AN ABSTRACT OF THE THESIS OF

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The Duzel Rock area includes four square miles in the eastern Paleozoic subprovince of the Klamath Mountains southeast of Fort Jones, California. The structurally complex terraine is composed of Silurian graywacke, post-Silurian phyllite, limestone and basalt associations, and minor andesitic intrusive rocks.

Three groups of sedimentary and extrusive rocks have been recognized; these are the Duzel Rock Group, the Spring Branch Group, and the White's Gulch Limestone. Each of these groups is composed of at least three members. Two limestone members of the Duzel Rock Group may be correlative with the Ordovician Facey Rock Limestone to the south. The White's Gulch Limestone contains a sequence of over 1,000 feet of micrites, oomicrites, oosparites, bedded chert, and green celadonitic sand. This unit is interpreted as a deposit formed in a periodically restricted inter-arc basin, and is tentatively assigned a Lower Paleozoic age. The graywacke,

which is a portion of the Moffett Creek Formation, is interpreted as a proximal fan turbidite deposit.

Carbonate-cemented breccias mantle limestone outcrops on the east and west flanks of Duzel Rock. These breccias are composed almost entirely of limestone clasts and are believed to be cemented by supersaturated groundwater.

All of the members of the Duzel Rock Group are separated from each other by thrust faults, and all three limestone groups have been thrust over the Moffett Creek Formation. A large thrust fault of a magnitude comparable to the pre-Cretaceous regional Mallethead Thrust, separates the Moffett Creek Formation from the Duzel Phyllite. There are numerous high-angle faults in the area post-dating the thrusts.

Geology of the Duzel Rock Area
Yreka Quadrangle, California

by

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GEOLOGY OF THE DUZEL ROCK AREA, YREKA QUADRANGLE, CALIFORNIA

INTRODUCTION

Location and Accessibility

The Duzel Rock area is located in the Klamath Mountains of northern California. The map area is situated in the southwest corner of the Yreka Quadrangle in Siskiyou County (see Figures 1 and 2). The area is a segment of a ridge that begins approximately three miles to the north-northeast and continues about four miles to the south where it merges with another ridge that is oriented east-west. These ridges are part of a larger range that trends generally north-south and divides the drainage of the Scott River and Shasta River valleys. Duzel Rock and the rest of its ridge are bordered on the north and east by Moffett Creek and on the west by Duzel Creek, a tributary of Moffett Creek. This ridge is incised by several gulches; those in and around the map area include White's Gulch, Telephone Gulch, Farley Gulch, and Spring Branch, the latter a tributary of Moffett Creek which flows year-round. Duzel Rock is the highest point on the ridge at 6,039 feet above sea level, and after Antelope Mountain to the northeast, it is the second highest point in the Yreka Quadrangle.

The area of study, which encompasses four square miles,

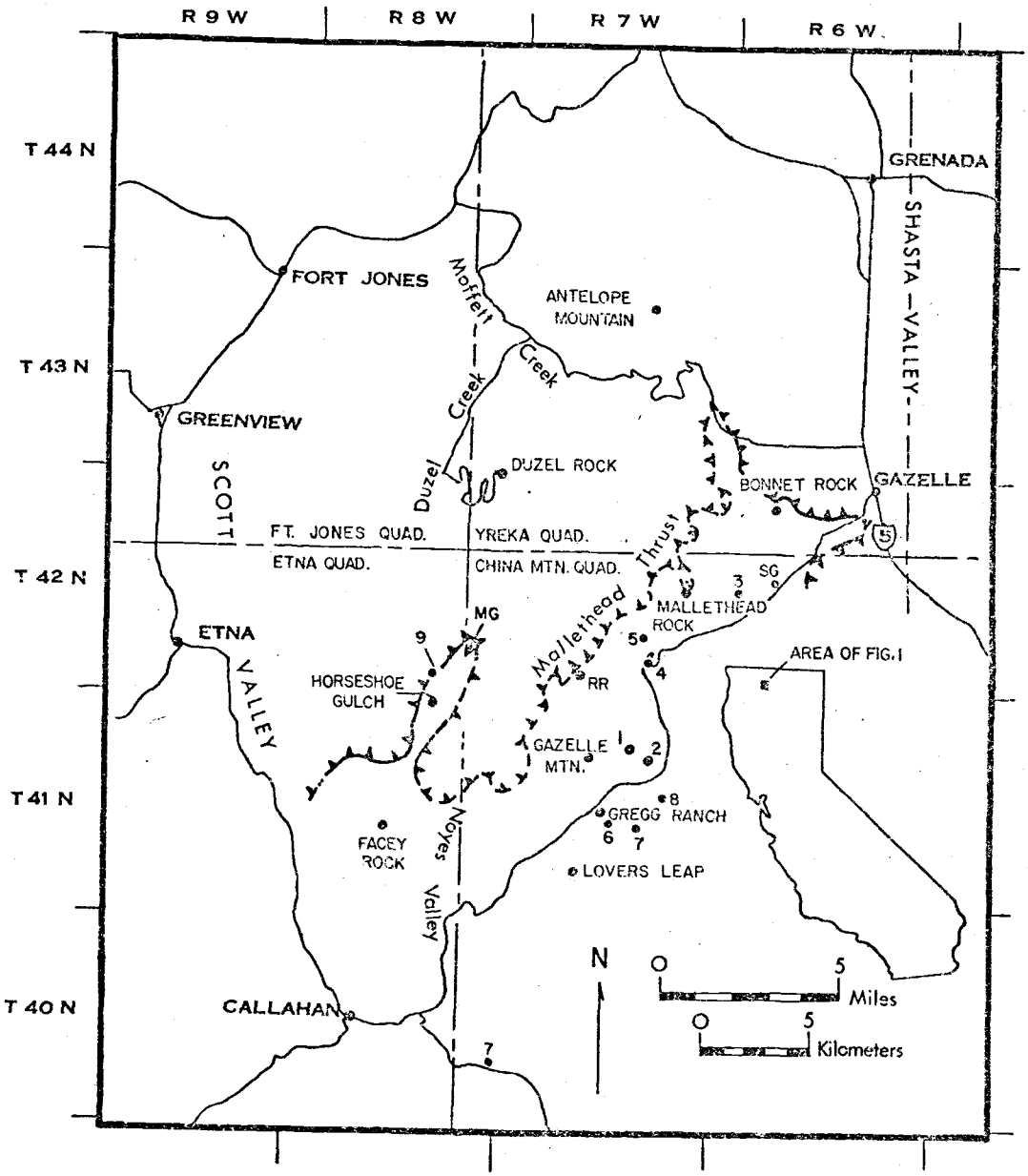

Figure 1. Index map of California showing the location of Duzel Rock and vicinity.

Figure 2. The north and west sides of Duzel Rock as viewed from the Moffett Creek Valley. The cliffs are composed of the Massive Limestone Member of the Duzel Rock Group.

includes all of sections 35 and 36, Township 43 North, Range 8 West, and all of sections 1 and 2, Township 42 North, Range 8 West. The area is federally owned and administered by the Bureau of Land Management. The State of California operates a look-out tower on top of Duzel Rock during the part of the year when fire danger is high, which is usually from June through October.

The map area is readily accessible both from the east and the west on dirt roads that provide access for the look-out tower and for logging operations. A road that branches off Duzel Creek road is parallel to White's Gulch to where it intersects a dirt road along the crest of the ridge. The eastern approach is along Spring Branch to where it also intersects with the dirt road along the ridge. The road that is on the crest of the ridge ends at the base of the cliffs of Duzel Rock; from this point a paved path winds up to the look-out tower on the top of Duzel Rock. The dirt roads that ascend to Duzel Rock connect with good gravel roads that in turn link up with California Route 3 south of Yreka. The remainder of the map area is accessible only by foot.

Purpose and Methods of Investigation

The purpose of this investigation was to determine the structure and stratigraphic relationships of the rocks in the area and interpret their paleodepositional environments. Samples for

micropaleontological study were to be collected of all the carbonate lithologies in an attempt to establish ages for their undated sedimentary units.

Field work began in the later part of June of 1975 and continued through the month of August. Geologic mapping was done on several copies of the U. S. Geologic Survey Yreka Quadrangle topographic map (1954) enlarged ten times photographically, to a scale of ten inches per mile. Data were consolidated on a copy of this enlarged map.

Locations were determined by triangulation, where possible, with a Brunton compass. The dense vegetation, steep slopes, and scarcity of pronounced landmarks made triangulation difficult, hence many locations were determined by pacing along a particular azimuth, using the Brunton compass, from known locations such as a road or a location determined by triangulation. In addition, a hand lens, a bottle of dilute hydrochloric acid, and a rock hammer were used for examination of rocks in the field.

Approximately 700 pounds (318 kg.) of rock samples were collected for laboratory studies. Thirty-five thin sections were made from among these samples for petrographic analysis. Twenty-four samples (each 100 to 200 grams) of the various limestone units were dissolved in concentrated hydrochloric acid. The insoluble residues derived were measured to determine their percentage of the total mass and studied under a binocular microscope to determine their

general character and mineralogical constituents. In mid-July of 1975, Dr. Stig Bergström, a conodont specialist, and Dr. Sven Laufeld, a specialist on chitonozoans, were given a tour of the map area by Dr. Arthur Boucot and the author. These two micropaleontologists collected over 30 samples, each weighing over 20 pounds (9 kg.), to crush up, dissolve, and examine for microfossils. On the same day Dr. Reuben Ross and Dr. Barney Poole assisted in the compilation of a measured section of the White's Gulch Limestone for which Dr. Ross' 100-foot tape measure and a Brunton compass were utilized.

Topography, Climate, Vegetation, and Drainage

The north end of Duzel Rock has an elevation of 6039 feet; this is the highest point in the area of study. A hill of lesser altitude dominates the southern half of the map area, located just west of the section boundary near the southern edge of the map (Plate 1). This hill has an elevation of about 5545 feet. The northwest corner of the area has an elevation immediately below 3600 feet; this is the lowest point in the four sections mapped in this study. Hence the total relief of the area from the northwest corner to the top of Duzel Rock is approximately 2360 feet.

The climate of the area is not extreme. Temperatures rarely dip below 0°F. or exceed 100°F. The average annual precipitation

occurs in the autumn and winter (Mack, 1958); the summers are generally hot and dry. The State of California maintains a continuous fire watch in the Duzel Rock look-out from early June until mid-October; during this period the fire hazard is often very high throughout the forests of the Klamath Mountains.

The vegetation is typical of a semi-arid climate--much of the ground is covered by sage, manzanita, scrub oak, mountain mahogany, and western juniper. Generally, northerly and westerly slopes are covered with dense brush and scrub trees or are heavily timbered with coniferous trees, the most common being Yellow (Ponderosa) Pine.

The lower portions of White's Gulch and Spring Branch have flowing water all year. Most springs and streams dry up in the summer months. The few springs located that had flowing water in late August are on the west slopes of Duzel Rock, at an elevation of approximately 5200 feet.

Exposures

In general the exposures are good. The best ones occur along ridge crests, on southerly and easterly slopes, and along the dirt roads, many of which were constructed along thrust contacts. Dense brush and regolith conceal outcrops locally. Many contacts in the central portion of the map area are obscured by talus from the

massive limestone cliffs.

Previous Work

Wells, Walker, and Merriam (1959) published a study of the Ordovician and Silurian rocks of the northern Klamath Mountains. The Duzel Rock area was mapped as the Duzel Formation, which was assigned an Ordovician age on the basis of fossils found in Horseshoe Gulch, six miles to the south-southwest. The authors described the Duzel Formation as being characterized by phyllitic wackes, and recognized a marble member that forms the cliffs of Duzel Rock. In 1974, Preston Hotz published a preliminary map of the Yreka Quadrangle on a scale of 1:62, 500. Hotz's map treats the complex "Duzel Formation" of Wells and others in greater detail, delineating at least six units where the 1959 map had one formation. The lithologies Hotz named in the Duzel Rock area are the phyllite of Scarface Ridge, the siltstone of Moffett Creek, and the limestone of Duzel Rock which includes interbedded volcanics and minor chert and shale. Hotz (in press) has since formalized the names of several of these units--his new names of Moffett Creek Formation (for the siltstone) and Duzel Phyllite will be used in this study. Richard Porter (1974) in an Oregon State University master's thesis, mapped and described the Facey Rock Limestone, which may be correlative with some of the limestone at Duzel Rock.

STRATIGRAPHY

Regional Stratigraphy

Duzel Rock is located within Irwin's (1966) eastern Klamath belt of the Klamath Mountain Province (Figure 3). Irwin (1960) had previously termed this sub-province the eastern Paleozoic belt of the Klamath Mountains, but rocks of Mesozoic age were later identified in the southern portion of this belt. The northern part of this region is dominated by Paleozoic geosynclinal rocks including graywacke, mudstone, greenstone, chert, limestone and low rank metamorphic rocks. The eastern Klamath belt is separated from the central metamorphic belt to the west by a thin band of ultramafic rocks, and is overlain to the east by the Tertiary and Quaternary volcanic rocks of the Cascade Range.

