GEOLOGY OF MOUNT KOBAY

by

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B.Sc., University of British Columbia, 1964

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THE REQUIREMENTS FOR THE DEGREE OF
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in the Department of
GEOLOGY

We accept this thesis as conforming
to the required standard

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Date 26 Sept. 1969
ABSTRACT

Rocks of the Kobau Group occur between the Okanagan and Similkameen Valleys in southern British Columbia and northern Washington. The Group consists of quartzite, phyllite, greenstone and minor limestone deposited within a synorogenic, eugeosynclinal environment in pre-Cretaceous, possibly post-Devonian, time. Intrusion and extrusion of basic igneous rocks accompanied deposition. The observed succession has been divided into nine units with total original thickness under 5,000 feet.

Earliest recognized deformation of the Group formed tight recumbent folds with easterly trending axes. Transposition of compositional layering to foliation and extensive shearing occurred at this time and was accompanied by regional dynamothermal metamorphism which attained the middle subfacies of the greenschist facies. Later (second phase) deformation produced overturned and normal folds with steep axial planes and south-easterly trending axes, and refolded early recumbent structures. Several quartz latite dykes cut the Group during or shortly after late folding.

Emplacement of granitic and dioritic stocks with radiometric ages of 144 x 10^6 years or less followed second phase
folding. Contact metamorphic zones of varying extent are present around larger intrusive bodies and attain the hornblende-hornfels facies. A latest (third) phase of deformation about poorly defined northerly trending axes may be related to this intrusive episode. A number of dacite and basalt dykes intruded both stock and country rocks.

Extensive fracturing during Tertiary time broke pre-existing structures into numerous blocks and wedges. Fractures parallel axial planes of early and late folds as well as faults in the Okanagan Valley.

Relationships between the Kobau Group and rocks in adjacent areas are unknown. The Group possesses lithologic and structural similarities to parts of the Shuswap Complex and may share some of the complex's history. Part of the southerly adjacent Anarchist Group may be correlative with the Kobau Group.
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FRONTISPIECE

MOUNT KOBAY SEEN FROM ANARCHIST MOUNTAIN, LOOKING TO THE NORTHWEST ACROSS OSOYOOS LAKE.
CHAPTER I

INTRODUCTION

GENERAL INTRODUCTION

Mount Kobau is located in south-central British Columbia between the Okanagan and Similkameen Valleys, six miles north of the forty-ninth parallel. Investigation of stratigraphy and structure on Mount Kobau was done to establish the geological history and age of the Kobau Group (Bostock, 1940) that crops out on Mount Kobau and to the south-east on Kruger Mountain. Mapping during three field seasons (1965 to 1967) was carried out on a scale of four inches to the mile on an enlarged version of an advance print of the Keremeos sheet (82 E/4), supplemented by aerial photographs (British Columbia Department of Lands and Forests, BC5123-001 to 044, November 5, 1964). An area of approximately seventy-five square miles between latitudes 49°04' and 49°13' north and longitudes 119°32½' and 119°44' west was examined. Specific geological and geographical limits to the map-area are the Okanagan and Similkameen Valleys to the east and west, respectively, Blind Creek and the Oliver Granite to the north, and Richter Pass to the south (Figure 1-1).
FIGURE 1-1.
LOCATION OF MAP-AREA
Access to Mount Kobau is provided by highways in the Okanagan and Similkameen Valleys and from Richter Pass (Figure 1-2). A good dirt road passes from Oliver along Blind Creek to Cawston and a poor British Columbia Forest Service road extends along the crest of the Mount Kobau ridge. Numerous lumber roads are open west of Oliver, and in 1966 the Federal Department of Public Works constructed an excellent road to the summit of Mount Kobau to provide access for the proposed Queen Elizabeth II Observatory. Sparse vegetation makes movement by foot easy over most of the area. Exceptions are burnt-off areas and canyons choked with second growth and heavy vegetation, and steep western slopes facing the Similkameen Valley.

Mount Kobau lies in the southwest corner of the Thompson Plateau just west of the Okanagan Highlands (Holland, 1964; p.71 and map). The plateau and highlands represent a late Tertiary erosion surface (Holland, 1964; p.71), a gently rolling area of low relief deeply dissected by the Okanagan and Similkameen Rivers and their tributaries. Structural control of these channels by fractures is postulated (see Chapter III). The most prominent valleys parallel north-trending faults.

Presence of ice-sheets during the Pleistocene is evidenced by prominent glacial striae at all elevations and drift
FIGURE 1-2
ACCESS ROUTES WITHIN THE MAP-AREA.
covered valley floors and lower slopes. Marginal meltwater channels and glacio-lacustrine silt beds are further indications of extensive glaciation. These silts, alluvial fans and other surficial deposits, together with the glacial history of the area, have been discussed by Nasmith (1962).

PREVIOUS WORK

Between 1859 and 1861, G. Gibbs and H. Bauerman, attached to a joint International Boundary Commission, surveyed lands near the forty-ninth parallel. In the area of the Ashnola, Similkameen and Okanagan Rivers Gibbs (1874) noted the presence of mica schist, lamellar quartz and syenite interstratified with slate. Bauerman (1884, p.19B) observed "...black siliceous slates...beds of gneissic mica-slate...". He considered these to be representative of the lower part of the Chilukweyuk (Chilliwack) slates.

G.M. Dawson (1879) while mapping physical and geological features of the southern interior of British Columbia, described some of the rocks observed on his traverse down the Similkameen Valley as follows:

"The wide valley...marks the appearance of schistose and often silvery rocks, in colour varying from blackish and greyish to greenish. These are frequently quartzites with division planes rendered lustrous by talc on imperfectly crystallized mica..."

"Further on, these are replaced by greener rocks, often feldspathic and occasionally
schistose, which become in places actually coarse green decomposed diorite, with veins of epidote, and so massive that it is impossible to ascertain their attitude. These rocks are again followed by pale banded cherty quartzite interbedded with greenish schistose rocks, hornblendic or chloritic..."

— Dawson, 1879, p.87B.

He included the folded and metamorphosed sediments in the Okanagan and Similkameen Valleys within the Cache Creek Group of Carboniferous age (Dawson, 1878). He states:

"The general impression conveyed by an examination of the rocks from Vermillion Forks (Princeton), by the Similkameen River to Osoyoos, is, that...there is represented but a single great formation; or, that if two or more of the greater geologic divisions are included, they have been folded together at one period."

— Dawson, 1879, p.87B.

Smith and Calkins (1904), from information gathered during geologic reconnaissance in northern Washington, correlated low grade metamorphic rocks near Oroville with Dawson's Cache Creek Group.

Umpleby (1911) observed deformed and metamorphosed shales, limestones, and volcanics in the United States near Oroville and Nighthawk. He subdivided these rocks into three parts (lower: clay slate and argillaceous schists; middle: slate and limestone; upper: siliceous argillites and volcanics) and correlated them on lithologic grounds with the Cache Creek
Group. He also observed west-dipping reverse and normal faults.

In 1912, Daly traversed the North American Cordillera at the forty-ninth parallel. He proposed the name Anarchist Series for a succession of meta-sediments of presumed Palaeozoic age found on Anarchist Mountain and near Bridesville. This series is made up of deformed quartzites, phyllitic slates, greenstones and limestone pods, and their more metamorphosed equivalents (schist, marble and amphibolite). Daly (1912, p. 389) believed that west of Osoyoos Lake the series: "...was probably represented by a yet more extensively metamorphosed group of rocks..." He correlated roof pendants of schist found on Chopaka, Snowy, and Horseshoe Mountains (west of the Similkameen Valley) with the Anarchist Series, and also discussed aspects of the Osoyoos Batholith, a gneissic granodiorite body lying west and east of Osoyoos Lake (Daly, 1912, pp. 439-443).

Campbell (1939) studied alkaline syenite rocks on Kruger Mountain and their relationship to surrounding intrusions and meta-sediments of the Anarchist Series. He found evidence that deformation of the Osoyoos Batholith postdated its intrusion.

Bostock (1940) first defined rocks on Mount Kobau as a distinct group and assigned a questionable Carboniferous age to them. Mapping on a scale of one inch to the mile, he was able to outline a series of northwesterly trending folds with
steeply dipping axial planes. These are shown diagrammatically on cross-sections accompanying his map (341 A). He described the group as follows:

"...a great thickness of metamorphosed, stratified rocks mainly of sedimentary origin. The quartzite members are thinly-bedded and commonly micaceous or graphitic. There are also fine-grained, siliceous, mica schists and others containing chlorite, hornblende, graphite and talc. The associated greenstones are variously sheared."

-- Bostock, 1940.

He believed the Kobau Group overlies a gneissic complex of unknown age east of the Okanagan Valley and dips under the Similkameen Valley and Mesozoic formations to the west.

Cairnes' (1940) map of the Kettle River (west half) area showed the Kobau Group as part of the Anarchist Series, both of which he considered to be Carboniferous (?) in age. He believed the Osoyoos Batholith and other granitic intrusions on Mount Kobau to have been highly metamorphosed and hence part of a series of highly metamorphosed rocks known as the Shuswap Complex (Dawson, 1898). He interpreted their state and that of the complex to be a result of extreme metamorphism by Mesozoic intrusions.

In 1941, Krauskopf investigated intrusive rocks of the Okanagan Valley near the forty-ninth parallel. He found that "...deformation (of the Osoyoos Batholith) preceded final
crystallization,...that gneissic and cataclastic structures developed by the same deforming stress." (Krauskopf, 1941, p.15). He also commented on altered quartz diorite north and east of Kruger Mountain, considering it to be a part of the Osoyoos Batholith because of similar contact metamorphic effects.

Waters and Krauskopf (1941) examined the Colville Batholith and surrounding intrusions and sediments east of the Okanagan Valley in Washington. They divided the Anarchist Series into three parts much as Umpleby (1911) did. On the basis of poorly preserved fossils they assigned a: "...probably Carboniferous and most likely Permian..." age. (Waters and Krauskopf, 1941, p. 1364).

Little (1961) revised Cairnes' map of 1940 and placed the Kobau Group below the Permian (Dunbar, 1932) Cache Creek Group, giving the former a possible Carboniferous age. He considered it impossible to distinguish the Anarchist Group* from similar groups ranging in age from Carboniferous to Upper Triassic. He believed granitic intrusions on Mount Kobau are related to Nelson intrusions of Mesozoic age. Subsequent mapping by Little and Thorpe (1965) enabled them to subdivide the Anarchist Group into six units Permian and/or younger in age. (Figure 1-3).

*The change in status from series to group is not explained.
FIGURE 1-3
GENERAL GEOLOGY OF THE SOUTHERN OKANAGAN.
White, et al. (1967) have dated radiometrically Mesozoic intrusions near Oliver and obtained ages of 140 and 110 million years for the Oliver Granite and Fairview Granodiorite, respectively.

GENERAL GEOLOGY

The Kobau Group in south-central British Columbia is a part of a body of pre-middle Mesozoic (Callovian) metamorphic rocks bounded for the most part by the "Okanagan Composite Batholith" (Daly, 1912), and in fault contact with Triassic volcanic rocks and sediments (Figure 1-3).

The type area of the Kobau Group is on Mount Kobau (Bostock, 1940). The predominant rock type is quartzite but schistose and massive greenstone and pelitic units including micaceous and amphibolitic schist and phyllite are also present. Marble forms a pure, highly recrystallized unit found in large and small lenticular bodies. Three phases of folding have extensively deformed the group. The first phase produced large, recumbent, tightly compressed nappes. Regional metamorphism associated with this folding attained the greenschist facies. The second phase deformed first phase structures and produced overturned and normal folds. Numerous small granitic and dioritic intrusions cut and deformed these earlier structures in Mesozoic time, and thermally metamorphosed country
rocks to the hornblende-hornfels facies within contact aureoles. The third phase of folding formed gentle folds and may be related to this episode of intrusion. Faulting in Tertiary time broke up pre-existing structures into numerous blocks and wedges.

Relationships between the Kobau Group and rocks in adjacent areas are unknown. The group possesses lithologic and structural similarities to parts of the easterly adjacent Shuswap Complex and may share some of the complex's history. The Anarchist Group is also lithologically similar to the Kobau and at the International Boundary the two groups have been mapped as one (Bostock, 1940; Campbell, 1939, Waters and Krauskopf, 1941). A member of the Cache Creek Group (Blind Creek limestone) of probable upper Carboniferous-lower Permian age (R.V. Best, W.R. Danner, personal communications, 1969) according to Bostock (1940) is in presumed fault contact with the Kobau Group. Triassic or older formations found west and northwest of Mount Kobau (Bostock, 1940) are also in presumed fault contact with both Blind Creek limestone and the Kobau Group.
CHAPTER II

STRATIGRAPHY

INTRODUCTION

The metamorphosed succession of sedimentary and basic igneous rocks which makes up the Kobau Group has been established by detailed mapping and integration of lithologic and structural data. Extreme shearing and transformation of bedding during early phases of folding have obscured the original stratigraphic succession and deformed compositional layering into lenticular bodies of limited lateral extent. Composites of lithologically distinct lenses comprise presently existing mappable units. Locally spatial relationships between such units have been determined and correlation between resulting structural sections has been established (Figure 2-1).

The observed sequence, of which neither top nor bottom has been observed, is illustrated in Figure 2-1. In no instance has it been possible to ascertain the original orientation of any part of the succession because sedimentary structures, if any existed, have been destroyed by severe deformation. Facies changes and unconformities, if any were present, are impossible to detect for the same reason.
STRUCTURAL LITHOLOGIC SUCCESSION WITHIN THE KOBAU GROUP.

Measured Sections

A

B

C

D

Faults.

Lithology

- Massive quartzite
- Foliated phyllitic quartzite
- Phyllite and schist
- Greenstone
- Marble and calcareous phyllite

FIGURE 2-1
Estimates of present thickness of the sequence have been made from local sections (Figure 2-1 and Appendix B, Figure 1) but do not reflect original thickness because of extreme deformation of the rock units.

The Kobau Group is composed of a mixed assemblage of metamorphosed sedimentary and basic igneous rocks. Psammitic and pelitic sediments make up most of the group but rocks of probable volcanic origin are common. Calcareous sediments are scarce, the primary occurrence being lenses of pure crystalline limestone within phyllite and greenstone.*

The most evident feature of most units of the group is foliation which parallels compositional layering. This penetrative structure was developed during the first phase of folding (see Chapter III) as a result of tight compression and shearing which forced original bedding and axial plane foliation into near parallelism. Platy and prismatic minerals developed along foliation planes in response to regional dynamothermal metamorphism which accompanied folding. Pelitic sediments, basic rocks, and argillaceous quartz sandstone that altered to phyllite, chlorite schist and micaceous quartzite, respectively, possess this foliation, the postulated formation of which is illustrated in Plates 3-1 to 3-6.

*See Appendix A for discussion and definition of this term.
Sections have been measured in a number of localities (Map 1). They are limited by megascopic structures which repeat and overturn parts of the succession and by late faults (see Chapters I and III), and must therefore be considered in relation to large scale folds which are described in later parts of this thesis.

