Low-angle faults that repeat these distinctive rocks at the western edge of the Volcanic Complex are thrusts within less competent strata of the Upper Harper Group (Chapter 5).

Black siltstone with dark sandstone interbeds and lenses of dolostone cobbles and concretions constitute the uppermost unit. It appears locally to be hundreds of metres thick, but may be tectonically thickened by thrust faults.

DEPOSITIONAL ENVIRONMENT

The general stratigraphy of the Upper Harper Group indicates a basin of increasing depth and decreasing depositional energy and clastic sediment supply. Basal paraconglomerate flows contain locally derived volcanic clasts and dolostone from older regional units, which indicates both subsidence of the volcanic structure and uplift of the shelf carbonate. The additional sedimentary succession at the base of sections 6 and 7 (Fig. 3.1) on Baldie outlier indicates that the Upper Harper Group basin probably transgressed over the Volcanic Complex (Fig. 3.5,b). Initial debris flows were succeeded upward by better-stratified sediments, deep-water carbonate and carbonaceous deposits that attest to poor circulation and sediment starvation. Volcanic eruptions were

prevalent during early stages. Northwest of the Volcanic Complex, cyclic sandstone-siltstone successions contain indicators of predominantly north-to-south paleocurrents (P. Mustard, pers. comm., 1986). Turbidites and occasional slumps constituted the principal deposits on the anoxic floor of a large basin.

STRUCTURAL SETTING

At the northwestern end of the Volcanic Complex, the Upper Harper Group rests directly on Fifteenmile dolostone of the Pinguicula Group. The contact is a low angle unconformity (P. Mustard, pers. comm., 1986). The same strata in outliers atop 1000 m of volcanic rocks indicate subsidence of the Complex, and imply that the original half-graben widened into a basin. Although the southern limit of the Upper Harper Group sedimentation is not recognized, the presence of subaerially erupted volcanic rocks south of Lower West Trib (Chapter 4), and CD_b directly overlying older units (Chapter 5), suggests that the Harper Basin was bounded by high ground along the line of the Early South Harper Fault. Silicic clasts in beds of Saddle outlier indicate subaerial erosion and the nearest source area appears to have been further south.

The Upper Harper Group is concordantly overlain by CD_b formation on the west side of the Volcanic Complex. Although not directly proven, the contact is interpreted as a thrust, which reconciles abrupt changes in thickness (sections 12-16 in Fig. 3.1), as well as the variety of units beneath the dolostone. For example, the greater thickness preserved near the North Harper Fault is interpreted as a pocket preserved on the downthrown side of this Mesozoic fault, and later truncated by a flat-lying thrust within the shaly beds beneath the CD_b dolostone (Chapter 5).

In summary, the Harper Group records two stages of subsidence. Initial movement on the Early South Harper Fault produced a half-graben, into which subaerial clastic rocks (Lower Harper Group) were deposited. These are overlain by Mount Harper volcanic flows which overstepped the fault onto the south side of the half-graben. Continued subsidence of the northern part of the Complex formed the Harper Basin, indicated by clastic and carbonate transgression (Upper Harper Group).

EXTENSIONAL TECTONICS AND THE HARPER GROUP

Carbonate and clastic platform deposits are common on continental shelves that are trailing edges left by earlier separation events. They constitute 'passive margin' successions which overlap and bury extended structures that represent initial stages of continental rifting. The early-formed structures have been investigated by geophysics and drilling, particularly beneath the continental shelves on both sides of the Atlantic Ocean (e.g. Burk and Drake, 1974; de Charpal *et al.*, 1978; de Graciansky *et al.*, 1985).

Current structural models for continental rifts are based upon the Phanerozoic examples, including young margins (the Red Sea, e.g. Bonatti, 1985) and active extension on continents (the Basin and Range province of the western United States, e.g. Wernicke, 1985; and the East African rift, e.g. Bosworth, 1985). Some modern and ancient examples are now recognized as asymmetric basins, formed by steep boundary faults. For lateral extension to take place, these boundary faults must flatten with depth. They are known as **detachment faults**. Lateral extension may result in domino-like rotation of overlying crustal blocks, while at deeper levels extension occurs by thinning of the lithosphere (Mackenzie, 1978). Detachment faults that underlie extending continental crust strike parallel to the rift axis, but commonly alternate in dip direction (Bally, 1982; Bosworth, 1985).

In the current model, the extending lithosphere is divided into an **upper plate** and **lower plate** by a major detachment fault (Fig. 3.8,a). These plates have contrasting internal structural geometry and overlying sedimentation regimes, and can therefore be recognized after rifting by their different stratigraphy, providing an indication of the dip direction of the previously active detachment fault. Where ancient rift successions are uplifted and eroded, the asymmetry of basement rocks remains: the lower plate margin is typically highly structured, consisting of tilted blocks separated by rotational normal faults, while the upper plate is comparatively intact, broken by a few steep faults (Lister *et al.*, 1986).

Although other models for graben formation such as 'piano-key' tectonics have been proposed (Eisbacher, 1985; Yeo, 1980), the shelf succession of the Ogilvie Mountains appears to be an upper plate margin. The detachment fault is therefore considered to underlie this region. The lower plate lay to the south; it has been rifted away, or may underlie the off-shelf succession as shown in Figure 3.8,d.

The Early South Harper Fault is steeply dipping with north-side-down movement (Fig. 3.8,c). It is unlikely that this fault was a major detachment because major elements of the stratigraphy can be traced across it. Because much older rocks are exposed in the South Block it is analogous to the 'outer basement high' of Schuepbach and Vail (1980).

CORRELATION WITH THE WINDERMERE SUPERGROUP

The Windermere Supergroup is the inclusive name for the Late Proterozoic clastic and volcanic assemblage in the southeastern Canadian Cordillera (Gabrielse, 1972; Stewart, 1972; F.G. Young *et al.*, 1973). It succeeds Middle and Late Proterozoic strata, commonly at a prominent

Fig. 3.8. Extension mechanisms applied to the Harper Group:

- a) Generalized elements of a dipping detachment fault model from Wernicke (1985). Upper plate and Lower plate are designations of Lister et al. (1986);
- b) sketch map of the Mount Harper area;
- c) initial half-graben setting (Lower Harper Group);
- d) structural relations in mid-Paleozoic time. The Harper Basin has been covered by stable platform carbonate while extension may have continued beneath the Selwyn Basin.

unconformity (the Goat River/Hayhook event; Fig. 2.6), and lies beneath Paleozoic quartzite, carbonate and shale. Additional elastic successions in Alaska, Idaho, Utah and California have been correlated with the Windermere on the basis of general stratigraphic position. The Harper Group, which overlies Late Proterozoic strata and is succeeded by lower Paleozoic CD_b formation is the Windermere Supergroup-equivalent unit in the southern Ogilvie Mountains.

Although the Windermere Supergroup term has been extended to encompass widely separated elastic deposits along the ancient western margin of North America, the unit appears to have been deposited in separated sedimentary basins. Deposition was at least partly controlled by normal faults, which are now deduced from stratigraphic discontinuities (e.g. Lis and Price, 1976; Root, 1987). Seismic and regional stratigraphic studies have outlined subsurface grabens in southwestern Alberta (Kanasewich *et al.*, 1969; McMechan and Price, 1982) and depositional paleo-highs in southeastern British Columbia (Reesor, 1973; Root, 1987).

In Yukon, two geographically distinct successions are correlated with the Windermere Supergroup (Fig. 3.9). The Hyland Group (Grit Unit; Gabrielse, 1972) occurs beneath Selwyn Basin strata from the Alaska-Yukon border to northeastern British Columbia. The other succession, which overlies Middle and Late Proterozoic shelf strata to the north and east of the Selwyn Basin, includes the Coates Lake and Rapitan groups (Fig. 2.5) as well as some younger formations (Eisbacher, 1981). The physical connection of this succession to the Hyland Group is obscured by Paleozoic strata and contraction faults; a rapid facies transition is reported by Eisbacher (1981) south of Pinguicula Lake (Fig. 2.2). West of the Mount Harper Complex overlying CD_b appears continuous with the middle carbonate of the Hyland Group.

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Glaciogenic deposits and iron formation are widely distributed in Windermere-equivalent successions (Yeo, 1980; Eisbacher, 1981; Young, 1982; Miller, 1985; Walker *et al.*, 1986). At least two glaciations are represented in the Windermere successions in Utah (Crittenden *et al.*, 1983), one of which occurred about 650 Ma (Narbonne *et al.*, 1985). Although locally distributed, their regional significance (e.g. Hambrey and Harland, 1982) may also be used to correlate Windermere Supergroup deposits. No unequivocal evidence has been found to indicate glacial influence among Lower Harper Group deposits.

The Coates Lake Group, which underlies the Rapitan Group in the Mackenzie Mountains, occurs in the analogous stratigraphic position to the Harper Group. Three formations constitute the Coates Lake Group: Thundercloud (volcanic conglomerate, dolostone and quartz sandstone), Redstone (evaporites and carbonate conglomerate) and Coppercap (stromatolitic limestone and interbedded turbidites), in sections up to 500 m thick (Jefferson and Ruelle, 1986). The Coates Lake Group unconformably overlies biostromal carbonate of the Little Dal Group, which includes local basaltic flows. It differs from the Harper Group by containing abundant evaporite beds and thin (maximum 30 m) conglomerate units, although paleo-fault scarps and half-graben structures are recognized (C. Jefferson, G. Yeo; pers. comm., 1985).

IGNEOUS ROCKS OF THE WINDERMERE SUPERGROUP

Local mafic intrusions and flows of Late Proterozoic age are associated with Windermere Supergroup-equivalent strata along the length of the western margin of cratonic North America, from Alaska (also in the Brooks Range; Churkin *et al.*, 1980) to Death Valley, California (Hammond, 1983). The most voluminous deposits are the Irene and Huckleberry formations that span the British Columbia-Washington border (Fig. 3.9). This unit is up to 1500 m thick and contains amygdaloidal flows, pillows and breccias (Daly, 1912; Little, 1960). Isolated basalt flows occur in

Fig. 3.9. Distribution of the Windermere Supergroup (pattered) in the Canadian Cordillera, showing igneous rock localities (black) and Upper Proterozoic dates (in millions of years; Table 3.1). Base map is modified from Eisbacher (1981), Evenchick (1985) and Pell and Simony (1987).

the Little Dal Formation of the Mackenzie Mountains and in the Tindir Group, but the Mount Harper Complex appears to be the best exposed northern occurrence.

Felsic volcanics are less common. Flows occur in the Mount Harper Complex, in structural isliers containing Late Proterozoic strata in the Monashee Mountains (south-central British Columbia; R. Scammell, pers. comm., 1987) and are indicated by rhyolite cobbles in conglomerate units near Monkman Pass (R. Parrish, pers. comm., 1983). A layer of possible alkalic tuff inferred to be 750 Ma occurs in the Monashee structural Complex (Scammell, 1986). In the Qmineca Belt, dyke swarms and amphibolite layers are common in the Horsethief Creek and Misinchinka groups. They are strongly altered tholeiites, interpreted as sill-sediment complexes (Einsele, 1985) and are evidence for regional extension (Evenchick, 1985; Scammell, 1986; Sevigny, 1987).

Isotopic dating helps to determine whether magmatism was simultaneous or protracted. Only two extrusive samples have been dated with reasonable certainty. A basalt flow from northeastern Washington state was dated using the Nd-Sm method on plagioclase and pyroxene mineral separates (762 ± 44 Ma; Devlin, 1986). Zircons extracted from a silicic flow in the Mount Harper Complex are $750 \pm 26/-18$ Ma (Chapter 6). Both volcanic units overlie coarse conglomerate at the base of their respective Windermere successions. Although the intrusive rocks of the Mount Harper Complex are less altered than volcanic rocks and could be dated more precisely, the two cannot be unambiguously correlated. Table 3.1 shows these and other similar dates from igneous rocks in the eastern belts of the Canadian Cordillera. The diabase intrusions and a quartz diorite plug in the Mackenzie Mountains are, within the limits of error, the same age as the Mount Harper volcanic rocks.

DATE (Ma)	Method	Source	Rock unit, relationships
728 +8/-7	U-Pb (zr)	Evenchick <u>et al.</u> , 1984	leucogranite non-conformably overlain by U. Prot. metaseds
741 +22/-20	U-Pb (zr)	R. Parrish, pers. comm.	Malton gneiss faulted against U. Prot. metaseds
746 +40/-39	U-Pb (zr)	Scammell and Brown, in prep.	Mt. Copeland nepheline syenite Monashee Mtns. B.C.
751 +26/-18	U-Pb (zr)	this thesis (Chap. 6)	Rhyolite flow, Mount Harper Complex, western Yukon
750 ±60	Rb-Sr (wr)	Duncan, 1984	Thor-Odin gneiss core Monashee Mtns. B.C.
762 ±44	Nd-Sm (pl+px)	Devlin, 1986	Huckleberry Volcanics, NE Washington state
766 ±24. 769 ±27	Rb-Sr (wr)	Armstrong <u>et al.</u> , 1982	Diabase Intrusions in mid-Prot Tsezotene Fm. Mackenzie Mtns.
778 +3/-2	U-Pb (zr)	Jefferson and Parrish, in prep.	Quartz diorite plug, Coates Lake, Mackenzie Mtns.

Table 3.1. Isotopic dates for igneous rocks associated with the Windermere Supergroup.
(zr) - zircon; (wr) - whole rock; (pl+px) - plagioclase and pyroxene separates

A granite in the Omineca Belt that yielded a U-Pb zircon date of $728 \pm 8/-7$ Ma (Table 3.1) is disconformably overlain by the clastic Misinchinka Group (Evenchick *et al.*, 1984; Fig. 3.9); the unroofing of this pluton implies at least 10 km of uplift and erosion occurred before initial sedimentation (Evenchick, 1985). This occurrence suggests that Windermere deposition began later in north-central British Columbia than in the northern or southern Canadian Cordillera; it also indicates uplift of rocks west of the intact ancient continental margin. A western source area is suggested for some Windermere sedimentary units, based on sediment transport directions in northern and southern British Columbia (Gabrielse, 1972 and pers. comm., 1981; Root, 1987) and in the Hyland Group of the Ogilvie Mountains (R.I. Thompson, pers. comm., 1985).

The Mount Harper Volcanic Complex is significant to the interpretation of the tectonic history of the Windermere Supergroup for three reasons. Firstly, the Harper Group can be correlated with other units of the Windermere Supergroup, because the age and local stratigraphic position of the volcanic rocks are known. Secondly, the sequence of faulting, half-graben formation, volcanism and subsidence is clear, providing a rare exposure of the rock record that probably lies buried beneath more extensive Windermere units. Thirdly, the Mount Harper volcanic rocks constitute a compositionally bimodal suite. Knowledge of the eruptive sequence and petrogenesis of the Volcanic Complex assists in determination of the tectonic setting and conditions of rifting for less well-exposed and more altered Late Proterozoic volcanic rocks in the Cordillera.

THE MOUNT HARPER VOLCANIC COMPLEX

GENERAL STATEMENT

The Mount Harper Volcanic Complex consists of extrusive and related intrusive rocks within the Harper Group, and covers an area 15 km (northwest to southeast) by 8 km wide (105 km^2). The distribution of the volcanic rocks is shown in Figure 4.1 (in pocket). Topographic relief near Mount Harper is 1325 m, but regional southwesterly tilt resulted in ample exposure of both the upper and lower units. The physiography and informal place names are shown on Figure 4.2.

The volcanic rocks have been buried, then uplifted and partly eroded since the Late Proterozoic, so that the remains of several large, superimposed volcanic structures can be discerned in the steep walls of cirques and valleys. Cliffs indicate removal by erosion of the volcanic rocks to the north and east of the Complex. The western and southern limits lie beneath Paleozoic strata. An inlier of basaltic flows and breccias 25 km west of Mount Harper may physically connect with the Volcanic Complex beneath mountains exposing Upper Harper Group and CD_b strata. Although the complete dimensions of the original volcanic centre are not known, vent areas for most extrusive rocks are probably located beneath the exposed volcanic rocks, as indicated by outward paleoslopes and flow directions, and by the rarity of dykes outside the area of the Complex.

Deep dissection by stream valleys, and differential erosion of rock types presents many cross-sectional views of the volcanic structure. The rocks are flat to moderately dipping and the present spatial distribution of volcanic units appears consistent with their depositional relationships. Pristine shapes of breccia fragments and textures of flows indicate that the Complex was not deformed by stretching or flattening. Most of the strain appears to have been accommodated along

Fig. 4.2. Sectioned block diagram of the Mount Harper area showing physiographic features and names. The volcanic rocks (long lines) have a synclinal form above the Lower Harper Group (circles) and Wernecke Supergroup (waves).

brittle faults whose vertical displacement is determined from the offset of layered rocks (Chapter 5).

The Mount Harper Complex is divided into six volcanic members on the basis of chemical composition and field relationships. Table 4.1 shows the general lithology and depositional environment of each member. Basalt flows and breccias constitute Member A and overlapping Member B. Member C comprises breccias of mixed pyroclastic and epiclastic origin. Members D and E are coeval rhyolite and andesite respectively. Member F, at the top, consists of quartz diabasic flows and shallow intrusions that are intercalated with the Upper Harper Group.

In the field, volcanic flows were described as basaltic or silicic because most are fine-grained, aphyric and contain few visible indicators of composition. Chemical analysis and microphenocryst identification have confirmed that the members recognized from hand sample identification and stratigraphy have a primary compositional basis (Chapter 7). Despite chemical alteration, original rock types can be estimated from morphology and textures, justifying the use of non-metamorphic rock names (basalt, for example, rather than meta-basalt or greenstone).

All the volcanic rocks are hydrated and contain low grade metamorphic mineral assemblages as a result of burial and fluid alteration. In the freshest flows and breccia fragments sub-isotropic sideromelane has replaced glass, but most mafic rocks contain abundant albite and chlorite. Formerly glassy silicic rocks have been reduced to quartz-feldspar-sericite intergrowths. Many have undergone Na_2O - K_2O and CaO - Na_2O exchange but contents of normally immobile refractory elements, such as Ti, Zr, Nb, as well as most siderophile elements, do not appear to have changed with alteration (Chapter 7).

Mem- ber	Lithology	Subareas of the Mount Harper Complex			Depositional environment
		Harper Ridge	West Harper	Southwest Harper	
Overlying units		(not present)	conform. overlain by UHG to west; overthrust by CD _b	overthrust by CD _b	deepening, stagnating basin; abruptly shallow platform; shelf edge
F	diaktytalic diabase, tuff- breccia	(not present)	tuff and sill complex 180 m pillow flows (10 m)	sills(?) 2-10 m thick (100 m)	subaerial tuff cones(?)
Upper Harper Group		mudstone w/ felsic cobbles, dolostone beds; pebbly siltstone (70 m)	hematitic siltstone, mudstone with mafic clasts (25 m)	(not present)	local uplift and erosion; subsidence, transgression of basin from north
E	andesitic floods, minor tuff	(not present)	tuff in NW (10 m) breccia (60 m)	flows 2-20? m thick, thicker where ponded	emergent vent; lava shattered in water, reworked tuff
D	quartz, feldspar- phyric rhyolite; crystal tuff	flows 5-30? m thick flow bx. (270 m)	Intrusion breccia crystal-litic dikes	lava flows, Ignimbrite and block breccia; crystal-tuff beds 20 cm	subaerial; mostly non- explosive; reworked tuff
C	epiclastic and pyroclastic breccias	(not present)	tuff-breccia (>200 m) conglom (30 m)	tuff-bx. flows (50 m) lapilli tuff layer (30 m) block/ash breccia (50 m)	subsiding, eroding edifice created shallow platform explosive basaltic eruptions
B	hematized, vesic. basalt; chlor- itized basalt	maroon lapillistone, cored bombs (200 m) amyq. flows (250 m)	margin flow bx. (70 m) cobble breccia (70 m) domed pillows (30 m)	pillow and breccia units 30 m thick (150 m)	decreasing water depth; ephemeral volcanic islands
A	chloritized basalt	flow-foot breccia (150 m); tubular pillows (400 m)	flow-top bx. (150 m) mostly pillow flows (225 m)	flow-foot breccia (150 m) aqueagene tuff, pillow and massive flows (250 m)	subaqueous shield volcano; fissure vents
Underlying units		conformable over stratified conglom.	disconform. on Pinguicula dolostone and conglom.	not exposed except in E; disconform. on Werneck dol.	subsiding subaerial debris fans; uplifted erosion surface

Table 4.1. General stratigraphy of the Mount Harper Volcanic Complex, outlined
by subareas. Maximum thicknesses are indicated.

Open spaces such as vesicles and spiracles are typically filled with calcite; textures of the original deposits are strikingly preserved on weathered surfaces where calcite has been dissolved (Plates 1,h; 2,j). Local alteration minerals include hematite, blue chlorite and agate. Veins of quartz and calcite are rare within the Complex.

The lithologic descriptions (Chapter 4) use terms for volcanic deposits that avoid connotations of source area or mode of emplacement (for example, 'block and ash breccia' is preferred to 'pyroclastic flow'). Table 4.2 lists some non-genetic terms for volcanic rock types, their diagnostic characteristics, and implications for composition and depositional environment.

The distribution of individual volcanic members is shown on Figures 4.3 (members A and B), 4.7 (Member C), 4.9 (Member D) and 4.12 (members E and F). These maps also show the location of analyzed samples (e.g. A124-3) and locations of particular outcrops (e.g. Loc. A1).

MEMBER A

DEFINITION AND DISTRIBUTION

Member A consists of basaltic flows and breccias. It is the lowest unit of the Mount Harper Complex, constitutes about a third of the exposed rock (Fig. 4.3) and extends an unknown distance west beneath younger volcanic members. The best exposures of Member A are cliffs on the lower slopes along Coal Creek.

Flow units in Member A generally consist of pillows tightly packed together, intercalated with monolithologic chloritized breccia and massive flows. In contrast to younger, more vesicular rocks in the Complex, flows of Member A are internally structureless or contain sparsely distributed spherical vesicles less than 1 mm in diameter.

	Diagnostic features	Interpretation
Pillow flows	comprise tubes with chill margins and concentric vesicular bands; associated hyaloclastite	subaqueous; relative water depth indicated by abundance of vesicles and amount of breccia
Massive flows	tabular or tapered, homogeneous; distinguished from sills by contacts, breccia margins	sheet flows (Ballard <i>et al.</i> 1979) blocky and aa types indicate subaerial environment.
Intrusion breccia	internally cross-cutting contacts	possible vent area
Isolated-pillow breccia	amoeboid, lenticular sacs, commonly prograding deposit (Carlisle, 1963)	subaqueous flow foot; insufficient lava supply at front
Broken-pillow breccia	Pillow fragments with concentric vesicle bands (Carlisle, 1963)	steep slope or breaks; thermal shock with water
Cobble breccia	Equant fragments of pillows, flows	as for bkn. pillow bx.
Crumble breccia	monolithologic, fractured <u>in situ</u>	commonly felsic flows or domes
Flow-top breccia	Monolithologic, vesicular fragments; grades into massive flow	viscous flow
Hyaloclastite	Between and surmounting pillows, angular chloritic shards	subaqueous, as for pillows
Aquagene tuff	Lenticular beds; chloritic globules and shards (Carlisle, 1963)	subaqueous basaltic effusions
Hydroclastic breccia	layered; chloritic shards and flow fragments, cross-stratified	formed by thermal shock of flows in littoral zone
Cinder	Scoriaceous lumps,	subaerial, Strombolian eruptions
Spatter	Agglutinated, vesicular clasts	subaerial, near vent
Tuff	Lithified ash - fine (< 1/16 mm) - coarse (to 2 mm); (Schmid, 1981)	probably subaqueous if thinly laminated
Lapilli-tuff	Most clasts 2-64 mm in tuff matrix	
Tuff-breccia and Block and ash breccia	Some clasts >64 mm; in tuff matrix	Typically subaerial, from collapsed flows and domes
Pyroclastic breccia	Monolithologic; dominantly blocks and bombs (> 64 mm)	Typically subaerial
Ignimbrite	Felsic, w/ fiamme, broken crystals	Explosive, emplaced hot, subaerial
Volc. epiclastic	Weathered volcanic clasts, seds. (volcanic conglomerate, volcarenite - Fisher, 1985)	Subaqueous
	Fine-grained	crystals or grains less than 1 mm
	Medium-grained	" " 1-5 mm
	Coarse-grained	" " greater than 5 mm

Table 4.2. Explanation of descriptive terms for volcanic deposits

Fig. 4.3. Distribution of members A and B, showing stations for sampling and localities mentioned in text. Blank areas within the Complex are covered by younger units (see Figs. 4.7, 4.9 and 4.12).

CONTACTS

Massive flows and breccias of Member A paraconformably overlie the Lower Harper Group along the northern and eastern edges of the Complex. The contact is always sharp, but undulates over contorted sandstone and conglomerate beds (Fig. 3.4). Where conglomerate underlies the Complex three kilometres south of Mount Harper, a 10 cm limestone bed is conformably overlain by isolated pillow breccias. At the northwest end of the conglomerate, the lowest of three pillowed flow units appears to fill a depression in the underlying dolostone (Fig. 3.2). Only one large clast, a 2 m dolostone boulder, was observed in the lowermost flows and no intercalated conglomerate layers were observed.

At the southernmost exposure of Member A (Loc. A3), thick pillow flow units unconformably overlie bedded limestone, and discontinuous volcanic sandstone (reworked tuff?) occurs at the contact. The clastic beds are evidence for earlier volcanism before the area was over-run by younger lava flows. The limestone varies in thickness and attitude beneath the flows and probably formed an erosion surface on the uplifted side of the Early South Harper Fault.

The upper contact of Member A is indistinct. Member B, distinguished by lower zirconium content than Member A, includes moderately to highly vesicular hematitic flows, as well as pyroclastic and explosively-derived breccias that are not present in Member A. The contact is obscure and possibly gradational in sections containing extensive breccia, but locally Member B breccias overlap erosionally truncated flows of Member A.

STRATIGRAPHY AND THICKNESS

Measured sections through members A and B (Fig. 4.4) demonstrated that few flow units could be traced laterally, although distinctive packages of flows and breccias were recognized. Because the volcano was

Fig. 4.4 (facing)

Fig. 4.4 (con't.)

Fig. 4.4. Volcanic stratigraphic columns in the Mount Harper Complex. Field station numbers for each section are in parenthesis, 'chem' indicates analyzed flows, described in Chapter 7.

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a constructional landform and most volcanic products were deposited on slopes, rapid lateral changes in rock type, unconformities and undulating paleosurfaces are common in the stratigraphy. Tuffaceous marker horizons are only present among rocks of members D and E.

Cliffs along the northeast side of the Complex expose the full thickness of Member A, which varies between 200 and 400 m thick. Only the upper part of Member A is revealed in valley bottoms of Coal Creek and West Trib. In the face south of Lower West Trib, Member A is 450 m thick, but contains numerous sills. Further south, across the Late South Harper Fault, it is less than 100 m thick.

LITHOLOGY

Member A contains pillowed and minor massive flow units 30-70 m thick, breccias derived from them, and subaqueous tuff deposits.

Pillow lavas are the most abundant rock type in the Mount Harper Complex, and where abundantly exposed, those of each member have characteristic morphology (Fig. 4.5). Pillows in Member A comprise networks of twisting, bifurcating and overlapping tubes, some at least 15 m long (Plate 2,a), with smooth exteriors and simple cross-sectional shapes 30-70 cm in diameter (Plate 2,b). Internally, pillows have thin (< 1 cm thick) darkened rims, shallow expansion cracks, evenly distributed microvesicles and rare cavities formed by contraction with cooling (Plate 2,c).

Hyaloclastite (glassy shards spalled from quenched lava surfaces; Fisher and Schminke, 1984) is rare in Member A: instead the matrix of pillow breccias consists of a fine-grained chloritized breccia. It contains compacted globules and rounded 1-3 mm granules with shiny black, sub-vitreous surfaces. In texture and depositional setting, this breccia matches auagene tuff (Carlisle, 1963) that was produced by disaggregation of fluid lava by turbulent water.

Member E

(E82 4)

1 m

Member B

(B118 13)

1 m

Member A

(A133)

1 m

Fig. 4.5. Representative pillow shapes in mafic members of the Mount Harper Complex. Sketches from photographs; localities shown on respective distribution maps.

Captions for Plate 2 (facing)

PLATE 2 - Members A and B

- a) Overlapping tubular basalt pillows (Member A), exposed in longitudinal section. Note bifurcation of tubes near hammer. From 1.9 km S of Mount Harper; A133, 3270' at NTS 524 715.
- b) Closely packed pillows of Member A, exposed in cross-section, decrease in diameter upwards. Dark selvage between pillows is chloritized aquagene tuff. Square encloses pack sack, about 50 cm tall. From 500m N of Mount Harper; 5800', NTS 544 729.
- c) Natural joint surface provides cross-section of a pillow tube, 30 cm in diameter (Member A). The thick darkened rim is mostly chlorite, replacing the original glassy rind. Evenly distributed microvesicles are filled with secondary celadonite. Calcite vug near centre may result from outward shrinkage during cooling. From 500 m NE of Mount Harper; 5500' at NTS 545 726.
- d) Isolated pillow breccia (Member A) contains lenticular to amoeboid lava (light coloured) suspended in aquagene tuff. It was formed by cascading of lava dribblets on the prograding apron of a major flow. From 2 m above base of Volcanic Complex, 2 km SSE of Mount Harper; 4800' at NTS 542 705.
- e) Cobble breccia (Member A) consists of fragments from cooled and fractured flows (fragment at top centre may be from radially jointed pillow). Clasts with bleached rinds and incipient fractures probably reacted with matrix fluids after emplacement. From 1.3 km NE of Mount Harper; 5300' at NTS 546 729.
- f) Rippled top surface of massive basalt flow (Member B), exposed by stream erosion, indicating a westward depositional slope. From 6 km WNW of Mount Harper; 4400' at NTS 483 745.
- g) Contact (dashed line) between massive, amygdaloidal basalt flows (Member B). Flow-top breccia beneath contact contains elongate protrusions and comminuted lava in a chloritized matrix. From 6.6 km WNW of Mount Harper; 4000' at NTS 473 741.
- h) Hyaloclastic breccia from between pillows (Member B). Matrix is replaced by epidote (naturally bleached in this sample) which highlights glass shards now replaced by sub-vitreous chlorite. This breccia formed by shattering and spalling of quenched lava. From float 7 km WNW of Mount Harper, 4700' at NTS 473 752.
- i) Indistinct top (dashed line) of subaerial flow (Member B). Strongly hematized basalt (lower half of photo) increases in vesicularity upward, grading into detached, monomictic flow breccia, formed by break-up of the cooled lava surface. Bright spots are lichen. From 1.2 km SE of Mount Harper; 5000' at NTS 546 715.
- j) Small cannonball-sized pyroclastic bomb consisting of hematized basaltic jacket (natural joint surface illuminated) around a scoriaceous core. Dark agglutinate contains vesicular basaltic lapilli indurated by heat and burial. Coin is 1 cm in diameter. From 500 m S of Mount Harper; 5600' at NTS 539 721.
- k) Cored bomb in moderately indurated lapillistone that suggests reverse size grading. This massive, 7 m thick bed was probably deposited subaerially. From 500 m S of Mount Harper; 5600' at NTS 539 721.

Plate 2. Members A and B

Two types of pillow breccia are well exposed in Member A. Isolated-pillow breccia (Carlisle, 1963) consists of elongate, amoeboid-shaped lava sacs resembling incompletely formed pillows enclosed in aquagene tuff (Plate 2,d). The sacs roughly overlap downward in the direction of flow advancement and form a 1-5 m thick base to pillowed flows. Such breccias form in the oceanic environment at the leading edge of large flows as a result of pillow lobes that detach and tumble down a prograding apron (Jones and Nelson, 1970; Lonsdale and Batiza, 1980). Commonly the breccia is subsequently over-run by the flow, and has been alternatively called 'flow-foot breccia' (Furnes and Sturt, 1976).

The second type, called broken-pillow breccia (Carlisle, 1963) contains fragments of former pillows (identified by portions of the original convex rind and concentric vesicle bands) chaotically mixed with aquagene tuff. This breccia forms in the same way as a more common fragmental rock type here called cobble breccia, that consists of angular, mostly polyhedral blocks (Plate 2,c). Clean outcrops reveal that cobble breccia was produced by disintegration of subaqueous flows as a continuous gradation from incipiently fractured lava toward progressively smaller clasts away from the intact flow. It is not clear whether changes in slope or cooling of the lava produced brecciation. A minimum clast size (5-8 cm) below which no clast contains incipient fractures suggests that internal stresses (probably thermal shock and shrinkage during cooling) cause the flow to disintegrate.

Massive flows are 3-10 m thick, with unbrecciated tops and bases. Some are columnar jointed and others are massive with internal convoluted flow banding indicated by modal variations of plagioclase crystals within largely glassy zones. Typically they contain microvesicles (<1 mm in diameter) sparsely distributed throughout the flow. Most were probably sheet flows, indicated by their smooth, rippled tops. Sheet

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flows on the Recent seafloor appear to result from higher flow rates and lower magma viscosity than pillow-producing lava (Ballard *et al.*, 1979).

Dykes and Sills are fine- to medium-grained chloritic diabase, whose genetic relationship to Member A is based upon field relationships and chemical analysis. The dykes are concentrated in a sub-circular area 1 km southwest of Mount Harper (described later). Sills are common near the base of the Complex, but are difficult to distinguish from massive flows without clear exposure of their contacts.

PETROGRAPHY

Lower flows of Member A are poorly crystallized. Thin sections from the interiors of pillows contain abundant fibrous mats of chlorite with albite and quartz which replace formerly glassy groundmass. Variolitic aggregates in many flows consist of fibrous feldspar and small opaque grains (Plate 3,a).

