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BRUCE FRANK HOUGHTON

GEOLOGY OF THE TAKITimu GROUP AND ASSOCIATED INTRUSIVE ROCKS, CENTRAL TAKITimu MOUNTAINS, WESTERN SOUTHLAND NEW ZEALAND.

Volume Two
6.1 Introduction

The MacKinnon Peak intrusive suite is defined as the swarm of porphyritic hornblende- and anorthite- and salitic pyroxene-bearing dykes of basaltic composition intruding the Takitimu Group in the Takitimu Mountains and Wairaki Hills. The dykes are petrographically distinct from the volcanic rocks of the Takitimu Group, and from the White Hill Intrusives. The dykes contain large (up to 6 cm) phenocrysts of tschermakitic hornblende, anorthite and rarely salitic clinopyroxene in a cryptocrystalline groundmass (Fig. 6:1). Many of the dykes are charged with hornblende-rich cumulate xenoliths (Fig. 6:2). Flow differentiation is common, with phenocrysts and xenoliths concentrated in the central portions of the dykes. The dykes range in thickness from 2 cm to 2 m, dip at shallow to moderate angles, and crosscut the steeply dipping strata of the Takitimu Group and sills of the White Hill Intrusives. Four hornblende concentrates from individual dykes have been dated at 231-242 m.y. (see section 6.7).

6.2 Distribution

Forty-two dykes have been located in the MacLean and MacKinnon Peaks in the headwaters of the Wairaki River (shown on the accompanying map, back pocket), and in the vicinity of the lower Wairaki River on Mt Linton and Beaumont Stations. These two areas are separated by a large area covered by the Takitimu Forest in which no exposure is known. Two predominant trends are present in the dyke swarm:

1. striking 345°-010° paralleling the strike of the host strata but dipping comparatively gently (50°-20°W), and
2. striking 060°-110° and dipping moderately to steeply.
Fig 6.1  Hornblende–anorthite–salite basalt, MacKinnon Peak intrusive suite, OU 38938. Photograph K.D. Mason.

Fig 6.2  Cumulate xenoliths in MacKinnon Peak basalt. Upper xenolith fine-grained hornblende–anorthite adcumulate (type 4, see section 6.6), lower xenolith coarse-grained hornblende cumulate (type 3), OU 38981. Photograph K.D. Mason.
6.3 Field Geology

Several members of the dyke swarm are strongly flow-differentiated, others maintain a uniform concentration of phenocrysts away from narrow (0.5-1.0 cm) chilled margins. The dykes are invariably chilled at the contact with country rock, and may possess a reddened margin. In the chilled margin phenocrysts are either absent or restricted to a few broken or fractured crystals. In the flow-differentiated members strong flow banding is present. The flow banding is pronounced in the outer 2-3 cm of the dykes, and consists of alternating white and black laminations within a grey groundmass (Fig. 6:3). In thin section the white and black laminations are seen to consist of trains of granulated and sheared phenocrysts of anorthite and hornblende. The banding is further emphasised by alignment of plagioclase microlites in the grey microcrystalline groundmass. Where the country rock consists of a massive lava flow or a crystalline intrusive rock the dykes maintain parallel walls. Dykes intruding fine-grained sediments are more irregular, with xenoliths of sediment commonly present in the dykes, and tongues of dyke material extending into the host sediment (e.g. S159/917797). At only two localities the termination of MacKinnon Peak dykes have been mapped. In particular in the McLean Peaks (S159/913825) a thin dyke branches into two 4 cm wide blunt-edged tongues diverging at 50° (Fig. 6:4).

6.4 Flow Differentiation

A complete transition is seen among the dykes from members showing an even distribution of phenocrysts to strongly flow-differentiated members. In general flow differentiation is most marked in the thinner dykes, having been observed in members only 4 cm wide. The concentration of phenocrysts in differentiated dykes increases rapidly inward from the margins reaching 65-70% in the
Fig 6.3 Flow-banded margin to MacKinnon Peak basalt, OU 38999C. Photograph K.D. Mason.
Fig 6.4 Branching termination of MacKinnon Peak dyke at S159/913825, MacLean Peaks, viewed southward bedding plane vertical and at right angles to plane of photograph.
cores of some dykes.

6.5 Petrography

The MacKinnon Peak dykes have the composition of high-alumina basalt, but their mineralogy is atypical of basalts in general. The dykes consist of large (2-60 mm) phenocrysts of tschermakitic hornblende, anorthite-rich plagioclase and rarely salitic pyroxene, together with microphenocrysts of anorthite, magnetite and rarely hornblende in a siliceous, cryptocrystalline to fine-grained groundmass (Fig. 6:2 to 6:9). Plagioclase microlites and granules of magnetite are conspicuous in the groundmass of the coarser-grained specimens. The dykes contain cumulus-textured xenoliths of comparable mineralogy to the host dykes (Fig. 6:2, 6:7 to 6:9). Table 6:1 lists the modal composition of representative specimens. In general phenocrysts and microphenocrysts form between 30% and 40% of dykes lacking conspicuous flow differentiation. In most flow-differentiated members axial regions of the dykes contain a high concentration of phenocrysts representing 50%-55% of the rock by volume. The aphyric margins to these dykes contain 5%-10% phenocrysts by volume. The dykes have mineralogy and colour indices more typical of andesites than basalts, but their basaltic chemistry is reflected in the high anorthite content of the plagioclase and the subsilicic nature of the amphiboles.

6.5.1 Hornblende

Hornblende phenocrysts are present in every specimen collected. These generally form a single size population, less commonly there is a complete size gradation from 1 cm to 0.1 mm (e.g. OU 39050, Table 6:1). The crystals are euhedral and commonly possess narrow optically-distinct rims (Fig. 6:5).
Table 6:1 Modal analyses of MacKinnon Peak basalts based on counting of 500 points/specimen

<table>
<thead>
<tr>
<th>Specimen</th>
<th>massive dykes</th>
<th>flow differentiated dykes</th>
</tr>
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<tbody>
<tr>
<td></td>
<td>38938</td>
<td>39050</td>
</tr>
<tr>
<td></td>
<td>interior</td>
<td>margin</td>
</tr>
<tr>
<td>hornblende phenocrysts</td>
<td>20.4</td>
<td>19.2</td>
</tr>
<tr>
<td>plagioclase &quot; &quot;</td>
<td>18.2</td>
<td>11.0</td>
</tr>
<tr>
<td>clinopyroxene &quot; &quot;</td>
<td>0.4</td>
<td>-</td>
</tr>
<tr>
<td>hornblende microphenocrysts</td>
<td>2.0</td>
<td>10.4</td>
</tr>
<tr>
<td>plagioclase &quot; &quot;</td>
<td>9.8</td>
<td>5.4</td>
</tr>
<tr>
<td>magnetite &quot; &quot;</td>
<td>2.0</td>
<td>1.4</td>
</tr>
<tr>
<td>clinopyroxene &quot; &quot;</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>groundmass</td>
<td>58.2</td>
<td>52.6</td>
</tr>
</tbody>
</table>

* contained 3.2% vesicles.
Fig 6.5 Zoned hornblende phenocryst, MacKinnon Peak basalt. Pale green relatively magnesium-rich core surrounded by deeper green, iron-rich rim. Partially albitised anorthite phenocryst to left of hornblende, OU 38981, plane polarised light.
Fig 6.6 Slabbed specimen of hornblende-salite adcumulate xenolith (type 1) with minor interstitial zeolite within MacKinnon Peak intrusive dyke. Host hornblende-anorthite basalt, OU 38981.

Fig 6.7 Hornblende-salite adcumulate xenolith(type 1), crossed Nicol prisms, OU 38981, MacKinnon Peak Intrusives.
The hornblende, with one exception is pleochroic, in shades of green and tan with absorption tints:

- tan-straw, 
- olive-green, and 
- deep-green.

In OU 38565 a hornblende-rich dyke, charged with anorthite-hornblende(-magnetite) inclusions, the phenocrysts consist of brown hornblende cores mantled by green hornblende. The former has an absorption scheme of

- tan-straw, 
- green-brown, and 
- green-brown.

Hornblende phenocrysts are rarely fringed by a discontinuous band of actinolite prisms. (These are considered to be deuteric in origin and are discussed in Chapter 7).

6.5.2 Plagioclase

Large plagioclase phenocrysts (An\textsubscript{78}-An\textsubscript{96}) are present throughout the dykes except for the margins of the most flow-differentiated dykes (OU 38981, OU 38999). The crystals are subhedral and show signs of resorption. The margins of the plagioclase crystals are altered to albite along cleavage planes and fractures. The crystals show predominantly simple, rather than polysynthetic, twinning and weak normal compositional zoning with a broad core either -

[1] of constant composition An\textsubscript{94-96}, or

[2] zoning gradationally from An\textsubscript{92-96} outwards to An\textsubscript{85} and occasionally narrow optically-discontinuous rims of more sodic composition. A complete size gradation from 1-2 mm to 0.4 or 0.2 mm is common in dykes. Groundmass feldspar compositions are similar to the compositions of the phenocrysts.

6.5.3 Clinopyroxene

Large (0.4 mm to 4 mm) salite-rich pyroxene phenocrysts occur in several dykes. The phenocrysts commonly show strong
optical zoning. Two forms are seen—

[1] numerous narrow zones, e.g. OU 39050, OU 38981
[2] diopside cores surrounded by wide salite rims,
   e.g. OU 38938.

6.5.4 Magnetite

Magnetite is an ubiquitous groundmass constituent of the MacKinnon Peak dykes and also commonly occurs as microphenocrysts (0.1-0.5 mm). Crystals are well formed and euhedral. Euhedral magnetite phenocrysts occur in OU 39050.

6.6. Xenoliths

The MacKinnon Peak dykes contain abundant coarse-grained xenoliths with mineralogy comparable to the host dykes. The xenoliths commonly show well-developed cumulus textures and are named in accordance with the terminology of Wager et al. (1960), rather than the conventional terminology of plutonic rocks. A variety of cumulus textures occur in the xenoliths with a full gradation from adcumulates to xenoliths possessing abundant intercumulus material. The intercumulus material is often similar in mineralogy and texture to the groundmass of the host dykes. Hornblende crystals in intercumulus xenoliths have perfect crystal outlines, and show marked zoning, features not observed in hornblendes within adcumulus xenoliths. Four types of xenoliths have been observed in the field and in thin section:

[1] coarse-grained (4-10 mm) hornblende ± salite and hornblende-salite-olivine ± magnetite adcumulates (Figs. 6:6, 6:7, 6:8);

[2] coarse-grained (2-5 mm) hornblende-anorthite ± magnetite (Figs. 6:4, 6:12) cumulates with small amounts of intercumulus material,

[3] coarse-grained (2-10 mm) hornblende cumulates with
Fig 6.8 Pseudomorphs of chlorite(?) - magnetite replacing olivine(?) in hornblende-salite-?olivine adcumulus xenolith, OU 38981C, MacKinnon Peak intrusive suite, plane polarised light.

Fig 6.9 Diverging trains of magnetite grains and exsolution on 010 of a fine-grained unknown phase within hornblende grain, OU 38981D, hornblende adcumulus xenolith, MacKinnon Peak intrusive suite, plane polarised light.
abundant intercumulus plagioclase-magnetite + stilbite + chabazite (Figs. 6:2, 6:11),

fine-grained (0.1-0.6 mm) hornblende-anorthite + salite adcumulates (Figs. 6:2, 6:13).

Each xenolith contains essentially a single size population of cumulus crystals, and there is a distinct contrast in size between type four xenoliths and xenoliths of types one, two and three.

6.6.1 Coarse-grained hornblende-salite adcumulates (type 1)

These xenoliths consist principally of moderately well-sorted crystals of salite and hornblende (Figs. 6:6, 6:7). The crystals are subhedral and grains of salite are frequently enclosed poikolitically by single hornblende crystals. Crystals of magnetite and pseudomorphs of magnetite-chlorite + actinolite, the latter probably after olivine, are present in some specimens, (Fig. 6:8). Intercumulus pore material is lacking. Hornblende predominates over salite in every xenolith examined and several are ultramafic, consisting only of hornblende and 1%-2% magnetite. Both hornblende and salite lack conspicuous compositional or optical zonation. A hornblende adcumulate xenolith in OU 38981 contains hornblende and salite in coarse-grained intergrowths in which rods of hornblende are aligned parallel to the c-axis of the host salite (Fig. 6:10). In another hornblende cumulate inclusion from this dyke, the cores of hornblende crystals contain delicate diverging trains of magnetite grains and schiller-like exsolution of exceptionally fine-grained lamellae, of an unknown phase, on 010 (Fig. 6:9). The magnetite trains are somewhat arcuate and occur both parallel to the a-axis and approximately at right angles to it. The inclusions and lamellae are absent in the margins of the crystals.

6.6.2 Coarse-grained hornblende-plagioclase cumulates (type 2)

Coarse-grained plagioclase-rich cumulate xenoliths occur in
Fig 6.10 Intergrowth of salitic pyroxene and hornblende in hornblende-salite adcumulate xenolith, OU 38981D, MacKinnon Peak intrusive suite. Rods of hornblende are aligned parallel to q axis of pyroxene, plane polarised light.

Fig 6.11 Hornblende cumulus xenolith, MacKinnon Peak intrusive suite, containing abundant intercumulus material, predominantly plagioclase and stilbite, OU 38981B. Photograph K.D.Mason.
two dykes, north of Spence Peak (S159/892848) and in the Wairaki Gorge (S159/910689). These contain the assemblage plagioclase-hornblende-magnetite ± salite (Fig. 6:12). Both cumulus and intercumulus phases and textures are present in any one xenolith, with large crystals of hornblende, plagioclase and occasionally rare salite and smaller magnetite in contact with each other, and with a fine-grained matrix. The matrix consists of elongate prisms of hornblende and subhedral magnetite granules set in plagioclase and secondary stilbite. Coarse-grained prisms of apatite are a common accessory phase. The large plagioclase phenocrysts show polysynthetic twinning and are occasionally mantled by untwinned rims of more sodic composition. Twinning is also common among the hornblende crystals.

6.6.3 Coarse-grained hornblende cumulates (type 3)

The hornblende cumulus xenoliths consist of a framework of large (up to 2 cm) hornblende crystals with interstitial plagioclase, magnetite and stilbite, (Fig. 6:11). The hornblende crystals differ in two respects from crystals in the adcumulates, in that they are euhedral and they commonly show marked zoning. Two forms of zoning are present:

[1] several broad zones of alternating Fe and Ca-Mg rich composition, each zone being approximately constant in composition across its width (Fig. 6:19);

[2] numerous thin zones showing striking oscillatory zoning with overall trends to iron-rich compositions at the rims (Fig. 6:20). The chemical nature of the zoning is discussed in section 6.10.

Crystals showing both types of zoning may be present in a single xenolith.
Fig 6.12 Polished thin-section of coarse-grained hornblende-anorthite-magnetite cumulus xenolith (type 3). Note absence of adcumulus texture, OU 39143, MacKinnon Peak intrusive suite. Photograph K.D. Mason.

Fig 6.13 Fine-grained hornblende-anorthite-(magnetite) adcumulus xenolith (type 4), MacKinnon Peak intrusive suite, OU 38981.
The intergranular space in these xenoliths is heavily zeolitised, but also contains small hornblende grains and equigranular fine-grained to microcrystalline aggregates of plagioclase similar to the matrix in the host rock. Where hornblende phenocrysts are in contact with intergranular material, elongate prisms of near-colourless actinolite extend from the crystal up to 0.5 mm into the groundmass. On some phenocrysts actinolite occurs as a continuous rim of constant thickness around one or more faces of the crystal.

Fine-grained hornblende-plagioclase adcumulates (Type 4)

These xenoliths consist of fine-grained aggregates of hornblende, plagioclase, rare K-feldspar and spinel. Hornblende in general predominates over plagioclase, but the distribution is highly variable and areas up to 2 mm in diameter consist entirely of plagioclase grains. Magnetite and K-feldspar are minor constituents (1-5%). The hornblende grains are in general larger (0.3-0.6 mm) than the associated plagioclase (0.1-0.3 mm). Larger hornblende crystals (<2 mm) occur infrequently, often sieved by inclusions of spinel and sphene. One xenolith (OU 38937) contains an ovoid area approximately 4 mm in diameter consisting of small 0.04-0.08 mm) subhedral grains of clinopyroxene and plagioclase which are set in coarser grained (0.2-0.4 mm) hornblende and plagioclase. Large plagioclase crystals (up to 1 mm in diameter) are seen rarely in the inclusions.

Noncumulate xenoliths

A single noncumulate xenolith was found in OU 39050. It consists of banded microcrystalline quartz and feldspar and probably represents a fragment of flow-banded acidic volcanic rock. The fragment is mantled with a reaction corona of hornblende.
6.7 Geochronology

K-Ar determinations were made of four samples from the MacKinnon Peak Intrusives. Hornblende and whole rock pairs were dated and the results are given below in Table 6:2.

Table 6:2  K-Ar age determinations for whole rock samples and hornblende concentrates from the MacKinnon dykes.

<table>
<thead>
<tr>
<th>Sample</th>
<th>Wt %</th>
<th>Ar40 rad nl/gm</th>
<th>Ar40 rad % total</th>
<th>Age</th>
</tr>
</thead>
<tbody>
<tr>
<td>38999A hornblende</td>
<td>0.412</td>
<td>4.028</td>
<td>81</td>
<td>231 + 2</td>
</tr>
<tr>
<td>38935A</td>
<td>0.345</td>
<td>3.415</td>
<td>80</td>
<td>233 + 2</td>
</tr>
<tr>
<td>38606A</td>
<td>0.226</td>
<td>2.328</td>
<td>77</td>
<td>242 + 2</td>
</tr>
<tr>
<td>38938A</td>
<td>0.520</td>
<td>5.336</td>
<td>78</td>
<td>242 + 2</td>
</tr>
<tr>
<td>38999A total rock</td>
<td>1.350</td>
<td>11.480</td>
<td>84</td>
<td>202 + 2</td>
</tr>
<tr>
<td>38935A</td>
<td>0.709</td>
<td>6.335</td>
<td>94</td>
<td>212 + 2</td>
</tr>
<tr>
<td>38606A</td>
<td>0.452</td>
<td>3.944</td>
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<td>207 + 2</td>
</tr>
<tr>
<td>38938A</td>
<td>0.623</td>
<td>5.582</td>
<td>78</td>
<td>213 + 2</td>
</tr>
</tbody>
</table>

All specimens are hornblende-anorthite basalts

The data have been plotted on a K/Ar\(^{40}\) diagram in Fig. 6:14. A line connecting the hornblende ages passes near to the origin suggesting the age represents a close approximation to a true age of emplacement for the dykes (Harper, 1970). The hornblende ages cluster on the Middle/Upper Permian boundary (Van Eysinga, 1972). The whole-rock data plot along a similar trend to the hornblende data, but displacement towards lower Ar contents strongly suggests a uniform loss of argon has occurred. A possible explanation for such trends is that 'argon diffuses from the mineral lattices according to conventional diffusion theory but remains at grain boundaries. During subsequent uplift and cooling the rocks...
Fig 6.14 Plot of potassium abundance against radiogenic argon$^{40}$ abundance for whole rock samples and associated hornblende concentrates, MacKinnon Peak Intrusives.

6.14

$^{40}\text{Ar}_{\text{RAD}}$

nl/g

hornblende

total rock

K wt %
"relax", and therefore release uniform quantities of argon' (C.J.D. Adams, written comm., 1976). If this hypothesis is correct it appears that argon diffusion is minimal from the MacKinnon Peak hornblende, but substantial diffusion occurred from other phases in the rock, plagioclase and the fine-grained groundmass. The apparent age from whole rock data is therefore likely to be a mixed age resulting from argon retention on uplift, in one phase, (hornblende) and argon loss in others.

MINERALOGY

6.8 Feldspar

6.8.1 Plagioclase

In general plagioclase phenocrysts show weak normal zoning over an interval of 5%-10% anorthite, although some rocks contain phenocrysts with a narrow optically-and compositionally-distinct (Na-rich) rim. The total range of composition for rimless phenocrysts and the cores of rimmed crystals is An_{85}-An_{97}. The rims have a restricted composition in the range An_{70} to An_{82}. Incipient albitization is present along cracks and cleavages. Groundmass and microphenocrystal plagioclase overlap in composition with the more sodic (outer) portions of the phenocrysts. The composition of plagioclase in hornblende-plagioclase xenoliths is similar to that of phenocrysts in the host rock. The analysed feldspars are plotted in Fig. 6:15 for individual rocks.

6.8.2 Alkali feldspar

One of the fine-grained xenoliths contains two feldspars, a calcic bytownite and an alkali feldspar close in composition to endmember potassium feldspar (Or_{97.3}Ab_{2.0}An_{0.8}-Or_{96.6}Ab_{2.5}An_{0.9}).
Fig 6.15 Partial analyses of plagioclase feldspars, MacKinnon Peak Intrusives closed circles cores, open circles rims and crosses groundmass.

To obtain Q.U. numbers add 38000 to numbers given.
The grains form less than 1% of the rock and proved too fine-grained for resolution of the potassium feldspar species.

6.8.3 Minor elements

Potassium is a minor element in all plagioclase feldspars analysed, and the crystals closely approximate simple albite-anorthite solid solution. More significant is the substitution of iron, reported here as $\text{Fe}_2\text{O}_3$, presumably for aluminium. Substitution remains at a relatively consistent level of 0.40–0.60 weight percent $\text{Fe}_2\text{O}_3$ without any significant variation with change in major element chemistry. The K-feldspar grains in OU 38981A contain appreciably less $\text{Fe}_2\text{O}_3$ than the co-existing plagioclase.

6.9 Clinopyroxenes

Clinopyroxene is rare in the MacKinnon Peak suite, occurring occasionally as phenocrysts but not in the groundmass, but clinopyroxene is an abundant constituent of the adcumulus xenoliths. The phenocrysts in two dykes were analysed using the electron microprobe (Table 6:3). In OU 38565 the phenocrysts are essentially unzoned, in OU 38938 the phenocrysts consist of two zones with a distinct break in chemical composition at the boundary. The rim is enriched in $\text{FeO}$ and particularly $\text{Al}_2\text{O}_3$ and depleted in $\text{MgO}$ and $\text{SiO}_2$ relative to the core. The pyroxenes in two hornblende-anorthite-salite xenoliths were also analysed.

6.9.1 Chemical variation

The analysed pyroxenes are salites or Ca-rich augites (Fig. 6:17, and Table 6.3) The principal substitution is iron for magnesium, with calcium approximately constant. In OU 38565 the xenolithic and phenocrystal pyroxenes have similar $\text{Fe}/\text{Mg}$ ratios, but the phenocrysts contain appreciably less calcium. The pyroxenes
Table 6.3 Analyses of clinopyroxenes from MacKinnon Peak basalts and xenoliths, determined by electron microprobe.

<table>
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<tr>
<th></th>
<th>38565</th>
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<th>38938</th>
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<th>39140</th>
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<tr>
<td></td>
<td>Phenocryst</td>
<td>Cognate</td>
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<td>Cognate</td>
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</tr>
<tr>
<td></td>
<td>Core</td>
<td>Rim</td>
<td>Xenolith</td>
<td>Core</td>
<td>Rim</td>
<td>Xenolith</td>
</tr>
<tr>
<td>SiO(_2)</td>
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<td>51.5</td>
<td>51.9</td>
<td>52.7</td>
<td>49.5</td>
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<tr>
<td>Al(_2)O(_3)</td>
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<td>2.48</td>
<td>1.69</td>
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<td>0.33</td>
<td>0.55</td>
<td>0.46</td>
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<tr>
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<td>0.33</td>
<td>0.10</td>
<td>0.12</td>
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<td>MgO</td>
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<tr>
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<td>0.01</td>
<td>0.01</td>
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</tr>
<tr>
<td>Cr(_2)O(_3)</td>
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<td>—</td>
<td>—</td>
<td>0.26</td>
<td>0.03</td>
<td>0.00</td>
</tr>
<tr>
<td>Total</td>
<td>99.9</td>
<td>99.1</td>
<td>99.6</td>
<td>99.4</td>
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<table>
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<th>Al(_{VI})</th>
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<th>Mg</th>
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OU 38565 Hornblende-anorthite-Ca-rich pyroxene basalt containing coarse-grained hornblende-anorthite ± Ca-rich pyroxene xenoliths

OU 38938 Hornblende-anorthite-Ca-rich pyroxene basalt.

OU 39140 Hornblende-anorthite xenolith (type 3).
Fig 6.16 Partial analyses of clinopyroxenes, MacKinnon Peak Intrusives, plotted on Ca-Mg-Fe partial triangle.

Open circles, specimen 565, are clinopyroxenes from a cognate xenolith.

To obtain Otago University numbers add 38000 to numbers present on diagram.