Local Stratigraphy

Eight major stratigraphic units are described in this study (see Figure 4). The Duzel Rock Group, the Spring Branch Group, and the White's Gulch Limestone are named in this study for the first time. Due to tectonism, none of the eight units are in depositional contact with each other. The uncertainty over the age of at least six of these units renders chronological determinations extremely difficult, but comparison with better-dated areas of similar structure in the

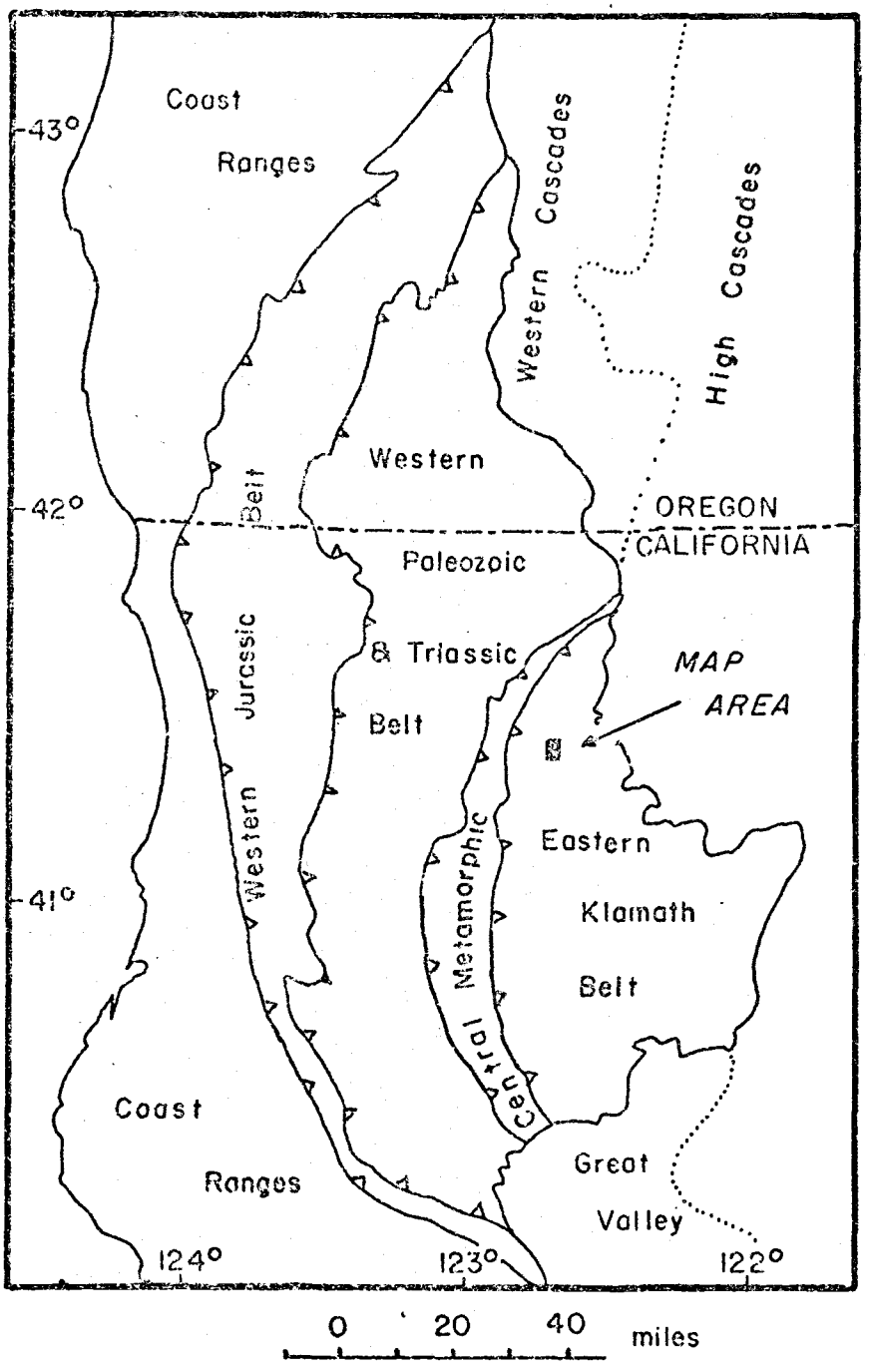

Figure 3. Index map showing study area, subprovinces of Klamath Mountains, and adjoining provinces (from Irwin, 1966).

	Massive Limestone Member	
	Duzel Rock Group	
	(Middle Ordovician?)	
	----------	Subsidiary Thrust
Spring Branch Group	Platy Limestone Member	
	Duzel Rock Group	
(Ordovician?)	(Late Ordovician?)	
	----------	Subsidiary Thrust
	Pillow Basalt Member	
	Duzel Rock Group	
	(Ordovician?)	
----------	----------	Subsidiary Thrust
Unnamed Red Shale		
		Subsidiary Thrust
(Silurian?)		

White's Gulch Limestone (Ordovician?)		
----------	----------	Subsidiary Thrust
Moffett Creek Formation		
(Silurian)		
		Major Thrust
----------	----------	(Subsidiary to Mallethead Thrust)
Duzel Phyllite		
(Devonian-Cretaceous)		

Figure 4. Structural relationship of the major stratigraphic units described in this study.

Klamath Mountains suggests that the older units have been thrust over the younger formations of greater areal extent. The oldest units, all of which have tentatively been assigned Ordovician ages, are the White's Gulch Limestone, the Spring Branch Group and the Duzel Rock Group, the latter of which is divided into three members; these are the massive limestone, platy limestone, and pillow basalt members. The younger units are the unnamed red shale, possibly Silurian in age, the Silurian Moffett Creek Formation and the post-Silurian Duzel Phyllite.

The term "group" is applied in this thesis to an aggregation of lithologic units that are structurally, and sometimes stratigraphically, related to each other. Each of these individual units within a "group" is designed as a "member," as in the Massive Limestone Member of the Duzel Rock Group. This usage is not in strict accordance with the Code of Stratigraphic Nomenclature (Krumbein and Sloss, 1963) in which the fundamental rock-stratigraphic unit is a "formation." Two or more formations comprise a group, and an individual formation can be divided into members. Since most of the lithologic units in the Duzel Rock area do not satisfy the definition of a formation as presented in the Code of Stratigraphic Nomenclature, this term is not applied to any of the units named for the first time in this thesis.

A large thrust fault, probably of a magnitude comparable to

the regional Mallethead Thrust, separates the Duzel Phyllite from the overlying Moffett Creek Formation. Smaller, subsidiary thrusts separate the Ordovician?, largely carbonate, units from the Moffett Creek Formation and from each other. The White's Gulch Limestone in part underlies the Spring Branch Group in the southern portion of the map area, and the Duzel Rock Group in the north. The latter two groups have thus been thrust over both the White's Gulch Limestone and the Moffett Creek Formation. The Duzel Rock Group is in turn divided into three members by thrust faults. Assuming that the platy limestone and massive limestone members of the Duzel Rock Group correlate with similar members of the Facey Rock Limestone, an inverted sequence of Middle Ordovician oomicrites on top of Late Ordovician micrites may be represented here. The unnamed red shale is a sliver separated from the White's Gulch Limestone and Spring Branch Group by thrust faults.

Duzel Rock Group

The name Duzel Rock Group is given to the cliff forming limestones that cap Duzel Rock in the SW 1/4 of section 36 and the NW 1/4 of section 1, and to the basalts and other rocks that lie on the southern and eastern flanks of Duzel Rock. It also includes the smaller block of limestone that lies to the east of Duzel Rock at an elevation of 4800 to 5100 feet, in the NE 1/4 of section 1.

Massive Limestone Member

Distribution and Type Locality

The massive limestone member occurs in two separate blocks; the larger of the two covers approximately 20% of the SW 1/4 of section 36 and extends into the NW 1/4 of section 1. (Much confusion surrounds the usage of the term "massive" in geologic literature. Throughout this thesis this term is applied as an adjective to sedimentary rocks that lack stratification and bedding and are generally devoid of large-scale sedimentary structures). The former is the highest unit topographically in the map area and forms the steep cliffs visible from the Scott River Valley, more than ten miles to the west-northwest. The maximum exposure of the massive limestone member is approximately 450 feet from the base of the unit upwards; this occurs on the northern part of the east slope of the larger block. The type locality is designated as the exposure due east of the upper end of the Forest Service road, located in the NE 1/4 SW 1/4 SW 1/4 section 1.

Lithology

The color of the massive limestone member is slightly variable. Fresh surfaces are commonly light gray (N7) to medium gray (N5) in color, but a few samples are darker, attaining a medium dark gray

(N4) color. Locally, the massive limestone assumes a light bluish gray (5B7/1) color. Weathered surfaces are usually lighter than fresh surfaces and range from very light gray (N8) to medium light gray (N6), but these are frequently stained by iron oxides, which impart a yellowish gray (5Y8/1), grayish orange (10YR7/4), pale yellowish brown (10YR6/2), or even a moderate brown (5YR4/4) color to the surface of exposures.

The weathered surface of the limestone is covered with solution pits, some of which are circular (up to 8 mm in diameter) and several that are elongate; these range up to 10 cm in length and are superimposed on some of the numerous fractures that transect the limestone. The weathered surface of the limestone has a rough texture to the touch; finely disseminated bits of silica that stand out in microrelief account for this.

Solution and brecciation along large scale fractures have produced two valleys in the larger block of the massive limestone (see cross-section). No caves were discovered, in marked contrast to the numerous caves described by Porter (1974) in similar limestones at Facey Rock, approximately eight miles to the south-southwest. It is possible that there were caves in the past where the valleys mentioned above are now located, but that these have since been unroofed by solution and collapse.

The massive limestone member is generally lacking in any

stratification or bedding features (see Figure 5). Some alternations of light and dark gray near the top of the exposure give a crude impression of bedding. This rare banding indicates the massive limestone member lies approximately horizontal in its present position. On the whole, however, stratification is virtually absent, which is characteristic of strata that have been subject to continuous bioturbation or have been deposited rapidly.

A widespread feature of the massive limestone member is the abundance of small fractures, most of which are filled with sparry calcite. Some of the larger fractures, however, are filled with a porous mass of carbonate and iron oxide minerals. The majority of these fractures are less than 2 mm thick, although a few attain a thickness of 9 mm. For the most part, the calcite-filled fractures do not extend more than 25 cm in any one direction. The fractures filled with the carbonate-oxide mixture frequently extend much farther and seem to be the youngest of the fractures, as they transect all other fractures. Many of the fractures filled with sparry calcite end abruptly at these larger, more porous fractures.

The lower contact of the massive limestone member is in most places obscured by talus (Figure 6). At three localities, however, laminated chert can be seen to underly the massive limestone. This chert attains its maximum discernible thickness of 19 feet near the southeast corner of the larger block of massive limestone; it ranges

Figure 5. The south face of Duzel Rock, looking east. Note the characteristic lack of stratification in the Massive Limestone Member of the Duzel Rock Group. Mt. Shasta (14, 162 ft.) and the Cascades Province in the background. NW 1/4 NE 1/4 NW 1/4 section 1.

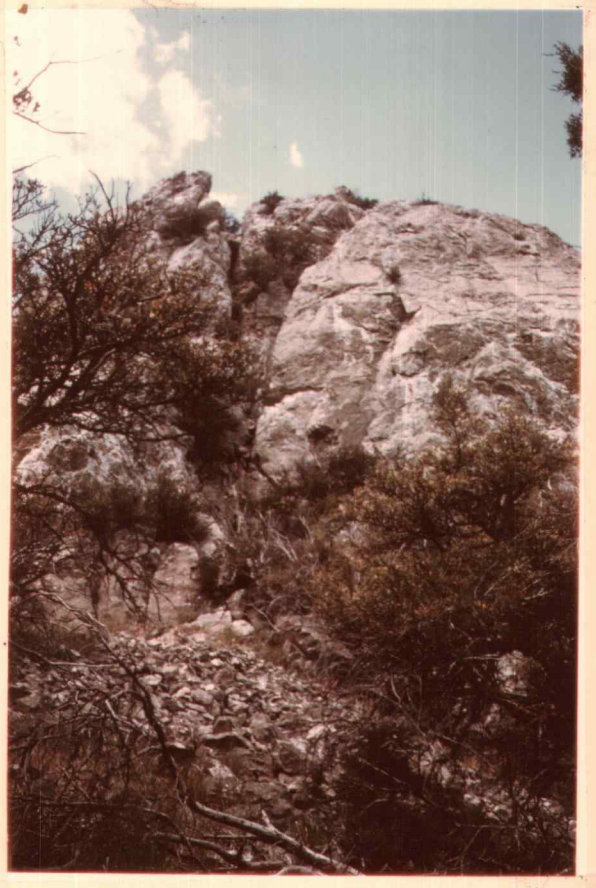

Figure 6. The south face of Duzel Rock, view to the north. Talus at the base of the cliff obscures the contact of the massive limestone with the underlying Platy Limestone Member of the Duzel Rock Group. NW 1/4 NE 1/4 NW 1/4 section 1.

in color from very light gray (N8) to a grayish blue (5PB5/2), and contains laminae stained grayish orange (10YR7/4) and grayish red (10R4/2) by iron oxides. The chert is extremely fractured and faulted on a small scale (Figure 7).