STRUCTURAL SUCCESSION

As the original orientation of the stratigraphic succession is unknown, it is necessary to establish a structural succession wherein the "oldest" unit lies within antiformal nappe cores. Northerly verging nappes have been defined as antiformal on the basis of present orientation. As original orientation is unknown, such a definition is arbitrary. What is believed to be the core of such a nappe is exposed about one mile northeast of the Testalinden Granodiorite (Map 2). Foliation and compositional layering in this area generally strike east and dip south at about 45 degrees. Section A, measured south (from nappe core upwards through its upper limb) is limited by faulting in Testalinden Creek, but contains Units one and part of two. Section B, measured north from the nappe core through an overturned sequence in its lower limb, contains these units as well as Units three, four, and five.
Unit one is a repetitive sequence of interfoliated quartzite and phyllite with minor amounts of greenstone and schist. Quartzite is foliated with biotite, chlorite, tremolite, and white mica in fine layers between siliceous bands. Phyllite is fine grained, distinctly foliated and composed of biotite and white mica with quartz and sericite. Sphene, plagioclase (An 6) and chlorite are present in small amounts. Greenstone is generally massive, yet schistose in some areas and is composed of actinolite, opaque ferruginous minerals, plagioclase (An 5), biotite and minor amounts of quartz. Estimates of thickness are unreliable because of complex folding within the nappe core, but the present thickness of Unit one is likely more than 1,000 feet.

Unit two contains light green chloritic phyllite and schist, phyllite with small lenses of dark green amphibolite and chloritic greenstone containing sheared lens-shaped bodies of possible pyroclastic origin. Chlorite, tremolite, biotite and plagioclase (An 5-7) are common constituents. Quartz, sphene, and epidote are present in small amounts. Calcite is present as thin stringers and foliae within chloritic phyllite. This calcareous phyllite forms a distinct sub-unit (2a). Foliation associated with early folding is well developed, consisting of closely spaced slaty (more properly phyllitic) cleavage and fine chloritic and micaceous
schistosity. Present thickness of Unit two, measured in section B, is about 400 feet.

Unit three consists of fine to coarse grained massive and indistinctly foliated quartzite, and massive, extremely fine grained, pure siliceous rock. The latter may be either metamorphosed chert or microcrystalline quartzite. Original texture has been obscured by extensive recrystallization. Where present, nearly coplanar foliation and compositional layering are outlined by biotite and rarely chlorite. Colors of quartzite range from white through grey to dark blue.

A band of chloritic phyllite similar to that of Unit two lies within Unit three and is designated sub-unit 3a. The total present thickness of Unit three is not known because an east-trending fault within Hester Creek cuts out or repeats parts of this section (Map 1). The greatest continuous exposure of Unit three is approximately 2,000 feet thick.

A sequence of foliated phyllitic quartzite and siliceous phyllite comprise Unit four. Colors are commonly grey, white and blue, although minor parts containing ferruginous impurities are yellow and reddish-brown. Quartz is fine to medium grained and highly recrystallized. Fine phyllitic foliae of biotite, chlorite, tremolite and ferruginous minerals outline lenticular siliceous layers. Parts of this unit are highly quartzose with phyllitic foliae under .01 inches
thick; other parts contain over 25 percent phyllite and a few bands are best termed siliceous phyllite. The last usually contains up to 20 to 30 percent pyrite, graphite, and other optically opaque minerals (unidentified) as well as minor amounts of chlorite and sericitized plagioclase (composition uncertain but possibly albite). Present thickness measured in section B is 700 feet.

Unit five contains mainly massive and irregularly foliated quartzite similar to that of Unit three. Foliated phyllitic quartzite comprises a number of minor subunits (not formally subdivided), one of which is the uppermost part of Unit five. Quartz is white and light grey in colour and medium to fine grained. Fine, often discontinous, micaceous and phyllitic foliae outline irregular lenticules of quartz. Biotite flakes, up to .04 inches across are developed within these foliae near intrusive bodies (see Chapter IV). Foliated quartzite contains a high proportion of phyllitic material that defines bands of fine grained quartz under .125 inches thick. The maximum present thickness of Unit five is about 600 feet.

Differentiation among siliceous units in sections A and B is not possible where no distinctive units such as Unit two are available for reference. Beyond Unit five to the north more siliceous units are encountered (Map 1) which are
indistinguishable from Units three, four, and five.

Section C is located just west of section B. Measured northwest, C includes the upper part of Unit four, Units five, six and part of Unit eight. Unit seven, present in other areas, is not found here. Farther to the northwest, Unit six is encountered again, followed by five, four and three. Such repetition implies the presence of a second macroscopic fold core which is delineated by Unit six and described in Chapter III (Maps 1 and 3). Outcrops of siliceous units north of Unit five in section B are now explained as repetition of Units three, four, and five by folding. Section C, like B, is overturned with respect to section A.

Unit six is composed mainly of light to dark green amphibolitic and chloritic phyllite and schist and some green metamorphosed basic igneous rocks. A distinct subunit (6a) of marble is seen in this section. Chlorite, tremolite/actinolite, lesser amounts of white mica and plagioclase (An 6), and minor quantities of biotite comprise most of Unit six. Some parts are siliceous, others contain ferruginous minerals. This unit is indistinguishable from Unit two except for the presence of subunit 6a, a white, light grey to blue, pure crystalline calcite marble found in two bands of sheared lenses 10 to 30 feet thick, less than 50 feet apart, with exposed lateral extents of under 100 feet. Extensions of these
lenses are thin calcite bands three to eighteen inches thick and thin calcareous stringers interfoliated with phyllite. Crystalline calcite is the predominant mineral; only minor amounts of dolomite, tremolite, and graphite are present. Complete recrystallization has obliterated any original features such as clastic particles or fossils. Crystals are coarse to medium grained in size, generally of the order of .05 inches across. The present thickness of Unit six in this section is about 350 feet.

In section C, Unit eight is composed of grey and white massive and irregularly foliated quartzite and microcrystalline quartzite or meta-chert similar to that of Unit three. Thickness of this unit as seen in section C is difficult to determine because of repetition by tight folding within the second nappe core.

Section D was measured within a part of the Kobau Group that is separated from other outcrops by extensive areas of Recent alluvium. Exposed thickness of all units in this area is 700 to 800 feet. The structurally uppermost unit, whose top is not seen, is composed of light to dark green phyllite and schist containing tremolite/actinolite, chlorite, and micas, and is similar to Units two and six. In three localities, pure calcite marble was observed in a band up to 20 feet thick and in thin stringers under one foot thick. The
mineralogy of this marble is very much like that of subunit 6a. On the basis of mineralogical and lithological similarities the uppermost unit of section D is correlated with Unit six of section C.

The succession below Unit six contains a number of relatively thin (under 200 feet) units of varying lithology. The first unit is composed of foliated quartzite containing variable amounts of phyllitic material. This unit is underlain by green chloritic and amphibolitic phyllite followed by massive quartzite, further green phyllite and lastly foliated phyllitic quartzite. This succession is quite distinct from that underlying Unit six in sections B and C (about 3,400 feet of massive and foliated quartzite) and is therefore placed structurally above Unit six. Section D is overturned with respect to the succession in section A.

Unit seven, immediately above six, is composed of two types of foliated quartzite. The first is a relatively pure, light grey quartzite with lenticular siliceous bands outlined by thin foliae containing white mica, biotite and rarely chlorite; the second is highly phyllitic (up to 50 percent) with thin bands of chlorite, tremolite/actinolite, and biotite. The present thickness of Unit seven is quite variable, ranging from zero feet in section C to 250 feet in the vicinity of section D.
Unit eight is made up of phyllite and massive quartzite. Most of this unit is light grey-green to dark green phyllite and schist containing fine to medium grained tremolite-actinolite and chlorite, and lesser amounts of biotite and quartz. Black ferruginous minerals, sphene and calcite are present in minor quantities. Some parts of this unit contain up to 30 percent bands and foliae of white quartzite. Massive and irregularly foliated quartzite comprising the remainder of Unit eight has been described earlier in section C. The present thickness of Unit eight is about 500 feet.

Unit nine contains grey and white foliated micaceous and phyllitic quartzite similar to that of Unit seven. Although Unit nine apparently overlies Unit eight, outcrops of it have been observed in only four localities and its true relationship to the previously described succession is uncertain. Present thickness is unknown but likely exceeds 200 feet.

OVERLYING SUCCESSIONS

Two small outcrops of brecciated siliceous sediment have been observed in the southeastern-most part of the map-area (Map 1). At the northern outcrop, the sediment lies unconformably over the Kobau Group but no contact was observed at the other. The age of both deposits is unknown.

Northwest of Blind Creek, limestone of the Blind Creek Formation is in apparent fault contact with the Kobau Group and may overlie the Group on a thrust faulted contact (R.V. Best, personal communication, 1969). Further discussion of this formation and its relationship to the Kobau Group is made in Chapter VI.
A structural succession containing nine units has been established by outcrop mapping of five rock types: massive quartzite, foliated phyllitic quartzite, chloritic phyllite and schist, greenstone and marble. The total present thickness measured in several sections is between 5,700 and 6,200 feet.

Assumptions have been made that are only partially supported by existing data. The most obvious of these is that the phyllite/marble assemblage of Unit six is unique within the map-area. Mineralogic properties of the marble (see Chapter IV and Appendix C) suggest but do not prove that this is so. The phyllite is also similar in different localities but often cannot by itself be differentiated from phyllites of supposed other units. Alternate hypotheses can be put forward, for example, that the marble unit is in fact two units, and a new model of the succession would be constructed on the basis of this assumption. The model presented is but the simplest of many possible models, one however which is consistent with available lithologic data.

It has also been assumed that no major facies changes and/or unconformities are present. Such changes in the normal sequence could invalidate interpretations of isolated parts of the established succession such as that described in
section D but because no evidence has been observed that confirms or disproves existence of these changes, their effect upon a synthesis of geological history of the map-area must be left as an (presently) insoluble problem.
CHAPTER III

STRUCTURE

INTRODUCTION

Geometry and history of deformation of the Kobau Group is based on field mapping and analysis of structural data supplemented by knowledge of the structural succession established in Chapter II and by measured sections described in Appendix B. Evidence of at least three phases of deformation has been observed on the mesoscopic scale. Tight, highly sheared near-isoclinal folds have been deformed into tight and normal folds which in turn have been cut by fractures and gently deformed by open folds. All of these structures have undergone later faulting. Parallel geometry of macroscopic structures has been confirmed by mapping and correlation of lithologic units. Measurement of linear and planar structures and determination of their age relationships was made in the field. Analysis of such data was aided by examination of microscopic structures and stereographic projection.

Structural features observed in the Kobau Group are outlined in Table 3-1.
TABLE 3-1

<table>
<thead>
<tr>
<th>Symbol</th>
<th>Structural Feature</th>
</tr>
</thead>
<tbody>
<tr>
<td>$F_0$</td>
<td>Compositional layering.</td>
</tr>
<tr>
<td>$F_1$</td>
<td>Phyllitic and schistose axial plane cleavage of near-isoclinal folds.</td>
</tr>
<tr>
<td>$L_1$</td>
<td>Fold axes, mineral lineations and foliation intersections ($F_0$ with $F_1$) associated with near-isoclinal folds.</td>
</tr>
<tr>
<td>$F_2$</td>
<td>Axial plane, strain-slip and other cleavage associated with folding affecting $F_0$ and $F_1$ surfaces.</td>
</tr>
<tr>
<td>$L_2$</td>
<td>Fold axes, crenulations and foliation intersections ($F_2$ with $F_0$ and $F_1$).</td>
</tr>
<tr>
<td>$F_3$</td>
<td>Axial planes of folds affecting $F_0$, $F_1$, and $F_2$ planes.</td>
</tr>
<tr>
<td>$L_3$</td>
<td>Fold axes and foliation intersections ($F_3$ with $F_0$, $F_1$, and $F_2$).</td>
</tr>
</tbody>
</table>

MESOSCOPIC STRUCTURES AND AGE RELATIONSHIPS

Two general classifications of folds and related structures have been observed in the map-area. The first has sub-parallel limbs and axial planes. Such tight folding has brought about transformation of compositional layering into foliation. This is accomplished by formation of axial plane cleavage ($F_1$) during initial stages of deformation (Plates 3-1 and 2) which is then brought into near parallelism with fold limbs ($F_0$) by continued deformation.
PLATE 3-1

TIGHT EARLY FOLD IN FOLIATED PHYLLITIC QUARTZITE OF UNIT SEVEN IN THE OLIVER SEPTUM. \( L_1 = 083^\circ/48^\circ \). \( F_1 = 154^\circ/49^\circ \) NE.

0 3
INCHES
TIGHT EARLY FOLD IN MASSIVE AND FOLIATED QUARTZITE OF UNIT SEVEN ONE MILE SOUTHEAST OF THE TESTALINDEN GRANODIORITE. L₁ = 133°38', F₁ = 072°40' SE, F₀ OF UPPER LIMB = 090°/45° S. HAMMER IS ONE FOOT LONG. INTER-LIMB ANGLE = 28°.

PLATE 3-2
and tightening of structures (Plates 3-3 and 4). Movement of material out of fold cores is accomplished by shearing and disharmonic folding. Limbs of tight folds are dislocated from crests (Plate 3-5). Clastic particles and crystals are sheared into lenses and recrystallized. In the extreme fold crests are sheared out and siliceous beds form extended lenticules of quartzite bounded by $F_0$ and $F_1$ surfaces and separated by thin foliae of phyllitic material, while pelitic sediments and pyroclastic volcanic rocks are generally sheared and recrystallized to finely foliated phyllitic and schistose units (Plate 3-6). Massive units having widely spaced or no bedding are tightly folded and recrystallized along $F_1$ surfaces. Foliation is well developed within basic igneous rocks whereas massive siliceous layers (originally possibly thick chert or sandstone beds) are merely recrystallized and have few visible mesoscopic structures within them.

Transformation of bedding appears to take place primarily on the mesoscopic scale. Macroscopic assemblages possess interfoliated contacts, the assemblages themselves can be traced in the field over considerable areas.

As shown in plates 3-1 to 3-6, the effects of this type of deformation were variable. Compositional layering was not everywhere transformed to foliation and relict fold
EARLY FOLD WITH SHEARED LOWER LIMB IN FOLIATED QUARTZITE OF UNIT FIVE SOUTH OF LITHOLOGIC SECTION D. PHOTOGRAPH FACES 100°.
PLATE 3-4
ISOCLINAL, SIMILAR EARLY FOLD IN SILICEOUS PHYLITE OF UNIT SIX ONE-HALF MILE NORTH OF THE MOUNT KOBAY SUMMIT.
$F_0/F_1 = 005^\circ/25^\circ W$. HALF INCH GRID ON PROTRACTOR.

PLATE 3-5
"FLOWED" SIMILAR EARLY FOLDS IN FOLIATED QUARTZITE OF UNIT 5 WITHIN THE EARLY NAPPE WEST OF THE FAIRVIEW GRANODIORITE.
PLATE 3-6

LENTICULAR FOLIAE AND REMNANTS OF EARLY FOLDS IN FOLIATED QUARTZITE OF UNIT FIVE NORTH OF THE SMALL LAKE NEAR THE MOUNT KOBAY SUMMIT. $F_o = 125^\circ/53^\circ$ SW (1), $139^\circ/41^\circ$ SW (2).
closures and even complete folds remain. These latter structures appear to be formed within competent foliated quartzite, less competent fine grained rocks are either crumpled disharmonically or sheared so as to destroy such structures (Plate 3-7).