The compositional trends for pyroxenes within the Takitimu Group lavas are shown as solid lines.
Fig 6.17 Trend of pyroxene composition for the MacKinnon Peak rocks as revealed by a plot of all available partial analyses on a Ca-Mg-Fe partial triangle. The compositional trend for pyroxenes within the Takitimu lavas is shown as solid lines.
from the MacKinnon Peak basalts contain appreciably more calcium than the pyroxenes of the Takitimu Group lava. The trends shown in Fig.6:17 resemble the typical compositional trends of alkaline intrusions and lavas (e.g. Wilkinson, 1956, 1958; Aoki, 1964) rather than the normal two-pyroxene trends of calcalkaline and tholeiitic magmas (e.g. Carmichael, 1967a; Nicholls, 1971). Similar pyroxenes, however, have been reported from calcalkaline basalts (Lewis, 1973) and from laboratory experiments studying the crystallization of basaltic magmas under conditions of high temperature and water pressure. Other elements present in the MacKinnon Peak clinopyroxenes are aluminium, titanium, manganese and sodium. The MacKinnon Peak clinopyroxenes contain equivalent amounts of alumina to the phenocrysts in the Takitimu Group lavas, with the exception of the phenocrystal rims in OU 38938. Alumina is principally present as Ca-tschermaks molecule since titanium, and hence the titanopyroxene component is uniformly low in analysed pyroxenes. All the pyroxenes contain only trace amounts of manganese and sodium, in this they are comparable to pyroxenes in the Takitimu Group lavas and other calcalkaline rock series (e.g. Solomon Islands, Stanton and Bell, 1969; Santorini Volcano, Nicholls, 1971; Lesser Antilles, Arculus, 1976).

6.9.2 Conditions of crystallization of clinopyroxene

As described earlier (sections 6:5 and 6:6) salitic pyroxene is abundant in adcumulate xenoliths with hornblende, magnetite and (?) olivine, but rare as phenocrysts in the host dykes. In the latter environment the pyroxene is commonly mantled by hornblende. If the xenoliths are cognate, a point discussed in more detail later, salitic pyroxene must be an early-formed phase. It is, therefore, principally found where it has been removed from contact with the
melt either by adcumulate growth or by formation of a mantle of hornblende.

6.10 Hornblende

Phenocrysts of hornblende are present in all the MacKinnon Peak dykes and hornblende is a conspicuous cumulus phase in all cumulate xenoliths. Thirteen full and nine partial analyses are presented here. All are subsilicic and high in alumina and magnesia.

In the following discussion the formula \[ A_{0.01-0.3}Y_{0.63-1.72}Z_{0.01-0.03}^{(OH,W)} \] is adopted for amphiboles with W representing a halide ion (Deer et al., 1966). No published procedure for assigning cations to this formula has widespread acceptance. In this thesis a procedure is followed which is proposed in a draft report, but as yet not formally published by the International Mineralogical Association Subcommittee of the Amphibole Group. This is:

1. Balance cations to 24(O, OH, F, Cl) or 23 oxygens.
2. Sum Z to 8.0 using Si, then Al, then \( \text{Cr}^{3+}, \text{Fe}^{3+}, \text{Ti}^{4+} \).
3. Sum Y to 5.0 using excess Al, Cr, \( \text{Fe}^{3+}, \text{Ti} \) from (2) then Mg, then \( \text{Fe}^{2+} \) and then Mn.
4. Sum X to 2.0 using excess from (3), then Li, then Ca, then Na.
5. Group excess from (4) with K in A.

6.10.1 Classification

The amphibole group is subdivided by the I.M.A. subcommittee on the basis of the occupancy of the X site. The analyses in Tables 6:4 and 6:5 indicate the X site in the MacKinnon Peak hornblende consists of \( \text{Fe}^{2+}_{0.15-0.32} \text{Mn}^{2+}_{0.01-0.03} \text{Ca}^{2+}_{1.65-1.72} \text{Na}^{+}_{0.00-0.03} \), i.e. the phenocrysts and the xenolith amphiboles are part of the calcic amphibole grouping for which \( (\text{Ca}+\text{Na})_x > 1.34 \) and \( \text{Na}_x < 0.67 \). The calcic amphibole field is subdivided on the basis of:
Table 6.4 Analyses of hornblende, MacKinnon Peak basalts, in each instance analyses of core and rim zones are given for each crystal.

<table>
<thead>
<tr>
<th>Phenocrysts</th>
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<th>38565</th>
<th>39050</th>
<th>38562</th>
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<td></td>
<td>1</td>
<td>2</td>
<td>3</td>
<td>4</td>
</tr>
<tr>
<td></td>
<td>core</td>
<td>rim</td>
<td>core</td>
<td>rim</td>
</tr>
<tr>
<td>SiO&lt;sub&gt;2&lt;/sub&gt;</td>
<td>41.3</td>
<td>42.1</td>
<td>43.8</td>
<td>43.8</td>
</tr>
<tr>
<td>Al&lt;sub&gt;2&lt;/sub&gt;O&lt;sub&gt;3&lt;/sub&gt;</td>
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<td>13.0</td>
<td>10.6</td>
<td>10.5</td>
</tr>
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<td>1.18</td>
<td>2.33</td>
<td>1.55</td>
</tr>
<tr>
<td>FeO*</td>
<td>10.3</td>
<td>13.0</td>
<td>13.1</td>
<td>12.7</td>
</tr>
<tr>
<td>MnO</td>
<td>0.08</td>
<td>0.18</td>
<td>0.25</td>
<td>0.27</td>
</tr>
<tr>
<td>MgO</td>
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<td>13.6</td>
<td>13.4</td>
<td>14.2</td>
</tr>
<tr>
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<td>11.3</td>
<td>11.0</td>
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<td>2.20</td>
</tr>
<tr>
<td>K&lt;sub&gt;2&lt;/sub&gt;O</td>
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<td>0.37</td>
<td>0.21</td>
</tr>
<tr>
<td>Cr&lt;sub&gt;2&lt;/sub&gt;O&lt;sub&gt;3&lt;/sub&gt;</td>
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<td>-</td>
<td>-</td>
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<tr>
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<td>97.0</td>
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<table>
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<tr>
<td>Ti</td>
</tr>
<tr>
<td>Fe</td>
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<td>Mn</td>
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<td>Mg</td>
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<td>Ca</td>
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<tr>
<td>Na</td>
</tr>
<tr>
<td>K</td>
</tr>
<tr>
<td>Total</td>
</tr>
<tr>
<td>Mg + Fe</td>
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</table>

OU 38938 basalt containing hornblende, anorthite and olivine phenocrysts
OU 38565 basalt containing hornblende & anorthite phenocrysts and hornblende-anorthite xenoliths
OU 39050 basalt containing hornblende, anorthite & magnetite phenocrysts
OU 38562 basalt containing hornblende and anorthite phenocrysts

* analyses 10 μm apart on outer margin of core and inner margin of rim.
Table 6.4 cont. Analyses of hornblende from xenoliths in the MacKinnon Peak basalts.

<table>
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<tr>
<th>Xenoliths</th>
<th>38999z</th>
<th>38981z</th>
<th>38565z</th>
<th>39143D</th>
<th>39143E</th>
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<td>core</td>
<td>rim</td>
<td>core</td>
<td>rim</td>
<td>core</td>
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<tr>
<td>SiO₂</td>
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<td>44.1</td>
<td>41.6</td>
<td>42.1</td>
<td>43.5</td>
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<tr>
<td>Al₂O₃</td>
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<td>12.4</td>
<td>14.8</td>
<td>13.3</td>
<td>10.6</td>
</tr>
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<td>1.66</td>
<td>1.83</td>
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<td>0.15</td>
<td>0.32</td>
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<td>14.2</td>
<td>13.4</td>
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<tr>
<td>CaO</td>
<td>12.0</td>
<td>12.0</td>
<td>11.8</td>
<td>12.1</td>
<td>11.5</td>
</tr>
<tr>
<td>Na₂O</td>
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<td>2.54</td>
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<td>0.41</td>
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<td>Cr₂O₃</td>
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<td>0.02</td>
<td>0.03</td>
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<tr>
<td>Total</td>
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<td>97.1</td>
<td>98.0</td>
<td>97.7</td>
<td>97.1</td>
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</table>

| Si         | 6.24   | 6.39   | 6.06   | 6.18   | 6.48   | 6.31   | 6.34   | 6.32   | 6.43   |
| Al         | 1.76   | 1.61   | 1.94   | 1.88   | 1.52   | 1.69   | 1.66   | 1.68   | 1.57   |
| Al₂       | 0.60   | 0.50   | 0.60   | 0.43   | 0.34   | 0.51   | 0.54   | 0.53   | 0.46   |
| Ti         | 0.10   | 0.10   | 0.18   | 0.18   | 0.24   | 0.15   | 0.16   | 0.15   | 0.14   |
| Fe         | 0.91   | 0.91   | 1.29   | 1.38   | 1.68   | 1.49   | 1.62   | 1.53   | 1.96   |
| Mn         | 0.01   | 0.01   | 0.02   | 0.02   | 0.04   | 0.03   | 0.02   | 0.03   | 0.04   |
| Mg         | 3.58   | 3.66   | 3.14   | 3.12   | 2.98   | 3.05   | 2.87   | 3.01   | 2.62   |
| Ca         | 1.86   | 1.86   | 1.85   | 1.90   | 1.83   | 1.88   | 1.88   | 1.87   | 1.90   |
| Na         | 0.65   | 0.68   | 0.72   | 0.71   | 0.57   | 0.48   | 0.54   | 0.55   | 0.49   |
| K          | 0.12   | 0.10   | 0.08   | 0.08   | 0.04   | 0.06   | 0.07   | 0.07   |        |
| Total      | 15.83  | 15.82  | 15.88  | 15.88  | 15.69  | 15.65  | 15.70  | 15.74  | 15.68  |
| Mg         | 0.80   | 0.80   | 0.71   | 0.69   | 0.64   | 0.67   | 0.64   | 0.66   | 0.57   |

OU 38999z fine-grained hornblende-anorthite-(salite) xenolith (type 4)
OU 38981z coarse-grained hornblende cumulate xenolith (type 3)
OU 38565z hornblende-anorthite-salite xenolith (type 2)
OU 39143D hornblende-anorthite-salite xenolith (type 2)
OU 39143E hornblende-anorthite xenolith (type 2)

(for xenolith types see section 6.6)
[1] the amount of Na and K in the A site,
[2] the ratio Mg: \((\text{Fe}^{2+} + \text{Mg})\), and
[3] the proportion of Si in the Z site.

The proportion of Si in the Z site of the MacKinnon Peak amphiboles varies from 6.0 to 6.5. All analyses listed in Table 6:4 were determined by electron microprobe, and without wet chemical determinations of FeO it is possible to obtain only a minimum value for \(\text{Mg}/\text{Fe}^{2+} + \text{Mg}\) based on \(\text{Mg}/\text{Fe}^* + \text{Mg}\). The minimum values, however, indicate the amphiboles are intermediate in composition between tschermakite (/tschermakitic hornblende) and pargasite (/pargasitic hornblende).

6.10.2 Composition

The MacKinnon Peak hornblende are relatively constant in composition with respect to CaO, Na\textsubscript{2}O, TiO\textsubscript{2}, and K\textsubscript{2}O. The exception to this is the TiO\textsubscript{2} content of the hornblende phenocrystals with a brown core in OU 38565 which is discussed later in more detail. The principal variations in composition are between Si and Al\textsuperscript{IV} and Fe, Mg, and Al\textsuperscript{VI}. In general, Si and Al exist in excess of the requirements of the Z site, Mg dominates the Y site which contains subordinate Fe and Al, Ca is the principal cation in the X site and Na is essentially confined to the A site (Table 6:5).

6.10.3 Zoning in phenocrysts

The majority of phenocrysts in the MacKinnon Peak basalts are optically-zoned, but a few rocks contain optically-unzoned phenocrysts. Zoning is invariably simple in form, consisting of two zones, a broad core and a narrow rim (Fig. 6:5). The boundary between the zones generally corresponds to a sharp change in composition. Two forms of compositional change occur, one of which

\[ \text{Fe}^* = \text{total iron expressed as Fe}^{2+} \]
Table 6:5  Formulae for amphiboles of MacKinnon Peak based on procedure outlined in text.

<table>
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<tr>
<th></th>
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<th>38938</th>
<th>38565</th>
<th>38850</th>
<th>38562</th>
<th>38999z</th>
<th>38981z</th>
<th>38565z</th>
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<tr>
<td></td>
<td>core</td>
<td>rim</td>
<td>core</td>
<td>core</td>
<td>core</td>
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<td>Si</td>
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<td>1.76</td>
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<td>Al</td>
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<td>0.54</td>
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<td>0.74</td>
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<td>0.60</td>
<td>0.60</td>
<td>0.34</td>
</tr>
<tr>
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<td>0.13</td>
<td>0.26</td>
<td>0.11</td>
<td>0.16</td>
<td>0.10</td>
<td>0.18</td>
<td>0.21</td>
</tr>
<tr>
<td>Mg</td>
<td>3.29</td>
<td>3.00</td>
<td>2.95</td>
<td>3.17</td>
<td>3.29</td>
<td>3.58</td>
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<td>2.98</td>
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<tr>
<td>Fe</td>
<td>1.02</td>
<td>1.33</td>
<td>1.46</td>
<td>0.98</td>
<td>1.01</td>
<td>0.72</td>
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<tr>
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<td>0.03</td>
<td>0.03</td>
<td>0.01</td>
<td>1.01</td>
<td>0.02</td>
<td>0.04</td>
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<tr>
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<td>0.08</td>
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<td>0.64</td>
<td>0.65</td>
<td>0.72</td>
<td>0.57</td>
</tr>
<tr>
<td>K</td>
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<td>0.07</td>
<td>0.07</td>
<td>0.09</td>
<td>0.05</td>
<td>0.12</td>
<td>0.08</td>
<td>0.04</td>
</tr>
<tr>
<td>A Total</td>
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<td>0.85</td>
<td>0.84</td>
<td>0.83</td>
<td>0.88</td>
<td>0.69</td>
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</tbody>
</table>

Descriptions of specimens as for Table 6:4
has only been observed in OU 38565, the sole basalt containing brown hornblende phenocrysts. In OU 38565 the principal difference between core and rim is the high content of TiO₂ in the core zone. This is atypical of the MacKinnon Peak hornblendes and may reflect some local condition during early crystallization such as delayed onset in the crystallization of titanomagnetite. In the other basalts Si increases sharply at the expense of Al from core to rim, and in general Fe increases at the expense of the Mg ion. The rim itself changes continuously towards an Fe-rich composition across its width, whereas the core remains constant in composition (Fig. 6:18).

6.10.4 Zoning in xenoliths

The zoning of cumulus hornblende in xenoliths is more complex than that of phenocrysts in the host dykes. In general hornblende in adcumulate xenoliths is not optically zoned and compositionally shows weak trends towards silica and occasionally iron enrichment at the margin (Table 6:4, analysis 9 & 10, 11 & 12). Xenoliths with abundant intercumulus material contain hornblendes with a variety of forms of optical zonation. The two extremes are:

[1] two to four broad zones with resorbed margins (Fig. 6:19).

[2] a large number of oscillatory zones with margins paralleling the crystal outline (Fig. 6:20).

In both cases individual zones remain approximately constant in composition across their widths (Tables 6:6a and 6:6b). The boundary between adjacent zones is marked by a sharp change in iron and magnesium and oscillatory zoned crystals show overall trends towards Fe-rich rims (Table 6:6b).
Fig 6.18 Electron microprobe scans across zoned hornblende phenocryst for calcium, magnesium and iron, OU 38981, MacKinnon Peak basalt. Crystal consists of a broad magnesian core and a narrow iron-rich rim.
Table 6:6A

<table>
<thead>
<tr>
<th>Crystal 2</th>
<th>rim</th>
<th></th>
<th></th>
<th></th>
<th>core</th>
</tr>
</thead>
<tbody>
<tr>
<td>Al&lt;sub&gt;2&lt;/sub&gt;O&lt;sub&gt;3&lt;/sub&gt;</td>
<td>14.0</td>
<td>12.6</td>
<td>12.2</td>
<td>12.7</td>
<td>12.7</td>
</tr>
<tr>
<td>TiO&lt;sub&gt;2&lt;/sub&gt;</td>
<td>1.22</td>
<td>1.27</td>
<td>1.06</td>
<td>1.00</td>
<td>1.07</td>
</tr>
<tr>
<td>FeO</td>
<td>12.4</td>
<td>12.4</td>
<td>8.00</td>
<td>8.04</td>
<td>11.6</td>
</tr>
<tr>
<td>MnO</td>
<td>0.10</td>
<td>0.13</td>
<td>0.07</td>
<td>0.05</td>
<td>0.09</td>
</tr>
<tr>
<td>MgO</td>
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<td>13.5</td>
<td>16.3</td>
<td>16.5</td>
<td>14.2</td>
</tr>
<tr>
<td>CaO</td>
<td>11.9</td>
<td>11.8</td>
<td>11.7</td>
<td>11.8</td>
<td>11.9</td>
</tr>
<tr>
<td>Na&lt;sub&gt;2&lt;/sub&gt;O</td>
<td>2.42</td>
<td>2.50</td>
<td>2.68</td>
<td>2.72</td>
<td>2.40</td>
</tr>
<tr>
<td>K&lt;sub&gt;2&lt;/sub&gt;O</td>
<td>0.32</td>
<td>0.31</td>
<td>0.42</td>
<td>0.50</td>
<td>0.33</td>
</tr>
</tbody>
</table>

**Fig. 6:19** Simply-zoned hornblende crystal OU 38981B containing a resorbed Fe-rich core (analysis 9) surrounded by an Mg-rich zone (analyses 7 & 8) and an Fe-rich rim (analyses 5 & 6).
Table 6:6B Partial analyses of zoned hornblende OU 38981B

<table>
<thead>
<tr>
<th>Crystal 1</th>
<th>1</th>
<th>2</th>
<th>3</th>
<th>4</th>
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</thead>
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<tr>
<td>Al₂O₃</td>
<td>14.3</td>
<td>12.7</td>
<td>13.6</td>
<td>13.2</td>
</tr>
<tr>
<td>TiO₂</td>
<td>1.17</td>
<td>1.11</td>
<td>1.06</td>
<td>1.07</td>
</tr>
<tr>
<td>FeO</td>
<td>11.5</td>
<td>10.6</td>
<td>12.1</td>
<td>12.0</td>
</tr>
<tr>
<td>MnO</td>
<td>0.08</td>
<td>0.09</td>
<td>0.10</td>
<td>0.10</td>
</tr>
<tr>
<td>MgO</td>
<td>13.1</td>
<td>14.1</td>
<td>12.9</td>
<td>13.3</td>
</tr>
<tr>
<td>CaO</td>
<td>11.9</td>
<td>12.0</td>
<td>11.9</td>
<td>11.9</td>
</tr>
<tr>
<td>Na₂O</td>
<td>2.36</td>
<td>2.43</td>
<td>2.40</td>
<td>2.34</td>
</tr>
<tr>
<td>K₂O</td>
<td>0.42</td>
<td>0.47</td>
<td>0.45</td>
<td>0.42</td>
</tr>
</tbody>
</table>

Fig. 6:20 Oscillatory-zoned hornblende crystal, OU 38981B location of analysis 1-4, Table 6:6 as indicated.
6.10.5 Conditions of crystallization of hornblende

The key substitutions in the MacKinnon Peak hornblendes are $\text{Al}^{\text{VI}}$ for Si and Fe for Mg. Factors which influence these substitutions during hydrous crystallization of basaltic magma are documented by Helz (1973). Helz states that the $\text{Al}$-tschermakite content of hornblendes in three basaltic rock compositions investigated experimentally is dependent on temperature, and not bulk composition or $F_0_2$, and that $\text{Al}^{\text{IV}}$ increases steadily with increasing temperature (Fig. 6:21a). If this relationship holds for the MacKinnon Peak basalts it follows that all the hornblende phenocrysts, with the exception of those in OU 38565 crystallized at high temperature (Fig. 6:21b).

Furthermore, the cores of the phenocrysts must have crystallized at approximately constant temperature, and a sharp drop in temperature then occurred prior to the crystallization of the rims. During crystallization of the rims temperature decreased steadily. Hornblende phenocrysts and the hornblende within xenoliths in OU 38565 must have crystallized at temperatures considerably below that of all other MacKinnon Peak hornblendes. Interpretation of chemical data for xenolithic hornblende is more complicated. Adcumulus hornblende crystals appear to have formed at approximately constant temperature. The simply-zoned crystals appear to have undergone a limited number of sharp changes in temperature. Oscillatory-zoned hornblendes appear to have undergone fluctuations in temperatures superimposed on an overall trend of declining temperature.

In Helz’s experiments the $\text{Mg/Mg+Fe}$ ratio of hornblende increases steadily with temperature until approximately $930^\circ C$. 
Fig 6.21 Caption on adjacent page

(a) [Graph showing Al(IV) per 230 vs. T °C]

(b) [Graph showing Al (IV) per 230 vs. T °C]
then decreases sharply. The ratio varies directly with the Mg/Mg+Fe ratio of the host melt at a given $f_{O_2}$ and temperature. No simple interpretation is possible then for changing values of this ratio in the hornblentes of the MacKinnon Peak basalts. However, if Mg/Mg+Fe remains constant across the width of a crystal or a single zone, then either both temperature and the bulk composition have remained constant during crystallization, or temperature and the Mg/Mg+Fe of the melt have varied inversely. The latter possibility is unlikely since, in general, melts become increasingly rich in iron relative to magnesium with increasing differentiation and decreasing temperature, and the evidence above for Al$^{IV}$ content of hornblentes suggests the temperature of each core zone is basically constant. It therefore seems likely that the cores of hornblende phenocrysts in each MacKinnon Peak basalt and adcumulus hornblende crystals in each of the cognate xenoliths crystallized from a melt which remained of approximate constant bulk composition during the time in which the crystal cores formed.

6.11 Iron-Titanium oxides

All the MacKinnon Peak basalts and the majority of the xenoliths contain titanium-bearing magnetite. In the dykes magnetite occurs as phenocrysts and also in the groundmass. Ilmenite was not detected in any of the specimens examined. Electron microprobe analyses of magnetites from a dyke and three xenoliths are presented here, recalculated according to the methods of Carmichael (1957b). The magnetites
Table 6.7 Composition of magnetites in dykes and xenoliths, MacKinnon Peak Intrusives.

<table>
<thead>
<tr>
<th></th>
<th>38981A 1</th>
<th>39143A 2</th>
<th>39050</th>
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</thead>
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<tr>
<td>SiO₂</td>
<td>0.10</td>
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<tr>
<td>Al₂O₃</td>
<td>7.61</td>
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<td>TiO₂</td>
<td>3.80</td>
<td>5.34</td>
<td>4.97</td>
</tr>
<tr>
<td>FeO</td>
<td>76.0</td>
<td>85.6</td>
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<td>MnO</td>
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<td>0.50</td>
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<tr>
<td>MgO</td>
<td>3.29</td>
<td>0.96</td>
<td>0.85</td>
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<td>CaO</td>
<td>0.10</td>
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<td>0.01</td>
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<tr>
<td>Na₂O</td>
<td>0.08</td>
<td>0.00</td>
<td>0.05</td>
</tr>
<tr>
<td>K₂O</td>
<td>0.01</td>
<td>0.07</td>
<td>0.09</td>
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<td>Cr₂O₃</td>
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<td>Total</td>
<td>93.6</td>
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Recalculated analyses
Ilmenite basis
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<th>38981A 1</th>
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</thead>
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<td>Fe₂O₃</td>
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<td>93.2</td>
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Ulvospinel basis
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</thead>
<tbody>
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<td>Fe₂O₃</td>
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</tr>
<tr>
<td>FeO</td>
<td>30.0</td>
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</tr>
<tr>
<td>Total</td>
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<td>101.2</td>
<td>101.2</td>
</tr>
<tr>
<td>mol.% Usp</td>
<td>10.7</td>
<td>15.2</td>
<td>14.3</td>
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</tbody>
</table>

38981A Magnetite in fine-grained adcumulate xenolith (type 4)
39143A 1. magnetite in hornblende-anorthite-salite cumulate(type 3)
        2. magnetite in hornblende-anorthite cumulate(type 3)
39050 Magnetite phenocryst in hornblende-anorthite basaltic dyke.
are low in ulvospinel relative to spinels within calcalkaline basalts of similar composition, e.g. Smith and Carmichael (1968), Lowder (1970), Heming (1974). The MacKinnon Peak magnetites are also appreciably richer in chromium and aluminium than spinels of the other basalts referred to above. The magnetite within OU 38981, a fine-grained hornblende-anorthite xenolith, is exceptionally high in both these elements and in magnesium.

PETROLOGY

6.12 Relationship of cumulate xenoliths to dykes

Several points suggest the cumulate xenoliths are cognate in origin and crystallized at approximately the same time as phenocrysts. In particular the cumulate xenoliths contain all the minerals recorded as phenocrysts in the host MacKinnon Peak dykes (plus olivine and potassium feldspar), and the chemistry of the equivalent phases is closely similar. The texture of the xenoliths with abundant intercumulus materials (type 3) is gradational to the textures of phenocryst-rich dykes.

The well-developed cumulus textures of the xenoliths are unlikely to have formed in the dykes following their emplacement. Similarly phenocrysts up to 1 cm long have been found in fine-grained dykes as narrow as 4 cm. Accordingly it is thought the xenoliths and the bulk of the phenocrysts crystallized prior to emplacement of the dyke swarm.

6.13 Petrochemistry

Chemical analyses of four dykes and a xenolith are presented in Table 6:8. Details of analytical techniques are given in the appendix. Two of the dykes analysed do not show flow
Table 6.8 Chemical analyses of MacKinnon Peak basalts and xenolith.