In thin section study of this chert, no trace of radiolaria or sponge spicules were discernible. The chert is composed of fine-grained chalcedony, and is being replaced in irregular splotches by calcite and very small (less than .01 mm) rhombs of dolomite. Brown and orange laminae in thin section are comprised of iron-bearing carbonates, oxides, and clay minerals. Much of the silica has been mobilized and carried by pore-fluids into the overlying massive limestone; an insoluble residue of a sample of the immediately overlying limestone (DZ20) contained an anomolously higher silica content (Appendix Table 1).

A primary depositional origin for these blue cherts seems most plausible. No regular laminations are found in the massive limestone and the contact between chert and massive limestone is a sharp one; it isn't likely that the chert is the result of silicification of the massive limestone member. The evidence from thin section study suggests the reverse: that carbonate minerals from the massive limestone member have been carried in solution along fractures and are irregularly replacing the chalcedony in the underlying chert. Some authors have noted that cherts fromed from biogenic oozes "have no necessary

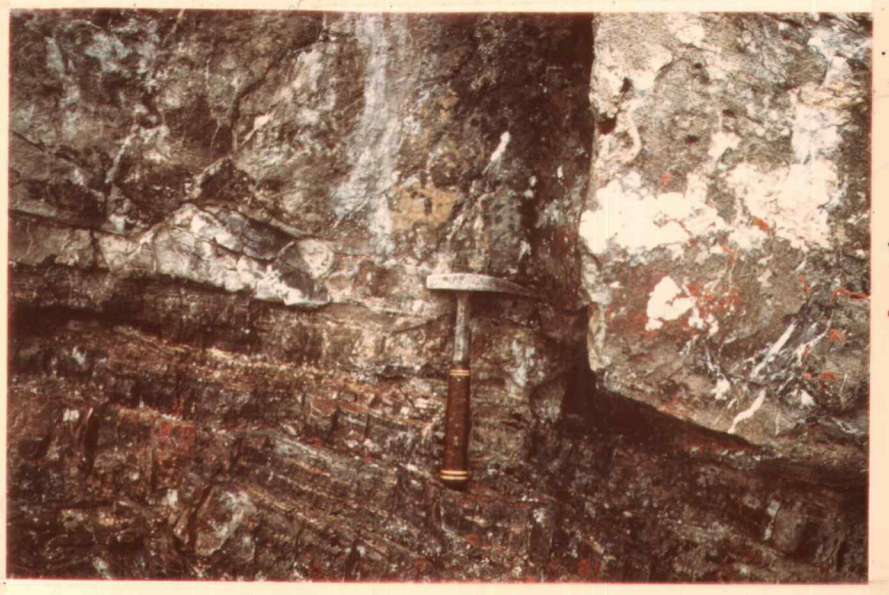

Figure 7. Laminated, thin-bedded blue chert underlying the Massive Limestone Member of the Duzel Rock Group. Note closely spaced fractures and small, nearly vertical fault at right. NW 1/4 NE 1/4 NW 1/4 section 1.

relationship to water depth" (Blatt, Middleton and Murray, 1972, p. 538); some Mesozoic spiculites have been useful shoreline indicators in deltaic sequences in the Appalachian Plateau (Cavorac and Ferm, 1968). In shallow water, the formation of chert by inorganic solution and precipitation would be most likely in an enclosed or semi-enclosed basin where evaporation could increase the pH of the water and corrode any available siliceous sediments or volcanic rocks; lowering of the pH by subsequent dilution would cause the precipitation of hydrous sodium silicates which with time convert to chert (Eugster, 1967). Hence it is at least plausible that these cherts were laid down in a shallow marine setting.

In thin section, the massive limestone is seen to contain a considerable percentage of allochemical constituents; visual estimates range from 40-60%. These allochems are supported by an original micrite matrix that is almost completely recrystallized to microspar and pseudospar. Visual estimates, using Dunham's (1962) classification, indicate that most samples are not in grain support, although portions of one thin section contain grains packed closely together. Large areas of the massive limestone are extensively recrystallized, and even samples in which textures are well preserved show several types of recrystallization features. In one thin section the outline of most allochems are well preserved, while many interior features have been obliterated or selectively replaced. Other samples are so

extensively recrystallized that all depositional textures have been obliterated. The most abundant allochems observed in the massive limestone are concentrically coated grains; portions of one thin section contain over 50% coated grains. Most of these allochems are spherical enough to be identified as ooids, but a few are irregular and are suggestive of algal encrustations.

The allochems range in size from smaller than 0.10 mm to about 2.0 mm for ooids and 6.0 mm for intraclasts. The average grain has a diameter of roughly 0.5 mm. The sorting of the allochems is not easily quantified because the grain diameters observed in thin section are not necessarily the maximum diameters. Visual estimates and comparison with other sedimentary rocks yield a description of moderate sorting, using Folk's (1968) parameters. The size and sorthing of these carbonate grains are not necessarily good indicators of the energy of the depositional environment as they would be for a clastic sediment; most allochems are derived from the same basin in which they were deposited and their size and sorting has not been affected by long periods of transport and reworking.

Over most of the outcrop area of the massive limestone, the silica content is very low. Total insoluble residue (most of which is silica, the remainder is detrital clay) is usually less than a half of one percent. One sample (DZ20), collected immediately above the banded blue chert described above, contained an insoluble residue

of over 13 percent, most of which was silt-size silica, presumably derived from the solution and reprecipitation of the underlying chert. The silica in the insoluble residues of the other samples (DQ4, DZ78) of the massive limestone is for the most part in the form of irregular blades and platelets up to 5 mm across. A few small (less than 1 mm across) silicified ooids also occur in the insoluble residues. It is likely that most of the platelets represent microvein fillings, since a few chert veins were observed in thin section.

Recrystallization in the massive limestone is extensive. The micrite matrix has undergone extensive neomorphism and is now a mosaic of microspar and mostly pseudospar. The average grain size of this neomorphic calcite is roughly 35-40 microns, which is close to the boundary of 30 microns between the sizes of microspar and pseudospar. This neomorphic calcite is distinguishable from pore-filling calcite by its irregular and indistinct boundaries, and by its darker color due to abundant suspended contaminants, such as clay and finely disseminated organic material. This recrystallization can be termed "aggrading, coalescive neomorphism" as described by Folk (1965).

All of the ooids have undergone various phases of neomorphic recrystallization as well. Most of these allochems have recrystallized to pseudospar of larger crystal size than the pseudospar in the matrix. In these ooids the concentric laminae are still easily discernible. In

some ooids, solution has taken place along certain laminae; the porosity developed has since been occupied by clean, pore-filling calcite. This type of selective "moldic porosity" has been described by Friedman (1964) in modern carbonate sediments from the Bahamas and Florida. Apparently this development of porosity and later destruction by pore-filling calcite can occur very early in the diagenic history of a limestone. A few of the ooids have had all but the outer one or two laminae dissolved, with the core later infilled with coarse sparry calcite.

There is a small amount of organic matter in the massive limestone member, most of which is concentrated along numerous microstylolites which transect the rock. During insoluble residue analysis this organic portion was apparent as an oily, black film which formed on the surface of the acid. Apparently, this insoluble organic matter has been concentrated along the stylolites by solution of the adjacent carbonate. In most instances, solution along these zones has removed only a small volume of rock; where the stylolites transect the larger allochems, a portion of the allochem is present on either side of the solution surface. The early concentration of organic matter and detrital clays along the microstylolites may have inhibited further solution.

The massive limestone member is riddled with microfractures that are filled with relatively coarse, clear sparry calcite. These

microfractures are related to the numerous systems of large fractures visible in outcrop. Generally, the higher the concentration of microfractures in a particular thin section, the more extensive the recrystallization and the less recognizable are any of the primary allochemical constituents. There is generally little or no displacement of allochems that are cut by these calcite-filled microfractures. Many of the microfractures end abruptly upon coming into contact with the solution surfaces (stylolites) described above; these fractures obviously predate the stylolitization. A late generation of fractures transect all features: other fractures, allochems and stylolites--apparently there have been several generations of fractures and infilling with calcite; these several sets of fractures are a manifestation of the intensity and longevity of the tectonic history of this region.

Idiomorphic rhombs of iron-bearing carbonates are present in very minor amounts, scattered through the matrix of the massive limestone member. These rhombs of ankerite, or iron-rich dolomite, are yellowish brown in color, with the color being darkest around the edges of the crystals. The iron was probably transported by pore fluids percolating along the numerous fractures that transect the rock.

The paragenetic sequence of the massive limestone member is apparently as follows: 1) Induration of a lime mud containing a high proportion of ooids and intraclasts. 2) Neomorphic inversion of the original aragonite in the ooids to coarsely crystalline calcite.

3) First episodes of tectonic fracturing and vein-filling. 4) Deposition of silica in pore spaces. 5) Solution and concentration of clays along stylolites. 6) Coalescive neomorphism of the matrix to microspar and pseudospar. 7) Last episodes of fracturing and vein-filling.

In Dunham's (1962) textural classification of carbonate rocks, this limestone, where primary textures are intact, would be classified as a grainstone. Using Folk's (1954) classification this limestone would be called a partially recrystallized oomicrite.

Age

No microfossils were recovered from the massive limestone member. The only way to assign an age to this member is by lithologic correlation with the massive limestone found at Facey Rock (Porter, 1974). At Facey Rock the same sequence is found as at Duzel Rock, that is, a massive limestone lying on top of a highly disturbed, laminated, platy limestone, the two units being separated by a thrust fault. The massive limestone at Facey Rock, which is lithologically similar in many respects to the massive limestone member of the Duzel Rock Group, has been dated by Dr. Carl Rexroad of the Indiana Geological Survey as Middle Ordovician on the basis of conodonts (Porter, 1974). It is likely then, that the similar massive limestone at Duzel Rock is also Middle Ordovician.

Structural Relationships

The lower contact of the massive limestone member is in most places covered by talus. At the northernmost exposure it can be seen to be lying unconformably over the Moffett Creek Formation. The contact has the appearance of a fault as both the limestone and the siltstone are highly fractured. The best evidence that the massive limestone has been thrust overy the platy limestone member and the Moffett Creek Formation is a large shear zone that occupies much of the NE 1/4 SW 1/4 SW 1/4 and the SE 1/4 SW 1/4 SW 1/4 of section 36 (Plate 1). This zone is well exposed along the dirt road, and in it blocks of platy limestone, calcareous siltstone and shale and porphyritic intrusives have been extremely sheared. Attitudes of beds are usually vertical in the thrust zone and many near vertical faults transect the zone. There is also a small outcrop of a tectonic breccia visible on the northeast side of Duzel Rock, below the lower contact of the massive limestone.

Thickness

A determination of the thickness of the massive limestone member is made difficult by the virtual lack of internal structure. Vague bedding visible in one locality near the summit of Duzel Rock indicates the limestone is in an approximately horizontal position. If this is

true of the entire member, then the maximum exposed thickness is roughly 500 feet (150 m).

Environmental Interpretation

The occurrence of ooids and intraclasts supported in a micrite matrix is indicative of a textural inversion. Folk (1962) states that this type of textural inversion is the result of mixing of sediments from different depositional environments, and while this mixing may be accomplished in several ways, it is usually indicative of a transition zone between high and low energy environments. Fluctuating high and low energy currents may lay down alternations of well sorted, coarse, allochem-dominated sediments and micritic sediments. Bioturbation or violent currents prior to lithification can mix these sediments, producing an oomicrite. Alternately, winnowed high energy carbonate sediments may be infilled with micrite and silt, but this process is usually characterized by grain support and sheltered porosity, both of which are generally absent in the massive limestone member of the Duzel Rock group, as is any evidence of bioturbation. Mixed sediments may also be produced by periodic, violent storms capable of transporting high energy sediments into a low energy environment; Folk's model for such a situation involves carbonate sand from on off-shore shoal or bar being washed into an adjacent sheltered, micrite-filled lagoon.

It is this last interpretation that Porter (1974) favored in interpreting the depositional environment of the massive limestone at Facey Rock. He invokes similar modern barrier bars and protected lagoons in the Bahamas as models, as described by Newell and others (1960). Porter observed numerous chert nodules and evidence of dolomitization within the massive limestone at Facey Rock (both of these features are absent at Duzel Rock); he envisions a lagoon with restricted circulation which may have generated hypersaline and/or silica-rich zones to explain these features. Such hypersaline conditions occur today on the Bahama Bank, as described by Cloud (1970).

