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Geology of lower Bass Creek Canyon Bitterroot Range Montana

R. Ernest Anderson
The University of Montana

GEOLOGY OF LOWER BASS CREEK CANYON

BITTERROOT RANGE, MONTANA

by

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ABSTRACT

Fifteen square miles of the northern Bitterroot Range of western Montana were mapped and studied. Emphasis was placed on microscopic analysis of selected samples in an attempt to establish plausable petrogenetic relationships between the recognizable lithologic units.

The principal rocks within the area are regionally metamorphosed pre-Cambrian Belt sediments which have undergone varying degrees of reconstitution and metasomatism, and varying types of structural deformation. A westward increase in metamorphic grade is recognized as the rocks pass from units which differ only slightly from the normal statically metamorphosed Belt sediments to coarse grained gneisses of high grade amphibolite facies. The transition is rapid and is believed to be a result of a rather steep temperature gradient which existed at a relatively shallow depth. Structural, mineralogical and petrographic evidence is presented in support of this view. The present mineral composition of most of the rocks appears to be closely related to the mineralogy of the original sediments. Notable exceptions

are (1) sheet-like bodies of massive to well foliated rock which approach the composition of a quartz monzonite, (2) steep walled bodies of massive to well foliated anorthosite (andesinite). Data are presented in support of both an igneous and metamorphic (metasomatic) genesis for these two rock types. The writer favors a metamorphic origin in both cases.

At least three periods of deformation are recognized. In sequence they are (1) folding which produced east-west trending folds, (2) folding which produced north-south trending folds accompanied by extensive shearing, cataclasis, contortion, crenulation and recrystallization, and concomittant with the major stage or stages of metamorphism, (3) high angle, en enchelon faulting along the range front. The age or intensity of the first period is not known. The intermediate and major period is doubtless associated with the emplacement of the Idaho batholith which may have a time span from early Late Cretaceous to early Tertiary. Two periods of movement are recognized along the later normal faults but the ages remain in question.

Three aerially restricted types of deformation are recognized and each is associated with a particular rock type. The structures and their rock associations are (1) shearing, cataclasis and some recrystallization--frontal zone gneiss, (2) shearing, recrystallization and contortion

--sillimanite gneiss, (3) plastic type deformation shown by tight isoclinal and ptigmatic folding--grey weathering gneiss.

INTRODUCTION

LOCATION AND ACCESSIBILITY

The mapped area is situated in the northern part of the Bitterroot Range within the drainage canyon of Bass Creek, 22 miles south of Missoula and 20 miles north of Hamilton, Montana. The area is bounded on the east by the Bitterroot Valley and comprises parts of Townships 9 and 10 North, Ranges 20 and 21 West. The meridian 114° 10' W. Longitude and the parallel 46° 35' N. Latitude pass through the area. The nine western sections of the area lie within the Bitterroot National Forest and the entire area lies within Ravalli County (See Figure 1.),

The Bitterroot Valley is served by U.S. Highway 93. Bass Creek Road, a U. S. Forest Service road, provides access to the mouth of Bass Creek Canyon from U.S. Highway 93. The entire length of Bass Creek Canyon is made accessible along its bottom by Bass Creek Trail, but motor vehicles are not permitted west of the Bitterroot National Forest boundary.

PURPOSE OF THE INVESTIGATION

The main objective of this thesis is a detailed study of the geology of the eastern half of the Bass Creek Canyon area. Because most of the previous geologic work in the Bitterroot Range has been of a reconnaissance nature, it was desirable to study a small area in detail; and in so

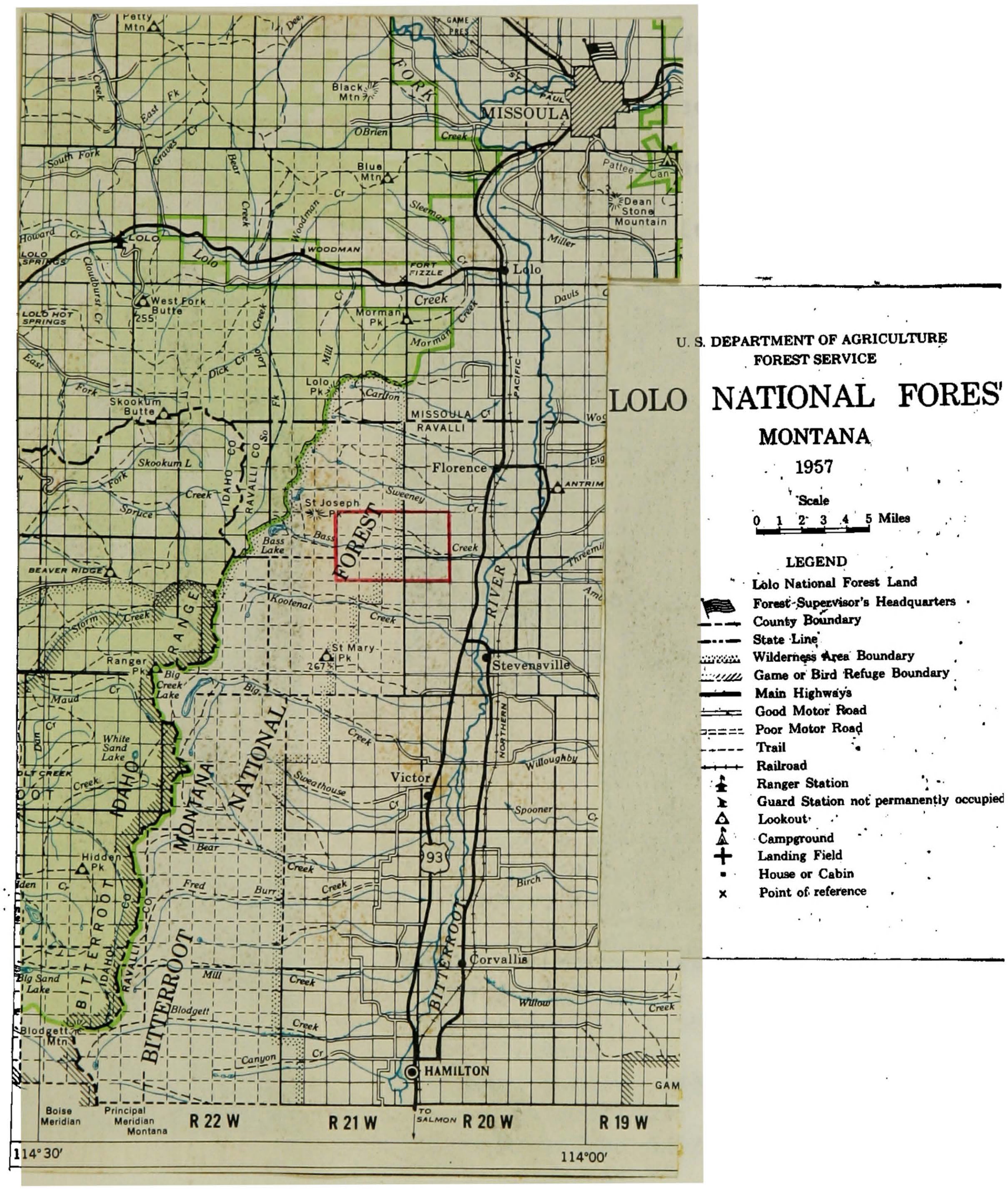

Figure 1. Index map of the northern Bitterroot Range a
vicinity. The area mapped is shown in red.

doing, obtain a more accurate delineation of the recognizable petrologic units and a clearer picture of their structural relationships. The Bass Creek Canyon area was selected because of its closeness to Montana State University and its accessibility by Bass Creek Trail. Also, it was known that a stock-like body is deeply dissected by Bass Creek Canyon and offers an excellent localle for the study of emplacement relationships.

PREVIOUS WORK

The only published geologic information concerning the mapped area is a small scale reconnaissance geologic map of the northern Bitterroot Range and vicinity compiled by Langton (1935). Langton's work has, in large part, been incorporated into the 1955 Montana State Geologic Map on a scale of 1:500,000. Groff (unpublished M.A. Thesis, 1954a; 1954b) mapped Kootenai Creek Canyon, the adjacent canyon to the south. On the basis of a brief study of Bass Creek Canyon, Groff compiled a reconnaissance geologic map on a scale of 1:62,500 extending his interpretations from Kootenai Creek Canyon northward to Bass Creek Canyon. The text of Groff's report (1954a) includes the only written description of the Bass Creek Canyon rocks prior to the present study. He describes the sequence of changes recognized on a westward traverse along the canyon and also briefly discusses the Bass Creek Canyon stock and its structural relationships.

Geologic investigations by Lindgren (1904), Pardee (1950), and Ross (1950) are mainly concerned with the genesis of the long, straight, eastern flank of the Bitterroot Range. Taken together, these reports represent a general survey of the range front from which the writer has drawn freely of the descriptions pertinent to the present study.

PRESENT STUDY

Field study was conducted during August and September, 1958 and several days during the spring of 1959. Thirty seven days were spent in the field.

Geologic data were plotted directly on aerial photographs having a scale of 1:20,000. The photographs were taken on July 10, 1937 and were secured from the regional office of the U.S. Forest Service, Missoula, Montana. Rugged terrain, limited time, and the plague of gradational boundaries did not permit the walking out of lithologic or isogradic contacts. Traverses were made from the bottom of the canyon to the ridges, and contacts were projected from traverse to traverse. In questionable areas lateral traverses were made along the canyon wall. Foliation, lineation and joint attitudes were recorded as often as expedient, and oriented rock samples for laboratory study were collected from key locations. Reconnaissance traverses were made westward to the head of Bass Creek Canyon and along the eastern parts of Kootenai and Sweeney Creek Canyons (See Figure 1 for all locations.).

Because no suitable base map was available for the area, a planimetric map was made directly from the aerial photographs by transferring all physiographic, cultural and geologic data from the photographs with a radial planimetric plotter. The scale of the new map is approximately 1:17,850. The section line grid was drawn in using known points in the eastern half of the area and projecting the lines westward. The map was then contoured by plotting known points of elevation taken from Map 21, Forest Atlas, Lolo Folio, Montana, 1927 and sketching the contours with the aid of physiographic features transferred from the photographs. The method certainly is not desirable and was employed only because it best preserved the accuracy of the geologic mapping. An east-west, restored, geologic structure section was constructed along the north wall of the canyon. Lithologic boundaries along with corrected foliation attitudes were projected into the line of section from the canyon wall. The upper limit of the section is the restored range profile as it now occurs along the spur between Sweeney and Bass Creek Canyons.

Laboratory investigations during the winter and spring of 1959 involved the following: (1) preparation and study of 52 thin sections,* (2) study of many rock samples with the binocular microscope, (3) modal analysis of 25 thin sections according to the method outlined by Chayes

* The writer is indebted to the Department of Geology, Montana State University, for funds covering the preparation of twelve thin sections.

(1949), (4) x-ray analysis of a suite of biotite grains, (5) x-ray analysis of plagioclase grains, (6) stereographic analysis of joint systems, (7) photomicrographing of selected petrographic relationships.