Better crystallized massive flows have pilotaxitic and intersertal texture. Prismatic plagioclase microlites, many with 'bow-tie' and 'belt-buckle' sections, form a loose mesh. Clinopyroxene (augite, estimated by extinction angle), apatite (?) and anhedral opaque grains (magnetite or pyrite; Brunet, 1986) fill interstices in the plagioclase network.

Microphenocrysts are widely distributed in flows of Member A. Those that may have been olivine, on the basis of hexagonal outline (001 sections), irregular fractures and lack of cleavage, are replaced by chlorite, sericite and, in some cases, silica (Plate 3,b). Some are skeletal or hopper pseudomorphs up to 1 mm long. The lowest flow on the north side of Mount Harper also contains 1 mm glomerophyric feldspars now completely replaced by sub-opaque clay minerals. The euhedral shape of these phenocrysts, in contrast to the poorly crystallized groundmass, implies that they were formed before extrusion of the lava flow.

Captions for Plate 3 (facing)

PLATE 3 - Petrography of members A,B,D,E and F

- a) Pillow core of Member A basalt has variolitic texture, shown by radiating feldspar (white) in a dark, chloritic matrix. Crystalline mass at left is a microphenocryst, possibly olivine, replaced by quartz, calcite (white) and chlorite. Dark circle is a bubble in thin section. Cross-polarized light, A128-4.
- b) Concentrically zoned microphenocryst probably originally olivine (distinctive 001 shape), with spinel(?) inclusion, replaced by serpentine and chlorite. Fine-grained intersertal groundmass consists of prismatic plagioclase with interstitial clinopyroxene (grey) and magnetite. From basalt dyke feeding lowest flows of Member A. Cross-polarized light, A128-3.
- c) Aggregate of former olivine microphenocrysts, rimmed and replaced by hematite (black) and chlorite, in a maroon-weathering basaltic flow of Member B. Groundmass contains sodic plagioclase (An_{25} ; white) with interstices overgrown by hematite. Fibrous grey amygdale along top consists of iron-rich chlorite and minor prehnite, with irregular calcite along rim. Plane-polarized light, B135-1.
- d) Subophitic, medium-grained basalt dyke is the freshest rock in the Mount Harper Volcanic Complex. Augite (cpx) crystals form interlocked clusters from which plagioclase (An_{60}) prisms project. Interstices contain clinopyroxene and magnetite. Cross-polarized light, B158-1.
- e) Felsic flow (Member D) consists of phenocrysts of quartz (qz) and garnet (dark) in recrystallized, sericitic matrix. Some phenocrysts provide nucleation for secondary quartz (lower right); detached slivers and concave edges on others indicate fracturing during cooling. Cross-polarized light, D138-1.
- f) Ash-flow of Member D contains dark chloritic fiamme (formerly pumice) and lithic fragments (quartz at lower left, others mafic) in a vitric ash matrix replaced by sericite. Plane-polarized light, D103.
- g) Member E andesite intrusion contains tabular sericitized plagioclase (grey), skeletal and hopper magnetite crystals, and interstitial quartz (white). Plane-polarized light, E73.
- h) Member F basalt flow has intergranular, diktytaxitic texture of sharp-tipped skeletal plagioclase with clinopyroxene (grey) and chlorite (dark). Amygdules of quartz (qz) rimmed with siderite are characteristic of this Member. Cross-polarized light, F79.

* All determinations of plagioclase composition by Michel-Levy method.

Plate 3. Petrography

Members A, B, D, E and F

PRIMARY STRUCTURES AND POSSIBLE VENT AREAS

In the northern half of the Complex, the upper contact suggests Member A formed a broad submarine shield 10 km or more across. The predominance of tubular pillow lavas over sheet flows, by comparison with Recent seafloor structures, indicates that the paleotopography was marked by fault scarps and walls, rather than broad axial valleys (c.f. Ballard and van Andel, 1977). In Coal Creek valley, most pillows indicate westward flow, implying that the source vent lies beneath Mount Harper, or further east (see Fig. 5.4).

Few intrusions that might have fed Member A are exposed, thus these structures must underlie the present outcrop. In surrounding older sedimentary rocks, eastward-trending dykes 3.5 km north of Showcap outlier may have fed Member A (they were not analysed), but the widely dispersed, north-trending dykes are compositionally related to Member B. Earliest eruptions probably issued from one or more centrally located vents.

A subcircular dyke zone 150 m across that is exposed 1 km southwest of Mount Harper is interpreted as the deeply eroded feeder pipe to the upper flows in Member A. It consists of mutually interpenetrating gabbroic intrusions that range from 10 cm to 7 m in width. Both aphanitic margins (A131-1) and medium-grained cores of dykes (A131-2) have compositional characteristics of Member A.

Some twenty sub-horizontal sills, 2-7 m thick, are exposed on the mountain faces south and west of the confluence of West Trib and Coal Creek (middle of section 144). Sills, indicated by 1-2 cm chill zones along contacts, appear to cross-cut each other and enclose slivers of pillowed flows up to 3 m thick and tens of metres long. Those near the middle of the section have medium-grained, subophitic diabase interiors petrographically similar to dykes in the feeder pipe 1.5 km to the northeast. The multiple injection of sills in this area appears to have

increased the thickness of Member A by about 150 m, raising the top of Member A in the southern half of the Complex. Inflation of seamounts by injection of dykes and sills has been deduced from cross-sectional exposures of uplifted examples (Staudigel and Schminke, 1984).

DEPOSITIONAL ENVIRONMENT AND ERUPTIVE STYLE

The predominance of aquagene tuff and pillows indicate that Member A was extruded underwater. The absence of explosively-derived breccias in Member A, and their abundance in Member B, may reflect decreased hydrostatic pressure upon the later extrusions, although increased magma viscosity or volatile content could have produced a similar change.

A 5 m thick columnar jointed flow northeast of Mount Harper has spiracles at the base where it flowed over unconsolidated sediment. Spiracles are epidote-lined cavities caused by steam trapped beneath the over-riding lava flows (Macdonald, 1972; p. 81). The absence of interdigitated conglomerate and volcanics near the base of Member A implies that either erosion ceased before volcanism (perhaps the uplifted fault block became submerged) or that the volcano rapidly became a positive topographic feature, and distal exposures where sediments may be intercalated are no longer exposed. Subsidence is implied because the conglomerates are inferred to have been subaerially deposited (Mustard, in prep.) but are everywhere overlain by pillow lavas and subaqueously deposited breccias.

The shape and size of the Member A seamount is incompletely known. Modern seamounts that developed above fracture systems are commonly elongated and have faulted margins (Basaltic Volcanism Study Project, 1981, p. 733). Ballard and van Andel (1977) described a seamount (2000 m relief) near the mid-Atlantic ridge that was constructed from fluid basalt flows, succeeded by more viscous and less extensive lavas that developed a shingle-like overlap. The seamount formed by Member A was

transformed into a more irregular-shaped structure by an increase in primary slopes with a greater proportion of breccia in Member B.

MEMBER B

DEFINITION

Basaltic flows and breccias that overlie Member A and the dykes considered to have fed them constitute Member B. This unit is characterized by very large pillows, as well as highly vesicular and locally hematitic flows and breccias. Most rocks weather orange, brown (limonite) or maroon (hematite). Pipe amygdules and cavities filled with translucent pink, orange and white agate are characteristic of Member B.

Dykes included in Member B do not cut overlying units and have chemical composition similar to the lavas. They constitute swarms that trend southeast. Faults parallel to the dykes cut overlying units (the Upper Harper Group in outliers) and were therefore active after emplacement of Member B.

The distribution of Member B is shown on Fig. 4.3. Best exposures are on buttresses of the main ridges, particularly beneath Snowcap, Saddle and Baldie outliers.

CONTACTS AND DIMENSIONS

The base of Member B is difficult to recognize because the compositional differences which distinguishes it from Member A are not reflected by an abrupt lithologic change. Where the boundary has been chemically defined, it also corresponds to the first appearance of orange-weathering or hematitic rocks.

Member B is disconformably overlain by three rock units: Member C (which contains mixed fragments of members A and B), hydroclastic tuff of Member E, and the sedimentary outliers of the Upper Harper Group. A very large volcanic edifice constructed by Member B appears to have been

eroded before deposition of these units; its remaining thickness varies greatly. This is revealed on the ridge south of West Trib where Member B ranges from less than 50 m thick at the eastern end to more than 500 m beneath Black Towers further west. Beneath Snowcap outlier a 560 m succession was measured (section 3 on Fig. 4.4); further west at Tango Tarn it is 160 m thick but the base is not exposed. Subhorizontally layered hematitic flows are 350 m thick south of Mount Harper (section 11 on Fig. 4.4).

LITHOLOGY

Volcanic rocks of Member B are grouped according to their physical and mineralogical features into the Chloritic facies and the Hematitic facies. The contact between the facies is transitional both laterally and vertically. The Hematitic facies, where preserved, always lies above the Chloritic facies.

The Chloritic facies includes massive and pillowed flows, as well as breccias formed by these flows and pyroclastic rocks.

Massive flows of Member B are fine-grained grey basalt with delicate diabasic textures visible on etched weathered surfaces. In general flows are tabular, more than 3 m thick, and extend hundreds of metres laterally. Many appear to have pooled in paleo-depressions. Some flows have brecciated tops consisting of sub-angular flow fragments in a compositionally similar matrix, but others are smooth, locally exhibiting rippled or lobate surfaces (Plate 2;f,g). Vesicles are commonly concentrated in bands, and some are stretched, suggesting that they formed before or during emplacement of the flow. Pipe vesicles that occur above the basal chilled margin of some flows and locally in pillows, are several millimetres in diameter and up to 10 cm long.

Pillowed flows typically form discrete mounds or layers less than

10 m thick. Some layers grade upward into massive flows, and outward into broken pillow breccia. Hyaloclastite jackets surround most pillows and commonly separate layers of pillows.

Two varieties of pillows are distinguished by their general forms. The first type, which is very common in the lower half of Member B, resembles elongate sacs with ellipsoidal cross sections (50-150 cm), has thick rims (2-4 cm), and rough, knobby surfaces. Most contain concentric bands of spherical vesicles and short pipes above the lower chilled margins. Where exposed parallel to the direction of flow, the pillows drape over one another, and are commonly overlain by thin sheet lava flows with rill-like corrugations.

The second type consists of larger (70-200 cm across) pillows that tend to be bulbous and ovate (Fig. 4.5). Some are shaped like mattresses or bolsters, and do not reveal a clear flow direction. Large interstices are filled with hyaloclastite, calcite and aquagene tuff.

Breccias related to subaqueous extrusion of Member B comprise several distinct types described below, with variations in texture and dimensions. Hyaloclastite is best preserved where resistant epidote has replaced the matrix and chloritic shards are in bold relief (Plate 2,h). The lack of crystallized lava fragments, lack of stratification and irregular distribution, typically around pillows, distinguish hyaloclastite from 'hydroclastic breccia', a deposit characteristic of Member E.

~~Spatially~~ and genetically associated with the first type of pillows are deposits that contain features of both isolated pillow breccia and broken pillow breccia types (previously described for Member A). In the hyaloclastite matrix, which constitutes most of the rocks, are individual pillows, pillow fragments, thin-dribble flows and feeder tubes (1 m diameter). Generally the deposits are part of a 30-50 m thick

succession grading from isolated pillow breccia upward to very large pillows and massive flows.

Maroon and green breccia is composed of amoeboid lava sacs distributed without preferred orientation in hyaloclastite. Some fragments have irregular 'arms' up to 2 m long and display reddish, oxidized rims. The lack of preferred orientation distinguishes them from isolated pillow breccia. Another breccia deposit contains twisted ribbons of flow rock in hyaloclastite. It may have formed from strips of chilled crust that became detached from flowing lava and floated by water or gases, similar to those produced by 'lava falls' at the Kilauea, Hawaii (H-U. Schminke, oral presentation, 1983). Unusual amoeboid pillow shapes are also produced by vesiculating basalt effusions into shallow, turbulent water (Dimroth *et al.*, 1985; Cas and Wright, 1987).

Breccias interpreted as pyroclastic in origin are irregularly distributed south of Mount Harper beneath the Hematitic facies. Most common are lenses 5 m thick by more than 20 m long, here called subaqueous tuff-breccia, in which cobble-sized, elongate clasts are embedded in a scaly chloritic matrix. Some clasts resemble cannonballs, with dark, spalling rinds and hematitic, flow-banded cores; others are egg-shaped and increase in vesicularity toward the center. They are pyroclastic bombs and armoured lapilli, resembling those illustrated in Fisher and Schminke (1984; p. 94).

The Hematitic facies consists of purple, red and brown highly vesicular flows and breccias that are not separated on the map. Five minor occurrences in the western part of the Complex signify the former extent of this facies, although it is not clear whether remnants of a single sheet deposit, numerous minor cones or individual slump blocks from a single source are represented.

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Hematitic lava flows are 1-5 m thick and form 20 m thick sections beneath Saddle outlier. Interflow contacts are obscure because flows are increasingly fractured toward their tops, grading into a carapace of monomictitic breccia (Plate 2,i). Vesicles also increase in size and abundance upward; 'dog-bone' and other angular shapes imply that they were deformed by viscous flow, a typical feature of aa flows.

Pyroclastic breccia consists of poorly stratified tuff and lapilli-bomb tuff. The latter contain scoriaceous basaltic lumps loosely distributed among hematitic granules. Some deposits contain cored bombs (Plate 2,j,k) and those with annealed clasts may be agglutinate deposited close to subaerial vents.

Subvolcanic intrusions are distinguished by their regular jointing perpendicular to contacts and are less altered than flows. Medium-grained interiors are ophi-mottled; on weathered surfaces the outlines of former pyroxene crystals are discernible. Sills are 1-5 m thick and difficult to distinguish among parallel-layered flows. Single dykes are 0.5 to 2 m wide, but single fissures up to 5 m wide typically contain multiple intrusions (Fig. 4.6,a). One dyke was observed to connect with a Member B flow (144-5) and another feeds a small laccolith (Fig. 4.6,b).

PETROGRAPHY.

Member B flows are better-crystallized than those of Member A. They have intersertal textures characterized by a dense mesh of plagioclase enclosing 1 mm augite grains. Some contain 1-2% microphenocrysts which may have been olivine, now pseudomorphed by yellow green chlorite and quartz with hematite at the edges (Plate 3,c). The groundmass contains 2-3% titanomagnetite and abundant secondary sphene.

Dykes of Member B are the freshest rocks in the Volcanic Complex. A mesh of well-formed, twinned plagioclase laths (An_{50-65}) hosts

Fig. 4.6. Intrusive relationships in Member B:
 a) layered flows cut by regularly spaced, multiple, interpenetrating dykes, near 144-5.
 b) Laccolith within breccia layer at top of Member A, near 144-11.
 c) Dykes cross-cutting basal contact of Volcanic Complex, 1.5 km N of Snowcap outlier.

subophitic, strongly coloured pyroxene. The coarse-grained centres of sills consist of 3 mm augite grains in mutual contact and partly enclosing small plagioclase prisms. Unlike flows, where mafic minerals were destroyed first, in dykes the plagioclase is more altered than pyroxene; it is albitized or replaced by clays and saussurite.

PRIMARY STRUCTURES

Although the upper contact of Member B is in most places disconformable, it provides clues to the former shape of the volcanic edifice. This is particularly apparent in a bench beneath the Snowcap outlier (Plate 4,a) and at the top of the cliffs south of Tango Tarn (Plate 4,c), where exhumed erosion surfaces are evidence of the flanks of the original volcano. Similarly, the planar southwestern slope 1 km west of Mount Harper is parallel to the former flow surface of a subaerial structure. These preserved remnants help to reconstruct original landforms (Chapter 9).

Concentrations of dykes at Localities B3, and B162 could represent the locations of eroded vents. They are generally consistent with outward flow directions determined from pillows and rippled surface features of the lavas (Fig. 5.6).

DEPOSITIONAL ENVIRONMENT AND ERUPTIVE STYLE

In the Chloritic facies, scoriaceous bombs, amoeboid pillow shapes and ribbon lava fragments indicate derivation from a shallow water vent. The thin flows and pillow breccia layers in the western part of the Complex were probably deposited as an apron of a volcanic centre located further east.

In the Hematitic facies, well-preserved aa flows, cinder-like breccias and agglutinate containing cored bombs indicate subaerial eruptions. All known occurrences of these deposits are strongly

Captions for Plate 4 (facing)

PLATE 4 - Volcanic structures

a) View eastward (from Baldie outlier) shows central part of the Mount Harper Volcanic Complex. Planar slopes (on left) are a paleo-erosion surface beneath a lava flow of Member F; cliffs below expose Member B pillows and breccias. Dark face of mountain in centre reveals a cross-section of the ancient Member B edifice which was partly eroded before deposition of the Upper Harper Group, preserved in Snowcap outlier.

b) View westward from Baldie outlier across the broad valley of Upper West Trib shows southward-dipping stratigraphy of Upper Harper Group. A thicker section of Upper Harper Group is preserved on north (right) side of the North Harper Fault. The CD₁ dolostone has been thrust northward upon incompetent rocks of underlying Upper Harper Group and Member C. Northwest of the Complex (top right corner) older Pinguicula dolostone is directly covered by Upper Harper Group and Member F, indicating that the initial half-graben (filled by Lower Harper conglomerate) did not extend northward. The Early South Harper Fault, the southern edge of the half-graben, underlies strata at extreme left. Width of view: 8 km.

c) View southward from slopes underlain by Member C, across Tango Tarn (bottom right corner). The cliff is a cross section of the Member B edifice, in which hematitic (subaerial) flows succeeded by pillow lavas imply synvolcanic subsidence. Maroon mudstone of the Upper Harper Group overlies Member B. The trace of Tango Tarn Fault (dashed line) is the southern limit of tuff-breccia (Member C). From this perspective, Member B appears thrust over Member C, but this fault is considered a reactivation of a synvolcanic fault scarp and Member C filled the trough (foreground area) before deposition of the Upper Harper Group (see Member C description and Fig. 4.8).

Plate 4 Structures

- a) West of Snowcap outlier
- b) Upper West Trib valley
- c) South of Tango Tarn

hematitic and no pillows or other subaqueously erupted rocks occur in this facies; therefore all maroon rocks are considered to have oxidized subaerially.

These two facies which overlie deep water volcanic rocks represent a shallowing subaqueous environment, resulting in an emergent volcanic landform. Because pyroclastic volcanic islands are quickly eroded, subaerial deposits are rarely preserved except by subsidence or slumping below water level. The deposits of Member B suggest a subaerial cinder cone was constructed during this phase of volcanism.

The subaqueous evolution of seamounts has been studied (Lonsdale and Batiza, 1980; Staudigel and Schminke (1984). To understand ancient volcanic deposits, accounts of modern eruptions leading to growth of volcanic islands such as Surtsey must be considered in terms of the potential of each deposit type for long term preservation (e.g. Kokelaar and Durant, 1983; Appendix II in Cas and Wright, 1987). Subaerially erupted products of island volcanoes are chiefly preserved by redeposition underwater (Ayers, 1982). Despite their subaqueous depositional setting, the extent and variety of oxidized deposits of Member B suggests considerable subaerial activity.

MEMBER C

DEFINITION AND DISTRIBUTION

Several types of volcanic breccia of inferred pyroclastic and epiclastic origin are grouped in Member C. This is a chronostratigraphic unit; although the breccias differ in texture and inferred mechanism of emplacement, they occupy a single stratigraphic interval. Member C succeeds members A and B but contains no clasts that were clearly derived from members D, E and F. Locally remnants of these younger members overlie Member C.

Member C breccias record both explosive activity at the close of the first phase of volcanism and simultaneous destruction of the volcanic structure composed of members A and B. They constitute a spectrum from wholly primary (pyroclastic) to completely reworked (epiclastic) breccias, but most outcrops reveal the latter range. They were basaltic, but the exact chemical composition of this phase of volcanic activity was not determined because massive flows or dykes suitable for chemical analysis could not be correlated with the tuffs. Hematitic lapilli and scoria in Member C indicate emplacement of the breccias before lithification of some Member B deposits.

The distribution of volcanic breccias included in Member C is shown on Figure 4.7. In the south part of the Complex, members A and B are locally tilted and unconformably overlain by discontinuous, coarse breccias (Member C) and Member D. Topographic benches around West Trib valley represent an exhumed erosion surface of the ancient edifice upon which a veneer of Member C breccias is locally preserved.

CONTACTS

Basal contacts of Member C are rarely exposed. Heterolithic conglomerate appears to drape over lenticular hyaloclastite and pillows of Member B at the south end of Upper West Trib valley (Plate 5,a). In

Fig. 4.7. Distribution of breccia units in Member C, and area of exposed paleo-erosion surface (top of edifice of members A and B).

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Micro-D

Captions for Plate 5 (facing)

PLATE 5 - Member C

a) (top centre) Abrupt contact (dashed line) of **volcanic conglomerate** with mudstone matrix above a basaltic flow with steep joints. Subangular cobbles are same rock type as underlying flow. From southwest corner of Complex, 82-4; 4500' at NTS 444 715.

b) Gradational contact (below dashed line) between Member C **tuff-breccia** and overlying paraconglomerate of the Upper Harper Group. The transition from tuffaceous basaltic matrix with subrounded mafic cobbles, to dolomitic mudstone matrix containing both **volcanic** (dark) and carbonate (light) clasts is interpreted as the result of ~~continuous debris~~ flow sedimentation from nearby volcanic and regional carbonate units. From 1 km N of Tango Tarn; 5000' at NTS 455 773.

c) **Block and ash breccia**, from 3 m thick, poorly sorted bed. White lichen grows preferentially on clasts of vesicular lava, such as the cluster below hammer head. Poorly consolidated matrix consists of chlorite (altered tuff) and comminuted lava. From 1 km SE of Black Towers; 5000' at NTS 462 713.

d) **Reworked volcanic ash** with probable breadcrust bomb (arrow, lower left) indicated by concentric vesicular bands and radial joints. This 3 m thick deposit is interpreted as a debris flow. From 1 km ESE of Tango Tarn; 5350' at NTS 465 762.

e) Poorly indurated **mafic tuff** underlying hydroclastic tuff (rough weathering upper layer). Undercut cliffs represent fine-grained, reworked tops of layers. Boxed area, 3 m high, is shown in Plate 5(f). From 300 m S of Soft Hills, C75; 4750' at NTS 477 707.

f) Detail of pyroclastic layers in Plate 5(e) shows mafic tuff with reverse size grading upward. Subangular lava clasts near top of bed (arrows) are overlain by reworked ash (light, with lichen spots). The deposit is interpreted to have resulted from a Plinian eruption toward the close of the first volcanic cycle. Location as above.

g) **Block and ash breccia** showing reverse size grading toward planar tops (dashed line), which indicates a non-erosive base of the subsequent flow. Note lava block greater than 1 m (arrow). From 5 km SW of Mount Harper, C50-1; 4400' at NTS 501 701.

h) Indurated, well bedded volcanic wackes, overlain (dashed line) by massive volcanic block and ash breccia. Bed in lower half is better sorted toward its reworked top. From 1 km SW of Black Towers; 3900' at NTS 463 710.

i) **Volcanic cobble conglomerate** containing rounded basaltic clasts, framework supported in mafic mudstone. From 1 km SW of Black Towers, 82-4; 4600' at NTS 444 714.

Plate 5. Member C

the northwestern part of the Complex (Loc. C125) Member B pillows are succeeded by block and ash breccias which have infilled cavities and contain broken pillow fragments.

Member C is overlain, in the western part of the Complex, by conglomerate of the Upper Harper Group. The upper contact varies, but north of Tango Tarn is a lithologic transition within a 5 m thick bed (Plate 5,b) to debris flows derived from sedimentary rocks. Near the Soft Hills locality, the breccias are conformably overlain by hydroclastic tuff (Member E).

LITHOLOGY

Member C is described below as four deposit types, here called tuff-breccia, lapilli-tuff, block and ash breccia and volcanic conglomerate. Their distribution and dimensions are indicated on Figure 4.7 and in Table 4.3 respectively.

Tuff-breccia consists of sub-angular basaltic fragments, loosely or densely packed in dark coloured tuff (Plate 5,c). Outcrops are distinctive because the crumbly chloritic matrix scales away, exposing fresh dark undercut cliffs in which each bed is outlined by contacts more resistant to weathering. North of Tango Tarn Fault it is homogeneous, more than 200 m thick and consists of uneven, laterally persistent 2-6 m beds that are internally stratified by clast trains. The matrix is weakly foliated oblique to bedding contacts, resulting from deformation of the poorly lithified, relatively incompetent deposits.

Monomictic basaltic clasts 10-50 cm in diameter constitute 10-40% of the beds. North of Tango Tarn many clasts are sub-spherical or lenticular, and most are vesicular with darkened rims (possible chill zones) within shells of scaly matrix. They could have been plastically deformed and rolled into loose ash following eruption and while still hot. Some clasts resemble bombs (Plate 5,d).

Location	Thickness total (Individ.)	Matrix % composition	Clasts shape size comp.	Fabric	Interpretation
Tuff-breccia					
N of Tango Tarn	>200 m (2-6 m)	>60% chloritic granules	sp., lent. non- vesic basalt	clast trains	pyroclastic flow
Headwaters Upper W Trib.	75? m (2-5 m)	80% volc-sand- stone, chert-tuff	cobbles, lapilli basalt	poorly sorted, no laminae	pyroclastic flow
Block and ash breccia					
SW of W. Trib confluence	45 m (5-18 m)	50% chlor. mud- stone, qtz chips	subang. pebbles, basalt, rhyolite	crude sorting reverse grading	lahar
S edge of Complex	50 m (10-15 m)	35% indur. basalt, mudstone w/ chips	ang., subdrd. vesic. basalt	poor sorting basal joints	hot? avalanche
Lapilli-tuff					
300 m S of Soft Hills	30 m (1-3 m)	10-90% vesic. granules, tuff	subang lapilli (mafic)	well-layered. normal grading	shallow water airfall
Volcanic conglomerate					
Baldie out- lier	30 m (0.5-3 m)	30% sand, volc. chips, siltstone	rounded, vesic. pebbles, cobbles	poorly sorted reworked tops	pebbly mudflow
west side of Complex	30? m (2-4 m)	15% hematitic volc canic sandstone	rounded cobbles angular pebbles	coarse strat.; reworked tops	debris. flow from across beach
SW of Black Towers	8? m (1-3 m)	15% chlor. ash	subround pebbles massive basalt	clast trains, mudstone parting	volcanic talus

Table 4.3. Dimensions and composition of major breccia deposits in Member C.

Lapilli-tuff exposed 6.5 km southwest of Mount Harper (Loc. C75) consists of 70 cm to 2.5 m thick beds with normally and reversely graded, poorly consolidated vesicular and dense mafic lapilli (Plate 5,e). Bedding contacts are sharp and undulating, and basaltic fragments up to 10 cm long are concentrated near the top (Plate 5,f). Fine laminae suggestive of quiet subaqueous conditions were not observed, but reworked lapillistone beds and pillow lavas indicate shallow water deposition.

Block and ash breccia differs from tuff-breccia by containing a mixture of rock types as clasts. Bedding is uneven and deposits are locally very thick (tens of metres) with subangular basaltic clasts from a centimetre to a metre across. They are poorly sorted and commonly reverse-graded at the top (Plate 5,g). Clasts have fractured edges, indicating derivation from pre-existing rocks including hematitic and chloritic flows and pillow breccias and uncommon dolostone. The matrix is lighter coloured, and contains more dolomitic mud and less chloritic ash than tuff-breccias. The basal contact is obscure for deposits that overlie lava flows and breccias, but for those that rest on fine-grained beds the contact is abrupt and indicates no basal scour (Plate 5,h). Block and ash breccias probably originated as debris flows from predominantly volcanic, but some uplifted sedimentary source areas.

Volcanic conglomerate layers contain a variety of clasts, from vesicular, hematitic to massive, chloritic basalt that are densely packed in a sand-sized volcanic matrix (Plate 5,i). The pebble- to cobble-sized, subrounded clasts are poorly sorted in tabular beds 1 to 2 m thick. Some beds are crudely stratified, with lenticular beds of larger clasts (cobble trains) or sandy matrix.

PRIMARY STRUCTURES

The distribution and lithologic variations in Member C yield clues to the paleogeography of the volcanic edifice. A steep, north-side-down

fault is inferred to lie in the valley of Tango Tarn, and forms the southern boundary of Member C (Fig. 4.8,a). Upper Harper Group conglomerate overlies the tuff-breccia to the north, but to the south the same unit directly onlaps an erosion surface on Member B. The breccia is inadequately exposed to determine lateral facies variations, but its thickness diminishes northward from the Tango Tarn Fault. These relationships suggest that the Tango Tarn Fault was active before deposition of the Upper Harper Group and possibly resulted in accumulation of the tuff-breccia in a depression to the north (Fig. 4.8,b). Other deposits, particularly south and east of the Soft Hills appear to have accumulated on steep primary slopes and probably represent accumulation on the lower flanks of the eroded volcanic structure.

DEPOSITIONAL MECHANISMS AND ENVIRONMENT

The tuff-breccia probably represents pyroclastic flows. Its matrix of intact shards and the high content of bomb-like clasts imply direct ejection from a vent, or slumps of recently-erupted debris. The large size and massive internal texture of the clasts is consistent with ejection during Plinian eruptions (L.D. Ayers, pers. comm., 1984). Although they could have issued from a subaerial vent, these breccias were probably deposited underwater.

The lapilli-tuff, which contains angular, internally fractured fragments and abundant tuff, is also typical of Plinian fallout (Fisher and Schminke, 1984; p. 117). The ash-filled debris flows plucked rock fragments and probably floated them in a thick slurry.

Heterolithic block and ash flows were probably volcanic debris flows (lahars). They may have begun as hot pyroclastic flows which incorporated broken rock and water, or could simply represent deposits formed as a result of slope failure on the flanks of the volcano. Typical laharic deposits are thin on steep slopes and much thicker on

Fig.4.8. Stratigraphic relationships of the Tango Tarn Fault.

- a) Sketch map showing distribution of Member C, mostly consisting of tuff-breccia;
- b) Schematic, true scale section illustrating the possible shape of tuff-breccia deposit if Tango Tarn was a synvolcanic normal fault.

gentle slopes and form broad aprons around volcanoes (Cas and Wright, 1987). Lahars do not disturb underlying beds, retain clasts up to 1 m in size within the flow, and are commonly reversely graded near the top (Fisher and Schminke, 1984, p. 307).

The heterolithic conglomerates were probably eroded from volcanic cliffs and reworked in shallow water. Subaerial erosion is implied by rounded clasts and layers of cobbles and pebbles.

The general distribution of breccias in Member C, from pyroclastic flows in the north and east, to laharc and conglomeratic beds in the southwest, may broadly reflect southwesterly progression of mixing, reworking, and distance from the eruptive source. The deposits appear to have formed an apron sloping westward from a large volcanic structure to the northeast.

MEMBER D

DEFINITION, DISTRIBUTION AND DIMENSIONS

Member D consists of massive silicic flows, tuffs and intrusions. Flows are light pinkish-grey or green and cherty; they are easily distinguished from mafic rocks. Because most rocks are columnar jointed and deeply fractured, only a few clean outcrops protrude from the rubble- and vegetation-covered slopes.

Extrusive rocks of Member D are restricted to the southern part of the Mount Harper Complex (Fig. 4.9). Several plagioclase-phyric flows constitute a 170 m section on the south side of the Late South Harper Fault east of Coal Creek. Texturally similar rocks west of the Creek and south of Black Towers are believed part of the same flow, and in this area the unit extends north of Late South Harper Fault, where it is about 150 m thick.

Fig. 4.9. Distribution of silicic Member D. Small occurrences are shown larger than scale.

Three mounds of silicic rubble, thought to be centred on feeder dykes, are situated on the high ridge south of Lower West Trib; the easternmost is called the Soft Hills. The mounds overlie hydroclastic breccia of Member E. Although at a higher stratigraphic level, these occurrences are included in Member D because they are compositionally similar. Two quartz-porphyry dykes and three silicic breccias that appear to be intrusive are probably contemporaneous with Member D.

CONTACTS

The basal contact of Member D is not exposed. Tilted amygdaloidal flows of members A and B are overlain by rhyolite with apparent angular unconformity (Plate 6,a). A green laminated sandstone containing detrital feldspar (reworked crystal tuff?) up to 30 cm thick is present at the contact at widely separated locations.

The upper contact of Member D is locally marked by a thin crystal tuff and is generally disconformable beneath the hydroclastic breccia and flows of Member E. Nothing remains of units that covered the upper group at the Soft Hills locality.

LITHOLOGY

Massive rhyolite dominates Member D, and is believed to represent one or more lava flows, although a pyroclastic origin cannot be ruled out. It contains flow bands (Plate 6,b), sparse but uniform phenocryst distribution and no accidental fragments, all features that are inconsistent with typical welded ash flows (Hausback, 1987). Most outcrops display jointed columns (15-25 cm sides) forming upright palisades or fans (Fig. 4.10). Fluted paleo-surfaces of these flows are overlain by fragmental deposits composed of polyhedral blocks broken along joint surfaces. Among the outcrops south of Mount Harper, variations in proportions of quartz to feldspar phenocrysts in the rock could reflect separate flows, which are each interpreted to be tens of metres thick.