While interpreting the massive limestones of Facey Rock and Duzel Rock as lagoonal micrites periodically receiving coarse barrier-bar sediments is certainly plausible, this author favors a deeper water origin, specifically, an environment similar to the deeps seaward of the escarpments of the present day Bahama Banks. In such a setting, tractive current activity is likely to be minimal and sporadic, allowing for the accumulation of micrite. As ooids commonly form along the margins of present-day carbonate banks, it is likely that periodically ooids would wash over the bank margins, and be deposited in deeper water. The advantages of such an interpretation over that of a lagoon-type setting is that the deep water seaward from a bank is a more permanent setting, which would allow for the continuous deposition of micritic sediments as observed on the thick

deposits of Facey Rock and Duzel Rock. In a shelf lagoon, facies may tend to migrate over long periods of time; one may expect beds of coarser, more winnowed sediments to be interstratified with finer ones. In addition, this deeper water environment would be more extensive areally, which would better explain two blocks of similar limestone surviving the extensive tectonism the Klamath Mountains have undergone. Finally, a deeper water model is not at all incompatible with the substrate of bedded chert that has been described above as underlying the massive limestone at Duzel Rock. While such cherts are commonly deposited in deep water, they have not been reported from the lagoonal facies of any modern day carbonate banks. The chert nodules and evidence of dolomitization that Porter described from the massive limestone at Facey Rock may be due to the effects of diagenesis that took place after these limestones were lithified and buried.

Platy Limestone Member

Distribution and Type Locality

The exposure of the platy limestone member of the Duzel Rock Group is continuous around the southern and eastern margins of Duzel Rock (see Plate 1). In several places this exposure is covered by talus and rockslides. There is no exposure of the platy limestone

on the north or east sides of Duzel Rock, except for a few slivers observed in a thrust zone below the lower contact of the northeast corner of the massive limestone member. The type locality of the platy limestone member of the Duzel Rock Group is designated as the exposure along the road in the NE 1/4 NW 1/4 NW 1/4 section 1.

Lithology

The platy limestone member of the Duzel Rock Group has a yellowish grey (5Y7/2), light olive gray (5Y6/1), pale yellowish brown (10YR6/2), or a dark yellowish orange (10YR6/6) color on a weathered surface. On a fresh surface, the most common colors are medium dark gray (N4), dark gray (N3) and olive gray (5Y4/1). Slopes developed on the platy limestone are gentle, as this unit has a low resistance to weathering. This limestone has a platy or flaggy fracture, splitting along thin argillaceous layers parallel to the bedding planes (Figure 8). Individual plates do not usually exceed 7 cm. in thickness.

The surfaces (bedding planes) of these limestone plates are laced with numerous intersecting sets of fractures. Most of these fractures are very thin, but a few exceed 1 cm. in thickness. These larger fractures are filled with sparry calcite which is usually white but occasionally will have a dark gray or black color.

This limestone is thinly bedded in most places, but in a few localities it is finely laminated. Thin laminae of argillaceous

Figure 8. Platy Limestone Member of the Duzel Rock Group. Looking NNW. Flaggy fracture along bedding planes is readily apparent. NW 1/4 NE 1/4 NW 1/4 section 1.

material separate thicker limestone strata up to 7 cm. in thickness. The argillaceous laminae are almost always fractions of a millimeter thick, although occasional laminae may exceed 2 mm. On a fresh surface these laminae are often not apparent; they show up best on surfaces that have been etched by differential weathering.

More than 90% of this limestone member is very fine grained. There are a few anomalous beds 1.5-15 cm. in thickness, of a medium gray (N5) conglomeratic limestone. Clasts range up to 2 cm. across and are predominantly composed of fine-grained mafic igneous rocks and perhaps pumice, most of which have been partially or totally weathered to brownish yellow clay. Intraclasts of micritic limestone abound, and a few concentric coated grains are present. The matrix is micritic limestone.

One bed of the platy limestone member was found that had well-defined bottom marks (Figure 9). These have the external appearance of load casts, but lack the characteristic trough-like internal laminae. Instead the proturberances contain truncated planar laminations that are parallel to those of the rest of the bed. It is possible that these are scour marks, formed by some erosive tractive current, and later infilled by subsequent sedimentation. The preservation of such scour marks until the time they were infilled would indicate an environment of relatively deep water, at least below wave base.

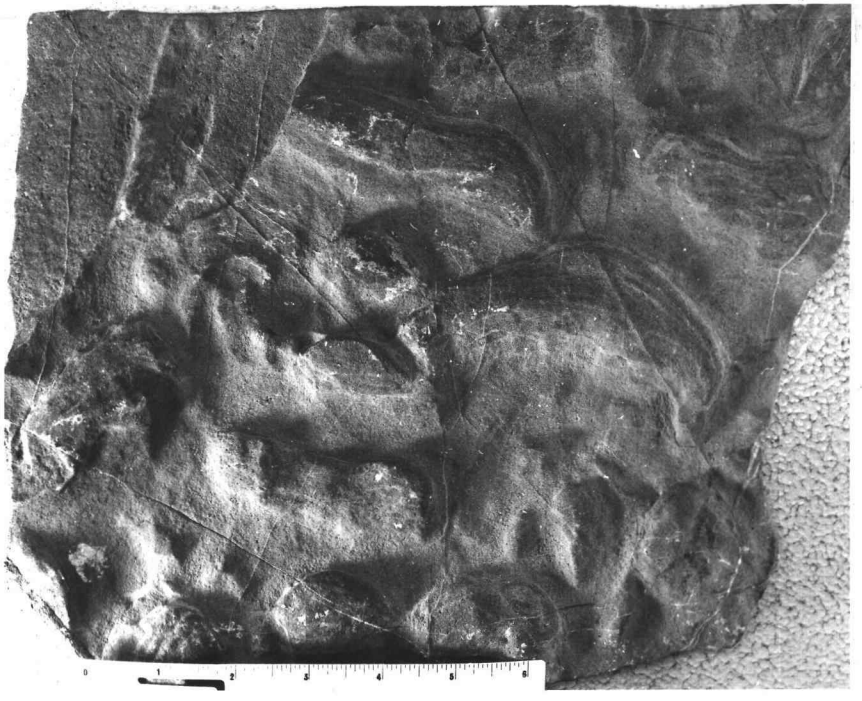

Figure 9. Bottom marks found in float of Platy Limestone Member of the Duzel Rock Group. Scale in inches. SW 1/4 NE 1/4 NW 1/4 section 1.

The attitudes of the platy limestone member are everywhere highly distorted. Large differences in the attitude of bedding over short areas are common, and strata that have been subjected to a high degree of flexure are abundant (Figure 10). The deformation is believed to be attributable to tectonic faulting and minor folding.

Features of the platy limestone that are discernible in thin section are clay laminations, microstylolites, calcite grains, and a few silt-size grains of chlorite and what appear to be shards of altered glass. The clay laminations are common; individual laminae are usually smaller than 0. 2 mm. in thickness. Examination of the thickest laminations reveals they are actually closely spaced multiple laminations. It is along the thicker laminations that the chlorite grains are discernible; these are sub-equant and may be derived from the alteration of biotite flakes or small amphibole crystals. Microstylolites are common, and there is frequently a film of clay minerals concentrated along them. Where microstylolites intersect fractures the stylolites are in most cases offset or terminated indicating that most of the tectonic fracturing post-dates the solution that created the stylolites. There are several sets of fractures, the oldest of which runs parallel to the bedding plane. It is along the margins of these that most of the altered, shard-like grains are located. A few of these glass shards contain tiny crystalites within them. Fractures that transect the bedding are for the most part

Figure 10. Small scale folds in Platy Limestone Member of the Duzel Rock Group. Pen for scale. SE 1/4 SW 1/4 SW 1/4 section 36.

younger than the filled fractures that parallel the bedding, although a few of these transecting fractures are severely offset, indicating they are the oldest. All sets of fractures have been offset by numerous microfaults. This small-scale faulting was the last event to affect the platy limestone member.

Insoluble residues of the platy limestone member range from 4 to 21% (this maximum is derived from a conglomerate layer and is comprised mostly of diagenic silica platelets). Insoluble residues from the micrite are almost entirely detrital clay and are of a dark gray (N3) color. Thin platelets of silica are common, and are probably the remnants of vein-fillings.

The high proportion of detrital clay and small grain size of the calcite suggests the platy limestone member was originally a micrite which has undergone considerable recrystallization. Most of the grains are in the pseudospar range and are fairly equant. The type of recrystallization is thought to be of an aggrading, coalescive nature (Folk, 1965).

A distinct limestone lithology has been included in the platy limestone member of the Duzel Rock limestone and is mapped as such. This limestone occurs in isolated blocks up to 100 ft. long along the strike of the bedding, and up to 40 ft. measured across strike; these blocks are all located in the highly sheared and faulted area south of the bulk of the platy limestone member (Plate 1). This

limestone is yellowish gray (5Y8/1) to light gray (N7) on a weathered surface and light gray (N7) or medium light gray (N6) on fresh surfaces. This limestone is thick bedded--beds range from 0. 5-2. 0 cm. thick. The beds are separated by relatively large stylolites and clay laminations. This limestone lacks the platy fracture of the of the typical platy limestone member. A striking feature of these blocks of limestone is the abundance of highly contorted beds (see Figure 11). This folding is believed to be a form of soft-sediment deformation, as in many instances beds above and below a folded layer are essentially undisturbed. Thin section examination reveals that this limestone contains considerably more clay than the typical platy limestone member. In addition to concentrations of clay along stylolites, there is a large amount of disseminated detrital clay throughout the rock in the form of small discontinuous microlayers and blebs. The insoluble residue of this limestone is high (approximately 25%) and is composed almost entirely of clay minerals.

Age

No fossils were recovered from the platy limestone. The platy limestone at Facey Rock has been dated by Dr. Rexroad as Late Ordovician (Porter, 1974), and on the basis of lithologic and structural similarities, the platy limestone member of the Duzel Rock Group may be of Late Ordovician age as well.

Figure 11. Isolated blocks of atypical lithology included in the Platy Limestone Member of the Duzel Rock Group. Note recumbent fold under rock hammer, believed to be a product of soft-sediment deformation. NE 1/4 SW 1/4 NW 1/4 section 1.

Structural Relationships

Presumed ages imply that the platy limestone and massive limestone members have been inverted by tectonism so that the older rock now rests on the younger, as is the case at Facey Rock. The upper contact of the platy limestone member is a thrust fault, as evidenced by the thrust zone described in the section on the massive limestone member. The extreme deformation visible in the platy limestone member suggests there has been considerable thrusting and folding within the unit itself. The lower contact is also a thrust fault, as the platy limestone is separated from the underlying pillow basalts below by a thick tectonic breccia which includes boulders of platy limestone, massive limestone, basalt and other exotic rocks. This breccia is extensive enough to merit description in a separate section.

Thickness

The extreme amount of internal faulting and folding within the platy limestone member makes an accurate determination of the thickness of the unit impossible. Individual beds may have been fractured into segments that were later thrust over each other so as to create a greater thickness than was originally present. The least deformed area of this member is in the northernmost exposures; here the

maximum thickness is approximately 250 feet (75 m.).

Environmental Interpretation

These limestones probably represent a deep water facies. The predominance of micrite and the clay laminations are characteristic of quiet water conditions. The occurrence of scour marks that were preserved until they were infilled is consistent with a deep water interpretation. The deep basin in which these limestones were deposited received periodic influxes of detrital sediment from distant volcanic highlands, as evidenced by the shards of altered glass, and abundant clay laminae. The thick, multiple laminations probably record periods of flooding or severe storms in the distant landmass. The few beds rich in basaltic rock fragments and intraclasts were probably derived from such a landmass, transported to the deep marine basin, where the platy limestone member was deposited by tractive currents.

Wilson (1969) studied deeper-water carbonate environments and cited as characteristic of these settings the predominance of lime mud and millimeter laminations, and the absence of evidence of bioturbation. All these features are characteristic of the platy limestone member of the Duzel Rock Group.

The distinct, thickly-bedded limestone lithology included in the platy limestone member is considered by me to have formed in a

transitional, slope environment, perhaps landward of the deep basin where the bulk of the platy limestone member formed. The thickly bedded limestone has considerably more detrital clay than the platy limestone, indicative of closer proximity to a sediment-supplying landmass. In addition, the occurrence of contorted layers that appear to have been deformed while the sediment was still soft is indicative of setting with a considerable slope, since such soft-sediment deformation is usually the result of some type of gravity sliding. Whether this distinct lithology is penecontemporaneous with the bulk of the platy limestone member is impossible to determine, as it is nowhere found in depositional contact with the platy limestone, but rather is always fault bounded. It is certainly possible that these two lithologies were produced at different times, in different basins distantly removed from each other, and have since been juxtaposed by the severe thrusting that has affected the entire Klamath Mountains Province.

Pillow Basalt Member

Distribution and Type Locality

The pillow basalt member lies on the southern flank of Duzel Rock, in the center of the NW 1/4 of section 1. A few small slivers of the basalt are situated in the platy limestone member in the

NE 1/4 NW 1/4 of section 1. The type locality of the pillow basalt member is the outcrop above the basalt breccia where the pillows are best developed, in the SW 1/4 NW 1/4 NW 1/4 of section 1.

Lithology

Most of the weathered surfaces of the pillow basalt member are stained by clays and iron oxides; dark yellowish brown (10YR 4/2), pale brown (5YR 5/2) and moderate brown (5YR 4/4) are the most common colors. On a fresh surface the basalt has a medium dark gray (N4) or a medium bluish gray (5B 5/1) color. The basalt is vesicular in places but is much more frequently amygdaloidal, the vesicles being filled with white sparry calcite.

