PETROLOGY AND EVOLUTION OF THE
RED MOUNTAIN OPHIOLITE COMPLEX,
NEW ZEALAND†

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ABSTRACT. The Red Mountain area comprises a relatively coherent ophiolite sequence with a maximum exposed "crustal" sequence of about 4 km. A basal zone of residual harzburgite is tectonically overlain by up to 1250 m of altered and cyclically interlayered cumulate dunite, wehlite, clinopyroxenite, troctolite, anorthositic gabbro, and clinopyroxene-hornblende gabbro. This unit grades upward into 200 to 400 m of hypautomorphic hornblende metagabbro and diorite. The plutonic units are cut by a complex of cross-cutting dikes of mafic and intermediate, with minor alkali, compositions. The percentage of dike material increases upward; gabbros and dikes are overlain by a series of metavolcanic units 600 to 1200 m thick. The metavolcanic rocks are in turn unconformably overlain by up to 600 m of well-bedded meta-hyaloclastic breccias. Mafic dikes intrude the metavolcanic and metasedimentary units. Relict igneous clinopyroxene and hornblende in the cumulate rocks show Fe-enrichment in the more evolved rocks, although many samples show significant metamorphic recrystallization. In view of the relations at Red Mountain, the "high-temperature contact" aureole around the Nelson Red Hills mass is reinterpreted as reflecting regional metamorphic partial recrystallization of mainly igneous plutonic rocks. Field relations and chemical compositional variations indicate a continuous magmatic history for the ophiolite that can be separated into three evolutionary stages. Stage I activity includes derivation of the harzburgite as a residue of partial melting and formation of the "crustal" cumulates. Stage II activity is dominated by intrusion of dikes and eruption of volcanic rocks that are compositionally similar to modern abyssal spreading center basalts. Stage III activity is dominated by more differentiated volcanism with compositions more typical of island arc environments. Overall petrology and stratigraphy indicate that Red Mountain ophiolite generation was mainly by processes similar to that producing abyssal oceanic basalts but in an environment proximal to an island arc.

INTRODUCTION

The area around Red Mountain in northwest Otago, New Zealand includes a portion of the Permian Dun Mountain ophiolite belt (Coombs and others, 1973; Coombs and others, 1976), a linear feature exposed for at least 170 km in the southern part of South Island, New Zealand (fig. 1). Along most of its strike, the Dun Mountain belt occurs as a series of mélanges and tectonic slices with complex structural relationships between mafic and ultramafic lithologies. In contrast, the Red Mountain area comprises a relatively coherent ophiolite sequence in fault contact with surrounding metasedimentary and metavolcanic units (fig. 2).

The Red Mountain ophiolite complex consists of rocks of the Humboldt Group (Waterhouse, 1964) and includes the Red Mountain Ultramafites (Red Mountain Ultramafics of Grindley, 1958) and the Livingstone Subgroup (Landis, 1974). The Red Mountain Ultramafites are dominantly harzburgite tectonites and are discussed in detail elsewhere (Sinton, ms and 1977). The Livingstone Subgroup in this area comprises a more or less continuous sequence of vertically dipping, to slightly overturned, altered basic igneous and volcaniclastic sedimentary rocks. Cyclic units of cumulate gabbros and clinopyroxene peridotites and

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pyroxenites at the base grade stratigraphically upward to less basic gabbros that are in turn overlain by metavolcanic and metasedimentary greenschists (fig. 3). A complex of cross-cutting dikes cuts the section with about 5 to 10 percent dike material in the harzburgite and lower layered gabbros and peridotites, increasing to almost 100 percent dikes in some places near the hornblende metagabbro-metavolcanic contact. The metavolcanic rocks are cut by dikes making up 5 to 15 percent of the outcrop area, whereas only about 1 to 3 percent of the area shown as metasediment on figure 2 consists of dikes (fig. 3).

The Red Mountain ophiolite complex has undergone two main periods of regional metamorphism (Sinton, ms). An early period of epidote-amphibolite facies metamorphism is best developed in the gabbros and peridotites and may have formed closely associated with magmatism. All parts of the ophiolite complex display evidence of a pervasive overprinting of pumpellyite-actinolite facies metamorphism with spasmodic development of soda amphibole. Lawsonite-albite-chlo-

Fig. 1. South Island, New Zealand, showing the location of the Dun Mountain ophiolite belt and other main geologic provinces (after Coombs and others, 1976). Red Mountain is in northwest Otago, near the northern limit of the portion of the ophiolite belt south of the Alpine Fault.
Fig. 2. Geologic map of the Red Mountain area. The ophiolite complex comprises the Humboldt Group in this area. Geologic mapping was undertaken by the author in 1971-1975.
rite assemblages dominate Bryneira Group sediments of the area (Landis, ms). The polymetamorphic history of most of the Dun Mountain belt has been demonstrated by Coombs and others (1976); the earlier high temperature phase has been correlated with ocean-floor metamorphism, whereas the later higher P/T event probably occurred with emplacement of the ophiolites during the Mesozoic Rangitata Orogeny. The rocks of the Red Mountain area have been variably affected by these metamorphic events, and, except for some tectonite harzburgites, there are no alteration-free samples in the area. The prefix meta-can therefore be taken as understood in much of the following discussion.

STRATIGRAPHY OF THE OPHIOLITE

**Peridotites and gabbros.**—The basal tectonite harzburgites of the Red Mountain Ultramafites are described elsewhere (Sinton, 1977). The Livingstone Subgroup comprises the “crustal” part of the ophiolite complex and is separated from Red Mountain Ultramafites by the Peanut Fault (Sinton, 1977), a shear zone with unknown displacement. Cumulate dunites and/or clinopyroxene-rich ultramafic rocks including clinopyroxenites, wehrlites, and feldspathic wehrlites form the stratigraphically lowest-exposed portion of the Livingstone Subgroup west and north of the Peanut Fault (fig. 2). Areas designated Livingstone peridotite on figure 2 are areas where ultramafic rocks are present, although commonly associated with gabbroic rocks. Some distinguishing characteristics between Livingstone peridotites and Red Mountain Ultramafites are summarized in table 1.

Livingstone peridotites are ratio layered with the amount of clino-
pyroxene controlling the aspect of the rocks. Layer boundaries are generally sharp but are locally gradational. Locally clinopyroxene oikocrysts up to 9 cm across enclose cumulus olivine grains, now altered to serpen-

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Fig. 3. Schematic structure section of the Red Mountain ophiolite complex. The distribution of igneous phases in the plutonic part of the Livingstone Subgroup is partly inferred from replacement minerals and textures. Rh: Red Mountain harzburgite, Rt: transitional peridotite, Lp/Lg: Livingstone peridotite and cumulate gabbro, Lhp: hornblende metagabbro, Lmv: metavolcanic units (upper—pillowed, lower—non-pillowed), Lms: metasediments.
TABLE 1
Distinctive characteristics of individual ultramafic units of the Red Mountain area

<table>
<thead>
<tr>
<th>Red Mountain Ultramafites</th>
<th>Livingstone Subgroup</th>
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<tr>
<td></td>
<td>Hornblende</td>
</tr>
<tr>
<td>Dominant lithologies</td>
<td>herzolite</td>
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<td>Textures</td>
<td>metamorphic tectonite</td>
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<td>Accessory lithologies</td>
<td>orthopyroxenite, dunite, mafic dikes</td>
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<tr>
<td>Accessory minerals</td>
<td>Cr-spinel, clinopyroxene</td>
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<tr>
<td>Clinopyroxene Cr2O3/(MgO+FeO)</td>
<td>0.94-0.55</td>
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</table>

tine (pl. I-A). Brown hornblende occurs as post-cumulus overgrowths on cumulus clinopyroxene and plagioclase in some feldspathic clinopyroxenites (for example, 37095) and gabbros but also occurs as cumulus grains in some clinopyroxene-hornblende gabbros (for example, 37041).