<table>
<thead>
<tr>
<th>Oxides</th>
<th>38636</th>
<th>38938B</th>
<th>38999B</th>
<th>38010</th>
<th>39143B</th>
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<tr>
<td>SiO$_2$</td>
<td>52.40</td>
<td>51.89</td>
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<td>TiO$_2$</td>
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<td>0.45</td>
<td>0.18</td>
<td>0.82</td>
<td>0.61</td>
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<tr>
<td>Al$_2$O$_3$</td>
<td>18.94</td>
<td>19.36</td>
<td>17.64</td>
<td>15.08</td>
<td>20.44</td>
</tr>
<tr>
<td>Fe$_2$O$_3$</td>
<td>3.81</td>
<td>4.34</td>
<td>0.83</td>
<td>4.42</td>
<td>4.56</td>
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<td>0.09</td>
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<td>MgO</td>
<td>4.16</td>
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<td>4.96</td>
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<td>CaO</td>
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<td>K$_2$O</td>
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<td>P$_2$O$_5$</td>
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<td>21.8</td>
<td>-</td>
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<tr>
<td>Or</td>
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<td>3.1</td>
<td>4.3</td>
<td>4.1</td>
</tr>
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<td>Il</td>
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<td>2.69</td>
<td>2.3</td>
<td>2.4</td>
</tr>
</tbody>
</table>
differentiation, one containing 42% phenocrysts (OU 38938B), the other containing less than 1% phenocrysts. The third specimen was taken from an aphyric margin to a differentiated dyke (OU 38999A) and the fourth is thought to be from the axial region of another differentiated dyke (OU 38010), and contains 56% hornblende phenocrysts. The one xenolith analysed, a coarse-grained hornblende-anorthite cumulate (type 2), was collected from a massive hornblende-anorthite basaltic dyke.

The compositions of the two fine-grained specimens, OU 38636 and OU 38999A, are probably approximations to liquid compositions. The two porphyritic specimens are almost certainly partly cumulate in origin. The composition of the cumulate xenoliths may be related to that of two distinct crystal fractions being removed from the magma, probably prior to intrusion.

6.13.1 Massive dykes

The two dykes lacking flow differentiation are high-alumina basalts similar in composition to the Takitimu Group basalts, and to many such basalts in calcalkaline associations. The two basalts are themselves similar in composition despite the contrast in phenocryst content. The massive dykes are quartz- and hypersthene-normative. The composition of normative feldspar for both dykes is considerably less calcic than the composition of feldspar phenocrysts.

6.13.2 Flow-differentiated dykes

The compositions of the two specimens collected from flow-differentiated dykes reflect the chemistry of phenocryst minerals in the dykes. Specimen OU 38010 is a cumulate rock containing 56%
of large hornblende phenocrysts which form a framework-supported network with minor interstitial cryptocrystalline material. Consequently its composition closely resembles the hornblende analyses presented in section 6.10, and the specimen is very basic and olivine- and nepheline-normative. The high \( \text{Fe}_2\text{O}_3 \) content probably reflects amphibole composition and not subsequent oxidation.

The aphyric margin to a differentiated dyke (OU 38999A) represents a residual liquid from which flow differentiation has removed a considerable quantity of anorthite and hornblende crystals. The two phenocryst minerals are both rich in calcium and depleted in silica and sodium relative to the massive basalts, and so OU 38999A is higher in \( \text{SiO}_2 \) and \( \text{Na}_2\text{O} \), and low in \( \text{CaO} \) relative to massive basalts. Anorthite contains more aluminium and less iron and magnesium than the massive basalts, and the reverse holds true for these three elements in tschermakitic hornblende. As a result all three elements are reduced in OU 38999A relative to the massive basalts. The result of this concentration of early-formed crystals is that specimen OU 38999A has an unusual composition equivalent in silica content to high-\( \text{SiO}_2 \) andesites of calcalkaline or island arc tholeiite lineages, (e.g. Cascades Province, Carmichael, 1964; Santorini Volcano, Nicholls, 1971; Tonga, Bryan et al., 1972), but dissimilar in other major elements. The aphyric margin is high in \( \text{Al}_2\text{O}_3 \) and \( \text{Na}_2\text{O} \), and low in \( \text{TiO}_2, \text{FeO}^* \), \( \text{MgO} \), \( \text{CaO} \), and \( \text{K}_2\text{O} \) relative to calcalkaline andesites and dacites.

6.13.3 Xenolith

The composition of the cognate xenolith analysed is probably the composition of a fraction which was removed from
MacKinnon Peak magma by crystal fractionation.

The small size of the specimens located during this study precluded bulk chemical analysis of the hornblende-salite adcumulus xenoliths. Specimen OU 39143B is a hornblende-anorthite-(magnetite) cumulate. The phases present in this xenolith are the principal phases which have been removed, as phenocrysts, from portions of flow-differentiated dykes to yield aphyric margins. Therefore the chemical compositions of the aphyric margins should closely resemble compositions of hypothetical residual liquids which would form by continual crystal fractionation of the type which produced the hornblende-anorthite cognate xenoliths. If this is true, then the composition of OU 38999A can be used to discuss fractionation trends of the MacKinnon Peak basalts.

The resemblance between OU 38999A and the hypothetical residual liquids can be tested on a Harker diagram. The composition of liquids produced by hornblende-anorthite fractionation should plot along a smooth continuation of the major element trends between xenolith OU 38143B and the massive dykes. The composition of the aphyric margin OU 38999A plots very close to these trends as shown in Fig. 6:22. The composition of OU 38999A has therefore been used in the discussion of fractionation trends in section 6.15.

6.13.4 Comparison of the MacKinnon Basalts with Takitimu Group and similar calcalkaline lavas.

Despite similarity in bulk composition between the massive MacKinnon Peak basalts and basaltic lavas of the Takitimu Group the phenocryst assemblages of the two sets of rocks differ greatly, as does the chemical compositions of more differentiated members of the suites. The differences principally reflect the importance of
hornblende crystallization in the MacKinnon Peak dykes, and the absence of hornblende or any other hydrous phase as phenocrysts in the overwhelming majority of the Takitimu Group lavas. This contrast in phenocryst assemblages is probably a function of higher water pressure during crystallization of the MacKinnon Peak phenocrysts relative to phenocrysts in the Takitimu lavas. No xenoliths have been found in the Takitimu lavas to compare with the cognate inclusions in the MacKinnon Peak dykes. Santorini Volcano, however, consists of a suite of calcalkaline lavas very closely similar to the Takitimu lavas in chemical and mineralogical composition and degree of iron enrichment (see Chapter 4), and the Santorini lavas contain abundant cognate xenoliths (Nicholls, 1971).

The main series of lavas contains 'gabbroic' xenoliths of bytownite-anorthite, olivine, augite, magnetite and occasionally hypersthene, which possess cumulus textures. Nicholls suggested a zoned magma chamber existed below the volcano prior to the eruption of these rocks with the following sequence of crystalline assemblages:

bytownite-olivine, bytownite-olivine-augite, labradorite-augite-olivine-hypersthene and andesine-hypersthene-augite-magnetite passing inward from the chamber walls. Each successive assemblage contained an increasing proportion of acidic magma filling the interstices between the crystals which is now represented by intercumulus material. These xenoliths therefore resemble the cumulate xenoliths of the MacKinnon Peak basalts in texture, but not mineral assemblage. The assemblages recorded in the Santorini xenoliths are identical with the phenocryst assemblages found in the Santorini and Takitimu lavas, and presumably the xenoliths crystallized under conditions of low $P_{H_2O}$. Again the MacKinnon Peak
assemblages indicate the phases crystallized under conditions of high $P_{H_2O}$.

6.14 Kindred Associations

Tschermakite hornblende and anorthite are reported as phenocrysts or constituents of xenoliths from several calcalkaline volcanoes. Yamazaki et al. (1966) report plagioclase-hornblende-salite-spinel gabbroic inclusions and large plagioclase and hornblende crystals in pyroclastic rocks at Kujiranami and Shiganami Volcanoes, N.E. Japan.

Lewis (1973) describes cumulus plutonic blocks ejected in pyroclastic flows by Soufriere Volcano, St Vincent, West Indies. The blocks consist of anorthite ($An_{97-89}$), olivine ($Fo_{79-67}$), aluminous salite, tschermakite hornblende and magnetite. An interstitial glass film commonly separates adcumulus grains in the xenoliths. The parent magma is considered to have been high-alumina basalt and olivine and anorthite are the first phases to have formed, suggesting the Soufriere magma crystallised under conditions of lower $P_{H_2O}$ than the MacKinnon Peak magma. Similar rocks have been described from other islands in the Lesser Antilles, St Kitts (Baker, 1968), Grenada and Carriacou (Lewis, 1973).

6.15 Fractionation trends

Fractionation trends resulting from removal of the phenocrysts from the MacKinnon Peak basalts are relevant to discussion of the genesis of calcalkaline magma lineages by amphibole fractionation (Holloway and Burnham, 1972; Eggler and Burnham, 1973; Boettcher, 1973; Cawthorn and O'Hara, 1976). Removal of hornblende-rich crystal fractions from a basaltic parent is an attractive hypothesis to explain formation of calcalkaline magma series because:

[1] most igneous amphiboles are subsilicic and their
removal from a basaltic parent causes a rapid increase in silica content of the residual liquids.

[2] Hornblende in basaltic melts (e.g. Helz, 1973) has an appreciably higher Fe/Fe+Mg coefficient than coexisting olivine and pyroxenes. Therefore liquids produced by amphibole fractionation of basalt show less iron enrichment than liquids produced by olivine and pyroxene fractionation.

It has been suggested that an andesitic liquid may be produced from a tholeiitic magma by removal of hornblende-olivine-clinopyroxene (Holloway and Burnham, 1972; Cawthorn and O'Hara, 1976), or removal of hornblende-plagioclase-orthopyroxene from magmas of intermediate composition (Cawthorn and O'Hara, 1976). Two principal assemblages have crystallized from the MacKinnon Peak basalts: hornblende-sanidine-(olivine-magnetite) and hornblende-plagioclase-(magnetite). Evidence summarised earlier suggests the former crystallized first. Removal of crystal fractions, corresponding to each of the assemblages, from the host basalts would produce two distinct fractionation trends. The removal of a crystal fraction equivalent to the second assemblage would produce a liquid similar in composition to that of specimen OU 38999A enriched in silica relative to the host basalt, but very different in composition to calcalkaline andesites. In contrast fractionation involving the first assemblage may produce a liquid similar in most major elements to calcalkaline andesite. However, all four minerals concerned contain less alumina than the host basalt, and therefore any differentiates will be enriched in alumina. The MacKinnon Peak basalts are high in alumina and removal of the first assemblage by crystal fractionation would produce a differentiate containing
greater than 19% alumina, unusually high for an andesite
(cf Chayes, 1969). This has important ramifications because if
calcalkaline magmas are generated by amphibole fractionation,
then either —

1] the parent basalt must be a tholeiite rather than
the typical calcalkaline high-alumina basalt, or

2] the amphibole must be unusually high in alumina.

6.16 Model for Crystallization
Evidence has been cited previously (section 6:12) to suggest
that the phenocrysts and cognate xenoliths of the MacKinnon Peak
dykes formed prior to the emplacement of the dyke swarm. If this is
so then the MacKinnon Peak magma has undergone two discrete periods
of crystallization. In the first period of crystallization the
phenocrysts and cumulate xenoliths formed. The groundmass of the
dykes and possibly some intercumulus material in the xenoliths are
the products of the second period of crystallization. The phases
produced in the second period are extremely fine-grained and
presumably formed during rapid cooling of the dykes. The dykes
are thought to have been emplaced at shallow levels, between the
two periods of crystallization, as a suspension of crystalline
materials in magma.

Mineral phases formed during the first period of
crystallization were concentrated by two processes; by crystal
settling to form cognate cumulate xenoliths and by flow differentiation
during emplacement. Flow differentiation produces aphyric margins to
the dykes which are probably very similar in composition to residual
liquids formed by crystal fractionation.
6.17 First period of crystallization

The large size and high concentration of phenocrysts and the concentration and cumulate textures of cognate xenoliths suggest that during the early period of crystallization the material forming the dykes and cognate xenoliths were part of one, or possibly more, larger bodies of differentiating basalt magma. The xenoliths showing well-developed adcumulus textures and constant composition of mineral phases, have presumably crystallized under an approximation to equilibrium conditions. In contrast the xenoliths containing euhedral, complexly-zoned crystals, and also zoned hornblende phenocrysts in the host dykes must have formed in a more unstable environment. The coarse-grained xenoliths contain the following cumulus assemblages: hornblende-salite+ olivine+ magnetite, hornblende-magnetite, hornblende-anorthite-salite, hornblende-anorthite, hornblende. Textural gradation exists from the first assemblage, marked by the most well-developed adcumulate textures, to the last with the best crystal outline and the greatest amount of intercumulus material.

Other observations relevant to the crystallization history of the first period are:

1. hornblende predominates in every xenolith assemblage,
2. salitic pyroxene and olivine occur in true adcumulus relationship to hornblende suggesting the crystallization intervals of the phases overlap at least in part, but occasionally salitic pyroxene and olivine grains are mantled and replaced by hornblende suggesting termination of crystallization of the anhydrous phases first. The reverse relationship has not been observed,
3. olivine and plagioclase have not been observed
coexisting in any xenolith, and coexisting anorthite and salite are rare.

Using this evidence, and assuming xenoliths with well-developed adcumulate texture crystallized before the xenoliths with abundant intercumulus material a probable sequence of crystallization has been derived.

<table>
<thead>
<tr>
<th>phase</th>
<th>first period of crystallization</th>
<th>second period of crystallization</th>
</tr>
</thead>
<tbody>
<tr>
<td>olivine</td>
<td></td>
<td></td>
</tr>
<tr>
<td>salitic pyroxene</td>
<td></td>
<td></td>
</tr>
<tr>
<td>hornblende</td>
<td></td>
<td></td>
</tr>
<tr>
<td>anorthite</td>
<td></td>
<td></td>
</tr>
<tr>
<td>magnetite</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

This sequence can be used to make some assumptions concerning $P_{H_2O}$ during crystallization utilising the experimental data of Yoder and Tilley (1962) and Green and Ringwood (1967, 1968).

The massive dykes are high alumina basalts in composition, as presumably was the parent melt from which the cognate xenoliths crystallized. Yoder and Tilley (1962) showed that in a high-alumina basalt the anhydrous near-liquidus assemblage of olivine-clinopyroxene-plagioclase-magnetite is replaced by assemblages dominated by calcic plagioclase and amphibole as $P_{H_2O}$ is increased (Fig. 6:23). The critical value of $P_{H_2O}$ above which amphibole crystallizes before plagioclase is approximately 2.7 kb. This places an approximate lower limit for $P_{H_2O}$ during crystallization of the adcumulate hornblende-salite-olivine xenoliths. Yoder and Tilley (op. cit.) and Green and Ringwood (1967, 1968) have shown independently that amphibole dominates the crystallization of high-alumina basalt at 10 kb $P_{H_2O}$ and plagioclase appears only near the solidus. The presence of
Fig 6.23 Pressure-temperature projection of a natural high alumina basalt-water system Yoder and Tilley (1962, p 451, fig. 28).
hornblende-plagioclase cumulate xenoliths, and the common occurrence of plagioclase as a phenocryst in the dykes, suggests that $P_{H_2O}$ during the first period of crystallization may be far less than 10 kb. As an approximation the values of 2.7 kb and 10 kb have been taken as limits to the initial value of $P_{H_2O}$ for the early period of crystallization.

INTERPRETATION

6.18 Introduction

The MacKinnon dyke swarm is thought to represent portions of a partially crystallized body (or bodies) of magma. The body was disrupted and emplaced as a series of narrow dykes. A model for the crystallization of shallow plutons is therefore necessary to understand the pre-emplacement history of the MacKinnon Peak intrusives. Evidence for such a model comes from study of shallow plutonic bodies (e.g. Ragland and Butler, 1972) or lava lakes (Ritcher and Moore, 1966).

Ritcher and Moore (1966) suggest the Kilauea Ika lava lake, Hawaii, consists of three zones: a lava crust, an underlying 'transient zone of crystallization' containing up to 30% interstitial fluid, and thirdly melt. The West Farrington Pluton, North Carolina studied by Ragland and Butler (1972) is a concentrically zoned body which shows increasing amounts of intercumulus material towards the centre of the body. Ragland and Butler consider that at any point in time the body consisted of three zones:

[1] magma containing a variable amount of crystals,

[2] 'liquidus minerals' forming a framework with interstitial magma,

The concept of a transitional zone of crystalline material with abundant interstitial fluid is fundamental to the models proposed in both these studies. The textures described by Ragland and Butler from the middle zone of the West Farrington Pluton closely resemble those of the MacKinnon Peak xenoliths with abundant intercumulus material. In the model given below the early stages of crystallization of the MacKinnon Peak magma closely resemble the equivalent stages of crystallization of the West Farrington Pluton.

6.19 Pre-emplacement History

Crystallization of the parent body (or bodies) of the MacKinnon Peak dykes probably commenced under conditions of high $P_{H_2O}$ and early-formed crystals, hornblende, clinopyroxene and olivine accumulated by crystal settling. The xenoliths containing well-developed adcumulus textures probably formed at this time. With increased differentiation a situation was probably reached where the body consisted of a stratified sequence passing from crystalline adcumulates through a transitional layer of cumulates with well-formed crystals and abundant intercumulus fluid, to a suspension of crystals in magma. Crystallization proceeded to a stage where salite and olivine ceased to crystallize, and anorthite had commenced to precipitate, accompanying hornblende. The body (bodies) was then disrupted and material from the body given access to high levels in the Takitimu Group volcanic pile. Rapid fluctuations of pressure due to the removal of magma may have produced the strongly oscillatory-zoned hornblende crystals in the intercumulus-rich xenoliths (type 3). Under such conditions the crystals may also have been brought rapidly into contact with either more or less differentiated magma than the liquid in which they were initially in equilibrium. Some of the magma escaping to higher levels may
have been largely fluid giving rise to the nonporphyritic dykes, but the crystalline and semi-crystalline layers were disrupted and most portions of the magma carried fragments of these layers upwards as xenoliths. In this model the adcumulate xenoliths probably represent broken portions of the crystalline layer, the intercumulus-rich xenolith represent portions of the transitional layer, and the phenocrysts represent crystals suspended in the melt.

6.20 Emplacement of the dykes

Members of the dyke swarm are relatively thin and contain high crystal concentrations in a siliceous groundmass. Despite these factors the dykes have penetrated considerable distances even through massive intrusive rocks (e.g. OU 38941). Relatively high mobility of the basalts is indicated by:

1. strong flow differentiation of some members,
2. well-developed flow-banding,
3. length to width ratios exceeding 100:1 even in the limited exposures of the dykes which are visible,
4. granulation of phenocrysts near the dyke margins,
5. narrow dykes splitting around resistant obstacles, e.g. pillow lavas and forming tongues and isolated bodies in fine-grained sediments.

Despite the overall basaltic composition of the dykes, prior to emplacement they consisted of a suspension of coarsely crystalline material in a comparatively siliceous fluid (66% SiO₂ in the aphyric margin of OU 38999). This system probably has much greater viscosity than a totally molten high-alumina basalt of identical bulk chemical composition.

Two factors are involved here:

1. the addition of solid particles to a fluid greatly increases its viscosity (Johnson and Pollard, 1973),
crystallization of hornblende and anorthite results in a siliceous residual liquid which even without the presence of crystals would be considerably more viscous than the parent basalt (Johnson and Pollard, op. cit.).

The combined result is to produce viscosities which greatly exceed theoretical values for suspensions formed simply by the addition of solid particles to a fluid of constant composition.

Such a relationship has been quantified for Hawaiian tholeiitic lavas and the viscosity of magma containing 25% phenocrysts is one thousand times the viscosity of the equivalent phenocryst-free magma (Johnson and Pollard, 1973).

Several facts suggest the fluid behaviour of the MacKinnon Peak magma may be a function of high volatile content namely, the initial $P_{H_2O}$ of the magma as deduced from the mineral assemblages, the widespread occurrence of small vesicles in the dykes, and the occurrence of hydrous minerals (zeolites and actinolite) as intercumulus phases in xenoliths. The effect of water content on viscosity of a melt has been documented for obsidian (Johnson and Pollard, 1973), and this indicates that the addition of even 2% of $H_2O$ lowers the viscosity of the melt by several orders of magnitude. The intrusion of the dykes may also be facilitated by formation of a separate water-rich phase and by volatile streaming during emplacement. During ascent of a body of hydrous magma total pressure decreases more rapidly than the equilibrium pressure of water ($P^e_{H_2O}$) in the melt because of low diffusion rates in magma, (Burnham, 1967). For this reason at some point $P^e_{H_2O}$ must exceed $P_{total}$ and a water-rich phase separate. This will occur regardless of whether or not early boiling was
produced by increasing $P_{H_2O}$ during fractionation as discussed previously. The water fraction thus evolved may provide a low viscosity precursor of the magma which facilitates intrusion as has been proposed for carbonatite magmas (Currie and Ferguson, 1970). The water fraction precedes the magma along a fissure partly opening and clearing a path for the advancing water-saturated magma. The loss of the water fraction causes a lowering of crystallization temperatures, and crystallization proceeds rapidly (Burnham, 1967). If plagioclase and magnetite crystallize this will increase $P_{H_2O}$ and so repeated boiling may occur. Loss of a volatile phase in this manner may explain the absence of hornblende in the groundmass of the dykes which contain the anhydrous assemblage plagioclase–magnetite.

6.21 Flow differentiation

Flow differentiation is produced during the laminar flow through a narrow conduit of a viscous fluid which contains suspended particles (Bhattacharji and Smith, 1964; Gibbs, 1968; Komar, 1972). Bhattacharji (1967) and Bhattacharji and Smith (1964) considered the migration of particles towards the centre of dykes was due to wall effects and forces acting on individual particles. Komar (1972) discarded these hypotheses relating the migration to dispersive pressures generated when shearing of the fluid produced particle interaction.

Particle dispersive pressures must be a major factor in the MacKinnon Peak dykes because of the high concentration of phenocrysts. For concentration above 8% by volume, particle interaction is inevitable because the average free distance between spherical grains is less than the average grain diameter (Bagnold, 1954 reported in Komar, 1972), and grain interactions must occur.
In massive MacKinnon Peak dykes concentrations of phenocrysts commonly exceed 30%. In addition the fractured phenocrysts aligned parallel to the margins of aphyric dykes indicate that shearing, considered by Komar to be a prerequisite for dispersive pressures, has taken place. The relationship between velocity profile and concentration of phenocrysts (by volume) is shown in Fig. 6:24(a & b), for two types of flow behaviour described by Komar (1972). During laminar flow the velocity profile across the conduit is parabolic (Fig. 6:24), during plug flow the central portion of the velocity curve is flattened (Komar, 1972). Komar relates dispersive pressure exerted on phenocrysts in a dyke directly to the velocity gradient across the dyke. For laminar flow the velocity and dispersive pressure gradients increase linearly across the dyke from zero at the centre to a maximum at the walls. Phenocrysts migrate from the walls of the dyke because of the pressure gradient, resulting in a high concentration of phenocrysts at the axis of the dyke.

During plug flow the velocity gradient is effectively zero over a large portion of the dyke and phenocryst migration is confined to the margin of the body. The central portion of the dyke therefore contains a uniform concentration of phenocrysts and the concentration decreases smoothly towards the dyke margins. Komar (1972) states that a plug flow velocity profile yields a phenocryst distribution very similar to those observed in flow differentiated dykes. Komar has established the relationship

\[ P_d = \frac{dv}{dy} \]

when \( P_d \) is dispersive pressure, \( v \) velocity and \( y \) dyke width.

It follows therefore, that the extent of flow differentiation will increase with increasing velocity or with decreasing dyke width.
Fig 6.24 Caption on adjacent page
width. This contrasts with the observation of Gibb (1968) in the ultrabasic dykes of Skye, that the maximum concentration of phenocrysts increases with dyke width. Gibb related this to a postulated increase in velocity of the thicker dykes. This, from the relationship above, is the case only if both velocity and velocity gradient increases with width. There are no conspicuous differences in thickness or any other aspect of morphology between massive and flow-differentiated MacKinnon Peak dykes. In fact the most striking flow-differentiation occurs in the thinnest dykes. Velocity of emplacement of the MacKinnon Peak dykes probably bears no relationship to dyke width and relates instead to volatile content. A dyke which has retained high volatile content and undergoes strong late stage retrogressive boiling will be emplaced at higher velocity than a comparable dyke of lower volatile content and is more likely to experience flow differentiation.
CHAPTER 7 - METAMORPHISM

INTRODUCTION

Grade of metamorphism and degree of mineral reconstitution increase with stratigraphic depth in the Takitimu section. Two broad metamorphic zones are distinguished, the upper (Zone 1) correlated with the zeolite facies, and the lower (Zone 2) with the prehnite-pumpellyite facies (Map 2.). Study of the area reveals many low-grade metamorphic features observed in other volcanic piles, especially:

1. gross overlap and sporadic distribution of hydrated calcium aluminosilicate phases,
2. absence of pervasive penetrative deformation,
3. extensive development of secondary minerals without destruction of primary igneous textures,
4. mineralogical readjustment varying from slight to near complete, on the scale of an outcrop, vertically and laterally, with relict phases particularly magnetite and clinopyroxene, present even in some of the most heavily altered rocks.

In originally porous rocks, void space (vesicles in lava flows, intergranular space in volcaniclastic strata) has been infilled by hydrated calcium aluminosilicates, calcite and quartz. Fractures in all rock types are infilled by these secondary minerals. Some dense impermeable horizons including glassy lava flows (e.g. OU 38950) show no visible signs of metamorphism in regions remote from fractures.