Lineations ($L_1$) associated with tight folding are seen as aligned minerals, foliation ($F_1$) - bedding ($F_0$) intersections, fine crenulations and axes of small folds. These are best seen in phyllitic and finely foliated greenstone units. In highly siliceous units thick foliation and absence of prismatic minerals does not favour development of obvious lineation.

The second type of mesoscopic fold is distinct from the first. Cleavage, schistosity and compositional layering are together deformed about axial surfaces generally at large angles to foliation. Resulting folds are of several related kinds; kink bands, chevron and box folds (Plate 3-8) assymetric similar folds (Plate 3-9), and, particularly in thickly foliated or massive sequences, overturned and normal flexural slip folds (Plate 3-10). Symmetry of these folds is monoclinic.

Associated foliation ($F_2$) is strain-slip cleavage (Plate 3-11) and possibly fracture or other cleavage which is usually parallel to axial planes of observed folds. Lineations ($L_2$) are commonly fine crenulations and axes of
PLATE 3-7

EARLY SIMILAR FOLD OF QUARTZITE BANDS WITHIN SHEARED AND FOLIATED CHLORITIC PHYLLITE OF UNIT SIX NORTHWEST OF LITHOLOGIC SECTION D. PHOTOGRAPH FACES 070°.

FEET

0 .................................. 1.
PLATE 3-8

CONJUGATE SETS OF LATE KINK BANDS AND MONOCLINIC BOX FOLDS IN FOLIATED GREENSTONE OF UNIT SIX NEAR THE SMALL DIORITE STOCK NORTHWEST OF THE MOUNT KOBAY SUMMIT. NOTE STEEPLY DIPPING FRACTURES PARALLEL TO $F_2$. PHOTOGRAPH FACES EAST.

INCHES

0 2
ASSYMETRIC SIMILAR LATE FOLDS IN PHYLLITE WITH SILICEOUS FOLIAE OF UNIT SIX NEAR LITHOLOGIC SECTION E. PHOTOGRAPH FACES EAST. SCALE: FULL SIZE.
OVERTURNED LATE FOLD IN MASSIVE AND FOLIATED QUARTZITE AND PHYLLITIC ARGILLITE OF UNIT SEVEN ON THE QUEEN ELIZABETH II OBSERVATORY ROAD SOUTH OF THE TESTALINDEN GRANODIORITE. LIMBS ($F_0/F_1$) = 105°20' SW, 095°78' SW. $F_2 = 090°42' N$, $L_2 = 275°5'$. NOTE VERTICAL FRACTURES IN LOWER LIMB.
PLATE 3-II

LATE FOLDS IN HAND SPECIMEN OF CHLORITIC PHYLLITE OF UNIT SIX ONE-HALF MILE WEST OF THE TESTALINDEN GRANODIORITE. TOP VIEW: AXES OF LATE CRENULATIONS TREND LEFT TO RIGHT \( (L_2 = 310^\circ 25') \), CUTTING EARLY LINEATION \( (L_1 = 343^\circ 7') \) SEEN AS FAINT LINES EXTENDING FROM UPPER RIGHT TO LOWER LEFT. SCALE: HALF SIZE. EDGE VIEW: LATE CRENULATIONS WITH STRAIN-SLIP CLEAVAGE \( (F_2) = 131^\circ 37' \) SW. SCALE: 1 1/2 TIMES FULL SIZE.
small folds (Plate 3-11) and intersections of cleavage \((F_2)\) and \(F_0/F_1\) surfaces.

Age relationships between the two types of folds are not always observed, however, where seen (Plates 3-11 to 3-14) indicate deformation of near-isoclinal folds and related structures by folds with axial planes at large angles to foliation and compositional layering. Tight folds are re-folded and cut by a cleavage (Plates 3-12 and 13), \(F_0/F_1\) surfaces are cut by closely spaced strain-slip cleavage (Plate 3-14) and two generations of lineation have been seen in a number of localities (Plate 3-11).

History of deformation of the Kobau Group can therefore be subdivided into earlier (first) and later (second) phases. Whether these phases are in fact two distinct periods of deformation or merely differing aspects of one period is discussed in a following section.

Joints and other fractures, as well as open minor folds which cut and deform the above mentioned structures have been observed. Fractures are the most common and occur in several sets. Both early (Plate 3-2) and late (Plate 3-10) folds are affected. Likewise, both early and late structures are deformed slightly by gentle flexures. These fractures and folds clearly post-date early and late structures and are therefore designated as a third or latest phase of deformation.
EARLY FOLD DEFORMED BY LATE STRUCTURES IN SILICEOUS PHYLLITE OF UNIT FIVE ONE-HALF MILE NORTH OF THE MOUNT KOBAY SUMMIT. STEEPLY DIPPING $F_2$ CLEAVAGE = 156°54' SW, $L_2 = 158°12'$. 
REFOLDED EARLY FOLD IN FOLIATED QUARTZITE OF SUB-UNIT 6c ONE AND ONE-HALF MILES NORTHWEST OF THE MOUNT KOBAU SUMMIT. HAMMER IS ONE FOOT LONG.

PLATE 3-13
PLATE 3-14

LATE STRAIN-SLIP CLEAVAGE CUTTING $F_0/F_1$ SURFACES IN FOLIATED SILICEOUS PHYLLITE OF UNIT SIX WEST OF THE TESTALINDEN GRANODIORITE. PHOTOGRAPH FACES SOUTHWEST.

$F_2 = 137^\circ/32^\circ$ SW, $L_2 = 284^\circ/24^\circ$. 
FOLD SETS

Fold sets will be described from latest to earliest because the latest folds are the most obvious macroscopic structures in the map-area and because knowledge of the geometry of latest folds is needed to understand the present orientation and geometry of earlier folds.

Latest (Third Phase) Folds

The major structure of the Mount Kobau area is a somewhat irregular double plunging antiform or dome centered immediately southwest of the Fairview Granodiorite (Map 4). This structure (the Fairview dome) is illustrated diagramatically on cross section E-F-G-H of Bostock's map (1940) and is shown on cross section A'-A" of Figure 3-1 (in pocket). Foliation ($F_0/F_1$) of the northeast quadrant of the dome, in the septum of meta-sediments northwest of Oliver (the Oliver Septum), dips steeply to the northeast and east at angles between 40 and 60 degrees, steepening to the northeast and becoming almost vertical within the contact zone of the Oliver Granite. In the southeast quadrant (southwest of Oliver) dips of foliation are generally south and southeast at 40 to 50 degrees and within the southwest quadrant, to west and southwest at about 50 degrees. The general trend of foliation in the northwest quadrant is near horizontal with
gentle easterly and westerly dips. The dome is closed on slightly over three-quarters of its circumference, only the northwest quadrant is open.

Smaller macroscopic folds believed to be formed during the third phase of deformation have been mapped in the southeast and northwest quadrants of the dome (Map 4). All are open folds with northerly-trending axial planes of uncertain dip. Fold axes, defined by intersection of latest axial planes with limbs of early and late folds, have a variable plunge depending upon the attitudes of these surfaces. In the southeast quadrant, plunges are to the south at 25° to 45°. A broad synform which forms part of the flat-lying segment of the northwest quadrant of the dome, plunges northerly at 20° to 25°.

Few mesoscopic structures formed in conjunction with latest folding have been observed. Small open folds with northerly-trending axes and axial planes and certain fracture sets (described in the section on faults and other fractures) are likely related to this phase of deformation.

Late (Second Phase) Folds

Late mesoscopic structures have been described briefly in a previous section. Their orientation, and that of late macroscopic structures is illustrated by stereographic projection of field data. In all figures presented below, lower
hemisphere equal area projection has been used. Contouring of data points (lineations and poles to foliations) was done by the Schmidt or grid method (Schmidt, in Turner and Weiss, pp. 58-62, 1963).

Because late structures are presently disposed at varying attitudes on limbs of latest folds, the map-area is divided into domains of approximate homogeniety with respect to the largest northerly-trending latest folds described in the previous section. Two domains are defined, one including the west, the other the east flank of the Fairview dome (Figure 3-2, in pocket).

Stereographic projection of late structures and foliation folded by them within each of these domains indicates some spread of data which is most likely caused by latest deformation within each domain, (Figure 3-2). Consistency in trend of late structures can still be seen, however. Most late lineations and fold axes ($L_2$) plunge at angles less than $40^\circ$ southeasterly in the east domain and at angles less than $30^\circ$ northwesterly in the west domain. Attitudes of $F_2$ surfaces are variable, but most planes dip in the order of $30^\circ$ to $40^\circ$ southwesterly in both eastern and western domains. A number of $F_2$ planes dipping $20^\circ$ to $30^\circ$ northeasterly have been observed, mostly in the western domain. These are believed to be a less well developed set of conjugate cleavage surfaces which in some cases, together with the southwesterly dipping
set, make up box folds and conjugate sets of kink bands (Plate 3-8). Distribution of poles to early foliation and compositional layering \((F_0//F_1)\) illustrates folding about axes plunging southeasterly at 25° to 30° and northwesterly at 10° to 15° in the eastern and western domains respectively. Two fold limbs can be seen in the eastern domain. Northeasterly dips were obtained primarily in the Oliver septum; the projection illustrates the major late antiform which constitutes a part of the Fairview dome. Preponderance of southerly dips in both domains is a result of greater expanse of the southwest limb of this antiform.

Macroscopic late folds are delineated by changes in orientation of \(F_0//F_1\) planes and by correlation of measured sections (Chapter II and Appendix B). The largest example of such folds is the antiform mentioned above. Its axial plane strikes southeasterly and dips 60° to 70° northeasterly. The folds axis plunges northwest at 10° to 20° and east at about 40° off the sides of the Fairview dome. This antiform is flanked to the northeast in the Oliver septum by a tight overturned synform with \(F_2\) striking about 125° and dipping 60° to the northeast, (Cross-section C-C', Figure 3-1). \(L_2\) is somewhat variable and plunges at low angles to the northwest and south-east.

Tight late folds are also found in the area of outcrop described in the structural succession measured in \(V\) (Chapter II) section D.
west of the Fairview Granodiorite. These folds form a composite box-like structure over a mile across plunging gently northwest nearly parallel to topography in the area (Cross-section B-B', Figure 3-1). Mesoscopic late lineations plunge at angles less than 20° toward 290° to 315°.

Axial planes and mesoscopic cleavage (F2) dip southwest and west at 15° to 30° and northeast at varying angles. The form of this structure and its mode of formation are not completely understood.

Late folds observed in south and southwest parts of the map-area are generally of more limited extent than those discussed above, rarely greater than one-half mile from one exposed limb to the next (Maps 2 and 3). All show a consistent southeasterly trend. L2 plunges northwest and southeast at angles under 20°, while south of the Testalinden Granodiorite plunge is slightly greater to 295° to 300°. F2 is quite variable. Most planes strike southeasterly and dip southwest at angles between 20° and 50°; some dip northeast at similar angles (Plate 3-10). In all instances L2 is contained within F2.

Late folds have been formed by both similar and concentric folding. The former is most commonly observed on the mesoscopic scale although the type formed often appears to be determined by degree of closure of the fold; initial flexure forming a concentric fold which develops into a similar fold with continued application of stress (Plate 3-15). Small
PLATE 3-15

CONCENTRIC AND SIMILAR LATE FOLDS IN PHYLLITIC QUARTZITE OF UNIT FIVE NEAR THE MOUNT KOBAYU SUMMIT. HAMMER HEAD IS FIVE INCHES LONG.
macroscopic folds may also be of both types, the type being determined by competence of the rocks involved. Concentric folding took place within massive and irregularly foliated quartzite, similar folding within massive and foliated calcareous and siliceous phyllite. Large macroscopic late folds likely are concentric.

**Early (First Phase) Folds**

Early structures are ubiquitous in the Kobau Group. In most localities, only transposed compositional layering and foliation are seen (Plates 3-6 and 3-14), however, a few near-isoclinal folds have been observed (Plates 3-1 to 3-7). They have been refolded by two succeeding phases of deformation and are presently disposed at various attitudes on limbs of late and latest folds.

As in the treatment of late folds, the map-area is divided into domains of approximate homogeniety with respect to late and latest structures. Six sub-domains are defined, three within each of the two previously established domains, each including the limb of a major macroscopic late fold, (Map 4, Figure 3-2, in pocket). Northern and central sub-domains contain northeast and southwest limbs of the late anitform lying over the Fairview Granodiorite. Southern sub-domains include areas generally southwest and west of the Testalinden Granodiorite, separated from the northern part of
the map-area by regions of Recent alluvium and intrusive bodies.

A macroscopic early fold has been mapped lying on the northeast limb of the overturned late synform within the Oliver septum (east northern sub-domain, Figure 3-2). Sub-units 6a and 6b (see section H, Appendix B) outline this fold which closes to the northwest and contains Unit five in its core (Maps 2 and 3, Cross-section C-C', Figure 3-1). Limbs and early axial plane foliation are sub-parallel, dipping northeast at angles between 40° and 70°. Early lineations and fold axes \( L_1 \) plunge at varying angles to east and northeast at angles between 35° and 55° except in shallow dipping parts of the trough of the late synform where plunge attitude is 0° to 15° toward 080° and 260°.

In the west northern sub-domain, a northerly-closing tight fold has been observed lying on the northeast limb of the large late antiform described earlier (Map 3, Cross-section B-B', Figure 3-1). \( L_1 \) plunges easterly at 10° to 30°. Correlation of lithologic sections F and G with H (Map 1, Appendix B) supports interpretation of this closure as being the same as that in the Oliver septum, brought into its present position by late refolding. Such refolding could leave both fold closures aligned along the original direction of \( L_1 \) as shown by Holmes and Reynolds (1954). Easterly alignment is evident on Map 3. Flat-lying parts of late folds have not been appreciably altered in orientation by
late folding and hence should contain $L_1$ at or near its orientation prior to such deformation (Ramsey, pp. 470-471, 1967). $L_1$ within near-horizontal surfaces in both east and west northern sub-domains trends easterly (Figure 3-2), parallel to alignment of macroscopic structures. This early macroscopic fold originally closed to the north, the northwesterly closure observed in the Oliver septum being produced by late refolding and overturning (Cross-section C-C', Figure 3-1).

Extension of limbs of this structure to the south is established by correlation of measured sections (Chapter II, Appendix B) and mapping of distinctive lithology. Part of the lower limb is exposed in the vicinity of section D (Map 1) on the major late antiform whose southwest limb defines the west central sub-domain (Figure 3-2). Here, the early lower limb is refolded by a box-like late structure (Cross-section B-B', Figure 3-1). $F_0/F_1$ planes dip northerly and westerly at angles under 30° to 40°. $L_1$ lying on these planes plunges at angles under 40° to the west and southwest.

The upper limb of the northerly-closing early fold, lying on the southwest flank of the Fairview dome, is also exposed along western slopes of Mount Kobau. The limb can be traced south over the summit and eastward (as it curves about the outer flanks of the dome) south of the Testalinden Granodiorite to the Okanagan Valley (Map 2, Appendix B, sections J & K).
The core and lower limb of this early fold are exposed north-east of the Testalinden Granodiorite, outlined by Units one to four (sections A, B and I). The lower limb here is an extension of that part exposed in the west central sub-domain described above. This fold can be seen to be a refolded, recumbent nappe-like structure closing northerly with axial surfaces sub-parallel to the limbs, dipping off the flanks of the Fairview dome (Figure 3-1). The fold axis originally had an easterly trend with unknown plunge, the axial plane may have been recumbent or have dipped to the north or south. The fold is defined as antiformal to assign a structural order to the lithologic succession (Chapter II).