With increase in feldspar, the peridotites grade into mafic cumulates (pl. I-B), including clinopyroxene gabbros, clinopyroxene-brown hornblende gabbros, and brown hornblende gabbros. Brown hornblende gabbros have adcumulus or heterocumulus interlocking textures. Inter-cumulus clinopyroxene and brown hornblende locally form oikocrysts, poikilitically enclosing cumulus olivine, feldspar, and/or clinopyroxene. Iron–titanium oxides (now variably altered to sphene) locally make up 10 to 15 percent of some cumulate gabbros, but no highly concentrated oxide segregations have been observed. Sulphides, especially pyrrhotite, are present in many gabbros and are abundant (up to 1-2 percent) in ultramafic rocks. Feldspar in gabbros has altered to clinozoisite and/or pumpellyite with no more than 1 to 2 percent albite in the rock, attesting to the original calcic nature of the parent plagioclase. Hence, many of these rocks were probably originally eucrites. Chlorite pseudomorphs indicate that olivine was a primary constituent of most of the more basic gabbros. Some gabbros show rhythmic layering over a few centimeters between clinopyroxene-rich and hornblende-rich variants.

Feldspathic cumulates also form part of the original section, particularly in the northern part of the area. These include plagioclase–olivine cumulates (troctolites) and plagioclase–clinopyroxene ± olivine cumulates (anorthositic gabbros). In some of these rocks, intercumulus clinopyroxene and/or hornblende form oikocrysts enclosing cumulus feldspar.

The layered gabbros and peridotites show pinch and swell structures and rapid changes in thickness. Minor irregular folding and other signs

3Five-digit numbers refer to samples catalogued in the Department of Geology, University of Otago.
A. Poikilitic wehlrite, Livingstone Subgroup peridotite. Dark areas are clinopyroxene oikocrysts now largely replaced by actinolite; lighter areas are serpentinized olivine. Pen scale is 14 cm long.

B. Interlayered wehlrite (dark layers) and clinopyroxene gabbro (37016). Livingstone Subgroup layered metagabbro. Base of photo spans 1.8 m.
of plastic deformation are common and were probably produced while the rocks were in a semisolid state. Lithologic sequences are not constant throughout the area, and individual layers are locally continuous for only up to a few hundred meters. Although such changes in the section from place to place may be partly due to early deformation before the rocks were fully crystallized, individual layers are not now nor were they probably ever continuous throughout the area. Although most lithologic sequences tend to be nearly constant over areas of only 1 to 3 km along strike, there is no direct evidence for separate magma chambers. The variations in section from place to place along strike may, in part, result from a single magma system that was continually being replenished by injections of new magma into a zone of lateral extension.

The sequence of clinopyroxene peridotite grading upward through various types of gabbros is repeated at least twice, and locally three or more times, in most places in the area (figs. 2, 3). Ultramafic horizons become progressively thinner in each successive repetition, and feldspar becomes progressively less calcic as indicated by an increasing amount of albite in the altered products. Hornblende also makes up an increasing portion of the assemblage within each successive cyclic unit. Post-cumulus growth on cumulus minerals becomes more pronounced up-section within cyclic units. Hornblende rims on pyroxene are increasingly abundant, and altered pseudomorphs after plagioclase have an outer rim rich in albite indicating late-stage growth of more sodic plagioclase on cumulus calcic feldspar. Increased adcumulus growth with structural height has also been noted for the Skaergaard Intrusion (Wager and Brown, 1967).

Interlayered gabbros and peridotites pass upward into a series of less basic gabbros and diorites (hornblende metagabbros in fig. 2). The contact between the layered series and the upper metagabbros is gradational but generally marks a boundary where cumulus textures are no longer evident, and the rocks take on a typically hypautomorphic texture. The overlying non-cumulus gabbro unit has been designated hornblende metagabbro (fig. 2) although hornblende and pyroxene are locally completely replaced by actinolite, and some rocks have intermediate compositions (see below). Hornblende, where present in the upper series, is typically green. Up to 5 percent quartz is present in some samples and, in contrast to the layered gabbros, metamorphic albite is an essential constituent.

Grain size varies over a few centimeters from medium grained to pegmatitic, both laterally and up-section. Pegmatitic rocks with amphiboles up to 7 to 10 cm long are relatively common in the upper parts of the section. Hornblende pegmatites have textures suggesting that the amphiboles have formed by replacement of pyroxene.

*Livingstone dike complex.*—A complex of mafic and intermediate composition dikes cuts other lithologies in the Red Mountain area. Dikes range from 20 cm up to 5 m with most from 1 to 1.5 m thick. The dikes tend to be cross-cutting (pl. 2-A), although a preference for northeast-southwest strikes is evident (Sinton, ms). Many dikes have
A. Cross-cutting dikes, Livingstone dike complex. A fine-grained greenstone dike trends parallel to the scale and cuts metabasic dikes. Scale is 14 cm long.

B. Light-colored albitite dike (37065) cutting metabasic dikes, Livingstone dike complex.
fine-grained (chilled) margins against country rock serpentinite, gabbro, or against earlier dikes. There is a tendency for the dikes to become progressively finer grained with decreasing age (as deduced from cross-cutting relationships) within a given outcrop area, but exceptions to this generalization occur locally. The percentage of dikes generally increases toward the metavolcanic contact, but the distribution of dikes is somewhat sporadic.

On field appearance, most of the dikes fit into three main groups. A series of medium-grained, amphibole-rich aphyric dikes with flow banded textures occur as highly deformed bodies within the gabbros. Some of these dikes contain relatively coarse, subhedral apatite grains which are otherwise rare in the area. Their occurrence in irregular-shaped bodies suggests that they represent disrupted early dikes. Amphibole-rich dikes intruded into the layered gabbros are typically rich in brown hornblende (37297), whereas those in the upper series of gabbros and diorites contain green hornblende. Amphibole-rich variants make up no more than a few percent of the dikes of the area.

A series of metabasaltic dikes are either medium grained and aphyric or feldspar-phyric and are typically gray in color. Nearly euhedral feldspar lath shapes (now altered to albite–clinozoisite or albite–pupeyellite) up to 7 mm long are in a medium grained, wholly recrystallized matrix.

Dikes of intermediate composition (table 6) alter to a green color and range from fine to medium grained, the finer grained varieties predominating. These dikes commonly include the youngest rocks as deduced from cross-cutting relationships in outcrop (pl. 2-A). They are completely recrystallized to greenschist facies assemblages and together with the metabasic dikes make up about 90 percent of the total dike population.

A few other exceptional dikes locally occur. Albitite and quartz–albitite dikes (plagiogranites) occur high in the section where dikes are particularly abundant. These rocks contain from 65 to 95 percent albite, with accessory quartz and generally less than 5 to 10 percent chlorite, actinolite, and/or epidote. The original composition is unknown, but the present mineralogical and chemical composition (analysis 7, table 6) suggests significant soda metasomatism. However, this metasomatism has apparently proceeded fairly selectively, since adjacent dikes do not show severe soda-enrichment, and albitite contacts are sharp (pl. 2-B). These observations plus the lack of any textures indicating replacement of mafic minerals indicate that these dikes probably had fairly acidic original compositions; they make up less than 1 percent of the crustal section.

Hornblende-phyric dikes are rare in the Livingstone Subgroup but a few examples occur in this area (for example, 37060 and 37021). Two virtually identical brown hornblende-phyric dikes (37060 and 37015) intrude metavolcanic greenschists. The brown hornblende phenocrysts are in a matrix now composed of phengitic mica, tremolite, and pumpel-
lyite. The mica content and high $K_2O$ (see analysis 8, table 6) are unusually high for rocks of this area.

Dikes cutting Livingstone metasediment south of the northwest trending fault in the southern part of the area are medium grained, high-iron metabasaltic rocks (for example, 37068). Although thoroughly recrystallized, the igneous texture of these rocks is still evident in thin section.

_Dikes intruded into Red Mountain Ultramafites._—Mafic and intermediate dikes are associated with serpentinized shear zones in the Red Mountain Ultramafites. These dikes are up to 4 m thick, and dike margins are everywhere separated by at least a few centimeters of serpentine from fresh peridotite. The thickest dikes tend to be medium grained with equigranular (doleritic) textures. Dikes less than 1 m thick locally have a porphyritic texture with feldspar and pyroxene phenocrysts up to 5 mm in a fine grained, holocrystalline groundmass. Dike thickness is generally correlated with the size of the serpentinized zone in which it is found, but dikes vary considerably in thickness along strike.