Many of the vesicles are elongated so as to give the appearance of tubes, the largest of which are 3 cm. long. In addition, the basalts are highly fractured, and many of the intersecting sets of joints have been enlarged so that they are up to 3 mm. wide. Many of these fractures are filled with white calcite like that in the amygdules.

The most striking feature in outcrop is the occurrence of pillow structures, which are well developed in a few localities and only vaguely discernible in others. Individual pillows average 1 m. in diameter, the largest measuring 1.7 m. across. Pillow structures in basalts are indicative of subaqueous extrusion, hence these basalts

were probably extruded either in an oceanic basin or in an arc or sea-mount setting.

In thin section, the basalts have a diabasic texture. Plagioclase laths oriented at random comprise 44% of the rock on the average; most of these have been severely altered to sericite, kaolinite, and calcite. Only a few laths with albite twins were discernible; these yielded an anorthite/albite ratio in the andesine range. Filling the interstices between plagioclase laths are subhedral clinopyroxene crystals, which comprise 24% of the basalt on the average, and irregular patches of magnetite, an average of 20%. There are a few crystals of olivine, which are altering to iddingsite, and abundant (as much as 3%) needles of ilmenite which intersect each other at right angles. Much of the pyroxene has altered to chlorophaeite and serpentine minerals. The groundmass between the larger crystals accounts for approximately 10% of the slides, and has been severely altered to clay minerals and hematite. Its original composition was probably a somewhat equal mixture of plagioclase and pyroxene with a considerable amount of magnetite.

Microveins and vesicles in thin section are for the most part filled with sparry calcite. A few are filled with chalcedony, and occasional veinlets are filled with chlorophaeite and other green alteration products.

Three samples of the pillow basalts were collected by Al Potter,

a doctoral candidate in geology at Oregon State University, and chemical analyses were done on these by Potter, and by Ken Scheidegger and Jack Corliss, both of the School of Oceanography at Oregon State University. Table 1 presents the percentages by weight of the major oxides, and the amounts of rare earth elements in parts per million. The ratio of yttrium to niobium can be used to indicate whether a basalt is a tholeiitic or an alkali olivine basalt, as these elements are relatively stable under metamorphic changes (Pierce and Cann, 1973). The Y/Nb ratios for the pillow basalts in the Duzel Rock Group were all below 2, indicative of an alkali olivine basalt. Similarly, the probable tectonic setting of a basalt can be derived by plotting the amounts of titanium, zirconium, and yttrium on a graph such as Figure 12 following the hypothesis of Pierce and Cann (1973). The pillow basalts all fall in field D, which are "within-plate basalts" such as those extruded on a sea-mount or oceanic island like Hawaii. Using this evidence, in conjunction with the presence of pillow structure, it may be inferred that these basalts were extruded in a sea-mount or shoal setting, removed from any convergent or divergent plate boundary.

Contacts

Both the upper and lower contacts of the pillow basalts are highly sheared and have all the appearances of thrust faults. The

Table 1. Chemical analysis of the Pillow Basalt Member of the Duzel Rock Group.

Constituent	Amount of sample (oxides in percentages, elements in ppm)		
	DR4	DR6	JG842-1-6
SiO_2	49.36	49.59	50.75
Al_2O_3	15.07	16.64	13.68
FeO	13.31	11.09	12.90
MgO	5.47	4.38	5.32
CaO	8.53	10.49	9.00
Na_2O	4.40	4.16	3.89
K_2O	0.73	0.69	0.53
TiO_2	3.13	2.96	3.60
Total	100.00	100.00	99.67
Ti	18764	17745	21582
Y	38.3	37.8	35.5
Nb	27.2	26.8	23.2
Zr	240.8	203.8	199.5
Cr	107	135	61
La	23.2	16.4	20.5
Ce	41.9	38.3	39.6
Sm	7.7	7.5	7.4
Eu	3.3	2.9	2.4
Tb	1.1	1.1	1.1
Yb	3.4	3.3	2.9
Lu	0.4	0.48	0.33

Localities DR4 DR6, and JG842-1-6: SW 1/4 NE 1/4 NW 1/4 section 1

Data provided by K. F. Scheidegger, A. W. Potter, and J. B. Corliss

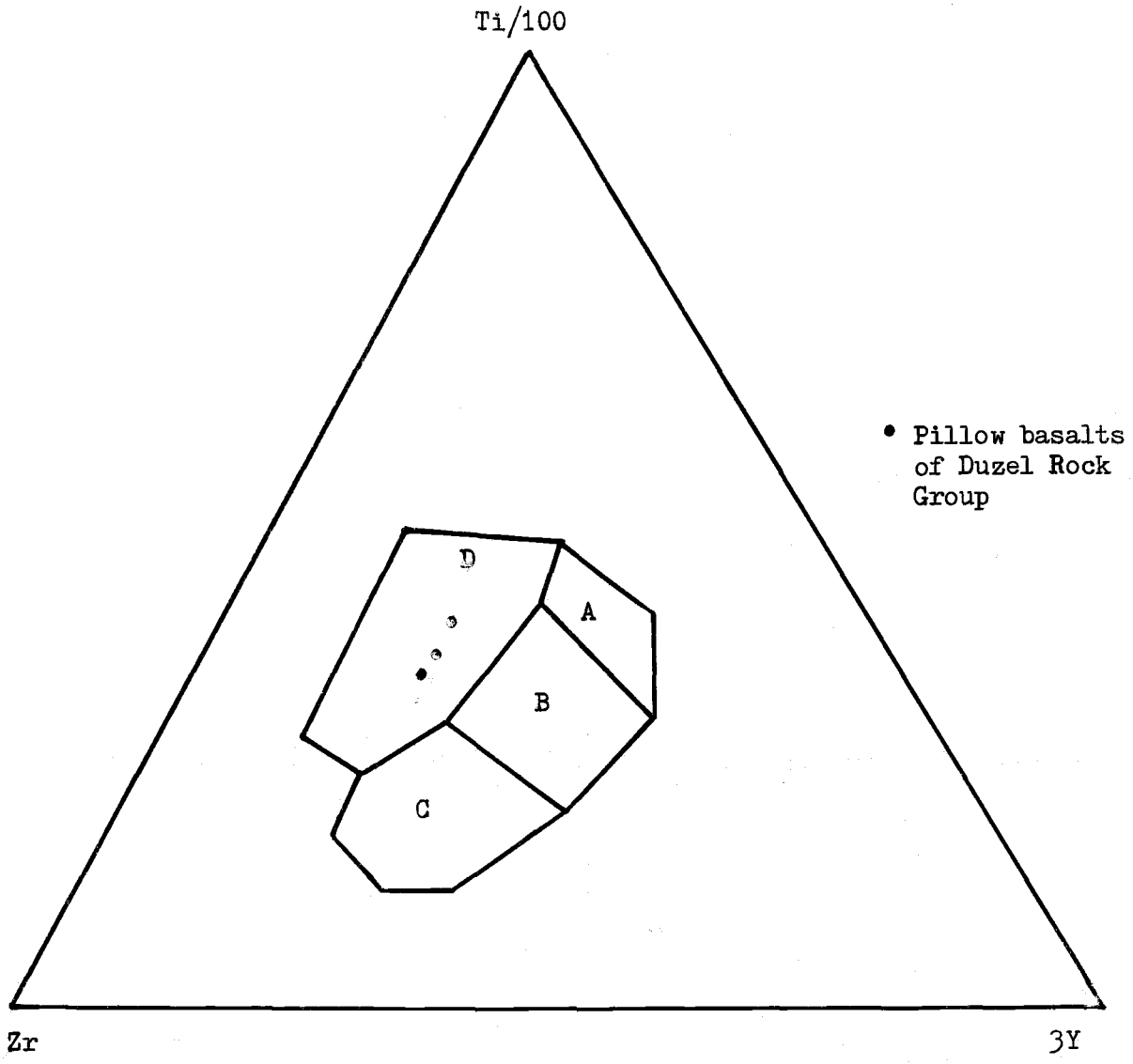

Figure 12. Differentiation diagram using Ti, Zr, and Y. A and B: low potassium (island arc) tholeiites. B: ocean floor basalts. C and B: calc-alkaline basalts. D: within plate (i.e. ocean island or continental) basalts. After Pearce and Cann (1973).

pillow basalts can be thought of as one of a series of fault slices resting on top of the Moffett Creek Formation, the most pervasive unit in the map area. Both the upper and lower contacts are bordered by basalt breccia, of a probable tectonic origin, and the eastern margin is a large thrust contact with the Moffett Creek Formation.

Age

The age of the pillow basalt is problematic, since it is not in depositional contact with any sedimentary strata that could be dated. It is certainly older than the tectonic event that faulted it into its present association, and much of this tectonism may have occurred in the Jurassic (see Geologic History below). It is tempting to assume that these basalts are of the same age as the limestones in the Duzel Rock Group and are Ordovician, but this is uncertain.

Basalt Breccia Member

Distribution and Type Locality

The basalt breccia member occurs in three bands running east-west along the south flank of Duzel Rock in the NW 1/4 of section 1 (Plate 1). The type locality is designated as the exposure below the basalt in the NE 1/4 SW 1/4 NW 1/4 of section 1.

Lithology

On a weathered surface, the most common colors the breccia exhibits are grayish olive green (5GY 3/2), dark greenish gray (5GY 4/1), light olive gray (5Y 5/2), and dark yellowish brown (10YR 4/2). The most common colors on a fresh surface are dark greenish gray (5GY 4/1) and dusky yellow green (5GY 5/2).

The breccia contains a wide variety of clasts, most of which are basaltic, and all are subangular to very angular. The clasts range in size from 1.0 m. by 1.0 m. by .5 m. for the largest boulders to microscopic fragments. Many clasts are surrounded by a rim of white sparry calcite, and the matrix is mostly sparry calcite with a small percentage of green alteration products (Figure 13). Large portions of the basalt breccia are extremely friable while other zones are much more cohesive.

In thin section, the majority of the clasts are finely crystalline, mafic volcanic rocks, many of which are similar to the pillow basalt member. The vast majority of these exhibit extensive alteration; some are completely altered to a mixture of chlorite, serpentine and clay minerals. Limestone clasts are second in abundance, and several types of limestones are present. Platy limestone and massive, oolitic limestone much like those that occur in place above the breccia are the most common carbonates, but a few exotic limestones

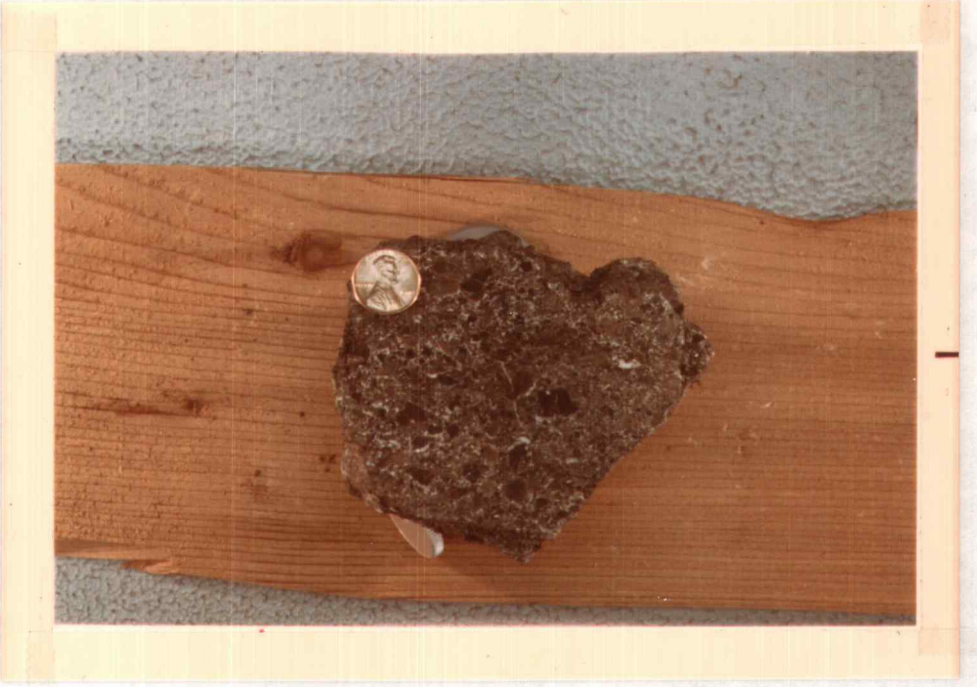

Figure 13. Sample of the Basalt Breccia Member of the Duzel Rock Group. Note angularity of clasts and the calcite rims that encompass many of the clasts. Penny for scale.

are represented, the most noteworthy being intrasparites in which the intraclasts are coated with hematite. Minor lithologies that occur as clasts include altered basaltic glass, black mudstones and porphyritic mafic rocks. As much as 20% of the clasts in many zones of the breccia are altered beyond recognition to a mixture of clay and green alteration products.

The breccia matrix in thin section is composed for the most part of clear, white sparry calcite. In addition to the calcite, chlorite, hematite, serpentine, chalcedony and a very minor amount of clay is also present. The angularity of the fragments, the abundance of clasts of adjacent lithologies and the predominance of calcite in the matrix are all suggestive of a tectonic origin for this breccia. Presumably when the limestones and basalt of the Duzel Rock Group were faulted into their present positions, the basalt breccia was formed by shearing along the contacts of the different lithologies, incorporating fragments of each, as well as exotic rocks that have since been eroded away or are not presently exposed in the immediate area.