Captions for Plate 6 (facing)

PLATE 6 - Members D, E and F

- a) View northward (from E72) of the southwestern part of the Complex, showing the upper volcanic units. Contacts dip gently toward viewer. A paleo-erosion surface on Member B is overlain by columnar-jointed rhyolite flows (Member D) which reduce to vegetated rubble, and are succeeded by massive flows (Member E) at top left. The base of Black Towers and crest of the ridge at right is underlain by resistant hydroclastic breccia (Member E).
- b) Refolded flow banding in rhyolite lava (Member D). White streaks are coalesced spherulites, indicating the previously glassy texture of this rock. From the north side of Soft Hills, 4970' at NTS 474 711.
- c) Flow breccia (Member D) containing comminuted rhyodacite lava. Dark specks are limonite replacing weathered ferromagnesian minerals. This deposit occurs on the margin of a rhyolite extrusion (exogeneous dome?) where it formed by comminution of viscous lava. From 1200 m NW of Soft Hills; 4650' at 461 713.
- d) Aerial view southward from east of Black Towers shows plateaux formed by resistant Member E flows. Vegetated slopes beneath cliffs are hydroclastic breccia and Member D rhyolite, but their contact is irregular. The planarity of Member E andesite flows indicates that they were lava floods.
- e) View westward of the top of the Complex south of Late South Harper Fault. Note that layered lava flows of Member E are truncated at the contact (dashed line) with gently dipping CDb dolostone, the only direct evidence that the contact is a fault.
- f) View northward from Soft Hills, showing smooth weathering slopes of Red Hill (Member F) and Baldie outlier (Upper Harper Group). Beneath these units and overlying Member B hydroclastic breccia (Member E) forms buttresses, which are offset by the fault that bisects Red Hill.
- g) Coarse hydroclastic breccia consisting of angular and sub-equant glass fragments of andesitic Member E flows. Two clast types are indicated by their different reactions to quenching and alteration. Formerly glassy rock shows concentric 'liesegang banding' (oxidation/reduction fronts) and spherulite growth (white spots). Fragments that appear to have been more crystallized (e.g. in left centre) are uniformly chloritized and crumble, without forming splinters. This breccia is an intermediate stage in the shattering of glassy flows upon contact with water. From 2 km SE of Black Towers, E44; 3800' at NTS 479 699.
- h) Fine-grained hydroclastic breccia contains well-sorted fragments of formerly glassy andesite, now whitened by spherulitic alteration. These breccias constitute a layer up to 60 m thick north of Lower West Trib, and were formed by shattering of quenched Member E flows, and not by pyroclastic activity. From south side of Snowcap outlier; 5200' at NTS 487 737.

Fig 4.10. Spatial relationship of rock types at the top of Member D on the south side of the Black Towers locality (sketch from photo).

Fig. 4.11. Rock units of Member B, C beneath massive flow of Member E, 600 m east of Soft Hills locality (sketch from photo).

In comparatively fresh flows, phenocrysts less than 2 mm in length include: potassic feldspar (1-3%, by volume), plagioclase (1-2%) and quartz (<2%). Rare spinel phenocrysts (or xenocrysts) were observed south of Mount Harper (Loc. 138-1; Plate 3,e). The paucity of phenocrysts and absence of broken crystals are uncharacteristic of ash flows, whereas several conformable breccias (Plate 6,c) support a lava flow interpretation. Strongly altered flows consist of crumbling limonitic clay, which locally reveals clusters of 0.5 mm spheres that may have been spherulites. All flows appear to have been glassy; no primary crystalline rocks have been found.

Ignimbrite occurs in a 4 m thick deposit at the base of the westernmost rubble mound (103). The rock is greenish, streaked with flattened formerly pumice fragments (5%) and speckled with 1 mm broken feldspar and quartz phenocrysts (2%; Plate 3,f). It is interpreted as a single cooling unit.

Felsic tuffs and arkose are discontinuous along two stratigraphic levels. Layers 1.0-1.5 m thick consist of laminated green mudstone interpreted as waterlain, formerly vitric and now recrystallized tuff. At or near the top of the main rhyolite flow the crystal-lithic tuff varies laterally to medium-grained arkose. The arkose is composed of broken feldspar crystals (40%), dark angular rock fragments (25%), and 1 mm quartz grains (10%) set in a red and green mudstone matrix.

Felsic dykes and subvolcanic intrusions comprise two types. At least eight sub-parallel, north-trending dykes 5 km southwest of Mount Harper (48-1) are compositionally similar to the massive flows. They are single or multiple, 3-5 m thick, and distinguished by broad columns normal to the vertical margins. Most consist of flow-banded quartz-phyric rhyolite that locally contain hematite, with vesicular zones filled with clay minerals (cladonite).

The second type, represented by dykes west of Black Towers (Loc. D3) and east of Tango Tarn (Loc. D4), are up to 3 m wide, and consist of phenocrysts of plagioclase, potassic feldspar and lesser quartz, densely packed, in a sparse chloritic matrix. Ragged patches of dark chlorite that may have been pumice fragments and angular 5-10 mm chips of basaltic rock that were probably torn from the walls. Grain size decreases toward the margins of these dykes, suggesting granulation during injection of a crystal mush. The dykes cut the hydroclastic tuff (Member E) and represent the last stage of silicic activity in the Complex.

PRIMARY STRUCTURES

An arcuate cliff outcrop 50 m in diameter and about 15 m high, could have been an explosive vent wall (Loc. D5). The scarp consists of massive, 1-20 cm angular silicic fragments in a dark tuffaceous matrix, strongly limonitized, that may represent a vent-facies deposit. The structure is truncated to the south by a south-side down fault, which could have followed the plane of weakness of the volcanic pipe.

The north-trending silicic dykes (e.g. 48-1) probably fed the rhyolite flows through fissures. Flows that were extruded with only minor explosive activity are documented in Inyo County, California (Eichelberger *et al.*, 1986), Baja California (Hausback, 1987) and under the Snake River Plain (Morgan *et al.*, 1983; Bonnicksen and Kauffman, 1987).

The conical rubble mounds of the Soft Hills either mimic original dome morphology or could be the result of deep erosion of a cylindrical plug. Significantly, the Soft Hills occurrence is bisected by a north-striking fault with about 5 m of vertical offset on the basal contact. The position of the fault implies a structurally weak zone, such as a feeder dyke.

Among basaltic flows on the lower slopes near West Trib are outcrops of pale green, quartz-phyric cobble breccia (Loc. D6). They are inferred to be intrusions that fed felsic flows or domes similar to those forming the rubble mounds.

PETROGRAPHY

Silicic flow rocks are strongly sericitized. The least altered flows contain arcuate streaks that may have been perlitic fractures, although the holohyaline groundmass has completely devitrified. Phenocrysts of potassic feldspar, plagioclase and quartz are present in flows. Those of plagioclase have tabular, euhedral outlines, but have been replaced by sericite, leaving cloudy or opaque cores. Phenocrysts of potassic feldspar show Carlsbad twinning in the replacement mineral (adularia?) and are filled with minute inclusions. Subhedral quartz phenocrysts, 0.2-0.4 mm in diameter, only occur in flows south of Mount Harper and in the Soft Hills. Some quartz has fringes of hematite dust. Single xenocrysts of iron-rich garnet (probably andradite; X-Ray diffraction determination) up to 1 mm across occur in one flow, and sphene, ilmenite and zircon, all up to 0.5 mm long, are common.

ERUPTIVE STYLE AND DEPOSITIONAL ENVIRONMENT

Rock types in Member D indicate mostly non-explosive extrusion; only a single ignimbrite and rare pumiceous clasts were discovered. Thick glassy flows are characterized by low contents of magmatic water and typically extruded from narrow fissures (Eichelberger *et al.*, 1986). Silicic volcanism at Mount Harper began explosively (ignimbrite) followed by extrusion of glassy flows.

Deposits of Member D contain few clues to the environment of deposition. Thin beds of arkose were probably deposited in shallow water, but the presence of rounded felsic clasts in Saddle outlier indicates subaerial erosion, as does the major unconformity beneath most

of Member D. The absence of fine-laminated beds and other indications of extrusion into water suggest that Member D was a subaerial unit.

MEMBER E

DEFINITION AND DISTRIBUTION

Member E consists mostly of massive flows, tuffs and breccia easily distinguished from the equivalent rock types in members A and B by petrography and chemical composition. Because they are aphanitic, rock names were applied from chemistry: Member E flows are andesites by silica content and icelandites in terms of iron enrichment. Flow rocks are very dark, and are not visibly affected by weathering. Distinctive dark blue chlorite alteration permeates brecciated rock, filling vugs and healing early fractures. Widely distributed hydroclastic breccia consists of small, shard-like fragments of Member E flows.

Remnants of flows of Member E (Fig. 4.12) are exposed in the southwestern part of the Mount Harper Complex as towers on ridge crests (Black Towers locality) and broad plateaux along the southern edge. Hydroclastic breccia occurs around the Black Towers, and as a continuous layer near the crest of the ridges north of Lower West Trib, beneath Red Hill, as well as Baldie and Snowcap outliers.

CONTACTS AND DIMENSIONS

Member E overlies felsic flows and crystal tuff of Member D (Fig. 4.11) and hydroclastic breccia disconformably overlies older units. Felsic flows comprising the Soft Hills locality intrude and overlie the hydroclastic breccia.

Along the southwest edge of the Complex the planar flows of Member E dip southwest and are covered by more gently dipping CD_b dolostone (Plate 6,c). The contact, marked by 1-5 m of black mudstone or gouge, is interpreted as a thrust fault based on the stratigraphic truncation and

Fig. 4.12. Distribution of volcanic members E and F.

structures beneath the CD₅ in Upper West Trib valley (see Plate 4,b), although older-over-younger strata relationships are not present here.

Member E is not exposed in a complete section, but south of Black Towers the flow succession appears to be 200 m thick. The distribution of flow remnants suggest they once covered at least 12 km². The hydroclastic breccia, up to 60 m thick beneath Snowcap outlier, probably extended over an area of 15 km² in the northern part of the Complex.

LITHOLOGY

Three rock types are identified in Member E: massive and pillowed lava flows, hydroclastic breccia and subvolcanic intrusions.

Massive flows are smooth-topped and 10-20 m thick, separated by thin or discontinuous layers of recessively weathering flow breccia. In one place the base of a flow reveals fragments of compositionally similar lava, and flow tops are wrinkled and ropy like pahoehoe. Vesicles are sparse, except in the upper few metres of some flows, which contain abundant stretched bubbles. Epidote alteration and calcite precipitate are typically concentrated along flow contacts.

Less extensive pillowed flows appear to be local accumulations where massive flows encountered small water bodies. The largest occurrence is 20 m thick and is associated with hyaloclastite and isolated pillow breccias. Pillows displaying bulbous mattress and bolster shapes (Fig. 4.5) are typically 1 m in diameter and reveal budding on top surfaces. They are not tightly fitted together. Unique to Member E pillows are vesicles, 1 to 3 mm in diameter, concentrated beneath thin rims and altered to bright blue chlorite. Some vugs are filled with celadonite and coloured agate. Breccias containing broken pillow fragments locally drape pillows, and volcanic conglomerate consisting of monolithic, subrounded cobbles indicate a high energy depositional environment.

Hydroclastic breccia is easily distinguished by delicately shaped black clasts, prominent in a blue, opalescent chalcedony matrix. Clasts are extremely angular and internally fractured, thus the breccia is named in accordance with the classification of Fisher and Schminke (1984, Chap. 12). Continuous cliff exposure along ridge spurs (e.g. Loc. E99) show that this breccia was formed by progressive disintegration of massive flows, probably as a result of thermal quenching as the lava came into contact with water. Although a continuous gradation exists, breccia textures are described as 'coarse' and 'fine' end-members.

Coarse hydroclastic breccia is exposed south of Lower West Trib, where massive flows grade laterally over several metres into large blocks that are chaotically jumbled together within a matrix of comminuted fragments and former glass shards. Farther from the intact flow, the blocks themselves are partly disaggregated, and are incipiently fractured (Plate 6,g). Crude layers are graded from large to small fragments, typically arranged like foreset bedding, with sloping trains of cobble-sized clasts.

Fine hydroclastic breccia consists of angular, 0.5-5 cm clasts of aphanitic vesicular and formerly glassy basalt, but includes rare, quartz-phyric, flow-banded siliceous fragments (Plate 6,h). Clasts are poorly sorted but planar cross-stratification and grading to smaller clasts occurs near the top of some of the 0.5-2.0 m beds.

Subvolcanic intrusions are compositionally correlated with Member E. Two outcrops in valley bottoms (Loc. E73 and E1) consist of black, aphanitic rock (diorite in thin section) compositionally similar to Member E flows. Although at least 330 m below the lowest gently dipping flows, these resistant exposures contrast with surrounding chloritized flows and breccias of Member B. The lack of vesicles and surface flow:

features, as well as the apparent sub-circular extent (contacts are recessive fracture zones) support an intrusive interpretation. They are considered deeply eroded feeders for overlying Member E flows.

Sills (Loc. E118-1, and E144-6), correlated with Member E by chemical analysis, are 5 m thick, massive, orange-weathering and columnar jointed. Although not recognized as Member E in the field, they are clearly distinguished from Member B rocks by thin section study.

STRUCTURE

Massive flows of Member E on spurs south of Black Towers are erosional remnants of broad tabular sheets or lava floods (Plate 6,d). Although the flows have been tilted, this area topographically resembles the 'Badlands' of the Columbia River Flood Basalt Province (Miocene) in central Washington state and other plateau lava occurrences.

PETROGRAPHY

Member E andesite is extensively crystallized, in contrast to basaltic members A and B, and also lacks olivine microphenocrysts and abundant mafic silicate minerals. Characteristically the flows are diktytaxitic, now filled by quartz and secondary clay minerals. The texture illustrates an advanced stage of crystallization (or draining of remaining fluid lava) as the flow cooled. It is typical in continental flood basalts (e.g. Schminke, 1967).

Massive flows are intersertal, consisting mostly of subhedral, tabular, 0.5-1 mm long plagioclase (An_{40-45}), some amphibole crystals and interstitial glass (Plate 3,g). Irregular 2-3 mm sub-isotropic patches, 2-3 mm across, could be melt pockets (e.g. Philpotts, 1982). Quartz blebs (< 1 mm) between plagioclase crystals are probably micro-amygdules. Idiomorphic and skeletal magnetite grains up to 0.3 mm (5 %) and minor ilmenite are primary accessory minerals, with minor secondary pyrite.

Pillowed flows have a hyalopilitic groundmass with skeletal, albitized plagioclase (An_{10}) that typically forms 1 mm radial clusters. Quartz blebs are arranged in axiolitic nodes, which suggest that they were formed by late stage crystallization, because they resemble textures described by Lambert (1974). Iron oxide and very dark chlorite obscure most primary textures in the groundmass.

In the sills, stubby plagioclase (An_{40} in 8 determinations) forms an intergranular matrix, with irregular-shaped grains of yellow-green pleochroic amphibole. Granophyric intergrowth is visible at the edges of some plagioclase, and anhedral quartz occupies interstices.

The dark blue alteration of glassy portions of flows, as well as shards in hyaloclastite breccia, was determined by X-ray diffraction of several samples to be iron-rich chlorite. White fibrous albite and prehnite fills vugs.

DEPOSITIONAL ENVIRONMENT AND ERUPTIVE STYLE

The extent of intact lava floods suggests that the southern part of the Volcanic Complex, represented by the deeply eroded Member B structure and rhyolite flows, was subaerially exposed during eruption of Member E. The fluid lavas spread southward on land, but to the north they came into contact with water. The hydroclastic breccia apron extended at least 4 km north of Lower West Trib, suggesting that a considerable volume of lava entered the Harper Basin; this implies pronounced topographic relief along the shoreline. Pillows characterized by abundant vesicles immediately beneath their chilled rinds are only found near the inferred vent areas and probably formed in shallow pools on the gently sloping land surface.

MEMBER F

DEFINITION, DISTRIBUTION, CONTACTS AND STRUCTURE

The youngest volcanic rocks in the Mount Harper Complex include mafic to intermediate massive and pillowed flows, breccias and tuffs, as well as compositionally similar dykes and sills. They weather orange to brown and are diabasic; etched surfaces show intergrowths of feldspar crystals. Intrusions of Member F physically resemble those of Member B, except that they contain minute quartz amygdules. Chemically they more closely resemble Member E (Chapter 7). Although many outcrops in the Complex were mapped by their appearance from a distance, only closely examined rocks and chemically analyzed specimens could be positively identified as Member F; thus some outcrops of this unit have probably been included in Member B.

In areas of poor and discontinuous exposure, flows are not easily distinguished from sills. Host rocks beneath and above sills are darkened and altered, whereas breccia layers and conformable contacts indicate flows.

The distribution of the known occurrences of Member F is shown in Figure 4.12. Layered tuffs were encountered in three isolated outcrops; in the extreme northwest part of the Complex (Loc. F1), on the plateau to the south of Upper West Trib valley (Loc. F2) and east of the Soft Hills (F161). West of Snowcap outlier, Member F flows form a resistant ridge-capping layer. Columnar-jointed sills of diabase are 2-10 m thick south of Tango Tarn (F126-3,5) and at Locality F5.

Dykes, correlated with Member F on the basis of chemistry, are vertical, north-trending and spaced 100 m apart in the central valley north of Lower West Trib. They outcrop below the tabular flows of Member F on the ridge crest to the north, but at Locality F4 a dyke appears adjacent to a 5 m thick pillowed flow, with abundant coarse breccia at the contact. It could represent an exhumed vent. Another intrusive-

extrusive centre is exposed in the southwest corner of the Complex (Loc. F3), where closely spaced dykes are surrounded by compositionally similar flows and breccias. Layering of pillows and rippled flow surfaces suggest outward flow from a 17 m wide basaltic dyke. They are included in Member F because flows overlie hydroclastic breccia of Member E. Silicified veinlets imply later hydrothermal alteration.

A sill (F121-7) up to 9 m thick in the paraconglomerate beds of the Upper Harper Group on Baldie outlier (Plate 1;e,f) pinches out westward over 150 m. Twenty metres above, a pillowed flow of Member F (F121-10) overlies maroon argillite and is overlain by tan-weathering, laminated dolomitic mudstone. Farther east, the same flow overlaps an older surface of chloritized pillows of Member B. Clearly the flow was extruded during deposition of the Upper Harper Group, and the sill may be an invasive flow (Wright *et al.* 1973) that intruded into unconsolidated sediment.

Red Hill is a 600 m high rubble mound north of West Trib (Plate 6,f) consisting mostly of breccia and probable intrusions chemically correlated with Member F. The smooth slopes and conical shape of Red Hill contrast with the typically craggy landforms produced by other mafic members of the Complex. Red Hill and a smaller conical hill 300 m farther east composed of petrographically similar breccia, are probably erosional remnants of a tuff cone. Underlying each is hydroclastic breccia (Member E), overlain at least locally by unaltered dolomitic conglomerate of the basal Upper Harper Group. Other tuff beds in Upper West Trib valley are parallel to layering of underlying block and ash breccia (Member C).

The total thickness of Member F was not determined. The accumulation of breccias at Red Hill is at least 180 m thick, but contains at least some sills. The tabular flow at the top of Baldie

outlier is up to 20 m thick. The duration and extent of Member F activity is unknown because tuff and pyroclastic breccia as well as thin lava flows occur in the Upper Harper Group beyond the study area. A red-weathering hill, about 6 km northwest of the North Harper Fault (outside the study area at $64^{\circ}47'N$; $140^{\circ}06'W$) contains similar diabasic rocks beneath or interbedded with the Upper Harper Group. If this occurrence is Member F (the texturally similar rocks have not been analyzed), it implies that volcanic rocks to the northwest and possibly flows in the Upper Tindir Group (Young, 1982) correlate with this unit of the Mount Harper Complex.

LITHOLOGY AND PETROGRAPHY

Flows east of Baldie Outlier are planar and locally columnar jointed. Individual flow units, separated by thin cracks between jointed palisades, are 2-7 m thick and locally reveal ellipsoidal vesicles concentrated near the top. Some flows are pillowed, consisting of short, thick tubes (60-80 cm diameter) that locally contain drainback cavities and shelves.

The rock is fine- to medium-grained, locally seriate and plagioclase-phyric (1-2%), with visible mafic crystals (augite). Most flows contain diktytaxitic vugs filled with quartz or celadonite (Plate 3,h). The flow on Baldie Outlier (F121-10) contains tabular plagioclase partly altered to clay, with uraltized subophitic pyroxene and 5-15% interstitial quartz in thin section. Some feldspars have albite twins (An_{10}). Moderately pleochroic, ragged-edged crystals with parallel extinction are probably biotite.

The breccia of Member F is poorly exposed. On Red Hill, rubble blocks consist of indurated vesicular lapillistone fragments characteristic of unwelded basaltic pyroclastic deposits. Other loose blocks, however, reveal diktytaxitic and subophitic textures suggestive

of intrusive rock. Red Hill is considered to be a layered accumulation of tuffs and sills.

Medium-grained samples from Red Hill (either flows or shallow intrusives) contain a felted mass of plagioclase with interstitial augite grains, quartz and ilmenite. Plagioclase crystals, 1-3 mm long, commonly have skeletal forms. Slender, ragged crystals with parallel extinction and vague sector zoning are tentatively identified as aegerine-augite. The brown, formerly glassy matrix is rich in opaque crystallites (magnetite?) and contains cavities into which plagioclase crystals project (Plate 3,h). This diktytaxitic texture suggests a high volatile content in the lava at the time of extrusion.

Tuffs are recrystallized, partly silicified and rendered bluish by chlorite alteration. They are distinguished from granular flows by broken crystals of feldspar (10%) and quartz (1%), and most contain fragments of broken lava and altered glass shards. Subtle layers 30-120 cm thick are distinguished by grain size variations (ash- to medium lapilli- size) and the relative proportion of chlorite to sericite in the matrix. At the base of the tuff (Loc. F1), a thinly laminated 10 cm layer, now cherty, with minute chute-and-pool structures and cross-stratification could be a surge deposit.

Tuff at Locality F3 consists of chloritic, formerly vitric shards and angular basaltic and flow-banded felsic fragments in a silicified matrix. Some clasts are bounded by broken vesicle walls, as if explosively fragmented. This tuff also contains irregular clasts of medium-grained mafic and felsic rocks, typically with thin alteration rims. The variety of exotic clasts in this pyroclastic deposit could have resulted from explosive eruption.

Dykes (typically 5-7 m wide) and Sills (up to 10 m thick) have grey-green fresh surfaces and are mostly fine-grained. Some medium-

grained cores display 1 mm spots that are probably altered pyroxene. Thin sections show equant plagioclase in subophitic clusters around birefringent pyroxene with interstitial opaques (probably titanomagnetite). Patches of matrix are mostly sericite, with microscopic quartz amygdules.

DEPOSITIONAL ENVIRONMENT

The various field relationships with sediments of the Upper Harper Group and eroded Member B edifice suggest that Member F was extruded during the period when the Harper Basin transgressed over the northern part of the Volcanic Complex. Lavas were probably subaerially erupted but flowed into water; the tuffs were likely produced in shallow water or from emergent cinder cones. The lateral continuity of the underlying hydroclastic tuff and tabular nature of the flows suggest gently sloping terrain and shallow depressions. The lack of clasts larger than pebble size in beds stratigraphically above Member F suggest that the ancient volcano had been extensively eroded. The wide distribution of north-trending dykes suggest westerly extension (and possibly subsidence) provoked these extrusions.

SUMMARY OF VOLCANIC ACTIVITY

Table 4.4 shows inferred landforms and eruptive style of each member deduced from their deposits.

Table 4.4 (facing)

Table 4.4. Summary of deposits, volcanic facies and inferred environment and activity.
Volcanic facies are after Williams and McBirney (1979).

Stage and Member	Rock type	Interp. deposit	facies	Remnants of orig. structure	Environment; Indicators	Type of volcanism (Macdonald, 1972)
Late stage (F)	Lithic tuffs	waterlain tephra	flank (proximal)	Red Hill: tuff and sill complex	shallow subaq.	effusive (preatomagmatic)
	Diabasic flows Sills	lava floods invasive flows		Tabular flow on ridge crest	some pillows	extruded from breaching dykes
Andesite cone and shield (E)	Massive flows	degassed lava floods	central to flank	plateaus in south of Complex	subaerial	Plains volcanism (Gretley, 1983)
	Hydroclastic breccia	shatters in contact with water	littoral debris apron	layer beneath Snowcap, Red Hill	shallow to deep subaq.	foreset bedding
Silicic complexes (D)	Massive lava flows Arkose	thick flows/domes crossbedded tuff	central flank	col. jtd. layers rubble cones horizon in south	subaer. erosive	Non-explosive pressure-fed extrusions
	Ignimbrite	welded ash flow			subaerial layering	Initial Plinian

Stage and member	Rock Type	Interp. deposit and environment	facies	Remnants of orig. structure	Environment	Environment Indicators	Type of volcanism (Macdonald, 1972)
Volcaniclastic /Pyroclastic	volc. conglom	beach slumps	distal	Plateau south of Upper West Trib.	subaerial	rounded clasts	
(C)	Block and ash tuff-breccia Lapillistone-tuff	pyroclastic flow tephra	flank flank	remnants on eroded edifice layers S of Soft Hills locality	shallow subaq.	foreset beds reverse size grading; vesicular	Vulcanian?
Island volcanism	hem. sandstone limestone	reworked tephra marginal reef	flank	West of Saddle	shallow		
(B _{H.M.})	Basaltic flows Hem. lapillistone, tuff	aa lavas cinder, agglutinate	(proximal-central?)	beneath Saddle outlier towers N of Saddle outlier	subaerial "	Irrreg. vesic. no laminae scoriaceous	Hawaiian to Strombolian
Shoaling seamount	Sills, dykes Large pillows Hyaloclastite	degassed flows	central satellite? eruptions	near Red Hill S of Soft Hills	shallow subaq.	pillows, hyaloclastite	
(B _{C.Nor.})	Cobble breccia Tuff	fragmented flows vesiculated lava	flank	Mount Harper			Surtseyan (preatomagmatic)
Submarine seamount	Sheet flows Pillow flows		flank	Layered cliffs; shield indicated by 'pillow benches'	subaq.		Submarine effusions
(A)	Aquegene tuff Isolated pillow breccia			E of Mount Harper	subaq.	minute vesicles	

Chapter 3

STRUCTURE OF THE MOUNT HARPER VOLCANIC COMPLEX

The Ogilvie Mountains evolved by northward thrusting and uplift during late Mesozoic contraction (Chapter 2). Shortening in the Mount Harper Complex was accommodated mostly along brittle faults. Primary textures are well preserved away from the faults, and there is no directional fabric or other indication of ductile strain in the volcanic rocks. Large areas of the Complex do not conform to the regional southwest dip, and faults appear to be more prevalent within the Complex than in surrounding sedimentary units.

Although the present structure of the Volcanic Complex represents a contraction regime, shortening is superimposed upon features developed in an extensional setting. Early-formed faults, as weak planes in the crust, have been reactivated during later contraction, sometimes in reverse of the original motion.

Figure 5.1 (in pocket) displays structural data for the Mount Harper Complex; inset diagrams and their captions describe evidence for the age and nature of most faults. This chapter describes significant folds and faults and introduces appropriate structural models. These steps are a prelude to inferring the sub-surface configuration of the Volcanic Complex so that cross-sections may be constructed (Appendix C). Because proximal volcanic units are deposited on slopes their dips and contacts are not planar, thereby hindering recognition of the shape and extent of buried faults.

FOLDS

Mesoscopic folds are rare because brittle deformation characterizes the Complex. Thick volcanic layers moved on faults and so remained internally undeformed. A small anticline (6 m wide, amplitude 4 m) in thin-bedded Gillespie Lake dolostone 2 km SW of Mount Harper

reflects uplift between two large fault blocks. Gentle folding of sedimentary strata in the outliers probably resulted from accommodation of offset along the numerous southeast-trending faults.

A broad syncline is apparent on the east side of Mount Harper (Fig. 5.1,h) because the basal contact of the volcanic rocks dips inward at both the north and south ends of the ridge. Synclinal structure is not apparent in the overlying volcanics because several faults divide strata into large tilted blocks. Steep westerly-striking fracture cleavage and rectilinear jointing of pillow tubes northeast of Mount Harper probably reflect gentle flexing. This syncline plunges gently west (about 6°), as indicated by progressively younger units preserved in the western part of the area.

FAULTS

Vertical fault sets oriented 90° , 140° and 20° with respect to north cut the volcanic rocks, and only the major faults can be traced into surrounding sedimentary strata. The brittle nature of deformation in the Complex, in contrast to broad folds in the overlying and underlying carbonates, suggests that the broadly lenticular Complex concentrated strain during northward translation and uplift of the otherwise well-stratified and homogeneous shelf succession.

Two principal episodes of faulting are recognized in the Complex. Differences in thickness or omission of volcanic and underlying strata indicate the older set; the younger faults have visible traces and offset. Typically the older faults have been reactivated, obscuring the magnitude and direction of their original displacement.

The amount of movement on known faults is limited by the evident continuity of the volcanic units. Because the cumulative stratigraphic thickness is about 1500 m, post-volcanic vertical displacement cannot

exceed this amount. Horizontal translation is undetermined because few distinctive steeply dipping units are presently exposed, and these could not be matched across faults. Where the orientation of the displacement vector is unknown, the measured offset of layered rocks is termed **apparent vertical displacement**.

The general sequence of faulting in the Complex has been determined from cross-cutting relationships (Table 5.1) and are numbered accordingly. Faults are discussed below, with the exception of low-angle thrusts [9] which extend across the region (Chapter 2) and the Early South Harper Fault [1], described in Chapter 3.

The **Late South Harper Fault** [8] cuts volcanic rocks in the southern part of the Complex and CD_b Formation in the east. To the west it is buried beneath an overthrust panel of CD_b dolostone. The fault trace is straight, trending west through gullies and notches across spur ridges; the fault contains black gouge with 10 m long lenticles of quartz and carbonate. Fault displacement is imprecisely determined because individual beds cannot be matched across the fault, but the unconformable contacts of major rock units indicate vertical offset of a few tens of metres (west of Coal Creek) to about 600 m southeast of Mount Harper (Fig. 3.6).

The Late South Harper Fault is parallel to the Early South Harper Fault, and represents reactivation of the same subsurface fault zone. In contrast to the earlier fault, however, its displacement is south-side-down, dropping the South Block (which had remained high-standing throughout deposition of the Harper Group). It probably occurred during Mesozoic contraction and could explain the additional vertical strain resulting from the differential thickness of graben fill, when the entire package was thrust northward. Alternatively, it may have resulted from post-compressive relaxation.

Age	South of West Trib	Central region	Northwest end of Complex
Early to Middle Cretaceous	<p>(19) <u>Low angle thrusts</u> - <u>Inferred</u>; at least 2 panels with CD_0 at the base; observed splays in the Upper Harper Group and truncate Member E flows in the S; thrust is higher in U.H.G. to NW; cuts Late SHF and NHF in W (Fig. 5.1,a,m)</p>		
Middle Jurassic to Cretaceous	<p>(8) <u>Late South Harper Fault (SHF)</u> - observed; drops S part of Complex and CD_0 (Fig. 3.4; 5.1,k,l) may dip northward (trace deflected N in valleys) and include sinistral oblique slip.</p>	<p>(6) <u>Multiple Southeast-striking faults</u> observed vertical offsets, <u>Inferred</u> dextral or oblique displacement. - truncate Member F and U.H.G. - Probably listric SW-dip and curve into westward-striking minor northward thrust - merge with, or cut by Late SHF and NHF.</p>	<p>(17) <u>North Harper Fault (NHF)</u> - observed; drops N side of Volcanic Complex and Upper Harper Group. - poss. listric dip to S; assoc. with or reactivated by thrust (Fig. 5.1,b)</p>
Late Proterozoic pre-Member D	<p>(13) <u>South Harper fault zone</u> - <u>Inferred</u> from locally tilted Member B flows on S side, overlain with angular unconformity by breccia and Member D flows (Fig. 5.1, g); trace obliterated by subsequent reactivation</p>	<p>(14) <u>Harper Ridge fault</u> - <u>Inferred</u> - truncation of hematized flows dropped on N side; - prob. vertical (Fig. 5.1, l) - cut by SE-striking faults,</p>	<p>(15) <u>Tango Tarn Fault</u> - <u>Inferred</u> from presence of Member C on N side only; truncates Member B' in cliffs on S side (Fig. 4.8); - prob. cut by NHF - reactivation unclear</p>
pre-Member C		<p>(2) <u>West Trib Fault</u> - <u>Inferred</u> from the greater exposed thickness of Member B on N side; no major offset of hydroclastic layer (Member E). - possible trace not exposed or reactivated.</p>	
pre-volcanic	<p>(11) <u>Early South Harper Fault</u> - <u>observed</u> - drops Pingulcula on N side, assoc. with S-derived conglomerate, overstepped by volcanic flows. - vertical to steep S dip when tilt restored; trace reactivated.</p>		

Table 5.1. Sequence of faulting and structural relationships in the Mount Harper Complex

The **North Harper Fault** [7] is the northwestern boundary of continuously exposed volcanic rocks. It crosses the high ridge in a 1-2 m wide cleft of black gouge with ankeritic dolostone stringers, forms a shallow rut on covered slopes and aligns with scarps of more resistant rock. North-side-down movement is indicated by the stratigraphically higher beds of Upper Harper Group strata on the north side, and volcanic inliers in valleys. The contact between these volcanics and overlying conglomerate, if projected southward to the fault, indicates about 350 m of vertical displacement from the equivalent contact at the base of Baldie outlier. Traced westward the North Harper Fault is linked with a north-side-down fault marked by a scarp on a CD_b -capped mountain (Fig. 5.1,a). Like the western end of the Late South Harper Fault, it is truncated upward by CD_b unit on a low angle thrust (Plate 4,c).

The **southeast-striking faults** [6] occupy narrow vertical clefts in outcrop and are aligned with many gullies in the highlands north of Lower West Trib. They reveal apparent vertical offsets of up to 40 m where they cut upper volcanic units or the outliers (the lateral slip component cannot be determined).

Numerous dykes are aligned with southeast-striking faults but are older because they do not penetrate the sedimentary outliers. The dyke swarms suggest extension during eruption of Member B, and the deep fractures could have been re-activated during later contraction. The best examples are the Soft Hills and Red Hill, both deduced to have been late-stage volcanic centres underlain by dykes. Both are faulted through their highest points (Plate 6,f). They attest to persistent planes of weakness remaining in the shallow crust from at least Late Proterozoic time.