7.1 Shear zones

Zones of severe alteration are present in and around shear and fault systems. The zones are among the most striking features of the
Fig 7.1 Metamorphic zonation in the central Takitimu Mountains. Boundary between zones one and two established on the basis of data from Map 2 (in back pocket).
Fig 7.2 Epidote metadomains formed in and adjacent to fractures in porphyritic microdiorite, OU 38866 upper gorge, Whare Creek. S159/ 864897.
metamorphic geology of the Takitimu Mountains. Many of the zones are developed at the margins of White Hill intrusive bodies. Large fault crush zones contain resistant 'knockers' in a matrix of pervasively sheared chlorite, whereas smaller shear zones may be infilled by slickensided quartz and epidote or prehnite without chlorite.

The larger shear zones are cut by fractures up to 10 cm wide infilled by prehnite-quartz (Zone 1) or epidote-quartz (Zone 2) often at irregular angles to shearing directions in the chloritised rock. The host rock is often cut by irregular anastomising networks of veins of laumontite (Zone 1) or prehnite (Zone 2).

Adjacent to the fractures in both large and small shear zones, the host rock is sometimes largely replaced by quartz and hydrated calcium aluminosilicate minerals without destruction of igneous textures. To illustrate this, table 7:1 lists the modal composition of three specimens collected adjacent to fractures in Zone 2. Photomicrographs of the thin sections analysed are included in Fig. 7:3 to show the relict igneous textures. These rocks now contain 75-80% hydrated calcium aluminosilicate minerals and quartz.

Smith (1968) described similar textures from the Ordovician Walli andesite in the Lachlan Geosyncline. Areas (generally less than 30 cm across) occur in a single outcrop where the primary igneous mineralogy of the rock has been replaced by epidote and/or pumpellyite. Smith calls these areas 'domains', a term later refined by Jolly and Smith (1972) to 'metadomains'. The metadomains described from the Takitimu Mountain sequence are discussed more fully in subsequent sections.

The larger shear zones resemble 'fossil' hydrothermal zones reported from the Tanzawa Mountains of Japan (Seki et al., 1969; D.S. Coombs, pers. comm.). They appear to have acted as channel-ways for substantial bodies of fluid during metamorphism. Where metadomains have developed
Table 7.1 Modal analyses based on 500 counts of three heavily altered igneous rocks adjacent to shear zones.

<table>
<thead>
<tr>
<th>mineral/rock</th>
<th>OU 38902</th>
<th>OU 38866A</th>
<th>OU 38866</th>
</tr>
</thead>
<tbody>
<tr>
<td>epi</td>
<td>71.2%</td>
<td>56.4%</td>
<td>54.2%</td>
</tr>
<tr>
<td>pre</td>
<td>-</td>
<td>-</td>
<td>12.4%</td>
</tr>
<tr>
<td>qz</td>
<td>-</td>
<td>24.0%</td>
<td>25.4%</td>
</tr>
<tr>
<td>plag micro.</td>
<td>16.6%</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>plag pheno.</td>
<td>12.2%</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>actinolite</td>
<td>-</td>
<td>7.2%</td>
<td>6.8%</td>
</tr>
<tr>
<td>magnetite</td>
<td>-</td>
<td>2.4%</td>
<td>1.2%</td>
</tr>
</tbody>
</table>

OU 38902 porphyritic microgabbro at margin of White Hill Intrusive sill

OU 38866 hornblende metadiorite, OU 38866A in contact with quartz-epidote filled fracture; OU 38866 10 cm from fracture.
Fig 7.3 Relict igneous textures preserved in epidote metadomains, modal analyses for these specimens are given in Table 7.1.

(a) Albitised plagioclase, actinolite (replacing hornblende) and magnetite set in equigranular epidote-quartz, OU 38666A (Table 7.1, no. 2).

(b) Albitised plagioclase phenocryst and microlites set in epidote, OU 38902 (Table 7.1, no. 1).
adjacent to fractures, the fluid has substantially modified the chemistry of the rock.

7.2 Metamorphic Zones

The distribution of key metamorphic minerals is indicated on Map 2 (back pocket). Laumontite and epistilbite are the dominant hydrated calcium aluminosilicates in the top six km of the Takitimu Mountains sequence. Prehnite is the key mineral in the lower nine km of the sequence, and occurs sporadically in the lower 500 m of Zone 1. The boundary between the two zones is taken as the lowest occurrence of laumontite. Epidote increases greatly in abundance below Zone 1.

<table>
<thead>
<tr>
<th>Phase</th>
<th>Zone 1</th>
<th>Zone 2</th>
</tr>
</thead>
<tbody>
<tr>
<td>laumontite</td>
<td></td>
<td></td>
</tr>
<tr>
<td>epistilbite</td>
<td></td>
<td></td>
</tr>
<tr>
<td>heulandite</td>
<td></td>
<td></td>
</tr>
<tr>
<td>stilbite</td>
<td></td>
<td></td>
</tr>
<tr>
<td>chabazite</td>
<td></td>
<td></td>
</tr>
<tr>
<td>analcime</td>
<td></td>
<td></td>
</tr>
<tr>
<td>prehnite</td>
<td></td>
<td></td>
</tr>
<tr>
<td>pumpellyite</td>
<td></td>
<td></td>
</tr>
<tr>
<td>epidote</td>
<td></td>
<td></td>
</tr>
<tr>
<td>albite</td>
<td></td>
<td></td>
</tr>
<tr>
<td>actinolite</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

Table 7:2 shows the approximate distribution of some metamorphic phases in the two zones.
MINERAL ASSEMBLAGES

Recognition of equilibrium assemblages in such low-grade rocks constitutes a problem (e.g. Zen, 1974). In the next sections tables 7:3 and 7:4 list assemblages recorded in Zones 1 and 2. In each instance the phases recorded were observed in contact with each other in a field of view of less than 4 mm diameter. The assumption that these assemblages formed under approximately equilibrium conditions is partly justified by:

(1) Uniformity of composition with respect to major elements of all phases (with the possible exception of actinolite, and in one specimen, prehnite) on the scale of a thin section.

(2) The lack of textural evidence for reaction relationships between phases, as in the presence of embayed grain boundaries or contrasting grainsizes within an assemblage (e.g. Glassley, 1974).

The assemblages are listed for a variety of local environments, e.g. vesicles, veins, fractures in shear zones. In general, compatible assemblages are seen in all local environments in each zone, except for shear zones and major fractures. In shear zones the characteristic assemblage is typically less hydrous than assemblages found in the host rock away from the shear zones (see tables 7:3 and 7:4).
ZONE ONE: ZEOLITE FACIES

Mineral reconstitution is seldom greatly advanced in Zone 1. In thin sections of volcanic, volcaniclastic and intrusive rocks from Zone 1 plagioclase feldspar is completely albitised in less than 20%, partly albitised in approximately 50%, and unaltered in the remainder, 1-2% show signs of replacement by chlorite. The original presence of olivine is recognised, with one exception (OU 38471) solely on the basis of distinctive pseudomorphs in thin section (Fig. 7:4). In OU 38471 olivine has been preserved because it is mantled by overgrowths of hypersthene. Augite is completely fresh throughout Zone 1, hypersthene commonly rimmed by iddingsite and/or chlorite, and magnetite shows patchy partial alteration to hematite and sphene. Volcanic glass is commonly palagonitised (especially in pillow lavas) or chloritised but rare specimens contain glass showing no visible signs of alteration (S159/932835, OU 38950). The most striking development of hydrated calcium-and sodium-aluminosilicates is within void spaces in the host rock.

Table 7:3 lists partial mineral assemblages recorded in Zone 1.
Table 7:3  Mineral Assemblages recorded in Zone 1, Takitimu Mountains

<table>
<thead>
<tr>
<th>Environment</th>
<th>Partial mineral assemblage ++</th>
</tr>
</thead>
<tbody>
<tr>
<td>vesicle</td>
<td></td>
</tr>
<tr>
<td>lau</td>
<td>epis</td>
</tr>
<tr>
<td>lau-epis</td>
<td>epis-cc</td>
</tr>
<tr>
<td>heu-cel</td>
<td>epis-chl</td>
</tr>
<tr>
<td>lau-qz</td>
<td>epis-co-pr</td>
</tr>
<tr>
<td>lau-chl</td>
<td>epis-chl-pr</td>
</tr>
<tr>
<td>lau-pr</td>
<td>epis-chl-anal</td>
</tr>
<tr>
<td>lau-co-(hm)</td>
<td>epis-chl-cel</td>
</tr>
<tr>
<td>lau-chl-anal</td>
<td>epis-lau-anal-cc</td>
</tr>
<tr>
<td>lau-chl-pu</td>
<td>epis-co-chl-(ep)</td>
</tr>
<tr>
<td>lau-stil*</td>
<td>epis-chl-pr-pu</td>
</tr>
<tr>
<td>stil-cc*</td>
<td>chl</td>
</tr>
<tr>
<td>lau-pr-pu-qz</td>
<td>chl-qz</td>
</tr>
<tr>
<td>lau-cc-chl</td>
<td>anal-nat-thom</td>
</tr>
<tr>
<td>vein</td>
<td></td>
</tr>
<tr>
<td>laz</td>
<td>anal</td>
</tr>
<tr>
<td>laz-epis</td>
<td>stil*</td>
</tr>
<tr>
<td>laz-cc</td>
<td>stil-chab*</td>
</tr>
<tr>
<td>laz-co(pr)</td>
<td>lau-(hm)</td>
</tr>
<tr>
<td>mineral cement</td>
<td>lau</td>
</tr>
<tr>
<td>replacing glass of acidic tuffs</td>
<td>heu-qz</td>
</tr>
<tr>
<td>replacing glassy basic rock fragments</td>
<td>chl</td>
</tr>
<tr>
<td>mudstones*</td>
<td>chl</td>
</tr>
<tr>
<td>replacing plagioclase</td>
<td>ab</td>
</tr>
<tr>
<td>shear zones and major fractures</td>
<td>pr</td>
</tr>
<tr>
<td></td>
<td>qz-pr-chl</td>
</tr>
<tr>
<td></td>
<td>qz-pr-chl(-pu)</td>
</tr>
</tbody>
</table>

* only in MacKinnon Peak intrusives  ( ) minor in abundance.

+ determined by X-ray diffraction studies of untreated, glycolated and heated samples. Whether the mont-chl assemblage represents two discrete phases or is in fact a regular interlayered corrensite has not been determined.

ab albite; anal analcime; cc calcite; cel celadonite; chab chabazite; chl chlorite; ep epidote; epis epistilbite; heu heulandite; hm hematite; lau laumontite; mont montmorillonite; nat natrolite; pr prehnite; pu pumpellyite; qz quartz; stil stilbite; thom thomsonite.

++ Note two columns are used only to conserve space, assemblages in both columns are equivalent in every respect.
7.3 Basic volcaniclastic rocks

In well-sorted coarse clastic rocks, including agglomerates and litharenites, coarse-grained prismatic laumontite commonly forms a mineral cement. In several specimens prisms form radiating aggregates extending inwards from clast boundaries, suggesting growth into a void (Fig. 7:4). In tuffaceous breccias and lapilli tuffs the ash-sized matrix may be reconstituted to chlorite. The glass of volcanic rock fragments shows a wide range of alteration, ranging from complete replacement by crystalline chlorite through weakly birefringent material to apparently unaltered isotropic glass. Glassy clasts within a single horizon show varying degrees of alteration. Alteration is most pronounced near the margin of rock fragments and decreases concentrically inward.

In rare instances (OU 38751 and OU 38945) the arenite portions in well bedded turbidites are largely reconstituted to laumontite. The beds are buff to tan in colour compared to the normal grey-brown litharenites, and are highly porous. In thin section an original volcaniclastic texture is apparent in plane polarised light but under crossed nicols the rock is seen to be a coarse-grained aggregate of equidimensional laumontite prisms. This is one of the few instances where volcanic glass is seen to be replaced by calcium alumino-silicate in the Takitimu sequence.

7.4 Acidic volcaniclastic rocks

Two acidic vitric tuffs present in the Heartbreak and MacLean Peaks formations show strong alteration. In OU 39069 (S159/908827) heulanditized shards and rare crystal debris are set in a matrix of finely-divided glass now completely replaced by heulandite. Shard outlines
Fig 7.4 Laumontite cement in tuff breccia, OU 38348, northern Takitimu Mountains.
Photograph K.D. Mason.
are sharp and sedimentary textures are preserved (see Section 3.4, page 47). Burrows infilled by well-sorted, relatively coarse crystal detritus are heulandite-poor, and rare entire bubbles are infilled by platy heulandite and celadonite. A second tuff band 500 m higher in the sequence (3159/926831, OU 38945) is recrystallised to an equigranular mosaic of hematite-studded laumontite (Fig. 7.5a and b).

7.5 Fine-grained sediments

Mudstones, argillites and fine-grained crystal-lithic tuffs generally show few mesoscopic signs of metamorphism. Colour mottling is occasionally present in these fine-grained rocks with ovoid leu­cocratic domains up to 5 mm across (Fig. 7.6). On close inspection, two textural forms of mottling can be distinguished:

1. Sharply bounded blebs or maculae with an appreciably higher content of laumontite than the host rock (e.g. OU 38788).

2. Diffuse maculae (e.g. OU 38872D) distinguished principally by a lower abundance of chlorite than the host rock. Boles (1971) reports similar mottling in Murihiku Supergroup sandstones in the Hokonui Hills.

Clay fractions were concentrated from four mudstones by hand-crushing and centrifuging. The concentrates were identified using standard X-ray diffraction techniques outlined by Carroll (1970). Two samples contain only chlorite and two give chlorite and montmorillonite peaks.

7.6 Pillow lava/pillow breccia

In pillow lava sequences pillow interstices are commonly infilled with quartz-prehnite or laumontite. Pillow margins are palagonitised. The typical textural sequence across a pillow is described in Chapter 3.
Fig 7.5 Laumontite-quartz metamorphic assemblage, developed in rhyodacitic vitric tuff, OU 38945, Zone 1.

(a) plane polarised light. Relict tuffaceous texture and large burrow evident.

(b) same view, crossed Nicol prisms showing tuff to consist of an equigranular mosaic of laumontite and quartz.
Fig 7.6 Laumontite-mottled siltstone, OU 38788, Zone 1, central Takitimu Mountains.
Photograph K.D.Mason.

Fig 7.7 Twinned epistilbite in segregation vesicle, porphyritic basalt, OU 38097, Zone 1, crossed Nicol prisms.
Elongate, lobate segregation vesicles are common in lava pillows and are infilled by zeolites, principally laumontite, as are radial fractures in the pillows.

7.7 Massive lava flows

In massive lava flows metamorphic reconstitution is localised about fractures and flow margins. Vesicles are infilled by a variety of assemblages listed in table 7:3, but simple one- and two-phase assemblages predominate. Where several phases are present in a thin section often only one or two phases infill individual vesicles. The following observations may be made about order of crystallisation:

1. Chlorite, where present, generally forms rims in the vesicles and thus crystallised before the phases it is found associated with.

2. Analcime crystallised before natrolite and thomsonite in OU 38117B, and the latter are not found in the same vesicle.

3. Celadonite where present, rims vesicles and has been observed rimming chlorite-infilled vesicles.

Prehnite and pumpellyte occur in association with laumontite in vesicular lava flows 0.5 km above the base of Zone 1. Chlorite is rarely found in contact with epistilbite in vesicles; where the two phases occur in the same thin section epistilbite commonly infills larger vesicles and chlorite smaller vesicles (Fig. 7.7).

Plagioclase is generally only partially albised; often severely altered plagioclases consist of a rim of fresh oligoclase preserved around an albised interior. Only one case of laumontite replacing plagioclase was observed (OU 38092, S159/913858) a highly altered porphyritic basalt cut by numerous veins of laumontite-calcite and
laumontite-prehnite-calcite). Olivine has been found in only one specimen in Zone 1, but intergrowths of celadonite-calcite and calcite-chlorite-hematite possessing rounded outlines and irregular fractures are interpreted as having replaced olivine.

7.8 White Hill Intrusives

The White Hill intrusive rocks show a similar pattern of alteration to the massive volcanic rocks. In addition, uralitic intergrowths of chlorite and actinolite replacing a prismatic igneous phase (? hornblende) are present in several intrusive bodies, these may be regarded as deuteritic in origin.

7.9 MacKinnon Peak Intrusives

In the MacKinnon Peak Intrusives plagioclase is partially albitised along cracks and cleavage traces, and vesicles are infilled by stilbite, stilbite-laumontite and stilbite-calcite. Veins in the dykes contain stilbite and stilbite-chabazite.

7.10 Veins

The Zone 1 sequence is cut by numerous, irregular minor fractures infilled by simple one- and two-phase vein assemblages. Laumontite is extremely common and assemblages are normally compatible with assemblages present in the host rock.

7.11 Shear zones

The shear zone assemblages have been discussed previously in section 7:1. Prehnite, the characteristic calcium aluminosilicate phase in shear zones, also commonly occurs in void space in strata adjacent to shear zones, e.g. prehnite infills vesicles in an auto-brecciated lava at OU 39193(S159/922880) whereas laumontite is the dominant phase (with minor prehnite) in vesicles 5 m higher in the lava (OU 39194).
ZONE TWO: PREHNITE-PUMPellyITE FACIES

Mineral reconstitution is more advanced in Zone 2 than in Zone 1, and increased with stratigraphic depth throughout Zone 2. Complete albitionization of plagioclase is common, even in granophyric plagioclase in the White Hill Intrusives. Hypersthene, where recognisable, occurs as small relics heavily mantled by chlorite. Magnetite is generally partially altered to hematite and only augite is still predominantly unaffected by metamorphism. Growth of new minerals is now common, not only in void space, but at the expense of phases in the host rock (cf. Zone 1). However new crystal growth rarely crosses original grain boundaries, and igneous textures are preserved in strongly reconstituted rocks (Fig. 7:3). Volcanic glass is replaced not only by chlorite, but also by pumpellyite, epidote and prehnite-calcite. Mineral assemblages in Zone 2 are listed in Table 7:4. Prehnite is the major hydrated calcium aluminosilicate phase present, and occurs in much greater abundance than epidote or pumpellyite.

7.12 Basic volcaniclastic rocks

In coarse-grained clastic rocks prehnite and/or calcite occurs as a mineral cement. Volcanic glass is commonly replaced by coarsely crystalline anomalous-blue birefringent chlorite as is the tuffaceous matrix of pyroclastic rocks. Volcanic glass is less commonly replaced by pumpellyite or epidote. Varying degrees of alteration occur within a single bed, pumiceous fragments are especially susceptible to replacement. Vesicles in clasts are infilled by the same range of assemblages as massive igneous rocks. Rarely crystals of the new-formed minerals extend from the vesicles out into the matrix,
Table 7:4  Mineral Assemblages recorded in Zone 2, Takitimu Mountains

<table>
<thead>
<tr>
<th>Environment</th>
<th>partial mineral assemblage</th>
</tr>
</thead>
<tbody>
<tr>
<td>vesicle</td>
<td>pr-chl</td>
</tr>
<tr>
<td></td>
<td>pr-qz</td>
</tr>
<tr>
<td></td>
<td>pr-pu-chl-qz</td>
</tr>
<tr>
<td></td>
<td>pr-pu-chl-cc-(ep)</td>
</tr>
<tr>
<td></td>
<td>pr-pu-ep-cc-qz</td>
</tr>
<tr>
<td></td>
<td>pr-ep-cc-chl-qz</td>
</tr>
<tr>
<td>groundmass of intrusive rocks</td>
<td>pr-chl</td>
</tr>
<tr>
<td></td>
<td>pr-chl-act-(ep)</td>
</tr>
<tr>
<td></td>
<td>pr-chl-act</td>
</tr>
<tr>
<td></td>
<td>pr-pu-act-chl-ep</td>
</tr>
<tr>
<td></td>
<td>± hm - sph</td>
</tr>
<tr>
<td>replacing pyroxene and amphibole</td>
<td>act</td>
</tr>
<tr>
<td>replacing olivine</td>
<td>pu-chl-and</td>
</tr>
<tr>
<td>replacing plagioclase</td>
<td>ab</td>
</tr>
<tr>
<td></td>
<td>ab-chl-(pu)</td>
</tr>
<tr>
<td></td>
<td>ser</td>
</tr>
<tr>
<td></td>
<td>pr</td>
</tr>
<tr>
<td>mineral cement in tuff breccias</td>
<td>pr</td>
</tr>
<tr>
<td>veins</td>
<td>pr</td>
</tr>
<tr>
<td></td>
<td>pr-qz</td>
</tr>
<tr>
<td></td>
<td>pr-ep-hm-(cc)</td>
</tr>
<tr>
<td></td>
<td>qz-ep-cc</td>
</tr>
<tr>
<td></td>
<td>qz-pr-(ep-hm)</td>
</tr>
<tr>
<td>mudstones</td>
<td>qz-chl</td>
</tr>
<tr>
<td></td>
<td>chl-epi</td>
</tr>
<tr>
<td>interstices of pillow lavas</td>
<td>pr-qz</td>
</tr>
<tr>
<td>shear zones</td>
<td>ep-chl-qz-(sph)</td>
</tr>
<tr>
<td></td>
<td>ep-chl-qz</td>
</tr>
</tbody>
</table>

( ) minor abundance; ab albite; act actinolite; and andradite; cc calcite; chl chlorite; ep epidote; hm hematite; pr prehnite; pu pumpellyite; qz quartz; ser sericite; sph sphene.
proving crystallization occurred after deposition. Plagioclase crystal fragments are strongly albitized and/or replaced by fine-grained aggregates of prehnite or pumpellyite.

7.13 Fine-grained sediments

Alteration of many fine sandstones and argillites in Zone 2 produced a distinct greenish colouration. This is caused by appreciable amounts of chlorite. In Whare Creek alteration of chlorite- and epidote-rich litharenites and hematite-rich mudstone often produces a strikingly red- and green-banded sequence (Fig. 3.20). Less commonly, zones within laminated siltstones are entirely replaced by prehnite or epidote. In thin section the sediments are strongly chloritised, and minor prehnite, pumpellyite and epidote may be present. Prehnite-mottling, reminiscent of zeolite-mottling in Zone 1 occurs in several exposures, notably in the upper and lower gorges of Whare Creek. X-ray diffraction of the clay fraction of four argillites revealed only chlorite.

7.14 Pillow lavas

Lava flows with pillowed margins are less common in Zone 2 than in Zone 1. Tuffaceous mud forming the matrix to a series of ovoid pillows up to 1 m in diameter S159/906882 is strongly reconstituted to prehnite-quartz.

7.15 Massive lava flows

Increasing mineral reconstitution between Zones 1 and 2 is most marked in the massive igneous rocks. Plagioclase is strongly altered in basic volcanic rocks, being replaced by albite, chlorite, prehnite, pumpellyite, sericite and epidote in order of decreasing abundance. Glassy inclusions in plagioclase are commonly replaced by pumpellyite. Glassy areas in the lava
flows are strongly altered to chlorite. Actinolite has replaced primary ferromagnesian minerals. Vesicles are infilled with a variety of 2-, 3-, and rarely 4- phase assemblages. Concentric zonation of the phases is uncommon, and frequently all phases occur in coarse-grained intergrowths (Fig. 7.8). An exception is segregation vesicles in a basaltic lava (OU 39284A, S159/802836) where small cavities in the glassy lining contain pumpellyite and the main void is infilled by a rim of prehnite-quartz and a core of interfingering sheafs of chalcedonic quartz (Fig. 7:10). In a pumiceous breccia (OU 39308A, S159/945726) a sequence of chlorite through pumpellyite and minor prehnite to a quartz core was observed within vesicles.

7.16 White Hill Intrusives

A pattern of alteration similar to the massive lava flows is present in the White Hill Intrusives. Alteration is most marked at the margins of the sills and in the vicinity of fractures. Plagioclase is commonly albitized, hypersthene altered to chlorite and incipient alteration of clinopyroxene has occurred. Olivine is replaced by chlorite and andradite and glassy segregations in microcrystalline and porphyritic bodies are severely altered and contain chlorite and hematite. Hornblende alters to pumpellyite and actinolite.

7.17 Veines

A wide variety of assemblages are recorded as vein fillings in minor fractures. Three-and four-phase assemblages are common (cf one- and two-phase vein-fillings in Zone 1). Commonly the phases within the veins are intergrown, but in OU 38909 veins with a prehnite lining and an epidote core are common. In one specimen (OU 38879C) a pumpellyite vein cuts an irregular pool of chlorite.
Fig 7.8 Vesicle containing quartz–prehnite–pumpellyite–epidote, in porphyritic microdiorite, OU 38354, northern Takitimu Mountains, plane polarised light.

- **quartz**: white, low relief
- **prehnite**: pale grey, high relief
- **pumpellyite**: dark grey
- **epidote**: not visible in field of view

0 0.5 1 mm
in a quartz microdiorite, other evidence for the sequence of crystallization of phases has not been observed.

7.18 Shear zones

Calcium aluminosilicate-rich shear zones have only been recorded from massive igneous rocks in Zone 2. The mineralogy of the zones has been described in section 7.1.