Such is the assumed gross structure of this fold. All parts are in fact made up of smaller macroscopic folds which have been deformed by later structures. Effects of these structures are marked within east and west southern sub-domains (Figure 3-2), in which \( F_0/F_1 \) surfaces are deformed about southeasterly and northwesterly trending axes \( L_2 = \pi F_0/F_1 \) respectively, and \( L_1 \) is spread about \( L_2 \).

The lower limb of the antiformal nappe can be traced east (sections B and C) into the east central sub-domain where repetition of the lithologic succession and exposure of a sheared early fold closure indicate the presence of a complementary synformal nappe lying structurally below the antiformal nappe (Maps 2 and 3). \( F_1 \) dips south-southeast at about 20° to 40°
and $L_1$ plunges southwesterly at about $10^\circ$ to $20^\circ$ (east-central sub-domain, Figure 3-2). The lower limb of this synformal nappe can be followed in turn west and northwest to a still lower small antiformal nappe whose closure is not seen but whose presence is indicated by further repetition of the lithologic succession (section E, Appendix B). $F_1$ planes of this lowest nappe have been deformed by latest folding (Maps 3 and 4) and presently dip to southeast, south and southwest at $30^\circ$ to $45^\circ$. $L_1$ plunges southwest and west at angles under $30^\circ$ (east-central sub-domain, Figure 3-2).

Mesoscopic structures of the east central sub-domain differ somewhat from those in areas discussed previously. As elsewhere, two generations of structures are observed, early near-isoclinal folds and sheared foliation are deformed by kink bands, chevron folds and asymmetric similar folds. However, in this sub-domain, fold axes and lineations formed by early folding are scattered within the southwest quadrant plunging at angles of less than $20^\circ$. Many linear structures formed by later folding possess similar attitudes while others plunge southeasterly at various angles. Age relationships between late folds trending southwesterly and those trending southeasterly are rarely seen. Some limited data suggest that the latter are younger and deform both early and late southwesterly-trending structures (Figure 3-3). It
FIGURE 3-3

SKETCH OF ORIENTED HAND SPECIMEN OF FOLIATED PHYLLITIC QUARTZITE OF UNIT FOUR NEAR THE MOUTH OF HESTER CREEK. FAINT EARLY LINEATION ($L_1 = 222^\circ/37^\circ$) IS CUT BY A DISTINCT MINERAL AND FOLIATION-INTERSECTION LINEATION ("LATER" $L_1 = 242^\circ/15^\circ$), BOTH OF WHICH ARE DEFORMED BY SMALL KINK BANDS ($L_2 = 098^\circ/33^\circ$). "LATER" $F_1$ CLEAVAGE ATTITUDE = $038^\circ/45^\circ$ NW. $F_0/F_1$ (TOP OF SPECIMEN) = $074^\circ/57^\circ$ SE. SCALE: FULL SIZE.
is possible that late southwest structures were formed by a final stage of deformation related to early folding.** As closure of early folds was completed, further stress could be accommodated only by folding of near-isoclinal early folds and $F_1/F_0$ surfaces about $L_1$. In the east central sub-domain (Figure 3-2), $L_1$ and "later" $L_1$ plunge southwesterly at $10^\circ$ to $20^\circ$.

In most instances, however, age relationships between southeasterly and southwesterly trending later structures are not observed, hence the above hypothesis must be considered tentative at best. As these structures are seen primarily within the core and lower limb of the major antiformal nappe and within underlying nappes described previously, it would appear that this southwesterly trending phase of deformation affected lower structural levels to a greater degree than higher ones. For the most part, the origin of these structures is not understood.

** These structures are designated as "later" $L_1$ and $F_1$ for convenience only. No genetic relationship to either early or late phase structures is implied.
FRACTURES

Faults

Rocks of the Kobau Group have been extensively fractured after completion of the three phases of deformation described in preceding sections. Many faults from small dislocations with movements of a few inches to major breaks with apparent strike-slip movements likely exceeding several hundred feet, have been observed. In many instances, large faults are seen only as lineaments across which lithologic contacts are displaced, hence, slickensides and other evidence of actual movement has not been obtained. Because of local variations in attitude of lithologic contacts it is often difficult even to estimate magnitude of apparent strike-slip displacement.

Mesoscopic faults have been observed in many localities in the map-area, particularly in fresh road cuts of the Queen Elizabeth II Observatory access road (Figure 2-2). Extensive faulting in these localities suggests the possibility of similar breakage having occurred in all parts of the map-area, however considerably less evidence has been observed in weathered out-crops. Stereographic projection of measured minor faults reveals a number of data point maxima suggesting that some of these mesoscopic breaks parallel northerly
trending faults in the Okanagan Valley (Little, 1961a) while others may represent splays from this major feature (Figure 3-4).

A number of lineaments detected in the field and on aerial photographs and topographic maps have been interpreted as macroscopic faults. Brown (1961) indicates that faults are accurately depicted as linears on aerial photographs. Such interpretation is aided by the presence of parallel minor faults and breccia zones, displaced lithologic units and structures, and topographic features which suggest fault scarps. Such evidence indicates that two well-developed sets of faults trending 150°-170° and 120°-130° have broken up folded structures of the Kobau Group.

South-southeast trending faults are parallel to faults in the Okanagan Valley (Little, 1961; Church, 1967). The Similkameen Valley may represent a smaller but similar fault-controlled valley. The best developed fault in the map-area parallel to the Okanagan system is one designated the Cawston-Richter fault passing through creeks of the same name. Evidence of faulting has been seen in Cawston Creek (Plate 3-16), steep canyon walls which may be fault controlled features are present in the upper part of Cawston Creek and the presence of a zone of fracturing and brecciation has been confirmed by diamond drill cores near the summit of Mount Kobau (J. Crawford, personal communication, 1967).
FIGURE 3-4

SYNOPTIC LOWER HEMISPHERE, EQUAL AREA STEREOSCOPIC PROJECTION OF POLES TO MINOR FAULT PLANES.
PLATE 3-16

SHEARED AND BRECCIATED FAULT ZONE SEPARATING PHYLLITE (UPPER RIGHT) FROM PHYLLITIC QUARTZITE (LOWER LEFT), BOTH OF UNIT SIX, NEAR THE MOUTH OF CAWSTON CREEK. PHOTOGRAPH FACES 150°. HAMMER IS ONE FOOT LONG.
Some displacement of lithologic units is evident but stratigraphic control is poor and the amount of movement cannot be estimated. Variations in the trend of the fault lineament suggest a dip of the fault plane in excess of $45^\circ$ to the west-southwest.

Other sub-parallel lineaments are present in the vicinity of the Cawston-Richter fault. Some show clear evidence of up to 500 feet apparent left-lateral strike-slip displacement (Map 4). Several minor faults striking approximately northerly and dipping west at about $40^\circ$ have been mapped and the steep slope of the west side of Mount Kobau may be a result of faulting along such a trend. Six closely spaced faults observed in the Similkameen Valley southwest of the map-area (Plate 3-17) also strike northerly and dip steeply to the west. In this locality, eastern blocks have moved upwards relative to western ones.

In eastern and central parts of the map-area, some south-southeast trending lineaments are also interpreted as faults primarily on the basis of lithologic changes near lineaments. Offsets derived from areas of patchy outcrop (common to much of the map-area) are uncertain. These and other lineaments of unknown affinity are illustrated on Map 4. Observed displacements of contacts are nowhere great, the largest has likely occurred along the Spotted Lake fault which extends from south of Spotted Lake to just east of the Testalinden Granodiorite. A lineament which represents a possible extension of this fault extends to a point just
PLATE 3-17

POLISHED FAULT SURFACE CUTTING MASSIVE AND FOLIATED QUARTZITE IN THE SIMILKAMEEN VALLEY WEST OF RICHTER MT. FAULT ATTITUDE = 009°/76° W. NOTE STEEPLY PLUNGING SLICKENSIDES.
west of the Fairview Granodiorite. The lineament defining the Spotted Lake fault is very distinct (Plate 3-18) and marked lithologic and structural discontinuities are present across it. It is possible that steep southeasterly plunges of late folds observed east of the fault result from rotational movement (clockwise when viewed eastward) on the fault plane.

Mapping of structures within the Shuswap Complex found on both sides of the Okanagan Valley indicates that movement along the Okanagan fault system has not been great (Ross and Christie, 1969). In the White Lake basin area north of Mount Kobau, movement along south-southwest trending faults has been described as:

"...a trapdoor-like downward rotation along west-dipping faults of the Okanagan system..."

(Church, 1967, pp. 96-99).

Faults in the Mount Kobau area presumably have similar displacements.

A number of other lineaments parallel to the Okanagan system have been interpreted as faults, but only if definite structural and/or lithologic discontinuities are present across them. All lineaments are shown on Map 4 and Figure 3-6.

Age differences among faults have not been observed. Cairnes (1940) reports faults older and younger than Mesozoic intrusions in the southern Okanagan region, however, these
PLATE 3-18

THE SPOTTED LAKE LINEAMENT, EXTENDING FROM LEFT MIDDLE DISTANCE TO RIGHT BACKGROUND NEAR HIGHWAY. TREND OF LINEAMENT = 140°. PHOTOGRAPH FACES SOUTHEAST. SPOTTED LAKE OUT OF PICTURE TO RIGHT NEAR RICHTER PASS HIGHWAY. OSOYOOS LAKE AND ANARCHIST MT. IN BACKGROUND.
PLATE 3-19

Crag and tail features on bench land at 2000 feet elevation northwest of Osoyoos Lake. Movement of ice from left to right (north to south). Strikes of foliation in area variable from northeast to southeast, dipping southerly.
have not been seen. Faulting in the map-area and surrounding areas (Church, 1967, pp. 96-99) is presumed to have taken place during one episode in early Tertiary time.

Joints

Planar joints have been observed over all the map-area. Stereographic projection of measured joint planes (Figure 3-5) indicates the presence of at least two main sets, one striking northerly, dipping steeply to east and west, the other near vertical, striking west to northwest.

Although in an area affected by polyphase deformation, numerous joint sets at various attitudes may be expected to form, observed joints appear to be of one age and are not affected by any folding. They are therefore, either related to latest folding or formed afterwards. If the former, observed joints represent sets parallel to axial planes and perpendicular to fold axes of northerly-trending folds. However, as no clearly defined age relationships have been observed among joints, it is not possible to relate any sets to other structures.

A comparison of strike frequencies of lineaments and joints is shown in Figure 3-6 and shows that lineaments striking 160° to 170° are approximately parallel to 20% of all measured joints in the map-area. Lineaments in the strike range
41 joints in the Oliver septum. Contours at 2, 4 and 8% per 1% area.

163 joints in the central third of the map-area. Contours at .5, 1, 2 and 4%/1% area.

III joints in the southern third of the map-area. Contours at 1, 2 and 4% per 1% area.

Synoptic diagram of 315 joints. Contours at .5, 1, 2 and 4% per 1% area.

Figure 3-5

Equal area projection of poles to joint planes.
FIGURE 3-6

ROSE DIAGRAMS OF STRIKE FREQUENCIES OF FRACTURES AND MACROSCOPIC LINEAMENTS.
120°-150° do not appear to correspond to any major concentration of joints, however, lineaments striking 120°-130° are approximately parallel to the strike of the late axial plane cleavage (F₂). Relationships between joints and minor faults are illustrated in Figures 3-4 and 3-6. Joint sets striking 000° and 100°-120° may be related to faults in the same strike ranges.

A number of northerly-trending lineaments within the map-area are likely produced by glacial action. Glacial striae (Plate 3-19) measured by the author and by Bostock (1940) trend 160°-180° and are therefore parallel to many lineaments.

STRUCTURAL HISTORY

Earliest recognizable folding produced large, near-isoclinal recumbent nappes, with northerly vergence and easterly trending axes (L₁). The attitude of F₁ at the end of early folding is unknown except that it was sub-parallel to limbs of the fold and had an easterly strike, and is presumed to have been relatively flat. As neither top nor bottom to the Kobau Group has been observed within the map-area, it is not possible to designate the major nappe as either anticlinal or synclinal.

At lowest observed structural levels, a post-early phase of deformation gave rise to refolding of F₁ and F₀ surfaces about axes sub-parallel to L₁.
A second phase of folding deformed early nappes into normal upright and tight overturned folds with moderate to steeply dipping axial planes \((F_2)\) and northwesterly-trending axes \((L_2)\). Separation of early and late phases of folding into distinct periods of deformation is suggested by differences in respective fold types and degree of metamorphism accompanying the two phases (Chapter IV) which imply differences in environment of deformation. Some time interval is presumably required to bring about such changes in environment as waning of metamorphism and/or erosion of overlying successions (ie. resulting in lower temperature and pressure conditions). However, transference of stress from northerly to northeasterly directions may have taken place during such changes in environment, thus, it is not possible to define these phases as distinct periods of deformation. Differences in environment may also be a result of position within the crust. Any differences in metamorphic grade at various structural levels as evidenced, perhaps, by mineralogical differences between Al.1 and Bl.1 (Winkler, 1967) sub-facies, are likely obscured by later contact metamorphism. Available information does not permit a choice between these two hypotheses.

A third (latest) phase of deformation resulted in gentle warping of early and late structures and the formation of the Fairview dome. Steep northerly-trending fractures parallel
axial planes of these folds whose axes lie at various attitudes on limbs of late structures. It is not possible to separate effects of this folding from possible effects of emplacement of Mesozoic igneous intrusions (Chapter V). The marked change in structural trend from early and late phases suggests that latest folding is a distinct period of deformation. Further evidence for this hypothesis depends upon dating of events and is presented in Chapter VI.

Faulting parallel to the Okanagan fault system affected all earlier structures. Stresses responsible for parallel faulting in the White Lake area were believed to be directed north and south (Church, 1967, p. 96) and were active in early Tertiary time. This fracturing is believed to be a distinct final period of deformation in the Kobau map-area.
CHAPTER IV

METAMORPHISM

INTRODUCTION

Study of metamorphism and microscopic structures of rocks of the Kobau Group was carried out by examination of hand specimens, thin sections, and determination of mineral content by X-ray diffractometry. Approximately fifty thin sections and fifty X-ray determinations were analyzed, some samples being tested by both methods. Temperatures of crystallization were obtained for five samples of crystalline limestone by measuring Mg/Ca ratios by X-ray diffractometry.