Although most southeast-striking faults show northeast-side-down vertical displacement, their movement vectors are not constrained by

field relations and estimates depend upon the overall interpretation of fault mechanism. Vertical motion may have been normal or reverse, because the present steep eastward dip of the faults reflects the regional southwestward tilting. The parallel pattern of these faults, however, implies transcurrent motion, best demonstrated by the opposing sense of vertical displacement on the Red Hill fault (Fig. 5.1). Volcanic layers that dip westward on the south side of Red Hill appear offset down to the west, but on the north side, the south-dipping pillow units indicate east-side-down offset. Although the fault plane is not exposed to reveal small scale movement indicators, a component of dextral slip is at least as likely as "scissors" faulting.

Although incompletely understood, the southeast-striking faults are possibly wrench faults, on which the southwest side of the Complex has moved northward, resulting in simultaneous thrust and dextral movement. The best exposed example is on Baldie outlier, where well bedded sediments are offset upward and northeastward on a fault splay (Fig. 5.1,c). Thrust motion is also deduced from portions of these faults which curve southwestward (e.g. Fig. 5.1,d) and large scale repetition of volcanic stratigraphy (Fig. 5.1,f).

The southeast-striking faults were not observed to cross the Late South Harper or North Harper faults. Their junctions are covered but they possibly merge with the easterly striking faults, thus presenting a rhombic fault pattern.

The **Tango Tarn Fault** [5] is not exposed in the talus- and moraine-covered floors of tarn-filled cirques, and it is difficult to locate among dykes crossing the ridge at the northwest end of the Complex. It is inferred from the discontinuous distribution of Member C (Chapter 4) and from the physiographic contrast of the cirque walls. Exposed in cliffs on the south side of the fault is a cross-section of

the ancient volcanic edifice (Plate 4,d) of Member B, but these rocks have no visible counterpart on the north side. This cirque is not so much an erosional feature as it is the exposure of an ancient scarp along which the northern part of the Volcanic Complex collapsed (Fig. 4.8). It is overlain by up to 250 m of Member C breccias dipping moderately southward on the north side of the Tango Tarn fault.

Segments of other synvolcanic faults are inferred from the discontinuous stratigraphy of Members A and B. The **Harper Ridge Fault** [4] is marked by a steep tributary draining westward into Coal Creek, where the sub-horizontally layered, hematized flows and breccia are set against steeply dipping older flows (Fig. 5.1,i).

A synvolcanic episode of movement on the **South Harper Fault** zone [3] indicates reactivation of the Early South Harper Fault. It separates two successions: steeply tilted Member B flows unconformably overlain by Members D and E on the south side (Fig. 5.1,j), from a conformable volcanic section including Member C on the north side. Although only remnants are exposed in valleys, the erosion of the older edifice on the southwest side (see also Plate 6,a) indicates that the South Block remained high while the northern area was down-faulted before extrusion of Member D.

The **West Trib Fault** [2] is assumed to lie beneath the deep valley of Lower West Trib, because the well-bedded Member A on the south side cannot be matched to the north. The north side subsided and was probably onlapped by an additional thickness of volcanic products so that no difference in elevation is apparent in the overlying hydroclastic breccia.

STRUCTURAL SYNTHESIS

Mesozoic faults (Fig. 5.2) show a distinct rhombic pattern within the Complex, and the central block, between the inward-dipping North Harper and Late South Harper faults, represents a synclinal horst. The structure could have been produced by uplift during contraction, or by lateral slip, or a combination; models are presented below.

The Coal Creek Inlier consists of Proterozoic strata that were uplifted as a result of northward translation during the Mesozoic. At the northern edge of the inlier (Fig. 2.3) uplift occurred before earliest Cambrian dolostone was deposited directly on Wernecke Supergroup rocks. This uplifted basement influenced the pattern of Paleozoic deposition and has been termed the Ogilvie Arch (e.g. Douglas, 1970) because it affected Paleozoic sedimentation. Typically in fold-and-thrust belts, basement rocks are not involved because cover rocks slide over them on sole faults. In this case the sole fault beneath the Ogilvie Mountains probably stepped upward, and the shelf strata were thrust upward by this basement ramp. In front of the ramp the succession was shortened and tectonically thickened, and the contraction could have accentuated the syncline at the eastern end of the Complex (Fig. 5.3).

At the west end of Coal Creek Inlier, Proterozoic strata shows less consistent regional tilt, and the Harper group has been overthrust by CD₀ dolostone. Most of the volcanic rocks remain buried, in contrast to the eastern part, where they appear to have been forced upward by contraction. The Volcanic Complex is therefore situated between two structural regimes. The northwest-trending axis of the Volcanic Complex, its westward plunge and the southeast-striking fault set might be a reflection of Mesozoic strain across this zone.

In the alternative model (Fig. 5.2, inset), the North Harper and South Harper faults have a component of transcurrent motion (inferred

Fig. 5.2. Mesozoic fault pattern for the Mount Harper Complex. Inset diagram shows mechanism for rotation of northwest-trending faults between the east-trending North Harper and South Harper faults.

Fig. 5.3. A model for Mesozoic contraction structures. a) Regional sketch map showing different strain on the eastern and western parts of the Volcanic Complex (stippled) by northward impingement against uplifted rocks of the Coal Creek inlier. b) Schematic block diagram shows uplift of central area of the Complex on inward-dipping, east-trending faults.

from their linear traces) during Mesozoic deformation. A conjugate shear couple between dextral transcurrent motion on the North Harper and Late South Harper faults imposes a counter-clockwise sense of rotation on the central part of the Complex. Under this model, the southeast-striking faults are splays in transcurrent fault geometry (e.g. Biddle and Christie-Blick, 1987) similar in space and shape to those bounding the Transverse Ranges of southern California (Luyendyk *et al.*, 1980; Hornafius *et al.*, 1987; Fig. 5.2, inset).

In these models, Mesozoic contraction has modified the primary shape and orientation of the Complex. For example, the presently southeast-striking faults could have originally been north-south extension faults (orthogonal to the half-graben), and regional contraction resulted in their rotation and the northwestern elongation of the Complex. The implications for paleomagnetic studies (e.g. Brunet, 1986) are great. Although the primary remnant component from magnetized flows may be useful to determine differential rotation between fault blocks within the Mount Harper Complex, overall paleo-pole comparisons with other Late Proterozoic igneous rocks (e.g. Park and Aitken, 1986) are unlikely to be valid.

Proterozoic extension structures have been considered previously (Fig. 3.8). Stratigraphy, unconformities and the fault pattern imply correlation with the Hayhook extension event (Young *et al.*, 1979; see Fig. 2.6). The tectonic model is presented in Chapter 8 but the spatial relationship between extension faulting and volcanism is discussed below.

Volcanic vents and flow directions are shown in relation to recognized segments of pre- and synvolcanic faults (incompletely exposed) that are north-side-down (Fig. 5.4). Vents of members D and E are on the trend of the Early South Harper Fault, while eruptive centres of earlier members were further north. Little consistency is shown by flow

Fig. 5.4. Exposed portions of syn-volcanic faults, present attitude of dykes and flow directions used to infer Late Proterozoic structure of the Mount Harper Complex. Numbers indicate the sequence of faulting (Table 5.1) and circled letters show location of vent areas inferred from concentrations of dikes and proximal deposits for respective volcanic members.

directions measured throughout members A and B, although they suggest that the vent area lay to the east, possibly above the dyke complex 1 km southeast of Mount Harper.

The Mount Harper Volcanic Complex overlaps the boundary of an asymmetric basin. Earliest extrusions appear to have been 1-2 km north of its South Harper Fault zone; they could have extruded from fractures in the floor of the half-graben or from the actual fault whose exposed trace has been modified by syn-volcanic erosion or reactivation. Eruptions clearly post-date faulting. Although magma ascent may have been triggered by initial faulting, the Lower Harper Group was deposited before magma reached the surface.

Passive or Atlantic-type margins contain extensional basins that are underlain by listric detachment faults (de Charpal *et al.*, 1978; Beach, 1983). Possible sites of magmatism include leaks along the vertical transfer faults that separate these detachments (Lister *et al.*, 1986; Bosworth, 1987; see Fig. 5.5). Member B dykes oblique to the primary extension direction could reflect the orientation of fractures in the floor of the half-graben. Aspects of the source and deep structure of the volcanics are addressed in Chapter 8.

Late Proterozoic extension geometry probably controlled the location of volcanism, as well as the distribution of depositional environments during evolution of the Harper Group. During the Cordilleran orogeny, some of the older extension faults were reactivated in a contraction regime; thus the position of Mesozoic faults and other structures was determined by Proterozoic elements. The following chapters address the age and composition of the volcanic rocks, which yield clues of the thickness and influence of the crust on magma evolution and tectonism.

Fig. 5.5. A model for Late Proterozoic extension during deposition of the Harper Group: a) sketch map showing present distribution of the Lower Harper and Upper Harper sediments and the possible extent of the fans; b) schematic block diagram showing listric detachment faults and transfer fault between them that could have localized volcanism.

Chapter Six

AGE of the MOUNT HARPER COMPLEX

PALEONTOLOGY

Fossil data for the Coal Creek inlier is recently acquired and only preliminary results are available. The Pinguicula carbonates contain stromatolites and microfossils that indicate its Middle to Late Proterozoic age (J.A. Donaldson, pers. comm., 1987; Mercier and Chauvel, 1987). Trace fossils such as Early Cambrian Cruziana (Mustard *et al.*, in press) occur in siltstone of the Upper Harper Group. The CD₆ dolostone, considered to be thrust over the Harper Group, contains Lower Cambrian Archaeocyathid fragments near the base about 20 km west of Mount Harper (R.I. Thompson and co-workers, pers. comm. 1982).

The trace fossil Oldhamia, of Late Hadrynian to Middle Cambrian age, is common in maroon and green argillite in east-central Alaska (Churkin and Brabb, 1965), in the Wernecke Mountains (Hofmann and Cecilie, 1981; Nowlan *et al.*, 1985) and in the Hyland Group 17 km south of Mount Harper (64°31'N; 139°56'W). The age of the upper part of the Hyland Group is therefore early Paleozoic, but the lower beds, including the limestone, are of unknown age. The Road River Formation has yielded Late Silurian Monograptus and Late Ordovician fauna debris (Green, 1972; p. 49-50) 13 km southwest of Mount Harper, and Middle(?) Ordovician gastropod fragments were found in carbonaceous limestone (M. Orchard and R.I. Thompson, pers. comm., 1983) 15 km southeast of Mount Harper (approx. location: 64°34'N; 139°38'W).

URANIUM-LEAD ZIRCON GEOCHRONOLOGY

The volcanic rocks lack fresh biotite and potassic feldspar, so that K-Ar and Rb-Sr isotopic systems probably no longer reflect the original age of the rock. The discovery during this study of felsic volcanic rocks (Member D) with a small quantity of accessory zircon prompted the effort to obtain an age using uranium-lead methods.

Analytic methods and Description of zircons

A 50 kg sample from a quartz-feldspar-phyric flow (about 70% SiO_2) was collected in 1983 from the ridge 5 km southwest of Mount Harper ($64^{\circ}39'\text{N}$, $140^{\circ}00'\text{W}$; the Soft Hills locality on Fig. 4D.1). Preparation was done by staff of the Geochronology Laboratory of the Geological Survey of Canada, including separation of zircon crystals using a Wilfley table, heavy liquids and a magnet, and was followed by a final cleaning in warm 3M HNO_3 to leach sulfides and impurities. The resulting concentrate consisted of about 20 milligrams of zircons.

Zircon crystals in this sample displayed four distinctive shapes (Fig. 6.1). 1) Long (up to 0.13 mm), narrow crystals (aspect ratio 5-10:1), comprised three-quarters of the total population. They were observed in thin sections of the groundmass of rhyolite flows, but during separation most were broken. Axial regions of some crystals contain minute cavities. 2) Euhedral, doubly terminated crystals, commonly with concentric zoning and clear central regions were probably formed by direct magmatic crystallization. 3) Less numerous, small equant crystals that appear octagonal and crystals with blunt ends (as opposed to the euhedral type with 'spear' tips) that have cores made cloudy by small inclusions. Whole equant crystals were combined with euhedral grains to achieve enough sample for analysis, but broken, cored grains were rejected. 4) A few grains with irregular shapes and rounded corners (top right in Fig. 6.1), may be xenocrysts.

Fig. 6.1. Zircon crystals from Sample 83-TW-D6, from photographs and microscope notes. The different shapes are discussed in the text.

Fig. 6.2. U-Pb concordia plot for zircons from the Mount Harper Complex. Dashed isochron (from York, 1969 regression equation) incorporates all seven points; solid line is best estimate for age of volcanism. Ellipses around points are 1 mean standard weighted deviation error.

The zircons were sieved into three size fractions, (>74, between 74 and 62, and <62 microns). Whole crystals were hand-selected by the author from the two larger splits into additional fractions according to shape (described below). Two of the fractions were air-abraded with pyrite grains, following the method of Krogh (1982).

The fractions were spiked with a mixed ^{208}Pb - ^{235}U - ^{233}U tracer (1984 analyses) or a ^{205}Pb - ^{233}U - ^{235}U tracer (1986 analyses). U and Pb isotopic analyses were performed on a Finnigan-MAT 261 multicollector mass spectrometer equipped with an electron multiplier at the Geological Survey of Canada in Ottawa. U was analysed as the metal ion on double Re filaments and Pb using a single filament silica gel- H_3O_4 technique. Electron multiplier bias corrections were made using calibrations of Loveridge (1986). The magnitude of errors of the $^{207}\text{Pb}/^{206}\text{Pb}$ and U/Pb determinations, as well as the mass fractionation of the mass spectrometer data were estimated from the Geochronology Laboratory records and reproducibility of standard solutions.

RESULTS

The analyses (Table 6.1) were made in the summer of 1984 and in the fall of 1986, using slightly different techniques (above) and different quantities of the same sample. The presence of lower Pb blanks in 1986 allowed analysis of much smaller samples. That portion of the sample remaining after selection of the 1984 fractions was sorted, abraded (R. Parrish, pers. comm., 1986) and analysed. These three fractions (prefixed 86-) contained far fewer crystals than any of the four earlier fractions (prefixed 84-). On the $^{206}\text{Pb}/^{238}\text{U}$ - $^{207}\text{Pb}/^{235}\text{U}$ concordia diagram (Fig. 6.2) points are variably discordant, but those nearest to concordia represent the fractions of the euhedral, sorted and abraded crystals.

Anal. no.	size ¹ , treatment ² weight (mg.)	U (ppm)	Pb ³ (ppm)	206Pb/204Pb Total		208Pb ³ (%)	Ratios ⁵				
				measured	common Pb ⁴ (pg.)		Ages (Ma)				
							206Pb/238U	207Pb/235U	207Pb/206Pb	207Pb/206Pb	207Pb/206Pb
86-1	<62 0.101	220.8	27.18	553	304	13.6	0.1161 ± .21 708.2	1.025 ± .36 716.2	0.06400 ± .30 741.7		
86-2	>74 abr. 0.063	184.2	22.65	263	344	13.8	0.1158 ± .26 706.3	1.021 ± .67 714.3	0.06394 ± .65 739.5		
86-3	<62 abr. 0.023	180.0	22.98	449	73	13.4	0.1206 ± .29 734.1	1.091 ± .41 748.9	0.06559 ± .30 793.4		
84-1	<62 unabr. 3.66	220.6	23.22	187.4	25400	15.1	0.09759 ± .35 600.2	0.8724 ± .6 636.9	0.06484 ± .5 769.0		
84-2	<74, >62 long 1.06	112.0	11.36	140.3	4806	16.5	0.09253 ± " 570.4	0.08193 ± " 607.7	0.06422 ± " 748.9		
84-3	<74, >62 euh, abr 2.49	148.5	18.79	299.2	8721	14.1	0.1181 ± " 719.9	1.0583 ± " 733.0	0.06497 ± " 773.3		
84-4	>74 euh, long 1.13	115.4	12.55	194.8	4070	14.23	0.1018 ± " 625.1	0.8996 ± .6 651.5	0.06407 ± .5 744.1		

¹ size refers to length aspect of zircon crystals in microns.
² abr = abraded; euh = euhedral, euhedral, euhedral; long = slender, prismatic crystals; fractions unlabelled contain clear crystals with slightly turbid or inclusion-filled central regions.
³ Radiogenic Pb from Stelger and Jaeger (1977)
⁴ Total common Pb in analysis corrected for fractionation and spike.
⁵ Corrected for blank Pb and U, common Pb; errors quoted are 1 standard error of the mean in percent.
⁶ Isotopic composition of common Pb based on 750 Ma Pb from the model of Stacey and Kramers (1975).

¹ size refers to length aspect of zircon crystals in microns.

² abr = abraded; euh = euhedral, euhedral crystals; long = slender, prismatic crystals; fractions unlabelled contain clear crystals with slightly turbid or inclusion-filled central regions.

³ Radiogenic Pb from Stelger and Jaeger (1977)

⁴ Total common Pb in analysis corrected for fractionation and spike.

⁵ Corrected for blank Pb and U, common Pb; errors quoted are 1 standard error of the mean in percent. Isotopic composition of common Pb based on 750 Ma Pb from the model of Stacey and Kramers (1975).

Table 6.1. U-Pb analytical data for zircons (Sample 83-TW-D6; Location 64°39'N 140°00'W)

INTERPRETATION

Several chords may be drawn through the data points in an effort to most accurately represent the age of the zircons (Fig. 6.2). The upper intercept of chords drawn through the points is interpreted to represent the age of crystallization and the lower intercept is an imprecise estimate of the date of Pb loss (Faure, 1986, Chapter 12).

Data from the 1984 analyses (Table 6.1), although discordant, yielded an upper intercept date of $777 \pm 27/-33$ Ma. The 1986 results are closer to concordia, but the Pb/Pb date for sample #86-3 is significantly older than any of the first set (793 Ma). If this sample is excluded, the six fractions combined yield a date of $751 \pm 26/-18$ Ma, which is taken as the best age estimate for the rhyolite flow of the Mount Harper Complex.

The abraded zircon fractions (#84-3 and #86-3) exhibit the highest $^{207}\text{Pb}/^{206}\text{Pb}$ ages and plot slightly below the chord drawn through other fractions. Abrasion, commonly applied to remove metamict rims on crystals, increases the proportion of zircon crystal cores, which in these fractions appear to have contained a greater concentration of radiogenic lead than that of whole crystals. Inherited or xenocrystic zircon cores were, however, not observed during sample preparation. If the non-abraded fractions from this sample also contained crystals with older lead in their cores, the upper-intercept ages would lie between the true age of these cores and the younger age when most of the zircon crystallized. Nevertheless, the felsic flow is unlikely to be less than about 700 Ma, because the points lie so close to concordia.

In comparison with other Late Proterozoic dates (Table 3.1) and because of their stratigraphically equivalent position, the Mount Harper volcanics are adequately dated by the U-Pb zircon date to place them among the early deposits of the Windermere Supergroup. It is sufficiently precise to indicate that the volcanics are older, and therefore not related to the continental separation if it occurred about 600 Ma (see Armin and Mayer, 1983; Bond and Kominz, 1984). Furthermore, the isotopic date indicates that the Lower Harper Group must be older than the glaciomarine Rapitan Group of the Mackenzie Mountains (see Yeo, 1981), which passes upward into Early Cambrian strata (Narbonne *et al.*, 1985). The Mount Harper volcanic rocks are essentially the same age as sills in the Mackenzie Mountains (Armstrong *et al.*, 1982) and the Huckleberry volcanics in northern Washington state (Devlin, 1986), suggesting a regional magmatic pulse at the start of Windermere deposition (Chapter 3).

Chapter Seven

WHOLE ROCK CHEMISTRY OF THE MOUNT HARPER COMPLEX

Two magmatic suites have been identified from field work in and chemistry of the Mount Harper Complex: 1) basaltic members A and B, and 2) a heterogeneous suite comprising members D, E, and F. Differences in their physical features extend to their chemistry, which provides clues to the volcanic evolution and tectonic setting.

This chapter discusses whole rock analyses of samples that were collected to determine the original nature of the volcanic rocks. The extent of alteration and primary compositional variations are described first, then the Complex is compared with published results from better known igneous suites and other Cordilleran occurrences associated with the Windermere Supergroup, as well as Ordovician volcanic rocks from the Ogilvie Mountains.

SAMPLING and ANALYTIC METHODS

This study is based on 106 hand samples that were collected to provide representative stratigraphic and spatial coverage of the Mount Harper Complex. Members A, B, D and E were recognized in the field and massive flows of each were deliberately sought. Member C, which consists mostly of breccias, was not sampled. Because Member F was identified later from petrographic study and its distinct chemical character; samples of this unit are probably less representative. The possibility of creating a bimodal suite through sampling bias was reduced by selection of rocks that appeared andesitic in colour or texture. In the basaltic Lower Suite, which constitutes the bulk of the Complex, systematic variation with stratigraphic height was investigated by sampling three sections (samples prefixed #118-, #128- and #144-; locations on Fig. 4.3) and other samples were collected to provide representative areal distribution.

Samples weighing 2-4 kg (largest for medium-grained rocks) were collected from freshest outcrops at pre-selected localities (shown on Figs. 4.3, 4.9 and 4.12, as well as sections). They include cores of pillows, micro- or non-vesicular flows and central zones of dykes and sills. Weathered surfaces, discoloration along fractures, secondary minerals in amygdules and veinlets were removed, and sampling of visible alteration (mostly oxidation and bleaching) was avoided. Thin sections enabled distinction of primary and secondary minerals, and most samples found to consist of silica or calcite were not analyzed.

Major oxide and trace element analyses were done in February, 1984 for the author at the Geological Survey of Canada by X-ray fluorescence (XRF) spectroscopy following standard procedures of the Analytical Chemistry Laboratory. One gram of sample (ground to -325 mesh) was mixed with 5 g of $\text{Li}_2\text{B}_7\text{O}_{14}$ and 0.3 g LiF, and fused to a glass disc within a platinum holder. Major elements were measured on this disc using the wavelength-dispersive method, with readings of 20 min/disc, starting with 100 seconds for $\text{K}\alpha\text{Na}$ and 100 seconds for background Na. Matrix effects resulting from absorption and enhancement caused by excitation of other elements within the specimen were corrected by alpha coefficients determined theoretically by D.M. Rousseau (pers. comm. 1987) of the Geochemistry Laboratory. Minor elements were measured on the powdered sample in a mylar-floored holder using the energy-dispersive method (G.R. Lachance, pers. comm., 1984).

Volatile fractions (CO_2 , H_2O and S) were determined by mixing the sample with oxides of vanadium and tungsten in nickel combustion boats, decomposing them at 950°C in a nitrogen atmosphere, and measuring evolved gases by non-dispersive infra-red absorptiometry (Bouvier and Abbey, 1980). FeO was determined by titration. Fe_2O_3 was calculated from: (Total Fe as Fe_2O_3 by XRF) - 1.1134 FeO (volumetric). Tables of accuracy

and precision are given in Appendix C.

C.I.P.W. normative calculations and numerous graphical presentations were performed by computer at the Geological Survey of Canada (IMALECA program); for these calculations, Fe_2O_3 in excess of $(\text{TiO}_2 + 1.5)$ was converted to ferrous iron (Irvine and Baragar, 1971). In order to minimize differences between samples that result from variations in water and carbon dioxide content, analyses have been recalculated to 100% on a volatile-free basis before plotting on graphs that compare samples. Rock nomenclature, except where noted in figures, follows that of Le Bas *et al.* (1987).

RESULTS

The composition of representative samples from members A, B, D, E and F (Table 7.1) and histograms for all samples (Fig. 7.1) show that the volcanic members have distinct signatures. For example, members D and E have lower MgO and Cr_2O_3 , as well as higher SiO_2 , K_2O , Nb, Y and Zr concentrations than other members. They are hereafter called the Upper Suite. Members A and B comprise the Lower Suite. (Member F, which is similar to Member B in some elements, and to Member E in others, is discussed later.) Complete analyses and normative minerals are listed in Appendix C.

Notwithstanding alkali metasomatism, no samples from the Mount Harper Complex are peralkaline (all contain molar $\text{Al}_2\text{O}_3 > [\text{K}_2\text{O} + \text{Na}_2\text{O}]$). All are hypersthene-normative and most contain normative quartz (Appendix C). Four samples (two basaltic dykes and two flows) from Member B contain normative olivine, but none contain normative nepheline.

Chemical variations with stratigraphic height in the Lower Suite are shown in Figure 7.2. Only the Mount Harper section (#128-) shows trends that could be attributed to compositional evolution. Samples from Member A show decreasing magnesium-to-iron ratios and Cr_2O_3 upward (Fig.

	Member A			Member B		Member D		Member E	Member F			OIT	Cont. thol.	CFB
	13-1 (1)	128-4 (2)	128-10 (3)	158-1 (1)	118-13 (2)	49-1 (1)	138-1 (2)	99 (1)	21-7 (1)	121-10 (2)	BHVO-1			
SiO ₂	53.41	57.23	50.83	48.43	51.36	70.61	77.03	59.51	61.23	53.60	49.9	52.96	54.53	
Al ₂ O ₃	16.07	15.42	14.67	16.46	15.87	14.72	16.85	13.25	14.58	16.31	13.84	13.16	13.72	
Fe ₂ O ₃	2.22	1.30	2.87	1.90	1.88	2.27	1.53	2.14	1.72	1.64	2.74	10.80	3.48	
FeO	7.61	7.60	9.68	8.44	6.89	2.74	1.84	9.08	5.58	6.99	8.55	7.66	8.96	
MgO	5.37	7.79	8.18	10.87	6.87	0.92	0.60	3.57	3.57	4.10	7.31	9.07	3.48	
CaO	11.84	4.74	9.86	11.50	12.63	0.93	1.13	7.51	8.29	12.23	11.33	2.53	6.97	
Na ₂ O	1.16	4.34	1.91	1.16	2.71	2.06	1.23	0.96	2.79	3.06	2.29	0.42	3.03	
K ₂ O	0.97	0.11	0.17	0.31	0.82	4.74	4.14	2.06	1.22	0.97	0.54	1.34	1.70	
TiO ₂	0.98	0.81	1.31	0.57	0.66	0.73	0.41	1.39	0.71	0.71	2.69	0.23	2.26	
P ₂ O ₅	0.11	0.11	0.13	0.04	0.04	0.24	0.20	0.30	0.12	0.11	0.28	0.18	0.36	
MnO	0.19	0.14	0.23	0.19	0.19	0.05	0.04	0.19	0.14	0.16	0.17	0.01	0.18	
S	0.03	0.31	0.15	0.03	0.0	0.0	0.0	0.02	0.02	0.01	0.01	0.03	0.04	
Cr ₂ O ₃	0.02	0.09	0.01	0.09	0.07	0.0	0.0	0.02	0.04	0.05	0.03			
As	4	0	4	1	4	0	5	1	1	7	135	208	680	
Ba	280	100	100	90	150	810	1010	1970	730	520			0.2	
Br	3	3	0	0	2	6	0	4	1	8			1.5	
Mo	2	2	2	7	2	0	0	0	0	4	1	16	19	
Nb	6	5	8	2	2	20	12	17	7	9	19		14	
Pb	12	10	12	7	13	16	9	18	13	16	10	42	47	
Rb	26	0	2	7	19	166	140	34	23	18	420	200	330	
Sr	268	364	111	63	208	52	79	182	119	124	1	4	6	
Th	6	5	4	1	5	20	14	12	11	15	0.4		1.7	
U	0	0	3	1	0	4	0	0	0	7	27	33	40	
Y	34	20	34	24	24	61	31	60	36	40	180	165	185	
Zr	98	73	111	28	31	305	222	296	138	120				

Table 7.1. Whole rock analyses of samples representative of the Less Altered Set (recalculated to 100% volatile-free). Complete analyses are in Appendix C. Symbols in first five columns refer to Fig. 7.5. Columns 6-8 are typical basaltic compositions of an ocean island tholeiite (from Hawaii; Abbey, 1983). continental tholeiite (from Morocco; Bertrand et al., 1982) and continental flood basalt (from Columbia River Group; Abbey, 1983).

Fig. 7.1 (facing)

Fig. 7.1, (con't.).

Fig. 7.1. Histogram of composition of all samples from the Mount Harper Complex. Solid bars represent the Less Altered set; stripes indicate dykes and sills. Samples from the three stratigraphic sections are arranged in stratigraphic order upward. Numbers beside columns are sample numbers in figures and Appendix C.

7.2,a), and MnO and CaO decrease sympathetically in successively erupted flows of Member B. In this and other sections, Nb and Y remain fairly constant throughout the stratigraphy, and variations in most element oxides probably resulted from dilution by secondary minerals, addition or depletion by alteration fluids, inhomogeneous concentrations of phenocrysts, or a combination of these. No differences in concentrations affecting entire sections are apparent, suggesting no regional variations.

The earliest flow in the Mount Harper section also exhibits major differences from the rest of Member A. The lowest sample (A128-1) has the highest TiO_2 , P_2O_5 , Nb, Rb, Sr, Y and Zr contents in the Lower Suite. The second flow above the base of the Complex (A128-2) has the highest magnesium-to-iron and Cr_2O_3 content. The implications of these results for magma evolution are described in a later section on chemical variations.

The boundary between members A and B in these sections cannot be compositionally defined. On average, flows of Member B above the contact are lower in TiO_2 , MnO, P_2O_5 and Zr than those of Member A beneath it. Hematitic flows are enriched in Na_2O and Fe_2O_3 , and correspondingly depleted in CaO and MnO.

In order to investigate differences between members, samples were classified according to their chemistry. Characteristic Zr and TiO_2 contents suitably distinguish most members (Fig. 7.3); initially those samples whose membership was determined by stratigraphic position or physical features were plotted to indicate general fields. Members D and E are enriched two to three times over the Lower Suite in Zr and TiO_2 respectively. Samples of Member A form a line of positive slope, and most of those in Member B have Zr content lower than 50 ppm. In the next step, samples that could not be assigned to members on a physical basis (those marked "x" on Fig. 7.3) were compared with the previously defined

Fig. 7.2. Compositional variation against stratigraphy for members A and B of the Mount Harper Complex. Upward lines connect sampled flows; data points for dykes and sills are connected by horizontal lines. See Fig. 4.3. for location of sections.

↑ a) Mount Harper section (128-)

c) West Harper section (118-)

↓ b) Southwest Harper section (144-)

Fig. 7.3. Ti:Zr diagram for the Mount Harper volcanic rocks, used to classify 'unknown' samples into members.

Fig. 7.4. Al_2O_3 : $\text{Mg}^\#$ diagram for Mount Harper volcanic rocks. Member F samples fall midway between those of the Lower Suite and Member E.

fields, and classified accordingly.

Secondly, a smaller group of less altered samples was selected from the Mount Harper analyses. The **Less Altered** set is a sub-set used to make relative estimates of the types and degree of alteration of all the samples, and in addition provides fairly rigorously chosen samples to represent the Mount Harper Complex in comparison with other extension-related igneous suites. The Less Altered set consists of samples containing lowest volatiles, after those with low ferrous iron [$\text{FeO}/(\text{FeO}+\text{Fe}_2\text{O}_3) < 0.6$] and abnormal $\text{Na}_2\text{O}/\text{K}_2\text{O}$ ratios were excluded. Because all basaltic rocks are hydrated, strict limits on maximum volatile content could not be applied without excluding all samples from some members. The samples and limits for alteration and volatiles are therefore listed in Table 7.2. On the diagrams to follow, Less Altered samples are indicated by filled symbols, and the remainder of the population denoted by open symbols.