7.19 Prehnite metadomains

In several rocks remote from fracture systems, wholesale replacement of igneous mineralogy by prehnite produces prehnite metadomains. Igneous textures are preserved as in the shear zone metadomains. The significance of the occurrence of prehnite, and not the less hydrous phase epidote which is found in shear zone metadomains is discussed later.
7.20 **Introduction**

Mineral phases were analysed in polished thin section using a JEOL JXA-5A electron microprobe analyser. Analyses of Na, Ca, Si, Al and Sr in zeolites were performed using an accelerating voltage of 15 kv, a specimen current of $1 \times 10^{-8}$ a and a 20µ wide beam. All other measurements were performed using an accelerating voltage of 15 kv, a specimen current of $1.5 \times 10^{-8}$ a and a 1µ wide beam. The Sr-rich heulandite quoted in table 7:9 was reanalysed for Sr using an accelerating voltage of 15 kv, a specimen current of $1.5 \times 10^{-8}$ a and a 1µ beam to minimise errors introduced by low count total. For zeolite phases and celadonite, the electron beam was shifted a distance of 10-15µ between counts to minimise alkali loss. On these minerals repeated counts on a single spot show slight decrease in Na$_2$O and possibly K$_2$O with time and a corresponding increase in CaO, SiO$_2$ and Al$_2$O$_3$.

7.21 **Actinolite**

Actinolite is a common phase in rocks of basic composition in the lower 4 km of Zone 2, and occurs sporadically above this. Actinolite forms principally fibrous overgrowths on, or pseudomorphs after, primary ferromagnesian phenocrysts. Actinolite typically has a deep green colouration comparable with chlorite in the same rocks. In OU 38011 hornblende phenocrysts are surrounded by two discrete actinolite rims of differing composition. The inner rim is wider (~ 50µ) and deep green in colour and the outer colourless and somewhat discontinuous. Five analyses of actinolites of probable metamorphic origin are
presented here (Table 7.5 analyses 2, 3, 4, 5 and 8). The formula used in discussion that follows is $A_{0-1}X_2Y_5Z_8O_{22}(OH)_2$ (Deer et al., 1963) where

\[ A = \text{Na, K, Ca} \]
\[ X = \text{Ca, Na, K, Mg, Fe} \]
\[ Y = \text{Mg, Fe}^{2+}, \text{Fe}^{3+}, \text{Al, Mn, Ti} \]
\[ Z = \text{Si, Al} \]

and the hydroxyl atoms may be replaced by halide ions or oxygen. Total Fe is reported here as FeO, the electron microprobe not distinguishing ferrous and ferric ion. Cationic totals greater than 15.00 for analyses 4, 5 and 8 suggest that appreciable Fe\(^{3+}\) may be present in these actinolites.

The analysed actinolites are similar in composition with Si occupancy of the Z site varying between 7.69 and 7.77 and the principal variations being in Al\(^{IV}\), Fe:Mg and the composition of the X site. The A site is essentially empty for all analyses. The principal differences between the two actinolites in OU 38011 are -

1. Fe:Mg ratios. The deep green inner rim having a composition equivalent to 61\% tremolite, and the pale green outer rim a composition equivalent to 75.5\% tremolite.

2. Ti, which is appreciably higher in the inner rim.

Actinolite occurring in Zone 1 as uralitic pseudomorphs with chlorite replaces primary ferromagnesian phenocrysts in the White Hill Intrusives.

Actinolite is occasionally present in the MacKinnon Peak
Table 7:5 Composition of amphiboles determined by electron microprobe analysis.

<table>
<thead>
<tr>
<th>Oxide</th>
<th>0.01 38011</th>
<th>0.01 38008</th>
<th>0.01 38565</th>
<th>0.01 38002</th>
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<tr>
<td></td>
<td>1</td>
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<tr>
<td>SiO₂</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Al₂O₃</td>
<td></td>
<td></td>
<td></td>
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</tr>
<tr>
<td>TiO₂</td>
<td></td>
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<td></td>
</tr>
<tr>
<td>FeO²</td>
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</tr>
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</tr>
<tr>
<td>Na₂O</td>
<td></td>
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</tr>
<tr>
<td>K₂O</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Total</td>
<td>96.7</td>
<td>98.6</td>
<td>97.7</td>
<td>98.5</td>
</tr>
</tbody>
</table>

| Total Si | 6.39     | 7.69     | 7.77     | 7.75     | 7.77     | 7.94     | 6.51     | 7.71 |
| Total Al | 2.00     | 0.36     | 0.57     | 0.28     | 0.33     | 0.34     | 0.15     | 0.47 |
| Total Ti | 0.14     | 0.04     | 0.00     | 0.03     | 0.05     | 0.00     | 0.08     | 0.07 |
| Total Fe | 1.35     | 1.85     | 1.11     | 2.24     | 1.70     | 2.50     | 1.64     | 1.68 |
| Total Mn | 0.04     | 0.09     | 0.06     | 0.09     | 0.08     | 0.34     | 0.05     | 0.02 |
| Total Mg | 3.35     | 2.83     | 3.43     | 2.93     | 3.42     | 2.21     | 3.05     | 3.20 |
| Total Ca | 1.60     | 1.93     | 1.69     | 1.49     | 1.49     | 1.81     | 1.64     | 1.69 |
| Total Na | 0.70     | 0.09     | 0.06     | 0.40     | 0.43     | 0.05     | 0.57     | 0.23 |
| Total K   | 0.10     | 0.03     | 0.02     | 0.04     | 0.03     | 0.01     | 0.06     | 0.02 |
| Total   | 15.67    | 15.05    | 14.99    | 15.27    | 15.20    | 15.01    | 15.71    | 15.11 |

- Al<sup>17</sup> = 8.00 - Si
- Y = Al<sup>17</sup> + Ti + Fe + Mn = Mg
- (X+4) = Ca + Na = K

1. Core of hornblende crystal in cumulate hornblende clinopyroxenite, White Hill Intrusive, Zone 2.
2. Optically discrete green actinolite rim to crystal in analysis 1.
3. Thin discontinuous band of pale green Fe-actinolite rimming.
4. Actinolite replacing hornblende in basic granophyre, White Hill Intrusive.
5. Actinolite formed in groundmass of 0.01 38008 (see §).
6. Actinolite rim to hornblende in amphibole-hornblende-augite cognate xenolith.
8. Actinolite in groundmass of gabbro.
Intrusives within Zone 1, forming discontinuous fringes on the margins of hornblende phenocrysts. In these instances the actinolite probably formed during late magmatic or deuteric alteration. Actinolite was analysed in one specimen of a MacKinnon Peak dyke (table 7.5, analysis 6). This amphibole contains more tetrahedral Si and less Al, Na and Ti than the metamorphic actinolites. It is also unusually rich in Mn.

On Hallimond's partial triangle all analyses plot in the vicinity of the tremolite-ferroactinolite apex (Fig. 7.9). Comparison with the field occupied by igneous amphiboles from the Takitimu Mountains suggests when actinolite replaces igneous hornblende SiO₂ is consumed and Na₂O and Al₂O₃ are liberated.

7.22 Albite

Albite, replacing plagioclase, is an extremely common constituent of the metamorphosed rocks in the Takitimu Mountains. Elsewhere albite is less common and confined to narrow veins associated with plagioclase phenocrysts. Albitized plagioclase typically has a cloudy texture produced by abundant minute (~1 μ) fluid inclusions, and small flecks of, possibly, hematite. Two complete analyses of albite are presented here, and partial analyses for CaO, Na₂O and K₂O were performed on several samples yielding a compositional range of Ab₉₇An₂Or₁ to Ab₉₄An₂Or₁. No systematic variation in the anorthite content of albite was observed.

7.23 Analcime

Analcime occurs as a vesicle infilling in Zone 1, where it occurs in association with calcium zeolites, chlorite and calcite. Two rocks containing small irregular veins of analcime have also been observed. The analcime is isotropic, colourless and possesses
strong negative relief and weak cleavage. Two samples were selected for analysis (Table 7:7), analysis 1 from a calcium zeolite-calcite assemblage, and analysis 2 from a sodium zeolite assemblage.

The analyses show no significant substitution of calcium for sodium, a feature of analcime group minerals in some other metamorphosed volcanic piles (Surdam, 1969; Seki, 1971, 1973). The analcimes are similar in composition to four specimens of analcime from burial metamorphic rocks quoted by Coombs and Whetten (1967), with an appreciably higher content of K₂O. The analcime analysis have slightly lower sodium content than conventional stoichiometry would imply. This may be partially due to alkali loss during analysis (cf. Tazaki and Hirano, 1973).

7.24 Andradite

Andradite is widespread as a minor mineral phase in more basic igneous rocks in Zone 2. It occurs as small, colourless granules, invariably associated with chlorite, replacing olivine phenocrysts or groundmass phases. Andradite replacing olivine in OU 38011 has a composition very close to end-member andradite. Its composition (Table 7:6, analyses 3) was recalculated using the method of Rickwood (1968). The composition is An₉₇•₂Py₁•₁Sp₀•₂Gr₁•₅ and 97% of the cations can be assigned to the end-member molecules. The majority of acceptable published analyses recalculated by Rickwood possess 95% or more assigned cations.

Recent studies suggest andradite garnet may be a widespread accessory phase in lowgrade metamorphic terrains (Kawachi, 1974).

7.25 Celadonite

Celadonite is restricted to Zone 1 and occurs infrequently
Table 7.6 Compositions of andradite, epidote, prehnite and pumpellyite.

<table>
<thead>
<tr>
<th></th>
<th>Andradite</th>
<th>Epidote</th>
<th>Prehnite</th>
<th>Pumpellyite</th>
</tr>
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<tbody>
<tr>
<td></td>
<td>1</td>
<td>2</td>
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<td>4</td>
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<tr>
<td>SiO₂</td>
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<tr>
<td>TiO₂</td>
<td>0.00</td>
<td>0.03</td>
<td>0.13</td>
<td>0.14</td>
</tr>
<tr>
<td>Al₂O₃</td>
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<td>23.1</td>
<td>23.7</td>
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Cations on the basis of 24(0), 12(0), 1(OH) recalculated to 16 cations

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Table 7.6

(1) OU 38011, andradite, with chlorite replacing olivine in cumulate hornblende clinopyroxenite, White Hill Intrusive, Zone 2.

(2) OU 38866A, epidote, shear zone metadomain containing quartz-epidote-actinolite, Zone 2.

(3) OU 38866 epidote, in altered hornblende diorite containing qz-prehnite-epidote-sphene, 10 cm from OU 38866A.

(4) OU 38340A epidote, within vesicle containing quartz-prehnite-pumpellyte-epidote-calcite, andesite lava flow, Zone 2.

(5) OU 38002 epidote, in groundmass of qz gabbro associated with prehnite, chlorite and actinolite, Zone 2.

(6) OU 38106A prehnite, quartz-prehnite metadomain, Zone 1.

(7) OU 38866 prehnite, as for (3)

(8) OU 38008 prehnite replacing groundmass in basic granophyre containing albite-prehnite-pumpellyte-chlorite.

(9) OU 38963 prehnite, within vesicle, containing quartz-laumontite-prehnite pumpellyte, Zone 1.

(10) Pumpellyite, 38008 as for (8).

(11) OU 38963 pumpellyite as for (9)

+ Total Fe as Fe₂O₃

0.00 below detectability limits
Table 7.7 Composition of sodium-bearing zeolites and albite determined by electron microprobe analysis.

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<th>Stilbite</th>
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Cations on the basis of 96(0) 10(0) 20(0) 72(0) 32(0)

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<tr>
<td>Ba</td>
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<td>0.00</td>
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<tr>
<td>Si/Al</td>
<td>2.43</td>
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</table>

1. OU 38085, analcime, in vesicle containing analcime-epistilbite-laumontite-calcite within andesitic lava flow.
2-4. OU 38117, phases filling vesicles in spilitic lava flow.
5. OU 38999A, stilbite, vesicle in basalt, Mackinnon Peak Intrusives.
6. OU 39011, albite replacing plagioclase, granophyre, White Hill Intrusives.
7. OU 39000, albite replacing groundmass plagioclase, porphyritic basalt.

* total Fe as FeO 0.00 below detectability limits. N.D. not determined
Fig 7.10 Vesicle in basaltic lava containing chalcedonic quartz, prehnite and pumpellyite, OU 39284A, Zone 2.

(a) plane polarised light. Note three zones:
   (i) glassy lining containing pumpellyite
   (ii) outer zone of quartz containing prismatic prehnite and numerous fine inclusions of possibly hematite.
   (iii) inner zone of clear chalcedonic quartz.

(b) crossed Nicol prisms showing habit of chalcedony.
in vesicles and replacing olivine and rarely occurs replacing basic volcanic glass. Individual crystals are small (less than 20μ), and crystal aggregates are common. Celadonite is deep green to blue green in colour, and very weakly pleochroic. An analysis of celadonite associated with chlorite and epistilbite from OU 38977 is presented in Table 7:8. The analysis is very similar to low-Al celadonites from the Murihiku Supergroup in the Hokonui Hills (Boles, 1971). The atomic proportions of Fe-Mg-Al for the celadonite, the coexisting chlorite and the Hokonui celadonites are plotted in Fig. 7.11.

7.26 Chlorite

Chlorite is the most common metamorphic phase in both Zone 1 and Zone 2 in the Takitimu Mountains. The mineral occurs:
in vesicles; as a cement; replacing volcanic glass, plagioclase and hypersthene and in major shear zones and fine-grained sediments.

Two discrete chlorites are present:

[1] pale green or green in colour with anomalous blue birefringence and uniformly length-slow elongation,


The second variety is probably oxidised. The atomic proportions of Al, Mg, Fe in the chlorites are plotted in Fig. 7.12. The analyses show a similar range of Fe:Fe+Mg values to chlorite in the overlying Murihiku Supergroup (Boles, 1971), but contain appreciably more Al (Table 7:8). This is accommodated principally in the tetrahedral site with a corresponding decrease in Si. The Takitimu analyses plot between the field shown by Boles (1971) and green-schist facies chlorites from East Otago (Brown, 1968). The
Fig 7.11 Atomic proportions of iron, magnesium and aluminium in celadonite.

Closed circle celadonite infilling vesicle in OU 38977, Zone 1, Takitimu Mountains
Open circle coexisting chlorite
Squares celadonites from zeolite facies rocks of the Hokonui Hills, data taken from Boles (1971)
Fig 7.12 Atomic proportions of iron, magnesium and aluminium in chlorite.

Closed circles chlorites from Zone 1, Takitimu Mountains
Open circles chlorites from Zone 2, Takitimu Mountains
Triangles chlorites from zeolite facies rocks of the Hokonui Hills, data from Boles (1971)
Dashed line field of greenschist facies chlorites from East Otago, taken from Boles (1971)
Table 7.8 Electron microprobe analyses of chlorite and celadonite, central Takitimu Mountains.

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<td>0.00</td>
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<td>17.89</td>
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Cations on the basis of 28

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N.D. not determined

0.00 below detectability limits

* total Fe as FeO₃.
Fe/(Fe+Mg) ratio of chlorite may be affected by the Fe/(Fe+Mg) ratio of the mineral precursor. A hornblende clinopyroxenite (OU 38011) contains very magnesium-rich chlorite which, with minor andradite, form pseudomorphs after olivine. Here the chlorite replaces an early-stage mineral in a rock formed early in the differentiation series of the White Hill Intrusives. Presumably the parent olivine was forsterite-rich (the coexisting clinopyroxene being of composition Wo₄₇•En₄₄•Fe₉) and some of the Fe²⁺ originally present in the olivine has been oxidised to Fe³⁺ and entered the andradite. The most Fe-rich chlorites are found in the most differentiated rocks, e.g. granophyre segregations in the White Hill Intrusives and rhyodacite lava flows.

Several analyses in Table 7:6 contain octahedral Al apparently in excess of tetrahedral Al. This anomaly is probably due to the presence of Fe in the form of Fe³⁺. Conversion of a portion of FeO to Fe₂O₃ reduces the proportions of all other cations, and this affects the allocation of Al in two ways—

[1] by reducing Si and hence increasing Al⁴⁺ since
\[ Al^{IV} = 8 - Si \]

[2] by reducing total Al, and hence decreasing Al⁶⁺ since
\[ Al^{VI} = Al^{total} - Al^{IV} \]

Boles (1971) considers some chlorites in Triassic rocks of the Murihiku Supergroup in the Hokonui Hills are oxychlorites and the discrepancy between Al⁴⁺ and Al⁶⁺ is even more pronounced in his analyses (Boles, 1971, Table 17). A similar explanation may therefore apply.

7.27 Epidote

Epidote is the dominant hydrated calcium aluminosilicate in shear zone fractures and metadomains of Zone 2. Epidote may
also infill void space in host rock immediately adjacent to the metadomain, but the abundance of epidote is low away from shear zones. Epidote occurs in a wide variety of habits: equigranular aggregates, stumpy prisms, elongate needles and wispy fibres (Fig. 7.13), and is typically yellow in thin section. Epidote compositions in Table 7:6 are listed with total Fe expressed as Fe₂O₃ (although a small amount may be in the form of FeO). The analyses correspond to 26-30% pistacite. Rare examples of colourless epidote rimming yellow epidote occur (OU 33402, Pringle, 1975) in epidote-rich shear zones.

7.28 **Epistilbite**

Epistilbite is the most common mineral phase present within vesicles in Zone 1, and is confined to that environment. A clear, poorly cleaved zeolite it is easily recognisable in thin section by its complex inter-penetration twinning (Fig. 7:7). The importance of epistilbite "a rather common but often mis-identified zeolite" (Galli and Rinaldi, 1974) in low grade metamorphism of volcanic sequences is probably understated by many workers. Only 25 localities (including 7 with identification uncertain) are reported by Galli and Rinaldi (1974) in a review of occurrences. The majority of these are within cavities in basaltic lavas. In contrast to the infrequent accounts of field occurrences epistilbite is a common product of short-term hydrothermal experiments utilising glassy starting materials (Coombs et al., 1959; Liou, 1970). In OU 38110 amygdaloidal epistilbite has replaced a mineral precursor with a radial growth pattern (Fig. 7:14).

7.29 **Heulandite**

Heulandite has a very restricted distribution in the
Fig 7.13 Typical habits of epidote, Zone 2.

(a) as radiating clusters of needles in calcite.
(b) as vein filling.
(c) replacing plagioclase (?).
Fig 7.14 Epistilbite replacing an earlier phase with radial growth pattern possibly chalcedonic quartz, vesicle in porphyritic basalt, Zone 1, OU 39046.

(a) plane polarised light, fine-grained inclusions outline 'ghost' radial habit of mineral precursor.

(b) crossed Nicol prisms, coarsely crystalline prismatic epistilbite transects original grain boundaries.
Table 7:9 Electron microprobe analyses of calcium zeolites, Zone 1, Takitimu Mountains.

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<tr>
<td>Ba</td>
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<td>0.00</td>
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</tr>
<tr>
<td>Total</td>
<td>27.34</td>
<td>27.15</td>
<td>27.30</td>
<td>7.02</td>
</tr>
</tbody>
</table>

| Si : Al                  | 3.00  | 2.69  | 4.07  | 2.93  |

1. O.U. 39097 epistilbite, in vesicle containing epistilbite-heulandite.
2. O.U. 39045 epistilbite, filling vesicle, andesitic lava flow.
3. O.U. 38110 epistilbite, filling vesicle, andesitic lava flow.
4. O.U. 38092 laumontite, associated with prehnite, calcite and chlorite, sheared vesicular andesite.
5. O.U. 38862A laumontite in monomineralic vein cross-cutting diorite containing prehnite-epidote-actinolite-sphene.
6. O.U. 38109 laumontite filling vesicle; permeable top of andesitic lava flow.
8. O.U. 38097 heulandite as for (1).
9. O.U. 39069 heulandite, infilling bubble in heulanditised dacite vitric tuff.
10. O.U. 39999A heulandite in vesicle containing stilbite-heulandite, hornblende basalt dyke, MacKinnon Peak Intrusive.
11. O.U. 38208 yugawaralite, as for (7).

* total Fe as FeO 0.00 below detectability limits  N.D. not determined
Takitimu Mountains occurring in segregation vesicles and replacing acidic glass in tuffs in the Heartbreak Formation. It is suggested under the discussion on fluid chemistry that heulandite may occur in the Takitimu Mountains out of its normal P,T stability field, due to special physiochemical restraints.

The heulandite-clinoptilolite series is perhaps the most chemically complex zeolite group. Three possible substitutional relationships have been suggested:

\[
\begin{align*}
[1] \text{CaAl} &= \text{NaSi} \\
[2] \text{NaAl} &= \text{Si} \\
[3] \text{Ca} &= 2\text{Na}
\end{align*}
\]


In addition strontian, barian and ferric heulandites have been reported (Gerny and Povondra, 1969; Miller and Ghent, 1973; Brown, Catt and Weir, 1969). Attempts at classification of heulandite and clinoptilolite have been based on the ratio Na+K:Ca (Mason and Sands, 1960), and on Si:Al ratios (Boles, 1972). On the basis of Boles' classification three specimens are Si-poor heulandites-Ca and one an Si-poor clinoptilolite-Ca.

Heulandite is the principal potassium-bearing zeolite in the metamorphic system (Table 7:9, analyses 7-10). Heulandite in OU 39069 contains 1.71% SrO, whereas coexisting epistilbite contains only 0.19% SrO, suggesting strontium is also preferentially taken up by heulandite.

7.30 Laumontite

Laumontite is the most important calcium aluminosilicate in Zone 1. It occurs with most other phases in vesicles, typically within a chlorite rim. It is also abundant in veins and
as a mineral cement. Less commonly it replaces acidic glass shards. Laumontite typically occurs as equidimensional aggregates of clear stout prisms, sometimes dusted with an opaque phase, probably hematite, producing a pink tint in hand-specimen. Laumontite contains less Na$_2$O and K$_2$O than the other Ca-zeolites in the Takitimu Mountains, and is close to the ideal composition CaAl$_2$Si$_4$O$_{12}$4H$_2$O (Table 7:9).

7.31 Natrolite and Thomsonite

A single occurrence of natrolite and thomsonite is reported. The phases occur in association with analcime within vesicles in a spilitic lava flow 400 m below the top of the section. Natrolite is present as radiating sheaves of well-cleaved prisms, set in analcime. Its composition is very similar to four representative analyses listed in Deer et al. (1963), (Table 7:7, No.3.).

Thomsonite forms felted masses of radially orientated needles. It is slightly more silica-rich and sodium-poor than analyses quoted in Deer et al. (1963), (Table 7:7, No.4.).

7.32 Prehnite

Prehnite is the most abundant hydrated calcium alumino-silicate in Zone 2. It occurs in small amounts in Zone 1 rocks, increasing in abundance in the Heartbreak Formation, 1 km above Zone 2.

In Zone 2 prehnite infills vesicles and small veins and forms a cement in tuff breccias. Volcanic sandstones commonly show conspicuous prehnite mottling, very similar to the zeolite mottling in Zone 1. In rare cases, e.g. vesicular lava flow (OU 39011, S159/854893), and a volcanic litharenite
(OU 39112, S159/851683) veined prehnite-rich (5-10%) regions coalesce to form metadomains with 50-70% prehnite. Prehnite also occurs in the groundmass of intrusive rocks commonly at points of high surface energy, such as triple junctions between primary grains of differing composition.

Prehnite occurs in two forms:

[1] elongate straight-extinguishing, length-fast tablets in void space (Fig. 7:15),


The chemistry of prehnite is comparatively simple (cf. pumpellyite). The unit cell contains 2 X Ca₂(AlFe)

\((AlSiO₃(OH))\), the principal substitution relationship is

\(Al \rightarrow Fe^{III}\), and so Fe is reported as \(Fe₂O₃\). Substitution in prehnite in the Takitimu sequence occurs up to a composition of

\(Ca₄FeAl₅Si₆O₂₀(OH)₄\). The iron contents are higher than any published values, except those listed by Surdam (1969).

Surdam reported an increase in \(Fe₂O₃\) content of prehnite with stratigraphic depth in the Karmutsen Group of British Columbia, and a wide range of values for each locality. Unlike the Karmutsen Group there is no evidence in the Takitimu sequence for increasing \(Fe₂O₃\) with depth of burial, the most Fe-rich prehnite occurring at the base of Zone 1 (Table 7:6, analysis 9). Only one polished thin section showed widely ranging values for \(Fe₂O₃\) (Table 7:10, analyses 1&2). This specimen is from a fault-zone metadomain in Zone 1, and ranges of 3.5 to 6.5 weight % \(Fe₂O₃\) were recorded from crystals in quartz-prehnite veins.
Table 7.10  Analyses of prehnite from veins in shear zone metadomain, OU 38106, Zone 1.

<table>
<thead>
<tr>
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<th>1</th>
<th>2</th>
</tr>
</thead>
<tbody>
<tr>
<td>SiO₂</td>
<td>43.0</td>
<td>42.8</td>
</tr>
<tr>
<td>Al₂O₃</td>
<td>21.3</td>
<td>19.7</td>
</tr>
<tr>
<td>TiO₂</td>
<td>0.06</td>
<td>0.22</td>
</tr>
<tr>
<td>Fe₂O₃*</td>
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<td>6.33</td>
</tr>
<tr>
<td>MnO</td>
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</tr>
<tr>
<td>MgO</td>
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<td>0.00</td>
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<tr>
<td>CaO</td>
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<tr>
<td>Na₂O</td>
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</tr>
<tr>
<td>K₂O</td>
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<td>0.04</td>
</tr>
<tr>
<td>Total</td>
<td>94.8</td>
<td>95.4</td>
</tr>
</tbody>
</table>

Fig 7.15  Typical tabular habit of prehnite, veins in highly sheared basalt, OU 38106, Zone 1. Location of analyses 1 and 2, Table 7.10 as shown, crossed Nicol prisms.
Individual tablets 0.02-0.05 mm wide are strongly zoned, and two analyses in Table 7:9 are from adjacent prisms (Fig. 7:15). Alumina varies sympathetically with Fe$_2$O$_3$, and the variation is attributed to rapid and metastable crystallization without re-equilibration between the crystal and fluid phases.