MINERAL ASSEMBLAGES

Mineral assemblages seen in various units of the Kobau Group are listed in Table 4-1 together with rock types in which they are found and possible parent lithologies. Without extensive chemical and mineralogical analysis, determination of parent lithologies of most metamorphic rocks can only be of a general nature. Particularly, differentiation among amphibolitic and chloritic greenstones derived from either basic intrusive or extrusive rocks, or from calcareous
<table>
<thead>
<tr>
<th>Rock Type</th>
<th>Mineral Content</th>
<th>Possible Parent Lithology</th>
</tr>
</thead>
<tbody>
<tr>
<td>Massive and foliated (phyllitic) quartzite</td>
<td>- quartz, white mica (muscovite, sericite ?), biotite</td>
<td>Quartz sandstone, chert, argillaceous sandstone.</td>
</tr>
<tr>
<td></td>
<td>- quartz, biotite, plagioclase (An 6), epidote</td>
<td></td>
</tr>
<tr>
<td></td>
<td>- quartz, biotite, garnet (spessartine ?), zoisite, sphene</td>
<td>likely peletic (shale, siltstone, greywacke).</td>
</tr>
<tr>
<td></td>
<td>- quartz, biotite, chlorite, ferruginous minerals</td>
<td>likely intrusive and extrusive igneous.</td>
</tr>
<tr>
<td>Phyllite, schist, greenstone</td>
<td>- biotite, plagioclase (An 6), sericite, chlorite, epidote</td>
<td>Limestone, limestone with basic igneous impurities, limestone-chert mixtures.</td>
</tr>
<tr>
<td></td>
<td>- chlorite, albite, quartz, biotite, calcite</td>
<td></td>
</tr>
<tr>
<td></td>
<td>- biotite, white mica, quartz, sphene, albite, chlorite</td>
<td></td>
</tr>
<tr>
<td></td>
<td>- biotite, clinozoisite, sphene, quartz, albite, sercite</td>
<td></td>
</tr>
<tr>
<td>Marble and calc-silicate</td>
<td>- tremolite/actinolite, plagioclase (An 5), biotite, chlorite</td>
<td></td>
</tr>
<tr>
<td></td>
<td>- chlorite, actinolite, calcite, quartz, albite, biotite</td>
<td></td>
</tr>
<tr>
<td></td>
<td>- hornblende, actinolite, chlorite, sphene</td>
<td></td>
</tr>
<tr>
<td></td>
<td>- albite, chlorite, tremolite/actinolite, calcite biotite</td>
<td></td>
</tr>
<tr>
<td></td>
<td>- tremolite/actinolite, epidote, albite, sphene</td>
<td></td>
</tr>
<tr>
<td>Meta-ultrabasic</td>
<td>- biotite, calcite, chlorite, tremolite, albite, sercite</td>
<td></td>
</tr>
<tr>
<td></td>
<td>- calcite, quartz, chlorite</td>
<td></td>
</tr>
<tr>
<td></td>
<td>- calcite, dolomite, tremolite</td>
<td></td>
</tr>
<tr>
<td></td>
<td>- calcite</td>
<td></td>
</tr>
<tr>
<td></td>
<td>- dolomite, calcite, quartz</td>
<td></td>
</tr>
<tr>
<td></td>
<td>- talc, magnesite, chlorite</td>
<td></td>
</tr>
</tbody>
</table>
pelitic sediments requires detailed knowledge of their chemistry. Quartzites are most often derived from quartz sandstone, chert and sometimes vein and "sweated" quartz. Where highly siliceous, obvious foliation does not always develop during deformation, however, the presence of pelitic or ferruginous material and the action of metamorphic differentiation (cf. Turner and Verhoogen, 1960) produces a distinct layering. Such layering may also be produced by folding of thin interbeds of original siliceous and argillaceous layers. Mineralogy of pelitic foliae is identical to that of meta-pelites.

Under conditions of low-grade regional dynamothermal metamorphism pelitic sediments give rise to phyllite and schist containing primarily quartz, albite, muscovite, chlorite, biotite, and epidote. Basic igneous rocks recrystallize to massive or schistose greenstones composed of chlorite, actinolite, epidote, albite, biotite, and sphene under similar conditions. Siliceous carbonates are metamorphosed to marbles and calc-silicates containing calcite, dolomite, tremolite, chlorite and quartz, depending upon what rock types may be present with limestone. Pure calcite limestone is not altered mineralogically by low grade metamorphism. (Plate 4-1). Only one outcrop of what is believed to be a metamorphosed ultrabasic intrusive rock was observed. The mineralogy of
PLATE 4-1

PHOTOMICROGRAPH OF PURE CRYSTALLINE CALCITE MARBLE (SAMPLE 11-4-107) OF UNIT 6a ONE MILE NORTH OF LITHOLOGIC SECTION D. MAGNIFICATION: 76 X. CROSSED NICOLS.

PLATE 4-2

PHOTOMICROGRAPH OF LATE KINK BANDS AFFECTING BIOTITE IN FOLIATED PHYLLITIC QUARTZITE OF UNIT FOUR NEAR THE OKANAGAN VALLEY SOUTH OF THE FAIRVIEW GRANODIORITE. QUARTZ, CHLORITE, BIOTITE, WHITE MICA. MAGNIFICATION: 18 X. CROSSED NICOLS.
such rock (Table 4-1) is explained by Winkler (P. 101, 1967):

"B 1.2 Quartz-Albite-Epidote-Biotite Subfacies...

"In metamorphised ultrabasic rocks: talc + actinolite + chlorite ± biotite ± quartz.

"In rocks undersaturated with silica, talc is replaced by serpentine in the green-schist facies. In the presence of a CO₂-bearing gas phase with an X_{CO₂} 0.05, serpentine reacts with CO₂ throughout the greenschist facies to give rise to talc + magnesite (see section 4.1)."

METAMORPHIC FACIES

Observed mineral assemblages indicate that regional metamorphism did not exceed the greenschist facies grade. Absence of andalusite or kyanite and, except in a few cases, (mentioned below) of garnet and hornblende, limit the grade to either the lower two Barrovian sub-facies (B 1.1 or B 1.2) or the lower Abukuma sub-facies (A 1.1) (Winkler, 1967). The presence of albite (An 7) provides another limit to the grade (Winkler, p. 101-102, 1967). Hornblende and one sample of garnet (likely spessartine) are restricted to areas near Mesozoic intrusive bodies and are believed to originate from contact metamorphism which attained the hornblende-hornfels facies within some inner parts, and the albite-epidote-hornfels facies in most outer parts, of contact aureoles. Mineralogy of the latter facies is generally similar to B 1.1, B 1.2, and A 1.1 sub-facies (Winkler, 1967) and it is
impossible either to define the extent of low-grade thermal metamorphism or to distinguish among these three subfacies of regional metamorphism.

Temperature and pressure conditions during metamorphism are defined by various reactions which gave rise to the observed mineral assemblages. The absence of andalusite or kyanite restricts temperatures to below 500°C at fluid (total) pressures between 1 and 4 kilobars. Further indication of temperature limits is provided by Mg/Ca ratios in calcite marble. Measured ratios indicate that temperatures during crystallization of the marble did not exceed 450°C (See Appendix C).

Contact metamorphism of the hornblende-hornfels facies grade takes place at temperatures as high as 580°C to 690°C (4 Kb.). In the Kobau Group, the limited extent of zones of this facies as well as the presence of chlorite coexisting with relatively small amounts of hornblende suggests that the reactions:

\[
\text{Chlorite + tremolite + quartz} \rightleftharpoons \text{hornblende + anthophyllite + H}_2\text{O}
\]

\[
\text{Chlorite + tremolite + epidote + quartz} \rightleftharpoons \text{hornblende + H}_2\text{O}
\]
did not go to completion and that temperatures did not remain above the lower limits of the hornblende-hornfels facies (ie. 520°C to 540°C) for long.
Metamorphic isograds within the map-area presumably exist between B 1.1 and B 1.2 subfacies (Winkler, 1967) and between albite-epidote-hornfels and hornblende-hornfels facies. Unfortunately, the B 1.1 and 1.2 sub-facies isograd is masked by albite-epidote-hornfels contact metamorphism which introduces mimetic biotite. The hornblende-hornfels facies isograd lies west of the Fairview Granodiorite, in the Oliver Septum, and south of the Testalinden Granodiorite (Figure 4-1). Elsewhere it must be presumed to lie very near the intrusive contacts.

RELATIONS BETWEEN STRUCTURE AND METAMORPHISM

With the exception noted below, available evidence indicates that regional metamorphism of the greenschist facies was coeval with early folding and shearing. Development of platey and prismatic minerals (micas, chlorite, tremolite/actinolite, etc.) is parallel to \( F_0/F_1 \) planes and \( L_1 \) directions. This development is not believed to be mimetic because minerals are aligned parallel to fold axes, rather than at random within foliation planes and because many minerals have been deformed by post-metamorphic but pre-intrusion stresses. Growth of some of these minerals at low angles to \( F_0/F_1 \) planes has taken place. This is attributed to crystallization along conjugate shear planes produced during closure of early structures, since other evidence (see Pg. 80)
FIGURE 4-1
HORNBLende-HORNfELS FACIES METAMORPHIC ISOGRAD,
MOUNT KOBAY.
indicates cessation of metamorphism prior to late folding. Elongate minerals deformed by late stresses have been observed (Plate 4-2), but no recrystallized minerals have been observed parallel to such new structures \((F_2, L_2)\); late micro-structures have been formed by mechanical, not chemical processes.

In a number of samples, particularly those from contact metamorphic aureoles of the larger granitic intrusions (see Chapter V) growth of biotite and hornblende has occurred (Plate 4-3). Biotite plates are seen to be aligned parallel to \(F_2\) as well as \(F_0/F_1\) planes. Such crystallization is believed to be largely mimetic as such mineral development is observed only near intrusive bodies and hornblende crystals are oriented within foliation planes but are not parallel to fold axes \((L_1\) or \(L_2\)).

Microscopic structures parallel and similar to larger features are commonly observed in thin section. Sheared lenticular foliae and near-isoclinal folds formed by early deformation, kinking of foliation by continued deformation about \(L_1\), as well as late folds and kink bands are illustrated in Plates 4-4, 5, and 6.
PLATE 4-3
PHOTOMICROGRAPH OF MIMETIC HORNBLENDE DEVELOPED ALONG F₁/F₂ SURFACES IN CHLORITIC PHYLLITE OF UNIT SIX IN THE CONTACT ZONE OF THE OLIVER GRANITE. CHLORITE, HORN­BLENDE, MINOR QUARTZ, FERRUGINOUS MINERALS. MAGNIFICATION: 18 X. PLANE POLARIZED LIGHT.

PLATE 4-4
PHOTOMICROGRAPH OF EARLY FOLD IN FOLIATED GREENSTONE OF UNIT SIX NEAR THE NORTHWESTERN-MOST PART OF THE FAIRVIEW GRANODIORITE. ACTINOLITE, HORNBLENDE, LENTICULAR FOLIAE OF CHLORITE AND QUARTZ, MINOR EPIDOTE AND BIOTITE. MAGNIFICATION: 18 X. PLANE POLARIZED LIGHT.
PLATE 4-5
PHOTOMICROGRAPH OF KINK BANDS AND BOX FOLDS FORMED BY FINAL STAGES OF EARLY FOLDING IN CHLORITIC PHYLLITE OF UNIT SIX, NEAR THE CLOSURE OF THE SYNFOMAL NAPPE ONE AND ONE-HALF MILES SOUTH OF THE FAIRVIEW GRANODIORITE. CHLORITE, FERRUGINOUS MINERALS, SPHENE, MINOR BIOTITE. MAGNIFICATION: 18X. PLANE POLARIZED LIGHT.

PLATE 4-6
PHOTOMICROGRAPH OF LATE SIMILAR FOLDS IN CALCAREOUS CHLORITIC PHYLLITE OF UNIT SIX ON THE QUEEN ELIZABETH II OBSERVATORY ROAD NEAR THE MOUNT KOBAU SUMMIT. BIOTITE, CHLORITE, CALCITE, QUARTZ AND FERRUGINOUS MINERALS. MAGNIFICATION: 18X. PLANE POLARIZED LIGHT.
CHAPTER V

IGNEOUS INTRUSIONS

INTRUSION PRECEDING EARLY DEFORMATION

Rocks of the Kobau Group have been invaded by igneous rocks during at least four episodes of intrusion (Map 2). The earliest of these were of two types, basic and ultrabasic. What are believed to have been basic intrusive rocks possibly of andesitic composition (Daly, 1912) were metamorphosed to actinolitic and chloritic, massive and schistose greenstones. Most are finely crystalline with platey and prismatic minerals aligned parallel to early foliation and lineation. Sphene commonly develops along such F1 planes together with pyrite, magnetite and other ferruginous minerals.

Contact relationships between intrusions and metamorphosed country rocks have been altered by early folding and shearing. Possible effects of such deformation are illustrated in Plates 3-1 to 3-6. Dykes and sills are folded and sheared into large lense-shaped bodies which are interfoliated with surrounding meta-sediments.

One outcrop of metamorphosed ultrabasic intrusive rock has been observed in apparently conformable contact with chloritic greenstone of Unit six. In hand specimen brown
rounded crystals of magnesite up to one quarter inch across are seen imbedded in a fine matrix of white and light green talc. Foliation is visible but not well developed. Thin section examination reveals the presence of chlorite in addition to talc in the matrix together with minor amounts of pyrite. Magnesite crystals are oxidized slightly to limonite and are presumably sideritic to some extent. Development of present mineralogy has been discussed in Chapter IV and original composition may have been that of a serpentinite.

It is possible that chloritic greenstone in contact with the ultrabasic intrusion has undergone reactions of retrograde metamorphism from hornfels to greenschist facies, but if this was the case, such reactions must have gone to completion and left no relic minerals as evidence.

INTRUSION FOLLOWING EARLY DEFORMATION

After the early phase of deformation but before at least part of the late phase, a number of dykes and sills were intruded into metamorphosed rocks of the Kobau Group. These intrusions have discordant contacts with foliated country rocks, contain xenoliths of foliated quartzite and have not undergone greenschist metamorphism. They were, therefore, emplaced after early deformation. On the other hand, they possess a faint foliation and, in at least one locality in the Oliver septum, have been mylonitized, indicating crystallization
prior to the onset of late folding (Plate 5-1).

These rocks contain feldspar and amphibole phenocrysts set in a finely crystalline matrix of feldspar, hornblende, biotite, quartz and ferruginous minerals. Determination of relative proportions of alkalic and calcic feldspars is hampered by extensive alteration to white mica and epidote, however, the presence of alkali feldspar phenocrysts and quartz suggests a probable classification of quartz latite.

INTRUSION FOLLOWING LATE DEFORMATION

Age

Rocks of the two satellite stocks of the Osoyoos Grano-diorite Batholith seen in the southeastern-most part of the map-area were emplaced after early folding yet apparently before the completion of subsequent deformation. Contacts with country rocks are concordant generally and locally sharply discordant. In the Osoyoos Batholith itself, which lies east of the Okanagan Valley, extensive shearing and dynamic metamorphism have been reported by Daly (1912), Campbell (1931) and Krauskopf (1941). Attitudes of resulting gneissic foliation observed by Krauskopf are similar to those of \( F_2 \) planes measured in the Mount Kobau area and may, therefore, be related to late stresses. However, Krauskopf (1941, p.16) believed that:
PLATE 5-1

PHOTOMICROGRAPHS OF QUARTZ LATITE DYKE LOCALLY MYLONITIZED BY LATE DEFORMATION IN THE OLIVER SEPTUM. MAGNIFICATION: 18 X. PLANE POLARIZED LIGHT (TOP) AND CROSSED NICOLS (BOTTOM).
"The general absence of linear structures, the frequent failure of platey structures to follow contacts, the gradational nature of the exposed contacts, all suggest that foliation in the Osoyoos batholith is not directly related to large-scale transitional movement but is due to partial granulation and recrystallization in place."