Among the major elements, the distinction between the Upper and Lower suites is best shown by the alumina and the magnesium to iron ratio [hereafter $\text{Mg}\# = \text{MgO}/(\text{MgO}+\text{FeO}+\text{Fe}_2\text{O}_3)$; Fig. 7.4]. All rocks of the Volcanic Complex are evolved, in comparison with a primitive magma ($\text{Mg}\# = 65$; Basalt Volcanism Study Project, 1981, p. 418). Members A and B occupy a diffuse field with $\text{Mg}\# = 35-55$ and 14-18% Al_2O_3 , which are typical values for oceanic basalts such as MORB (Mid-ocean ridge basalt; 15.4% Al_2O_3) and OIT (Ocean Island tholeiite; 14% Al_2O_3). Samples from members D and E, with $\text{Mg}\# = 18-32$, have 10-13% Al_2O_3 , in the range of CFB (continental flood basalts; 13.4% Al_2O_3) and continental rhyolites (13% Al_2O_3 ; all representative analyses from Best, 1982, p. 40, 184 and 200). Figure 7.4 depicts primary compositional differences using constituents that constitute a significant proportion of the rock: similar separation between members and trends are indicated by minor and trace element contents in the following figures. The magnesium depletion indicated by

Table 7.2 (facing)

Sample Rock H₂O CO₂ FeO[#] +/- Mg#
type

Member B

129	S	3.6	0.1	.49		48
158-1	D	3.6	0.5	.81		51
97A	D	2.7	0.8	.71		43
128-15	D	1.7	1.3	.80	Na-	45
162-1	D	3.0	1.7	.81	Na-	50
118-13	F	3.0	1.8	.79		43
128-5	D	3.5	2.7	.82		49
118-12	F	3.4	2.8	.77		47
144-5	B	3.1	3.1	.79		46
135-2	Fh	3.4	2.8	.25	Na+	42
128-18	Fh	3.7	2.1	.35	Na+	39
-17	Fh	2.3	0.5	.34	Na+	39
-16	Fh	3.4	0.5	.37	Na+	41
-14	Fh	3.1	5.5	.44	Na+	40
-13	D	4.5	10.2	.81	Cc	49
-12	F	4.5	10.6	.54	Cc	40
124-2x	Fh	4.9	1.0	.48		38
-1x	F	2.8	0.9	.82		45
126-6x	P	4.4	2.8	.62	K-, Na-	48
-4x	F	4.0	10.2	.42	Cc	36
-2x	Fh	2.6	1.9	.29	Na+	28
-1x	F	3.3	0.5	.81	Na+	52
121-3	F	4.8	3.7	.85		40
118-14	P	3.9	1.6	.80	Sl	36
-10	F	3.6	11.4	.88	Cc	44
-9	F	5.8	7.3	.84	Na-	56
117-5x	F	4.4	3.4	.82	K+	35
-4x	F	4.0	6.8	1.00	K+	47
162-2x	P	5.2	7.1	.86	K-, Na-	53
144-15x	F	3.5	4.4	.47		45
-14x	F	6.9	7.8	.86	K-, Na-, Cc	59
-13x	Fh	3.8	10.4	.19	K-, Na-, Cc	42
-12x	F	5.0	11.4	.89	K-, Na-, Cc	51

Member AA

128-3	D	3.9	0.2	.60	K-, Na+	53
127-2	S	4.0	0.3	.40		54
128-2	S	4.2	0.4	.35		55

Sample Rock H₂O CO₂ FeO[#] +/- Mg#
type

Member A

130-2	D	3.5	2.0	.80		44
131-1	D	3.0	2.1	.77		35
128-10	F	4.3	2.1	.71		39
128-4	D	4.6	2.1	.85	Na+	47
118-6	F	3.2	3.4	.77		43
130-1	D	3.7	2.5	.75		41
131-2	D	3.7	2.8	.79	Na-	46
144-11x	S	3.4	3.9	.83	Na-, K-	43
133	P	3.9	4.3	.74		51
117-3	P	4.0	4.7	.82		40
132	P	3.6	5.4	.80		48
128-11	D	4.6	4.2	.77		40
-9	P	3.2	8.8	.86	Cc	40
-8	F	3.5	9.0	.83	Cc	44
-7	P	4.4	13.0	.90	Cc	44
-6	P	4.5	3.9	.79		49
128-1	F	4.2	1.5	.68		34
127-1	F	4.6	6.7	.62		33
97b	P	4.2	8.4	.80	Cc	47
15	P	3.9	12.7	.80	Cc	48
8	P	4.3	6.5	.88		42
124-3x	F	4.1	7.1	.78	Cc	39
118-8	F	4.8	5.4	.80		44
-7	F	4.1	2.9	.77		46
-5	F	5.1	11.6	.88	Cc	51
-4	F	4.9	7.7	.85	Cc	42
-3	F	4.9	9.1	.86	Cc	39
-2	P	3.8	5.0	.91	Sl	44
117-2	P	4.6	3.9	.87		39
-1	F	5.4	2.8	.79		38
158-2x	P	6.8	8.5	.87	Na-, Cc	48
144-10x	F	4.0	6.9	.79		40
-9x	F	5.1	5.5	.85		42
-8x	F	3.3	9.0	.83	Cc	45
-7	P	4.6	7.6	.84	Cc	49
-2	F	4.3	6.4	.85		46
-1	P	5.0	7.6	.89	Cc	46
164-1x	Fh	4.6	4.2	.63		50
-2x	F	4.7	5.6	.60		50
-3x	F	4.3	6.6	.58	Cc	47

Rock type: P = pillow core D = dike
F = massive flow S = sill
Fh = hematitic flow
Bx = indurated aphanitic breccia

+/-: Na+ >3.8% Na₂O; Na- = <1.0% Na₂O
K+ >5.0% K₂O; K- = <0.2% K₂O
Cc = >7.5% CO₂, (calcite; Fig 7.6)

Sample Rock H₂O CO₂ FeO# +/- Mg#
type

Member F

143	P	2.5	1.9	.59	Na-	35
21-7	D	3.2	2.5	.76		33
121-10	P	3.7	4.7	.81		32
126-3	S	3.2	5.6	.83		37
126-5	S	3.1	6.4	.79		25
121-7	F	3.8	6.5	.83		46
161x	Bx	4.6	0.8	.70		40
72-1x	F	2.2	7.2	.67	Na-, SI	25
112x	F	6.0	8.7	.87	Cc	59

Sample Rock H₂O CO₂ FeO# +/- Mg#
type

Member E

99	F	2.9	1.7	.81		24
79x	Bx?	3.2	1.8	.73	Na-	25
71	P	2.6	2.5	.69	SI	17
144-6x	S	2.9	3.3	.77		17
72-2	F	4.2	.8	.72	SI	31
82-4	P	3.3	1.1	.62	SI	22
70	F	4.5	1.7	.81		23
118-1x	S	4.4	4.4	.85		28
73	D	4.2	5.5	.82		23
102x	F	3.8	7.6	.90	Cc	19
104x	Bx	4.1	11.8	.84	Cc	41

Sample Rock H₂O CO₂ FeO# +/- Mg#
type

Member D

138-4	F	2.1	.1	.69	K+, Na-	18
48	D	2.1	.4	.32		15
103	F	2.1	.7	.85	Na-, SI	22
138-1	F	1.8	.9	.54		15
138-2	F	1.7	1.8	.72	K+, Na-	19
166	D	1.9	2.8	.88	K+	12
138-3	F	2.1	1.3	.84	K+, Na-	16
44	F	2.6	1.3	.70		16
154	F	2.1	1.8	.83		14
136	Fx	3.3	2.6	.75		41

+/-: Na+ = >3.8% Na₂O; K+ = >5.0% K₂O; SI = silicified trend, Fig. 7.4.

Na- = <1.0% " K- = <0.2% " ; Cc = >7.5% CO₂, indicates calcite: Fig. 7.6.

$$\text{FeO\#} = \text{Fe}_2\text{O}_3 / (\text{FeO} + \text{Fe}_2\text{O}_3)$$

$$\text{Mg\#} = \text{MgO} / (\text{MgO} + \text{Fe}_2\text{O}_3 + \text{FeO})$$

x = samples (mostly intrusions) placed into members by Ti:Zr content (Fig. 7.3)

p = pillow core

bx = breccia

d = dyke

s = sill

Table 7.2. H₂O and CO₂O content, FeO#, mobilized elements and Mg# for all samples of the Mount Harper Complex. Samples are arranged as: **Less Altered set** (by increasing volatile content), and - Remainder by area (east of Coal Creek; north of Lower West Trib, and south of Lower West Trib, in stratigraphic order.

lower Mg# in the Upper Suite reveals its evolved character and is discussed in a later section.

Harker diagrams (Fig. 7.5) are used to qualitatively demonstrate trends by plotting element oxide contents as a function of varying silica content. The Lower Suite samples contain between 50 and 58% SiO_2 , whereas the Upper Suite has a wide silica distribution (Member D is 70-77%, Member E has 58-64% SiO_2). Although Fe^{Tot} , TiO_2 , Y and Zr remain relatively constant for individual members, Al_2O_3 , MgO, CaO and Cr_2O_3 generally decrease with increasing silica. The broadest scatter in Na_2O occurs at 55-60% SiO_2 , in contrast to K_2O , which varies considerably in samples with $>70\%$ SiO_2 .

ALTERATION

All volcanic rocks of the Mount Harper Complex are altered and contain varying amounts of low grade metamorphic minerals. Glassy portions of mafic and felsic rocks have crystallized to fine-grained quartz and feldspar and phenocrysts in basaltic rocks are pseudomorphed by different minerals.

In mafic rocks, olivine has been completely replaced by quartz and hematite (Plate 3,c), and plagioclase microphenocrysts are replaced by clay minerals. Prismatic crystals in the groundmass are albitized and filled with hematite specks. The matrix of maroon flows is obscured by hematite, and some vesicles unavoidably included in samples contain chlorite and prehnite. Andesitic flows of Member E consist of sericitized plagioclase, and interstices between crystals are filled with quartz.

The freshest felsic rocks are cherty, and thin sections reveal the formerly glassy groundmass consists of cryptocrystalline silica and sericite. Strongly altered rocks, which typically crumbled in the field, were not analysed. Feldspar phenocrysts are cloudy, obscuring primary characteristics such as twinning. Some quartz phenocrysts are partly

Fig. 7.5. Harker diagrams for all samples of the Mount Harper Complex. Solid and numbered symbols denote the Less Altered set. The $\text{Al}_2\text{O}_3\text{-SiO}_2$ plot is Fig. 7.8.

resorbed, but fringes of secondary quartz (Plate 3,c) and spherulites suggest silicification.

Although the type of alteration depends upon the original composition of the rocks and that of the permeating fluids during burial and metamorphism, its intensity appears to reflect primary porosity. Thin flows (<1 m thick), vesicular and fractured interflow zones are visibly more altered than cores of thicker flows, sills and dykes. Quartz veins are rare and no defined alteration zone or hydrothermally altered rock was encountered to indicate an underlying intrusion. Alteration around dykes and sills is confined to sheared contacts.

Petrographic study suggests that these reactions occurred:

Basaltic glass $\rightarrow$ quartz + feldspar + hematite $\pm$ chlorite

[after Deer *et al.* (1966), Jolly and Smith (1972)]

The mafic rocks contain substantial H_2O and CO_2 (see Table 7.2), which implies at least local remobilization of alkalis and other elements. Principal alteration effects suggested by petrography and chemistry are hydration, carbonatization, iron oxidation and sodium-potassium metasomatism.

Hydration involves the formation of hydrous minerals, accompanying devitrification of metastable volcanic glass, as well as low grade metamorphism. Hydrous minerals contain most of the water in metabasalts. Chlorite (13 wt.% H_2O) and sericite (about 9 wt.% H_2O ; Deer *et al.*, 1966) constitute from 5% in some diabase dykes of Member B to 35% (volume

visually estimated in thin sections) of poorly crystallized, formerly glassy and devitrified pillows at the base of Member A. In contrast, felsic flows contain only slightly more water than is normally present in fresh rhyolites.

Many mafic lavas in modern rift zones are water-rich (up to 3.5%; Bailey, 1974). The saturation or migration of water trapped within the volcanic pile may have resulted in pervasive internal alteration during cooling, rather than reflecting a regional metamorphic event. Lithic clasts containing sub-isotropic basaltic glass and fresh plagioclase microlites occur in conglomeratic sandstone of the Upper Harper Group, which implies that alteration was more pronounced within the Complex than in the surrounding region.

Carbonatization is indicated by calcite replacement of mafic minerals and patches of matrix in many altered samples. Medium-grained diabase has both high CaO and high MgO, indicating abundant augite, but other samples show a positive correlation between CaO and CO₂, indicating calcite (Fig. 7.6). High calcite content dilutes the concentration of other elements, but attempts to restore the original composition of some samples by recalculating the analysis "calcite-free" (in addition to the H₂O- and CO₂-free already calculated) were unsuccessful. The attempts failed because the recalculated concentration of other mobile element oxides was higher than that expected for normal basaltic compositions, perhaps because Ca⁺⁺ ion was partly derived from albitization of plagioclase and hematization of pyroxene, rather than introduced by later fluids.

It is important to know the site of the Ca⁺⁺ ion because Murphy and Hynes (1986) showed that normally stable elements such as Ti, P, Zr, Y and Nb are variably depleted from rocks under low CO₂ partial pressures. Formation of metasomatic minerals, such as titanite, release CO₂ (Mason, 1980) and therefore indicate that the HFS (high field

Fig. 7.6. Integrated CaO:MgO, CaO:CO₂ and H₂O:CO₂ diagrams for all samples from the Mount Harper Complex. Samples joined by dashed lines are augite-bearing dykes and sills; "calcite line" indicates the proportion of CaO and CO₂ in calcite.

strength) elements have been altered (Murphy and Hynes, 1986). In the Mount Harper volcanic rocks, these incompatible elements appear to have remained immobile because their contents are reasonably uniform among altered and Less Altered samples from the same member, possibly because the formation of calcite has removed available CO_2 from alteration fluids. These minor elements may therefore be used to assess evolutionary trends in the Mount Harper Complex.

Iron oxidation of Lower Suite rocks occurred during emplacement of some lavas (hematized facies) and also during secondary mobilization. Interstitial hematite is present in massive flows associated with highly vesicular breccias and lapillistone that were probably subaerially erupted. It could have formed in susceptible lavas during eruption or later during weathering. Both synvolcanic hematite formation and secondary coatings on minor pyrite and in amygdules were described by Brunet (1986).

With the exception of the hematized flows, Lower Suite samples exhibit normal proportions of ferrous and ferric iron [$\text{FeO}/(\text{FeO}+\text{Fe}_2\text{O}_3) = 0.7-0.8$; Table 7.2], in comparison with ocean and continental basalts. Hematized rocks contain up to five times more Fe_2O_3 than FeO , but indicate no marked increase in total iron content. Strong iron oxidation is a visible indication of element mobility, however, because flows rich in hematite have higher Na_2O as well as lower K_2O , CaO , Sr and Rb contents than chloritized and Less Altered basalts in the Complex.

Sodic and potassic metasomatism is the major process which affected these volcanic rocks. It has different effects in mafic and felsic samples. Although some hematized basalts are sodic (up to 5.8% Na_2O), most are below average Na_2O content for unaltered basalt (Fig. 7.7). Several basalts, whose matrix is obscured or replaced by hematite, are below the detection limits in both Na_2O and K_2O (0.5% and 0.05% respectively).

Felsic rocks are also depleted in Na_2O but are enriched in K_2O (up

Fig. 7.7. Na₂O:K₂O diagram shows sodic enrichment of hematized samples of Member B; all other samples have lost Na₂O and most have gained K₂O.

Fig. 7.8. Al₂O₃:SiO₂ diagram shows silicified samples with lower alumina than other samples of the same member. Symbols as in Fig. 7.7.

to 7%) relative to fresh rhyolite from the Mount Harper Complex (Fig. 7.7).

Potassic alteration is commonly noted in ancient felsic rocks (e.g. Annells, 1974; Reichenbach, 1986). Secondary potassium enrichment of sea-floor rhyolite leads to K-feldspar + hematite + quartz + illite-montmorillonite assemblages (Munha *et al.*, 1980). Sheppard and Gude (1968) described rhyolite tuff converted to K-feldspar by diagenesis in saline lake beds and cited experimental work that also suggests that under saline conditions silicic rocks devitrify to zeolites and K-feldspar rather than clay minerals. Possibly the felsic rocks at Mount Harper were altered in similar fashion during diagenesis by infiltration of waters from the later marine basin of the Upper Harper Group. Andesitic flows have also lost Na_2O and may contain secondary K_2O formed under similar conditions; thus the major elements should not be used to classify these rocks.

Silicification has not noticeably affected the Mount Harper Complex. Several samples of Member E, however, have low $\text{Al}_2\text{O}_3:\text{SiO}_2$ ratio relative to normal rock types (Fig. 7.8). Chloritized basalt, and samples containing interstitial quartz blebs, are not significantly different in SiO_2 . This suggests that the visible quartz formed from breakdown of within-rock constituents. Replacement of delicate primary minerals by secondary minerals indicates that little volume increase took place during alteration.

In summary, the diagrams show that the Less Altered samples are: 1) more uniform in major mobile element contents, and 2) cluster nearer to the composition of unaltered rocks than the mean of the entire population. Although all samples are altered, some primary mineralogical and chemical features have not been obliterated, and analyses of the Less Altered set are useful in assessing magmatic trends within the Volcanic Complex.

PRIMARY COMPOSITION OF THE MOUNT HARPER COMPLEX

To minimize the effects of alteration, the rocks have been classified using trace and minor constituents that are relatively immobile under low grade alteration (e.g. Pearce and Cann, 1973; Morrison, 1978). These elements show a comparatively narrow spread between contents in samples from the Less Altered set and other samples. The Zr/TiO_2 versus Nb/Y diagram (Fig. 7.9) illustrates their consistent relationship within members; A and B are sub-alkaline basalt to andesitic basalt; Member D is rhyodacite and members E and F are andesites. An alkalic chemical suite is not indicated, in contrast to volcanic rocks related to the Windermere Supergroup in Utah and Idaho, as well as compared to early Paleozoic volcanics of the Selwyn Basin (e.g. Harper and Link, 1986; Cecile, 1980; Roots, in press).

Although classification based upon plots using alkali elements is suspect for altered rocks, the Na_2O+K_2O versus SiO_2 (Fig. 7.10) and AFM (Fig. 7.11) diagrams for Mount Harper samples are consistent with the results of the trace element plots, albeit more scattered. As suitable trace element data are not available for most other Cordilleran volcanic rocks, these diagrams offer the only concise means for their comparison with Mount Harper.

The distribution of members on the Jensen Cation plot (Fig. 7.11,c) is similar to that shown by the AFM diagram. Some of the Less Altered samples from the Lower Suite contrast with Member E by their separation into 'Mg-rich tholeiitic basalts' and 'Fe-rich tholeiitic basalt' respectively.

The iron enrichment of Member E actually represents depletion in magnesium content (Fig. 7.4). The $Mg^{\#}$ provides a relative estimate of the extent to which a basaltic magma has evolved (e.g. Cox et al., 1979); primary magma derived from the lower crust or mantle is estimated to have

Fig. 7.9. Zr/TiO₂:Nb/Y diagram (Winchester and Floyd, 1977) for Mount Harper volcanic rocks. Other data is from Harper and Link (1986; 'Bannock') and Roots (in press; 'C-O volcanics').

Fig. 7.10. Total alkali : silica diagram (Le Bas, 1986) for Mount Harper volcanic rocks. Note the similar distribution of members as in Figure 7.9.

MOUNT HARPER COMPLEX

Fig. 7.11. AFM and Jensen diagrams for the Mount Harper volcanic rocks. The dividing line between calc-alkalic and tholeiitic fields is from Irvine and Baragar (1971) and Jensen (1976). The proportions are in wt % on the AFM diagram and cations on the Jensen plot. Note the similar distribution of all members on the two diagrams.

Mg#=65 (Basaltic Volcanism Study Project, 1981; p. 418). The lowest flows of Member A are Mg# 54, which is a relatively primitive value. The average for the Lower Suite is Mg# 45, in contrast to the Less Altered samples of Member E which have Mg# 13-25. Samples of Member F have intermediate Mg# between the Lower Suite and the more evolved Member E.

Samples from the two suites show marked contrast in incompatible element contents, which is illustrated by chondrite-normalized abundance diagrams or **spidergrams**. Incompatible elements are those constituents less easily incorporated into early rock-forming minerals, and as a result they are more regularly distributed in various extrusions from the same magma source, or show slight increase in residual phases. Although some of these elements are mobile in fluids during alteration (K, Rb, Sr and Th), and the very low concentrations of other elements strain the detection limits of analysis by the XRF technique (see Appendix C for discussion), the Less Altered samples define a distinctive spidergram pattern for each member (Figs. 7.12, a,b,c,d and e). Together the patterns indicate that the Mount Harper Complex is enriched in large-ion elements 20-100 times over a chondrite standard (chosen to represent the primitive earth composition; Cox *et al.*, 1979; Thompson *et al.*, 1983). The patterns are not smooth, but contain 'peaks' in Th, La and Sm, and 'troughs' in P and Ti. The Upper Suite is more enriched in Ba, Sr, K and Th, as well as in Sm and Zr, and samples from Member F (Fig. 7.12,c) have intermediate values between those of the Upper and Lower suites. The significance of the patterns for various members, and comparison with those of other Cordilleran suites, as well as representative oceanic and continental basalts (Fig. 7.12,f,g,h) are addressed in the following discussions.

Fig. 7.12. Chondrite-normalized, ($Y_n=15$) incompatible element abundance diagrams (spidergrams; see Thompson *et al.*, 1983) (p.1 of 2) for some Mount Harper volcanic rocks. See Appendix C for notes on construction. Note that the Upper Suite members have similarly shaped diagrams and are more enriched than the Lower Suite.

Fig. 7.12, con't. Spidergrams for Mount Harper and other basaltic rocks (p.2 of 2) for comparison: e) Member F (lines) falls between shaded envelopes for Upper and Lower suites; f) Comparison of Member E pattern with other volcanic rocks related to the Windermere Supergroup; g) representative oceanic basalts, and h) representative continental basalts.

A. CHEMICAL VARIATION AMONG LAVA FLOWS

Three aspects of the primary composition of rocks of the Mount Harper Complex are addressed here: 1) variation with stratigraphic position in the Lower Suite; 2) similarity in minor element concentrations among members D and E of the Upper Suite; and 3) immobile element content of Member F, which is compositionally intermediate between the two suites.

1. In a simple fractional crystallization model, the most primitive flow would be expected at the base, but in the Mount Harper section (Fig. 7.2,a) the lowest flow contains greater K_2O , TiO_2 , P_2O_5 , Nb, Y and Zr, as well as lower $Mg\#$ and Cr_2O_3 contents than subsequent flows. Alternatively the higher lithophile element content may reflect a relatively small percentage of partial melting at the outset of magmatism. Subsequent flows that ascended in established conduits are more primitive in composition than the initial extrusion, and were likely of higher temperature due to the insulating effect of chilled magma along the walls.

The most significant trend in Member A is that of decreasing $Mg\#$ (from 54 to 38, in nine flows over 370 m) with height in the Mount Harper section (Fig 7.2,a). The depletion suggests olivine crystallization depleted magnesium from later lavas.

The other sections of the Lower Suite show minor trends of decreasing Zr in Member A (Fig. 7.2,c) but no clear variations. Lowest flows have $Mg\#$ and Cr_2O_3 contents similar to those well above the base of the Mount Harper section and lack olivine microphenocrysts. The distribution of feeder dykes and flow directions (Fig. 5.6) indicate that the sections to the west and southwest of Mount Harper were probably more distant from the main vent. They therefore could represent later

eruptions that travelled farther than initial extrusions. Where flows are small in volume, compositional variations would be expected to be most pronounced among earliest extrusions, especially where sampled close to the source.

The boundary between members A and B reflects changes in eruptive style, water depth, and probably magma supply rate and degree of crystallization. As the edifice developed, magmas became increasingly crystallized before extrusion.

2. In the Upper Suite, the spatial association of vents and interlayered flows of members D and E indicates that these extrusions were related in time. Additional sampling is needed, the present analyses suggest a bimodal series. Of 21 Upper Suite samples, only one lies between 64% SiO_2 (Member E) and 70% SiO_2 (Member D; Fig. 7.5).

Despite their major element differences, samples from members D and E have similarly enriched spidergram patterns (Fig. 7.12 c,d). It is unlikely that the patterns could have resulted from parallel crystallization histories because the members have different primary mineral assemblages. The concentrations of most of these elements are not controlled by crystallization of the early-forming mineral phases (Thompson *et al.*, 1983), and therefore are more indicative of original or parental composition. The high Zr and K_2O , as well as similar Nb and Y contents imply a genetic relationship between the two members; they could be differentiated components of a single magma. Temperature or mineral density gradients in the magma chamber might have resulted in early felsic extrusions from the upper part and later Member E flows from deeper levels.

3. Although rocks of Member F physically resemble those of Member B, the presence of intervening Member E indicates that these extrusions were separated in time. Samples from Member F are more evolved (Mg# 25) than the Lower Suite, and akin to Member E because they contain a low proportion of mafic minerals and have enriched contents of Rb, Th, K, Na, Sm and Zr.

The combination of characteristics of members B and E may have resulted from mixing between a relatively primitive magma (similar to that which fed the Lower Suite) and evolved magma (such as that which produced Member E). The hybrid magma interpretation is supported by the spidergrams for Member F samples, that lie between the general patterns for the Upper and Lower Suites (Fig. 7.12,e). Member F is thought to be a product of a small, infrequently replenished magma chamber, such as are typical of those in rift zones (Sanford and Einarsson, 1982; Mahood and Baker, 1986) and characterize resurgent volcanism (e.g. Barberi *et al.*, 1977). Repeated extensional faulting and an additional magma pulse could have promoted eruption of Member F.

In conclusion, magma mixing cannot be confirmed or disproved on the basis of the present analyses, but a single source region beneath the Mount Harper Complex would be consistent with the available data. In order to attempt quantitative models of magmatic evolution, a better understanding of primary composition and chemical variation of members D, E and F would be required. Models based on the behaviour of a few trace elements are likely to yield ambiguous results (e.g. Watson, 1982).

B. MAGMA TYPES INFERRED FROM CHEMISTRY

On the basis of regional geology and stratigraphy, some tectonic settings, and therefore magma types, are clearly inappropriate for the Mount Harper Complex. It is unlikely that the Complex was deposited on oceanic crust, because shallow water carbonate both underlies and succeeds the volcanic rocks. Furthermore, most members include features indicating local subaerial exposure. It is also unlikely that the magmas were arc-related, because there is neither evidence of sedimentary assemblages typical of island arcs nor of contraction deformation and uplift. Extensive products of explosive volcanism and subduction-related rocks of Late Proterozoic age have yet to be found in the Canadian Cordillera.

Discriminant diagrams (e.g. Pierce and Cann, 1973; Floyd and Winchester, 1975; Mullen, 1983; Thompson *et al.*, 1983) that employ the abundances or ratios of trace and minor elements have been widely used to determine the tectonic setting of ancient basaltic rocks. Their application, however, to continental basalts and to those related to extension has produced conflicting or misleading results (Morrison, 1978; Holm, 1982; Arculus, 1987). The wide variations among magmatic suites in continental extensional settings reflect inhomogeneous source rocks and differing depths of formation, paths of fractionation, degrees of partial melting and contamination (Mohr, 1982; Hildreth *et al.* 1985; Duncan, 1987), which may obscure primary chemical characteristics of this particular tectonic setting. Chemistry is therefore only one of many factors upon which the recognition or interpretation of ancient rift settings should be based (Sengor and Burke, 1978; Pearce, 1987).

Spidergrams, however, are useful for distinguishing major rock compositional groups. Figures 7.12,g,h illustrate typical spidergram patterns for representative samples of oceanic and continental

tholeiites. Mid-oceanic ridge basalts (MORBs) are the least enriched rock group, with less than ten times chondrite abundance of lithophile elements (Ba, Rb, Th and K) and about ten times enrichment of elements whose ionic radius exceeds that of Sr (this contrasts with the spidergram pattern is common among those examples discussed here; thus the elements to the right of Sr are hereafter called the HFS (high field strength) elements; see Murphy and Hynes, 1986). MORBs generally display a convex upward, or humped, pattern.

In contrast to MORBs, ocean-island tholeiites (OITs) are more enriched, with lithophile and HFS elements 10-20 times chondrite and higher Nb, La and Sr. The resulting spidergram pattern is also humped, but lacks the positive slope of MORBs. Similar enrichment occurs in continental flood basalts (CFBs), but the general pattern differs from OITs in two ways: 1) it has a negative slope, and 2) a trough at Nb is commonly present. The negative Nb anomaly, found throughout large suites like the Deccan flood basalts as well as in widely dispersed dykes such as those related to opening of the Atlantic (e.g. Moroccan tholeiites) is considered indicative of magma-interaction with continental crust (Dupuy and Dostal, 1983).

Envelopes enclosing the Mount Harper samples do not produce smooth patterns, but all have generally negative slopes, and the Lower Suite shows a pronounced Nb trough. On this basis, as well as on structural and stratigraphic evidence, the Mount Harper volcanic rocks indicate a continental, rather than an oceanic setting (differences in the pattern for the Upper Suite reflect more siliceous rock types that are inappropriate to compare with basalts on these diagrams). Lower Suite basalts have lower HFS element contents, particularly P and Ti, than MORBs, which may indicate derivation from a comparatively depleted mantle source.

Although spidergrams are effective in displaying basaltic compositions, the temporal progression of rock types is also important to magma genesis (Marsh, 1987). In particular, the sequence from alkalic to transitional and tholeiitic rocks in some rift suites has implications for the evolution of source regions (e.g. Wendlandt and Morgan, 1982; Kampunzu *et al.*, 1986, and references therein).

At Mount Harper, the following changes in magma types are inferred from the succession of Lower Suite to Upper Suite lava flows:

- comparatively primitive (high Mg#) to evolved (low Mg#) magmas
- overall increase in silica-saturation (olivine basalt to andesite)
- increase in lithophile and HFS elements.

Initial extrusions (Member A), high in Al_2O_3 , constructed a thick pillowed section, suggestive of oceanic character, whereas final flows of Member E, in physiography, petrography, low Al_2O_3 and higher TiO_2 , resemble continental floods.

Perhaps equally important are trends not indicated by the chemistry of the Mount Harper Complex. In the East African rift, the appearance of MORB-like basalts in later stages implies an approach to oceanic conditions, but without continental break-up (Mohr, 1982). Magmatic evolution at Mount Harper appears to trends away from depleted oceanic magma types. Furthermore, alkali enrichment, although difficult to assess because mobile elements have been disturbed by alteration, is not indicated by primary mineralogy or Nb/Y ratio of the Mount Harper rocks (Fig. 7.9). Neither the tholeiitic-to-alkalic trend shown by some island volcanoes (e.g. Hawaii; Chen and Frey, 1983) nor general increase in silica-undersaturation that is characteristic of some continental rifts (e.g. East Africa; Bailey, 1974; Mohr, 1982) are indicated for Mount Harper.

C. COMPARISON WITH VOLCANIC SUITES RELATED TO THE WINDERMERE SUPERGROUP

In Chapter 3, contemporaneous igneous rocks in Alaska, the Mackenzie Mountains, the southeastern Canadian Cordillera and in Washington, Idaho and Utah were described because their geologic setting and stratigraphic position are analogous to those of the Mount Harper Complex. Published analyses of these rocks are few, typically incomplete, and no occurrences have been systematically documented, so that the full compositional variation remains undetermined (e.g. felsic tuff and ultramafic intrusions have been reported but not analysed; R. Scammell, pers. comm., 1987). Because these occurrences, now known to be similar in age and volcanic style, are probably part of an igneous province related to extension during initial deposition of the Windermere Supergroup, it is worthwhile to compare them chemically with the Mount Harper Complex.

Despite the shortcomings of discriminant diagrams for the identification of extension-related igneous suites (cited in the previous section), bivariate and triangular plots are useful to concisely display chemical variations among rocks that have a common geographic or tectonic setting. The AFM diagram (Fig. 7.13) shows Late Proterozoic Cordilleran occurrences compared with Less Altered samples of the Mount Harper Complex and the Keweenaw volcanic suite (a 1.1-Ga mid-continental rift succession exposed around Lake Superior; Green, 1983). The sills in the Tsezotene Formation in the Mackenzie Mountains (Armstrong *et al.*, 1982), as well as flows from the Tindir Group (Alaska; Stewart, 1972) and the Irene and Huckleberry formations (southeastern British Columbia and Washington state; Devlin *et al.*, 1985) are iron-rich tholeiites, distributed in the general field of Member E of the Mount Harper Complex. The Keweenaw volcanic rocks are wholly tholeiitic with a lithologic

Fig. 7.13. AFM diagrams for continental extension-related volcanic suites; all proportions in wt.%. Late Proterozoic Cordilleran data from Armstrong et al. (1982), Young (1982) and Devlin et al. (1985).

distribution (from rhyolite to olivine tholeiite) resembling that of the Less Altered rocks of the Mount Harper Complex.

Among minor elements, lower TiO_2 distinguishes the Mount Harper Complex (up to 1.4%) from extension-related basalts (1.6 - 4%; Fig. 7.14,a). Only Member E samples are included in the field for "within-plate lavas" designated by Pearce (1980). The Zr-Ti-Y and $\text{MnO-TiO}_2\text{-P}_2\text{O}_5$ discriminant diagrams (Figs. 7.14,b,c) show similarity among Huckleberry/Irene and Tindir groups, but only samples from Utah-Idaho and the Paleozoic volcanics from Ogilvie Mountains occupy the "within plate" field outlined by Pearce and Cann (1973) for samples from the Karoo and Deccan floods as well as the East African rift. Note also that the low Zr content of some samples from Member B results in their falling outside the fields for normal rocks on Figure 7.14,b. These samples fall in the same area that is occupied by metamorphosed volcanics whose so-called immobile element (Ti, Y and Zr) contents have been disturbed by alteration (Murphy and Hynes, 1986). It is most unlikely that this has occurred at Mount Harper because these diabase samples are among the freshest rocks obtained. The reason for their distribution is unknown, but the problem illustrates the hazards of using discriminant diagrams for basalts related to continental extension.

Spidergrams, available for the Huckleberry/Irene and Utah-Idaho samples (Fig. 7.12,f) do not fit a single pattern recognized for one of the major rock groups. The spread in relative abundance of lithophile elements reflects varying degrees of alteration, although many have HFS element patterns similar to continental tholeiites. Like Mount Harper samples, most show a negative P anomaly; in contrast, they lack the Nb trough. On the basis of spidergrams it is not clear which member of the Mount Harper Complex most resembles these southern occurrences.

Fig. 7.14. Discriminant diagrams, used here to show distribution of volcanic units in the Windermere Supergroup. No Tindir data is available for b), nor Idaho-Utah data for c).

From the limited data available on the composition of the volcanic rocks related to the Windermere Supergroup (Figs. 7.9, 7.12, 7.13 and 7.14), the following conclusions may be drawn:

1. Although none of the occurrences exactly matches a major magma type, they appear to be most closely related to continental tholeiites (see also Devlin *et al.*, 1985).
2. On the basis of low Mg#, high TiO₂ and other HFS elements, most Windermere-related igneous rocks were derived from relatively evolved magmas.
3. The Lower Suite at Mount Harper appears to be unique among the occurrences, by virtue of a) its relatively primitive origin (as indicated by higher Mg# and less enrichment of incompatible elements), and b) its large variation in composition.
4. Most occurrences, with the exception of the Mount Harper Complex, reveal indications of alkalic affinity. Carbonatite complexes (Hoy and Pell, 1986) and alkalic volcanic tuff (Scammell, 1986) in south-central British Columbia are local evidence of strongly undersaturated magmatism in Late Proterozoic time.

In conclusion, the chemistry of the Mount Harper Complex is consistent with either a marginal basin or intracontinental rift setting.