ACKNOWLEDGEMENTS

The writer gratefully acknowledges the assistance, both in the field and laboratory of Dr. J. P. Wehrenberg. Also, the comments and criticisms of Doctors R. M. Weidman and J. Hower have aided greatly in the completion of this thesis. Thanks are extended to Doctor R. L. Konizeski for giving the writer access to his aerial photograph mosaics and his unpublished maps of the Bitterroot Valley.

PHYSIOGRAPHY

TOPOGRAPHY

Physiographic boundaries for the Bitterroot Range were established by Lindgren (1904, p. 13) and have been accepted in later geologic reports. The range is bounded on the north by Lolo Creek and extends southward for 60 miles to the Nez Perce Fork of the Bitterroot River. The Bitterroot Valley forms the eastern boundary and the range merges westward with the Clearwater Mountains.

The eastern part of the range is characterized by a north-south trending, east sloping front which is deeply incised by some twenty east draining canyons, of which Bass Creek Canyon is one. The higher regions of the range show effects of severe alpine glaciation accomplished during Wisconsin time (Alden, 1953). In the northern part of the range the eastern ends of the canyons have a typical stream eroded cross profile, while their western ends are characterized by U-shaped troughs. The lower 1000 feet or so of the mouths of some of the canyons (vis. Kootenai, Sweeney and Carlton Creek Canyons) are characterized by steep-walled notches which give the impression of having been formed by relatively recent downcutting (See Groff, 1954a, p. 37). Most of the mountain summits near the heads of the canyons have elevations between 9,000 and 10,000 feet, significantly higher than neighboring mountains.

There is a maximum relief of 5,400 feet within the mapped area. The highest point is located along the north margin of Section 26; the elevation is 9,000 feet. The most pronounced topographic feature in the area is the serrate ridge which forms the spur between Kootenai and Bass Creek Canyons. Several small glacial cirques with headwalls over 1,000 feet in height are carved in the quartz rich gneisses of the spur. Why these cirques are so well developed in this one small area is an interesting geomorphologic problem for which no solution is offered here. In addition to the cirques, numerous gullies cut the steeply sloping canyon walls; their positions are often joint controlled.

CLIMATE AND VEGETATION

Annual precipitation varies considerably within the mapped area. The higher regions have an estimated annual precipitation of about 25 inches which decreases to about 15 inches at the lower elevations of the Bitterroot Valley, according to data obtained from the regional office of the U.S. Forest Service, Missoula, Montana. Precipitation occurs largely during the winter months as snow. No measured temperature data were obtained but the temperature range probably approximates that characteristic of the Northern Rocky Mountain physiographic province. In general, this province has a cool-temperate, semi-arid climate.

The area has a forest vegetative cover comprised of many types of coniferous trees, some of which are ponderosa, lodgepole and white pine, spruce and larch. Forestation is absent on the mountain summits and also on the steeper slopes of the canyon walls.

GENERAL GEOLOGY

The mapped area lies entirely within the Bitterroot Range which is a part of the northern Rocky Mountain physiographic province. Metamorphic rocks compose most of the eastern part of the range from its straight, steeply dipping front westward for some nine miles to the crest of the range where the rocks grade westward into the massive phases of the Idaho batholith. The massive rocks of the batholith are not exposed within the mapped area. Tertiary sediments and Quaternary alluvium of the Bitterroot Valley extend up to the main range front except for some bedrock exposures in the low foothills which hug the toe of the range. The rocks within the area are essentially metamorphosed pre-Cambrian Belt sediments which have undergone various degrees of structural deformation, metasomatism, and granitization but still reflect, to a considerable extent, the compositions of the original sediments.

On the basis of limited identification of scattered Beltian outcrops on the western flank of the Bitterroot Valley (Langton, 1935; Ross, 1950), and on the basis of the petrology and gross mineral composition of the metamorphics, it is believed that the metamorphics are equivalent to either the Ravalli Group or to the Pritchard formation or both.

Three major metamorphic rock types are recognized within the area mapped. The most abundant is a sheared and contorted, red weathering, biotite-sillimanite (bic muscovite) gneiss located in the central and northweste parts of the area. In places it has been partially grε tized to form a more massive rock with a quartz monzoni composition. The sillimanite gneiss is also the princi rock type recognized on a reconnaissance traverse from western boundary of the mapped area to the crest of the Groff (1954) also recognized the sillimanite gneiss as principal rock type in the western portion of the Koote Creek drainage as did the writer in the Sweeney Creek d age. It was not noted by Ross (1950) in his mapping of Hamilton quadrangle to the south. Ross, however, was m concerned with the eastern front of the range and may n have studied the rocks to the west in great enough deta The second major rock type is located in the southwest portion of the area. It is a grey weathering gneiss which is more quartzose and reflects a different type of structural deformation than the sillimanite gneiss. Isoclinal and ptygmatic folding are dominant and shear fects are nominal. It is, in general, more equigranula and more evenly banded than the sillimanite gneiss. Th least abundant of the major rock types is the frontal zone gneiss which forms a relatively thin, east dipping sheet of foliated, sheared and cataclastized rock in th eastern part of the area. These rocks show considerabl petrologic variation. The principal units are (1) part ally reconstituted quartz rich Belt sediments, (2) coar grained, mica rich gneisses and schists, (3) quartz ric diopside bearing metasediments.

Several anorthosite (andesinite) bodies are exp(ed within the thesis area. The two larger bodies appeai to be bun-shaped in east-west cross section. The struc- tural outline of the smaller bodies is not known. Some- what similar rocks have been described by Hietenen (195(from the northern margin of the Idaho batholith in Idahc Their occurence has not been recorded in other areas in the Bitterroot Range nor has Ross observed them elsewhei in the range (personal communication, 1959).

Three periods of structural deformation are recc nized. The intermediate and major period is responsible for the existing major structure and physiography of the Bitterroot Range. Certainly the emplacement of the Idah batholith has been the critical factor in this deformati This great batholith lies at the junction of structures related to both Nevadan and Laramide orogeny (Eardley, 1951, p. 313) and thus its age is controversial. From a study of the southern portion of the batholith between Boise and Hailey, Idaho, A. L. Anderson (1952) concludes that the batholith is composed of discrete masses of gra itic rock assignable to one or the other of two distinct age groups. The older group represents deep-seated intr sions of the Nevadan orogeny and the younger represents shallower intrusions of probable Laramide age. Lead alp age determinations obtained by Larsen et al., (1958) on 16 samples from the batholith give a mean age of 108±12 million years which is equivalent to the great batholi of the Nevadan orogeny. On the basis of these values l sen assigns an early Late Cretaceous age to the emplac ment. Values of 53 million years (Chapman et al., 195 60 million years (Larsen et al., 1958) and 42 million years (Hayden and Wehrenberg, 1959) have been recorded from rocks of the border zone of the Idaho batholith wi the Bitterroot Range. These data serve to point out th unsolved state of the age problem, which will not be di cussed further here.

The major structures within the thesis area are shown on Plate 1 and are discussed, together with minor structures and textures, in a later section.

PETROLOGY AND PETROGRAPHY

Frontal Zone Gneiss

The eastern front of the Bitterroot Range is co posed of an east dipping sheet of laminated and foliate metamorphic rock. Previous workers have been in consta disagreement on the genesis of the eastern front of the range, as well as the central portion, and thus the lit erature is filled with a complex nomenclature relating the rocks of the eastern front. In the present paper t term "frontal zone gneiss" is applied as adapted from Groff (1954a, p. 21). The frontal zone gneiss is equivalent to the "border zone gneiss" of Ross (1950, p. 15 and probably is equivalent to the "older gneiss" of Lan ton (1935, p. 39).

The minimum thickness of the frontal zone gneiss in the vicinity of Bass Creek Canyon is 1000 feet as measured normal to 25° east dipping planar features. Groff (1954a) estimated a thickness of 2700 feet in Kootenai Creek Canyon.

Lindgren's much quoted description of equivalent rocks from Mill Creek [now called Kootenai Creek] is worthy of restatement here:

> "The normal gneiss from Mill Creek, two miles above the mouth of the canyon, is a plainly schistose rock with large orthoclase crystals pressed into partly lenticular shape. Biotite and a little muscovite lie in flat aggregates between streaks of pressed feldspar and quartz. There is much cataclastic action and formation of new allotriomorphic aggregates along wavy lines, which indicate schistosity. Large feldspar and quartz grains, when crossed by these lines are greatly crushed." (Lindgren, 1904, p. 21).

To adapt this well-phrased description to the rocks of Bass Creek Canyon the term orthoclase should be replaced by oligoclase because the principal porphyroblasts are plagioclase of that composition. Ross (1950, p. 154) was impressed by the sedimentary features of the frontal zone. In his words, "The impression that the outer slopes and spur ends are stratified persists and, in most localities on the border of Bitterroot Valley, is strengthened when the range is approached and finally entered... . The major laminae are generally several inches to a foot or, rarely somewhat more in width. They are subparallel but tend to lense out along the strike somewhat like the bedding in the quartzitic argillite and argillitic quartzite that make up much of the Belt series." Ross also noted the presence of relict cross bedding (p. 154), oval, detrital grains of plagioclase and quartz (p. 163), and gradations between rock with an obvious sedimentary texture and that which has been completely recrystallized (p. 164). Groff (1954a) and Ross (1950) both noted that the quartz content is too high for the rocks to be anything but metasedimentary. Data collected in the present study serve as additional confirmatory evidence of the observations and conclusions of previous workers as stated above. The writer has observed all of the features mentioned except oval, detrital grains of plagioclase and quartz.

Groff (1954a) placed considerable emphasis on rounded zircon grains as indicators of metasedimentary origin. Whether or not zircon grains can be used as petrogenetic indicators is a controversial subject (See bibliography in Vintinage, 1957.). Because of the refractive and resistive properties of zircon, the writer believes that an extensive comparative study of grain morphology and color of the Bitterroot zircons and zircons of equivalent Belt sediments would be necessary before the criterion could be used.

The most convincing evidence for a metasedimentary origin of the entire frontal zone gneiss obtained in the present study is the gradational relationship from rocks which are definitely metasediments to rocks which retain no sedimentary characteristics. As a westward traverse made along the north wall of Bass Creek Canyon the sequ of changes which the rocks reflect is charcaterized chi ly by the development of a more pronounced gneissic tex ture as muscovite and biotite increase in abundance and segregated character. A westward decrease in quartz an increase in mica is recognized. The modal analyses for rocks of the frontal zone gneiss presented in Table 1 il lustrate these trends. Samples 8-4-1 and 9-16-6* are fı the mouths of Bass and Sweeney Creeks respectively. Th values for quartz in both samples is strongly suggestiv of a sedimentary rock, as is the texture, which resembl that of a partially recrystallized para-quartzite. Sutı quartz boundaries characteristic of quartzites are a coı feature. Samples 8-6-2 and 8-13-6 are from the western portion of the frontal zone and sample 9-1-7A is an equi valent rock from west of the frontal zone on the upper portion of the spur. They show a pronounced increase iı mica content and a reduction in quartz over the samples to the east. In the first two samples the quartz conteı is higher than one would expect in an igneous rock. Th abrubt westward increase in mica content from samples 8-4-1 and 9-16-6 to the three samples to the west was partially resolved by studies made with the binocular microscope on samples collected from intermediate loca-

* The locations of all samples discussed in the text aı shown on Plate 2.