Altered, granulated and weakly foliated quartz diorite and granodiorite have been observed in the Osoyoos Granodiorite satellite stocks, but insufficient data is available to decide upon the origin of any structural features.

Both the Osoyoos and Fairview Granodiorites were considered by Bostock (1940) to be among the older major intrusions in the Keremeos map-area because they are more sheared and altered than either the Testalinden of Similkameen Granodiorites (Map 2), which present a fresh structureless appearance. Radiometric age determinations (White, et. al, 1967, and White, et. al, 1968) give ages of 110 ± 5 million years to the Fairview Granodiorite and 136-144 ± 6 million years to the Oliver Granite. The latter is also slightly foliated near contacts with Kobau rocks and altered, and furthermore intrudes a body of granodiorite similar to the Testalinden pluton north of the map-area. It seems probable that presence of alteration and foliation may be due more to conditions present during and after emplacement rather than age.

Numerous small stocks of dioritic composition crop out in western parts of the map-area (Map 2). They are similar to
dioritic phases of the Osoyoos and Fairview plutons and may be contemporaneous with them (Bostock, 1940).

Generally, all major acidic intrusions in the Kobau map-area are believed to be related to Triassic (?) – Jurassic Nelson intrusions (Little, 1961).

**Composition**

As little detailed study has been made of intrusive rocks in the map-area much information has been drawn from work by Daly, Campbell, Bostock, Krauskopf and Richards.

The Osoyoos satellite plutons are comprised of quartz diorite and granodiorite with some dioritic phases. Outcrops in the map-area are primarily altered quartz diorite containing feldspar, pseudomorphs of amphibole and quartz. Extensive hydrothermal alteration has produced chlorite and epidote from mafic minerals and albite with fine sericite, calcite and epidote from other feldspars. Sphene, magnetite and apatite are common accessory minerals. Krauskopf (1941, p. 20) stated that this quartz diorite:

"...resembles the hydrothermally altered rocks occasionally found along the margin of the Osoyoos batholith, but (that) it differs markedly from normal Osoyoos types."

however he nonetheless believed (p. 52) like Daly (1912) and Bostock (1940) that these intrusions were genetically related.

The Fairview Granodiorite is similar texturally and mineralogically to the Osoyoos stocks and although generally
less altered and foliated, is also considered to be related to the Osoyoos Batholith.

The Testalinden Granodiorite is composed of unaltered, homophanous, fine to medium grained granodiorite containing zoned plagioclase (oligoclase-andesine), microcline and quartz with large crystals of biotite and hornblende. Accessory minerals are sphene, epidote, apatite and iron ore. Both aschistic and diachistic dykes, including pegmatite, lamprophyre and diorite and granodiorite porphyry have been noted in the field but not mapped in detail. The Testalinden stock is compositionally and texturally very similar to the Similkameen Granodiorite which crops out south of Richter Mountain and west of the Similkameen Valley. Contact metamorphism of country rocks by these two intrusions is also similar, resulting in hard, black hornfelses of the hornblende-hornfels facies grade, and they are believed to be genetically related (Bostock, 1940).

The Oliver Granite is a three phase intrusion of granite and/or quartz monzonite and associated dykes. In a recent study, Richards (1968) states:

"... the Oliver quartz monzonite was formed by intrusions of granitic magma in the mesozone. The intrusion assimilated much of the metamorphosed Kobau sediments (sic) and produced the contact biotite-hornblende quartz monzonite and the more central porphyritic quartz monzonite. A resurgence of the magma which formed the above two rock types produced the central muscovite-garnet quartz monzonite."
Dykes of younger phases cut older phases. In addition, large and small quartz veins and lamprophyre dykes cut earlier intrusions.

Mode of Emplacement

All of the observed major igneous bodies within the map-area were formed by emplacement and crystallization of magma. Evidence of stoping, cross-cutting and interfingering of country rocks by intrusions has been observed along most exposed contacts (Figure 5-1, Plates 5-2,3). In a few localities, particularly near some small diorite stocks, molten material has intimately mixed with and metasomatised phyllitic country rocks. Recrystallization of phyllite and schist has taken place and in such cases the contact is a zone progressing from diorite to contaminated intrusive, partially recrystallized phyllite and finally hornfelsed meta-sediment. Assimilation of country rock by magma has taken place in the Oliver Granite (Richards, 1968).

Igneous intrusion resulted in tectonic interaction between deformed and metamorphosed Kobau Group rocks and intruding magma. Movement of magma may have been controlled in part by northwesterly-trending late structures. Such control could have produced stocks elongated (in plan view) in this direction. Such elongation is most evident in the Testalinden and Osoyoos
FIGURE 5-1.

SKETCH FROM PHOTOGRAPH OF CONTACT OF THE TESTALINDEN GRANODIORITE WITH BLACK AND LIGHT GREEN PHYLLITE OF UNIT SIX. NOTE THE STOPED BLOCK AND INTERFOLIATION OF INTRUSION AND METASOMATISED COUNTRY ROCK (STIPPLED). HAMMER IS ONE FOOT LONG.
PLATE 5-3.

CONTACT OF A LARGE GRANODIORITE DYKE NORTHWEST OF THE TESTALINDEN GRANODIORITE WITH SILICEOUS FOLIATED GREENSTONE (A MINOR MEMBER OF UNIT FIVE). NOTE SMALL D2 FOLDS TO RIGHT OF HAMMER, INJECTED APLITE VEINLETS (LOWER RIGHT CORNER), AND LATE JOINTS (DIPPING TO LEFT) WHICH CUT COUNTRY ROCK AND INTRUSIVE.
satellite bodies and is also seen in the straight west-northwest trending border of the Oliver Granite and the sub-parallel northeastern and southwestern contacts of the Fairview Granodiorite. The effect of early structures is difficult to assess since these were reoriented by late folding prior to intrusion. In some areas, however, interfingering of intrusion and country rock has taken place along $F_0/F_1$ planes. This is particularly evident in the Oliver Granite contact zone (Map 2).

Igneous intrusion also produced new structures in country rocks. In many areas early foliation ($F_0/F_1$) is concordant with intrusive contacts suggesting warping of these planes to conform to the igneous mass. Concordant contacts are present around the Osoyoos Granodiorite stocks, the south and southwest parts of the Testalinden Granodiorite, the southeast and northeast contacts of the Fairview stock and along the whole of the south-southeastern contact of the Oliver Granite.

Only a part of the contact zone of the northern Osoyoos Granodiorite stock is exposed. Where seen, however, no accommodation structures such as fans of tensional cross joints or steeply plunging folds have been observed. $F_0/F_1$ planes northeast and south-southwest of this stock strike east to east-southeast, dipping southerly and the exposed form of the pluton
suggests emplacement as a partially concordant intrusion.

No obvious accommodation structures appear to be associated with the Testalinden Granodiorite. Discordant contacts suggest movement by melting, stoping or assimilation controlled only in part by structures within country rocks. A distinct lineament, which may represent a post-intrusion fault, marks the western limit of the Testalinden pluton. Contacts have not been observed in this area and structural relationships are unknown.

Generally concordant relationships between the Fairview Granodiorite and Kobau rocks imply emplacement at exposed levels by deformation of surrounding rocks and gross movement of magma parallel to existing structures. Locally, foliation within the granodiorite is parallel to contacts indicating flow of partially crystallized material parallel to them. Such movement may be caused by convective flow in boundary layers of the pluton (Lacy, 1960). Formation of the Fairview dome (Chapter III) may be related in part to emplacement of this stock, however apart from possibly fortuitous coincidence of intrusion and dome, no evidence has been observed that supports such a relationship.

The Oliver Granite was emplaced by assimilation of country rock, upward movement of magma along $F_0/F_1$ planes and deformation of bordering metamorphic rocks. Richards (1968) indicated that assimilation produced a border phase of biotite-horn-
blende quartz monzonite from a parent muscovite quartz monzonite. Considerable interfoliation of intrusion and country rock within one-quarter mile of the southern margin of the stock has been observed.

Stringers of metasediment are found within the pluton and sills of quartz monzonite intrude Kobau rocks. Accompanying such emplacement was considerable forceful intrusion resulting in southerly directed stress. A number of examples of steeply plunging accommodation folds have been observed and the overturned late synform (Chapter III) in the Oliver septum is believed to have been tightened by a combination of vise-like forces stemming from emplacement of both Oliver and Fairview plutons.

Structural effects of many small diorite stocks seen in western parts of the map-area are difficult to assess. Near the summit of Mount Kobau early and late mesoscopic structures are reoriented in a manner which cannot be analysed with available data. Such deformation may be a result of doming in response to intrusion of the diorite body which crops out just northwest of the summit. Presence of numerous small stocks of similar composition suggests that they belong to one or more larger parent plutons underlying the Kobau Group. If this is the case, relatively little structural evidence of such bodies exists.
General spread of structural data (Chapter III) which could be caused by subsurface arching and accommodation, is more likely explained by post-tectonic faulting and original variations in early and late structures.

Intrusion and crystallization of all major granitic stocks in the map-area reached its culmination at levels in the crust where surrounding temperature and pressure conditions correspond to depths of 5 to 9 miles, i.e. the mesozone. With the exception of the Testalinden Granodiorite, which may have been intruded at somewhat higher levels (upper mesozone - lower epizone), intrusions discussed previously possess characteristics typical of mesozonal plutons as defined by Buddington (1959):

"The degree of metamorphism of the regional country rock is not more intense that the green-schist and epidote-amphibolite facies. The inferred temperature of the country rock at the time of intrusion is generally no higher than 400°-500° C. The characteristic plutons have complex emplacement relationships to the country rock - in part discordant, in part concordant. Locally there may be some replacement. Planar foliation is often well developed, especially in outer portions of the pluton. Uplift of the roof may be inferred."

Basalt dykes cutting Kobau rocks have been observed in one outcrop immediately southwest of the map-area in the Similkameen Valley (Plate 5-4). These are dark grey, unaltered sub-porphyritic olivine basalt with small plagioclase
phenocrysts and may be feeders for Triassic or older basalt flows and sills of the Old Tom Formation or for basaltic lavas of the Marron Formation of Eocene age (Bostock, 1940). The latter is more likely as these dykes have not undergone any deformation. However, no olivine basalt is reported in sub-divisions of the Marron Formation (Church, 1967) and their age and correlation must await further study.

YOUNGEST INTRUSION

Youngest intrusions in the map-area are dacite and lamprophyre dykes. The first of these, and the most common, cut Kobau metamorphic rocks and Mesozoic intrusions and often appear to parallel late faults of probably Tertiary age. Dykes are distinctive in outcrop (Plate 5-5); massive, white to light grey in colour and contain small hornblende, plagioclase and occasionally quartz phenocrysts in a matrix of plagioclase and chlorite microlites set in muscovite, leucoxene and quartz. Plagioclase (albite ?) is zoned and together with other minerals is highly altered. Feldspars are altered to white mica, mafic minerals to chlorite, epidote and iron ore. The matrix is cut by carbonate veinlets. The origin of these dykes is unknown, however, it is possible that they are related to late phases of
PLATE 5-4

TERTIARY(?) BASALT DYKE CUTTING FOLIATED ARGILLACEOUS QUARTZITE OF THE KOBAU GROUP WEST OF RICHTER MOUNTAIN. FOLIATION ATTITUDE 098°/79° S. DYKE ATTITUDE 032°/55° NW.
Mesozoic intrusions described earlier.

Only one example of a young lamprophyre dyke is known. This biotite-augite lamprophyre cuts the Oliver Granite and has been dated radiometrically at $53 \pm 2$ million years (White, et al., 1968). Other lamprophyre dykes, believed to be related to Mesozoic intrusions may in fact be of similar age but have not as yet been radiometrically dated.
OUTCROP OF DACITE DYKE ON THE NORTHWEST SLOPE OF MOUNT KOBAN. BLIND CREEK IN THE BACKGROUND. NOTE THE MASSIVE APPEARANCE AND BLOCKY JOINTING. EXPOSED THICKNESS ABOUT TEN FEET, ATTITUDE 175°/35° W.
CHAPTER VI
INTERPRETATION AND CONCLUSIONS

AGE AND CORRELATION

Introduction

The history of the Kobau Group begins with deposition of a thick succession of quartz sandstone, greywacke, siltstone and shale. Deposition was accompanied by intrusion and extrusion of basic igneous rock. Chert may have formed together with submarine extrusion and minor amounts of limestone were deposited either as detrital material or as small scale reef growth. The total thickness of the observed succession, from shales and impure quartz sandstone of Unit one to similar sandstone of Unit nine, is difficult to estimate. The present thickness obtained from various measured sections (Chapter II and Appendix) is between 5500 and 6500 feet. Any relation to pre-tectonic thickness is highly conjectural but the combined actions of thickening because of tight folding and thinning as a result of shearing and flow folding are not likely to change the order of magnitude. Original thickness quite probably did not exceed 5000 feet.
Prior to or concurrent with deformation and metamorphism, intrusion of a small ultrabasic body took place. Emplacement may have been by cold intrusion as no contact metamorphic effects have been observed. Such effects possibly were obliterated by later metamorphism.

Time of deposition can only be determined on the basis of structural and lithologic evidence. Taken by itself, the Kobau Group is known to be older than 140 million years (pre-Cretaceous) because of radiometric ages of intrusions that cut the group. Northwest of Blind Creek (Figure 1-3) the group is in fault contact with fossiliferous limestone (Angold, 1957) of the Blind Creek Formation. Little detailed work has been done here; fossil ages are given as either Permian (Bostock, 1940) and/or Pennsylvanian (McGugan, et. al., 1963). Brachiopods and fusilinids have been tentatively dated as lower Pennsylvanian (K. Sada, R.V. Best, personal communication, 1969), whereas Danner (McGugan, et. al., 1963) suggests a probable Upper Carboniferous to lower Permian age. Presence of fossils, faulting and brecciation as well as virtual absence of folds (R.V. Best, personal communication, 1969) suggests that these limestones did not undergo extreme deformation and metamorphism that affected the Kobau Group. Thus, the latter is presumably at least pre-Pennsylvanian in age. Support for this age limit comes from possibly
correlative successions found south and north of Mount Kobau.

The Anarchist Group

Previous mapping suggests that the Kobau and Anarchist Groups, at least in the Okanagan Valley, are one and the same. Lithologies are very similar and both groups have been deformed and metamorphosed. South and southeast of Kruger Mountain the two names have been applied to the same rocks. Réconnaissance traverses by the author in this area confirm this interpretation. Fossils obtained from the Anarchist Group are likely Permian and possibly Carboniferous in age (Waters and Krauskopf, 1941) and other fossils found near Greenwood and Grand Forks give Permian and/or earlier ages for lower Anarchist formations.