Chapter 8

TECTONIC IMPLICATIONS of the MOUNT HARPER COMPLEX

The Late Proterozoic igneous rocks related to deposition of the Windermere Supergroup are discussed here with respect to: 1) the extension setting, and 2) possible mechanism of extension-related magmatism.

TYPE OF EXTENSION

Several lines of evidence link the Mount Harper Complex with an extensional event; results from the present study are summarized in Table 8.1. These rocks can be correlated with clastic sediments and contemporaneous igneous rocks along the length of the Cordillera (Devlin *et al.*, 1985), indicating that the event affected an area of continental proportions. Three aspects are considered here: a) the probable nature of the underlying crust; b) the tectonic setting of the extension basin (intra-continental or marginal); and c) whether rifting was of passive or active type.

a) The Harper Basin (the structure filled by the Harper Group) has the key elements of an asymmetric rift: an uplifted side (the South Block; Chapter 3), a steep boundary fault and clastic deposits whose thickness tapers northward away from the fault. As Burke (1980) and others have noted, well-preserved continental rifts are those that have failed to develop completely or have been abandoned, because rifting is merely an intermediate stage in continental break-up. After separation, abandoned rifts along new continental margins are buried beneath thick shelf-edge deposits. The structure of the Harper Group is an example of a minor rift setting that is spatially intact and has been later well exposed.

Table 8.1. Evidence for Extension in the Harper Group

STRUCTURES

Early South Harper Fault: Steeply dipping, north-side-down, 1600-3500 m cumulative vertical displacement; active between Pinguicula deposition and initial volcanism.

Derivation of conglomerate wedge from uplifted terrain across fault.

Dyke swarms: parallel dykes indicate tensile stress in lithosphere.

Vent areas may be aligned with faults suggest planes of crustal weakness.

CHANGES IN DEPOSITIONAL ENVIRONMENT

Disruption of stable shelf setting: Change from initial shallow water and subaerial deposition to deposits reflecting deeper water and sediment-starved, euxinic conditions.

Continuous subsidence: subaerially erupted volcanic rocks buried by pillow flows and subaqueous clastic sediments derived from uplifted older rocks.

COMPOSITION OF IGNEOUS ROCKS

Tholeiitic magmas: relatively primitive initial composition and rapid ascent indicated; negative Nb anomaly caused by influence of continental crust (Dupuy and Dostal, 1984).

Bimodal suite: differentiation into rhyolite and andesite may have occurred in magma chamber in relatively thick crust.

The occurrence of shallow water carbonate overlapping the syn-rift volcanic rocks demonstrates the continued existence of continental crust beneath the area, after extension led to formation of the Harper Basin. It indicates that rifting ceased before continental crust separated, as does the absence of MORBs (which would indicate formation of new oceanic crust). In contrast to the Harper Basin, the region south of Mount Harper later became the Selwyn Basin. It is underlain by extensively thinned continental or transitional crust, as indicated by its structural response to Mesozoic contraction; it was appreciably shortened by isoclinal folds and thrust faults (Chapter 2).

b) The distribution of tectonic elements or regional paleogeography of the western margin of North America during the Proterozoic has been determined chiefly from stratigraphy and sedimentology (Stewart, 1972; Young *et al.*, 1973, Young, 1977; McMechan, 1981). A divergent western margin of cratonic North America may have developed in latest Proterozoic to Cambrian time (Van der Voo *et al.*, 1984; Root, 1987). The date of continental separation, 575 ± 25 Ma, has been calculated from subsidence rates based upon post-rift cooling (Bond *et al.*, 1984). The present location of the continental land mass (or sliver) that separated from North America has not been established, it may have been incorporated into Siberia (Sears and Price, 1978), or central Australia (Bell and Jeffers, 1987) or included among rocks later accreted to the western margin of North America (Tempelman-Kluit, 1979). The lack of distinctive units and difficulty of matching asymmetric rift margins (e.g. Lister *et al.*, 1986), as well as obliteration by later deformation and units ensure that paleo-reconstructions remain speculative for some time. Toward elucidation of this problem, minor rifts such as the Harper Basin that may have been precursors to continental separation but were later abandoned provide indications of the age and nature of syn-extensional deposits.

From data presented in chapters three and seven the Windermere Supergroup is considered to have accumulated in an intracontinental setting prior to continental separation. That it was flanked by an opposite margin (to the west, using present-day coordinates) is indicated by northeastward paleocurrents both in the Ogilvie Mountains (R.I. Thompson and co-workers, pers. comm., 1986) and in British Columbia (Gabrielse, 1972; Root, 1987) as well as abundant quartz grit, which must have had an alternate source than the carbonate shelf succession in the northeast. Detrital zircon of Middle Proterozoic age (Erdmer and Baadsgaard, in prep.) and 2.0-Ga crustal rocks to the west of present Windermere deposits (Evenchick *et al.*, 1984) are further evidence that the Windermere Supergroup was both derived from, and situated upon, uplifted continental basement.

Because it appears to have been bounded to the west, the Windermere basin might have been 1) an crustal sag not related to regional tectonic event; 2) a back-arc, or fore-deep basin; or 3) one or more rifts. The first possibility is eliminated because syndepositional tectonism is widely indicated, both by block faulting (Lis and Price, 1976; Root, 1987 and the Early South Harper Fault), and synchronous volcanism over 3000 km along strike. Although a back-arc basin setting cannot be disproved with present evidence, it is less likely because of the length of the basin (>4500 km) and its paucity of volcanoclastic sediment; there is no evidence for a nearby volcanic arc. The third tectonic setting, that of intracontinental rifts, is supported by asymmetric clastic deposits, such as the Harper Group and its less-well-exposed equivalents in southeastern British Columbia (Root, 1987), as well as syndepositional faults. Furthermore, the Harper Basin lies parallel to the margin of the Selwyn Basin, and could have been its precursor that was later abandoned as extension localized further outboard (southward) from the craton.

c) The variety of modern and Phanerozoic continental rifts have been qualitatively described according to two end-members. These are active (or high-volcanicity) rifts and passive (or low-volcanicity) rifts, based upon the temporal relationship of volcanism to rift structure, as well as the distribution and abundance of volcanic rocks (Sengor and Burke, 1978, Neumann and Ramberg, 1978, Palmason, 1982, Kaz'min, 1984). Active rifts, for which the East African (Bailey, 1974) and Rio Grande (Riecker, 1979) rifts are examples, appear to be driven by transmission of thermal energy into the lithosphere by rising asthenosphere (e.g. Burke, 1980; Wendlandt and Morgan, 1982). Rift valleys of the active type and their shoulders tend to be elevated continental regions, emphasized by tholeiitic lava plateaus. Active rifts are characterized by high heat flow and magma reservoirs at shallow crustal levels (e.g. Sanford and Einarsson, 1982).

In contrast, passive and failed rifts appear to result from tensile lithospheric stress (Burke, 1980). These features reflect continuous subsidence, and generally contain isolated volcanic centres that were generally preceded by faulting and clastic sedimentation. Volcanism appears to be a passive response to rifting, as magma rises in fractures through the crust. These rifts occur within continents (Baikal rift: Kisyelev *et al.*, 1978), as well as both at high angles and parallel to their margins (Benue Trough of French West Africa; Fitton and Dunlop, 1985 and along the Atlantic coastline; Burk and Drake, 1974; Bedard, 1985).

Both active and passive rifting can lead to continental separation, as demonstrated by the East African-Red Sea (Bonatti, 1985) and Atlantic (Manspeizer, 1978) examples, and both contain bimodal volcanic suites and alkalic rocks.

Sedimentary facies and the volcanic style indicate that the Harper Basin was a passive rift. Mount Harper Complex was an isolated volcanic centre that clearly post-dated faulting and initial clastic deposition.

3

Major unconformities and evidence of uplift indicative of active rifts is absent. Other Windermere deposits have also been interpreted as passive rifts (Scammell, 1986). The spatial and temporal relationship of volcanic to sedimentary rocks of the Windermere Supergroup further imply that volcanism was a passive response to extension. Although volumetrically insignificant, volcanic rocks are synchronous and widely distributed, demonstrating that basaltic magma was mobilized beneath this zone of extension.

EXTENSION-RELATED MAGMATISM: MECHANISM AND TIMING

Rift-related volcanism is not confined to rift valley floors and boundary faults. For example, the larger volcanoes in Africa, including Kilimanjaro and Mount Cameroon, are situated hundreds of kilometres from the Kenyan rift and Benue Trough respectively. They are termed *off-axis volcanoes* and thought to be related to rifts by underlying detachment faults (Bosworth, 1987). A schematic cross-section (Fig. 8.1,a) indicates the spatial relationship between the rift valley and off-axis volcanoes, which conforms to the profile of asymmetric rift basins (compare with Fig. 3.8). Magmatism occurs in two general locations; in the rift valley itself, where the crust is under extension, and through lithosphere subjected to tensile stress above the detachment fault.

Off-axis magmatism could result from shear heating along the detachment fault (Wernicke, 1985), melting or underplating of lithosphere exposed to hot asthenosphere by the fault (Lister *et al.*, 1986; Bosworth, 1987) or decompression of the asthenosphere (Bonatti, 1985). Rise of the magma is localized by tensile fractures or leaky transfer zones (Lister *et al.*, 1986; Sykes, 1978) and fault-controlled sub-basins commonly occur beneath off-axis volcanoes (Bosworth, 1987).

b) Syn-rift section, Ogilvie Mountains

Fig. 8.1. Schematic cross-sections showing magmatism related to extension for a) northern Kenya rift, and b) similar structures of Late Proterozoic age inferred to lie beneath the Ogilvie Mountains, as they may have looked before mesozoic shortening. The original distance between the Harper half-graben and the Hyland Basin is speculative.

Although the Mount Harper Complex occurs in a half-graben, in terms of the regional extension model it is situated northeast of the zone where separation eventually occurred (either beneath the Selwyn Basin or further outboard). In this sense, volcanism at Mount Harper may be considered an off-axis centre. The steep dip of the precursor Early South Harper Fault, and the limited amount of extension indicated beneath the Harper Basin, support this interpretation.

Figure 8.1,b depicts the Mount Harper Complex as an off-axis volcano situated above the intersection of a detachment fault with the lithosphere-asthenosphere boundary. The Early South Harper Fault bounding the Harper half-graben provided a zone of lower lithostatic pressure for ascending magma.

The chemical evolution of the Complex, from relatively primitive flows to more evolved compositions, implies deep penetration of the fault, allowing initial rapid rise of asthenospheric magma. Later the Harper Basin appears to have stagnated, indicated by cessation of activity on the Early South Harper Fault, and the increasingly lithospheric character of the magma (implying little or no replenishment by mantle-derived magma). Other igneous occurrences related to basal deposits of the Windermere Supergroup are evolved magma types, suggestive of slow ascent or seepage along extension fractures, rather than the deep penetration of a fault.

The discrepancy in the age of rifting, between that isotopically determined for initial Windermere sedimentation (750 ± 30 Ma; see Table 3.1) and the time of continental separation calculated from back-stripping (550-600 Ma; Armin and Mayer, 1983; Bond and Kominz, 1984) has been treated extensively (Devlin *et al.*, 1985; Devlin, 1986). It is unlikely that rifting occurred continuously from 750 Ma, and the

existence of two discrete extension events cannot be evaluated without further regional study. The structural and sedimentological features of the Windermere Supergroup are suggestive of multiple episodes of extension overlapping in time and along the intracratonic basin.

Abandoned rifts, such as that filled by the Harper Group, were probably common and contain initial Windermere deposits, but most lie covered by middle Windermere sedimentary units. In some areas, such as the Deserters Range of north-central British Columbia, sedimentation began later (< 700 Ma; Evenchick, 1985). Episodic extension is also implied by the onlap of major depositional units, such as the Kaza Group over the Horsethief Creek Group (Pell and Simony, 1987; see also Figs. 2.5, and 3.9), and similar to the shifting depocentres indicated by the spatial relationship of the Harper and Hyland groups. Marginal sub-basins, such as that beneath the Mount Harper Volcanic Complex, were probably abandoned as rifting became more localized toward a zone of later continental separation.

Chapter Nine

GEOLOGIC HISTORY AND CONCLUSIONS

The Harper Group, which consists of a lower conglomeratic wedge, a bimodal volcanic complex and an upper transgressive sedimentary succession, records the development of an isolated, extension-related basin atop the Late Proterozoic carbonate platform. The Mount Harper Volcanic Complex includes six members. The first three are basaltic and record the growth of a subaqueous to subaerial volcano, and were followed by rhyolite domes, andesitic floods and diabase intrusions. The Harper Group accumulated in a rift basin. The rift was earlier than, or coeval with continental separation that formed the divergent western margin of cratonic North America further west, but at this location did not extend to the extent that oceanic crust was produced.

The evolution of the Mount Harper Volcanic Complex is deduced from the sequence and composition of the volcanic rocks; this allows inferences to be made about underlying structural and magmatic processes. The completeness of the geologic record, however, depends upon the type and quality of deposits that are preserved. Although a continuously subsiding rift environment affords excellent preservation potential, subaqueous deposits are more likely to be preserved than are unconsolidated, subaerially erupted rocks such as pyroclastic ejecta. Furthermore, inferences about types of volcanic activity are dependant on deposits that vary in volume and likelihood of preservation.

Phreatomagmatic eruptions, for example, produce huge volumes of subaqueous breccia from relatively small, subaerial vents; the physical abundance of these deposits belie the brief activity causing them when compared to the life of the volcano (see Ayers, 1982). Similarly, the products of a brief catastrophic eruption could overshadow, or wholly destroy the thinner accumulation of a longer and more significant time period.

This account summarizes interpretations based upon the stratigraphy, structure and chemistry of the Mount Harper Volcanic Complex. The succession of volcanic landforms is schematically shown in Figure 9.1.

The geologic history of the Harper Group has four stages:

- I. Regional faulting and deposition of the Lower Harper Group;
- II. Seamount and island basalt volcanism (Lower Suite);
- III. Felsic and andesitic subaerial volcanism (Upper Suite); and
- IV. Deposition of Upper Harper Group.

In comparison with more recent analogues of the rock types, the first three stages are inferred to have had a duration of less than 30 Ma, while the Upper Harper Group probably represents more than 100 Ma. The fourth stage, is examined in Mustard (in prep.). The first three stages are described here according to a) depositional record; b) structural regime; and c) magmatic evolution. Evidence for them is summarized in Table 9.1.

Stage I: Faulting and deposition of the Lower Harper Group

Depositional Record: The broad continental platform, capped by Middle to Late Proterozoic Pinguicula carbonate, was block faulted, and local uplifts shed debris flows into adjacent depositional basins. The Early South Harper Fault was repeatedly active, and a clastic fan system, the Lower Harper Group, developed northward from the fault scarp. Deposition ceased when the uplifted South Block was reduced by erosion and the entire area became submerged.

Structural regime: The Early South Harper Fault was probably a steep or north-dipping, normal fault developed along older structures. It and other faults inferred to lie further south preceded deposition of the Lower Harper Group and Mount Harper volcanic rocks. With respect to a constant water level, structural movement appears to have been initial regional uplift, followed by local subsidence, resulting in the Harper half-graben.

Fig. 9.1. Sequence of volcanic landforms in the evolution of the Mount Harper Volcanic Complex.

Stages	Synpositional Structures and Paleogeography	Sedimentary and Volcanic deposits	Petrologic and Geochemical evolution
Upper Harper Group	Subsidence to deep water, euxinic conditions Uplift of northern margin indicated by southward deposition of clasts of Ping-uicula and older units; this conglomerate overlies eroded Lower Suite rocks.	Euxinic shale basin, sand flows. Paraconglomerate of local derivation, mostly debris flows	
III. Bimodal volcanism (Upper Suite)	Upper Suite subaerial in south; subaqueous hydroclastic apron to north; indicates south margin of half-graben remained high.	Member F diabasic flows intercalated with paraconglomerate and mudstone Tholeiitic floods (Member E) and rhyolite flows and domes (Member D), erupted on south flank of Complex.	Upper Suite enriched in lithophile and incompatible elements; low MgO Member E and F plagioclase and magnetite crystals; diktytaxitic Member D rhyolite is glassy with quartz, feldspar, spinel phenocrysts
II. Seamount volcanism (Lower suite)	Hundreds of metres of subaqueous basaltic pillows; overlain locally by subaerial cinder cones and vesicular flows. Progressive subsidence; lack of intercalated sediments indicates adjacent fault scarps. Basalt pillow flows overstep Early South Harper Fault.	Hematized vesicular flows, lapillistone breccias Cobble and broken-pillow breccias Pillow and massive flow units Basal flows over-run sandstone with sediment deformation; on south side of fault unconformably overlies Wernecke dolostone	Member B: pl- (ol)- aug- pl Latest diabase has lowest lithophile element content. Member A: ol megacryst- pl- (aug)- glass First extrusion high in lithophile elements; contains olivine megacrysts
I. Half-graben formation (Lower Harper Group)	Early South Harper Fault indicated by uplift and progressive erosion of units on south side, adjacent conglomerate wedge on north side.	Lower Harper conglomerate is proximal fan conglomerate, mostly debris flows derived from southern scarp, alternating silt-sandstone beds near top are more distal.	

Table 9.1. Summary of evidence for paleo-geographic and magmatic history

Magmatic evolution: The Early South Harper Fault could have triggered volcanism by its penetration to mantle depths; its interpretation as an normal fault suggests that the crust was subjected to tensional strain.

The causal relationship between faulting and volcanism is demonstrated by their coeval activity and the superposition of the Complex on the western extent of the Early South Harper Fault. The steep plane of the upper part of the fault could have provided a route for ascent of magma, which rose through the crust during deposition of the Lower Harper Group. The basaltic composition of the magma implies its derivation from the mantle, but the earliest extrusion has a higher lithophile element content than later flows, either as a result of contamination during ascent or a smaller degree of partial melting at its source.

STAGE 2: Seamount and island volcanism (Lower Suite)

Depositional history: Basaltic volcanism constitutes four phases which varied in duration and were not synchronous at different vents in the area. The earliest highly fluid flows overlapped submerged clastic fans and formed subaqueous shields of tubular pillows, breccia and aquagene tuff (Member A). Pillowed units 200 m thick that comprise the base of the western part of the Complex were erupted later than those in the east.

The second eruptive phase consisted of shallow water effusions that produced short flows, and cobble breccias that resulted from their disintegration (Chloritic facies of Member B). It succeeded the first phase by predominance of central vents, and steeper sided cones developed. Large blocks in breccias were possibly mobilized by explosive erupted from vents choked by lava rubble.

In the third phase, Surtseyan eruptions are represented by lapillistone, vesicular pillows and breccia (Member B). Explosive activity in shallow water, and rapid erosion of emergent deposits resulted in broad subaqueous platforms around vents.

The fourth phase of activity produced aa flows, agglomerate and cinder-like lapilli (Hematitic facies of Member B). Vulcanian and Strombolian eruptions inferred from these deposits produced short-lived volcanic islands from subaerial vents. Lavas that flowed into water were transformed into large pillows, hyaloclastite and breccia containing hematized clasts.

The maximum height and shape of the large volcanic structure is unknown because it underwent erosion concurrent with volcanism. Volcanic conglomerate and breccias (Member C) represent mass wasting and explosive eruptions.

Structural regime: The earliest known flows are located several kilometres north of the Early South Harper Fault, suggesting extrusion began on the floor of the half-graben. The subaerially deposited conglomerate is overlain by thick pillows lava successions, indicating subsidence during volcanism. Islands were formed when volcanic accumulation outstripped the rate of subsidence, but preservation of lithified cinder attests to rapid subsidence and burial. Toward the end of Lower Suite volcanism, two contrasting depositional regimes began. The area south of Lower West Trib was uplifted, possibly as a result of inflation of the volcanic structure by sills, while the northern part subsided, forming the Harper Basin.

Magmatic evolution: The Lower Suite lavas were derived from a primitive basaltic parent magma which evolved by fractionation of olivine. Early flows were glassy; later flows contain plagioclase, pyroxene and magnetite. The higher proportion of breccias and greater thickness of later flows could have resulted from the increasingly crystallized magma, with higher viscosity and reduced supply rates. Explosive eruptions occurred in shallow water from exsolving gases or mixing with steam, and subaerially erupted flows and breccias were hematized.

Stage 3: Bimodal volcanism (Upper Suite) and deposition of the Upper Harper Group

Depositional Record: Thick rhyolite flows (Member D) were extruded unconformably over basalts in the southern part of the Complex. Arkosic sediments, crystal tuff and a single ash flow occurrence at the contact indicate initially explosive eruptions. Thick flows and possible domes appear to have been subaerial. They were covered by andesitic floods (Member E) that flowed east and southeast. Large vesicular pillows indicate the presence of shallow pools, but flows also disintegrated in larger water bodies, forming hydroclastic breccia. The Harper Basin, which transgressed the northern part of the Complex, accumulated debris flows derived from both volcanic and uplifted carbonate source areas, as well as dolomitic sandstone, and carbonaceous siltstone. The Upper Harper Group succession represents increasing depth and low energy conditions.

Structural regime: The contrast in depositional record between the north and south parts of the Complex indicates a topographic discontinuity. The high-standing southern area resulted from inflation by sills, rise of Upper Suite magmas or reactivation of the Early South Harper Fault. Silicic dykes and inferred vents for Member E occur along the buried trace of the fault, evidence of a tectonically weakened zone, although fault offset cannot be documented. Hydroclastic deposits which spread northward from disintegrating Upper Suite flows in the south emphasize topographic relief along the shoreline of the Harper Basin.

Magmatic regime: After the Lower Suite volcanism a dormant period, indicated by unconformities and sedimentary intervals, resulted in differentiation in a shallow magma chamber. Initial explosive eruptions of the Upper Suite consisted of volatile-rich rhyolite that was probably concentrated in the upper part of the chamber. Fissure-erupted felsic flows (Member D) and andesitic floods (Member E) have similar

incompatible element abundances, suggestive of derivation from the same magma source. The final pulse (Member F) produced pyroclastic eruptions and emplaced sills. Its composition indicates that relatively primitive magma mixed with magma that might have yielded the earlier extrusions.

In conclusion, the rocks of the Harper Group record depositional environments that varied in water depth and are consistent with passive extensional tectonics. Unlike clastic sedimentary successions which accumulate in response to uplifted source areas, however, volcanic products are extruded both on land and underwater, and therefore record relative changes in water level. The geologic history of this area may be considered a dynamic balance between six natural forces. Three of these give rise to topographic uplift: 1) faulting, 2) accumulation of volcanic rocks, and 3) inflation by sub-volcanic magma, while three resulted in preservation of the sedimentary successions: 4) erosion and burial, 5) downfaulting, and 6) local downwarping from lithostatic loading or evacuation of underlying magma chambers. The rock record at Mount Harper indicates a depositional setting that became more complex with time, until volcanism and extension ceased.

Most of the Mount Harper Volcanic Complex is considered the product of a single igneous pulse (Member F could be a minor resurgent phase). The ascent of magma in a deep crustal break is indicated by the linear concentration of inferred eruptive centres along a linear zone. Relatively primitive early flows ascended directly, while later products were more crystallized, and were probably slower to reach the surface. There is no evidence for assimilation of crustal rocks.

The South Harper Fault zone appears to have been fundamental to the development and preservation of the Harper Group. Its initial movement resulted in deposition of the basal conglomerate wedge, provided a weak zone for magma ascent and possibly hosted magma chambers that gave

rise to the Upper Suite. Little if any lateral extension is indicated, but it became the northern limit of a zone of extended or thinned crust, later overlain by the Selwyn Basin. During Mesozoic contraction much of the South Harper Fault zone was reactivated, and may have served to deflect deformation around the Volcanic Complex, thus preserving its primary structures. The South Harper Fault is therefore implicated in the creation of the rift, the volcanic and clastic succession within it, their preservation, and uplift to present view.

Chapter 10

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1977: Stratigraphic correlation of Upper Proterozoic rocks of north-western Canada; *Canadian Journal of Earth Sciences*, v. 14, p. 1771-1787.
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1979: Middle and Late Proterozoic evolution of the northern Canadian Cordillera and Shield; *Geology*, v. 7, p. 125-128.

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Appendix A. Geological maps (1:250,000 scale) for northern Cordillera

N.T.S. map-area Name (see Fig. 2.2)	G.S.C. map	Author or Publication*
95 L Flat River	1314 A	Gabrielse <u>et al.</u> , 1973; Mem. 366
95 M Wrigley Lake	1315 A	"
96 D Carcajou Canyon	1390 A	Aitken and Cook, 1974; Pap. 74-13
105 I Nahanni	8-1967 O.F. 689	Blusson <u>et al.</u> , 1968 Gordey, 1980; Mem. in prep.
105 J Sheldon	12-1961	Roddick and Green, 1961
105 K Tay River	13-1961 O.F. 212	"
105 L Glenlyon	1221 A, 1222 A	Tempelman-Kluit Campbell, 1967; Mem. 352
105 M Mayo	890 A	Bostock, 1947
105 N Lansing	O.F. 205	Blusson, 1974
105 O Middy Lake	O.F. 765 O.F. 807	Cecile, 1981 Gordey, 1981
105 P Sekwi Mtn.	1333 A	Blusson, 1971. Pap. 74-13
106 A Mt. Eduni	O.F. 221	Aitken and Cooke, 1974; Pap. 74-13
106 B Bonnet Plume R.	O.F. 205	Blusson, 1974
106 C Nadaleen R.	O.F. 206	Blusson, 1974
106 D Nash Creek	1282 A	Green, 1972; Mem. 364
106 E Wind River	1528 A	Norris, 1982
106 F Snake River	1529 A	"
106 G Rampart River	1452 A	Aitken <u>et al.</u> , 1982; Mem. 388
106 H Sans Sault Rapids	1453 A	"
115 N, O Stewart River	711 A 18-1973	Bostock, 1942 Tempelman-Kluit, Pap. 73-41
115 P McQuesten	1143 A	Bostock, 1964
116 A Larsen Creek	1283 A	Green, 1972; Mem. 364
116 B, C Dawson	1284 A	"
116 F, G Ogilvie River	1526 A	Norris, 1982
116 H Hart River	1527 A	"

* some publications not listed in Chapter 10, but available by number from libraries containing publications of the Geological Survey of Canada. (O.F. = Open File; Pap. = Paper; Mem. = Memoir)

East-central AlaskaCharley River:

Brabb, E.E.

1967: Stratigraphy of the Cambrian and Ordovician rocks of east-central Alaska; United States Geological Survey, Professional Paper 559-A.

Brabb, E.E. and Churkin, M.J.

1979: Geological map of Charley River Quadrangle, east-central Alaska; United States Geological Survey, Miscellaneous Geological Investigations, Map I-573.

Eagle Quadrangle

Foster, H.L., Weber, F.R., Forbes, R.B. and Brabb, E.E.

1973: Regional geology of the Yukon-Tanana Upland, Alaska; in Arctic Geology, ed. M.G. Pitcher; American Association of Petroleum Geologists, Memoir 19, p. 388-395.

also: Foster et al., 1975; United States Geological Survey, Professional Paper 1271-6; Map I-922.

References for Regional Stratigraphic Correlations (Fig. 2.5)

STRATIGRAPHIC STUDIES

- 1 Brabb, 1967; Payne and Allison, 1981
- 2 Green, 1972; Norris, 1976
- 3 Cecile, 1982 (Misty Creek embayment)
- 4 Mertie, 1933; Young, 1982
- 5 Eisbacher, 1981; discussion in Young *et al.* 1979
- 6 Eisbacher, 1981
- 7 Yeo, 1981
- 8 Jefferson and Ruelle, 1986
- 9 Pell and Simony, 1987
- 10 Little, 1936; Root, 1987
- 11 Thompson and Roots, 1981
- 13 Gabrielse *et al.*, 1975
- 14 Aitken, 1981
- 15 Delaney, 1981
- 16 McMechan, 1981

AGE DETERMINATIONS

- F1 Microbiota, (Allison, 1980)
- F2 Archaeocyathids (R.I. Thompson, pers. comm., 1982)
- F3 Oldhamia (Hoffmann and Cecile, 1981)
- F4 (Nowlan *et al.*, 1985)
- F5 (Narbonne *et al.*, 1985)

Isotopic dates (indicated on Fig. 2.5):

Section B: U-Pb 751 $\pm 26/-18$ zircons in felsic flow (Chapter 6)

Section E: U-Pb 778 $\pm 3/-2$ quartz diabase intrusion (Jefferson and Parrish, in prep.)

Section F: Nd-Sm 762 ± 44 plag./pyx. separates from meta-basaltic flow (Devlin, 1986)

Rb-Sr 1320 $\pm ?$ Bishop, D.T.; Petrology and geochemistry of the Purcell sills in Boundary county, Idaho; in Belt Symposium, vol. 1; Department of Geology, University of Idaho and Bureau of Mines and Geology, Moscow, Idaho.

References above not included in Chapter 10:

- Payne M.W. and Allison, C.W.
1981: Paleozoic continental margin sedimentation in east-central Alaska: *Geology*, v. 9, p. 274-279.
- Mertie, J.B.
1933: The Tatonduk-Nation district; United States geological Survey, Bulletin 836-E, p. 347-443.
- Allison, C.W.
1980: Siliceous microfossils from the Lower Cambrian of northwest Canada: Possible source for biogenic chert; *Science*, v. 211, p. 53-55.

Appendix B. Construction of Cross-sections (Figs. B-1, in pocket)

The Mount Harper Complex is interpreted to occupy a westerly plunging syncline that has been uplifted by regional shortening. The limbs of the syncline have been dropped by the North Harper and Late South Harper faults; earlier faults trend northeasterly, easterly and southeasterly. Vertical offset of the faults is generally visible, but strike-slip displacement cannot be determined. The faults and their relative age are indicated in Figures 5.2 and 5.4, as well as Table 5.1.

The cross-sections (Fig. B-1, in pocket) show the actual surface and large underlying structures. They are preliminary attempts, and unconstrained because they have not been balanced. A principal difficulty in balancing the sections is taking into account the transcurrent motion on some southeast-trending faults; normal balanced sections only resolve movement that occurred in the plane of the section.

The rock units of the Harper Group are not tabular, and are therefore projected to depth with considerable uncertainty. From estimated thicknesses in several exposed sections, the general shapes of some volcanic units has been deduced. Furthermore, buried erosion surfaces that outline the remnants of older volcanic structures.

The Lower Harper Group is interpreted as a northward-tapering wedge. Its present exposure is an oblique cross-section from 1500 m thick (Section U) to 25 m thick further northwest (Section J). Although it probably consists of coalesced debris fans abutting the Early South Harper fault, it is depicted in the cross sections as a wedge-like slab of even dimensions.

The Early South Harper Fault is a stepped normal fault, locally reactivated, with an arbitrary 2000 m of vertical displacement.

Member A is a low-angle shield with a maximum thickness of 1500 m in Section 144, and tapers to south, west and north from that area. Member B, an erosionally truncated remnant, has its greatest exposed thickness beneath Black Towers (about 1300 m). Member C is probably more extensive than presently exposed, and shown here as clastic wedges beneath the Tango Tarn Fault (Fig. 4.8) as well as in the upper West Trib valley. Member D is assumed to be a thick, tabular flow. The andesitic flows of Member E are floods that may have ponded to the south of the present Complex; the hydroclastic breccia is considered a broad apron that tapered northward from approximately the Lower West Trib area. Figure B-2 shows possible pre-deformational paleogeography for the Lower Harper Group and earlier volcanic units.

Fig. B.2 Morphology of Late Proterozoic deposits, before erosion, burial and deformation: a) Lower Harper Group, interpreted as overlapping clastic fans; b) Volcanic edifice comprising a shield (Member A) surmounted by steeper sided cones (Member B). Sections L, O and Q of Fig. 5.2 are shown as they might have appeared before erosion and deformation.

Accuracy and precision of chemical analyses

Details of the X-ray fluorescence spectroscopy method in the Analytic Chemistry Division laboratory of the Geological Survey of Canada (chief analyst: D.M. Rousseau) are given near the beginning of Chapter 7.

Detection limits for major and minor elements, given by the laboratory, are as follows: 0.50% for Na_2O ,

0.40% for SiO_2 and Al_2O_3 ;

0.2% for TiO_2 , Cr_2O_3 , P_2O_5 , S and CO_2 ;

0.1% for Fe_2O_3 , MgO and CaO;

0.05% for K_2O and H_2O ;

0.03% for FeO and 0.01% for MnO.

The laboratory states that "analyses by optical emission are expected to be more accurate at lower concentrations, i.e. Si, Al, Fe, Mg, Ca and Na when less than 1.0%, and Ti, Mn when less than 0.5%"

Detection limits for trace elements, also given by the laboratory, were:

30 ppm for As;

20 ppm for Mo, Pb, Th and U, and

10 ppm for Nb, Rb, Sr, Y and Zr.

Typical trace element abundances in basaltic rocks from Mount Harper are low, and only Nb, Rb, Sr, Y and Zr contents are consistently higher than the detection limit and so are used here.

Results of tests for accuracy and precision are presented in Tables C-1 and C-2. The laboratory also ran 5 Geological Survey standards for internal verification: these results were not distributed.