Rock type	Sample Number	Loc.	Modal Percentages								
			Qtz	Orth	Plag	Bio	Musc	Sill	Opq	Zir	Gar
Sillimanite Gneiss	*M-7	K.C.	39	1	24	20	5	10	Tr.	Tr.	-
	*M-9	K.C.	40	13	10	20	< 1	15	Tr.	Tr.	Tr.
	*M-61	K.C.	40	5	19	18	6	11	Tr.	Tr.	-
	8-14-1	B.C.	38	5	12	21	8	15	1	Tr.	-
	*B-3	B.C.	34	1	25	19	2	16	2	Tr.	-
Frontal Zone Gneiss	8-4-1	B.C.	63	10	23	< 1	4	-	2	Tr.	-
	9-16-6	B.C.	54	5	23	-	6	10	1	Tr.	-
	9-9-1	B.C.	53	Tr.	4	20	22	-	1	< 1	-
	8-6-2	B.C.	43	4	21	13	19	-	1	Tr.	-
	8-13-6	B.C.	49	1	14	17	19	-	Tr.	Tr.	Tr.
	9-1-7A	B.C.	33	1	20	16	11	-	Tr.	Tr.	-

*Samples collected and prepared by Groff, 1954.

K.C. - Kootenai Creek Canyon
B.C. - Bass Creek Canyon

Table I. Modal analyses for frontal zone and sillimanite gneisses.

tions. These intermediate samples have mica contents which are somewhere between the extremes indicated by the modal analyses. The mineralogy of sample 9-9-1 is anomalous when considered in the light of these observations. It was collected from the foothills to the east of the main range front. Poorly understood fault relationships in this area may be responsible for its presence. From these observations it is concluded that the frontal zone gneiss in the vicinity of Bass Creek Canyon comprises a transition sequence representing a westward increase in metamorphic intensity which has impressed textural and mineralogical variations on the original sediments. The transition is not an unbroken serial gradation but rather a general trend. Variations in the mineralogy of the original sediments have doubtless been critical in producing the existing mineralogy. Indicative of this fact is the presence of diopside bearing rocks about 1000 feet west of the canyon mouth. These rocks must represent lime-silicate zones in the original siliceous sediment. They are discussed in more detail in a later section on special rock types (p. 39).

Because the frontal zone gneiss is transitional, it is difficult to establish a suitable western boundary for the unit. The problem is largely one of locating an isograd line between the quartzofeldspathic - biotite-muscovite gneiss of the western portion of the frontal zone and the sillimanite gneiss. This problem can be appreciated by comparing the modal analyses (Table 1) of samples 8-6-2, 8-13-6 and 9-1-7A from the frontal zone with values for the sillimanite gneiss. Practically the only difference is between the percentage of muscovite and that of sillimanite. Texturally the two rock types are very similar.

Typical unzoned, metamorphic pegmatites are common in the frontal zone. They are generally small concordant lenses or pods composed principally of quartz and sodic plagioclase. According to Groff (1954a, p. 25), "Pegmatites have been intruded along and across the foliation of the gneiss of the frontal zone (Figures 6 & 7)." Groff has pictured cross cutting pegmatites in Kootenai Canyon which are unlike any observed by the writer in Bass Creek Canyon. He also described sheets of igneous rock within the frontal zone gneiss. It is possible that igneous intrusion has given rise to some of the pegmatite material described by Groff and that equivalent material is not present in Bass Creek Canyon. As stated above, a much greater thickness of frontal zone gneiss occurs in Kootenai Creek Canyon than in Bass Creek Canyon, and therefore it is not surprising to find rock types exposed there which are absent in Bass Creek Canyon. A reconnaissance traverse up Kootenai Creek Canyon revealed a striking difference in lithology between the frontal zone rocks exposed there and the frontal zone rocks of Bass Creek Canyon.

Sillimanite Gneiss

The principal rock type in the area is a red weathering, quartzofeldspathic biotite-sillimanite gneiss of high grade amphibolite facies. The unit is equivalent to the "granitic rocks of the Idaho batholith" of Ross (1950), and to the "contorted gneiss" of Groff (1954). These rocks are well foliated, sheared, coarse to medium grained, porphyroblastic, augen gneisses which are generally contorted and crenulated. Slickensides and sillimanite needles and bundles are aligned in an east-west direction, and most of the small fold and crenulation axes are aligned in a general north-south direction.

Quartz-feldspar segregations in the form of augen, pods, bands, and lenses are abundant. Their size range is from pea-sized augen or clots, up to unzoned pegmatite bodies a few hundred feet in length (Figure 3). Some are foliated and some are not. One of the larger pegmatite bodies was studied and found to be composed of quartz 20%, oligoclase 75% and orthoclase 5%. A thin section cut across an augen has a similar composition; quartz 25%, oligoclase 70%, and orthoclase 5%. A striking characteristic of the sillimanite gneiss is the absence of basic segregations in the form of amphibolite bodies. Such segregations commonly occur in rocks of similar description from many points on the globe.

Modal analyses of five thin sections of the sillimanite gneiss from Kootenai and Bass Creek Canyons are given in Table 1. Note the consistent values for quartz, biotite and muscovite + sillimanite and the paucity of garnet. Values for quartz and plagioclase are probably high because of selective sectioning of well foliated rocks. That is, if a proportionate number of counts were made on sections of the quartz-feldspar segregations just discussed, the values for quartz and plagioclase would be increased significantly with respect to the other major components. No quantitative correction could be made for this.

Field study indicated that the sillimanite content decreases upward from the canyon floor and is zero at or near the upper regions of the spurs where the rocks are quartzofeldspathic-biotite-muscovite gneisses equivalent to the frontal zone gneiss. Because of difficult terrain and limited time no attempt was made to map an isograd boundary between these units. The lower grade rocks contain about 18 percent muscovite which is approximately equal to muscovite + sillimanite in the higher grade rocks. Also, quartz is somewhat higher in the non-sillimanite bearing rocks.

Certain petrographic features are worthy of note. In most thin sections shearing is evidenced by alignment of sillimanite needles, elongation of quartz, and by cataclastic effects. Quartz commonly shows undulatory extinction. Sillimanite is best developed in bundles along biotite rich foliation planes where shearing is evident but it also occurs as discreet needles or hair-like aggregates disseminated through quartz and feldspar grains (Figure 2). Oligoclase feldspar is rarely zoned but it is generally polysynthetically twinned and often shows undulatory extinction and bent twin lamallae. Where zoning is observed it is either normal or in the form of a calcic core rimmed with a sodic border which becomes more calcic outward.

Figure 2. Section across crenulation in sillimanite gneiss showing shear planes with sillimanite. Crossed nicols, x 17.

Figure 3. Sodic segregations in sillimanite gneiss. Note size range; largest is about 1' thick.

Sillimanite bearing metamorphosed Belt sediments have been described from the western border of the Idaho batholith by Leonard (1957), and from the northern border by Hietenen (1956) and Anderson (1940). On the basis of structural and mineralogical evidence, Groff (1954a) concluded that the sillimanite gneiss of the northern Bitterroot Range is equivalent to the Prichard formation of lowermost Belt Series.

Partially Granitized Metasediments

The rocks grouped under this heading are, for the most part, moderately foliated metasediments which show evidence of an approach toward a graditoid texture and a quartz monzonitic composition. Their main occurence is within the sillimanite gneiss as large, sheet-like bodies a few feet to several hundred feet thick and as much as two miles in length. Plate 1 shows the aerial distribution of several of the bodies. It was only possible to delineate the larger ones, and the boundaries of these are in many cases questionable. A great many more are known to exist within the mapped area but delineation of these would be a very tedious and time consuming project. The vegetal cover of the south canyon wall would make such a project impossible there. The rock is distinguished in the field by a more granitoid texture and by a rather smoothly weathered surface much different from the platy and slabby weathering surface produced on the sillimanite gneiss (Figures 4 & 5). All stages of gneissic structure from a fine- to medium-grained, essentially massive rock to a coarser grained, well foliated porphyroblastic gneiss were observed. Only rarely are these rocks devoid of foliation and in most cases the massive or slightly foliated phases occur towards the center of the bodies. The upper and lower contacts are concordant, and in many cases, especially in the larger sheets, the texture grades so impercep-

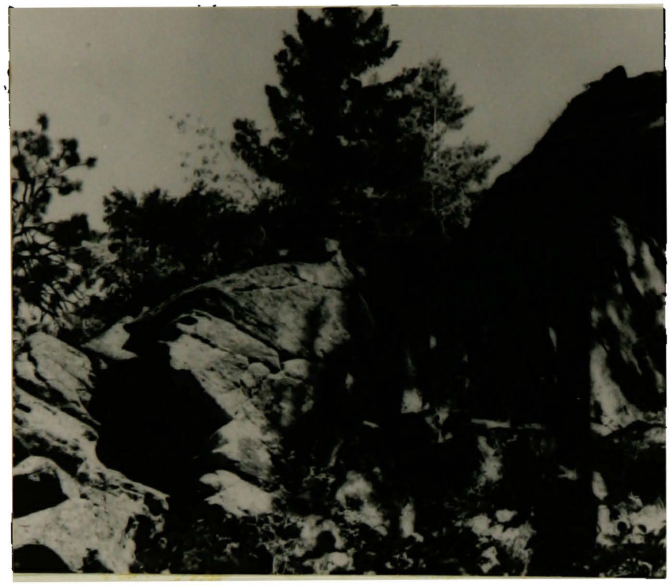

Figure 4. Typical jointing and weathering developed on partially granitized sediments.

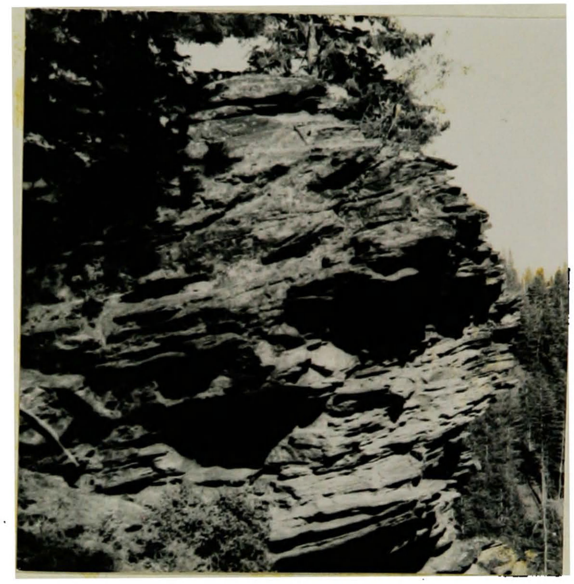

Figure 5. Typical slabby weathering surface of sillimanite gneiss.

Figure 6. Gneissic inclusion in faintly foliated partially granitized rock.

Figure 7. Early stage of inclusion development in partially granitized rock.

tably into the enclosing gneiss that no contacts can be found. The larger sheets grade laterally in at least one direction into the sillimanite gneiss. It is impossible to locate a definite contact in these gradational areas. There is no observable structural dislocation of the enclosing gneiss in the vicinity of the terminal portions of the sheets.