Dike alteration also varies as a function of the size of the zone it intrudes. Dikes in wide zones are invariably altered to a variety of metamorphic and metasomatic assemblages dominated by vesuvianite, diopside, and grossular, in contrast to the prehnite-rich rodinolites in other parts of the Dun Mountain belt (Challis, 1965; Coleman, 1966). Thin dikes separated by only a few centimeters of serpentine from fresh peridotite are commonly nearly free from alteration. Such evidence may indicate that at least some of the mafic dikes were intruded after the Red Mountain Ultramafites had undergone a period of serpentinization.

Unaltered porphyritic and doleritic dikes are dominated mineralogically by plagioclase and clinopyroxene. Plagioclase phenocrysts commonly show normal or oscillatory zoning. Clinopyroxenes have well-developed sector zoning, especially in porphyritic rocks. Many doleritic rocks are also slightly porphyritic with plagioclase and clinopyroxene phenocrysts. Brown hornblende is a minor groundmass phase in some dolerites. Neither olivine nor Ca-poor pyroxene has been observed. Other groundmass phases are labradorite, clinopyroxene, and Fe-Ti oxides.

_Livingstone greenstones and greenschists._—Massive greenstones overlie the gabbros and dikes with an unfaul ted contact. Locally the gabbros and dikes are overlain by a 10 to 20 m altered volcanic breccia layer which is in turn overlain by highly deformed pillow lavas. In some places, the gabbros are directly overlain by metasediments (fig. 2).

Where it is well-exposed the upper part of the Livingstone Subgroup at Red Mountain can be divided into three main units. A lower, more or less continuous, section of metamorphosed massive volcanics and dolerites, locally interbedded with metasediments, is overlain by a series of altered pillow volcanics with minor interbedded volcanioclastic mudstone and siltstone. This metavolcanic unit is in turn overlain by a well-bedded sequence of metamorphosed volcanioclastic breccias, silstones, and sand-
stones (figs. 2, 3). The lithologies present in this part of the Livingstone Subgroup at Red Mountain are similar to the Lee River Group at Red Hill in Nelson Province (fig. 1), described by Walcott (1969). However, the upper Livingstone sequence at Red Mountain is metamorphosed to higher grade than the assemblages at Nelson. More important is the apparent reversal of sequence where pillow lavas at Red Mountain overlie the massive volcanics and dolerites, a reverse relationship to that apparently present at Red Hill (Walcott, 1969).

Unequivocal interpretations of relationships within the metavolcanic sequence are hindered by the thorough recrystallization and textural alteration of these rocks. Near the base of the sequence, there is a high proportion of medium-grained rocks, which probably represent altered dolerites, and are possibly sills. Finer-grained rocks are spatially associated, but whether or not intrusive rocks are present is not clear from the field exposures.

A few volcaniclastic mudstones and siltstones are interbedded in the volcanic sequence. One of these is 30 to 60 m thick and separates the lower massive metavolcanics and dolerites from a sequence of up to 600 m of pillow lavas locally intruded by dikes (fig. 3). The upper series of volcanics are moderately foliated and contain highly deformed pillow structures. No radial fractures or variolitic textures have been observed in the outcrops interpreted as pillow lavas, and very little inter-pillow material was apparently present in the original piles. Some textural variations are observed in thin sections from cores and rims of inferred pillows, but metamorphic recrystallization has largely obliterated original textures. Facing directions based on the pillow structures are highly dubious due to their deformed nature but generally indicate younging toward the west.

In the northern part of the area, metasediments locally lie in direct depositional contact with gabbroic rocks, and, with the possible exception of the southernmost part of the area, the lower contact of the metasediments is apparently unconformable throughout the area (fig. 2). Livingstone metasediments are well-bedded with individual layers from 1 cm up to 2 m thick. Sediments interbedded with volcanics tend to be very fine grained, whereas those overlying gabbros and volcanics vary from fine to medium grained. Volcaniclastic breccias have clasts up to 1 m long of volcanic material (pl. 3). Volcanic fragments in breccias are petrographically indistinguishable from lithologies of the underlying volcanic units.

The field relations indicate that the sediments may be, at least in part, derived from the volcanic units that may pass laterally into them. Although these rocks now resemble metamorphosed tuffs, the close association with submarine volcanics suggests that they are altered hyaloclastic rocks. The well-bedded nature of the Red Mountain hyaloclastites indicates that they are probably redeposited volcanic material and do not simply represent in situ fragmentation products of submarine erup-
tion. The metasedimentary sequence has a maximum thickness of more than 600 m in this area, but the top of the sequence is not exposed. Livingstone metasediments are in fault contact with Bryneira Group and Barrier Formation metasediments to the west (fig. 2).

**IGNEOUS MINERALOGY**

All mineral analyses reported here were made on the University of Otago JEOL JX-5A electron probe microanalyzer, using an accelerating voltage of 15 kV, specimen current of about 0.03 μA, and 1-2 μm beam diameter. At least five countings of 10 sec integration time were averaged for each analysis. Count rates were standardized against a combination of natural and synthetic standards, as listed in Nakamura and Coombs (1973). Data correction was according to the method of Bence and Albee (1968).

Although orthopyroxene is abundant in the harzburgitic Red Mountain Ultramaftes, it was apparently not present in the mafic rocks of this area. However, orthopyroxene has been reported from what is probably a gabbro in related rocks from Nelson Province (Challis, 1965). This is

**PLATE 3**

Schistose volcaniclastic breccia with a large metavolcanic clast. Livingstone Subgroup metasediments. 15 cm scale rests on the metavolcanic clast which contains abundant quartz veins.
the only known occurrence of Ca-poor pyroxene from Livingstone Subgroup or correlative rocks in New Zealand. Pseudomorphs after olivine are common in some layered gabbros and peridotites, but, in contrast to Red Mountain Ultramafites, no relict olivine grains have been observed in Livingstone rocks of this area. Rieh plagioclase is present only in some of the mafic dikes that intrude Red Mountain Ultramafites. Igneous brown hornblende and clinopyroxene have survived alteration in some dikes and in many Livingstone gabbros and peridotites.

Plagioclase.—Plagioclase is preserved only in dikes intruding Red Mountain Ultramafites. Anorthite component varies from An 78 to An 74 in the cores of large plagioclase grains. Rims and groundmass grains range to An 52 (fig. 4). K₂O (0.01-0.05 wt percent) and Or (0.3 mole percent) contents of these plagioclases are low but probably increase with decreasing An (fig. 4). Low Or content in plagioclase probably reflects low total K₂O of the parent magma (for example, Ridley and others, 1974).

Clinopyroxene.—Analyzed clinopyroxene from Livingstone layered rocks are presented in table 2. In general over the range of Fe/Mg represented by the analyzed samples, MnO, FeO, 1/CaO, and 1/MgO vary sympathetically with Fe/Mg in the clinopyroxenes of the layered rocks. Since it is problematic to correlate specific lithologies across faults, the variation in Ca–Fe–Mg of Livingstone clinopyroxene (fig. 5) can, however, only tentatively be correlated with structural height in the section.

Clinopyroxene cores tend to have little compositional variation within each plutonic rock and are not zoned, but clinopyroxenes in the three most basic analyzed rocks do have narrow, optically indistinguishable, compositional rims that are in sharp contact with the homogeneous cores. The rims are relatively lower in Al₂O₃, Cr₂O₃, and TiO₂ and higher in MgO and Mg/Fe compared to the cores. The Ti/Al ratio is nearly constant at about 1:8 in both cores and rims. Since these pyroxenes are low in Ti and Cr, the Al can be assumed to be present as CaTs (CaAl₂SiO₆) component. Several possible alternative interpretations can be considered as explanations for formation of the rims.

The rims may reflect the appearance of brown hornblende or plagioclase as crystallizing phases. However, some hornblende apparently formed after the clinoxyroxene rims and, as such, was not responsible for

![Graph](image)

**Fig. 4.** Or versus An mole percent of plagioclase in dikes intruded into Red Mountain Ultramafites: phenocryst cores (solid triangles), rims (open triangles), groundmass grains (circles).
the changing pyroxene composition. Plagioclase crystallization should raise the Ti/Al ratio of coexisting pyroxene, and this result is not observed.