Accuracy is estimated from analyses of three samples by different workers at different laboratories. The fine-grained volcanic rocks, powdered to -180 mesh, were procured from the Saskatchewan Geological

	SGS-1				SK	SGS-2				SK	SGS-3		
	This work	EK	Mo	RE		This work	EK	RE	KB		This work	RE	SK
Total max. error													
abs. ± rel. dev.													
SiO ₂	76.8	74.7	76.3	76.34	76.2	51.2	51.7	51.85	51.85	51.5	47.4	48.47	47.7
TiO ₂	0.13	0.14	0.14	0.12	0.13	1.15	1.14	1.12	1.12	1.12	0.82	0.78	0.79
Al ₂ O ₃	11.7	12.0	12.1	11.75	12.0	19.8	20.1	20.30	20.30	20.1	15.10	15.47	14.9
Fe Tot	0.70	2.21	2.17	2.42	2.24	2.60	8.55	8.75	8.72	8.61	3.50	17.38	16.8
Fe ₂ O ₃	1.40	1.0				5.40	5.3				12.20		
MnO	0.05	0.03	0.05	0.04	0.03	0.13	0.12	0.12	0.12	0.12	0.30	0.27	0.28
MgO	0.54	0.46	0.43	0.30	0.50	2.60	2.56	2.43	2.43	2.59	6.04	6.21	6.27
CaO	1.83	1.77	1.83	1.71	1.83	8.32	8.28	8.05	8.05	8.21	7.18	6.84	6.78
Na ₂ O	3.40	4.11	4.20	3.84	4.40	2.90	3.84	3.98	3.98	3.79	1.10	1.97	1.81
K ₂ O	1.67	1.62	1.67	1.63	1.62	0.99	1.00	1.00	1.00	0.99	0.27	0.26	0.27
P ₂ O ₅	0.01	0.03	0.03	0.00	0.02	0.39	0.39	0.29	0.29	0.38	0.13	0.05	0.14
S	0.0	0.01		0.0		0.28	0.21	0.16	0.16		0.13	0.06	
B ₂ O ₃	0.049					0.038					0.011		
L.O.I.		1.23	1.31		1.16	0.10*	2.62	2.3		2.23	4.40		3.26
H ₂ O	0.90					2.60					0.40		
CO ₂	0.50					0.70					0.40		
As	18	<5			2	4	<5			5	11		9
Mo	490	570	560	478	600	380	320	366		410	110	129	170
Nb	5	20	20	2	10	1	10	0		<10	1	0	<10
Pb	5	12			12	18	41			22			14
Nb	29	<10	* 30	30	40	9	10	12		30	3	0	20
Sr	116	120	150	119	150	584	660	616		616	90	90	70
Th	0	6			3	3	<1			0.7	3		0.5
U	1	<10			1.8	0	<10			0.7	3		0.5
Y	48				30	0	<10			20	35	32	<10
Zr	177	130	150	148	140	36	58	40		50	39	32	<10

EK - XRF analyzed by X-Ray Assay Laboratories, Toronto, Canada, 1983

L.O.I. = Loss on Ignition

RE - XRF analyzed at University of Ottawa, Dept. of Geology, 1984

SK - XRF analyzed by X-Ray Assay Laboratories, Toronto, Canada, 1982

* = all compounds lost on Ignition other than CO₂ and H₂O.

Table C-1. Analyses of three "standard samples" to estimate accuracy.

	F104	DUP	8131-1	DUP	A128-7	DUP	F161	DUP
SiO ₂	38.9	39.3	50.5	50.1	34.8	35.1	52.5	51.6
TiO ₂	1.05	1.05	0.93	0.93	0.45	0.46	0.59	0.58
Al ₂ O ₃	10.4	10.1	15.2	15.4	13.9	13.9	13.6	13.6
Fe ₂ O ₃	1.7	1.9	2.10	1.70	0.60	0.30	2.7	2.4
FeO	9.2	9.2	7.2	7.3	5.8	6.0	6.5	6.6
MnO	0.18	0.19	0.18	0.18	0.23	0.24	0.19	0.20
MgO	7.45	7.60	5.08	5.20	5.10	5.12	6.02	5.91
CaO	12.4	12.5	11.2	11.3	17.6	17.5	9.42	9.67
Na ₂ O	0.3	0.3	1.1	1.3	1.9	2.1	1.0	0.8
K ₂ O	2.25	2.30	0.92	0.90	0.67	0.72	0.88	0.86
P ₂ O ₅	0.11	0.10	0.10	0.11	0.03	0.07	0.04	0.05
Cr ₂ O ₃	0.01	0.01	0.02	0.01	0.07	0.08	0.03	0.02
S	0.08	0.04	0.03	0.01	0.69	0.74	0.02	0.04
Ba	.045	.043	.028	.028	.017	.021	.022	.023
H ₂ O	4.1	4.0	3.0	3.0	4.4	3.9	4.6	5.0
CO ₂	11.8	11.8	2.1	2.0	13.0	12.4	0.8	1.1
Total	99.93	100.40	99.66	98.45	99.24	98.63	98.89	98.45
As	9	5	8	0	6	1	3	6
Br	4	2	1	4	3	2	2	1
Mo	0	2	0	3	3	2	4	2
Nb	7	9	4	8	2	1	5	5
Pb	16	13	10	13	16	13	3	4
Rb	43	46	27	24	16	15	19	18
Sr	224	237	270	266	75	73	32	33
Th	6	6	4	8	0	2	5	0
U	0	1	0	0	0	4	2	0
Y	45	43	34	33	23	23	31	31
Zr	118	122	99	97	43	40	83	82

Table C-2. Analyses of duplicate samples to estimate precision.

Survey (Regina, Saskatchewan; C. Dunn, D.J. Thomas; pers. comm. to K. Bell and J. Richardson, Carleton University, 1980, 1983 and 1986). A 'best value' has not been calculated for these samples (additional analyses are in progress), so that variations in results could reflect different procedures in sample preparation or measurement.

Table C-1 shows reasonable agreement in SiO_2 , TiO_2 , MnO , CaO and K_2O , with lower Al_2O_3 and higher MgO , except in basalt. The greatest discrepancy is the considerably lower Na_2O (20% or more) from the G.S.C. laboratory. In conjunction with the high detection limit, the dozen samples with very low or no recorded Na_2O could contain up to 0.7%. Na_2O is commonly reanalyzed by the atomic absorption method.

Precision was estimated by analysis of four sets of duplicate samples, prepared as splits after crushing and submitted blind to the laboratory. All major and minor elements show close agreement between analyses of the same rock, except Fe_2O_3 , which in two samples differs by more than 20%. Among trace elements, all variations are within the range given by the 'estimate of validity', although the concentrations are below the detection limits for As, Br, Mo, Th and U.

In conclusion, the analyses are considered reliable, except for Na_2O , Fe_2O_3 , and trace elements where lower than detection limits.

Table C-3. Whole-rock analyses from Member A

	131-1	118-6	144-11	150-2	132	130-1	131-2	128-3	133	127-2
SiO ₂	50.50	50.20	46.10	49.80	47.70	52.30	48.40	47.30	53.50	47.70
Al ₂ O ₃	15.20	14.00	15.30	15.90	13.00	13.60	13.60	14.00	14.10	13.60
Fe ₂ O ₃	2.10	2.10	1.50	1.60	1.50	1.40	2.00	1.90	1.60	2.00
FeO	7.20	7.20	7.40	6.60	6.00	5.60	7.40	7.70	4.50	7.60
MgO	5.08	7.04	6.79	6.35	6.94	6.51	8.17	10.80	6.39	11.30
CaO	11.20	10.40	12.00	11.20	11.50	8.53	11.30	11.40	8.08	11.10
Na ₂ O	1.10	2.10	0.90	1.70	1.60	2.70	1.00	0.20	2.60	0.40
K ₂ O	0.92	0.74	0.12	0.62	0.30	0.22	0.38	0.12	0.28	0.16
TiO ₂	0.93	0.67	0.69	0.83	0.61	0.76	0.79	0.69	0.61	0.66
P ₂ O ₅	0.10	0.06	0.09	0.09	0.07	0.08	0.08	0.07	0.09	0.07
MnO	0.18	0.19	0.17	0.16	0.15	0.17	0.19	0.18	0.15	0.20
S	0.03	0.08	0.0	0.01	0.0	0.0	0.0	0.31	0.0	0.04
Ba	0.028	0.070	0.008	0.029	0.008	0.011	0.017	0.010	0.010	0.013
Cr ₂ O ₃	0.02	0.04	0.08	0.06	0.14	0.13	0.16	0.13	0.10	0.13
CO ₂	2.10	2.40	3.90	2.00	5.40	2.50	2.80	0.20	4.30	0.30
H ₂ O	<u>3.00</u>	<u>3.20</u>	<u>3.40</u>	<u>3.50</u>	<u>3.60</u>	<u>3.70</u>	<u>3.70</u>	<u>3.90</u>	<u>3.90</u>	<u>4.00</u>
Total	99.66	100.47	98.44	100.42	98.51	98.70	99.97	98.90	100.30	99.26
As	4	4	12	7	21	0	12	12	1	4
Br	3	4	1	1	4	5	0	4	1	1
Mo	2	0	1	2	3	0	0	7	0	4
Nb	6	2	6	6	4	5	6	6	3	5
Pb	12	9	10	6	14	13	13	11	11	11
Rb	26	12	4	15	12	6	9	2	8	3
Sr	268	142	102	218	66	144	121	91	72	125
Th	6	6	7	4	3	5	1	2	2	5
U	0	0	0	0	0	0	0	1	3	0
Y	34	32	23	23	20	29	24	25	21	22
Zr	98	71	70	83	52	79	77	61	55	59
Q	12.48	6.44	10.79	7.44	13.84	12.31	9.39	4.99	20.43	4.01
C	0.0	0.0	1.18	0.0	1.91	0.0	0.0	0.0	5.20	0.0
OR	5.63	4.30	0.75	3.78	1.87	1.37	2.33	0.75	1.72	0.99
AB	9.63	18.27	8.01	14.84	14.26	24.05	8.79	1.78	22.84	3.55
AN	34.99	27.34	36.08	35.00	23.66	25.62	32.72	38.90	12.79	39.57
DI	3.89	5.19	0.0	4.61	0.0	0.99	3.64	11.17	0.0	11.16
HE	2.74	2.70	0.0	2.45	0.0	0.41	1.69	3.99	0.0	3.94
EN	11.28	15.74	17.79	14.18	18.21	16.60	19.44	23.13	16.52	24.36
FS	9.10	9.39	12.05	8.65	9.41	7.85	10.36	9.47	6.36	9.86
FO	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0
FA	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0
MT	3.15	3.13	2.29	2.39	2.29	2.90	3.01	2.90	2.40	3.04
IL	1.83	1.30	1.38	1.63	1.22	1.52	1.56	1.38	1.20	1.31
CR	0.03	0.06	0.12	0.09	0.22	0.20	0.24	0.20	0.15	0.20
AP	0.24	0.14	0.22	0.21	0.17	0.19	0.19	0.17	0.22	0.17
PY	0.06	0.17	0.0	0.21	0.0	0.0	0.0	0.67	0.0	0.08
CC	4.94	5.61	9.32	4.69	12.94	5.98	6.61	0.48	10.15	0.71

Major elements in wt.%; minor elements in ppm; normative calculations in wt.%

Table C-3 (con't.) Whole-rock analyses from Member A (p.2)

	128-11	128-9	128-8	128-7	128-6	97b	15	8	124-3	118-8	118-7
SiO ₂	47.70	48.50	45.90	34.80	52.00	44.80	38.90	46.80	43.40	45.50	49.80
Al ₂ O ₃	14.00	13.90	14.00	13.90	14.40	14.30	11.80	14.10	16.00	19.00	18.70
Fe ₂ O ₃	2.30	0.80	1.00	0.60	1.50	1.30	1.20	1.30	2.00	1.30	1.40
FeO	7.90	4.90	4.80	5.80	5.70	5.40	4.90	6.40	6.90	5.30	5.00
MgO	6.25	3.84	4.57	5.10	6.86	5.93	5.68	5.55	5.69	5.24	5.62
CaO	8.64	11.40	12.00	17.60	8.09	10.80	17.50	9.15	9.30	9.28	9.50
Na ₂ O	2.20	3.50	3.30	1.90	2.10	2.80	1.60	3.30	3.30	1.30	1.10
K ₂ O	0.28	0.33	0.91	0.67	0.24	0.37	0.26	0.46	0.71	2.10	0.49
TiO ₂	1.30	0.74	0.79	0.45	0.68	0.65	0.59	0.72	0.57	0.49	0.51
P ₂ O ₅	0.13	0.08	0.10	0.03	0.09	0.08	0.04	0.08	0.06	0.04	0.07
MnO	0.16	0.18	0.14	0.23	0.17	0.14	0.17	0.14	0.16	0.10	0.10
S	0.16	0.04	0.0	0.69	0.12	0.01	0.0	0.05	0.0	0.01	0.0
Ba	0.010	0.033	0.0019	0.017	0.010	0.006	0.006	0.023	0.013	0.062	0.022
Cr ₂ O ₃	0.02	0.08	0.09	0.07	0.08	0.08	0.07	0.08	0.13	0.07	0.06
CO ₂	4.20	8.80	9.00	13.00	3.90	8.40	12.70	6.50	7.10	5.40	2.90
H ₂ O	<u>4.60</u>	<u>3.20</u>	<u>3.50</u>	<u>4.40</u>	<u>4.50</u>	<u>4.20</u>	<u>3.90</u>	<u>4.30</u>	<u>4.10</u>	<u>4.80</u>	<u>4.10</u>
Total	99.84	99.79	100.70	99.24	100.43	98.71	99.31	99.03	99.45	99.99	100.05
As	0	9	0	4	4	0	0	0	4	2	0
Br	1	0	5	3	2	1	1	0	2	0	4
Mo	0	0	0	3	0	1	2	1	0	1	2
Nb	6	4	6	2	3	5	4	3	3	3	4
Pb	9	10	10	15	12	8	10	6	12	11	9
Rb	6	8	20	16	5	11	5	15	15	35	6
Sr	102	182	111	74	102	47	59	68	103	89	100
Th	2	4	4	1	5	0	0	1	0	0	1
U	2	2	1	2	0	0	3	0	0	0	0
Y	30	23	29	23	22	19	18	18	14	15	15
Zr	116	68	82	42	65	66	58	71	51	47	48
Q	13.75	18.28	12.78	8.06	19.31	14.79	14.78	12.32	8.28	13.87	14.07
C	4.63	7.39	7.67	8.69	5.45	9.83	6.90	7.18	9.89	10.77	5.02
OR	1.74	2.02	5.54	4.18	1.48	2.31	1.61	2.87	4.40	13.42	3.02
AB	19.54	30.66	28.72	16.96	18.52	25.07	14.19	29.47	29.28	11.55	15.87
AM	16.24	0.42	2.04	5.21	15.52	0.0	6.57	3.99	1.06	12.23	29.53
DI	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0
HE	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0
EM	16.34	9.9	11.71	13.39	17.81	15.74	14.82	14.59	14.86	13.70	14.58
FS	10.93	7.55	7.06	8.12	8.45	8.39	7.64	10.11	10.76	8.35	7.63
FO	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0
FA	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0
MT	3.50	1.20	1.49	.92	2.26	1.99	1.82	1.99	3.04	1.98	2.11
IL	2.59	1.45	1.54	0.90	1.35	1.31	1.17	1.44	1.13	0.98	1.00
OR	0.03	0.12	0.14	0.11	0.12	0.12	0.11	0.12	0.20	0.11	0.09
AP	0.32	0.19	0.24	0.07	0.22	0.19	0.09	0.19	0.15	0.09	0.17
PY	0.35	0.85	0.0	1.49	0.26	0.02	0.0	0.11	0.0	0.02	0.0
CC	10.35	20.72	21.06	31.19	9.25	20.21	30.27	15.60	16.93	12.90	6.87

Major elements in wt.%; minor elements in ppm; normative calculations in wt.%.

Table C-3 (con't.) Whole-rock analyses from Member A (p.3)

	118-5	118-4	118-3	118-3	117-2	117-1	158-2	144-10	144-9	144-8	144-7	144-2
SiO ₂	35.20	45.80	42.80	58.00	48.90	47.20	39.60	45.00	45.90	46.00	45.10	49.20
Al ₂ O ₃	15.40	13.50	14.70	11.50	16.20	16.50	15.80	14.90	14.60	14.00	14.50	14.60
Fe ₂ O ₃	0.90	1.30	1.20	0.50	1.20	2.40	1.10	1.70	1.40	0.90	1.20	1.00
FeO	6.40	7.50	7.60	5.20	7.90	8.00	7.60	6.60	7.70	4.80	6.40	5.90
MgO	7.56	6.28	5.63	4.55	5.83	6.48	7.97	5.62	6.70	4.33	7.23	5.83
CaO	15.50	9.59	10.40	6.02	5.86	8.14	10.80	11.20	8.18	13.90	10.00	9.65
Na ₂ O	0.60	0.70	0.70	0.50	3.10	2.40	0.0	1.40	2.00	1.30	1.10	1.80
K ₂ O	1.24	1.54	2.57	3.17	0.67	0.32	1.11	0.27	0.47	1.39	1.05	1.29
TiO ₂	0.41	0.71	0.72	0.59	0.77	0.79	0.81	0.76	0.83	0.55	0.77	0.79
P ₂ O ₅	0.04	0.09	0.09	0.08	0.08	0.08	0.07	0.10	0.06	0.10	0.07	0.09
MnO	0.17	0.16	0.17	0.15	0.14	0.18	0.11	0.15	0.13	0.18	0.14	0.18
S	0.0	0.0	0.0	0.33	0.08	0.05	0.0	0.0	0.02	0.04	0.06	0.11
Ba	0.044	0.021	0.026	0.075	0.0	0.019	0.018	0.010	0.026	0.021	0.020	0.019
Cr ₂ O ₃	0.05	0.11	0.08	0.06	0.05	0.06	0.06	0.06	0.06	0.11	0.06	0.07
CO ₂	11.60	7.70	9.10	5.00	3.90	2.80	8.50	6.90	5.50	9.00	7.60	6.40
H ₂ O	3.10	4.90	4.90	3.80	4.60	5.40	6.80	4.00	5.10	3.30	4.60	4.30
Total	100.17	99.88	100.66	99.45	99.28	100.79	100.36	98.62	98.69	99.47	99.64	100.61
As	0	6	1	6	1	6	0	0	0	4	0	0
Br	5	1	0	4	4	1	1	2	0	5	2	2
Mo	2	0	3	0	1	7	1	4	1	1	0	0
Nb	3	5	6	3	5	8	5	6	5	3	5	4
Pb	13	10	13	9	11	15	9	9	9	10	10	10
Rb	14	23	31	30	16	7	37	7	14	44	30	36
Sr	111	86	122	48	106	123	21	87	56	63	52	59
Th	4	1	2	3	5	5	3	3	2	0	0	2
U	0	0	0	0	1	1	0	0	0	0	0	2
Y	12	19	23	22	19	29	24	26	24	16	23	35
Zr	39	76	80	71	70	76	89	88	88	46	82	83
Q	9.92	22.61	17.14	34.64	13.03	7.72	19.49	19.07	15.92	19.39	20.02	18.86
C	12.47	11.24	11.23	7.57	9.45	4.26	15.60	8.53	9.52	6.35	11.75	7.41
OR	7.71	9.59	15.86	19.59	4.19	1.98	7.02	1.69	2.97	8.55	6.52	7.92
AB	5.34	6.23	6.18	4.42	27.70	21.29	0.0	12.52	18.08	11.43	9.77	15.81
AN	3.47	0.0	0.0	0.0	4.11	23.30	0.0	12.21	5.51	12.07	1.03	7.09
DI	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0
HE	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0
EN	19.80	16.05	13.14	11.28	15.33	16.92	21.00	14.79	17.83	11.21	18.90	15.07
FS	11.15	12.04	11.27	7.67	12.99	12.43	12.55	10.24	12.59	6.85	10.18	9.09
FO	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0
FA	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0
MT	1.37	1.98	1.81	0.76	1.84	3.48	1.70	2.60	2.17	1.36	1.83	1.56
IL	0.82	1.42	1.43	1.17	1.54	1.57	1.69	1.52	1.68	1.09	1.53	1.56
OR	0.08	0.17	0.12	0.09	0.08	0.09	0.09	0.09	0.09	0.17	0.09	0.10
AP	0.09	0.22	0.22	0.19	0.19	0.17	0.25	0.15	0.25	0.17	0.22	0.23
PY	0.0	0.0	0.0	0.71	0.17	0.11	0.0	0.0	0.04	0.09	0.0	0.23
OC	27.75	18.43	21.59	11.89	9.37	6.68	20.66	16.58	13.36	21.28	18.15	15.11

Major elements in wt.%, minor elements in ppm; normative calculations in wt.%,

Table C-4. Whole-rock analyses from Member B (p.1)

	129	158-1	128-15	97a	162-1	118-13	144-5	118-12	128-5	128-17
SiO ₂	49.70	45.90	48.40	48.70	48.10	49.20	43.40	48.60	47.90	64.80
Al ₂ O ₃	15.80	15.60	14.70	14.30	15.60	15.20	15.80	16.10	13.80	14.50
Fe ₂ O ₃	4.10	1.80	1.90	3.10	1.70	1.80	2.00	1.70	1.50	4.10
FeO	4.00	8.00	7.60	7.70	7.20	6.60	7.70	5.70	7.00	2.10
MgO	7.62	10.30	7.80	8.87	8.91	6.58	8.28	6.69	8.08	4.03
CaO	10.80	10.90	13.30	11.20	11.00	12.10	14.20	8.99	11.00	1.59
Na ₂ O	2.20	1.10	0.90	1.30	0.40	2.60	1.10	3.10	2.40	5.80
K ₂ O	0.34	0.29	0.39	0.33	0.29	0.79	0.36	0.48	0.48	0.17
TiO ₂	0.72	0.54	0.65	0.71	0.59	0.63	0.58	0.64	0.62	0.52
P ₂ O ₅	0.05	0.04	0.05	0.05	0.07	0.04	0.03	0.06	0.04	0.02
MnO	0.17	0.18	0.20	0.21	0.18	0.18	0.18	0.13	0.18	0.04
S	0.0	0.03	0.02	0.02	0.0	0.0	0.0	0.0	0.23	0.01
Ba	0.016	0.009	0.014	0.015	0.014	0.015	0.015	0.012	0.025	0.010
Cr ₂ O ₃	0.05	0.09	0.07	0.06	0.08	0.07	0.08	0.08	0.06	0.09
CO ₂	0.10	0.50	1.30	0.80	1.70	1.80	3.10	2.80	2.20	0.50
H ₂ O	3.60	3.60	1.70	2.70	3.00	3.00	3.10	3.40	3.50	3.60
Total	99.25	98.87	98.98	100.06	98.82	100.59	99.91	98.47	9.99	100.55
As	2	1	10	0	8	4	0	0	6	0
Br	3	0	0	2	4	2	0	1	3	4
Mo	3	7	6	5	1	2	4	0	2	0
Nb	3	2	5	2	4	2	3	0	0	3
Pb	9	7	10	11	12	13	11	7	10	6
Rb	11	7	11	8	8	19	8	7	9	1
Sr	236	63	55	87	88	208	56	87	36	73
Th	1	1	0	0	5	5	0	2	1	0
U	0	0	0	0	0	0	0	0	0	0
Y	22	24	26	29	22	24	21	15	28	7
Zr	43	28	34	32	55	31	30	46	32	38
Q	2.54	0.0	4.96	3.08	8.70	0.38	0.0	3.77	1.15	20.68
C	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.81	0.0	3.15
OR	2.10	1.80	2.37	2.00	1.79	4.79	2.20	2.99	2.97	1.02
AB	19.50	9.77	7.83	11.31	3.53	22.54	9.61	27.59	21.27	50.06
AM	33.77	38.59	35.89	33.12	41.66	28.15	38.33	27.88	26.67	4.69
DI	12.72	8.27	12.39	9.96	2.03	11.44	7.59	0.0	8.74	0.0
HE	9.48	3.51	6.44	5.06	0.89	6.01	3.78	0.0	3.94	0.0
EN	13.94	19.15	14.22	18.09	22.21	11.49	15.27	17.52	17.02	10.24
FS	5.65	9.33	8.48	10.54	11.12	6.92	8.73	8.60	8.79	4.81
FO	0.0	1.48	0.0	0.0	0.0	0.0	1.76	0.0	0.0	0.0
FA	0.0	0.0	0.0	0.0	0.0	0.0	1.10	0.0	0.0	0.0
MT	3.37	2.74	2.83	3.29	2.57	2.67	2.99	2.59	2.28	2.99
IL	1.43	1.08	1.27	1.38	1.17	1.23	1.14	1.28	1.23	1.01
CR	0.07	0.14	0.10	0.09	0.12	0.11	1.22	0.12	0.09	0.10
AP	0.12	0.09	0.12	0.12	0.17	0.09	0.07	0.15	0.09	0.05
PY	0.0	0.06	0.04	0.06	0.0	0.0	0.0	0.0	0.49	0.02
CC	0.30	1.94	3.04	1.87	4.03	4.19	7.28	6.69	5.24	1.16

Major elements in wt. %; minor elements in ppm; norm. calculations in wt. %.

Table C-4 (con't.) Whole-rock analyses from Member B (p.2) 212

	126-2	124-1	128-14	126-1	144-8	128-16	135	144-15	118-10	128-18	144-13	144-12
SiO ₂	53.40	70.80	51.60	71.20	46.00	49.20	51.80	43.90	41.30	53.20	47.70	41.00
Al ₂ O ₃	17.00	14.00	15.00	11.80	14.00	17.30	13.60	16.80	13.00	15.30	10.90	10.40
Fe ₂ O ₃	7.40	0.50	3.80	0.80	0.90	6.00	6.50	4.90	0.60	6.30	6.00	0.80
FeO	3.00	2.30	3.00	3.50	4.40	3.60	2.20	4.40	4.40	3.40	1.40	6.90
MgO	4.12	2.30	4.40	4.59	4.33	6.54	6.77	7.27	4.23	6.16	4.75	8.10
CaO	2.77	1.16	6.10	0.59	13.90	4.25	5.91	11.40	16.40	4.64	13.30	14.00
Nb ₂ O ₅	7.00	2.90	4.40	3.10	1.30	5.30	3.80	0.90	1.80	4.40	0.0	0.0
K ₂ O	0.17	1.95	0.63	0.11	1.39	0.28	0.28	1.12	1.07	0.18	0.15	0.32
TiO ₂	0.60	0.38	0.72	0.38	0.55	0.56	0.61	0.50	0.44	0.46	0.32	0.32
P ₂ O ₅	0.06	0.03	0.08	0.0	0.07	0.05	0.06	0.02	0.04	0.04	0.03	0.03
MnO	0.10	0.02	0.11	0.03	0.18	0.07	0.07	0.13	0.15	0.10	0.16	0.19
S	0.0	0.0	0.0	0.0	0.04	0.0	0.02	0.0	0.06	0.0	0.0	0.28
Ba	0.018	0.075	0.030	0.011	0.021	0.018	0.016	0.030	0.011	0.013	0.008	0.005
Cr ₂ O ₃	0.12	0.07	0.08	0.07	0.11	0.08	0.09	0.07	0.09	0.06	0.04	0.05
CO ₂	1.90	0.90	5.50	0.50	9.0	2.10	3.90	4.40	11.40	2.10	10.40	11.40
H ₂ O	2.60	2.80	3.10	3.30	3.30	3.40	3.40	3.50	3.60	3.70	3.80	5.00
Total	99.24	100.11	100.56	99.97	99.47	98.73	98.99	99.31	98.63	100.04	98.95	98.51
As	3	1	18	7	4	5	11	12	0	3	10	0
Br	3	2	6	5	5	1	2	3	4	2	4	0
Mo	2	0	0	0	1	1	2	1	2	1	4	3
Nb	3	3	4	4	3	3	6	3	3	3	3	3
Pb	9	10	7	11	10	11	13	13	9	9	10	4
Rb	2	43	19	3	44	7	7	44	32	5	5	1
Sr	75	63	246	27	63	169	168	86	80	94	31	43
Tb	4	0	1	1	0	2	2	4	4	0	4	0
U	3	0	1	0	0	0	0	3	0	0	2	1
Y	18	3	17	3	16	20	17	20	14	18	14	15
Zr	63	46	70	3	46	41	66	35	32	39	27	27
Q	0.0	42.77	12.81	54.12	19.39	0.0	13.47	6.69	13.99	8.96	38.60	25.94
C	5.0	7.31	5.34	6.81	6.35	5.83	5.76	3.79	5.86	4.58	11.32	11.07
OR	1.05	11.85	3.83	0.67	8.55	1.74	1.74	6.94	6.66	1.11	0.94	0.25
AB	61.66	25.21	38.26	27.13	11.44	47.25	33.80	7.97	16.03	38.83	0.0	0.0
AN	1.39	0.0	5.28	0.0	12.07	7.88	4.49	29.96	9.50	9.84	0.04	0.0
DI	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0
ME	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0
EN	8.43	5.85	4.24	11.76	11.21	16.10	17.73	18.96	11.21	16.03	12.49	20.80
FS	10.04	3.23	5.35	5.28	6.85	10.36	8.95	11.02	7.29	11.57	8.06	11.56
FO	1.22	0.0	0.0	0.0	0.0	0.75	0.0	0.0	0.0	0.0	0.0	0.0
FA	1.51	0.0	0.0	0.0	0.0	0.53	0.0	0.0	0.0	0.0	0.0	0.0
MT	3.17	0.74	3.31	1.20	1.36	3.15	3.22	3.04	0.91	2.96	2.78	1.24
IL	1.19	0.74	1.41	0.75	1.09	1.12	1.22	0.99	0.88	0.91	0.64	0.65
OR	0.18	0.10	0.12	0.11	0.17	0.12	0.14	0.11	0.14	0.09	0.06	0.08
AP	0.14	0.07	0.19	0.0	0.17	0.12	0.15	0.05	0.09	0.09	0.07	0.07
PY	0.0	0.0	0.0	0.0	0.17	0.0	0.0	0.0	0.13	0.0	0.0	0.61
CC	4.49	2.10	12.86	1.18	21.28	5.03	9.33	10.48	27.28	4.98	24.98	27.71

Major elements in wt.%; minor elements in ppm; normative calculations in wt.%.