Contacts

The basalt breccia serves as the contact between the pillow basalt member and the platy limestone above and below it. The breccia along its margin is highly sheared, as are zones within the

breccia itself. The lowermost (southernmost) exposure of the breccia is the most highly sheared; isolated blocks of the breccia and other lithologies are scattered about below the lowermost mapped outcrops.

Age

The basalt breccia was formed during the major tectonic event that faulted the Duzel Rock Group and juxtaposed slivers one on top of the other. Most, if not all, of the calcite in the matrix could have been derived from the limestones above and below it. This tectonic event may have been initiated in the Jurassic (see Historical Geology below); hence these breccias may be as young as Jurassic. Certainly the extensive alteration visible in most of the clasts suggests the breccia was not formed by the later tectonic activity that has affected this area, but rather that a Mesozoic orogeny created the breccia and it has since been exposed to a variety of chemical and physical weathering processes.

White's Gulch Limestone

Distribution and Type Locality

The White's Gulch limestone occurs in two discrete thrust blocks within the map area (Plate 1). The southern block occupies much of the eastern half of section 2, comprising the western half

of the hill to the south of Duzel Rock. The best exposures of this block are in its center; the northern and southern ends are obscured by dense trees and manzanita bushes. The northern block occupies the area around the junction point of the four sections of the map area, where it's extensively silicified, and extends north from there to occupy much of the southeast 1/4 of section 35. The type section is designated as an east-west line through the southern block where the stratigraphy is best exposed (Line F-F' on Plate 1).

Lithology

The White's Gulch limestone, in particular the southern block, is the only unit in the map area where a stratigraphic sequence is preserved. There are 1,159 feet of relatively undeformed section exposed--the beds strike roughly North-South and dip into the hill (towards the east) at angles varying from 40 to 72 degrees. The unit can be subdivided into three members: a sequence 638 feet thick of platy micrites and massive silty micrites and oosparites, a series of bedded green cherts 163 feet thick, and a sequence 358 feet thick of massive recrystallized micrites and oomicrites with extensive silicification along certain beds (Plate 3).

The lowest member includes units DS1 through DS10, and consists of an alternation of argillaceous, platy micrites and massive oosparites. In the lowermost part of the section, basaltic flows are

intercalated with platy micrites; these are designated as DS2 in the type section. These aphanitic basalts are commonly olive gray (5Y 4/1) on a weathered surface and medium gray (N5) on a fresh surface. These basalts occur in lenticular pods, are highly fractured and have undergone severe chemical alteration. Thin section examination reveals these volcanics were originally comprised of plagioclase and orthopyroxene crystals with a diabasic texture; there is a high proportion of subhedral magnetite and/or ilmenite crystals. Chemical weathering has altered much of the feldspars to sericite, clay minerals, and calcite, almost all of the pyroxene to chlorophar-ite, and all of the oxides have been altered to a white opaque substance, possibly leucoxene. The micrites in the lowest member (units DS1, 3, 4, 6, 8 and 10) are a dark to medium light gray (N3-N6) color on a fresh surface and break up into plates two to ten inches thick. They contain insoluble residues ranging from 8.12% to 38.9%--these insoluble residues are for the most part detrital clay minerals except for the residue from unit DS10, which is mostly finely disseminated silica, presumably originally derived from the mobilization and reprecipitation of the overlying chert. Thin section examination reveals these rocks are comprised entirely of recrystallized micrite and clay minerals, the clay occurring both as discontinuous, disseminated splotches and as concentrations along laminae and microstylolites. There are a few circular allochems that have been

dissolved and infilled with sparry calcite, and there are numerous microfractures filled with sparry calcite, most of which predate the microstylolites and microfaults. The numerous microfaults are a manifestation in thin section of the numerous large and small scale faults that can be seen to transect the bedding in outcrop; these faults are most numerous in the lowermost part of the sequence (units DS1-3). The micrite has undergone total neomorphism to microspar and pseudospar. The grain size of this neomorphic calcite varies through the thin section, and is seen to be related to the clay content of individual laminae--the clay-rich laminae are bordered by calcite in the microspar range while the clay-free layers are comprised of calcite grains that are coarse pseudospar. There is a minor amount of chalcedony in the lower micrites, most of which is along stylolites as cavity fillings, and in unit DS8 a few grains of authigenic pyrite were discernible, indicative of a reducing environment. The massive limestones in the lowest member (units DS5, 7, and 9) range in color from a light to a medium gray (N7-5). The lower massive limestones (unit DS5 and interlayers in DS4) are silty and sandy micrites with a few ooids and intraclasts that stand out in relief due to selective silicification. Much of the "silt" in these micrites is actually finely disseminated silica. The massive limestones higher in this member (DS7 and 9) are oosparites that contain a high proportion (up to 40%) of concentric grains, a few intraclasts, and, in some

beds, a considerable amount of terrigeneous detritus. Some of the concentric grains may be identified as ooids, but there are several that are highly asymmetrical; these are suggestive of algal encrustations. Two thin sections (one from DS7, one from DS9) were sent to Dr. Donald Toomey of the University of Texas of the Permian Basin in Odessa for identification. Because of recrystallization Dr. Toomey couldn't make a positive identification--"They may be recrystallized algal nodules--but this is a gut opinion! I really don't know what they are." (Dr. Donald Toomey, per. comm., 1975.) There are a few intraclasts of micritic limestone and algal(?) grains cemented together by sparry calcite and surrounded by concentric laminae of calcite. The highest concentrations of terrigenous detritus occur in the lower parts of unit DS7. Most of this detritus is silt and sand sized fragments of basaltic volcanic rocks that have been extensively altered and leached. There are a few quartz grains, and a few fragments of chlorite schists. These sand grains, which in subunit DS7A make up approximately 50% of the rock, are poorly sorted and angular, indicating rapid influx of sediments from a source area of considerable relief. This unit is in many ways similar to the Middle Eocene Rickreall Limestone, of the Oregon Coast Range (Boggs and others, 1973). There is a minor amount of clay in these sparites in addition to frequently occurring cavity fillings of iron oxides and chalcedony. Some subunits of the massive limestones have a porosity as high as 2%.

The insoluble residues are generally low (3. 17-3. 93%), except for subunit DS7A which has an insoluble residue of 32. 37%, comprised chiefly of silica and basaltic rock fragments.

The second member in this stratigraphic sequence includes units DS11-13 and is 163 feet thick. It consists primarily of bedded cherts that have a wide range of colors, the most prevalent ones being grayish blue green (5BG 5/2), dusky yellow green (5GY 5/2), dusky blue (5BP 3/2), pale blue (5B 6/2) and yellowish gray (5Y 8/1). Beds of chert stand out conspicuously above the less resistant limestones; some beds are as much as 12 feet higher than the surrounding hillside (Figure 14). These chert units are extensively fractured, in sets both parallel to and perpendicular to the bedding, and it is along the chert beds that the offset along the many faults that transect the entire stratigraphic sequence is most readily measured. Many of the smaller fractures have been filled with clear white sparry calcite, while many are unfilled, creating a considerable amount of linear porosity in these cherts. Upon thin section examination, the bedding in the chert can be attributed to a variation in the grain size of the chalcedony; laminae comprised of chalcedony of sub-microscopic grain size alternate with laminae of chalcedony in which the crystal size is between . 005 and . 05 mm. and roughly equant in shape. There are a few discontinuous laminae that are made up of a much coarser chalcedony of a radiating fibrous nature; in some instances

Figure 14. Resistant beds of green chert of White's Gulch Limestone. View along strike (looking SSE). NW 1/4 SE 1/4 SE 1/4 section 2.

clusters of fibers radiate out from an aggregate of granular quartz up to 2.5 mm. long. There were no traces of radiolaria, sponge spicules, or diatoms discernible in thin section. There are scattered cubes of pyrite and rhombs of dolomite and siderite in the submicroscopic chalcedony--all of the pyrite and most of the carbonate rhombs are very small (less than 0.05 mm. across) but a few are as large as 0.2 mm. along the edge of the rhomb. A thin bed (2 feet thick) of light bluish gray (5B 7/1) micrite (DS12) occurs between the two chert units DS11 and DS13. This limestone bed shows no evidence of silicification, but the micrite is extensively recrystallized to a coarse pseudospar. Chert unit DS13 differs from DS11 in that thin (0.5-4.0 mm.) laminae of carbonate are frequent; these limestone laminae are often discontinuous along bedding.

The uppermost member (units DS14-23) is a 358 feet thick sequence of micrites and oomicrites. Recrystallization in these micrites is extensive, and many members have been completely recrystallized to a coarse neomorphic calcite. The predominant colors of these limestones are medium light to very light gray (N6-8) but some members have a yellowish gray (5Y 8/1), pinkish gray (5YR 8/1) or a light olive gray (5Y 6/1) color. Four units (DS 14, 15, 20 and 23) are extensively silicified. Silicified beds are as much as 4 cm. thick and may extend one or two meters before merging, along the bedding plane, with limestone (Figure 15). Many of these

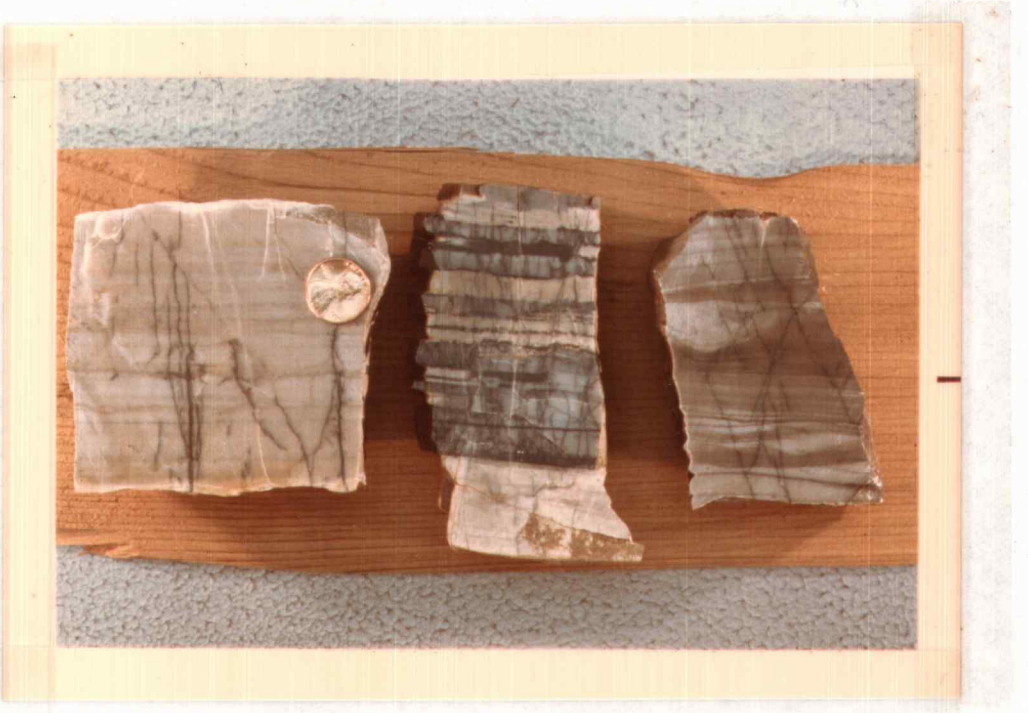

Figure 15. Micrites, some selectively silicified, of the upper member of the White's Gulch Limestone. Samples derived from units DS 17, DS 20, and DS 21, from left to right. Penny for scale.