As pressure increases, CaTs solubility in clinopyroxene coexisting with plagioclase increases in silica-saturated environments (Kushiro, 1965). These rocks probably did not form from silica-saturated liquids, and similar core-rim relationships occur in sample 37095 (without cumulus feldspar) as in the gabbroic rocks. Although these rims may have formed at lower pressures than the cores, pressure change probably is not a major contributing factor to rim formation.

Kushiro (1960) and Le Bas (1962) demonstrated that Al and Ti contents of clinopyroxene decrease as the silica activity of the host magmas increases. Although silica activity of the liquid would be expected to increase with advancing crystallization of mafic minerals with decreasing temperature, these minerals should also become more iron-rich. However, it has been shown experimentally that a reverse trend, that is a decrease in Fe/Mg ratio, occurs in olivine and pyroxene if oxygen fugacity increases with falling temperature (Speidel and Osborn, 1967). Oxygen fugacity and silica activity are interdependent in olivine-bearing rocks by the reaction

\[ 3 \text{Fe}_2\text{SiO}_4 + \text{O}_2 \rightleftharpoons 2 \text{Fe}_3\text{O}_4 + 3 \text{SiO}_2. \]

It is notable that rims were not detected in the clinopyroxenes of sample 37066 which probably did not originally contain olivine. The clinopyroxenes are not continuously zoned but rather have rims that apparently formed as overgrowths on the original cumulus cores. This supports the conclusion that the rims formed by adcumulus growth.

![Diagram](image)

*Fig. 5. Ca-Fe-Mg compositions of clinopyroxenes (solid symbols) and brown hornblendes (open symbols) in Livingstone Subgroup clinopyroxenite and clinopyroxene gabbros (squares) and hornblende gabbro and hornblende-phyric dike (circles). The lines connect analyses of minerals from individual samples.*
Table 2
Analyses of representative clinopyroxene enes from Livingstone Subgroup rocks

<table>
<thead>
<tr>
<th>Lithology</th>
<th>Clinopyroxenite</th>
<th>Gabbro</th>
<th>Mafic dike</th>
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<tr>
<td>Sample</td>
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<td>37041</td>
<td>37036</td>
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<td>Anal.</td>
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<td>100.7</td>
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</table>

Cations on the basis of 6 oxygens

| Si  | 1.934 | 1.930 | 1.915 | 1.956 | 1.893 | 1.917 | 1.931 | 1.896 | 1.934 | 1.879 | 1.941 |
| Al  | 0.138 | 0.118 | 0.136 | 0.099 | 0.156 | 0.171 | 0.175 | 0.179 | 0.098 | 0.177 | 0.101 |
| Ti  | 0.010 | 0.006 | 0.017 | 0.011 | 0.023 | 0.019 | 0.013 | 0.024 | 0.013 | 0.020 | 0.013 |
| Fe  | 0.132 | 0.122 | 0.211 | 0.197 | 0.218 | 0.224 | 0.263 | 0.182 | 0.171 | 0.170 | 0.208 |
| Mn  | 0.003 | 0.004 | 0.008 | 0.007 | 0.007 | 0.008 | 0.009 | 0.005 | 0.004 | 0.004 | 0.005 |
| Mg  | 0.851 | 0.884 | 0.797 | 0.823 | 0.791 | 0.800 | 0.797 | 0.852 | 0.957 | 0.879 | 0.962 |
| Ca  | 0.902 | 0.918 | 0.891 | 0.894 | 0.902 | 0.889 | 0.873 | 0.818 | 0.782 | 0.839 | 0.750 |
| Na  | 0.021 | 0.016 | 0.034 | 0.030 | 0.032 | 0.022 | 0.025 | 0.017 | 0.017 | 0.019 | 0.018 |
| Cr  | 0.019 | 0.011 | 0.006 | 0.000 | --     | --     | --     | 0.016 | 0.017 | 0.024 | 0.004 |
| Total | 4.010 | 4.007 | 4.015 | 4.007 | 4.022 | 4.010 | 4.016 | 3.991 | 3.999 | 4.011 | 4.003 |

| Mg/(Mg+Fe) | 0.866 | 0.879 | 0.791 | 0.807 | 0.784 | 0.781 | 0.752 | 0.824 | 0.848 | 0.838 | 0.822 |

*Total Fe as FeO.  a, b. Opposing sectors in sector zoned phenocrysts.
from trapped interstitial liquid of higher silica activity and oxygen fugacity than the original liquid from which the cumulus grains (cores) crystallized. The narrowness of the rims attests to the limited amount of adcumulus growth in rocks from lower parts of the section.

Clinopyroxene phenocrysts in unaltered dikes cutting Red Mountain peridotites typically have stubby prismatic habit and are commonly sector zoned. The (100) sectors are enriched in Mg and depleted in Al, Ca, and Ti relative to (010) and (110) sectors (table 2), similar to those described by Nakamura and Coombs (1973). In general, the Al content increases with Ca, Ti, Mg, and Fe/Mg. Fe and Cr contents show random variation in the sector zoned grains.

Normative compositions of clinopyroxenes from the mafic dikes are variable due to zoning and range from slightly Ne-normative to 30 percent Hy-normative (fig. 6). These pyroxenes fall within the tholeiitic field of Coombs (1963).

Brown hornblende.—Igneous brown hornblende forms post-cumulus overgrowths on cumulus pyroxene, olivine, and plagioclase, and

Fig. 6. Clinopyroxene (solid symbols) and brown hornblende (open symbols) from Livingstone Subgroup clinopyroxenite and gabbros (squares), hornblende gabbro and hornblende-phyric dike (circles) and from dikes intruded into Red Mountain Ultramafites (triangles), plotted in terms of their normative compositions. Tie lines connect coexisting cpx-hbl pairs. Compositional fields for pyroxenes from alkalic (a) and tholeiitic (t) basalts are after Coombs (1963).
possible cumulus grains in some gabbros. Analyses of hornblendes from some cumulate gabbros, peridotites, and dikes are given in table 3. Progressive increase in Fe/Mg of hornblende is correlated with a similar increase in coexisting clinopyroxene (fig. 5). Compositional changes in hornblende (as in pyroxene) apparently monitor progressive changes in the liquids from which they form.

Helz (1973) has shown that Ti and Al\(^{IV}\) increase with temperature in quartz–magnetite–fayalite buffered hornblendes crystallized in experiments on natural basalts. Kretz (1960), Onuki (1966), and Helz (1973) have demonstrated a correlation between temperature and Fe/Mg

### Table 3
Analyses of representative igneous hornblendes from Livingstone Subgroup rocks

<table>
<thead>
<tr>
<th>Lithology</th>
<th>Clinopyroxene</th>
<th>Gabbro</th>
</tr>
</thead>
<tbody>
<tr>
<td>Sample</td>
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<td>31852</td>
</tr>
<tr>
<td>Anal.</td>
<td>1</td>
<td>2</td>
</tr>
<tr>
<td>SiO(_2)</td>
<td>45.6</td>
<td>44.9</td>
</tr>
<tr>
<td>Al(_2)O(_3)</td>
<td>12.1</td>
<td>12.0</td>
</tr>
<tr>
<td>TiO(_2)</td>
<td>1.06</td>
<td>1.20</td>
</tr>
<tr>
<td>FeO*</td>
<td>6.47</td>
<td>6.63</td>
</tr>
<tr>
<td>MnO</td>
<td>0.20</td>
<td>0.12</td>
</tr>
<tr>
<td>MgO</td>
<td>16.8</td>
<td>16.8</td>
</tr>
<tr>
<td>CaO</td>
<td>11.9</td>
<td>12.3</td>
</tr>
<tr>
<td>Na(_2)O</td>
<td>2.56</td>
<td>2.56</td>
</tr>
<tr>
<td>K(_2)O</td>
<td>0.08</td>
<td>0.08</td>
</tr>
<tr>
<td>Cr(_2)O(_3)</td>
<td>0.98</td>
<td>1.00</td>
</tr>
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</table>