Three-fold sub-divisions of the Anarchist Group have been made (Umpleby, 1911; Waters and Krauskopf, 1941) but no structural mapping or analysis accompanies these and correlation of each sub-division is therefore unreliable. Calcareous shale, bedded limestone and greenstone of Middle Anarchist rocks have been observed by the author near Blue Lake west of Oroville (Figure 1-3). These have been folded at least once and bedding, not foliation is observed (Plate 6-1). Part of the Anarchist Group has undergone at least two periods of deformation (the
RECUMBENT CLOSE FOLDS IN THIN AND THICK BEDDED ARGILLACEOUS LIMESTONE AND CALCAREOUS SHALE OF THE ANARCHIST GROUP NEAR BLUE LAKE ABOUT SEVEN MILES WEST-SOUTHWEST OF OROVILLE, WASHINGTON. PHOTOGRAPHS FACE NORTHWesterly, APPROXIMATELY DOWN-AXIS.
first very severe, i.e. early Kobau) while the rest has undergone but one (presumably late Kobau). More detailed mapping should reveal an unconformity within successions presently known as Anarchist Group. Erosion is suggested by the presence of conglomerate within middle and lower Anarchist rocks (Waters and Krauskopf, 1941). It is likely, although not certain, that available fossils have been found in less highly deformed parts and that these were deposited over already deformed Kobau rocks. The age of the Kobau Group is therefore again implied to be pre-Permian, possibly pre-Carboniferous.

The Shuswap Complex

Parts of the highly deformed and metamorphosed Shuswap complex (Vaseaux Formation) crop out within two to three miles of the exposed northern limit of the Kobau Group. This formation, correlated with the Monashee Group found near Vernon (Little, 1961; Jones, 1959) has been deformed at least five times and undergone major intrusion three times (Ross and Christie, 1969). Comparison of deformation, associated metamorphism and igneous intrusion in the Kobau Group and Vaseaux Formation (Table 6-1) suggests that the Kobau Group was deposited over deformed and at least partially metamorphosed Vaseaux rocks prior to second
deformation of that formation. No separation of metamorphism has been observed between first and second phases of deformation in Vaseaux rocks (Ross and Christie, 1969) however such separation could well be obscured by high grade metamorphism accompanying the second phase.

What may be the stratigraphically uppermost group of the Shuswap Complex crops out in a northerly-trending zone approximately 25 miles west of Vernon (Figure 3-1). Jones (pp. 27, 28, 1959) describes the lithology of this, the Chapperon Group thusly:

"Argillaceous sedimentary rocks or their metamorphosed equivalents comprise the greatest part of the assemblage but rocks probably of volcanic origin are also common."

"... argillaceous rocks are ... black ... coarsely bedded ... other(s) ... are grey ... thin bedded or platey ... commonly calcareous ... all ... contain fine flakes of sericite."

"... quartzites are argillaceous ... thin bedded ... very fine grained ... light to dark grey ... others are thick bedded, fine grained and colour banded ... grey and blue-grey."

"... limestone is white, massive, thick bedded, and blue-grey weathering ... in beds as much as five feet thick ... distributed among argillaceous rocks and green schists."

"... green schists ... are finely bedded ... others contain no sign of bedding ... highly calcareous ... contain thin carbonate layers between foliation planes."
### Table 6-1

<table>
<thead>
<tr>
<th>Kobau Group</th>
<th>Vaseaux Formation</th>
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<tbody>
<tr>
<td>Deposition of the Kobau Group</td>
<td>Near isoclinal folding and associated metamorphism.</td>
</tr>
<tr>
<td></td>
<td>Trend of (L_1): northerly</td>
</tr>
<tr>
<td>Near isoclinal folding and shearing. Metamorphism to greenschist facies. Trend of (L_1): easterly.</td>
<td>Near isoclinal folding and mylonitization.</td>
</tr>
<tr>
<td></td>
<td>Metamorphism to uppermost almandine amphibolite facies. Trend of (L_2): 110°-135°.</td>
</tr>
<tr>
<td></td>
<td>Inter- or syntectonic intrusion of granite gneiss</td>
</tr>
<tr>
<td>Overturned and normal folds. Trend of (L_2): 120°-140°.</td>
<td>Overturned and normal folds. Trend of (L_2): 090°-130°.</td>
</tr>
<tr>
<td></td>
<td>Jurassic - Cretaceous intrusion.</td>
</tr>
</tbody>
</table>
Jones also states (op. cit., p. 28):

"The Chapperon group has not been subdivided into formations because of the complexity of internal structures..."

and:

"... the Chapperon group (is) ... cut by serpentinized ultramafic dykes of the Old Dave intrusions."

Lithologic similarities between Kobau and Chapperon Groups are marked, and support possible correlation. No structural analysis of Chapperon Group structures has been done and its relationships to the Mount Ida and Monashee Groups is uncertain, although Jones (1959) postulated a possible equivalence with the uppermost formation of the Mount Ida Group. A tentative Carboniferous or Permian age has been assigned this group on the basis of observed crinoid stems (Campbell, 1966). The Chapperon Group is overlain unconformably by argillite of what may be the lowermost division of the Permian and Pennsylvanian (?) Cache Creek Group (Jones, 1959; Preto, 1964; Schau, 1968). If correlation can be extended between Kobau and Chapperon Groups, such a relationship provides additional indication of a pre-Permian, Carboniferous (?) age for the former.

Origin and ages of strata, metamorphism and deformation of the Shuswap Complex are not as yet agreed upon. Ages of
strata are believed by some to be pre-Cambrian (Dawson, 1879; Daly, 1915; Jones, 1959), by others to range from late pre-Cambrian to possibly Mesozoic (Brock, 1934; Cairnes, 1939; Wheeler, 1965, 1966; Hyndman, 1964, 1968) or Palaeozoic (McConnell and Brock, 1904; Gunning, 1928; Reesor, 1957; Ross, 1968). Estimates of time of deformation and metamorphism are also in dispute. Culmination of syntectonic metamorphism may have taken place during the first period of folding (Jones, 1959; Wheeler, 1966) and/or with the second period (Ross, 1968). Some metamorphism may have post-dated deformation (Gabrielse and Reesor, 1964). Folds apparently similar to Shuswap earliest structures have been observed within Mesozoic successions (Hyndman, 1964, 1968; Cairnes, 1939) and radiometric ages of mantled gneiss domes within the complex (Gabrielse and Reesor, 1964; Reesor, 1965), indicate Mesozoic or younger metamorphism possibly accompanied by extensive deformation. Wheeler, (p. 37, 1966) states:

"The earliest recognized structures in the complex are warps and isoclinal, recumbent structures developed along east-west axes. Similarly oriented structures also occur in low-grade metamorphic rocks of Triassic and Lower Jurassic age just east of the complex ... therefore ... early structures in the Shuswap developed in post Early Jurassic time."
Preto (1967) supported such an interpretation of structural history for rocks of the Shuswap Complex in the Grand Forks area but hinted at the possibility of a still earlier (mid-Palaeozoic) period of deformation.

In the Vaseaux Formation, immediately across the Okanagan Valley from outcrops of the Kobau Group, easterly-trending structures are not the earliest, therefore earliest folding in the Shuswap Complex may have taken place in pre-Permian (Mississippian (?)) time, resulting structures (recumbent nappes) being refolded sometime during the Permian and again in the early Jurassic (Ross and Kellerhals, 1968). Such differences in interpretation of regional tectonics suggests that correlation of periods of deformation over large areas by comparison of structural trend may be unreliable, especially when two periods of near-isoclinal folding are present and when trends of early structures possibly vary over large areas. Solution of such regional inconsistencies is beyond the scope of this thesis and must be left until further detailed structural analysis has been made of the areas in question.

The Cache Creek Complex

Parts of the Cache Creek Group found in western and central parts of the Cordillera may be of Pennsylvanian age (McGugan,
et. al., 1963) and are therefore possible correlatives of the Kobau Group. Little structural analysis of the Cache Creek Group has been done and relationships with rocks of similar age are generally unknown. In the Kamloops area, Cache Creek rocks of Pennsylvanian (Morrowan) age (Danner and Nestell, 1966) are believed to contain earliest isoclinal Shuswap structures (Ross and Kellerhals, 1968) indicating a post-Early Pennsylvanian age for the Kobau Group, but little is known of these and they may in fact be early Kobau isoclinal folds.

DEPOSITIONAL ENVIRONMENT

Deposition of a mixed succession of pure and impure clastic sediments and volcanic rock can take place in a number of possible environments. In addition, tectonic "blurring" of sedimentary environments (Krumbein and Sloss, pp. 428, 429, 1963) makes interpretation somewhat unreliable. Nonetheless, comparisons with successions typical of various basins of sedimentation suggests the Kobau Group has some affinity with synorogenic geosynclinal sediments (greywacke suite). These are characterized by greywacke, shale and turbidite sandstone deposited in generally shallow to moderately deep marine basins. Spilitic and pillowed greenstones and basic tuffs, formed during sedimentation, are associated with greywackes. Lime-
stones are rare. Upper parts of typical successions contain a higher proportion of sandstone, and grade into deposits of continental origin (Pettijohn, pp. 615-622, 1957). Such a general classification is of limited use in specific areas but indicates possible conditions present during deposition. The Kobau Group resembles upper parts of the greywacke suite and was presumably deposited in water of shallow to moderate depth in a synorogenic environment.

Broad regional studies of geotectonics of the Western Cordillera suggest eugeosynclinal deposition east of the Okanagan Valley during late Palaeozoic-earliest Mesozoic time (Yates, et al., 1966). Late Palaeozoic subsidence in central British Columbia was believed to have initially produced "...local and probably isolated basins..." (White, p. 70, 1959) which coalesced to form the depositional site of the Cache Creek Complex. Interpretation of the course of deformation of the Kobau Group (Chapter III) and of the adjacent Shuswap Complex (Ross and Christie, 1969) suggests sporadic development of basins of deposition coeval with folding, within a large scale eugeosynclinal setting. Kobau sediments may have been derived from rising nappe structures developed during earliest Shuswap folding east of the Okanagan Valley and possibly from western sources now buried by Permian and Mesozoic successions.
STRUCTURAL HISTORY

The Problem of Early Deformation

Presence of extreme shearing and tight folding in the Kobau Group and its proximity to the Vaseaux Formation suggest the possibility of a phase of deformation prior to early Kobau (i.e. earliest Shuswap) having affected the former. Evidence for such a phase would necessarily be scanty because of destructive effects of early Shuswap folding at high structural levels. No attenuated folds with northerly trending axes have been observed, except where they are the result of reorientation of easterly structures by refolding. (e.g. East-northern and central sub-domains, Figure 3-2). No isoclinally refolded tight folds have been seen. North-northwesterly trending lineations related to tight folding are likely caused by reorientation as a result of later folding. Such negative evidence is naturally not conclusive, however on the basis of data gathered in this study, the Kobau Group is not believed to have participated in earliest Shuswap deformation.

Early Folding

Deformation of the Kobau Group likely began during deposition of the observed succession. The apparent pattern of continued and simultaneous deformation and deposition evident
in this area during late Palaeozoic time suggests a relatively short chronological break between earliest Shuswap folding and deposition of the Kobau Group over deformed (and likely deforming) rocks of the Vaseaux Formation.

Variation in structural trend from one period of deformation to another is difficult to explain. Northerly-trending earliest Shuswap structures are folded by easterly-trending early structures which are in turn refolded about late southeasterly axes. The first trend may be related to structures of the pre-Cambrian basement (Ross, 1968) and late structures parallel a persistent trend in most Mesozoic and younger rocks of the Western Cordillera. What then of the intervening easterly trend? It may be related to extremely large scale events which brought about changes in stress fields of provincial or even continental dimension. Such a hypothesis is too all-embracing to be of much use, besides, the time interval between earliest and early folding appears to be too short to allow great changes in geotectonic pattern. It is possible that interaction of several phases of deformation may produce variations in trend of early structures over large areas, however detailed structural data is presently insufficient to support such a hypothesis.

Whatever the cause, after deposition of the Kobau Group, presumably over a basement of the Vaseaux Formation (Table 6-1)
the tectonic framework of the region was altered so as to produce tightly appressed nappes with relatively flat-lying axial surfaces and easterly-trending axes within both successions. Metamorphism accompanied deformation, reaching uppermost almandine-amphibolite and granulite facies in more deeply buried Vaseaux rocks and middle greenschist facies in the Kobau Group. Syntectonic recrystallization and extensive shearing are characteristic of early Kobau folding which may have initiated as flexural slip but likely altered to flexural flow folding as rise of temperature reduced ductility of deforming rocks (cf. Donath and Parker, 1964). Transformation of bedding \( F_0 \) to foliation \( F_1 \), described in Chapter III, accompanied deformation.

Deformation after early nappes were fully closed and foliation fully developed resulted in refolding of \( F_0/F_1 \) surfaces about \( L_1 \) forming assymetric similar folds not unlike those produced by later tectonism.

Uplift and erosion of overlying rocks and possibly upper parts of the Kobau Group may have accompanied growth of nappe structures. No evidence of unconformities have been found, although one may exist between Anarchist/Kobau equivalent successions and younger members of the Anarchist Group. Erosion
may not have taken place until after later deformation.

Late Folding

Major refolding of early Kobau nappe structures followed their development and is believed to have affected rocks of the Cache Creek Group, portions of the Anarchist Group south of the forty-ninth parallel and the Barslow, Independence, Shoemaker and Old Tom Formations found west and northwest of Mount Kobau. Normal to tight, upright and overturned folds have been reported from most of these successions (Waters and Krauskopf, 1941; Bostock, 1940). Late Kobau deformation is presumably early Mesozoic in age, in any case older than the Oliver Granite which has been dated at $144 \times 10^6$ years (Chapter V).

Refolding of $F_0/F_1$ planes occurred about $L_2$ plunging at low angles to northwest and southeast. Folding took place in a number of styles depending upon the character of the rocks in question. Quartzite, which in some areas retains original bedding, deformed by flexural slip producing cylindrical appearing folds, which by virtue of presence of earlier folds must in fact have been slightly conic. Most commonly, however, in chloritic phyllites and schists and foliated quartzites, flexural flow folding and passive slip and flow took place. Assymetric
similar folds, chevron and box folds, and kink bands are characteristic of late Kobau folding.

Macroscopic structures are of similar type. Overturned synforms and antiforms are illustrated on Map 4 and accompanying cross-sections (Figure 3-1, in pocket). All late folding is a part of the north-westerly-trending regional structural pattern established in the western Cordillera during the Mesozoic.

Pre- or syntectonic intrusion of subporphyritic fine grained dykes of quartz latite took place at this time.

Widespread uplift and erosion followed early Mesozoic deformation in the central fold belt of the Cordillera. Approximately 65 miles north of the map-area rocks of the Nicola Group were deposited unconformably on upper formations of the Cache Creek Group (Schau, 1968). In the Okanagan Valley south of Oroville, Washington, fossiliferous upper Triassic limestone was laid down on the Anarchist Group (Waters and Krauskopf, 1941) and in the Grand Forks area middle Triassic conglomerate is found overlying the Permian Knob Hill Formation (Little and Thorpe, 1965). Minor outcrops of brecciated siliceous sediment in apparent unconformable contact with the Kobau Group have been observed in southeastern-most parts of the map-area (Map 1) but their age is unknown.
Latest Structures

Open folding about poorly defined northerly axes (latest Kobau) took place in Jurassic or later time. Few mesoscopic folds related to this period of deformation have been observed with the exception of rare open flexures with northerly axes and possibly related joint sets striking approximately north. Effects of this deformation are difficult to separate from those arising from emplacement of the Okanagan Batholith and related stocks which occurred during the late Jurassic and Cretaceous. Accommodation to igneous intrusion also produced tightening of some late structures. Associated contact metamorphism was relatively mild, reaching lower-most hornblende-hornfels facies in narrow contact aureoles. Diachistic and aschistic dykes cut Kobau rocks at this time.