An interesting feature observed at location 8-6-4 and pictured in Figure 6 is the presence of a well foliated, rounded inclusion contained within the more massive partially granitized rock. The planar features in the inclusion are at a high angle to the foliation in the host. Figure 7 is included to show how such a relationship might arise in a rock which was undergoing deformation accompanyiny granitization. The block of gneissic material delimited by the dotted line shows a similar relationship captured at an earlier stage of formation. It is bounded on only two sides by more massive rock. It is easy to visualize the intermediate stages which would exist between the two photos. In fact, it does not seem unreasonable to visualize the entire mechanism operating in an essentially crystalline state where the host rock was capable of yielding plastically.

The partially granitized rocks at the western boundary of the larger andesinite body and also in the sandwich of varied rock types in the center of the smaller andesinite body are different in certain respects from the sheets which occur in the sillimanite gneiss. The contacts are commonly sharp, the texture is slightly more granitoid, and the composition is more sodic. Structural relationships suggest that these rocks were not injected but rather that they represent more extensively granitized and soda metasomatized equivalents of the slightly dissimilar rocks in the sillimanite gneiss. Figure 8 shows a sharp contact with the grey weathering gneiss. It is indeed difficult to conceive of a metamorphic or even a metasomatic contact being as sharp as the one pictured. However, certain characteristics of the contact are inharmonious with intrusive interpretation. For example, the gneissic rock shows no increase in deformed character as the contact is approached. In fact, the deformed banding of the gneiss continues into the more massive rock as a faint preferred orientation of biotite grains. Careful examinations of excellent exposures show that the faint foliation in the quartz monzonite possesses a deformed structure identical with that of the gneiss. Differences in mineral composition could also be traced laterally across the contact in some outcrops. In addition to these features, thin stringers of gneissic rock an inch or so in thickness extend for several feet into the quartz monzonitic rock. These stringers show no evidence of deformation which might be related to intrusion of the quartz monzonite.

On the other hand, there is evidence which suggests that at least some parts of the bodies were largely molten.

A thin section cut across a knife sharp contact between the quartz monzonitic rock and the andesinite (Figure 9) shows a striking disequilibrium relationship which can not be interpreted as either metamorphic or metasomatic. In the monzonitic portion the plagioclase is An_{18} and the mica is biotite and in the andesinite portion the plagioclase ranges from An_{33} to An_{54} and the mica is chlorite. One must conclude that the quartz monzonite was brought into contact with the andesinite and therefore is later than the andesinite. Emmons (1953) describes a somewhat similar condition where a gabbro was cut by a liquid granite apophysis. A labradorite crystal contained in the gabbro continued to grow into the granite as oligoclase. Such a condition could only develop between liquid and solid phases where the temperature of the liquid was too low to produce equilibrium.

Petrographic considerations will be confined to the studies of the more massive phases because increasing gneissic texture only represents a change to characteristics similar to the sillimanite gneiss which has been considered above. The mineralogy is probably dependant in part on the composition of the original sediment. Two samples have 73 and 58 percent quartz with minor orthoclase and two others have the approximate composition of a quartz monzonite. Three others are recognized as being more sodic. Modal analyses are given in Table 2. The texture is medium grained, allotriomorphic, inequigranular and invariably granoblastic (Figure 10). Large (ca. 5mm) grains of slightly perthitic orthoclase with

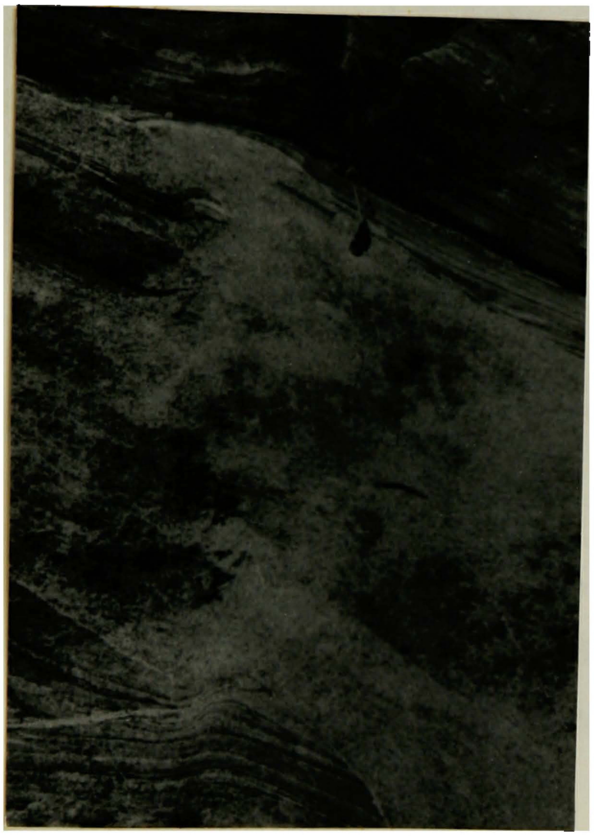

Figure 8. Contact between extensively granitized rock and migmatized grey weathering gneiss at Loc. 8-28-2.

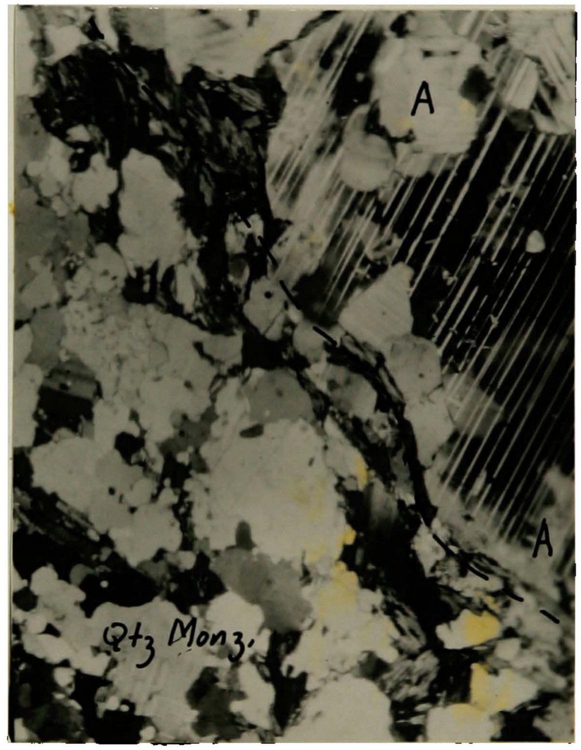

Figure 9. Contact between sodic quartz monzonite and andesinite (A). Loc. 9-2-8. x 17. Crossed nicols.

well developed poikilitic texture are common. Inclusions are biotite, quartz and plagioclase. The plagioclase is oligoclase and is commonly rimmed with albite which has grown during the exolution which formed the perthite. Some potash feldspar grains have a core of oligoclase. Plagioclase grains commonly exhibit faint progressive or regeneration zoning. (Figure 11). They often have a core of orientated, lath shaped inclusions of muscovite and biotite. The laths were probably incorporated within the growing plagioclase and oriented by forces related to the

	Sample	Modal Percentages				
		Qtz	Orth	Plag	Bio	Musc
Quartz Rich	8-12-7	73	3	9	7	8
	8-6-3B	58	2	15	9	16
Quartz Monzonitic	8-13-10	31	28	33	7	1
	9-2-9	27	31	33	8	
Sodic Quartz Monzonitic	8-28-1	24	16	41	19	
	8-28-1C	24	22	42	11	
	9-2-8	30	18	40	11	

Common accessories are apatite, zircon, chlorite, sphene sericite and garnet.
Note: Estimated mineral compositions were obtained on several other thin sections. Each of these could be classified as either quartz monzonitic or sodic quartz monzonitic thus substantiating the division made above.

Table 2. Modal analyses of partially granitized rocks.

crystallography. Where the rock is slightly foliated the folia are generally short, discontinuous, and are not accompanied by shearing effects. One thin section contain-

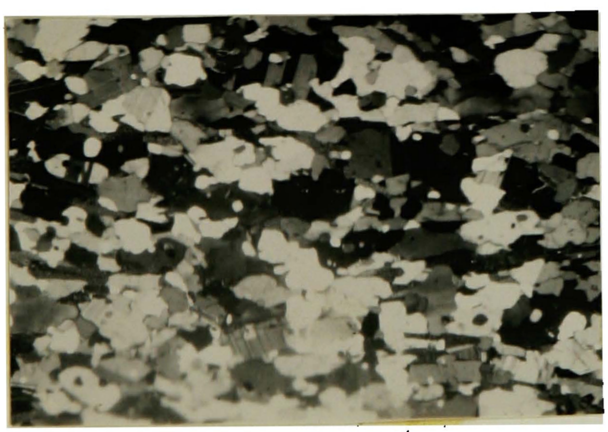

Figure 10. Typical texture of partially granitized metasediments. Crossed nicols, x 17.

Figure 11. Zoning of plagioclase in partially granitized metasediment. Crossed nicols, x 17.

ed two well formed, relict porphyroblasts of garnet. The majority of these petrographic observations point to a metamorphic origin. The presence of zoned structures in some plagioclase grains may be a result of a relatively rapid decline in temperature at the close of metamorphism. (See page 57.).

Grey Weathering Gneiss

The entire southwestern portion of the mapped area is composed of a quartz rich, fine- to medium-grained, grey, evenly banded, contorted and ptygmatically folded gneiss interlayered with more massive granitized or injected (?) rock. Outcrops of the grey gneiss extend from the bottom of Bass Creek Canyon to the ridge top, thus representing a minimum thickness of approximately 2500 feet. Groff (1954a) described a grey weathering gneiss in Kootenai Creek Canyon which contains less quartz than the sillimanite gneiss. The present study revealed a much greater abundance of quartz, in fact, twice as much, in the grey weathering gneiss as in the sillimanite gneiss. The modal analyses of three thin sections from widely spaced locations are given in Table 3. It is obvious from these data that the rock is a metasediment which is consistently high in SiO_2. Certainly this rock is not equivalent to Groff's grey weathering gneiss which he interpreted as a concordant igneous intrusion (Groff 1954a, p. 31). Because Groff did not map his grey weathering gneiss, the terminology is adapted here in spite of the dissimilarity of the two rock types.

In addition to color and quartz content the grey weathering gneiss differs from the sillimanite gneiss in several respects. It is a banded rather than a foliated rock. Structural deformation has not been accomplished by extensive shearing and crenulation but by tight isoclinal and ptygmatic folding. Under the microscope sillimanite is extremely rare and is not found in well formed bundles. Also, the iron-magnesium ratio of the biotite is lower in the grey rock. Sericitization is common as is the reaction biotite→chlorite, both indicating the presence of significant amounts of water. Potassium feldspar occurs as very irregular amoeba shaped blebs interstitial between quartz grains. Another major difference is the presence of continuous bands and lenses of a greenish colored rock composed of quartz 30%, orthoclase 11%, andesine feldspar 19%, calcic amphibole 26%, phlogopite 6%, sericite 7% and minor sphene, apatite and zircon. These bands apparently represent rather impure, calcic zones in the original quartz rich sediment. They range from a few inches to a few feet in thickness.

	Modal Percentages						
Sample	Qtz	Orth	Plag	Bio	Musc	Chl	Ser
8-19-7	65	11	2	5	10	5	
8-27-11	68	5	12	4	5	3	2
8-19-2c	68	5	8	2	5	4	6

Chl. - chlorite Ser. - sericite
Common accessories are sillimanite, apatite, rutile, sphene and opaque minerals.