Table C-4 (con't.) Whole-rock analyses from Member B (p.3)

	128-13	128-12	124-2	126-6	126-4	121-3	118-14	118-9	117-5	117-4	162-2	144-14
SiO ₂	39.30	38.30	51.10	42.80	41.60	45.10	65.60	47.40	52.70	44.70	49.10	39.50
Al ₂ O ₃	14.90	14.10	18.30	14.00	11.10	16.00	9.00	12.90	15.20	13.60	12.10	17.00
Fe ₂ O ₃	1.20	4.20	5.40	2.90	6.30	1.50	1.70	1.00	1.40	0.0	1.00	1.10
FeO	5.20	4.90	4.90	4.70	4.60	8.40	6.80	5.40	6.40	9.10	6.10	6.90
MgO	6.21	6.11	6.16	7.03	6.20	6.67	4.82	8.02	4.84	6.56	8.10	11.70
CaO	15.20	14.50	1.41	20.20	13.20	8.79	3.49	9.59	4.28	12.40	9.05	8.53
Na ₂ O	1.40	1.70	3.70	0.0	1.00	2.00	1.10	0.40	0.20	0.80	0.0	0.0
K ₂ O	1.73	0.87	1.68	0.04	0.47	0.53	0.13	0.58	5.03	1.36	0.18	0.16
TiO ₂	0.41	0.38	0.52	0.52	0.34	0.74	0.52	0.45	0.79	0.61	0.51	0.61
P ₂ O ₅	0.04	0.05	0.04	0.03	0.05	0.09	0.05	0.04	0.10	0.06	0.05	0.07
MnO	0.22	0.17	0.06	0.14	0.15	0.18	0.11	0.11	0.07	0.18	0.13	0.13
S	0.02	0.0	0.04	0.0	0.0	0.02	0.11	0.0	0.13	0.32	0.01	0.0
Ba	0.044	0.027	0.342	0.006	0.009	0.018	0.006	0.009	0.089	0.022	0.006	0.009
Cr ₂ O ₃	0.08	0.07	0.09	0.06	0.07	0.06	0.01	0.10	0.10	0.05	0.15	0.08
CO ₂	10.20	10.60	1.00	2.80	10.20	3.70	1.60	7.30	3.40	6.80	7.10	6.90
H ₂ O	4.50	4.50	4.90	4.40	4.00	4.80	3.90	5.80	4.40	4.00	5.20	7.80
Total	100.61	100.45	99.30	99.62	99.28	99.58	98.94	99.09	99.04	100.54	98.78	100.48
As	2	0	2	4	2	2	0	0	2	4	3	7
Br	0	1	1	0	1	2	1	4	0	0	1	2
Mo	3	1	1	4	6	3	0	4	1	1	0	2
Nb	1	0	3	3	4	4	19	2	6	0	3	3
Pb	7	9	10	11	5	10	9	8	8	13	7	7
Rb	49	26	38	0	15	11	0	11	41	17	4	4
Sr	73	51	138	18	63	111	53	39	62	95	18	20
Th	0	0	0	3	0	1	5	2	1	0	0	4
U	0	0	0	0	0	0	0	0	0	1	0	0
Y	14	10	7	22	13	25	18	15	28	24	14	13
Zr	29	33	50	31	35	74	82	33	88	36	47	56
Q	7.22	9.45	8.87	2.77	19.13	7.47	46.20	28.45	22.31	10.61	34.22	18.17
C	7.09	9.07	10.89	0.0	9.17	5.27	4.77	11.94	9.95	4.31	12.72	18.14
OR	10.65	5.38	10.57	0.25	2.93	3.34	0.81	3.68	31.41	8.33	1.14	1.02
AB	12.32	15.03	33.29	0.0	8.92	18.04	9.79	3.63	1.79	7.01	0.0	0.0
AN	11.09	4.80	0.44	40.03	0.71	20.93	7.23	1.25	0.0	18.79	0.0	0.0
DI	0.0	0.0	0.0	27.08	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0
HE	0.0	0.0	0.0	10.45	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0
EN	16.09	15.90	16.31	5.85	16.29	17.71	12.63	21.41	12.51	15.92	21.46	30.89
FS	8.51	11.35	12.70	2.59	14.63	14.08	10.73	9.07	9.41	15.89	10.24	11.58
FO	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0
FA	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0
MT	1.81	2.85	3.11	3.08	2.81	2.32	2.59	1.55	2.14	0.0	1.55	1.72
IL	0.81	0.35	1.05	1.04	0.68	1.50	1.04	0.92	1.58	1.20	1.03	1.25
CR	0.12	0.11	0.14	0.09	0.11	0.09	0.15	0.16	0.15	0.07	0.24	0.13
AP	0.09	0.12	0.10	0.07	0.12	0.22	0.12	0.10	0.24	0.14	0.12	0.17
PY	0.04	0.0	0.04	0.0	0.0	0.04	0.24	0.0	0.28	0.68	0.02	0.02
CC	24.14	25.19	2.42	6.69	24.47	8.97	3.83	17.79	8.17	16.02	17.25	16.92

Major elements in wt.%; minor elements in ppm; normative calculations in wt.%. 5

Table C-5. Whole-rock analyses from Member D

	138-1	138-2	166	48-1	103	138-4	138-3	154	44	136-1
SiO ₂	75.40	70.10	66.60	68.60	76.70	69.10	68.50	67.50	70.50	67.10
Al ₂ O ₃	11.60	12.70	13.10	14.30	8.40	13.50	12.90	14.10	11.80	12.90
Fe ₂ O ₃	1.50	0.80	0.50	3.40	0.80	1.20	0.80	0.80	1.40	0.90
FeO	1.80	2.10	3.90	1.60	4.60	2.70	4.20	3.80	3.30	2.70
MgO	0.59	0.67	0.60	0.89	1.55	0.84	0.94	0.73	0.88	2.76
CaO	1.11	2.08	3.79	0.90	0.87	1.33	1.60	2.59	2.07	3.83
Na ₂ O	1.20	0.00	0.30	2.00	0.00	0.0	0.00	2.70	1.20	0.40
K ₂ O	4.05	7.17	5.59	4.61	4.32	7.40	7.05	2.33	2.59	2.81
TiO ₂	0.40	0.52	0.50	0.71	0.26	0.52	0.52	0.55	0.48	0.62
P ₂ O ₅	0.20	0.19	0.19	0.23	0.11	0.17	0.17	0.19	0.17	0.22
MnO	0.04	0.08	0.11	0.05	0.03	0.03	0.03	0.09	0.06	0.02
S	0.00	0.00	0.00	0.00	0.02	0.00	0.00	0.00	0.00	0.0
Ba	0.101	0.204	0.065	0.081	0.030	0.225	0.205	0.075	0.120	0.168
Cr ₂ O ₃	0.00	0.01	0.01	0.00	0.01	0.01	0.00	0.01	0.01	0.0
CO ₂	0.90	1.80	2.80	0.40	0.70	1.00	1.30	1.80	1.30	2.60
H ₂ O	1.80	1.70	1.90	2.10	2.10	2.10	2.10	2.10	2.60	3.30
Total	100.59	99.92	99.89	99.79	100.47	99.9	100.09	99.29	98.36	100.16
As	5	14	3	0	0	6	8	0	1	2
Br	0	8	2	6	0	0	3	0	0	3
Mo	0	0	0	0	2	0	0	0	0	0
Nb	12	20	20	20	13	22	20	22	17	17
Pb	9	26	23	16	9	12	15	12	23	18
Rb	140	179	199	166	111	161	138	119	120	91
Sr	79	86	23	52	9	48	57	121	83	62
Th	14	27	19	20	15	25	25	26	21	17
U	0	8	13	4	1	3	4	4	0	3
Y	31	77	77	61	21	83	75	88	60	63
Zr	222	379	384	305	194	421	411	453	386	266
Q	41.70	52.30	40.60	37.80	55.75	39.10	38.40	40.28	52.17	49.49
C	5.02	5.30	6.64	6.00	3.78	5.60	5.39	7.25	6.97	9.07
OR	43.20	24.20	33.70	27.90	25.97	44.75	42.44	14.18	15.99	17.16
AB	0.0	10.20	2.58	17.30	0.0	0.0	0.0	23.50	10.60	3.49
AN	0.0	0.0	0.0	0.44	0.0	0.0	0.0	0.23	0.98	1.16
DI	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0
HE	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0
EN	1.30	1.18	1.51	2.27	3.80	1.99	2.22	1.87	2.29	7.09
FS	1.93	1.18	6.19	2.02	7.25	3.01	5.93	5.72	4.40	3.33
FO	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0
FA	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0
MT	1.81	2.20	0.74	3.28	1.18	1.78	1.18	1.19	2.12	1.34
IL	1.00	0.77	0.969	1.38	0.50	1.01	1.01	1.07	0.95	1.22
CR	0.01	0.0	0.15	0.0	0.01	0.01	0.0	0.01	0.01	0.0
AP	0.45	0.47	0.45	0.54	0.26	0.40	0.40	0.45	0.41	0.52
PY	0.0	0.0	0.0	0.0	0.42	0.0	0.0	0.0	0.0	0.0
CC	4.17	2.07	6.49	0.93	1.16	2.32	3.02	4.21	3.08	6.10

Major elements in wt.%; minor elements in ppm; normative calculations in wt.%.

Table C-6. Whole-rock analyses from Member E

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	71	99	144-6	79-1	82-4	102	104	72-2	73	118-1
SiO ₂	69.00	55.70	54.60	56.10	64.90	48.10	38.90	60.70	48.80	53.00
Al ₂ O ₃	9.60	12.90	12.20	12.80	10.30	12.90	10.40	12.10	12.40	12.80
Fe ₂ O ₃	2.00	2.00	2.30	2.90	4.00	1.20	1.70	2.80	2.30	1.60
FeO	4.50	8.50	7.80	7.80	6.60	11.40	9.20	7.20	10.40	9.10
MgO	1.35	3.34	2.08	3.59	3.03	2.88	7.45	4.43	3.75	4.18
CaO	3.51	7.03	6.62	6.95	1.98	7.13	12.4	2.94	7.48	6.24
Na ₂ O	0.50	0.90	1.50	1.10	1.50	1.10	0.30	1.50	1.50	0.50
K ₂ O	3.74	1.93	3.02	1.93	1.26	2.01	2.25	2.10	1.24	1.81
TiO ₂	0.39	1.30	1.47	1.31	1.51	1.48	1.05	1.17	1.30	1.41
P ₂ O ₅	0.13	0.28	0.29	0.29	0.38	0.15	0.11	0.22	0.13	0.28
MnO	0.10	0.18	0.18	0.18	0.07	0.20	0.18	0.08	0.19	0.15
S	0.03	0.02	0.0	0.05	0.04	0.17	0.08	0.17	0.09	0.03
Ba	0.089	0.197	0.094	0.191	0.014	0.028	0.045	0.046	0.018	0.025
Cr ₂ O ₃	0.01	0.02	0.02	0.03	0.01	0.02	0.01	0.02	0.02	0.01
CO ₂	2.30	1.70	3.30	1.80	1.10	7.60	11.80	0.80	5.50	4.40
H ₂ O	<u>2.60</u>	<u>2.90</u>	<u>2.90</u>	<u>3.20</u>	<u>3.30</u>	<u>3.80</u>	<u>4.10</u>	<u>4.20</u>	<u>4.20</u>	<u>4.40</u>
Total	99.76	98.20	98.28	100.03	99.98	100.14	99.93	100.43	99.30	99.91
As	1	1	9	4	0	4	7	10	2	3
Br	0	4	5	2	8	3	3	0	0	6
Mo	0	0	0	0	0	1	1	0	0	0
Nb	15	17	16	13	13	10	8	12	7	15
Pb	17	18	22	25	20	13	15	21	9	9
Rb	185	34	92	32	57	56	45	103	54	70
Sr	59	182	127	176	41	80	228	66	54	31
Th	16	12	14	11	12	13	6	8	3	15
V	0	0	7	5	5	2	1	0	0	2
Y	46	60	54	62	56	67	44	51	53	58
Zr	318	296	262	289	183	165	120	266	136	275
Q	47.33	24.09	23.22	23.92	44.17	25.69	14.65	30.87	23.31	31.74
C	4.10	0.69	2.90	1.17	6.35	9.27	7.78	4.58	8.46	9.98
OR	22.77	11.98	18.73	11.79	7.72	12.37	13.86	12.91	7.71	11.21
AB	4.35	7.99	13.31	9.61	13.14	9.68	2.64	13.19	13.34	4.43
AN	2.08	23.39	10.57	21.90	0.39	0.0	0.0	8.41	1.56	1.37
DI	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0
HE	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0
EN	3.46	8.73	5.43	9.23	7.81	5.67	16.02	11.47	9.82	10.89
FS	6.54	12.68	10.81	0.0	9.11	13.83	12.01	9.44	15.98	13.89
FO	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0
FA	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0
MT	2.81	3.04	3.49	4.21	4.52	1.81	2.57	4.02	3.51	2.43
IL	0.76	2.59	2.93	2.57	2.97	2.92	2.07	2.31	2.59	2.80
CR	0.01	0.03	0.03	0.01	0.01	0.01	0.01	0.03	0.03	0.01
AP	0.31	0.68	0.70	0.69	0.91	0.36	0.26	0.53	0.32	0.68
PY	0.06	0.04	0.0	0.10	0.08	0.36	0.17	0.36	0.19	0.06
CC	5.38	4.05	7.87	4.23	2.59	17.98	27.94	1.89	13.15	10.48

Major elements in wt.%; minor elements in ppm; normative calculations in wt.%

Table C-7. Whole-rock analyses from Member F

	143	126-5	21-7	126-3	121-10	121-7	161	112-1	72-1	Member E 70
SiO ₂	54.90	50.20	57.10	50.60	49.10	46.50	52.50	42.60	64.30	53.70
Al ₂ O ₃	12.90	13.60	13.60	13.60	15.00	15.60	13.60	13.90	4.80	13.50
Fe ₂ O ₃	3.80	1.60	1.60	1.30	1.50	1.00	2.70	0.70	1.80	2.50
FeO	5.60	6.20	5.20	6.30	6.40	5.10	6.50	4.60	3.60	11.00
MgO	4.94	2.66	3.33	4.39	3.76	5.24	6.02	7.90	1.82	3.99
CaO	7.94	9.92	7.33	9.62	11.20	10.30	9.42	11.80	9.22	3.31
Na ₂ O	4.40	3.10	2.60	3.30	2.80	3.20	1.00	0.80	0.00	2.40
K ₂ O	0.05	1.23	1.14	1.08	0.89	0.87	0.88	1.56	3.29	0.89
TiO ₂	0.65	0.61	0.60	0.62	0.65	0.75	0.59	0.63	0.48	1.54
P ₂ O ₅	0.07	0.10	0.11	0.12	0.10	0.11	0.04	0.09	0.11	0.30
MnO	0.14	0.15	0.13	0.17	0.15	0.13	0.19	0.07	0.09	0.20
S	0.0	0.04	0.02	0.03	0.01	0.0	0.02	0.03	0.05	0.16
Ba	0.008	0.068	0.073	0.066	0.052	0.027	0.022	0.012	0.007	0.029
Cr ₂ O ₃	0.02	0.04	0.04	0.05	0.05	0.03	0.03	0.05	0.01	0.02
CO ₂	1.90	6.40	2.50	5.60	4.70	6.50	0.80	8.70	7.20	1.70
H ₂ O	2.50	3.10	3.20	3.20	3.70	2.80	4.60	6.00	2.20	4.50
Total	95.66	98.95	98.96	99.99	100.01	99.13	98.89	99.43	98.97	99.71
As	0	3	1	3	7	0	4	0	5	15
Br	1	1	1	3	8	0	2	2	4	2
Mo	0	0	0	0	4	1	4	0	1	2
Nb	6	9	7	6	9	8	5	6	5	17
Pb	11	11	13	11	16	4	11	18	12	19
Rb	1	27	23	24	18	17	19	38	252	32
Sr	51	54	119	81	124	78	32	32	17	88
Th	4	5	11	9	15	3	4	4	1	16
U	0	2	0	6	7	0	2	5	0	3
Y	33	31	36	35	40	36	31	28	21	67
Zr	100	21	138	17	120	140	82	109	81	272
Q	8.32	16.98	20.77	11.71	8.85	9.76	14.98	16.91	48.66	22.76
C	0.0	4.38	0.09	2.87	0.20	6.28	0.0	10.50	1.27	7.59
OR	0.30	7.59	7.04	6.60	5.47	5.39	5.52	9.87	20.10	5.53
AB	38.38	27.36	21.97	28.85	24.59	28.40	8.98	7.24	0.0	21.33
AN	15.78	8.45	22.74	11.92	26.16	9.74	31.86	3.16	0.0	3.89
DI	5.31	0.0	0.0	0.0	0.0	0.0	6.14	0.0	0.0	0.0
NE	3.74	0.0	0.0	0.0	0.0	0.0	3.75	0.0	0.0	0.0
EN	10.22	6.91	8.66	11.29	9.72	13.69	13.07	21.05	4.57	10.44
FS	8.62	9.61	7.62	10.00	10.02	7.88	9.17	7.33	4.43	16.40
FO	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0
FA	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0
MT	3.21	2.47	2.42	1.95	2.26	1.52	3.21	1.09	2.69	3.81
IL	1.27	1.20	1.31	1.22	1.28	1.49	1.19	1.28	0.94	3.07
CR	0.03	0.06	0.06	0.07	0.07	0.05	0.05	0.08	0.01	0.03
AP	0.17	0.24	0.26	0.29	0.24	0.27	0.09	0.22	0.26	0.73
PY	4.43	0.08	0.04	0.06	0.02	0.0	0.04	0.07	0.10	0.35
CC	4.45	15.18	5.43	13.16	11.09	15.51	1.43	21.18	16.92	4.06

Major elements in wt.%; minor elements in ppm; normative calculations in wt.%.

Appendix C. Method of construction and validity of spidergrams

Spidergrams are used to compare igneous rocks by displaying the proportions of lithophile and incompatible elements in a manner similar to rare-earth-element diagrams. Each element is divided by its abundance in chondritic meteorite (representative of undifferentiated earth composition) and these points are then connected by a single line. The shape of this profile is the signature of that sample. A straight line at sample/chondrite = 1 indicates no depletion or concentration of elements. Most terrestrial rocks, however, reveal enrichment of lithophile elements. Particular shapes of convex-upward curves are distinctive of broad classes of igneous rocks (see, for example, Thompson *et al.*, 1983).

The method of construction of spidergrams in Figure 7.12 follows that of Thompson *et al.* (1983), with two modifications. 1) Most rare-earth elements were not analyzed and are not shown (Ta, Ce, Nd, Hf, Tb, Tm and Yb). The analyzed elements are shown in the correct order (general decrease in incompatibility to the right), but the patterns are less constrained or more 'coarsely tuned' than if the full suite was available; 2) Spidergrams are generally normalized (to minimize absolute differences in element abundance) so that Yb = 10 (Thompson *et al.* 1982, p. 169). Because Yb was not analyzed for these samples, the slightly more incompatible element yttrium was used. Among these Mount-Harper samples, Y varies from 18 to 37, and other elements were recalculated on the basis of $Y = 30$ (which, after division by chondrites, shows as 15 on the spidergrams). This simplification does not alter the shape, but affects the relative abundance indicated by the patterns. This is a prime distinction between average patterns of the Upper and Lower suites. Because the proportion of Y relative to other elements is greater in the Upper Suite, the normalizing procedure increases the apparent difference between the two suites.

A problem in the use of spidergrams for the Mount Harper rocks is the analytic uncertainty resulting from very low concentrations of these elements in basic rocks, which approach the detection limits of XRF analysis. These errors in analytic precision are exacerbated by the logarithmic scale in plotting. Figure C-1 shows error bars for the limit of uncertainty given in the analyses (see Table C-1). The greatest potential error occurs for Rb, Th and Nb. Despite the latitude that could be expected for these elements, the similar values for samples of the same member collected from separate locations in the Complex suggests these diagrams are valid for hydrous, low-grade metamorphic volcanic rocks. If a systematic error was introduced by the analysis or calculations, the distinctive patterns of various members, and the contrast between them, would not be so apparent. Thompson *et al.* (1982) further demonstrated the robustness of the spidergram approach by noting the similarity of patterns of dolerite samples from different continents.

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Fig. C-1. Spidergram of Less Altered dyke sample (B158-1), showing potential uncertainty for each element. Error bars are $\pm$ (total absolute error) or (estimate of validity) from Table C-1, plotted on the logarithmic abscissa.

Table C-8. Data and Methods for Spidergram construction

Chondrite factors:	Th	0.042	K	120	K ₂ O*	0.21
	Nb	0.35	Ti	620	TiO ₂	0.103
	Rb	0.35	P	46	P ₂ O ₅	0.021
	Sr	11.8	Ba	6.9	880	7.7
	Zr	6.4	Y	2		

Point on graph = element content / chondrite value $\times \frac{(15)}{Y/2}$
(in ppm)

* Oxides given in % are recalculated with oxygens removed:
= oxide(%) / chondrite value $\times 0.00001 \times \frac{(\text{atom. wt. oxide}) \times (15)}{\text{atom. wt. element} \times Y/2}$

Precision error for Sample B158-1 (Medium-grained basaltic dyke)

element	Analyzed value	Precision	Minimum	Maximum
Ba	0.009	$\pm 0.002 \%$	0.007	0.011
Rb	7	$\pm 10 \text{ ppm}$	0	17
Th	1	$\pm 20 \text{ ppm}$	0	21
K ₂ O	0.29	$\pm 0.05 \%$	0.24	0.34
Nb	2	$\pm 10 \text{ ppm}$	0	12
Sr	63	$\pm 10 \text{ ppm}$	53	73
P ₂ O ₅	0.04	$\pm 0.02 \%$	0.02	0.06
Zr	28	$\pm 10 \text{ ppm}$	18	38
TiO ₂	0.54	$\pm 0.02 \%$	0.52	0.56
Y	24	$\pm 10 \text{ ppm}$	14	34

4

NEC
NATIONAL ELECTRONIC CORPORATION
1000 LEXINGTON AVENUE, SUITE 1000, NEW YORK, N.Y. 10017-4803

H

Upper West Trib valley

H'

Fig. B-1. (2 sheets)
Serial Cross sections
of
Mount Harper Volcanic
Complex

1000 2000m

No vertical exaggeration. Dikes, faults and
isolated outcrops enlarged for clarity

Sections located on Fig. 5.1 (also illustrations)

H Upper West Trib valley

H'' 2000 m

P"

2000 m

2000 m

P''
2000 m

Q'
2000 m

M

Soft Hills

M'

Legend

ODRR Road River Formation

ED_h "CD_h" Formation

Hy Hyland Group

Harper Group

UHG Upper Harper Group

Mount Harper Volcanic Complex
Member

F F
E E, hydroclastic breccia
D D, crystal tuff, dikes
C C, paleo-erosion surface
B, B_h B, hematized, related intrusion
A, A_p A, pillow facies

LHG Lower Harper Group

P Pinguicula Group

W Wernecke Supergroup
(Gillespie Lake Gp. above dashed line)

Present erosion surface
Dip (apparent) of layered rocks
in plane of section

Direction of movement on faults
(in plane of section only)

Out of plane, not plane of section

Late Mesozoic Contraction Faults

- 9 Low angle thrusts, incl "sub-CD_h"
- 8 Late South Harper Fault
- 7 North Harper Fault
- 6 Southeast-striking faults

Late Proterozoic Extension Faults

- 5 Tango Tarn Fault
- 4 Harper Ridge Fault
- 3 South Harper Fault (syn volcanic)
- 2 West Trib fault
- 1 Early South Harper Fault

U
SOUTH

5000'

South Harper
Fault
Late 1 Early

8: Late South Harper Fault

Late Mesozoic Contraction Faults

- 9 Low angle thrusts, incl "sub-CD_b"
- 8 Late South Harper Fault
- 7 North Harper Fault
- 6 Southeast-striking faults

Late Proterozoic Extension Faults

- 5 Tango Tarn Fault
- 4 Harper Ridge Fault
- 3 South Harper Fault (syn-volcanic)
- 2 West Trib fault
- 1 Early South Harper Fault

U
SOUTH

5000'

South Harper
Fault
Late Early

U'
NORTH 2000 m

8 Late South Harper Fault

Harper Group

UHG

Upper Harper Group

Mount Harper Volcanic Complex

Member

F

F

E

E, hydroclastic breccia

D

D, crystal tuff, dikes

C

C, paleo-erosion surface

B, Bh

B, hematized, related intrusion

A, Ap

A, rhyolite facies

LHG

Lower Harper Group

P

Pinguicula Group

W

Wernecke Supergroup

(Gillespie Lake Gp. above dashed line)

Present erosion surface

Dip (apparent) of layered rocks
in plane of section

Direction of movement on faults
(in plane of section only)

out of plane, into plane of section

8 Late South Harper Fault

7 North Harper Fault

6 Southeast-striking faults

Late

Proterozoic Extension Faults

5 Tango Tarn Fault

4 Harper Ridge Fault

3 South Harper Fault (syn-volcanic)

2 West Trib fault

1 Early South Harper Fault

5000'

3000'

0

U

SOUTH

5000'

3000'

0

8 splay

South Harper
Fault
Late Early

LHG

8

1

8 Late South Harper Fault

7 North Harper Fault

6 Southeast-striking faults

Late Proterozoic Extension Faults

5 Tongo Tarn Fault

4 Harper Ridge Fault

3 South Harper Fault (syn-volcanic)

2 West Trib fault

1 Early South Harper Fault

5 of/de 5

Outliers correlated with the Upper Harper Group.

"BALDIE" outlier

"SADDLE" outlier

SYMBOLS

Limit of continuously exposed rock, strike and dip indicate small, isolated outcrops

Stratigraphic contacts

- — — — — between units defined, approximate
- — — — — between volcanic members observed, inferred

Faults

- — — — — observed, inferred
- — — — — ball on downthrown side (steep, normal or reverse)
- — — — — inferred strike-slip
- — — — — low and moderate-dipping thrusts teeth on over-riding plate

↗ Strike and dip of bedding (strata formerly horizontal)

↗ Strike and dip of volcanic layers (depositional slope unknown)

↗ Strike and dip of pillowed flows (bulb shows present dip direction, arrow indicates local flow, not corrected for tilt)

/// Dikes and sills (commonly clustered and not drawn to scale strike and dip where noted)

Early Paleozoic

Road River For

CDb Formation

Lower Proterozoic

HARPER GROUP

Upper Harpe

Mount Harper

Member F

Member E

Member D

Member C

Member B

Bh

B₁

Member A

Ap

Lower Harpe

Middle and late Proterozoic

Pinguicula Gr

Wernecke Super
Gillespie L

W₆

W₀

Quartermaster

SYMBOLS

Limit of continuously exposed rock, strike and dip indicate small, isolated outcrops

Stratigraphic contacts

- — — — — between units defined, approximate
- — — — — between volcanic members observed, inferred

Faults

- — — — — observed, inferred
- ⌋ ball on downthrown side (steep, normal or reverse)
- ⌋ inferred strike-slip
- ⌋ low and moderate-dipping thrusts teeth on over-riding plate

✓ Strike and dip of bedding (strata formerly horizontal)

✓ Strike and dip of volcanic layers (depositional slope unknown)

✓ Strike and dip of pillowed flows (bulb shows present dip direction, arrow indicates local flow, not corrected for tilt)

/// Dikes and sills (commonly clustered and not drawn to scale; strike and dip where noted)

UNITS

Early Paleozoic

Road River Formation

Black argillite and chert; siltstone, quartzite and limestone in western part

"CDb*" Formation

Light grey dolostone; minor sandy dolomitic limestone

Late Proterozoic

HARPER GROUP

Upper Harper Group

Siltstone, dolomitic sandstone, para-conglomerate with volcanic clasts

Mount Harper Volcanic Complex

Member F

Basaltic sills and massive flows; typically orange weathering, diktytaxitic

Member E

Andesitic basalt flows, breccia and tuff; Hydroclastic breccia facies

Member D

Rhyolitic flows, breccia, and ignimbrite; Crystal tuff facies; includes volcanic arkose

Member C

Basaltic breccia; includes block and ash flows, tuff-breccia and volcanic conglomerate

Member B

Basaltic flows, breccia, dikes and sills; typically brown weathering, columnar jointed Hematized facies: highly vesicular, oxidized flows and breccia

B_h

Amygdaloidal facies: massive and pillowed flows with pipe amygdules, breccia

B_i

Intrusive facies: vesicular basaltic breccia

Member A

Basaltic flows, breccia and dikes; typically dark green weathering

A_p

Pillow facies: tubular pillows, aquagene tuff

Lower Harper Group

Conglomerate; minor sandstone, siltstone and carbonate breccia

Middle and Late Proterozoic

Pinguicula Group

Thick-bedded dolostone, minor dolomitic sandstone

Wernecke Supergroup

Gillespie Lakes Group Thick-bedded, deformed dolostone

W_G

W_O

Quartet Group

Phyllite, minor laminated meta-mudstone

8 Volcanic Complex

"SNOWCAP" outlier

Section Symbols

- Fines upward, downward
 limestones
PALEOSLOPE INDICATORS
 sl Slump folds
 xb Cross-beds
 (azimuth corrected for dip of beds)

10

11

Lower Harper Group (underlies the Volcanic Complex)

Pinguicula dolostone

Pinguicula dolostone

30 25

Fig. 4.1

GEOLOGY

drawn to scale strike and dip where noted)

Legend

B₁ Member

Ap

Lower Harpe

Middle and Late Proterozoic

P Pingicula Group

W Wernecke Supergroup

W₆ Gillespie Lake

W₆ Quartzite Group

W₀ Quartet Group

Phyllite, minor laminated meta-mudstone

50

drawn to scale strike and dip where noted

B

Intrusive facies: vesicular basaltic breccia

Member A

Basaltic flows, breccia and dikes:
typically dark green weathering

A_p

Pillow facies: tubular pillows, aquagene tuff

Lower Harper Group

Conglomerate, minor sandstone,
siltstone and carbonate breccia

Middle and late Proterozoic

Pingicula Group

Thick-bedded dolomite, minor dolomitic
sandstone

Wernecke Supergroup

Gillespie Lakes Group Thick-bedded, deformed dolomite

W_G

W₀

Quartet Group

Phyllite, minor laminated meta-mudstone

16
overlies Volcanic Complex

MAP LEGEND

- €D_b fm.
- Upper Harper Gp.
 - Dark shale, siltstone
 - Thin-bedded dolostone
 - Diamictite
- Volcanic Complex
 - Members D,E,F
 - Members A,B,C
- Lower Harper Gp.
 - Conglomerate
- Pinguicula Gp.
- Werneckie Sgp.

- Dolomite
- Siltstone
- Mudstone
- Diamictite
- Stratified
- Unstratified
- Conglomerate
- Stratified
- Unstratified

Fig. 3.1. HARPER GROUP
Sedimentary rocks associated with the Mount Harper Volcanic-Complex

C. Roots with assistance of J. Dennet, H. Smit, D. Handel, additional information from R. Thompson, P. Mustard

MAP LEGEND

ϵD_b fm.

Upper Harper Gp.

- Dark shale, siltstone
- Thin-bedded dolostone
- Diamictite

Volcanic Complex

- Members D,E,F
- Members A,B,C
- Lower Harper Gp. Conglomerate
- Pingicula Gp.
- Wernecke Sgp.

LEGEND for SECTIONS

Dolostone	Weathering colour
t tan
w grey
wh white	Contacts between beds
Limestone		— sharp, non-erosive
Sandstone		~ sharp, undulating
Siltstone		~ probably erosive
Mudstone		--- gradational
Diamictite		
stratified matrix, mud size		
unstratified matrix		
Conglomerate		
framework-supported cobbles, mudstone, sand matrix		
stratified pebbles, cobbles in sandy matrix		
Lapillistone		
Tuff		
Mafic intrusion		
		Clast Types
		rounded, subrounded
		sub-angular
		sedimentary
		mafic volcanic
		felsic
		jasper detritus

UPPER GROUP
 associated with
 Volcanic Complex

Fig. 4.1

GEOLOGY
of the
Mount Harper Volcanic Complex
Ogilvie Mountains, Yukon

Contour interval: 100 feet (30.5 m)

Geology by C. F. Roots, 1981, 1982, 1986

Base: Provisional topographic maps 116 B/12 and 116 C/9
produced 1968 by Surveys and Mapping Branch,
Department of Energy, Mines and Resources,
Canada.

(a)

(b)

(c)

(d)

(e)

(f)

(g)

(h)

(i)

(j)

(k)

(l)

(m)

CAPTIONS FOR INSET FIGURES

(a) North Harper Fault (NHF) separates S-dipping heterolithic volcanic breccias (C) from sediments (UHG), but is overlapped by, and therefore predates deposition of, Early Paleozoic dolostone (CD₁). The NHF may be syndepositional with UHG, because the black siltstone changes in thickness across the fault. Faults in background are post-Early Paleozoic and some are minor listric thrusts within UHG.

(b) North Harper Fault juxtaposes basal paraconglomerate of UHG (to the S) against younger black siltstone on the N. Repetition of tan and grey dolostone beds indicates minor thrust displacement on the S-dipping splay.

(c) The thrust fault, marked by offset of the unconformable base of Baldie outlier, and thin dolostone beds within it, appears to be connected with the vertical, SE-striking fault set to the south. Thrust movement shown here implies that oblique slip must have occurred on steeply dipping segments of the fault set.

(d) South-dipping volcanic units of irregular primary thickness are cut by faults that appear to steepen up-section. When volcanic units are restored to approximate primary dip (gently N), the dikes (left) are vertical and faults appear to have been listric.

(e) East-facing volcanic wall is exposed by erosion along steep-dipping, SE-striking faults, marked by deep gullies (oblique to this view) and dislocated layering. The apparent curve of fault splays shown on the map in part reflects their dip angle and high topographic relief.

(f) Similarity of two N-facing cliff sections of A capped by B suggests a S-dipping thrust in the valley between them. SE-striking faults at the head of the valley do not continue north of the thrust and appear to connect with it.

(g) Dolosiltstone and volcanic paraconglomerate beds are preserved by N-side down displacement on SE-striking fault offset. On the other side of the saddle (beyond this view), only a few metres of vertical offset can be measured on these faults, whereas adjacent fault to the NE (not visible in this view) shows tens of metres displacement. Differences in the amount of apparent vertical offset along faults suggests that the SE-striking set is a coherent transcurrent system.

(h) The ridge crest extending S of Mount Harper (here in profile) consists of three or more 200+ m thick sections of A and B that may be imbricated by SE-striking faults in gullies oblique to this view.

(i) Late South Harper Fault (LSHF) drops rhyolitic flows (D) on S side against older flows (A) to the north. Further north, prominent gullies and truncation of layered maroon flows (B) suggest a N-side-down fault. It may be coincident with the pre-volcanic Early South Harper Fault and suggests reactivation of the half-graben margin during or after volcanism.

(j) North-dipping LSHF brings basaltic breccias of B against rhyolitic D flows. If the fault predates folding, restoration of tilted beds indicates that it moved as a steep extension fault; alternatively it is a post-deformation S-directed reverse fault. Note also the S-tilted B flows, unconformably overlain by D, which indicates uplift and erosion of southern parts of the Complex before felsic extrusions (during deposition of C).

(k) Early Paleozoic dolostone (CD₁) appears to onlap more steeply dipping flows of K and E. Alternatively, several metres of black mudstone at the

(h) Early Paleozoic dolostone (CD₁) appears to overlap more steeply dipping flows of E and F. Alternatively, several metres of black mudstone at the contact indicate a possible fault; the nature of movement is unclear but it may be connected with thrusts that overlie the western edge of the Volcanic Complex (in UHG; see inset a). The LSHF crosses the foreground spur, but the lack of visible vertical offset across the saddle in background suggests that it is not or changes to shallow dip.

(i) Late South Harper Fault (LSHF) crops rhyolitic flows (D) on S side against older flows (A) to the north. Further north, prominent gullies and truncation of layered basalt flow (B) suggest a N-tilt-down fault. It may be coincident with the pre-volcanic Early South Harper Fault and suggests reactivation of the rift-shaft margin during or after volcanism.

(j) North-sipping LSHF crops basaltic breccias of B against rhyolitic D flows. If the fault predates folding, restoration of tilted beds indicates that it moved as a steep extension fault; alternatively it is a post-deformation S-tipped reverse fault. Note also the S-tilted B flows, unconformably overlain by D, which indicate uplift and erosion of southern parts of the Complex before felsic extrusions (during deposition of C).

(k) Early Paleozoic dolostone (CD₁) appears to onlap more steeply dipping flows of E and F. Alternatively, several metres of black mudstone at the contact indicate a possible fault; the nature of movement is unclear but it may be connected with thrusts that overlie the western edge of the Volcanic Complex (in UHG; see inset a). The LSHF crosses the foreground spur, but the lack of visible vertical offset across the saddle in background suggests that it is not or changes to shallow dip.

Fig. 5.1.

STRUCTURE of the Mount Harper Volcanic Complex

- Faults: observed, projected, assumed
(dip of fault plane)
- U/D
upthrown/downthrown side
- (100 m)
estimated vertical offset;
- ↗ suspected oblique slip and strike
slip displacement - see text)
- Mesozoic
- Thrust faults (older over younger
strata; dip of fault plane)
- Probable thrusts (low-angle truncation of units)
- Folds, meso-scale: strike and
dip of axial plane, plunge
(excluding igneous flow folding)
- Cleavage: inclined
- Joints: inclined, vertical
(excluding igneous cooling joints)
- Lineation: plunge; ss = slickensides
- Representative strike and dip of
bedding, volcanic layering: inclined
- Major drainage, hillside spring

Line of cross-section

General location of inset figure²
with direction of view

- ODRR Road River Fm.
- ↗ ↘ Lower Paleozoic dolostone
autochthon/allochthonous
- NHF North Harper Fault
- LSHF Late South Harper Fault
- A, B, C, D, E, F Volcanic members
- N, S, E, W Cardinal points in inset figures
- Pp Pinguicula dolostone
- W Supergroup

¹ See Fig 5.2 (pocket)

² sketched from photos

Area and scale same as in Figs. 3.1 and 4.2 (in pocket)

END

20.08.88

FIN