<table>
<thead>
<tr>
<th>Anhydrous Total</th>
<th>97.7</th>
<th>97.6</th>
<th>96.9</th>
<th>97.7</th>
<th>98.2</th>
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<th>97.8</th>
<th>98.0</th>
<th>97.6</th>
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<tr>
<td>Al(^{IV})</td>
<td>1.500</td>
<td>1.567</td>
<td>1.185</td>
<td>1.887</td>
<td>1.882</td>
<td>1.664</td>
<td>1.572</td>
<td>1.565</td>
<td>1.585</td>
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<tr>
<td>Al(^{V})</td>
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<td>0.255</td>
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<td>0.380</td>
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<tr>
<td>Ti</td>
<td>0.114</td>
<td>0.129</td>
<td>0.172</td>
<td>0.419</td>
<td>0.416</td>
<td>0.273</td>
<td>0.292</td>
<td>0.342</td>
<td>0.335</td>
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<tr>
<td>Fe</td>
<td>0.771</td>
<td>0.795</td>
<td>1.342</td>
<td>1.290</td>
<td>1.283</td>
<td>1.304</td>
<td>1.330</td>
<td>1.534</td>
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<tr>
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<td>0.027</td>
<td>0.018</td>
<td>0.020</td>
<td>0.020</td>
<td>0.020</td>
<td>0.031</td>
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<tr>
<td>Mg</td>
<td>3.588</td>
<td>3.601</td>
<td>3.373</td>
<td>3.048</td>
<td>3.027</td>
<td>3.164</td>
<td>3.038</td>
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<tr>
<td>Ca</td>
<td>1.822</td>
<td>1.895</td>
<td>1.770</td>
<td>1.822</td>
<td>1.812</td>
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<td>1.840</td>
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</tr>
<tr>
<td>Na</td>
<td>0.706</td>
<td>0.711</td>
<td>0.562</td>
<td>0.961</td>
<td>0.936</td>
<td>0.746</td>
<td>0.707</td>
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<tr>
<td>K</td>
<td>0.015</td>
<td>0.015</td>
<td>0.016</td>
<td>0.013</td>
<td>0.011</td>
<td>0.012</td>
<td>0.011</td>
<td>0.022</td>
<td>0.024</td>
</tr>
<tr>
<td>Cr</td>
<td>0.110</td>
<td>0.114</td>
<td>--</td>
<td>0.054</td>
<td>0.072</td>
<td>--</td>
<td>--</td>
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<td>Total</td>
<td>15.675</td>
<td>15.742</td>
<td>15.560</td>
<td>15.866</td>
<td>15.832</td>
<td>15.750</td>
<td>15.663</td>
<td>15.688</td>
<td>15.709</td>
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<tr>
<td>atomic Mg/(Mg+Fe)</td>
<td>0.823</td>
<td>0.819</td>
<td>0.715</td>
<td>0.703</td>
<td>0.702</td>
<td>0.708</td>
<td>0.696</td>
<td>0.658</td>
<td>0.651</td>
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</tbody>
</table>

*Total Fe as FeO.

n.d. — not determined.
Fig. 7. Variation of hornblende Ti and Al cations on the basis of 23 oxygens per unit formula with $K_0 (\text{hbd/cpx}) = \frac{X_{\text{hbd}}}{X_{\text{epx}}}^{\text{hbd}} \frac{X_{\text{epx}}^{\text{cpx}}}{X_{\text{epx}}^{\text{cpx}}}$ for coexisting cpx–hbd pairs in Livingstone Subgroup plutonic rocks. $K_0 (\text{hbd/cpx})$ has been shown to increase with temperature (Kretz, 1960; Onuki, 1966; Helz, 1973).

distribution in hornblende and coexisting clinopyroxene. Fe/Mg distribution data for Red Mountain coexisting pairs are listed in Table 4. There is a linear increase in Ti content of hornblende with $K_0 (\text{hbd/cpx})$ (fig. 7). Al$^{IV}$ increases and Al$^{VI}$ decreases with temperature, features previously noted in experimentally crystallized hornblendes (Helz, 1973).

Helz (1973) has also shown that hornblendes from a given bulk composition fall on a well-defined line in plots of Al$^{IV}$ versus A-site occupancy. Red Mountain hornblendes show a nearly linear positive correlation between Al$^{IV}$ and A-site occupancy (fig. 8), indicating that A-site occupancy increases with temperature, and that the Red Mountain hornblende compositions are consistent with crystallization from liquids of similar bulk compositions.

Analyses of pale brown hornblende from some mafic dikes in Red Mountain Ultramaafites are also listed in Table 3. Compared to hornblendes in Livingstone Subgroup samples, they have slightly lower Al contents and significantly lower Ca and Mg/Fe; they tend toward subcalcic compositions. Brown hornblendes in these rocks tend to form as rare interstitial grains in doleritic rocks (for example, 37070) and as rims on clinopyroxene. Extensively altered rocks have abundant hornblende, whereas it is rare in fresher samples. Hornblende in these rocks

**Table 4**

<table>
<thead>
<tr>
<th>Sample</th>
<th>$K_0$</th>
</tr>
</thead>
<tbody>
<tr>
<td>37095</td>
<td>1.12</td>
</tr>
<tr>
<td>37041</td>
<td>1.77</td>
</tr>
<tr>
<td>37036</td>
<td>1.49</td>
</tr>
<tr>
<td>37066</td>
<td>1.57</td>
</tr>
</tbody>
</table>
is probably a late or post-magmatic mineral, and much of it probably formed by deuteritic replacement of clinopyroxene. It is therefore not truly comparable in its growth history to the cumulus and post-cumulus hornblendes of the Livingstone rocks previously described. The difference in crystallization histories may be reflected in the rather different compositions. It is notable, however, that the normative compositions of the late magmatic-deuteritic brown hornblendes in Red Mountain dikes occupy a similar compositional field to the cumulus and post-cumulus hornblendes of the Livingstone gabbros and dikes on figure 6.

REINTERPRETATION OF THE RED HILLS "CONTACT METAMORPHIC" AUREOLE

Challis (1965) described mineralogical zonation around the Red Hills ultramafic mass in Nelson which she interpreted as a contact metamorphic aureole. Many of the features she described are also present at Red Mountain but need not have formed by contact metamorphic processes. The mineralogical zonation at Red Hills comprises away from the ultramafic contact (1) a pyroxene zone about 4 m wide with both ortho- and clinopyroxene, (2) a brown hornblende zone up to 70 m wide with minor pyroxene, (3) a brown-green hornblende zone about 80 m wide (Challis, 1965). Rodinmites composed variably of diopside, grossular, hydrogrossular, vesuvianite, prehnite, wollastonite, and pectolite occur in the brown hornblende and pyroxene zones.

Similar zones are crudely represented at Red Mountain within each cyclic unit. Although at least some orthopyroxene is present at Red Hills (Challis, 1965), the clinopyroxenes and brown hornblendes are essentially identical in composition to those at Red Mountain (compare Challis, 1965; and tables 2, 3). At Red Mountain, brown hornblende typically mantles clinopyroxene or other minerals and is

---

**Fig. 8.** $Al^{IV}$ versus total cations in the A-site of igneous hornblende in Livingstone Subgroup clinopyroxenite and gabbros (open squares), clinopyroxene–hornblende gabbro (37041—solid squares), hornblende gabbro (31852—circles) and hornblende–phyric dike (37060—triangles). A-site occupancy calculated after Helz (1973) as total cations per 25 oxygens minus 15.0 where all Fe is calculated as FeO. End member components tschermakite (Ts), pargasite (Pa), and edenite (Ed) are designated. See text for discussion.
in turn locally mantled by metamorphic green, blue-green, or colorless amphibole. Pyroxene and hornblende commonly occur as cumulate grains, and the rocks called hornfels by Challis (1965) can alternatively be interpreted as having igneous textures. Although Challis suggests the rocks near the Red Hill mass were originally tuffs and spilitic volcanics, the author has observed altered gabbros in significant amounts along the western contact of the body. The zonal arrangement of rocks dominated by clinopyroxene passing into brown hornblende-rich rocks away from the ultramafic contact is repeated in each cyclic unit at Red Mountain, and the mineral compositions reflect crystallization from a differentiating magma. Green and colorless amphiboles form post-magmatic overgrowths on earlier formed minerals throughout the Livingstone sequence, although green hornblende is most common in the section overlying the layered series. Processes producing green hornblende at Red Mountain apparently affected the entire section. A clinopyroxene-brown hornblende gabbroic specimen personally collected from the western contact of the Nelson Red Hills body also contains metamorphic blue-green amphibole (37017).