chert beds bifurcate, while others thin and thicken along bedding. Most are too long and have overly straight contacts to be described as nodular. Many chert beds are connected with other ones above or below them by chert "bridges" that transect micritic layers. The silicified beds stand out considerably in relief on a weathered surface, creating a blocky appearance. The chert beds are medium light gray (N6) or a grayish blue (5PB 5/2) color and usually stand out in sharp contrast from the limestone layers. In outcrop, the micrites in the uppermost member have either an even, "massive" texture or are finely laminated. One unit, DS21, has swirled laminations that resemble load casts. The oolitic micrites have a nobbly weathering on exposed surfaces, due to recrystallization and/or silicification of ooids. The entire member is transected by numerous microfaults and fractures; some fractures are filled with white sparry calcite, but in several members all the fractures are filled with black metallic oxides. Identification of original constituents via thin section examination proved difficult because of the extensive recrystallization that has taken place. In general, recrystallization has been minimal in the lower units (DS14 and DS15) and increases towards the top of this member. The least recrystallized limestones are composed of mostly intact micrite with a few laminae which have undergone neomorphic recrystallization to microspar. The remainder of the micrites have been extensively recrystallized--all or most of the

micrite has been recrystallized to microspar and pseudospar and traces of any original allochems have been destroyed. Exceptions to this are the oolitic micrites (DS16 and DS18). These may contain as much as 20% allochems, all of which have been dissolved and filled in with clean sparry calcite or chalcedony. A few have not been infilled, creating an oomoldic porosity of 1-2% in these limestones. The vast majority of these allochems are spherical and were probably ooids; a few retain the outer laminae. A few of the allochems that have been dissolved and refilled have angular outlines; these were probably intraclasts of limestone. The chert in the silicified laminae and lenses is composed of fine-grained chalcedony with an even texture. The contact between the chert and micrite is usually sharp, although occasionally stringers of carbonate extend from the micrite into the chalcedony. Diagenic rhombs of iron-rich dolomite are quite common in both the intact (not recrystallized) micrites and all of the silicified layers. Clay content of all the micrites in the uppermost member is consistently low; in thin section clays are seen to concentrate along the stylolite-like boundaries between micrite and chert, along microfractures or as large, irregular blebs in chert lenses. Insoluble residues from this member range from 1.44-1.72%; these are mostly detrital clay, except from the ooolitic micrites, whose insoluble residues are comprised for the most part of silicified ooids. In the lowermost unit of this member, DS14, there are small lenses

of green sandstone that are discontinuously interbedded with the micrite. These sandstone lenses are no more than 20 cm. thick and occur near the base of unit DS14. These sandstones are moderately sorted and composed of subangular very fine sand and silt. The sand is held together by a matrix of neomorphic calcite and organic material, with sparry calcite filling some larger spaces. The most common framework constituent is celadonite, a common alteration of volcanic glass. In a few celadonite grains vesicles are still discernible; in others crystalites forming from the devitrification of the original glass are visible. Other framework constituents, in order of their abundance (most abundant first), are basaltic rock fragments, detrital calcite (intraclasts), quartz, chlorite, chert and feldspar. A few cavities have been filled with chalcedony, even fewer with a zeolite mineral, and hematite as a cement is quite common in portions of the thin section. The organic-rich calcite matrix has a reddish brown color to it, and comprises as much as 45% of the rock. From their mineralogy and texture, it can be supposed that these sands were derived from a predominantly basic volcanic source terrain, and that they were not subjected to any extensive abrasion or chemical weathering process of long duration. Their fine size does not preclude aeolian transport to account for their deposition in this relatively clastic-free basin.

The most northerly thrust block that contains the White's Gulch

Limestone has experienced much more internal fracturing than the more southerly block in which the type section is located (Plate 1). The stratigraphic sequence is not nearly as well preserved in the more northerly block, and extensive tree cover severely limited the area of exposure. In addition, large areas of the more northerly block are covered with a Cenozoic limestone breccia, making stratigraphic correlation all the more difficult. Despite these difficulties, the limestone in this block can be included in the White's Gulch Limestone for several reasons. The single most convincing piece of evidence is the occurrence of bedded green chert near the base of the more northerly fault block. This chert is virtually identical to member DS11 in the type section, and is as much as 50 feet thick. The chert does not occur as a continuous bed, but is faulted apart so that segments of the same chert bed are separated by several feet, or tens of feet. Further evidence is obtained from the petrography of the limestones above the chert. These resemble very closely the micrites and oomicrites that lie above the bedded chert in the type section, units DS14-23. One thin section (DQ2) is comprised of micrite that has undergone considerable recrystallization to microspar and pseudospar; it differs from the micrites in the type section in that it has a higher clay content (the insoluble residue obtained from this sample was 4.65%, almost all of which was clay). The other thin section observed was an oomicrite similar to unit DS16--all

of the allochems were either dissolved out and infilled with sparry calcite or were internally recrystallized to neomorphic pseudospar. As in DS16, the ooids were the predominant allochem, but there was a slightly higher proportion of intraclasts.

Over 50% of the northern block of White's Gulch Limestone has been completely silicified to a hard, white chert. Similar silicified limestone has been reported by Porter (1974, p. 33) from the Massive Limestone Member of the Facey Rock Limestone. On a sawed surface, some samples of this white chert have a brecciated appearance. In thin section, the outlines of the reddish brown fragments are reminiscent of intraclasts and coated grains found in the White's Gulch Limestone. These allochems are composed of much coarser quartz grains than the surrounding matrix. It is probable that this portion of the White's Gulch Limestone being "sandwiched" between two thrust faults, was highly fractured, and hence susceptible to permeation by low temperature, silica-rich solutions. There are a few patches in this silicified zone where the limestone has not been replaced; these may be isolated cores of limestone that escaped the extreme brecciation the surrounding areas experienced.

Age

No fossils have been found in the White's Gulch Limestone. As no similar stratigraphic section has been described from the Klamath

Mountains, no age can be assigned on the basis of lithologic correlation. The only line of reasoning, and a tenuous one at that, to follow in order to obtain an estimate of the age is to note the pattern in other thrust blocks of limestone in the vicinity. Typically, as at Facey Rock and presumably with the Duzel Rock Group, Ordovician limestones have been thrust over the extensive Moffett Creek Formation, which has been dated as Silurian by Al Potter (1976, pers. comm.). As the White's Gulch Limestone has similarly been thrust on top of the Moffett Creek Formation, it is certainly plausible that it is older than Silurian; perhaps it is Ordovician in age, perhaps even older. It is hoped that further examination of the samples collected by Drs. Bergström and Laufeld will in the future provide substantive evidence as to the age of these limestones, as detailed knowledge of this sequence would shed considerable light on the unraveling of the history of the early depositional and tectonic environments in the Klamath Mountains Province.

Structural Relationships

The northern block of White's Gulch Limestone is completely encompassed by thrust faults. The western contact with the Moffett Creek Formation is a severely faulted and sheared zone as is the eastern contact with the platy limestone. While these two contacts can often be determined to an accuracy of a few meters, the northern

and southern contacts are covered by thick rubble and soil, as well as considerable vegetation, and can only be approximately determined. This northern block has experienced extensive internal fracturing and brecciation in addition to the thrusting at its margins, as evidenced by the extreme silicification the block has undergone.

The southern block is bordered on three sides (north, west, and south) by a thrust fault; a small lens of breccia and the extensive fracturing of the western edge of the block are evidence of the thrusting. The eastern margin is a nearly vertical fault which separates the White's Gulch Limestone from an extremely fractured red shale and the massive limestone member of the Spring Branch Group. It is the author's opinion that the Spring Branch Group and White's Gulch Limestone comprise separate slivers that have been thrus into their present positions, the red shale being squeezed between them and becoming extremely fractured as a result (see Plate 2). It is also plausible that the two groups were juxtaposed by faulting prior to the main episodes of thrusting and were thrust as one coherent block.

Thickness

The preserved thickness of the White's Gulch Limestone is, as measured in the type section, 1,159 feet. The northern block of this lithology is extensively faulted, making correlation of individual units

with those described from the type section in the southern block difficult. It is likely that since most of the northern block is probably correlative with the uppermost member of the type section, that several hundred feet could be added to the thickness measured, but this is not certain, since in most of the northern block the attitudes of strata are extremely disturbed.

Environmental Interpretation

The bedded cherts of units DS11 and DS13 are believed to be primary siliceous deposits, despite the absence of remains of any siliceous organisms. The extremely low carbonate content of the cherts and the abruptness of the contacts with the adjacent bedded micrites are suggestive of a primary origin. The association of bedded cherts and limestones is relatively rare, as bedded cherts are commonly associated with shales in non-carbonate sequences (Blatt, Middleton, and Murray, 1972, p. 536). Drilling in the Western Pacific reveals that in Cretaceous and Tertiary deposits in this region all cherts associated with carbonate sequences are nodular, while bedded cherts are restricted to non-carbonate deposits (Heath and Moberly, 1971).

The deposition of siliceous sediments requires no special explanation involving extreme volcanic input or concentration by silica-concentrating organisms. According to Ketner (1969),

> The large volume of dissolved silica entering the sea (from streams and rivers) compels constant precipitation of silica. This can be accomplished by organic precipitation of opal and by inorganic precipitation of quartz. . .

The deposition of purely siliceous deposits is dependent on two conditions: abundant dissolved silica and sparse detritus. The problem one faces in explaining the occurrence of nearly pure chert in the sequence of the White's Gulch Limestone is accounting for the cessation of carbonate deposition during the formation of the chert, when carbonates and silica were simultaneously deposited both before and after the deposition of the chert.

I am of the opinion that the White's Gulch Limestone was laid down in a relatively shallow inter-arc basin as envisioned in Churkin and McKee's (1974) reconstruction of the early Paleozoic in the Klamath Mountains. The basalt flows near the base of the sequence represent vulcanogenic crust from the island arc system. This basin would have been considerably restricted during the deposition of units DS1 through DS6--the water must have been relatively calm to allow for the deposition of micrites and detrital clays that compose these platy limestones. The restriction of the basin is best explained by tectonic silling--submarine block faulting is often responsible for the creation of basins that are cut off from sediments supplies that may exist in the vicinity. During the deposition of units DS7 through DS10, the basin was periodically subjected to higher energy currents.

Gaps may have been eroded in the fault-block barriers to allow for the influx of normal oceanic currents. Ooid formation and the lack of detrital clay in massive limestones DS7 and DS9 are indicators of fairly constant shallow marine current activity. Occasionally the basin received angular terrigeneous detritus from an emergent basic volcanic source in the vicinity--between 40 and 50 percent of subunit DS7A is composed of mafic volcanic rock fragments. The cross-beds in unit DS10 are also indicative of at least periodic, high-energy conditions.

During the deposition of the cherts in units DS11 and DS13, the basin must have become severely restricted again, perhaps by renewed block faulting of the island-arc and oceanic crust. Detrital input was extremely low, as there is virtually no clay or volcanic rock fragments discernible in thin sections of the chert. While the bottom waters may have periodically stagnated, the basin must have been replenished by circulating marine water, ensuring a continuous supply of dissolved silica. The question that persists is why carbonates were not deposited simultaneously with the chert, since limestones are so prevalent both before and after. It is conceivable that water temperatures at the bottom of the basin dropped sufficiently during the deposition of the chert to maintain all of the calcium carbonate in solution. Changes in the pH of the bottom waters may have excluded carbonate deposition; overly acidic waters would

effectively dissolve all carbonate while allowing for the deposition of siliceous sediments (Mason, 1966, Chapter 6).

After the deposition of the chert, changes in the conditions of the bottom waters in the basin occurred to allow for another cycle of carbonate sedimentation. Perhaps renewed current activity effected a more thorough circulation of the basin's water. In any event, calcium carbonate and silica were deposited simultaneously during the episode represented by units DS14 through DS23; finely disseminated silica later migrated in solution and precipitated during diagenesis to form the discontinuous layers of chert that are so conspicuous in units DS14, 15, 20, and 23. The angular detritus, rich in altered glass shards, that forms the small lenses of sandstone in unit DS14 may be aeolian sediments transported from emergent volcanoes during violent eruptions, later concentrated by currents on the basin's floor. While the extensive neomorphic recrystallization in the uppermost member of the sequence makes environmental interpretation difficult, the predominance of micrite and the general absence of terrigeneous detritus suggests that the basin was as yet considerably restricted. The occurrence of ooids in units DS16 and DS18 indicates that portions of the basin were extremely shallow; ooids formed in shallow areas may have been transported to deeper, calmer portions of the basin to form the oomicrites that occur in the uppermost member of the sequence.

Spring Branch Group

The name Spring Branch Group is applied to the limestones, basalts, and chert that occupy the eastern half of the hill south of Duxel Rock. The limestones, basalts and cherts comprise a thrust block that abuts the White's Gulch Limestone to the west and lies on top of the Moffett Creek Formation. The entire group lies in the easternmost 1/8 of section 2, except for a small part of the massive limestone member of the group, which extends into the SW 1/4 NW 1/4 of section 1.

Massive Limestone Member

Distribution and Type Locality

There are two blocks of the massive limestone member of the Spring Branch Group, the smaller of which lies in fault contact with the White's Gulch Limestone, in the southern half of the NE 1/4 SE 1/4 of section 2. The larger block is bordered on two sides by platy limestone and basalt of the Spring Branch Group and on the other two sides by a thrust-fault contact with the Moffett Creek Formation. This larger block occupies most of the eastern half of the SE 1/4 NE 1/4 of section 2, and extends into the NE 1/4 NE 1/4 SE 1/4 of section 2 and into the SW 1/4 NW 1/4 of section 1. The type locality of the massive limestone member is designated as the outcrop directly

overlying the platy limestone member in the NE 1/4 NE 1/4 SE 1/4 of section 2.

Lithology

The color of the massive limestone member is very variable. The most common colors on fresh surfaces are light brownish gray (5YR 6/1) and pale red purple (5RP 6/2). Much of the massive limestone member is extensively brecciated and colors on fresh surfaces of brecciated samples are generally very light gray to medium light gray (N8-6). Weathered surfaces range from very light gray (N8) to yellowish gray (5Y 8/1) to grayish orange (10YR 7/4); the yellow and orange hues are the result of iron oxide stains. Slopes developed on the massive limestone are uneven and often steep; large outcropping blocks up to 50 feet long, 20 feet wide and 30 feet high stick out from the surrounding hillside and have nearly vertical sides.