7.33 Pumpellyite

Pumpellyite has widespread distribution in Zone 2, but is seldom abundant. Like prehnite it appears in significant amounts at the base of Zone 1. Pumpellyite principally occurs as a vesicle-filling or replacing volcanic glass, and less commonly it is found replacing plagioclase phenocrysts (Fig. 7:16). Two forms of pumpellyite occur in the Takitimu sequence:

1. elongate prisms with marked pleochroism from blue/green to colourless, generally restricted to vesicles,

2. spongy masses of anhedral grains, characterized by green-brown to light green pleochroism, which occur both in vesicles and replacing glass.

The pumpellyite (s.l.) group has complex mineral chemistry. Its formula has been summarised as \( W_{x} X_{y} Y_{z} Z_{2} O_{58-n} (OH)_{n} \) by Passaglia and Gottardi (1973), where

\[
\begin{align*}
W &= Ca, Na, K \\
Y &= Al, Fe^{III}, Ti \\
X &= Mn, Mg, Fe^{II}, Fe^{III}, Al \\
Z &= Al, Si
\end{align*}
\]

The \( W \) site is of seven-fold co-ordination, \( Z \) four-fold, and \( X \) and \( Y \) eight-fold.

In practice substitutions of alkalis for Ca, and Al for Si are very slight, and Ti contents are low. The principal variations are between Mg, Fe$^{II}$, Al$^{III}$, and Fe$^{III}$ in the \( X \) site, and Al and Fe.
in the Y site. On this basis two classifications have been proposed as follows:

Table 7:11 Pumpellyite classification.

<table>
<thead>
<tr>
<th>Passaglia and Gottardi</th>
<th>X</th>
<th>Y</th>
<th>Moore (1971)</th>
</tr>
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<td>pumpellyite - Al</td>
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<tr>
<td>pumpellyite - (Mg)</td>
<td>Mg</td>
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</tr>
<tr>
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<tr>
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<tr>
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<td>Fe\textsuperscript{III}</td>
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</table>

The Passaglia and Gottardi classification is preferred here for the reasons given in their paper. A procedure is developed here to classify electron microprobe analyses of pumpellyite. Since $\text{Fe}^{\text{II}}/\text{Fe}^{\text{III}}$ ratios are not known, the analyses presented here have been recalculated to 16.00 cations. To estimate $\text{Fe}^{\text{III}}/\text{Fe}^{\text{III}}$ it is necessary to make an arbitrary assumption about the number of (OH) units present. Available analyses indicate that this is not justified (Passaglia and Gottardi, 1973). The author suggests the following procedure should be followed for electron microprobe analyses of pumpellyite:

1. recalculate the analysis to 16.00 cations,
2. allocate all Si to the Z site and remedy any deficiency with Al,
3. allocate remaining Al to the Y site,
4. if an excess of Al occurs the excess is placed with Fe, Mg and Mn in the X site,
Fig 7.16 Albite and pumpellyite replacing plagioclase, porphyritic basalt, OU 38980. Large areas of dark (low birefringent) pumpellyite are apparent under crossed Nicol prisms. Note also irregular pools of chlorite in the phenocryst perhaps replacing glassy inclusions.

(a) plane polarised light

(b) crossed Nicol prisms
in the Y site. On this basis two classifications have been proposed as follows:

Table 7.11 Pumpellyite classification.

<table>
<thead>
<tr>
<th>Passaglia and Gottardi (1973)</th>
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<th>Moore (1971)</th>
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</table>

The Passaglia and Gottardi classification is preferred here for the reasons given in their paper. A procedure is developed here to classify electron microprobe analyses of pumpellyite. Since Fe\textsuperscript{II}/Fe\textsuperscript{III} ratios are not known, the analyses presented here have been recalculated to 16.00 cations. To estimate Fe\textsuperscript{III}/Fe\textsuperscript{III} it is necessary to make an arbitrary assumption about the number of (OH)\textsubscript{6} units present. Available analyses indicate that this is not justified (Passaglia and Gottardi, 1973). The author suggests the following procedure should be followed for electron microprobe analyses of pumpellyite:

1. recalculate the analysis to 16.00 cations,
2. allocate all Si to the Z site and remedy any deficiency with Al,
3. allocate remaining Al to the Y site,
4. if an excess of Al occurs the excess is placed with Fe, Mg and Mn in the X site,
[5] if insufficient Al exists to fill the Y-site remedy the deficiency with Fe and group the remaining Fe with Mg in the X site.

The pumpellyites can then be classified as:

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</tr>
<tr>
<td>pumpellyite - (Fe)</td>
<td>Fe</td>
<td>Al</td>
</tr>
<tr>
<td>jülgoldite - (Mg)</td>
<td>Mg</td>
<td>Fe</td>
</tr>
<tr>
<td>jülgoldite - (Fe)</td>
<td>Fe</td>
<td>Fe</td>
</tr>
</tbody>
</table>

If ferrous iron is subsequently determined pumpellyite-(Fe) and jülgoldite - (Fe) can be subdivided into (FeII) and (FeIII) varieties. On this basis the Takitimu pumpellyite may be classified as pumpellyite -(Fe). The composition of the pumpellyites suggests Mg is relatively constant and variation is principally of the form Fe = Al.

7.34 Quartz

Quartz occurs in two forms in the metamorphosed rocks. In veins and shear zones it occurs as coarse-grained aggregates associated with prehnite and epidote. Quartz is found less commonly in vesicles where a chalcedonic variety occurs as radiating and sinuous groups of fibres.

Quartz is not abundant in the andesitic to basaltic lava flows and rock fragments where its low abundance may relate to original bulk composition. In contrast it is common in the rhyodacitic lava flows of the Chimney Peaks Formation and as a constituent of very rare acidic rock fragments found in arenites and rudites.
7.35 **Sphene**

Small granules of highly birefringent sphene are common, replacing primary Fe-Ti oxides and occurring in the groundmass of most of the rocks studied.

7.36 **Stilbite**

Stilbite occurs in veins and vesicles associated with laumontite and heulandite in the upper 3 km of the Takitimu Mountains sequence (Campbell, 1975). Below this, stilbite has only recorded from the MacKinnon Peak Intrusives where it is the predominant zeolite associated with laumontite and calcite in vesicles and chabazite in veins. Stilbite has parallel extinction and occurs as platey aggregates. A single specimen was analyzed and is closer in composition to Ca-endmember stilbite. (Table 7:9).

7.37 **Stilpnomelane**

Stilpnomelane, a common phase in prehnite-pumpellyite and pumpellyite-actinolite facies rocks elsewhere, has not been detected in the central Takitimu Mountains. It is present in minor amounts in rocks from the Princhester fault block (Scott, 1974 and pers. observation).

7.38 **Sulphides (Fe-Cu)/Oxides/Native Copper and alteration products.**

At several localities minor base metal mineralisation is associated with late-stage faulting. Discontinuous veins of pyrite, chalcopyrite-pyrite, chalcopyrite-pyrite-siderite and chalocite occur, and secondary malachite is abundant. The veins cross-cut prehnite-or laumontite-veins and occasionally possess silicified margins.
Pringle (1975) reports skeletal grains of native copper rimmed by malachite in prehnite-laumontite and laumontite veins (OU 33483, S150/961949) in the Princhester fault block. Antiferite, a hydrated copper sulphate, was identified at this locality by Hutton and McCraw (1950).

Jolly (1974) suggests Cu may be concentrated in metamorphic fluids in volcanic/volcaniclastic piles. The low concentration of Cu in the Takitimu rocks may be a factor in limiting copper-bearing phases to the late stage of metamorphism.

7.39 Yugawaralite

Yugawaralite occurs, in association with heulandite, in vesicles within doleritic sills in the Princhester fault block (S150/973999). A poorly-cleaved zeolite, it has a dusty appearance in thin section. Yugawaralite has previously been reported from geothermal areas (Seki and Okumura, 1968; Sameshima, 1969) and from low-grade metamorphic rocks (Miyashiro and Shido, 1970). The yugawaralite from the Takimtu Mountains (Table 7:9, analysis 11) is slightly more silica-rich than yugawaralite from the type locality (Yugawara Hot Spring) or from Shido, Shizuoku Prefecture, Japan (Sameshima, 1969).
Divergent chemical trends arising during production of metadomains.

Marked changes in chemistry have occurred where primary igneous assemblages have been almost totally replaced by hydrated calcium aluminosilicates. Localised redistribution of components is an essential aspect of formation of metadomains (Smith, 1969). CaO, SiO₂ and Al₂O₃ are concentrated in and adjacent to void space in the rock, and the host rock being enriched in Na₂O, FeO and MgO. This produces divergent trends on ACF and ACN diagrams (Fig. 7:17). This pattern of diverging chemical trends was first reported by Smith (1968) from the Ordovician Walli andesite of New South Wales. Subsequently a similar pattern was observed in the Portage Lake lava series of the Keweenawan Peninsula, Michigan, by Jolly and Smith (1972). Here three zones are recognised in which the vesicular tops to basaltic lava flows has been transformed to chlorite, pumpellyite and epidote rich metadomains. In the Takitimu Mountains chlorite metadomains have not been recognised, and there would appear to be a fundamental difference between metadomains formed (1) adjacent to, and (2) remote from, major fractures. This is considered in depth in the following sections.

Metamorphism and Igneous Mineralogy

The original system of igneous phases is essentially simple and consists of plagioclase-clinopyroxene-magnetite-volcanic glass with orthopyroxene, hornblende, olivine and quartz as minor components. Table 7:12 lists the occurrences of the primary phases.
Fig 7.17 Metasomatic trends produced by formation of calcium aluminosilicate metadomains, modified from Smith (1968). Alteration of primary rocks of basaltic composition produces two divergent chemical trends towards calcium enrichment in the metadomains and enrichment in iron and magnesium in the host rock.

Dotted field  unaltered basalts
Crossed field  calcium aluminosilicate metadomains
Stripped field  albite-chlorite enriched host basalts
Trends indicated by solid line

ch  chlorite  
ep  epidote  
pr  prehnite  
pu  pumpellyite
Table 7:12 % of total thin sections containing a given phase and % of that number completely reconstituted under burial metamorphic conditions.

<table>
<thead>
<tr>
<th>Phase</th>
<th>% Present⁺</th>
<th>% Altered</th>
</tr>
</thead>
<tbody>
<tr>
<td>plagioclase</td>
<td>99</td>
<td>72</td>
</tr>
<tr>
<td>augite</td>
<td>79</td>
<td>6</td>
</tr>
<tr>
<td>magnetite</td>
<td>94</td>
<td>11</td>
</tr>
<tr>
<td>glass</td>
<td>75</td>
<td>98</td>
</tr>
<tr>
<td>hornblende</td>
<td>3</td>
<td>12</td>
</tr>
<tr>
<td>orthopyroxene</td>
<td>8</td>
<td>40</td>
</tr>
<tr>
<td>quartz</td>
<td>3</td>
<td>-</td>
</tr>
<tr>
<td>olivine *</td>
<td>8</td>
<td>100</td>
</tr>
<tr>
<td>alkali feldspar</td>
<td>3</td>
<td>40</td>
</tr>
<tr>
<td>biotite</td>
<td>2</td>
<td>-</td>
</tr>
<tr>
<td>apatite</td>
<td>2</td>
<td>-</td>
</tr>
</tbody>
</table>

⁺ on the basis of 200 thin sections.

* inferred from pseudomorphs of chlorite, calcite, magnetite or iddingsite.

The limited distribution of the minor phases severely restricts the role they play in contributing mobile components to the metamorphic system. Of the major phases clinopyroxene and magnetite remain as relics even in some of the most severely altered rocks, and this suggests breakdown of these minerals was not the principal source of mobile components, either. Plagioclase and volcanic glass are the only common phases normally
showing marked alteration. It is suggested that two principal reactions are responsible for liberation of the mobile components,

\[1\] hydration and alteration of glass (including palagonitisation), forming chlorite and montmorillonite,

\[2\] albitization of plagioclase.

7:42 Nature of metamorphism

Several authors in recent years have argued persuasively that low-grade alteration of glass is a solution-microprecipitation phenomenon, e.g. Hay (1962, 1963), Hay and Iijima (1968), Hoover (1968), Sheppard and Gude (1969), Iijima (1971), Surdam (1972). Hay (1962, 1963) states that in the Oligocene-Lower Miocene John Day Formation of Oregon, the boundary between fresh- and zeolitised-tuff is marked by a cavernous zone in which vitric particles are represented by unfilled cavities in a montmorillonite-rich groundmass. Textural evidence suggests glass alteration within Zone 1 in the Takitimu Mountains is integrally related to a fluid phase. In pillow lavas and glassy rock fragments, alteration is most severe at the particle margins, and decreases rapidly inwards. Commonly a sharp surface, interpreted as a relict solution interface, separates relatively fresh glass from a phyllosilicate-or palagonite-rim. In some cases the interface is marked by a narrow zeolite band. Similarly the alteration of plagioclase proceeds along grain boundaries and cracks, and relict fresh plagioclase occurs immediately adjacent to sharply-outlined secondary albite zones. Less direct evidence for the importance of a fluid phase is (1) a strong correlation between original porosity and degree of mineralogical reconstitution during metamorphism, and (2) the rapid decrease in degree of alteration
of impermeable dykes away from dyke margins, joints and fractures. Quartz within shear zones in Zone 2 contains abundant small (<1 μ) fluid inclusions.

The chemical nature of solutions involved in the reactions can be gauged from experimental studies, and studies of the alteration of Quaternary volcanic deposits. Hay and Iijima (1968) in the study of the Pleistocene tephra of Oahu concluded that conditions of high pH and salinity in the metamorphic fluid were essential for palagonitisation.

Surdam (1973) showed that alkaline solutions at 80°C dissolve CaO, SiO₂ and Al₂O₃ from basaltic glass. On cooling, the solutions form gels which crystallize to phillipsite on reheating. Hay (1966) states that the reaction of rhyolitic glass is more rapid in a saline lacustrine environment than in the less saline marine environment. The evidence implies fluids causing alteration of volcanic glass are characterized by high pH and salinity.

7.43 Origin of the metamorphic fluids

There are two major sources of H₂O in the metamorphic fluids in a submarine volcanic pile, (1) entrapped seawater and (2) solutions of a deep-seated origin derived from dehydration at depth. The fluid compositions may undergo modification in two ways, by interaction with the host rock and by addition of volatiles from residual fluid associated with intrusive bodies. If clay minerals act as semi-permeable membranes during expulsion of water by compaction, the salinity of fluids may also be increased.

Seawater has been considered the principal source of the metamorphic fluid phase in several studies (Brown and Thayer, 1963; Surdam, 1973). The Takitimu Group has accumulated in a
submarine environment and shows an alternation of porous and non-porous horizons which places restrictions on the migration of pore fluids. However, the presence of entrapped seawater should not be regarded as an essential factor for low grade metamorphism of volcanic rocks, because the Precambrian Keeweenawan lavas of Michigan show signs of a burial metamorphic event closely resembling that described in the present study and are considered to be subaerial in origin (Jolly and Smith, 1972). Relatively high rock/fluid ratios in volcanic/volcaniclastic systems probably ensure that fluid compositions are buffered by the rock chemistry, and so initial fluid composition is not of crucial importance.

Fyfe (1974) emphasises the importance of deep-seated solutions as a source of fluid during low grade metamorphism. He estimates for every km$^3$ of rock under near-surface conditions, a volume of 0.5 km$^3$ of water will be lost by the time the rock equilibrates under greenschist facies conditions. This fluid, during its passage to the surface, may react with and modify mineral assemblages at higher levels. Jolly (1972) has applied this relationship, i.e. dehydration reactions at depth, with complimentary hydration reactions at shallow levels, to the Keeweenawan basaltic lavas of Michigan.

The White Hill intrusive bodies contain abundant late-stage aplite and pegmatoidal granophyre segregations, the latter often forming approximately 10% by area of the thickest bodies, implying considerable differentiation of the melt. Exchange of volatiles between late-stage residual melt and groundwater may have occurred and increased salinity of the fluid. To assess the relative importance of the various influences on the initial composition of the metamorphic fluid is beyond the
Evolution of the metamorphic fluids

The following evidence suggests chlorite is often the first phase to form from the metamorphic fluids:

1. Where present in vesicles chlorite occurs as a cavity lining or rim,
2. Chlorite often occurs alone at the sides of epidote-and prehnite-infilled fractures.

With increasing time and depth of burial, the metamorphic fluid changes in composition by virtue of interaction with the host rock. The major steps are:

1. Reaction of volcanic glass to form chlorite and montmorillonite liberating $\text{CaO, Na}_2\text{O, K}_2\text{O, SiO}_2$ and $\text{Al}_2\text{O}_3$, and consuming $\text{FeO}$ and $\text{MgO}$. Hay and Iijima (1968) have shown that a similar solution and hydration of glass in vitric tuffs of the Pleistocene Honolulu series of Hawaii liberated one third of the $\text{SiO}_2$, half of the $\text{Al}_2\text{O}_3$ and three quarters of the $\text{CaO}$ and $\text{Na}_2\text{O}$ originally present in the volcanic glass.

2. Reaction of plagioclase to form albite consuming $\text{Na}_2\text{O}$ and $\text{SiO}_2$ and liberating more $\text{CaO}$.

The chemical trend is in accordance with data recently available from the Deep Sea Drilling Programme (Sayles and Manheim, 1975). Pore fluids from two drill holes through rapidly deposited terrigenous sediment (0-660 m site 102, and 0-148 m site 147) show a rapid decrease in ($\text{Mg}^{2+}$) and increase in ($\text{Ca}^{2+}$) with depth.

The metamorphic assemblages of the Pliocene-Miocene Esmeralda Formation of Nevada may reflect a similar pattern of
increasing alkalinity and salinity with time and depth. Moiola (1970) suggests montmorillonite formed initially prior to significant glass dissolution under conditions of low pH. Formation of phillipsite at shallow depths then occurred under inferred conditions of low $a_{SiO_2}/a_{Al_2O_3}$, and $a_{Na^+}$ and $a_{K^+}$, and these phases are succeeded below 265 m (865') by clinoptilolite-opal forming under conditions of high $a_{SiO_2}/a_{Al_2O_3}$, $a_{Na^+}$ and $a_{K^+}$.

7:45 Palagonitisation and metamorphism

Palagonite rinds on basaltic pillow lavas are common in the Takitimu Group. The process of palagonitisation is initiated soon after deposition of subaqueous material, and proceeds at rates of up to 50 μ/1000γ (Moore, 1966). Furnes (1975) suggests reaction commences as soon as water gains access to the system. Furnes (1975) also shows there is a rapid acceleration in the rate of palagonitisation at between 40°C and 70°C. Several studies have also revealed that major chemical changes occur during the transformation of basaltic glass to palagonite.

The evidence from the literature suggests:

[1] Palagonitisation in volcanic piles is initiated before low grade metamorphism immediately after deposition (Moore, 1966; Furnes, 1975) and,


There is therefore, a strong possibility that palagonitisation may, in part, contribute the saline fluid which plays an integral part in low grade metamorphism. As a result the author has investigated the palagonitisation of a basaltic pillow lava (OU 38821) from the Telford Ridge to determine the nature of the chemical changes.
produced in the glassy pillow margin. The location of the analyses in Table 7:13 are given in Fig. 7:18. The analyses suggest that FeO and MgO are enriched in the palagonite relative to the pillow interior and sideromelane and that SiO₂, Na₂O, Al₂O₃ and to some extent CaO are depleted. It is generally thought that palagonitisation is isovolumetric (Moore, 1966; Hay and Iijima, 1968). Under these circumstances a mass balance equation can be used to evaluate the percentage changes during reaction. An equation has been derived from the analyses in Table 7:13 using as an approximation the specific gravities for palagonite and sideromelane measured by Hay and Iijima (1968). Since the sideromelane may itself be partially altered a computation was also made from the analysis of the pillow interior. This analysis probably reflects most closely the original composition of the magma because (1) its composition is reasonably representative of the analysed aphyric basalts, although low in K₂O, (2) it contains less than 2% phenocrysts, (3) it is not extremely hydrated, having a loss of ignition value of only 2.34%. The isovolumetric transformation of the glassy basaltic pillow rim into palagonite results in a net loss of approximately 2/5 of SiO₂; 1/2 of Al₂O₃; 1/5 of CaO; and 4/5 of Na₂O present in the original glass. A small amount of this material is localised in the zeolite rims to the sideromelane fragments, and as zeolite infilling the radial fractures. However, the total volume of zeolite in the immediate vicinity of the palagonite is slight (see Fig. 7:24), and so a considerable amount of material is lost from the system (by implication) into a pore fluid. With compositional changes of this magnitude relatively small amounts of palagonitisation may greatly increase the salinity of the pore fluids.
Table 7:13  Chemical compositions for pillow interior, sideromelane and palagonite basaltic pillow lava, O.U.38821 and computed composition changes for the reactions sideromelane-palagonite and pillow interior-palagonite.

<table>
<thead>
<tr>
<th></th>
<th>1.</th>
<th>2.</th>
<th>3.</th>
<th>4.</th>
<th>5.</th>
<th>6.</th>
<th>7.</th>
<th>8.</th>
<th>9.</th>
</tr>
</thead>
<tbody>
<tr>
<td>SiO₂</td>
<td>51.59</td>
<td>35.5</td>
<td>53.9</td>
<td>40.6</td>
<td>40.9</td>
<td>-44.5%</td>
<td>-40.8</td>
<td></td>
<td></td>
</tr>
<tr>
<td>TiO₂</td>
<td>1.18</td>
<td>1.34</td>
<td>0.60</td>
<td>0.61</td>
<td>0.88</td>
<td>-42.3%</td>
<td>-52.4%</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Al₂O₃</td>
<td>14.54</td>
<td>12.6</td>
<td>12.5</td>
<td>10.9</td>
<td>10.1</td>
<td>-37.3%</td>
<td>-46.2%</td>
<td></td>
<td></td>
</tr>
<tr>
<td>FeO*</td>
<td>13.17</td>
<td>12.3</td>
<td>15.0</td>
<td>20.3</td>
<td>19.8</td>
<td>+17.8%</td>
<td>+13.7%</td>
<td></td>
<td></td>
</tr>
<tr>
<td>MnO</td>
<td>0.21</td>
<td>0.31</td>
<td>0.30</td>
<td>0.43</td>
<td>0.44</td>
<td>+7.2%</td>
<td>+45.9%</td>
<td></td>
<td></td>
</tr>
<tr>
<td>MgO</td>
<td>4.89</td>
<td>6.26</td>
<td>6.26</td>
<td>9.2</td>
<td>10.2</td>
<td>+16.0%</td>
<td>+48.5%</td>
<td></td>
<td></td>
</tr>
<tr>
<td>CaO</td>
<td>6.38</td>
<td>7.58</td>
<td>8.50</td>
<td>7.21</td>
<td>6.73</td>
<td>-35.6%</td>
<td>-19.1%</td>
<td>6.04</td>
<td>5.89</td>
</tr>
<tr>
<td>Na₂O</td>
<td>4.45</td>
<td>4.25</td>
<td>3.58</td>
<td>0.60</td>
<td>0.51</td>
<td>-57.6%</td>
<td>-89.1%</td>
<td>8.12</td>
<td>8.36</td>
</tr>
<tr>
<td>K₂O</td>
<td>0.33</td>
<td>0.50</td>
<td>0.36</td>
<td>0.94</td>
<td>0.88</td>
<td>+57.5% +105.4%</td>
<td>0.37</td>
<td>0.22</td>
<td></td>
</tr>
<tr>
<td>P₂O₅</td>
<td>0.27</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>Total</td>
<td>99.3</td>
<td>100.6</td>
<td>99.0</td>
<td>90.8</td>
<td>90.4</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

* includes loss on ignition 2.34

1. X-ray fluorescence analysis of aphyric core to pillow lava, O.U. 38821A
2. Electron microprobe analyses of relict sideromelane in rim of pillow lava, O.U. 38821A
3. Electron microprobe analyses of palagonite adjacent to relict sideromelane in rim of pillow lava O.U. 38821A
4. Percentage changes, by weight, in composition accompanying transformation of sideromelane to palagonite based on average composition of sideromelane from analyses 2 and 3 and average composition of palagonite from analyses 4 and 5, assuming isovolumetric replacement and specific gravity as above.
5. Percentage changes by weight of composition during the transformation from pillow core to average palagonite composition, conditions as for (6).
6. Partial analyses of zeolite rim to sideromelane fragment.
Fig 7.18 Palagonite rim to basaltic pillow lava. Large sideromelane relics in top half of picture are rimmed by zeolite and set in palagonite (light grey). Isolated varioles (dark grey) are present in lower half of picture, OU 38821, southern Takitimu Mountains.
The Takitimu terrain is regarded to be the main source area for the detrital component of the overlying Middle-Upper Permian Productus Creek Group (Landis, 1969; Force, 1972; Begg, 1974). Included in the volcanic detrital component are epidote-pumpellyite- and actinolite-bearing metavolcanic rock fragments and detrital epidote. Landis (1969) states that the epidote- and pumpellyite-bearing metavolcanic rock fragments were probably derived from the western Takitimu Mountains. If this is correct, then the lower Takitimu Group had already undergone mineral reconstitution under prehnite-pumpellyite facies conditions, uplift and erosion by the Upper Permian. Evidence given in chapters 3 to 6 suggests the Lower-Middle Permian was also a period of extensive intrusive activity in the Takitimu terrain, during which emplacement of the MacKinnon Peak Intrusives and early members of the White Hill Intrusive suite occurred. A model is therefore proposed involving rapid and overlapping volcanism, intrusive activity, and burial in Lower Permian time, with uplift and erosion of at least the older rocks of the volcanic pile in Middle to Upper Permian time. The contact between the Takitimu Group and the Productus Creek Group is poorly exposed, but Begg (1974) considers the contact is a low-angle unconformity.
EFFECT OF PENECONTemporaneous Intrusive Activity
ON METAMORPHISM.