Minerals associated with rodingites, including clinopyroxene, at both Red Mountain and Red Hills almost certainly formed coincident with serpentinization (Coleman, 1966; Sinton, ms) and therefore do not reflect high-temperature contact metamorphism. In general, evidence for high temperature contact metamorphism is lacking at Red Mountain, and except for the minor presence of orthopyroxene and the lesser amount of exposed gabbro, relations at Red Hills do not appear to be different. It is therefore likely that the relationships at Red Hills also simply reflect regional metamorphic partial recrystallization of primarily igneous plutonic rocks.

**Composition of the Parent Liquid(s)**

The cumulus mineral assemblage of olivine, clinopyroxene, plagioclase, and hornblende, without Ca-poor pyroxene could suggest that parent liquids to the plutonic sequence were silica undersaturated. However, as shown below, these liquids fractionated to relatively silica-rich differentiates, indicating that initial compositions must have been on the SiO$_2$ side of the critical plane of silica undersaturation as defined by Yoder and Tilley (1962). Because of high water contents, differentiated liquids crystallized hornblende rather than orthopyroxene, similar to the plutonic sequence of the Point Sal ophiolite (Hopson and Franco, 1977). Olivine was on the liquidus of magmas giving rise to the layered plutonic rocks which have a tholeiitic iron enrichment trend; these magmas were probably olivine tholeiitic in composition.

As shown by Kushiro (1960) and more extensively by Coombs (1963), the compositions of pyroxenes in basaltic magmas can give an indication of the nature of the liquids from which they crystallize. On the basis of the Ti and Al contents of the pyroxenes in table 2, these minerals probably formed from transitional tholeiitic to mildly alkalic magmas. Normative compositions of clinopyroxenes from the
Livingstone layered series are slightly undersaturated and mainly plot on the low SiO$_2$ side of the diopside–forsterite join on figure 6. An analysis of one of the rims is Hy-normative, supporting the concept of growth from a liquid of relatively higher silica activity. All the analyzed pyroxenes fall within the tholeiitic or transitional fields of Coombs (1963) and are generally less Hy-normative than Hawaiian tholeiitic clinopyroxenes. Low alkali, high normative Di/Hy basalts similar in composition to ocean floor olivine tholeiites were unknown when Coombs (1963) defined his “transitional” type, and Livingstone Subgroup clinopyroxene compositions are consistent with crystallization from such liquids.

Brown hornblendes from the area have also been plotted in terms of their normative components on figure 6, and it is apparent that they tend to show corresponding degrees of silica saturation to coexisting clinopyroxenes. The variation of silica saturation (deviation away from Di–Ol join in fig. 6) is independent of iron-enrichment.

The widespread presence of magmatic amphibole in the cumulate sequence indicates that the parent liquids were significantly hydrated. It is not known whether this water was introduced by interaction of the magmas with sea water or whether it was primary magmatic. Nonetheless, the Livingstone parental magma apparently contained significant amounts of water, well before final solidification.

It should be noted that magmatic amphiboles are apparently fairly common in published accounts of other ophiolite complexes, for example the Point Sal ophiolite of California (Hopson and Frano, 1977). The Red Mountain ophiolite complex is similar to the Point Sal complex in other respects, most notably the abundance of clinopyroxene and lack of orthopyroxene in the cumulate sequence. Miyashiro, Shido, and Ewing (1970) noted the widespread presence of brown hornblende in tholeiitic gabbros from mid-ocean ridges, and similar rocks have been collected from the Romanche Fracture Zone (Prinz and others, 1976; J. M. Sinton, unpub. data). Hornblende is very rare in continental and arc tholeiitic suites (Miyashiro, Shido, and Ewing, 1970).

The plagioclase-rich cumulites in some parts of the Livingstone section indicate that, in contrast to most of the cumulate sequence, plagioclase locally accumulated before or with clinopyroxene. Although plagioclase starts to crystallize before clinopyroxene in many abyssal tholeiitic (for example, Sinton and Byerly, in press) and other basaltic magmas, due to differential settling rates, accumulation sequences may not accurately reflect crystallization sequences (Bryan and Moore, 1977; Hopson and Frano, 1977).

In summary, field relations, mineral compositions, and crystallization sequences indicate that the parental liquid(s) from which the Livingstone cumulate rocks crystallized were probably hydrous olivine tholeiitic in composition, similar to high normative Di/Hy ocean ridge tholeiites. Compositional data presented below show that these liquids were also very low in K$_2$O and P$_2$O$_5$ and relatively low in TiO$_2$. 

CRYSTALLIZATION CONDITIONS OF LIVINGSTONE PLUTONIC ROCKS

In each cycle, olivine appears to have been the first phase to crystallize, followed by clinopyroxene and/or plagioclase. In some samples, hornblende may have started to crystallize before plagioclase. Cohen, Ito, and Kennedy (1967) and Green and Ringwood (1967) found respectively that plagioclase + olivine and plagioclase are not stable solidus assemblages for several basaltic systems above 9 kb. Yoder (1965) also demonstrated a reaction relationship between plagioclase and olivine producing orthopyroxene above 8 to 9 kb. In view of these considerations, an upper limit of 8 to 10 kb total pressure can probably be assumed for the Livingstone plutonic assemblages.

Field relations at Red Mountain indicate that the plutonic sequence crystallized at a minimum depth of 2 to 3.5 km. Plagioclase starts to crystallize before amphibole at water pressures less than 2 to 2.5 kb in hydrous olivine tholeiite and alkali basalt (Yoder and Tilley, 1962; Holloway and Burnham, 1972), and at less than about 3 kb $P_{Na}$ in high alumina basalt from Medicine Lake Highland, Calif. (Yoder and Tilley, 1962). Several samples (for example, poikilitic gabro with cumulus feldspar, intercumulus amphibole) preserve this sequence, and much of the cumulate sequence can be considered to have formed below 2 to 3 kb. In some of the observed assemblages, amphibole may have started to crystallize before or coeval with plagioclase, suggesting that these sequences formed under prevailing water pressures above this limiting condition.

CHEMICAL COMPOSITION OF LIVINGSTONE SUBGROUP ROCKS

Chemical analyses for major elements have been obtained by X-ray fluorescence spectrometry at the University of Otago. Microprobe analyses of triplicate strip-melted glasses (Jezeck and others, 1978) have been averaged into some of the analyses. $Na_2O$ and $MgO$ were determined by atomic absorption, $FeO$ by titration, and $H_2O$ by fusion in Penfield tubes. $Cr$ and $Ni$ values were obtained colorimetrically by J. Norberg at the Smithsonian Institution, Washington, D.C.

Analyses of representative Livingstone Subgroup rocks from the Red Mountain area are presented in tables 5, 6, and 7. Except for analyses 1 and 2, table 6, all samples are at least partially recrystallized and contain water contents significantly above those that can reasonably be expected to reflect igneous processes. $Na_2O$ is also enriched in the metavolcanic rocks and in many of the dikes, probably indicating an early spilitic alteration of these samples. Other possible alteration effects are harder to assess. Consistent variations of particular element abundances with $H_2O$ contents are not apparent. Recent work on a variety of mafic samples by Melson, Thompson, and van Andel (1968), Smith (1968), Pearce and Cann (1973), Vallance (1974), Frey, Bryan, and Thompson (1974), Pearce (1975), Hajash (1975), Floyd (1976), Coish (1977) and Humphris and Thompson (1978a,b) indicate that few major elements can be considered to be immobile during alteration. Furthermore, these
references indicate that, of the elements listed in tables 5, 6, and 7, the elements Ti, Cr, Ni, and P are likely to be least affected by various hydrothermal and/or higher grade alteration processes.

Although the Livingstone Subgroup rocks of this area locally show variations in the degree of recrystallization, sharp metamorphic discontinuities are not present. However, chemical discontinuities commonly occur between layers in the cumulate sequence and across dike boundaries throughout the section; these discontinuities probably reflect compositional differences in the original igneous rocks. Although probable alter-

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<tr>
<th></th>
<th>1</th>
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<th>3</th>
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** Atomic

\[
\text{Mg} = \frac{0.892}{0.854} \text{Mg+Fe}_{\text{total}}
\]

*Analysis by microprobe.