Table 3. Modal analyses of grey weathering gneiss.

Langton (1935) described rocks of sedimentary character on the upper regions of the spur between Bass and Kootenai Creek Canyons. The aerial distribution does not agree with the present mapping nor does the structural interpretation offered by Langton. Langton interpreted the rocks as a large zenolith. Structural data obtained by the writer on the grey weathering gneiss agrees very well with the regional structural pattern, thus precluding any necessity for a complicated explanation of its occurence. It is simply a large tract of metamorphosed Belt sediments in normal structural position.

Andesinite

Several anorthosite bodies are exposed in Bass Creek Canyon. The outlines of the two larger bodies in east-west cross section are shown in Plate 1. The larger body has an outcrop area of approximately 1½ square miles. The aerial extent of the small bodies was not determined. The bodies are composed almost entirely of a non-resistant, very leucocratic, monomineralic rock to which the term andesinite is applied. Plagioclase feldspar, ranging in composition from An_{50} to An_{30}, and minor chlorite (penninite) are the principal, and in most cases the only mineral constituents. Plagioclase forms from 95 to 100 percent of the rock.

Bass Creek Canyon symmetrically dissects the larger body and offers an excellent area for three dimensional study by exposing the upper portion to a depth of 1800 feet. Groff (1954a) made a reconnaissance study of Bass Creek Canyon and briefly described the "Bass Canyon intrusive" (p. 34). He recognized an eastward gradation from a grey, medium grained, granitoid quartz monzonite through alaskite (my andesinite) to the sillimanite gneiss. In his words, "The gradual gradation outward through alaskite to the contorted gneiss may indicate the extensive assimilation of the gneiss or, more probably, its alteration and the driving out of femic materials by emanations from the intruding [quartz monzonite] magma." (p. 39-40). Groff's application of the term "alaskite" was indeed a misnomer for he recognized that the rock is composed almost entirely of plagioclase. His structural interpretation is also incorrect because the main mass is andesinite (his alaskite) bounded on the west by quartz monzonite rather than a quartz monzonite core surrounded by an aureole of andesinite.

As the upper and eastern borders of the larger body are approached a distinct foliation is recognized and it is always concordant with the enclosing gneiss. The thickness of the foliated zone ranges from approximately 50 to 200 feet. Chlorite, phlogopite, biotite and in some cases muscovite and sillimanite are contained in bands which alternate with bands composed wholly of plagioclase. The rock is termed a foliated andesinite. In general, foliation becomes more distinct and coarser in texture as the enclosing gneiss is approached. Near the contact the foliation in the two rocks is very similar. In the more faintly foliated inner zone, zoned and twinned plagioclase with a composition range from An_{47} to An_{29}

composes 85 to 98 percent of the rock.

A dominant feature of the foliated andesinite zone is the ubiquitous occurence of concordant lenses and pods of amphibolite which are commonly enveloped by masses of pegmatitic material. The amphibolite bodies range from one foot to thirty feet in thickness and may be several hundred feet long or, where they occur as pods, only a few feet long. In several outcrops where they appeared pod shaped in east-west cross section they were seen to be elongate in a north-south direction, thus resembling boudinage type structures with a north-south axis. Commonly, they are evenly foliated and exhibit east-west alignment of hornblende prisms. Where they are cut by pegmatite material the foliation continues unaffected up to the pegmatite contact. The amphibolite bodies may be found anywhere within the foliated zone but are most common near the contact with the enclosing gneiss. Some occur in the enclosing gneiss near the contact. An average of three thin sections gave the following estimated mineral composition: quartz 18%, plagioclase (An_{45}) 15%, hornblende 55%, biotite 7%, chlorite 2%, garnet 3%. Chlorite and garnet are not always present and minor accessories are rutile, ilmenite and pyrite. At one location garnet porphyroblasts increased in abundance toward the center of a pod where they composed about 20 percent of the rock. Under the microscope the texture is granoblastic where quartz and feldspar are abundant, but otherwise it is typically poikilitic with both hornblende and garnet riddled with inclusions. Garnet porphyroblasts are generally euhedral and hornblende is anhedral. The texture and structure of the amphibolite bodies strongly indicates a metamorphic origin and because they occur at random positions within the foliated border zone of the andesinite a metamorphic origin is also suggested for this zone.

X-ray diffraction studies of the basal reflections of biotite (Gower, 1957) separated from six samples collected at various distances from the massive andesinite revealed an increase in the iron-magnesium ratio as the enclosing gneiss is approached and entered. The increase is also clearly visable in the field by a color change from light colored phlogopite to red-brown biotite characteristic of the sillimanite gneiss. The thesis is that the biotite has lost much of its iron content which has been segregated into amphibolite bodies.

The central mass of the andesinite bodies is white plagioclase rock, with specks of green chlorite, and is generally devoid of texture, although a faint preferred orientation of the chlorite is discernable in some outcrops. The petrography of the andesinite and its foliated border facies is principally a consideration of the texture, composition, zoning and twinning of the plagioclase feldspars. The texture is allotriomorphic granular and typically granoblastic. It bears a striking resemblance to a clastic sedimentary rock. The grain size ranges from 0.2 mm to 3.5 mm. Grain size studies revealed a concentration around the larger and smaller figures with intermediate sizes being seriate but less abundant. Commonly, the smaller grains are concentrated in clots which serve to accentuate the size difference. This feature may indicate two generations of plagioclase. Figure 12 shows all of these features rather closely. The southwest corner of the main body is coarser grained than average with some plagioclase grains attaining lengths of 2 cm. Grain breakage is indicated by zone boundaries which do not follow the grain outlines. This, however, is not a distinctive feature in any of the sections studied. In general, protoclastic effects are absent.

Figure 12. Typical texture of the massive andesinite. Crossed nicols, x 17.

A wide compositional range between individual grains of plagioclase is characteristic of all thin sections studied. Compositions were determined by the Michael-Levy method, and for two samples the data were checked and verified by refractive index measurement made with the four axis universal stage. The compositional ranges for the plagioclase of seven thin sections, in terms of weight percent anorthite, are as follows:

Sample Number	Percent High	An Low	Range	Rock Type
8-8-6A	50	29	21	Foliated andesinite
8-19-2B	50	31	19	" "
9-2-3	53	37	16	" "
8-21-4A	50	31	19	" "
9-2-8	54	33	21	Massive andesinite
8-20-3	43	28	15	" "
8-21-1	50	30	20	" "

Four zoning relationships are characteristic of the plagioclase: (1) a calcic core bounded rather sharply against a more sodic rim (Figure 13), (2) a calcic core in sharp contact with a more sodic rim which becomes progressively calcic outward (Figure 14), (3) a progressive sodic increase toward the grain boundary through as many as four distinct zones (Figure 15), (4) a simple progressive zoning with no observable contacts. The first two are probably the most abundant. Two or three of these zone relationships may be present in a single thin section together with a large number of grains which show no zoning. The twinning is principally a well developed, fine, polysynthetic (albite) type. Some grains show cross twinning and offset, bent and irregular twin lamallae, all of which are features indicating that cataclasis has been operative to some extent.

The composition and zoning of the andesinite certainly indicates a rock which was arrested in a state of

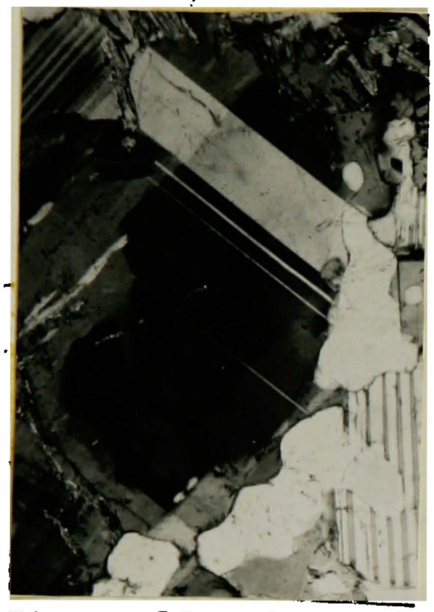

Figure 13. Crossed nicols, x 50.

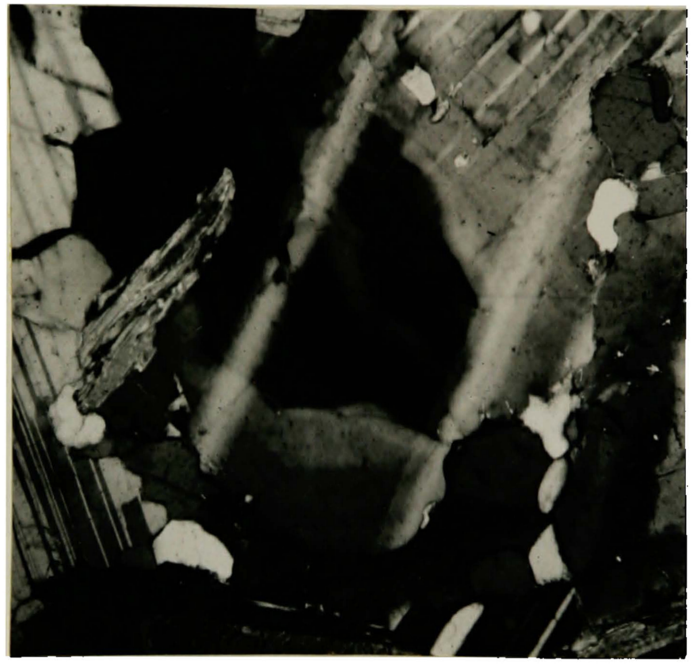

Figure 14. Crossed nicols, x 50.

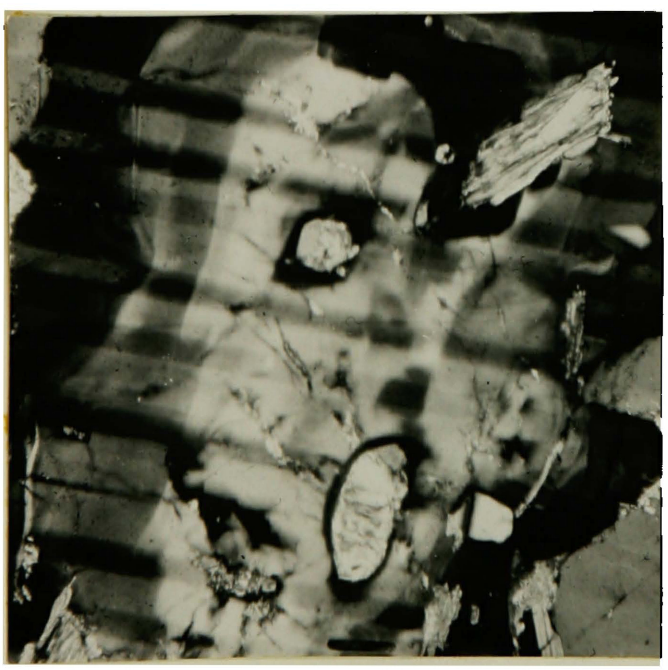

Figure 15. Crossed nicols, x 50.