The effects of normal burial metamorphism and intrusive phenomena are tightly interwoven in the Takitimu Mountains. The White Hill Intrusives comprise 20% by area of the exposed sequence in Zone 2, and approximately 15% in Zone 1. The bodies are lensoid and emplaced almost exclusively along bedding planes. The margins of the larger bodies in the western Takitimu Mountains are marked by extensive shearing and fracturing of both host and intrusive rock, and these and other fractures commonly form the site of extensive development of metadomains. The larger bodies are either multiple intrusions, or have differentiated in situ. The metamorphic system can be pictured as transitional between normal burial metamorphism of the 'Taringatura type', where assemblages have reacted with fluids derived from seawater under a moderate geothermal gradient remote from external hydrothermal influences related to igneous activity, and metamorphism of the 'Tanzawa type', where zonation can be directly related to the distribution of large intrusive masses up to 15 km in width (Seki et al., 1969). In the Takitimu Mountains the intrusive bodies are small by comparison with Tanzawa, and metamorphic aureoles are not well developed. The regular zonation about large intrusive bodies intruded under shallow conditions will probably be obscured when a comparable volume of magma invades a volcanic pile as numerous small masses emplaced at depths up to 10-12 km, where under a moderate geothermal gradient of 25°/km temperatures of 250-300° could already have been reached, and re-equilibration of the igneous assemblages already commenced. It is difficult, and perhaps misleading, to delineate the products of purely hydrothermal
metamorphism under these conditions of overlapping volcanism, rapid burial and shallow-level plutonism. Franks (1974) describes a 2000 m sequence of Lower Ordovician volcanogenic sediments comprising the New Bay Formation in Central Newfoundland. The sequence is intruded by numerous basic sills and dykes comprising between 21 and 31% of the exposure of the formation. The formation is overlain by the Lawrence Head Volcanics, a 1000 m sequence of pillow lavas, and Franks cites evidence to suggest that the intrusive bodies in the New Bay Formation have acted as feeders for the Lawrence Head volcanics. The New Bay Formation has been metamorphosed to prehnite-pumpellyite facies conditions, only a narrow (5-10 cm) hornfels zone is present adjacent to the intrusions, but in at least one locality prehnite increased in abundance within tuffaceous sandstones approaching the intrusion. The major effect of shallow level plutonism is addition of heat to the system, and this is considered in more detail in the section 7.52. However, it is possible to list other probable effects of the pervasive invasion of the sequence of strata by the swarm of intrusive bodies, such as formation of channel-ways along fault systems (providing increased access for hydrothermal fluids) and interaction of possibly alkali-rich late stage residual magmatic fluids and the metamorphic fluid.
7.46 Introduction

The metamorphic assemblages developed in the Takitimu Mountains were governed by a complex interplay of factors. Most of these factors emphasize the crucial role played by the fluid phase during metamorphism. The influences considered below are:

[1] temperature and load pressure,
[2] bulk composition,
[3] varying permeability,
[4] varying ratios of $P_{H_2O}/P_{load}$
[5] varying fluid chemistry and salinity,

Zone 1 of the Takitimu Mountains may be correlated with the zeolite facies, and Zone 2 with the prehnite-pumpellyite facies. Zone 1 stratigraphically overlies Zone 2, and thus mineral assemblages become increasingly less hydrous with depth, presumably in response to increasing temperature.

7.47 Temperature and load pressure

Since the overlying Productus Creek Group contains prehnite-pumpellyite grade Takitimu detritus, it is reasonable to assume the Takitimu Group underwent low grade metamorphism at load pressures generated solely by the weight of overlying Takitimu Group material. This is equivalent to a maximum burial depth of 5.6 km for the base of Zone 1, and 14.4 km for the base of the section. Using an average wet density of 2.64 recorded by Hatherton and Leopard (1964) for a volcanic sandstone from the Takitimu Group, this corresponds to load pressures of 2.1 kb and 5.4 kb respectively. Maximum and minimum temperatures may be
The evidence suggests a geothermal gradient in the range 27.5°-50°/km in the Takitimu Mountains. As discussed later the author suggests short-lived elevated thermal gradients may have occurred, perhaps in the upper portion of this range, superimposed on a lower geothermal gradient.

7.48 Bulk composition

Within the Takitimu Mountains there are three broad compositional groups each of which have developed a distinctive metamorphic mineralogy. These are as follows:

[1] Andesite-basalt. In this group hydrated calcium aluminosilicates are abundant, hematite is a common accessory mineral, and quartz rare.

[2] Rhyodacite. The metamorphic mineralogy of rhyodacite lavas and rock fragments is dominated by quartz and albite, and they contain little or no hydrated calcium aluminosilicate.

[3] Mudstone-argillite. In the fine-grained sediments...
chlorite and chlorite-montmorillonite are abundant and quartz and the hydrated calcium aluminosilicates rare.

The contrast between [1] and [2] may be seen on the scale of a thin section in the rare volcanogenic sediments containing acidic and basic rock fragments (OU 39159). The contrast between [1] and [3] may be observed in a single sedimentary unit where volcanogenic litharenite containing prehnite- and pumpellyite-bearing clasts grades into chlorite-rich mudstone. These differences are attributed to difference in bulk composition of the host rock.

Typical assemblages of the prehnite-pumpellyite and pumpellyite-actinolite facies are characteristically developed in rocks of basic to intermediate igneous composition. The Takitimu Mountains provide an excellent section for study of prehnite-pumpellyite assemblages with its succession of massive volcanic, volcaniclastic and intrusive rocks of predominantly andesitic to basaltic composition. Within this compositional group the effect of slight variations in bulk composition on mineral distribution is examined in more detail in the following sections.

Difficulties in graphically depicting prehnite- and pumpellyite-bearing assemblages have been detailed by Coombs et al. (1970) and Zen (1974). Recognizing the limitations that $\text{Al}_2\text{O}_3$, $\text{CaO}, \text{MgO}, \text{FeO}, \text{Fe}_2\text{O}_3$ and possibly $\text{CO}_2$ and $\text{H}_2\text{O}$ must be regarded as independent variables, Coombs et al. plot on an $\text{Al}_2\text{O}_3-\text{CaO}-(\text{Fe, Mg})$ diagram partial mineral assemblages of prehnite-pumpellyite grade from the Appalachian Belt of Northern Maine. An average basaltic composition plots close to the prehnite-chlorite join (Fig. 7:19). Only slight compositional variation is
Fig 7.19 Changes in the bulk composition of a metamorphic assemblage produced by 
1. segregation or addition of chemical components, 
2. relict phases.

A residual alumina after formation of micas, albite and K-feldspar 
C CaO 
FeO MgO MgO and total iron expressed as FeO 
A average basalt composition taken from Coombs et al. (1970)

Trend A, produced by the removal of calcium and aluminium to form metadomains, and trend B, produced by removal of aluminium or addition of Na₂O and K₂O to form albite, K-feldspar or micas, both displace bulk composition across the calcite-chlorite tie line.

Trend C, produced by the presence of roughly equal proportions of relict magnetite and clinopyroxene, may displace bulk composition across the prehnite-chlorite or pumpellyite-chlorite tie lines.
necessary to displace a rock of this composition from the prehnite-calcite-chlorite field to the prehnite-pumpellyite-chlorite field.

Microprobe data for the Takitimu Sequence suggests that prehnite contains substantial Fe and pumpellyite contains Fe$^{II}$, Fe$^{III}$, and Mg, and these three components must be regarded as independent variables. Zen (1974) considered the 8-phase system actinolite, chlorite, prehnite, pumpellyite, epidote, hematite, stilpnomelane and calcite, and the components FeO, MgO, Al$_2$O$_3$, Fe$_2$O$_3$, CaO, H$_2$O, CO$_2$ (excess quartz present). A 6-phase assemblage with externally controlled chemical potentials of water and carbon dioxide (boundary value) must be interpreted as univariant. Zen considers this unlikely and prefers the alternative interpretation that µCO$_2$ was determined by bulk rock composition.

In the Takitimu Mountains quartz is far from ubiquitous, and so the analogous system to that above is quartz and the 8 phases listed above, and the components FeO, MgO, Al$_2$O$_3$, Fe$_2$O$_3$, CaO, H$_2$O, CO$_2$ and SiO$_2$. With 8 components a 6-phase assemblage is divariant, which means CO$_2$ may be a boundary-value component. It is also important to realise that even in isochemical reconstitution of these rocks, the chemical composition of the metamorphic assemblage is not that of the bulk rock, but that of the bulk rock minus the relict phases. In the Takitimu sequence where clinopyroxene and magnetite are commonly unaltered, the metamorphic assemblage is rich in CaO, Na$_2$O, and Al$_2$O$_3$ and depleted in FeO and MgO relative to the host rock.

7.49 Permeability/Availability of fluids

The importance of a fluid phase to low-grade metamorphism is emphasised throughout this study. Permeability principally

affects the degree of alteration of the parent rock. Two points illustrate this:

[1] Laumontite or prehnite cements are common in coarse litharenites and breccias but rare in fine litharenites and siltstone.

[2] Fresh plagioclase is more common in siltstones and fine litharenites than in coarse litharenites.

Another possible case where permeability may have affected the development of a metamorphic assemblage is the rhydacitic tuffs. There are two possible interpretations for the different assemblages; (1) possible differences in original glass composition. If the zeolite chemistry reflects the chemistry of the glass precursor, then the heulandite-rich tuff (OU 39069) may have been rhydacitic or rhyolitic in composition, and the laumontite-rich tuff dacitic. (2) If present permeability is a reflection of original porosity, then the pore fluid in OU 38945 may have equilibrated with entrapped water which has been in contact with andesitic rock fragments in the surrounding strata, while the pore fluid in OU 39046 remains in a stagnant closed-system environment in contact only with rhydacitic glass. The laumontite-tuff (OU 38945) occurs in a series of laumontite-cemented litharenites and is still highly permeable, readily absorbing moisture on even newly-broken surfaces. Tuff OU 39069 is now completely impermeable.

If the statement made earlier is the case, the fluid in OU 38945 may have been buffered by Ca-rich andesitic glass in the rocks surrounding the tuff, and that in OU 39069 by K-rich acidic glass. Under these circumstances Ca-rich laumontite is more likely to form in OU 38945 and K-bearing heulandite in OU 39046.
The consistent development of differing assemblages in shear zone metadomains compared with host rock metadomains requires further explanation. Contrasting assemblages occur in both zones, and within a single lithology on the scale of an outcrop and cannot be explained easily by variations in $P_{\text{load}}$, $T$, or bulk composition. To simplify further discussion, metadomains developed in and adjacent to vesicular flow tops and void space in permeable strata will be referred to as Type A, and metadomains developed in and adjacent to shear zones as Type B. Metadomain assemblages are summarized below.

<table>
<thead>
<tr>
<th>Metadomains</th>
<th>Type A</th>
<th>Type B</th>
</tr>
</thead>
<tbody>
<tr>
<td>Zone 1</td>
<td>laumontite + calcite</td>
<td>prehnite-quartz</td>
</tr>
<tr>
<td>Zone 2</td>
<td>prehnite</td>
<td>epidote-quartz</td>
</tr>
</tbody>
</table>

The phases may be linked in the equations:

$\text{CaAl}_2\text{Si}_4\text{O}_{12} \cdot 4\text{H}_2\text{O} + \text{Ca}^{2+} + 2(\text{OH})^- = \text{Ca}_2\text{Al}_2\text{Si}_3\text{O}_{10}(\text{OH})_2 + \text{SiO}_2 + 4\text{H}_2\text{O}$

$\text{laumontite} \quad \text{prehnite} \quad \text{quartz}$

$\text{Ca}_2\text{Al}_2\text{Si}_3\text{O}_{10}(\text{OH})_2 + \text{Fe}^{3+} + 3\text{OH}^- = \text{Ca}_2(\text{Al}_2\text{Fe})\text{Si}_3\text{O}_{12}(\text{OH})_2 + 2\text{H}_2\text{O}$

$\text{prehnite} \quad \text{epidote}$

Both reactions involve dehydration of the Type A assemblage. The occurrence of less hydrous assemblages within fracture systems is the reverse of sequences formed during retrogressive metamorphism where successive generations of veins contain progressively more hydrous phases formed as metamorphism wanes, e.g., in mid-Canterbury, New Zealand, certain greywackes of the Torlesse Supergroup contain prehnite-pumpellyite veins crosscut by laumontite-infilled fractures in turn veined by stilbite, (Coombs et al., 1959).
The presence of less hydrous assemblages within fracture systems suggests perhaps different values of $P_{H_2O}/P_{load}$ existed during metamorphism. Two factors may have produced reduced $P_{H_2O}$:

[1] Varying $\mu_{H_2O}/\mu_{CO_2}$. Zen's pioneering work led to recognition of the role of varying $\mu_{CO_2}/\mu_{H_2O}$ in low grade metamorphism. Subsequently various authors have postulated contrasts in $\mu_{CO_2}/\mu_{H_2O}$ to explain contrasting metamorphic parageneses (Coombs et al., 1970; Surdam, 1973; Ghent and Miller, 1974; Pepezić, 1974; Richter and Roy, 1974). Evidence does not favour varying $\mu_{CO_2}/\mu_{H_2O}$ in the Takitimu section. The petrographic evidence, especially the absence of clay-carbonate assemblages, suggests a consistently low value of $\mu_{CO_2}$ for the metamorphic fluids.

[2] Varying $P_{fluid}/P_{total}$. At shallow depths of burial, shear zones and major fracture systems may be open to surficial levels in the pile, and therefore $P_{H_2O}$ may approximate hydrostatic pressure, i.e. $P_{H_2O} \approx 1/3 P_{load}$. In contrast in a vesicle cavity of an impermeable horizon remote from such fractures, calcium aluminosilicates form in an environment where $P_{H_2O}$ closely approximates $P_{load}$, i.e. $P_{H_2O} \approx P_{load}$.

It is proposed that relatively high $P_{H_2O}$ in the essentially 'closed system' Type A metadomains favours formation of the more hydrous phase under conditions where the less hydrous phase is stable in the 'open system' Type B metadomains. Evidence from drilling programmes indicates that conditions of reduced fluid pressure exist in nature. In Miocene–Eocene sediments in the San Joaquin Valley laumontite appears at a depth of 2900 m and is widespread at depths below 3200 m. At 3200 m the recorded temper-
ature was 200°C and $P_{H_2O} = 0.3$ kb, less than half the value of the lithostatic pressure at that depth (Castano and Sparks, 1974).

Shear zone metadomains replacing wall-rock in Zone 2 are largely confined to volcanic flows and intrusive horizons and this can be explained in terms of a strength restriction placed on the relationship $P_{H_2O} \approx 1/3 P_{load}$. To maintain an open fissure at reduced fluid pressure the strength of the rock must be sufficient to prevent failure and consequent closure of the fissure. The approximate relationship is therefore:

$$P_{load} = P_{H_2O} + e \text{ wallrock}$$

where $e$ is the strength of the wallrock under these conditions.

We can modify this to $P_{H_2O}/P_{load} = 1 - e \text{ wallrock}$ to derive Table 7.14.

Table 7.14 Relationship between load pressure, crushing strength and minimum values of $P_{H_2O}/P_{load}$.

<table>
<thead>
<tr>
<th>$P_{load}$</th>
<th>0.5</th>
<th>1</th>
<th>2</th>
<th>3</th>
<th>4</th>
<th>5</th>
<th>6</th>
<th>10</th>
</tr>
</thead>
<tbody>
<tr>
<td>$e$</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>0.1</td>
<td>0.80</td>
<td>0.90</td>
<td>0.95</td>
<td>0.97</td>
<td>0.97</td>
<td>0.98</td>
<td>0.99</td>
<td>0.99</td>
</tr>
<tr>
<td>0.5</td>
<td>0.00</td>
<td>0.50</td>
<td>0.75</td>
<td>0.84</td>
<td>0.88</td>
<td>0.90</td>
<td>0.92</td>
<td>0.95</td>
</tr>
<tr>
<td>1</td>
<td>0.00</td>
<td>0.00</td>
<td>0.50</td>
<td>0.66</td>
<td>0.75</td>
<td>0.80</td>
<td>0.84</td>
<td>0.90</td>
</tr>
<tr>
<td>2</td>
<td>0.00</td>
<td>0.00</td>
<td>0.00</td>
<td>0.33</td>
<td>0.50</td>
<td>0.60</td>
<td>0.66</td>
<td>0.80</td>
</tr>
</tbody>
</table>

conditions under which rock strength will permit $P_{H_2O} = 1/3 P_{load}$ in open fissures.
It is difficult to choose which measurable value of strength of a rock is applicable under the conditions encountered. As an order of magnitude approximation we can assume $e$ to be the value of crushing strength for the rock. The following crushing strengths are listed by Handin (1966).

Table 7:15 Crushing strengths for selected rock types.

<table>
<thead>
<tr>
<th>material</th>
<th>crushing strength (bars)</th>
</tr>
</thead>
<tbody>
<tr>
<td>hypersthene andesite Idaho</td>
<td>1290-1330</td>
</tr>
<tr>
<td>basalt Oregon</td>
<td>1690-2190</td>
</tr>
<tr>
<td>granodiorite Australia</td>
<td>1270</td>
</tr>
<tr>
<td>monzonite Washington</td>
<td>1750</td>
</tr>
<tr>
<td>granite Washington</td>
<td>1490</td>
</tr>
<tr>
<td>lithic tuff Oregon</td>
<td>35</td>
</tr>
<tr>
<td>siltstone California</td>
<td>250</td>
</tr>
<tr>
<td>subgreywacke India</td>
<td>900</td>
</tr>
</tbody>
</table>

Clearly massive volcanic and intrusive rocks have considerably greater strength ($1.5 - 2$ kb) than fresh unmetamorphosed sedimentary rocks ($0.05 - 1$ kb). Thus, while fissures maintaining low $P_{H_2O}$ may remain open in clastic strata buried under zeolite facies
conditions, this is unlikely under conditions of prehnite-
pumpellyite or higher grade metamorphism. Fissures with low
\( P_{H_2O} \) (approximating to hydrostatic pressure) may however persist
in massive volcanic and intrusive rocks during prehnite-pumpellyite
metamorphism.

7.51 Fluid chemistry

The activity of the following components are important in
determining mineral distributions and relationships in the
Takitimu sequence:

\[ \text{SiO}_2^*, \text{Al}_2\text{O}_3^*, \text{Fe}^{3+}, \text{Fe}^{2+}, \text{Mg}^{2+}, \text{Ca}^{2+}, \text{Na}^+, \text{K}^+, \text{H}^+, \text{H}_2\text{O}, \text{CO}_2. \]

The activities of the components in solution are in part a function
of:

[1] initial fluid composition,

The initial fluid composition in all but the shear zones in the
Takitimu Mountains and other volcanic sequences is probably buffered
rapidly by rock chemistry. Thus similar metamorphic assemblages are
developed in submarine sequences where initial pore fluid
composition is that of sea-water and in sub-aerial volcanic sequences

The presence of volcanic glass is probably critical to
several aspects of the distribution of metamorphic phases in the
Takitimu Mountains.

A combination of rapid burial and abundance of volcanic glass
produces a situation in which rapid evolution of pore fluid
composition occurs. The effects of varying activities of several

* principally as \( \text{H}_4\text{SiO}_4 \)
* as hydroxide or chloride complexes in solution.
components are considered in the following section.

Silica activity: In nature silica activity is likely to be buffered in one of three general fields and this may affect zeolite equilibration. In order of increasing value of $a_{SiO_2}$ the buffers are:

(a) other silicate minerals, (b) quartz and (c) amorphous silica or siliceous glass. Bischoff and Dickson (1975) have shown that silica activity of sea-water in contact with volcanic glass rises rapidly to near saturation with respect to amorphous silica.

Simple reactions link various phases in Zone 1. Reactions relating laumontite, yugawaralite and epistilbite are:

\[
\text{CaAl}_2\text{Si}_4\text{O}_{12} \cdot 4\text{H}_2\text{O} + \text{SiO}_2 = \text{CaAl}_2\text{Si}_6\text{O}_{16} \cdot 5\text{H}_2\text{O} \\
\text{laumontite} \quad \text{yugawaralite}
\]

\[
\text{CaAl}_2\text{Si}_4\text{O}_{12} \cdot 4\text{H}_2\text{O} + \text{SiO}_2 = \text{CaAl}_2\text{Si}_6\text{O}_{16} \cdot 5\text{H}_2\text{O} \\
\text{laumontite} \quad \text{epistilbite}
\]

Increasing $a_{H_4SiO_4}$ will favour formation of the silica-rich zeolite in each reaction. Fig. 7:20 shows the stability field of laumontite relative to non-zeolite calcium aluminosilicates in $a_{H_4SiO_4} : a_{Ca^2+}/(a_{H^+})^2$ space at 200°C at 1 kb P_{fluid}. Insufficient thermodynamic data exist to quantitatively plot epistilbite and yugawaralite stability fields. However, their position can be estimated by utilizing the available information from the metamorphic assemblages.

The assemblage laumontite-quartz is relatively common in the Takitimu section and laumontite also occurs less commonly in association with siliceous glass. Epistilbite and yugawaralite occur solely in segregation vesicles where prolonged contact with the glassy lining would have afforded maximum opportunity for equilibration under condition (c). With this information it is possible to tentatively plot stability fields of yugawaralite and epistilbite to the left of the amorphous silica saturation.
Fig 7.20 Activity diagram for the system CaO-Al₂O₃-H₂O at 200°C and 1 kb pressure. Equilibrium values of $a_{H_4SiO_4}$ for quartz from Helgeson (1969) and amorphous silica from Holland (1967).
boundary (Fig. 7:21). Laumontite infilled vesicles commonly possess an outer lining of chlorite. This is very rare for epistilbite-infilled vesicles, larger vesicles commonly being completely infilled by epistilbite, whereas smaller vesicles contain chlorite only (e.g. OU 38110, 38096, see Table 7:3.). A fluid in contact with a chlorite-lined vesicle will evolve in a direction away from the quartz saturation buffer and not approach the stability fields of yugawaralite and epistilbite. In a segregation vesicle, a fluid of $a_{\text{SiO}_4}$ approximating that of the quartz saturation buffer may evolve towards the amorphous silica saturation buffer and enter the yugawaralite and epistilbite stability fields, due to stagnant fluid : volcanic glass interaction. Coombs et al. (1959) show ready synthesis of epistilbite from glass, oxide mixes and natural minerals, the synthesis field being expanded when amorphous silica or glass is present. This study would appear to be an experimental analogue to the field occurrence of epistilbite in the Takitimu Mountains.

The sequence chlorite-pumpellyite-quartz as observed in a vesicle in OU 39308A may be produced by crystallization of a fluid in which $a_{\text{SiO}_4}$ increases steadily.

**Sodium activity:** Several occurrences of contrasting assemblages in adjacent outcrops suggest gradients of $a_{\text{Na}^+}$ may have been established on a scale of metres. In Zone 1 a vesicular top to a spilitic lava (OU 38117) contains the assemblage analcime-natrolite-thomsonite. Adjacent to this lava flow laumontite is the dominant zeolite phase in both later veins, and as a mineral cement in clastic sediments. The simplest explanation is that the Na-rich assemblage equilibrated with an Na-rich pore fluid, derived from sea-water but in a closed system.
Fig 7.21 Activity diagram for the system CaO-Al₂O₃-SiO₂-H₂O plotting hypothetical stability fields for epistilbite and yugawaralite; data as for Fig 7.20.
The distribution of analcime may in fact relate to local gradients in $a_{Na^+}/a_{Ca^{2+}}$ produced by (a) varying initial fluid composition and (b) varying degree of equilibration of the fluid with pre-existing solid phases.

Heulandite distribution: The occurrence of heulandite in vesicles probably reflects the ability of this zeolite to accept diverse cations by comparison with laumontite and epistilbite. As the major K-bearing new-formed phase in the system heulandite is favoured by increasing $a_{K^+}/a_{Ca^{2+}}$ in metamorphic fluids, perhaps as a result of depletion of Ca$^{2+}$ (and Na$^+$) due to prior crystallization of other phases. Chabazite, the other K-rich zeolite present has very limited distribution. Heulandite also readily accepts Sr$^{2+}$ and probably Ba$^{2+}$ (Table 7:8, Nos. 8 and 9) which occur in only limited amounts in all other phases. Under favourable conditions of fluid composition strontian heulandite may be expected to persist with other calcium zeolites beyond the normal T,P field (stable or metastable) of (Ca-Na) heulandites. Being a relatively silica-rich zeolite crystallization of heulandite will also be favoured by relatively high values of $a_{SiO_2}$. Thus the occurrence of heulandite replacing glass shards in impermeable acidic airfall tuffs adjacent to laumontite-cemented litharenites (OU 39069 and 39255) probably reflects: (1) high $a_{K^+}$ and $a_{Na^+}$ and (2) high $a_{SiO_2}$ in the pore fluid. These conditions will result if there is prolonged and stagnant contact of the fluid and the acidic glass of the tuff. Read and Eisbacher (1974) record an analogue in the Sustut Group of British Columbia where tuffs contain heulandite-albite-quartz-clay, whereas laumontite-albite-calcite-quartz-clay occurs in non-tuffaceous sediment.
Oxygen fugacity: The widespread occurrence of hematite and the presence of nearly pure end-member andradite garnet indicate a relatively high $f_{O_2}$. This is compatible with the widespread distribution of $\text{Fe}^{3+}$-rich epidote and prehnite relative to $\text{Fe}^{2+}$-rich pumpellyite.