**Cr and Ni colorimetric analyses by J. Norberg, Smithsonian Institution.

n.d.: Element not determined by this method.

1. 31846 olivine clinopyroxenite
2. 37187 poikilitic peridotite
3. 37095 feldspathic olivine clinopyroxenite
4. 37054 troctolite
5. 37048 clinopyroxene gabro
6. 37055 clinopyroxene-brown hornblende gabro with abundant cumulus Fe-Ti oxide
7. 37064 pegmatitic hornblende metagabbro
8. 37057 hornblende metagabbro
tion effects preclude complete quantitative definition of these differences, several generalizations concerning igneous petrogenesis can be inferred from the listed analyses.

*The plutonic suite.*—Analyses 1 to 6, table 5, are of cumulate samples from the Livingstone Subgroup, and their MgO, CaO, Al₂O₃, and TiO₂ values generally reflect the amount of olivine, clinopyroxene, plagioclase, and/or oxide in the original rocks. Cumulate samples from the ultramafic zones (analyses 1-4, table 5) have constant high Mg/(Mg+Fe) values, consistent with their formation as the early crystallization products of the parent magma(s). Analyses 7 and 8 (table 5) are of non-cumulate gabbros and are generally similar to low TiO₂, K₂O, and P₂O₅ basaltic liquid compositions.

*Dikes.*—Analyses of dike samples from the Red Mountain area indicate rather variable SiO₂, CaO, and MgO contents; much of this

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<td>0.12</td>
<td>0.20</td>
<td>0.12</td>
<td>0.09</td>
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<td>0.31</td>
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Total      | 100.8 | 100.3 | 99.5 | 100.5 | 100.6 | 100.3 | 99.0 | 99.9 | 99.9 |

Cr (ppm)⁴     | 216 | 200 | 305 | 220 | 20 | 1 | 1313 | 6 |
Ni (ppm)      | 85 | 84 | 149 | 101 | 25 | 4 | 559 | 41 |

Atomic Mg Mg+Fe⁵ | 0.585 | 0.577 | 0.650 | 0.574 | 0.442 | 0.415 | 0.176 | 0.797 | 0.399 |

*Cr and Ni colorimetric analyses by J. Norberg, Smithsonian Institution.

1. 37032 porphyritic basalt dike intruded into Red Mountain Ultramafites.
2. 37070 porphyritic dolerite dike intruded into Red Mountain Ultramafites.
3. 37035 altered fine-grained maﬁc dike, Livingstone dike complex.
4. 37040 altered feldspar–phryic maﬁc dike, Livingstone dike complex.
5. 37269 medium-grained greenschist dike, Livingstone dike complex.
6. 37189 fine-grained greenschist dike, Livingstone dike complex.
7. 37085 albite dike, Livingstone dike complex.
8. 37060 altered brown hornblende–phryic dike, intruded into pillow metavolcanics.
9. 37063 altered ferrobasalt dike, intruded into volcanioclastic metasediments.
variation may be related to alteration processes. However, the gray-altering mafic dikes (analyses 1 to 4, table 6) are chemically distinct from the greenstone dikes (analyses 5 and 6, table 6), even though these samples indicate identical metamorphic grades and are juxtaposed along sharp contacts. Unless alteration processes have operated in a highly selective manner, these observations support the conclusion that these compositional differences are original, that is, reflect differences in the original igneous rocks. The gray mafic dikes have low SiO$_2$ and relatively high MgO, Cr, and Ni contents compared to the greenstones which have intermediate silica and low MgO, Cr, and Ni contents. The albitite dike (analysis 7, table 6) has high SiO$_2$ and Al$_2$O$_3$, and extremely high Na$_2$O. Although this composition almost certainly reflects extensive soda metasomatism, albitite contacts with surrounding rocks are very sharp (fig. 7), and these rocks probably originally had fairly acidic compositions. The hornblende–phyric dike (37060, analysis 8, table 6) has an unusual composition including high MgO and K$_2$O. Although K$_2$O is typically considered to be fairly mobile during alteration processes, this rock is petrographically identical to another hornblende–phyric, mica-rich dike from another part of the area; both dikes have sharp contacts with country rocks, and at least some of the high K$_2$O should probably be considered to be primary. This sample also has very high Cr and Ni contents (table 6). With the exception of sample 37060, all the dikes have very low K$_2$O and P$_2$O$_5$ contents; TiO$_2$ is quite variable.

Metavolcanics.—A massive greenstone from the base of the metavolcanic sequence has a composition (analysis 1, table 7) similar to greenstone dikes in the dike complex. Except for high Na$_2$O and H$_2$O, all the other analyzed metavolcanic samples have basaltic compositions with low K$_2$O, TiO$_2$, and P$_2$O$_5$. The high Na$_2$O and H$_2$O values probably reflect alteration processes. Note that there do not appear to be significant compositional differences between the upper (pillow) metavolcanics and the lower (non-pillow) metavolcanics. Besides having higher SiO$_2$ and lower MgO than the other metavolcanics, the basal greenstone also has significantly lower Cr and Ni contents; P$_2$O$_5$ and TiO$_2$ contents are low in all the analyzed metavolcanic samples.

Discussion.—On the basis of major element compositions, the analyzed dikes and metavolcanics can be separated into two compositional groups. Most of the metavolcanics and the gray dikes have basaltic silica and MgO contents broadly similar to group I ocean-floor basalts (Bryan and others, 1976). Many of the other samples comprise a second, more heterogeneous, group of more differentiated compositions. However, since major elements are subject to mobilization during alteration, it is probably only worth noting that these compositional groupings broadly confirm differences based on field evidence. Gray, metabasaltic dikes tend to be the older dikes in the area.

The separation of Livingstone Subgroup dikes and metavolcanics into two compositional groups is supported by variations in the less
mobile elements Cr, Ni, and Ti (fig. 9). The basaltic samples have high Ni and Cr contents relative to Ti, whereas the other samples are strongly depleted in these compatible elements. Although possible alteration effects preclude quantitative analysis of fractionation trends, it is clear from figures 9, 10, and 11 that the two groups are not likely to be related by any simple fractionation mechanism. Hence, these two groups are here treated as having quite separate petrogenetic histories.

Pearce (1975) has shown that ocean-floor basalts can be distinguished from low K arc tholeiites in terms of their Ti–Cr distributions. The Livingstone Subgroup basaltic samples appear to have ocean-floor affinities (fig. 10), whereas the other Livingstone samples have Ti and Cr contents more like primitive arc volcanics. On the basis of these compositional characteristics it can be concluded that most of the

**Table 7**

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<tr>
<td>Ni (ppm)*</td>
<td>17</td>
<td>227</td>
<td>81</td>
</tr>
</tbody>
</table>

**Atomic**

$$\frac{\text{Mg}}{\text{Mg+Fe}}_{\text{total}} = 0.484 \quad 0.659 \quad 0.623 \quad 0.588 \quad 0.579$$

*Cr and Ni colorimetric analyses by J. Norberg, Smithsonian Institution.

1. 37068 basal metavolcanic gneissone
2. 37059 non-pillowed metavolcanic
3. 37053 pillow metavolcanic
4. 37051 pillow metavolcanic
5. 37006 pillow metavolcanic
analyzed metavolcanic samples and the earlier dikes of the Livingstone Subgroup at Red Mountain have compositions indicating the original samples were similar to basalts from modern mid-ocean spreading centers. This conclusion is generally supported by rare earth and other trace element data to be presented elsewhere (Sinton and Rankin, unpub. data; see also Sinton, ms).

Younger dikes and at least one greenstone from the lower metavolcanic sequence have major and minor element compositions different from those of modern spreading center lavas. They are, however, similar in Ni and MgO contents to the dominantly siliceous rocks from the more southerly Lintley-Oтаномомо sector of the Dun Mountain Belt and from Eua in the Tonga arc (fig. 11; Coombs and others, 1976). Although the pre-Upper Eocene Eua volcanic succession is apparently ophiolitic (Ewart and Bryan, 1972, 1973), the tectonic regime in which these rocks formed is not clear. The present structural settings of Eua and of the igneous rocks of the Lintley-Oтаномомо sector of the Dun

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**Fig. 9.** Livingstone Subgroup dikes and metavolcanic rock compositions plotted in terms of their Cr–Ti–Ni ratios. Symbols refer to dikes (circles), metavolcanic rocks (squares), high K-O dike (37060—upright triangle) and ferrobasaltic dike (37063—inverted triangle). Solid symbols are high compatible element masic rocks; open symbols are more highly differentiated rocks.