Figures 13, 14, and 15. Selected zonal relationships of plagioclase from andesinite. See text for descriptions.

strong disequilibrium. Such disequilibrium relationships are indeed strange to metamorphic rocks. On the other hand, if one evaluates each factor at its face value, a complex crystallization history, in terms of igneous genesis, must be invoked for an adequate explanation of all the factors. In a later section on petrogenesis the writer will consider these factors in the light of metasomatic transfer as a possible operative mechanism.

Minor Rock Types

Thin sections of three diopside bearing rocks from the frontal zone were studied and found to be extremely variable in texture and mineralogy. Samples 8-6-1G and B-2 were collected from the vicinity of a west facing cliff about 1,000 feet west of the canyon mouth. The first is composed of about 45 percent large, poikioblastic prisms of diopside with about 25 percent of smaller, subrounded inclusions and clots of plagioclase (An_{25}). A light green amphibole (about 20 percent) occurs as fracture fillings and is seen to replace the pyroxene. Minor minerals are quartz and sphene. Sample B-2 is similar in texture but quite different in mineralogy. Its approximate composition is diopside 30%, plagioclase (An_{60}) 25%, quartz 40%, and sphene 5%. The third sample 9-9-3, was collected from the low foothills east of the main range front. It differs from the above in both texture and mineralogy. The texture is subequigranular with subrounded, anhedral grains of diopside and plagioclase set in a matrix of elongated quartz grains (Figure 16). The approximate mineral composition is diopside 45%, plagioclase (An_{57}) 4%, quartz 55%, and sphene 1%. Sample 8-6-1F is a related rock which was collected from the same outcrop as 8-6-1G. It does not contain pyroxene but is characterized by bands rich in green amphibole alternating with biotite rich bands. The modal composition is quartz 41%, plagioclase (An_{36}) 30, biotite 13%, amphibole 12%, and minor sphene, zircon and opaques.

A second group of minor rocks is associated with the contact zones of the andesinite bodies. In the vicinity of the upper, eastern boundary of the larger andesinite body, the rocks bear evidence of severe deformation. Small folds and fracture zones are common. Under the microscope the deformation is evidenced by severe grain breakage and development of coarse crystalline sillimanite (Figure 17) or by extensive foliation plane shearing with the development of large bundles of sillimanite (Figure 18). Chlorite is commonly a major constituent of these rocks. Another rock type observed at the contact of the smaller andesinite body is a highly sericitized and chloritized gneiss which has the following modal composition: quartz 41%, plagioclase (variable composition) 7%, sericite 21%, muscovite 8%, chlorite 14%, sillimanite 4%, with minor rutile, ilmenite, pyrite and sphene (Figure 19). The significant feature of all of these rocks is the high percentages of hydrated minerals. It is possible that water has played a significant role in reactions which occured at or near the contact of the ande-

Figure 16. Pyroxene-quartz-plagioclase rock with well formed sphene prisms. Loc. 9-9-3. Plane light x 58.

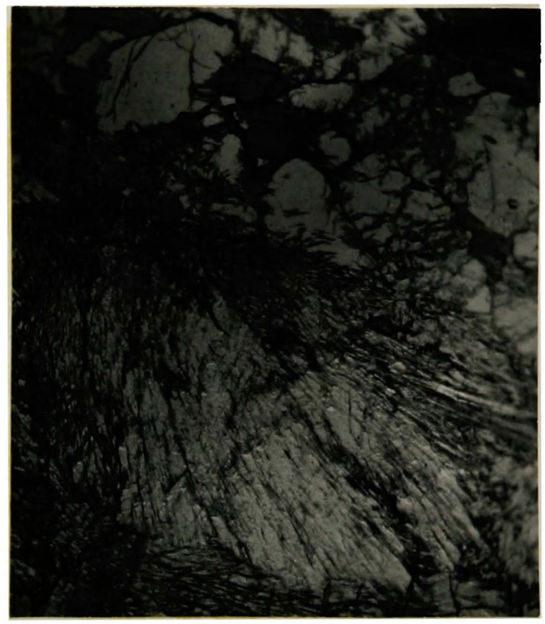

Figure 17. Coarsely crystalline sillimanite and broken plagioclase. Loc. 8-8-6c. Plane light x 17.

Figure 18. Sillimanite-chlorite gneiss. Loc. 8-8-6d. Crossed nicols x 12.

Figure 19. Chloritized and sericitized gneiss. Loc. 9-2-1. Plane light x 17.

sinite.

STRUCTURE

Previous structural interpretations relating to the Bitterroot Range are concerned mainly with the genesis of the long, straight, eastern front (Lindgren, 1904, Langton, 1935, and Ross, 1950). Existing structural data on the main portion of the range are very scarce. The present thesis marks the first attempt to construct a geologic structure section which is not hypothetical or idealized to a great extent. Also, it marks the first* detailed mapping which is accompanied by extensive structural data (Plate 1). It must be emphasized however, that the Bass Creek Canyon area is probably very atypical when compared with the rest of the Bitterroot Range. In addition, it must be realized that the mapped area is an extremely small segment of a very extensive structural province. With these limitations, it is apparent that generalized conclusions regarding regional structural relationships can not be made with certainty.

Three periods of deformation are recognized. Whether they represent distinct episodes in a related sequence of

* Leischner (1959) has mapped in detail about 20 square miles of the border zone of the Idaho batholith in the vicinity of Lolo Hot Springs to the northwest of Bass Creek Canyon. However, his area is structurally very different from the Bass Creek Canyon area and as such, cannot be brought to bear on the present study. A study of the intermediate area would be most interesting and would probably be of great aid in solving existing structural problems.

deformation or unrelated periods of deformation is not known. The first period is of unknown intensity and age. Evidence for this deformation is represented by east-west trending small folds and contortions. For example, on the north wall of Kootenai Creek Canyon about two miles west of the canyon mouth, excellent examples of mullion structure are present. The individual rods are an inch to a few feet in diameter and attain lengths of several yards. They bear east-west and plunge to the east at about 20°, parallel to the foliation in the area. The rods were formed by the shearing off of the limbs of very small east-west folds during a later period of east-west shearing. The anticlinal and synclinal portions sheared past one another producing a series of rod-like structures which, when viewed from the south, resemble a stack of various sized pipes resting on the canyon wall. Small east-west folds were recognized in several other exposures in Kootenai and Bass Creek Canyons. The apparently anomalous foliation attitudes recorded directly west and southwest of the larger andesinite body may be a result of this earlier period of deformation. That is, the north dipping attitudes in the south west corner of section 35 and the steeper southwest dips recorded from the west central part of the same section may represent two limbs of an earlier east-west syncline. East-west folding has been noted in the drainage of Carlton Creek north of the mapped area (Wehrenberg, J. P., 1959, personal communication).

The second and major period of structural deformation is responsible for the gross structural and physiographic pattern of the Bitterroot Range and for the texture and fabric of most of the rocks which make up the range. There is little doubt that this period of deformation is related to the emplacement of the Idaho batholith. Ross (1936 and 1950) was the first to state clearly the genetic association, although his interpretation regarding the boundary of the batholith is questionable. Referring to the east dipping frontal zone gneiss, as adapted here, Ross states (1950, p. 170), "The explanation of the attitude of the gneiss slab that most simply and satisfactorily accounts for the known facts is that this slab is part of the rocks that were invaded by the Idaho batholith and were both metamorphosed and domed during the intrusion (Figure 56)." The structure of the northern part of the range is far more complex than this statement indicates, and if the thesis of a metasedimentary origin for the sillimanite gneiss is accepted, Ross's interpretation is inadequate. A simple domal relationship is precluded by the west dipping foliation attitudes recorded by Langton (1935, Plate 1), Ross (1950, Plate 4), and in the present paper (Plate 1). Langton (1935, p. 39) pictured the northern part of the range as a broad, north-south, asymmetrical, anticlinal-like-structure, the western limb of which is the steeper and dips from 50° to 80° to the west, into, and maybe under, the batholith. The writer is in essential agreement with this interpretation although it is doubtful that the relationships are as simple as Langton postulated. The Idaho batholith is thus viewed as a discordant body which has invaded, deformed, and granitized large volumes of pre Cambrian sediment. In addition to structural data obtained within the mapped area, the writer has based these conclusions on a reconnaissance traverse made to the crest of the range along Bass Creek Canyon. The observations made on that traverse are in agreement with observations by Anderson (1930) on the border zone of the batholith in the vicinity if Orofino, Idaho. As noted by Anderson, it is difficult to recognize the boundary of the batholith where it is in contact with Belt sediments "because the intense metamorphism induced by the batholith has changed the sediments to gneisses, mainly as the result of injection of igneous material along bedding or schistosity planes. The gneissic shell is many miles broad in places and it is exceedingly difficult to know whether the rock should be classed as sheared batholith or as gneissic sediments as in many places there is apparently a gradational transition from one to the other.".

Within the mapped area the second period of deformation has produced three distinct types of structures. (1) Extensive normal shearing in an east-west direction along the foliation planes of the red weathering sillimanite gneiss and the frontal zone gneiss. In the latter, extensive cataclasis and crenulation accompanied the shearing. Shearing culminated in extensive contortion, crenulation and crinkling of much of the red weathering gneiss. (2) An equivalent deformation of the grey weathering gneiss accomplish-by tight isoclinal and ptigmatic folding and contortion without significant shearing. (3) Broad, north-south folding superimposed upon and probably concomitant with the first two. This folding, and its relation to the andesinite bodies, is discussed on page 50.

A rather naive attempt to determine the direction of movement along foliation planes was made by feeling the slickensided surfaces of unoriented samples broken from outcrops and then reorienting the samples. In all cases the movement was found to be normal. Also, drag folded pegmatite stringers in the frontal zone indicated a normal movement.

Evidence of the third and last period of deformation is indicated by northeast trending, high angle, normal, faulting along the range front and by a related (?) set of joints within the range proper. The faults have physiographic expression in the form of shallow trenches which separate low foothills from the main range. Trenches have been noted near the mouths of Big Creek (Lindgren, 1904, p. 49), Sweathouse Creek (Groff, 1954a, p. 40), and the writer has observed trench-like features near the mouths of Sweeney, One horse, and Carlton Creeks. Also, Pardee (1950) has noted the presence of notches in the spurs south of Roaring Lion and Blodgett Creeks and Groff (1954a) has described high angle faults which cut the spur between

Kootenai and Bass Creeks.* The writer had the opportunity to study aerial photograph mosaics of the Bitterroot Valley at the U.S. Geological Survey office, Montana State University. The photographs indicate that each of the trenches represents the trace of a separate fault. Taken together, they form a north-south band of en enchelon faults which strike N 17° to 30° E and are thought to dip steeply to the southeast (?). Two factors indicate a steep dip and the second of these suggests that the dip is to the southeast. (1) The aerial photograph mosaics show that the traces of the faults are only slightly affected by changes in topography. (2) An analysis of 75 joint attitudes recorded in the present study revealed a dominant set which strikes N 27° E and dips 81° SE. Two small faults in the west central part of the area have similar attitudes. A second and minor joint set strikes N 57° W and dips 78° NE. The similarity in strike between the dominant joint set and the en enchelon faults suggests that they might be related and have similar dips.