The overlap in the stability fields of laumontite or epistilbite and prehnite may be increased if prehnite is rich in $\text{Fe}^{3+}$. Thus high $f_{O_2}$ may be a factor in stabilizing the rare assemblage zeolite-prehnite in the Takitimu Mountains, as suggested by Liou (1971 b).
DISCUSSION

7:52 Facies series concept

Seki (1969) first interpreted the low grade metamorphic facies in terms of facies series. The low pressure facies series was shown to comprise a direct transition from zeolite and prehnite-pumpellyite facies to greenschist facies. Several terrains characterised by the low pressure facies series contain a significant volume of shallow-level intrusive rock comagmatic with and overlapping in age with the volcanic/volcaniclastic rocks they intrude, e.g. Tanzawa Mountains central Japan (Seki et al., 1969); Lachlan Geosyncline, Australia (Smith, 1966, 1969); Iberian Pyrite Belt, Spain and South Portugal (Schermerhorn, 1975). In such instances, where overlap occurs between volcanism, subvolcanic plutonism and rapid burial, magmatic heat may be a factor in establishing short-lived elevated geothermal gradients causing low-pressure, high-temperature metamorphism. It has been suggested that the pumpellyite-actinolite facies appears to be suppressed in a low P-T metamorphic regime (Smith, 1969; Coombs et al., 1970). The intermediate facies series zeolite facies to prehnite-pumpellyite facies to pumpellyite-actinolite facies to greenschist facies is typically developed in sequences of metavolcanic and lithic clastic rocks in areas devoid of intrusive horizons, e.g. Dansey Pass area, New Zealand (Bishop, 1972); Kii Peninsula, Japan (Seki et al., 1964; Seki et al., 1971).

Three case studies of low pressure facies series terrains are reviewed here.

Tanzawa Mountains, Japan

In the Tanzawa Mountains, central Japan, the Tanzawa Group a 10000 m Lower-Middle Miocene pile of massive, pyroclastic and epiclastic volcanic rocks, is intruded by a quartz-diorite body exposed over an
area of 240 sq. km. Seki et al. (1969) consider the body to be Middle Miocene in age contemporaneous with the middle portions of the Tanzawa Group. Five metamorphic zones are distinguished in the Tanzawa Group, and in the later Ashigara Group (Late Miocene–Early Pliocene) separated by the following succession of isograds; stilbite-out, laumontite-out, actinolite-in and oligoclase-in. Prehnite and pumpellyite have equivalent distribution, first occurring at the base of the laumontite zone (Zone 2) and persisting intermittently into the actinolite zone (Zone 4). Zone boundaries are concentric to the outline of the quartz diorite mass, with metamorphic grade increasing towards the body.

Lachlan Geosyncline, New South Wales, Australia

Smith (1969) describes a 3600–9000 m sequence of Ordovician–Silurian strata within the Lachlan Geosyncline in south-eastern Australia. The sequence consists of massive, intermediate to mafic lavas, breccias, tuffs and epiclastic sediments. Smith distinguishes five metamorphic zones on the basis of the appearances of prehnite, pumpellyite, actinolite (coincident with the disappearance of pumpellyite) and biotite respectively. Prehnite overlaps in distribution with actinolite in Zone 4. The sequence contains several small sub-volcanic intrusives with comparable metamorphic assemblages to the host rock (Smith, 1966). These may have acted as a heat source for short-lived, elevated temperature gradients (Smith, written comm., 1975).

Iberian Pyrite Belt

Schermerhorn (1970, 1975) describes the petrology of an upper Paleozoic Geosyncline in the south-west of the Iberian Peninsula. His study of metamorphism is centred on the 700–800 m thick Carboniferous Volcanic-siliceous Complex, forming the middle part of the succession. The sequences consist of spilitic and quartz keratophyric metavolcanic rocks and sediments; associated with the volcanics are albite diabase
sills. In the north of the belt repetitions of diabase sills make up the bulk of the complex (up to a thickness of 500 m). Schermerhorn (1970) considers the sills are high level intrusives intruded prior to and perhaps overlapping with spilitic volcanism. The Iberian Pyrite Belt was metamorphosed in Upper Carboniferous times to prehnite-pumpellyite and greenschist facies assemblages (Schermerhorn, 1975). The two zones corresponding to the two facies are separated by a pumpellyite-out isograd. Prehnite disappears close to the pumpellyite isograd and overlaps actinolite in distribution. In the Iberian Peninsula, however, there is a zone up to 8 km wide in the vicinity of the pumpellyite-out isograd which contains no mafic igneous rocks. This zone may have been metamorphosed under conditions equivalent to the pumpellyite-actinolite facies, although bulk composition precluded the development of pumpellyite-actinolite assemblages.

Similar short-lived high temperature gradients may have prevailed in the prehnite-rich, pervasively intruded metavolcanic sequence of the Takitimu Mountains, and in other volcanic arc terrains where a transition from prehnite-pumpellyite facies to greenschist facies or alternatively, pumpellyite-actinolite facies is not seen. The Takitimu sequence probably experienced a variety of geothermal gradients with an initial but short-lived period of high temperature gradients during Middle to Upper Permian subvolcanic plutonism, followed by a longer period of burial under a moderate geothermal gradient. Mixed thermal gradients produced by this type of poly-metamorphism may contribute to the overlapping pattern of distribution of the metamorphic phases.

7.53 Status of Prehnite-pumpellyite facies.

De Roever (1947) describing low-grade metamorphic rocks in the Celebes, first related pumpellyite to a discrete metamorphic facies; he considered the facies a precursor of the 'lawsonite-glaucophane
sub-facies. Coombs (1961) then defined a prehnite-pumpellyite meta-
greywacke facies to include those suites of assemblages produced under
physical conditions in which the assemblages quartz-prehnite-chlorite
or quartz-albite-pumpellyite-chlorite are commonly formed without the
presence of quartz-laumontite, quartz-heulandite, lawsonite or jadeite.
He distinguished two zones or subfacies, a lower grade zone characterised
by combinations of quartz, albite, prehnite, pumpellyite, chlorite,
calcite, sphene, orthoclase and muscovite, and a higher grade zone with
combinations of quartz, albite, chlorite, sphene, actinolite, muscovite,
calcite, stilpnomelane, pumpellyite and epidote.

Hashimoto (1966) defined the pumpellyite-actinolite schist facies
characterised by the occurrence of pumpellyite-actinolite and
stilpnomelane, and the absence of prehnite, lawsonite and glaucophane.
This in effect raised the two original zones or subfacies of Coombs
(1961) to facies status. Hashimoto linked the two facies by the
following discontinuous reaction:

\[
\text{prehnite} + \text{chlorite} + \text{SiO}_2 = \text{pumpellyite} + \text{actinolite} + \text{H}_2\text{O}.
\]

The facies boundary is then marked by the appearance of the
partial assemblage pumpellyite-actinolite, in principle at the expense
of prehnite-chlorite. In practice in many low grade metamorphic
terrains of the low pressure facies series it is very difficult to
judge if this reaction occurs. Varying degrees of reconstitution,
and the presence of deuteric alteration or mineral assemblages
produced under mixed thermal gradients, may make it difficult to
judge if metamorphic equilibrium is attained. In discussing metamorphic
assemblages in the Takitimu Mountains and similar terrains, this reaction
can be adequately characterised only by the five component system
\[\text{Al}_2\text{O}_7\text{Fe}_2\text{O}_5\text{CaO-FeO-MgO (with excess SiO}_2\text{ and H}_2\text{O-C}_2\text{ fluid phase), and special constraints on composition are required for the joins prehnite-}\]
chlorite and pumpellyite–actinolite to intersect. For other combinations of phase compositions all four phases may coexist.

Subsequent writers have mapped zonal sequences utilizing either an actinolite-in (Seki et al., 1969) or a prehnite–out isograde (Bishop, 1972). Problems arise in relating zonal sequences to metamorphic facies where there is a broad overlap in distribution between prehnite and actinolite, as in the Iberian Pyrite Belt (Schermerhorn, 1975) and the Takitimu Mountains. Also, one must be cautious in applying the concept of progressive metamorphic zonation to situations such as the Takitimu Mountains where the assemblages may be the products of superimposed local and regional thermal gradients.
CHAPTER 8 - SUMMARY AND CONCLUSIONS

REGIONAL SETTING

8.1 Lower Permian paleogeography of southern New Zealand

Evolution of the Takitimu Group followed stages recognised in modern volcanic arc systems overlying present day consuming plate margins. Major and trace element chemistry and mineralogy of the rocks exposed in the central Takitimu Mountains are typically calcalkaline. Paleotectonic reconstructions leave little doubt that the group and its correlatives of the Brook Street terrain formed in part of an ancient island arc-trench system (Wellman, 1956; Landis & Coombs, 1967; Coombs et al., 1975). The Dun Mountain ophiolite belt lying to the east of the Brook Street terrain probably originated as oceanic floor either at an active plate margin or in a back-arc basin (Coombs et al., 1975). In the first model the Takitimu Group formed in an east-facing arc and on the margin of a fore arc basin. In the second model the group accumulated in a west-facing arc and possibly on the arc-ward flank of a marginal basin.

Various paleotectonic reconstructions suggest that in pre-Late Cretaceous times southern New Zealand lay to the southwest of Tasmania (Griffiths & Varne, 1972; Hayes & Ringis, 1973). In Silurian-Triassic times calcalkaline volcanism occurred in eastern Australia within a series of eastward-migrating, eastward-facing island arc-trench systems (Solomon & Griffiths, 1972; Scheiber, 1973; Leitch, 1975). If the paleotectonic reconstructions cited above are correct the Takitimu Group probably formed in a similar and possibly related eastward-facing island arc system.
8.2 Introduction

The geological history of the Takitimu Group is marked by close relationships, and overlaps in time, between sedimentation and volcanism, shallow level plutonism and low grade metamorphism. Tilting and block-faulting of the strata probably also partially overlapped in time with the afore-mentioned processes. A major aim of this study has been to document each aspect to the development of the Takitimu Group as a complete description of the history of an ancient volcanic arc sequence. The rocks of the Takitimu Group are better exposed than most similar sequences in and adjacent to modern island arcs and aspects of the geological history of the group have been examined in more detail than is found in existing publications describing modern analogues. For this reason study of such ancient volcanic arc terrains can give insight into processes active at the present day.

8.3 Tectonism

"Uplift of the volcanic arc may occur either towards the close or following the cessation of volcanism ... Many island arcs contain very thick stratigraphic successions consisting largely of calcalkaline volcanic and volcaniclastic rocks which are steeply dipping and cut by numerous high-angle faults." - Mitchell & Reading (1971), Evolution of Island Arches, p259 & p271

Deformation style of the Takitimu Group closely resembles that of older (principally Tertiary) volcanic rocks in modern island arcs (Mitchell & Reading, 1971). Takitimu Group strata are steeply dipping and locally block-faulted. The precise timing of tectonic movements affecting the group is not established. The major movements may have accompanied the later-Mesozoic Rangitata orogeny. However, the initial dip of strata in volcaniclastic
aprons may be relatively steep (Parsons, 1969), and it is clear that prior to deposition of the overlying middle–Upper Permian Productus Creek Group sufficient uplift had occurred to expose at least portions of the deeper part of the Takitimu pile.

8.4 Sedimentation and Volcanism

"The resulting magma erupts as submarine volcanoes ... Continued eruptions and volcanic upbuilding results in volcanic islands ••• Continued eruptions and volcanic upbuilding results in volcanic islands ••• Erosion of the volcanic islands ... leads to the accumulation of large amounts of epiclastic volcanic detritus mixed and inter-bedded with pyroclastic and autoclastic deposits."


Active volcanism persisted throughout the period of deposition of the Takitimu Group. The chemical composition of the erupted volcanic rocks changed several times from exclusively basalt to basalt and rhyodacite back to basalt and finally to basalt and andesite.

The oldest rocks exposed consist of predominantly fine-grained volcaniclastic sediment and red-brown mudstone. A similar association is found at the axis of modern marginal basins (Karig, 1970; Hawkins, 1974) and on the trenchward side of fore arc basins (Karig, 1971). Basaltic and rhyodacitic lavas were erupted towards the end of the Brunel times. In the overlying Chimney Peaks Formation basaltic arenites and rudites predominate alternating with lenzoid accumulations of basaltic and rhyodacitic lava flows and tuffs and agglomerates and pillow and pyroclastic breccias of basaltic composition. Individual volcanic vents have not been recognised but the lenzes of primary volcanic rock constitute a vent facies. The clastic rocks probably formed in volcaniclastic aprons fringing these lenzes.

Deposition of the Chimney Peaks Formation was characterised by bimodal volcanism. Explosive and extrusive basaltic volcanism alternated but rhyodacitic volcanism principally consisted of
passive extrusion. Rhyodacitic, pyroclastic and volcaniclastic rocks are, in fact, rare throughout the Takitimu Group.

A thick sequence of basaltic agglomerate, pillow breccia and lava and microgabbro then formed which constitutes the Heartbreak Formation. The formation is thought to have formed almost exclusively in the vicinity of active submarine vents although vent breccias and agglomerates have not been recorded.

The MacLean Peaks Formation then accumulated dominated by basaltic to andesitic arenites and rudites. Thickly-bedded arenites and rudites alternate with massive and pillow lava flows and tephra in the lower half of the formation. This association probably formed in the proximal portions of a volcaniclastic apron(s) (by analogy with archipelagic aprons fringing oceanic volcanoes, Lonsdale, 1975). The upper portion of the formation contains thinly- and thickly-bedded arenites forming couplets with red-brown mudstone and basaltic and andesitic tuffs. A similar sequence was encountered at D.S.D.P. site 60 in the Marianas fore arc basin and in the distal portions of the Samoan archipelagic apron (Lonsdale, 1975).

The Elbow Formation was deposited at the close of the Takitimu times when subsidence is thought to have failed to keep pace with deposition. It contains a high proportion of rudites including well-rounded conglomerates.

The Takitimu Group is overlain by the Middle and Upper Permian Productus Creek Group, a sequence of bioclastic limestones and calcareous volcaniclastic sediments. The transition between the groups was marked by a substantial diminution in volcanic activity. The Takitimu Group was apparently a source of Productus Creek Group detritus and possibly formed an inactive frontal arc in Productus Creek times.
8.5 Plutonism

"Plutons possibly formed by anatectic melting at the base of the pile intrude the sedimentary prism ..."

Intrusive rocks exposed in the central Takitimu Mountains are subdivided here into the White Hill and MacKinnon Peak Intrusives. The suites may be distinguished on the basis of mineralogy and field occurrence.

The White Hill Intrusives probably formed from several distinct but petrologically similar magma batches emplaced over a wide time interval. The older members of the suite predate the MacKinnon Peak Intrusives and are probably Middle to lower Upper Permian in age. Other members of the suite were probably emplaced as late as the Early Triassic.

The White Hill Intrusives show strong similarity in mineralogy and chemistry to the Takitimu Group. Large bodies are differentiated and at least part of the differentiation took place in situ. Concordant intrusion prevailed and unlike similar intrusive suites elsewhere (e.g. Western Cascades, Fiske et al., 1963) there is no evidence that the White Hill bodies ever broke to the surface to form volcanic vents. The White Hill Intrusives closely resemble the more basic intrusive rocks of the Longwood complex, the MacKay Intrusives and the Tasman Intrusives which intrude strata correlative of the Takitimu Group in south western Southland, West Otago and Nelson respectively (Grindley, 1958; Beck, 1964; Wood, 1969; Williams, 1975). The period of time in which the White Hill Intrusives were emplaced was therefore a period of extensive, and protracted, intrusive activity throughout the Brook Street terrain.

The MacKinnon Peak Intrusive rocks are mineralogically distinct from all other igneous rocks in the Takitimu Mountains. Correlatives
have not been reported from any other portion of the Brook Street terrain. The distinctive phenocrystal mineralogy of the dykes suggests the phenocrystal phases formed under conditions of high $P_{H_2O}$.

A model for the origin of the dykes is presented (Chapter 6) postulating extensive crystal fractionation in a shallow magma chamber(s) leading to formation of adcumulate layers overlain by a transitional zone of close-packed crystals with interstitial fluid. The chamber and cumulus layers were disrupted, and the MacKinnon basalts emplaced in a 'crystal-mush'. Late-stage retrogressive boiling and volatile streaming probably played a role in dyke emplacement.

$K/Ar$ cooling ages for hornblends from the dykes (231-242 my) suggests disruption of the chamber(s) and dyke emplacement probably coincided with tectonic movements which exposed lower portions of the Takitimu Group to erosion during deposition of the Productus Creek Group.

8.6 Low grade metamorphism

"Isostatic adjustment of the crust causes subsidence and because of the high temperature gradient, the lower part of the volcanic and sedimentary pile may be metamorphosed in the zeolite facies. Prolonged burial at a considerable depth may result in low pressure/high temperature metamorphism . . . ."


The Takitimu Group is subdivided into two metamorphic zones. The upper zone contains zeolite facies assemblages, the lower prehnite-pumpellyite facies assemblages. Degree of mineralogic reconstitution is variable and nowhere complete. Maximum reconstitution post-dated emplacement of large White Hill Intrusive bodies but incipient alteration of each horizon probably commenced soon after deposition. Submarine volcanic rocks are susceptible to early alteration because of:
(1) the abundance of volcanic glass,
(2) the presence of abundant pore fluid, derived from seawater, with the capacity for rapid reaction with glass,
(3) rapid burial which is common in submarine volcanoes.
Volcanic glass initially reacted with seawater to form chlorite and/or montmorillonite and a fluid enriched in $\text{SiO}_2$, $\text{Al}_2\text{O}_3$, $\text{Na}_2\text{O}$ and $\text{CaO}$. Ultimately the fluid evolved to the stage where albite and hydrated calcium- and sodium-aluminosilicates formed.

The White Hill sills probably contributed heat to the system during metamorphism. Several other volcanogenic terrains characterised by the low pressure facies series of Seki (1969) also contain significant amounts of shallow level plutonic rock and magmatic heat high in the crust may be a factor in establishing elevated geothermal gradients responsible for the low pressure metamorphism.

8.7 Petrogenesis

The Takitimu Group shows the following characteristic features of calcalkaline rocks:

(1) the suite lacks strong iron-enrichment,
(2) strong enrichment in light REE relative to heavy REE in analysed specimens,
(3) porphyritic rocks are common and calcic plagioclase is the dominant phenocryst phase,
(4) basaltic rocks are low in nickel and chromium.

Strongly plagioclase-phyric basalts, which form the predominant rock type in the group, lack positive europium anomalies. This suggests that the porphyritic basalts are unlikely to be partly cumulus unless the plagioclase phenocrysts in these rocks are not enriched in europium. Experimental work shows unrealistically
high values of $f_{O_2}$ are required to preclude europium concentration in plagioclase (Drake & Weill, 1975). The absence of europium anomalies in plagioclase-phyric basalts has major implications for the definition of high-alumina basalts. This term was introduced by Kuno (1960) for exclusively aphyric basalts. In suites where porphyritic basalts lack an europium anomaly they can be justifiably included in the category of high-alumina basalts.

The petrogenetic model favoured by the writer for Takitimu Group igneous activity derives basaltic andesite and andesite from parental high-alumina basalt by crystal fractionation. The major element composition of rhyodacitic lavas cannot be derived by removing observed phenocryst minerals from the siliceous andesites. Trace element geochemistry and the field association of rhyodacite with basalt but not andesite also suggests a separate origin for the rhyodacites. Two-stage partial melting to yield separate primitive basaltic and rhyodacitic magmas is a possible mechanism.

Crystal fractionation of plagioclase-rich assemblages can successfully generate a series of basic to intermediate calcalkaline liquids as demonstrated for Santorini Volcano, Rabaul Caldera and the Takitimu suites. It probably can also be applied to other suites with a predominance of high-alumina basalts and basaltic andesites, and little modal amphibole. It is unlikely to be viable for calcalkaline suites with a predominance of hornblende andesite e.g. Lesser Antilles. In such suites hydrous partial melting of amphibolite or peridotite or amphibole fractionation are more likely to have occurred.
8.6 Synopsis

The Lower Permian paleogeography inferred for the central Takitimu Mountains resembles that about present-day island arcs. The post-depositional history of the group, outlined in this study, closely resembles that of the older strata in Tertiary-Recent volcanic arcs in the vicinity of modern plate boundaries (see Mitchell & Reading, 1971).

The oldest strata exposed formed in the axial portions of a basin or basins flanking an active calcalkaline volcanic chain. The thick sequence exposed in the central Takitimu Mountains accumulated partly in this basin(s) and partly in the volcanic chain, in Lower Permian times. Shallow-level plutonism and low grade metamorphism followed rapidly on from sedimentation and volcanism. By Upper Permian times metamorphosed lower portions of the pile had been uplifted and partly eroded to contribute detritus to the Productus Creek Group.
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*
Sample Preparation

Rock samples were split into 1-2 cm cubes using a hardened steel rock splitter after removal of weathered surfaces and veins. The cubes were crushed in batches for 4 seconds in a tungsten-carbide 'Tema' mill and sieved and recrushed until the sample passed through 60 mesh bolting cloth. A split was then removed for FeO determination and the remainder crushed for 45 seconds to one minute.

Sample size varied from 500 g to 1 kg depending on grain size of the rock.

Samples for K/Ar analysis were handcrushed to pass through 120 mesh bolting cloth, to minimise argon loss through heating.

Whole rock analysis - X-ray fluorescence spectrography

Whole rock analysis for all major oxides except FeO was performed at Victoria University on a Siemens Kristalloflex X-ray generator with T-type electronics. A fusion mixture was prepared consisting of 0.5 g sample, 2.68 g flux and 0.035 g NH₄NO₃, fused and quenched on 38 mm Dural Plattens at 230°C. Concentrations were determined relative to an artificial standard FSL4A. Major element concentrations for various international rock standards determined relative to FSL4A are given in Table A1. Two Victoria University standard rocks VUW-1 and L-44 were run as internal standards and results are given in Table A3 as an indication of consistency. The analyses and matrix correction procedures follow the methods of Norrish and Hutton(1969).

Whole rock analysis - spark source mass spectrometry

Rare earth elements (REE) and other trace elements were determined by spark source mass spectrometry by P.C.Rankin, Soil Bureau, Lower Hutt, using methods similar to those of Taylor (1965, 1971).
samples were sieved through 50 μm nylon cloth and mixed in a 1:1 ratio with ultra-pure graphite which contained 500 ppm spectrographically-pure lutetium oxide as an internal standard. Based on the repeated analyses of international standards, the precision of the method is considered to be 15% for the REE, U, Th, Pb and up to 25% for the other elements. Further description of the technique, and the results of the analysis of two international rock-standards are given by Howorth and Rankin (1975).

Whole rock analysis - chemical

Back-titration against ammonium metavanadate was used to determine ferrous iron (FeO) following the method of Wilson (1955). Loss of ignition (LOI) was determined as the weight loss of a sample heated for 1-2 hours at 1000°C.

Mineral analysis - electron microprobe

Analyses were made with a Japan Electron Optics Laboratory electron probe X-ray microanalyser Model JXA-5A. Instrument conditions for analyses of all phases with the exception of zeolites were 15 kv accelerating potential, approximately 0.03 μA specimen current and 1-2 μm electron beam diameter. For zeolites operating conditions were

- **accelerating potential**
  - Si, Al, Na, Mg, Ca: 8 kv
  - Fe, Mn, K: 15 kv

- **specimen current**
  - 0.03 μA

- **beam diameter**
  - 10-20 μm

Three to five countings of 10 seconds integration time were averaged for each element. Standard materials were Amelia albite (Na), Gotthard adularia (K), Sennin Mine hematite (Fe), Nodatamagawa Mine manganosite (Mn), Acoje Mine chromite (Cr), synthetic periclase (Mg), corundum (Al), Wollastonite (Ca, Si) and rutile (Ti). The correction procedures used
were those of Bence and Albee (1968) for most phases using correction factors given by Nakamura and Kushiro (1970) and Sweatman and Long (1969) for zeolites.

K/Ar age determination

Samples selected for K/Ar age determinations were crushed as described previously. Individual hornblende crystals were hand-picked from the MacKinnon Peaks basalts and purified using heavy liquids and a magnetic separator. Experimental details of potassium and argon analysis techniques are given in Adams (1975) but in this project argon analyses were measured on an AEI MS-10 mass spectrometer recently coupled to the argon extraction system. The accuracy of potassium and argon analyses was assessed by measurement of international standards P207 muscovite and Arvonia slate total rock.
Table A1 Whole rock analyses of various international rock standards using artificial standard FSL-4A.

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<th>G-2</th>
<th>BCR</th>
<th>AGV</th>
<th>GSP</th>
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Table A2 Generally acceptable analyses of above standards listed by Victoria University taken from Flanagan (1974).

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<th>GSP</th>
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GEOL OGY OF CENTRAL TAKITIMU MOUNTAINS; cross-sections