**Fig. 10.** Anhydrous Ti versus Cr contents in metavolcanic and dike rocks of the Red Mountain ophiolite complex. Symbols as in figure 9.
Mountain Belt are consistent with formation either in an arc-trench gap or perhaps within a primitive volcanic arc. Evidence supporting the presence of igneous activity in the accretionary prism of several Pacific arc-trench systems has been presented by Marshak and Karig (1977), and such a setting may be appropriate for the younger igneous activity at Red Mountain.

RELATIVE CHRONOLOGY OF IGNEOUS ACTIVITY IN THE OPHIOLITE COMPLEX

Igneous activity giving rise to the Red Mountain ophiolite complex can be separated into three main stages on the basis of several criteria. The earliest stage (stage I) involves derivation of the harzburgitic Red Mountain Ultramafites as a residue of 20 to 30 percent partial melting in the upper mantle (Sinton, 1977). This basal peridotite underwent a complex pre-emplacement history but can be related to the overlying cumulate sequence by the presence of a transitional peridotite near the harzburgite-basal Livingstone contact. As emphasized earlier (Sinton, 1977), the olivine-clinopyroxene dominated transitional peridotite formed by recrystallization, accompanying shallow level re-equilibration of the ultramafic residuum with olivine basaltic melt similar to the parent magma for the lower Livingstone cumulate sequence. This interpretation implies that the residual and magmatic parts of the ophiolite at Red Mountain apparently represent complementary parts of the same partial melting event. Stage I therefore includes formation of the Livingstone gabbro-peridotite sequence and the accompanying re-equilibration of the upper portions of the underlying residual peridotite. Red Mountain Ultramafites and transitional peridotite formed the floor to the cumulates. This activity was accompanied by significant deformation as

Fig. 11. Anhydrous MgO versus Ni contents in metavolcanic and dike rocks of the Red Mountain ophiolite complex. Fields for Group I ocean floor basalts (OF3), selected North Island, New Zealand and Japan andesites (CA) and for data for Eua, Tongan Islands and the Lintley-Otanomomo segment of the Dun Mountain ophiolite belt (E, L-O) are after Coombs and others (1976). Symbols as in figure 9.
seen in the hyper-solidus deformation of the cumulates. The roof to
the stage I magmatism is equivocal, but the upper gabbros probably
gave way to erupted volcanics near the surface.

In contrast to the dominantly plutonic stage I magmatism, stage II
activity is mainly volcanic. Although stage II activity may have partially
overlapped that of stage I in time, evidence for separation of much of
the dike intrusion and volcanism into a stage separate from the formation
of the cumulate sequence consists mainly of (1) chilled contacts of dikes
against country rock residual peridotite or Livingstone Subgroup
gabbros and peridotites, and (2) compositional coherence of these dikes
with much of the metavolcanic sequence. Although the absolute timing
of stage II igneous activity is largely unknown, the chilled margins of
the dikes may indicate periods of cooling interrupting periods of intru-
sion. Alternatively the fine-grained margins may have formed by rapid
decrease in \( P_{H_2O} \) as fractures penetrated to shallow levels (C. A. Hopson,
personal commun., 1978). Such “pressure-quench” margins do not require
prolonged periods of cooling. Nonetheless, at least one (sub)stage of
quiescence is indicated by the 30 to 60 m thick, fine-grained interbedded
metasedimentary unit that separates the lower and upper metavolcanic
units that are both here considered to represent stage II activity. Dikes
intruded into Red Mountain Ultramafiles have alteration-resistant
element contents identical to mafic dikes and metavolcanics in the
Livingstone Subgroup (figs. 9, 10, 11), and, for this reason, these dikes
are considered to have acted as feeders to the overlying Livingstone
dikes and flows.

Stage III igneous activity is dominated by more siliceous magmatism
which, on field evidence, is predominantly younger than the stage II
mafic activity. The various dike relations are important evidence for
interpreting the evolution and tectonic setting of the ophiolite. The
common presence of chilled stage II dikes in serpentinized zones in
residual peridotite is corroborating evidence that some of these dikes
may have been emplaced after a period of seepentinization (see above).
This suggests that significant magmatic activity continued after the
harzburgites had moved both upward and outward, away from the
zone of partial melting and intense magmatism. Similarly the predomi-
nantly finer grained character of stage III dikes, compared to the older
stage II dikes, probably indicates that this latter activity was mainly
concentrated in a setting of lower regional temperatures. However, it
must be emphasized that there is no discernible discrete break between
the various stages, and production of the liquids of the two composi-
tional groups defined above probably, at least in part, overlapped in
time. The definition of the three evolutionary stages is important for
understanding how the ophiolite evolved but does not necessarily
require temporally isolated magmatic events. Indeed, the three stages
described above may simply represent different parts of a single
evolving magma system.
TECTONIC SETTING OF THE RED MOUNTAIN OPHIOLITE COMPLEX

The following considerations are pertinent to any tectonic reconstruction of the Red Mountain area, and proposed models of origin must satisfy these observations.

1. There is abundant evidence for formation in an area where magmatism and tectonism were apparently closely related. Hypersolidus deformation of the cumulates and the general lack of dike preferred orientations indicate that deformation continued throughout much of the magmatic history of the ophiolite.

2. Field evidence and crystallization sequences of the cumulates imply relatively shallow depths (pressures mainly less than 2-3 kb) for the initial magmatic differentiation. Early amphibolite metamorphism (Sinton, ms; Coombs and others, 1976) also indicates low prevailing pressures.

3. The amount of dike material locally approaches 100 percent by volume, and the dike complex must have formed in a regime generally characterized by tensional tectonics.

4. The upper (pillow) metavolcanic sequence was subaqueous and probably marine. The marine fossil *Atomodesma* has been identified in sediments interbedded with Livingstone Subgroup lavas near Mossburn (fig. 1) to the south of Red Mountain (G. Hyden, written commun., 1978) and with correlatives of the Livingstone Subgroup in Nelson Province (Walcott, 1969).

5. Abyssal sediments are not present, and spatially associated metasediments are mainly well-bedded volcaniclastic breccias. These sediments indicate a proximal volcanic source, but the probable presence of an unconformity above the igneous members of the ophiolite relieves the possible constraint imposed by the nature of the overlying sediments. Metasediments interbedded with the metavolcanics are mainly fine-grained volcaniclastic mudstones.

6. The contrasting chemical compositional groups present in the complex require special attention. The parent liquids to the cumulates and the mafic dikes and lavas were olivine tholeiitic similar to modern ocean-ridge magmas. Stage I magmas that produced the cumulates were also hydrous, but it is not clear whether the water was primary magmatic or sea water introduced during crystallization. Stage III magmas were not typical of modern spreading ridges and probably require much of the later activity to be spatially proximal to an active volcanic arc.

Individually the above considerations are not necessarily uniquely indicative of a specific tectonic environment for formation of the Red Mountain ophiolite complex. However, taken together these observations and interpretations make a strong case for generation and early evolution in a spreading environment. Modern spreading ridges and marginal basins are equally consistent with such an environment. The non-sheeted nature of the dike complex may indicate that the ophiolite formed in
an environment where spreading was geometrically irregular. The apparently close temporal association with highly differentiated magma, and the nature of the overlying sediments, may indicate that the spreading environment was close to an active volcanic arc. Taken together, the above evidence favors a marginal basin environment. However, near-arc spreading on the fore-arc side cannot be precluded. The latest subduction of the Rangitata Orogeny was dipping westward, and Red Mountain is located on the upper plate above the suggested subduction zone to the east (Coombs and others, 1976). However, pre-Cretaceous tectonic history of this region is equivocal and absolute definition of the tectonic setting during formation of the ophiolite complex awaits further regional studies.

ACKNOWLEDGMENTS

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Petrology and evolution of the Red Mountain ophiolite complex


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