Two of the faults are present in the northeast corner of the mapped area and their approximate locations, as indicated on the aerial photographs, are shown on Plate 1. The northern ends are represented by shallow trenches

* The faults mapped by Groff were not seen by the writer on a reconnaissance traverse along the north wall of Kootenai Creek Canyon. Groff used west facing scarps as the chief evidence for recent high angle faulting. Such scarps are common along the walls of most of the canyons of the Bitterroot Mountains and do not indicate faulting.

but the southern ends are buried beneath the valley alluvium. The fault on the west marks the eastern limit of the main range between Bass and Carlton Creeks. Its trace north of the mapped area intersects the canyons of both Sweeney and Onehorse Creeks very near their mouths. The fault to the east is well exposed in an outcrop along Sweeney Creek Road about one half mile east of the mouth of Sweeney Creek Canyon. At this location, a zone, at least 200 feet wide, of crushing and brecciation of a light green quartzitic rock is present. A thin section of the brecciated rock from this exposure showed an early fracturing which has been thoroughly healed by quartz and feldspar. The present brecciated condition of the rock must be a later feature, and thus, at least two periods of movement are recognized. Other faults in the en enchelon pattern parallel the range, or intersect it at very low angles, in the vicinity of the mouths of Kootenai, Big and Sweathouse Creek Canyons. A zone of severe crushing and brecciation was observed at the mouth of Kootenai Creek Canyon. Study of a thin section from this location indicated two stages of movement as mentioned above. Konizeski, R. L. (personal communication) noted a crush zone more than 100 feet wide at Sweathouse Creek, and Lindgren (1904) reported a fracture zone one-half mile in width at the Curlew Mine south of Big Creek.

From the data presented, it is apparent that the band of known en enchelon faults extends for at least twenty five miles along the northern part of the range. It is reasonable to assume that the faults are related to a tear zone along the range front. The mountain block has moved northward relative to the valley block. Figure 19 is an east-west cross section through the faulted area (See Plate 1 for location.). It shows the structural interpretation which best explains the westerly dips recorded in the western portion of section 28. This interpretation suggests dip-slip movement of rather large magnitude along faults which border rotated fault blocks. The wide fracture zones characteristic of the faults may also indicate a rather large movement. An evaluation of the role which this faulting has played in the building of the Bitterroot Range, or a consideration of its relation to regional tectonics, is beyond the scope of the present work.

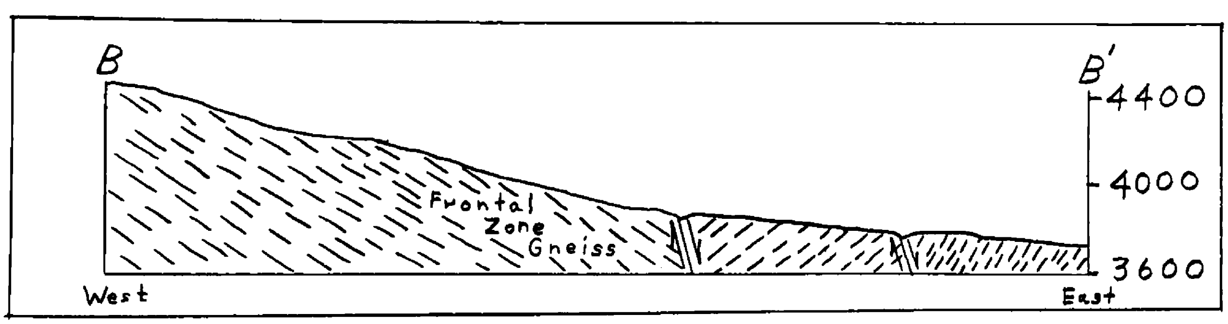

Figure 20. East-west cross section along line B-B' showing interpretation of faulting along range front (See Plate 1). Scale and datum plane same as Plate 1.

Structures Related to the Andesinite Bodies

In the bottom of Bass Creek Canyon, approximately one mile west of its mouth, foliation attitudes change from the 25° to 30° eastward dips characteristic of the range front to westerly dips of approximately the same magnitude.

From this point westward, along the bottom of the canyon, the dips steepen to the west until they are practically vertical at the eastern border of the larger andesinite body. The same increase in magnitude of westerly dip is observed as a westward traverse is made between the two andesinite bodies. In addition to being bounded on the east by vertical dipping gneisses, the smaller andesinite body has a septum of vertical dipping gneissic rocks in its center. The rocks of the higher regions of the canyon walls are not involved in the folding, although a gentle, north-south, synclinal-like, structure occurs high on the canyon wall near the eastern boundary of the larger andesinite body. Such a striking difference in structure within a relatively limited vertical distance suggests that extensive shearing has played a significant role in allowing the deeper rocks to be folded while the more shallow rocks remained virtually undeformed.

The structure at the western boundary of the larger andesinite body is complicated by minor faulting, possible earlier east-west folding, and the presence of bodies of rather massive quartz monzonitic rock. The faults are well exposed on the canyon wall as shear zones about one foot in width. They cut the quartz-monzonite and thus are younger than this rock. Possible evidence of rather recent movement along the faults is suggested by the apparent left-lateral offset of the canyon bottom in the vicinity of the faults. The interpretation of the faulting shown on Plate 1 is highly speculative because extremely steep terrain made much of the area inaccessible and large talus slopes conceal a large part of the lower canyon.

PETROGENESIS

Genesis of Metasediments

The sedimentary characteristics of the frontal zone gneiss and the grey weathering gneiss compel one to conclude that they are metamorphosed pre-Cambrian sediments, probably of the lower Belt Series. Both the Ravalli group and the Pritchard formation of the lower Belt are composed mainly of argillites and argillitic quartzites (Ross, 1950, p. 144). The writer is not aware of the existence of any quantitative petrologic data on these sediments, and because of this, it is not known how important the role of metasomatism has been in producing existing mineralogical assemblages. The percentage of Na feldspar, both in the gneisses and in the pegmatitic material, represents a greater amount of Na than one would expect in a normal sediment. Because of this, it is believed that at least this one element was metasomatically introduced in significant quantities.

Mineralogically the sillimanite gneiss has the approximate composition of a quartz monzonite. It thus might be argued that the gneiss is the metamorphosed border facies of the quartz monzonitic rocks of the Idaho batholith. However, the high percentage values for biotite and sillimanite in the gneiss show that it is probably higher in Fe and Al than the rocks of the batholith. Also, some samples collected from within the sillimanite zone are too rich in quartz to have had an igneous origin. In addition, gradations are recognized from both the frontal zone gneiss and the grey weathering gneiss to the sillimanite bearing rocks. In the former case the gradation appears to represent an increase in metamorphic intensity. The upper stability limit of muscovite is exceeded very near the P-T conditions favoring the stability of sillimanite. In the latter case the gradation is a northward transition from the grey weathering gneiss to the sillimanite gneiss along the strike of the foliation in the vicinity of the southern boundary of the smaller andesinite body. The grey gneiss is higher in quartz content by a factor of two. Such a gradation is best explained by a facies change in the original sediment. Another possible explanation of the lateral change is that selective metasomatism has given rise to a compositional difference. An indirect piece of evidence which can be invoked in support of a metasedimentary genesis is the similarity between the metamorphics of the northern Bitterroot Range and high grade metamorphosed Belt sediments reported from the contact zone of the Idaho batholith in Idaho (Anderson 1930 & 1940, Hietenen 1956, Leonard 1957).

The presence of pyroxene bearing rocks in the frontal zone gneiss, and hornblende bearing bands in the grey weathering gneiss point to the control which the original sediment composition has exercised in shaping the existing mineralogy. In both cases the rocks represent stratigraphic beds or zones with a significantly high lime content. According to Harker (1939), one would not expect pyroxene to be a stable phase in the higher grades of regional metamorphism of calcareous, argillaceous sediments. A green hornblende and plagioclase are the stable minerals. This assemblage is characteristic of the grey weathering gneiss. Harker points out that pyroxene will only develop where the lime content of the initial sediment is high enough to reduce shearing stresses by its yielding nature. Under these conditions a limestone or dolomite, "with a sufficient admixture of siliceous and aluminous material ... will suffer complete de-carbonation, and we will find accordingly silicate rocks, not merely as nodules or inconstant bands in a silicate bearing limestone but occuring in mass." (Harker p. 260). He states further that if there is an excess of silica, diopside-quartz rocks, with or without other significant minerals, will develop. The diopside bearing rocks of the frontal zone probably formed in this manner. Indeed, a thin section of sample 9-9-3 has the identical mineralogy and texture of a diopside-quartz rock pictured by Harker (Figure 128) (Compare Figure 16, this paper).

Depth

There are several lines of evidence which suggest that the metamorphism was accomplished at a relatively shallow depth. The first is the paucity of almandite garnet in the sillimanite bearing gneiss. Iron garnet is a stable mineral phase in coarse grained sillimanite gneisses from so many metamorphic terrains that its virtual absence must bear some significance. It is not likely that the stability range of garnet was reduced by the extensive shearing which is characteristic of the Bass Creek Canyon gneisses because garnet porphyroblasts often show rotation during growth and are thus used as an indicator of foliation plane shearing (Knopf and Ingerson, 1938). Also, it can not be said that a deficiency of Fe in the sillimanite gneiss prevented the formation of iron garnet. The average modal percentage of biotite for these rocks is 19 percent which is rather high and certainly represents a sufficient source of iron. Garnet is a dense mineral and its stability range is enhanced by high confining pressures. (See Yoder, "The Crust of the Earth," 1955, p. 520). It has not been shown that garnet is unstable at low pressure but perhaps such conditions are not favorable for its development. At any rate, deep burial favors its formation and where it is not present, under otherwise favorable conditions, it seems valid to assume a comparatively shallow burial.

A second line of evidence is the cataclastic effects which have resulted from extensive shearing of the frontal zone gneiss. Handin (1958) reports that microscopic examinations of sandstones which had been subjected to triaxial compression tests revealed that, as pore pressure increased, grain breakage became less important until there was none, and deformation was due entirely to intergranular movements. In addition, Yoder (1955) states that at the outset of the first metamorphic reaction of a regionally metamorphosed terrain pore pressure is approximately equal to rock pressure. These two considerations suggest that cataclasis of large volumes of rock would not be expected under the normal depths of burial, for it would be impeded by high pore pressure. If the frontal zone gneiss was never deeply buried then it follows that the underlying sillimanite gneiss was buried at no greater depth than the normal layered sequence indicates.

A major discrepancy arises if it is concluded that metamorphism was relatively shallow. In many exposures the grey weathering gneiss has a typical migmatoid structure accompanied by ptigmatic folding. Such relationships are generally interpreted as characteristic of deep-seated intrusive contacts (See Turner, 1948, p. 5). An interesting problem thus arises, because within a rather restricted area we have the co-existence of deformation by cataclasis and deformation by "plastic flow" at no great difference in depth. Perhaps the explanation is simple. If the frontal zone gneiss was shallow enough so that its pore fluids were free to be lost to the surface and the pore fluids of the grey weathering gneiss were restrained from free migration, the resulting structures may reflect the control of such a difference. The presence of significant quantities of chlorite and sericite in the grey gneiss and their absence in the frontal zone gneiss supports this hypothesis. Also, the total decarbonation of lime rich zones mentioned above suggests that volatile or fluid phases of the frontal zone rocks were free to migrate.