Development of a conjugate set of gentle folds about northerly and easterly-trending axes has been observed in the Vaseaux Formation and also in the Nicola Group northwest of Mount Kobau. In the latter area, related uplift may have resulted in deposition of the Clapperton Conglomerate in post-early Jurassic time (Schau, 1968, pp. 13-15).

Final deformation of the Kobau Group took place during Cretaceous and Tertiary time in the form of faulting along
northerly and easterly directions. Associated minor faults have extensively brecciated many parts of the map-area. Faulting is believed to have occurred along reactivated planes of weakness associated with early and late structures and along directions parallel to faults in the Okanagan Valley (Little, 1961a). Similar faulting is likely present in the Similkameen Valley. Northerly faulting in this area may be a southern extension of the Vernon-Sicamous fault (Jones, 1959).

CONTRIBUTION

This study has provided additional knowledge of the geologic history of the Kobau Group and shown it to be a succession related structurally to both Shuswap and Cache Creek Complexes. Rocks of the group have been subdivided into nine mappable units and have been shown to possess a three phase history of deformation which has formed a ubiquitous foliation from original compositional layering. Intrusions of various types have been shown to cut the group on at least six occasions.

Indirectly, this study has suggested the possibility of a major mid-Palaeozoic deformation of the Shuswap Complex and suggested a somewhat more complex history for the Anarchist Group than was previously suspected. Detailed information has been provided for another part of the East Cordilleran Fold Belt, a part whose history also bears directly on events in the Cordillera as a whole.
Much additional study could be made of intrusive rocks in the map-area, the present work having dealt only with properties of contact zones. Detailed mapping of the Oliver Granite has revealed complex internal relationships. Similar features may be found in the Fairview, Testalinden and Osoyoos stocks. Additional radiometric ages for these intrusions should be obtained. Compositional comparisons could be made between small diorite stocks and larger granodiorite intrusions to determine any possible genetic relationships. Studies of related dykes could also be made.

Additional detailed mapping of the Kobau Group itself could be done to clarify structures in complex areas such as in the core of the major early nappe northeast of the Testalinden Granodiorite. Extension of quarter-mile mapping could be made to southern parts of the map-area and south of Richter Pass on Kruger Mountain. Analysis of many fracture sets exposed on the Queen Elizabeth II Observatory road should provide greater knowledge of geometry of late and latest structures.

Outside the map-area, investigation of relationships between Shuswap and Anarchist rocks east of the Okanagan Valley should be made, as well as structural analysis of the Anarchist Group south of the forty-ninth parallel. Finally, much work
must be done before the geologic history of Triassic formations west of Mount Kobau is understood.
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APPENDIX A

Definition of Greenstones

The term greenstone, as applied to green altered basic igneous rocks is often considered too general a classification to be useful (Glossary of Geology, 1957). As defined by Moorhouse (1959), however, it serves as a convenient label for basic igneous rocks whose extrusive, intrusive or pyroclastic origin is obscured by low to medium grade metamorphism and attendant shearing. Such rocks are sometimes termed greenschists, but as both schistose and massive metamorphosed basic igneous rocks are observed in the Kobau area such classification is not accurate, besides the term is also applied to a metamorphic facies and hence open to misinterpretation.

In this thesis, greenstones are basic igneous rocks possibly of intrusive, extrusive and/or pyroclastic origin, metamorphosed to lower and middle greenschist facies and either massive and banded or schistose. Of greenstones observed in the map-area, principal minerals are tremolite/actinolite, and albite. Hornblende, biotite, quartz, chlorite, white micas, sphene and ferruginous minerals are usually present. Primary textures have not been observed.
APPENDIX B

Description of Correlative Structural Successions within the Kobau Group

To establish continuity of the structural succession described in Chapter II and to supplement illustration of macroscopic structures (Chapter III), seven additional sections (E to K) have been measured. These sections are described below and correlated with the established structural succession (Figure B-1).

As described previously (Chapter II, p. 21), a repeated section, implying the presence of a second fold core, crops out along the northwest continuation of the line of section C. Here, Unit six is "underlain" by Units five, four and three. Still, farther to the northwest (Maps 1,2), a green phyllitic unit containing amphibolite and chlorite and including a band of marble, is observed. This unit is correlated with Unit Six and the section (E) is a further repetition of the succession described in section C. Such additional repetition suggests that a third macroscopic fold core lies between the base of section C and the top of section E. Both sections C and E are overturned with respect to Section A.

Section F lies west of the Fairview Granodiorite and contains a succession of units similar to that of section D. The structurally lowermost unit of foliated phyllitic quartzite is
FIGURE B-1
CORRELATIVE STRUCTURAL
LITHOLOGIC SECTIONS WITHIN
THE KOBAU GROUP.

Lithology.
- Massive quartzite
- Foliated phyllitic quartzite
- Phyllite and schist
- Greenstone
- Marble and calcareous phyllite
- Igneous intrusion
- Faults

Measured Sections: E F G H I J K

0 1000 5000 Feet
correlated with Unit nine. Units eight to five crop out to the east. The gross similarities between the two sections as well as the presence of distinctive pure crystalline limestone within green chloritic and amphibolitic phyllite (Unit six) support this correlation. In this area, Unit six contains a second sub-unit (6b) which is a green and black actinolite schist. On the mesoscopic scale this sub-unit consists of black and light green compositional layers deformed into sheared lenses and rootless folds. Black layers contain tremolite/actinolite and chlorite, biotite, and minor amounts of epidote and albite. Green layers are similar in composition but contain, in addition, white mica and calcite. All minerals are fine grained.

Eastward, beyond outcrops of Unit five in section F, marble and schist crop out again over a broad area bounded to the north and northeast by the Oliver Granite (Map 2). As in previous instances (Sections B, C, and E), such repetition is believed to result from tight folding on the megascopic scale, and this hypothesis is confirmed by the presence of a tight fold core delineated by Unit five.

Section G illustrates the repeated sequence which is upright with respect to section A. Marble of sub-unit 6a is further repeated three times in this section but no evidence has been found that supports the existence of more megascopic tight folding in this area. Presence of faults and parallel lineaments (Map 4)
suggests that repetition may have resulted from movement on northwesterly-trending faults (Chapter III).

Section H is measured northeast from a green phyllitic unit through foliated phyllitic quartzite into green phyllite containing marble and schist. On lithologic ground this latter phyllite is correlated with Unit six, the marble with sub-unit 6a and the schist with sub-unit 6b. The underlying succession is most like that seen in Units seven and eight. Section H is therefore overturned with respect to section A. Northeast of section H are outcrops of foliated quartzite comprising the upper-most part of Unit five, followed by more outcrops of phyllite, marble and schist of Unit six. In this area a large tight fold is delineated by mapping of the distinctive lithology of Unit six. Continuation of the succession to the northeast is terminated by the Oliver Granite.

Sections I, J, and K are southern continuations of section A and contain Units one to eight. Identification of units in this area is uncertain because of thermal metamorphism near the Testalinden Granodiorite, repetition by local folding and interruption of the sections by faults. Much of Unit four is either buried by alluvium or hidden by faulting. Unit five is thicker in this section than in section B and may be repeated either by faulting or folding, or both. Lateral changes in thickness may be caused by differences in original thickness or varying effects
of shearing and internal folding.

In this section, Unit six contains thin stringers of marble less than one foot thick which likely represent sub-unit 6a. A third sub-unit (6c) is observed. It is composed of several bands of massive and foliated quartzite under 20 feet thick within a zone about 400 feet thick. Unit six and its contained sub-units has a present thickness of about 550 feet in section I.

Dark grey, faintly foliated amphibolitic and micaceous schist seen in a number of bands is associated with Units six and eight. It is not observed in sections A to H and may be present because of a facies change. Thickness of any one band is under 250 feet and generally variable.

Outcrops in the southwest quarter of the map-area are separated from rocks of sections A to K by areas covered wholly or in part by alluvium, or intruded by later granitic stocks (Maps 1 and 2). A number of outcrops of green chloritic and amphibolitic phyllite and schist containing stringers and bands of calcite marble as well as several relatively thin (20 feet thick) bands of massive and foliated quartzite have been observed. These have been correlated with Unit six on the basis of general lithologic similarity. This unit can be traced from just south of
Cawston Creek in the Similkameen Valley to the summit of Mount Kobau (Maps 1 and 2). South and west of the summit Unit six is lost beneath alluvium and can only be tentatively traced to outcrops seen in sections J and K. Over much of this area exposures are on a foliation dip-slope, and in the rest, successions are repeated and displaced by folding and extensive faulting, thus it is impractical to set up extensive local sections for comparison with the established succession.
APPENDIX C

Determination of Temperature of Crystallization of Crystalline Calcite

Following the procedure of Morgan (1967), five samples taken from crystalline limestones in the northern half of the map-area were analysed for mineral content and then for Mg/Ca ratios by a Phillips-Norelco X-ray diffractometer using CuK\textsubscript{α} radiation and a Ni filter. Each sample was scanned four times at a rate of 2°/minute. An internal standard of silicon was used to determine angles of 2θ between the Si standard and the strong (104) peak of calcite. The curve of Goldsmith and Graf (1958), which is based on displacement of the 104 calcite peak, was used to derive Mg/Ca ratios and hence, from the curve of Graf and Goldsmith (1958) (see Figure C-1), temperatures of crystallization.

Since iron affects the 104 calcite peak more than magnesium, only iron-free limestones were used (Table C-1). These were determined by staining with a solution of potassium ferricyanide in dilute HCl; iron-bearing limestones stained blue, iron-free limestones are unaffected. This test is very sensitive for small amounts of iron.
Errors in measurement of 2θ are ± .025°, producing errors of ± .0025 Å in d spacing and ± .8 mol % in MgCO₃ content. This produces errors in temperature of ± 30° C., however, data provided by Graf and Goldsmith (1958) do not cover the temperatures below 500° C. and extrapolations to 400° C. (for MgCO₃ mol percentages below 4.0) give a broad range of possible temperatures of crystallization which exceed any experimental errors (Figure C-1).

Table C-1

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<tr>
<th>Sample No.*</th>
<th>Iron Test</th>
<th>Mineral Content</th>
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<tr>
<td>11-4-107</td>
<td>-</td>
<td>calcite, dolomite (?)</td>
</tr>
<tr>
<td>12-1-x</td>
<td>+</td>
<td>dolomite, calcite, tremolite</td>
</tr>
<tr>
<td>13-2-101</td>
<td>+</td>
<td>calcite</td>
</tr>
<tr>
<td>13-4-29a</td>
<td>+</td>
<td>calcite</td>
</tr>
<tr>
<td>13-4-44a</td>
<td>+</td>
<td>calcite, quartz</td>
</tr>
<tr>
<td>14-3-10</td>
<td>-</td>
<td>calcite</td>
</tr>
<tr>
<td>14-5-16</td>
<td>-</td>
<td>tremolite, calcite</td>
</tr>
<tr>
<td>22-1-208</td>
<td>trace (?)</td>
<td>calcite, dolomite, quartz</td>
</tr>
<tr>
<td>22-3-8</td>
<td>-</td>
<td>dolomite, calcite, quartz</td>
</tr>
</tbody>
</table>

* See Map 1 for sample location.
Only those samples (22-3-8, 22-1-208 and possibly 11-4-107) which contain dolomite (i.e. an excess of Mg) can be expected to give reliable minimum temperatures of crystallization. Morgan (1967) observed a maximum of 3.5 mol % MgCO₃ in calcites of the greenschist facies indicating minimum temperatures of 415° to 480° C. Kobau Group calcites contain up to 2.6 mol % MgCO₃ corresponding to minimum temperatures of crystallization of less than 450° C. The number of samples is admittedly small but results are not contradicted by other mineralogical data which indicate crystallization under conditions of lower greenschist subfacies.

Table C-2

<table>
<thead>
<tr>
<th>Sample No.</th>
<th>Avg. 2θ between 104 calcite and standard</th>
<th>2θ 104 calcite peak</th>
<th>dA</th>
<th>mol % MgCO₃</th>
<th>T°C</th>
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<td>11-4-107</td>
<td>1.007</td>
<td>29.473</td>
<td>3.0306</td>
<td>1.6</td>
<td>&lt; 420</td>
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<td>14-3-10</td>
<td>.986</td>
<td>29.452</td>
<td>3.0327</td>
<td>.8</td>
<td>&lt; 400</td>
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<tr>
<td>22-3-8²</td>
<td>1.043</td>
<td>29.509</td>
<td>3.0269</td>
<td>2.6</td>
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<tr>
<td></td>
<td>.949</td>
<td>29.415</td>
<td>3.0364</td>
<td>0.1</td>
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<tr>
<td>22-1-208</td>
<td>1.038</td>
<td>29.504</td>
<td>3.0274</td>
<td>2.2</td>
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<td>14-5-16</td>
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1 Silicon CuKα 2θ = 28.466
2 Two magnesium calcites present in this sample.
MINIMUM TEMPERATURES OF CRYSTALLIZATION OF CALCITE INDICATED BY MOl % OF INCLUDED MgCO$_3$.
Figure 3-1
Vertical Cross-sections, Mount Kobau Area.

Section B-B
Section C-C
Section D-D
Section A-A'

Lithologic Contacts, Faults and Macroscopic Structures Projected onto Planes of Sections to Aid Illustration of Geology.
GEOLOGY OF MOUNT KOBAU
BRITISH COLUMBIA
GEOLOGY BY A. V. OMULITCH
1965-1967

MAP 4

GEOLOGIC LITHOLOGY
- Igneous intrusions
- Limestone
- Faulted granite
- Chloritic, schistitic, phyllite
- Foliated phyllite
- Chloritic phyllite & schist
- Marble
- Massive quartzite
- Foliated quartzite
- Chloritic phyllite
- Foliated quartzite

LEGEND
- Compositional layering and early foliation.
- Late main foliation and axial plane cleavage.
- Late axial planes and axial plane cleavage.
- Axial traces of major folds.
- Faults
- Lineaments

GEOGRAPHIC LEGEND
- CENTURIES 100' INTERVALS.
- CREEKS INT/INT/INT/INT.
- MAJOR ROADS & HIGHWAYS.
- LAKES INT/INT.
- SCALE: METRIC

TOPOGRAPHIC INFORMATION FROM DEPARTMENT OF MINES AND TECHNICAL SURVEYS.
MAP 82 E/4 (E) KEREMEOS- ADVANCE PRINT 1964.
EAST DOMAIN, UPPER NET: 809 POLES TO F0/F, SURFACES AND EARLY AXIAL PLANES CONTOURED AT 0, 1, 2, 4 AND 6% PER 1% AREA.

WEST DOMAIN, UPPER NET: 746 POLES TO F0/F, SURFACES AND EARLY AXIAL PLANES CONTOURED AT 0, 1, 2, 4 AND 6% PER 1% AREA.

WEST DOMAIN, LOWER NET: 184 POLES TO F0 SURFACES CONTOURED AT 1, 2, 4 AND 6% PER 1% AREA. 230 LATE LINEATIONS AND FOLD AXES (L,= -).