<u>Temperature</u>

A rough estimate of the temperature of metamorphism can be made on the basis of (1) the muscovite-sillimanite isograd and (2) x-ray analyses of the lattice spacings of plagioclase. According to Yoder and Eugster (1955) the upper stability limit of muscovite at the sillimanite - muscovite isograd is approximately 650° C. Fyfe, W. S., et al. (1958, p. 164) point out that this value of 650° C. can not be applied directly to the amphibolite facies because the data were obtained on a silica deficient system. These writers suggest that the temperature is significantly lower in a natural system. X-ray powder diffraction studies of the difference in two theta values between the ($1\bar{3}1$) and (131) lattice planes of plagioclase of known composition can be used as a measurement of the degree of inversion from high temperature to low temperature forms (Smith and Yoder, 1956). Such measurements were made on five samples, the compositions of which were determined by imersion oil measurements. A sample from the sillimanite gneiss and one from the frontal zone gneiss showed a significantly lower state of inversion than samples of similar composition from normal granites analysed by Smith and Yoder. Also, three samples of plagioclase from the andesinite showed a state of inversion similar to volcanic rocks analysed by Smith and Yoder. These data indicate that the rocks cooled rather rapidly. It is also apparent that during the late stages of the metamorphic history temperatures must have reached values high enough to produce high temperature forms of plagioclase. There is little data available on the inversion temperature of plagioclase but Tuttle and Bowen (1958, p. 15) speculatively conclude that for albite the temperature is approximately 720° C. The sluggishness of the inversion renders laboratory measurement extremely difficult. Because of this, it is likely that the inversion temperature is significantly lower in a geological environment. From what has been said, it is apparent that one can not pinpoint a temperature or even give a definite temperature range based on experimental data. However, it is likely that the maximum temperature obtained during the metamorphism was close to a value of 600° C. To depress this value furthur requires a greater departure from data obtained experimentally. This value is in essential agreement with normal temperature values which are assumed for the amphibolite facies. Ramberg (1952), for example, suggests a temperature range of 400° to 600° C. for the amphibolite facies (Figure 73, p. 137).

Kinetics

In a previous section (p. 26) the writer has presented evidence which supports the view that the slightly foliated and massive sheets of quartz monzonitic rock which occur in the sillimanite gneiss are partially granitized metasediments. An hypothesis will now be advanced for their genesis. It is believed that the level of regional metamorphism attained in the area was such that appropriate increases in energy caused by shearing were sufficient to allow the development of high energy volatile or fluid phases which migrated and acted as granitizing agents. The interaction between these materials and solid phases, under the P-T conditions which prevailed, partially or totally obliterated the coarse grained gneissic texture but never allowed the development of a granitoid or igneous texture. Bennington, K. O. (personal communication) has shown that large amounts of energy in the form of heat are liberated in the shearing of solid material. If sufficient quantities of this heat are absorbed by the crystalline phases present, bond energies may be surpassed and certain elemental components may be free to migrate. At the site of deposition of the materials thus released, the reactions are exothermic and significantly large quantities of heat are released. Bennington concludes that if shearing is of large enough magnitude the energy transfer will be quantitatively sufficient to granitize large volumes of rock at the site of deposition. The modal analyses obtained in the present study show that the partially granitized areas have been enriched in K with respect to the enclosing sillimanite gneiss. The enrichment may be a result of an energy gradient transfer as outlined by Bennington.

Genesis of Andesinite

In this section the salient features associated with the two larger andesinite bodies will be enumerated under two subheadings, (1) those which can be interpreted as evidence in support or origin in situ, (2) those which can be interpreted as evidence of intrusive emplacement. In some cases a dual interpretation is recognized.

Features indicative of origin in situ:

1) Thermal metamorphism at the contact is absent. This factor precludes the high temperatures which are required to have a melt of andesinitic composition.

2) Protoclastic textures are absent. This factor precludes the possibility of intrusion as a crystal mush. It msut be noted, however, that recrystallization subsequent to intrusion could eliminate such structure.

3) There is no structural dislocation of the overlying gneiss. This factor, together with the sense of folding at the eastern margin of both bodies, indicates strongly that the bodies could not have been forcefully emplaced from below.

4) Where foliated border facies are present they are concordant with the enclosing gneiss and in some places they contain concordant remnants of the enclosing gneiss. The foliation displayed within these borders can not be interpreted as a primary flow structure. It is often similar in structure to the characteristic foliation of the enclos-

ing gneiss. Several stages of eliminated foliation can be recognized.

5) The foliated borders contain large masses of amphibolite which are best interpreted as segregated mafic material originally contained in the replaced country rock. X-ray analysis of a suite of biotite grains showed an increase in iron content across the border zone of the andesinite into the enclosing gneiss. This factor serves to strengthen the argument that iron, and probably magnesium,* was lost from the enclosing gneiss. The amphibolite masses are a likely site of deposition.

6) The texture of the andesinite and also the enclosed amphibolite, is a typical metamorphic, granoblastic texture. There are no textural indications of crystallization from a melt. It is possible, however, that the metamorphic texture developed as a recrystallization product subsequent to intrusion.

Features indicative of intrusive emplacement:

1) Zoned plagioclase grains are characteristic of all the andesinite which was studied. It is significant that features (2) and (6) above require recrystallization subsequent to intrusion in order to be reconciled with an intrusive genesis. Such recrystallization would erase any zonal structures developed during primary crystallization and thus weakens this argument.

* The total percentage of biotite decreases in the same direction, thus indicating that Fe and Mg were both lost, and that Fe started its migration earlier.

2) One might conclude that the amphibolite masses represent a metamorphosed sheet of mafic, dioritic material which was intruded around the andesinite bodies. However, the composition of the amphibolite (See p. 34) does not suggest a rock which would fit into the normal differentiation sequence which might give rise to andesinite (the andesinite is a leucocratic diorite and the amphibolite a melanocratic quartz diorite). In fact, the absence of rocks which are normally recognized as consanguineous components of the differentiation sequence associated with anorthosite bodies can be invoked as indirect evidence against intrusive emplacement.

3) The fact that structural deformation in the form of folding is restricted to the areas surrounding the andesinite bodies suggests a genetic relationship. It is possible that the bodies were intruded laterally, either from the north or from the west. As such, they may have folded the gneissic rock down and to the east causing the present structure. However, it is difficult to visualize this as being the easiest release of intrusive forces. Deformation of the overlying rocks seems more reasonable.

4) It is very unlikely that metasomatic replacement of a metasediment containing zircon and shpene could give rise to a rock in which these minerals are virtually absent. Zircon is a common mineral constituent of the sillimanite gneiss and shpene is common in the lime-silicate rocks. This is certainly the most convincing argument

favoring intrusive emplacement of the andesinite.

The majority of the enumerated features suggests origin in situ. The conclusion is that the bodies were formed by the metasomatic transfer of material upward along the steeply dipping foliation planes and shear surfaces. Na, Ca, and Al are the elements which must have been introduced and K and Si are the elements which must have been removed in greatest quantity. However, if the original rock was rich in Ca, a lime rich sediment for example, this element might have been in sufficient supply prior to metasomatism. Assuming this, the process may appropriately be defined as feldspathization, a process which is generally believed to be quite common in the shallower regions of the earth's crust, although it is generally not visualized as going to completion. Indeed, it is difficult to concieve of an adequate supply of Al to completely feldspathize large volumes of rock. Alumina metasomatism has been reported from some localities (See for example, Haung, 1957.). A theoretical consideration of the problem is beyond the scope of this study.

In general, metamorphic rocks show a distinct lack of zoned structures in crystals of isomorphic series. This feature is generally taken to indicate the sluggishness with which the maximum T & P conditions of metamorphism in a particular area are relieved. The low state of inversion of plagioclase mentioned above is certainly unusual in this respect because it suggests a rather rapid decline of temperature. This condition is adequate to arrest any zonal structures which existed during the latest stages of metasomatism. The question arises, what conditions were responsible for producing the zoning? A hypothetical solution, predicated on the premise of a shallow depth of metamorphism, is advanced. Perhaps either liquid or vapor pressures could be built up at the site of metasomatism. If these pressures were relieved at certain stages of growth by the tapping off of more mobile fluid or gaseous phases to the surface, or to the enclosing rocks, the changes in the physico-chemical character of the residual material should be reflected in the plagioclase as zoning. Emmons, R. C., et al. (1953, p. 41) show that in plagioclase feldspars zoning may develop as a result of reactions between an essentially crystalline mass and the interstitial liquid material very late in the crystallization history and relatively rapidly.

CONCLUSIONS

Pre Cambrian Belt sediments are the only rocks which have been reported as being in contact with and involved in the emplacement of the Idaho Batholith. There is no doubt that the frontal zone gneiss and the grey weathering gneiss exposed in Bass Creek Canyon are the metamorphosed equivalents of this stratigraphic unit. Both rock types possess distinctive sedimentary features and are separated from the main mass of the batholith by a coarse grained sillimanite bearing gneiss which is also believed to be a metasedimentary rock, probably the regionally metamorphosed and soda metasomatized equivalent of the lowermost Belt Series. The existing mineral assemblages of the major metamorphic units are at least partially controlled by the original sediment compositions. Metamorphism is believed to have been accomplished at a relatively shallow depth under conditions of a comparatively steep temperature gradient. Theoretical considerations suggest that the maximum temperature of metamorphism was at least 600° c. Extensive shearing may have been a factor in elevating temperatures sufficiently to partially granitize large portions of the sillimanite gneiss. The sillimanite gneiss is rather unusual when compared with the normal sillimanite bearing rocks of high grade amphibolite facies; it shows a striking paucity of almandite garnet and amphibolite material.

Each of the major metamorphic units possesses a distinctly different type of deformed structure. The structures and their rock associations are (1) shearing, cataclasis and some recrystallization — frontal zone gneiss, (2) shearing, recrystallization and contortion — sillimanite gneiss, (3) plastic type deformation shown by tight isoclinal and ptigmatic folding — grey weathering gneiss.

Several concordant andesinite bodies are exposed within the mapped area. They vary greatly in size and appear to be either bun-shaped or lensoid in cross sectional outline. Certain features of these bodies indicate that they are igneous intrusions but the majority of the features indicate that they were formed in situ by the metasomatic feldspathization of existing rocks. The latter is the mode of origin favored by the writer.

The main stage or stages of metamorphism and deformation are doubtlessly associated with the emplacement of the Idaho batholith. This batholith is thought to have a discordant relationship with the enclosing gneiss in the northern Bitterroot Range. The attitudes of the gneissic foliation within the mapped area indicate a broad, north-south, anticlinal-like structure with the more gentle eastern limb forming the eastern boundary of the range. The steeper western limb is seen only in the deeper rocks of the gneissic sequence. The shallower rocks do not appear to be involved in the anticlinal folding except on the eastern limb.

At least three periods of deformation are recognized. In sequence they are (1) folding which produced east-west trending folds, (2) folding which produced north-south trending folds accompanied by extensive shearing, cataclasis, contortion, crenulation and recrystallization, and concomittant with the major stage or stages of metamorphism, (3) high angle, en enchelon faulting along the range front.

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