Much of the massive limestone member is extensively recrystallized. No allochems are visible in the majority of hand samples; the limestone has a fine-grained granular texture similar to that of a marble. One of the more noticeable features in hand sample is the high porosity of the rock. There are as many as 15 macroscopic cavities per square centimeter on a sawed surface; the vugs range from microscopic size to 7 mm. in diameter. Most of the vugs are

nearly spherical and are lined with drusy calcite stained yellow by limonite; several cavities are completely filled with the same material. In addition to the considerable porosity, the massive limestone member is laced with numerous intersecting sets of fractures. The fractures never exceed 0.5 mm. in thickness, and most of them are much finer than that. There has been some small scale displacement along some of these fractures as some of the fractures and vugs are offset as much as 2 mm.

In several localities the massive limestone member has been extremely brecciated; most of these localities are in close proximity to the lower thrust contact of the entire group. The clasts range from microscopic size to 3 cm. along the longest axis; most are subequant but several are triangular or blade-shaped. For the most part the clasts are subangular to angular, although a few have somewhat rounded corners. The matrix is made up of pulverized limestone and sparry calcite, in places stained yellow by iron oxides. The brecciated localities are not characterized by the large and numerous pores that are prevalent in the bulk of this member, but the fractures are generally wider (up to 0.75 mm. wide) and usually lined with calcite and yellow iron oxides. It is in the clasts within the brecciated samples that the least amount of recrystallization has taken place and the allochemical constituents are most easily discernible. Thin section examination reveals that considerable recrystallization has

taken place within the clasts, but allochems are still identifiable. The overwhelming majority of the allochems are coated grains, perhaps originally ooids. These range from 1.5 mm. to 0.01 mm. in diameter. Most of these coated grains are spherical in shape but some are elongated and a few appear to be composite, that is, composed of two or more coated grains that have been incorporated in one mass by a surrounding coating. All of the coated grains have experienced some degree of internal solution, and most of them have had all but the outer few layers dissolved away. The majority of these have been infilled with sparry calcite, but a few remain as cavities, lending a microscopic moldic porosity to the rock. Many of the coated grains are characterized by a dark, somewhat diffuse micrite rim around their perimeters. The dark color is probably due to organic matter; these micrite rims resemble those formed by the boring of photosynthetic algae (Bathurst, 1966).

Some authors feel, notably Swinchatt (1969), that most of the micrite rims observed in allochemical grains are the products of boring by photosynthetic algae, and hence micrite envelopes can be used as an indication of formation at depths shallower than 200 m. If the micrite rims observed in the vast majority of the coated grains in the massive limestone member were in fact produced by the boring of photosynthetic algae, this would provide an environmental parameter for the shallow water (i.e., photic zone) depositional origin of

this limestone. This assumption may not be valid, however, since there are other boring organisms capable of producing such rims around the allochemical and biotic constituents in a carbonate sediment. Friedman, Gebelein, and Sanders (1971) point out that fungi, bacteria and heterotrophic algae are known to be active borers. Friedman has observed micrite envelopes on carbonate grains from depths in considerable excess of 200 m., and the authors state,

> At present we do not know how to distinguish the results of boring activity of photosynthetic algae from those of algae living heterotrophically.

At best, the high incidence of micrite rims in this member may sugtest a considerable contribution from the activity of a concentrated population of photosynthetic algae, and hence this limestone may have formed in relatively shallow water, but it is also conceivable that all of the boring was accomplished by heterotrophic algae and/or fungi living at greater depths.

In addition to the abundance of coated grains, there are a few intraclasts of a micritic limestone with an abundance of opaque, microscopic spots. Use of reflected light shows this mottling to be due to abundant concentrations of organic matter in these micrites. These intraclasts amount to less than 1% of the allochems preserved in the brecciated samples of this limestone. The matrix between the allochems within the clasts of the brecciated limestone was originally of a micritic nature is its dark, mottled color in contrast to the

white sparry calcite that occurs as pore fillings and the irregular boundaries of the pseudospar grains. The matrix between clasts of the brecciated portions of the massive limestone member is primarily composed of finely granulated limestone, much of which was sparry calcite prior to brecciation. The grain sizes of the crystals in this granulated limestone range from 0. 25 mm. to extremely small crystals which are not discernible with the highest magnification (320X) on the petrographic microscope used in this study. Some of this matrix is stained by finely disseminated black organic matter or a mixture of hematite and limonite that have apparently seeped along grain boundaries within the granulated matrix. A considerable amount of recrystallization and diagenic replacement has taken place both in the clasts and in the matrix of these brecciated portions of the massive limestone. Much of the rock has been replaced by coarse, clear sparry calcite. In thin section, patches of sparry calcite can be seen to have replaced large portions of granulated matrix and allochem-rich clasts, so as to obliterate the boundary between the two. Sparry calcite also occurs along irregular veinlets, as do a few stringers of amorphous silica.

In naming the massive limestone member of the Spring Branch Group, I am assuming that the clasts in the brecciated samples is representative of the entire member, since it is only in these samples that preserved depositional textures are discernible. By Dunham's

(1962) classification, this limestone would be classed as a wackestone. By Folk's (1959) classification, this member would be an oomicrite.

The insoluble residue of the massive limestone member is a very small percentage of the rock; two samples (DQ18 and DZ44) yielded percentages of 0.42% and 0.47%. Approximately 65% of this is medium light gray clay. The remainder of the residue is composed of irregular clumps and blades of silica, quartz crystals and quartz silt. Some of the clumps of silica are composed of silicified ooids cemented together; the largest of these clumps was a rough cylinder 5 mm. long and 2 mm. in diameter. The few quartz crystals that were present in DZ44 were prisms up to 1.5 mm. long and 0.5 mm. perpendicular to the c axis; they were very angular and had fresh crystal faces. Usually quartz crystals of this nature would imply a silicic volcanic or a plutonic origin, but it is hard to envision how they would be transported by tractive currents out to a basin where very little clay was settling without suffering considerably more rounding due to abrasion than they exhibit. Perhaps these crystals are volcanic in origin, and settled out after becoming air-borne due to a violent eruption, but the absence of any glass shards or fragments of volcanic flow rocks is not supportive of an extrusive origin. It is conceivable that these small, euhedral quartz prisms are of an authigenic origin, precipitated out of silica-rich solutions in a cavity

in the limestone.

Age

No fossils were found in the massive limestone member of the Spring Branch Group. As with the White's Gulch Limestone, there is virtually no evidence that would indicate a particular age for this unit. Again, the tenuous assumption can be made that since most of the dateable limestone blocks that have been thrust on top of the Silurian Moffett Creek Formation are Ordovician in age, the massive limestone member of the Spring Branch Group may also be Ordovician, perhaps older.

Structural Relationships

The larger of the massive limestone blocks is bordered on the west and southwest by the platy limestone and basalt members of the same group; these borders have all the appearances of depositional contacts, despite the virtual lack of recognizable bedding in the massive limestone. The smaller block is bordered on the west by a near vertical fault contact with the White's Gulch limestone (Plate 1). This contact is clearly visible in only two small localities; along most of its length it is obscured by rubble and trees. The high degree of brecciation of both lithologies in the vicinity of the contact is indicative of its fault nature. The eastern contact of both blocks of massive

limestone is a thrust contact with the underlying Moffett Creek Formation; it is in close proximity to this contact that the brecciated samples described above were located.

Thickness

The high degree of recrystallization and brecciation in the massive limestone member has obliterated most of any original bedding that may have existed. The few localities where bedding features can be discerned indicate that there has been considerable tectonic disturbance. Hence, the observed thickness, assuming that the lower contact of the member is a bedding plane, is in all likelihood an inaccurate figure. By a rough estimate, the exposed thickness comprises a minimum of 200 feet and a maximum of 350 feet.

Environmental Interpretation

The pervasive, total recrystallization that has affected the massive limestone member of the Spring Branch group has obliterated almost all paleoenvironmental indicators. Only the brecciated samples collected near the eastern thrust margin provide some clues as to the nature of the original depositional setting of these limestones. The predominance of micrite rims around the perimeters of most allochems may or may not be indicative of shallow water depths, as discussed previously. The predominance of round coated

grains, highly suggestive of ooids, however, is a strong indicator for an initial shallow depth of formation under the influence of waves or tidal currents for at least the allochemical constituents.

As with the Massive Limestone Member of the Duzel Rock Group, this massive limestone is characterized by a textural inversion; abundant coated grains, presumably formed in a turbulent, shallow environment, are supported by an originally micritic matrix, which requires fairly calm water to settle to the sea floor. Folk (1962) states that such a textural inversion is the result of the mixing of sediments from different depositional environments; such mixing would commonly occur at the transition zone between fluctuating high and low energy environments. The various mechanisms whereby such mixing can take place were reviewed in the paleoenvironmental interpretation of the massive limestone of the Duzel Rock Group; to avoid repetition the reader is referred to that section.

As with the massive limestone atop Duzel Rock, the author favors a deeper marine environment, offshore from a carbonate bank, as the interpretation of the origin of the massive limestone member of the Spring Branch Group. The black chert and fine-grained basalt that underly the entire group are consistent with a deep marine setting. The mottled, micritic platy limestones that directly underly the massive limestone member similarly invoke a bathyal setting. In this interpretation, ooids formed at the edge of a shallow carbonate

bank, similar to the Bahamas Bank today (Bathurst, 1975), would periodically be carried out to deeper waters through channels in the bank by offshore currents. Once out in the deep waters off the bank, they would settle out along with the micrite that was continuously, or periodically, deposited in the relatively quiet water. The few intraclasts that occur in the massive limestone are very similar to the platy limestone that underlies the massive oomicrite. They probably represent rip-ups of semilithified micrite, perhaps torn up by the same tractive currents that carried ooids from the edge of a shallow bank out into the deep sea.

Platy Limestone Member

Distribution and Type Locality

The Platy Limestone Member of the Spring Branch Group occurs in two distinct slivers around the base of the southern and southwestern exposures of the massive limestone member. The smaller of the two outcrops lies in the SE 1/4 SE 1/4 NE 1/4 of section 2 while the larger is in the NE 1/4 NE 1/4 SE 1/4 of section 2. No exposures of this lithology was found along the eastern, northern, or northwestern contact of the massive limestone member. The type locality is designated as the exposures northwest of the road in the larger of the two slivers, in the NE 1/4 NE 1/4 SE 1/4 of section 2.

Lithology

The platy limestone member contains a pale yellowish brown (10YR 6/2), dark yellowish orange (10YR 6/6) or light olive gray (5Y 6/1) color on a weathered surface. On a fresh surface, the most common colors range between medium light gray (N6) and light olive gray (5Y 6/1). The limestone has a flaggy appearance, breaking up into plates between 3 and 7 cm. thick, the average thickness centering around 4. 5 cm.

The surfaces of these plates are crossed by several sets of fractures. Some of these are very fine but the most numerous and most conspicuous are between 2 and 3 mm. thick and occur in a number of intersecting sets that run perpendicular to the bedding. All of these thicker fractures are filled with coarse sparry calcite, which in a few cases is white or pale orange, but in most instances is stained black.

The platy limestone member of the Spring Branch Group is not finely laminated as is the platy limestone member of the Duzel Rock Group, but rather has a "massive" appearance on a surface sawed through an individual plate. The only indication that the bedding does run parallel to the direction of fracture between plates of this limestone is the presence of rare, poorly defined, discontinuous clay concentrations that have weathered to a pale orange and are approximately 0. 75 mm. thick. This limestone is extremely fine-grained

throughout the exposure; no coarse-grained or clastic-rich layers or lenses were found.

The beds of platy limestone are considerably disturbed. While they dip below the massive limestone member (to the east and north) for the most part, substantial differences in the strike of the beds over short distances are quite common. In contrast with the platy limestone member of the Duzel Rock Group, no highly flexed or folded beds were observed in the platy limestone member of the Spring Branch Group due to faulting.

Features observed in thin section examination of the platy limestone member include finely disseminated organic matter and clay, detrital grains of hematite, angular pore spaces, and calcite veins. The limestone has a mottled appearance in thin section; large portions of the slide are stained yellowish brown, reddish brown and dark red by organic matter and iron oxides. Yellow, brown and gray patches of clay are also scattered throughout the slide, and characteristically have diffuse boundaries. There are virtually no concentrations of clay along laminations or stylolites as is common in other platy limestones in the area of study. Silt size grains of hematite are common throughout the thin section of this limestone; most of them are sub-angular to angular, but a few are subrounded. A few of these hematite grains have round, bubble-like pores, a few have tiny inclusions that appear to be lath-like crystals, suggesting that the

hematite grains are the totally weathered products of volcanic glass or scoria and basaltic fragments. There are many dispersed, unconnected, silt-size pore spaces, and almost all of these are highly angular. Some of these pore spaces are filled with amorphous silica, and in a few the silica has filled in around the outline of a circular bubble with tiny crystallites. All this suggests that these pore spaces were originally glass shards, and most of them have weathered and been leached out, while a few devitrified or were replaced by amorphous silica.

There are many sets of calcite veins visible in thin section; the largest veins are 4 mm. wide. They are all filled with coarse white sparry calcite, in marked contrast with the mottled appearance of the rest of the limestone. Most of the veins are offset by microfaults (most of which run parallel to the bedding) and in a few the coarse calcite veins show evidence of severe shearing. The micrite that makes up 98% of this limestone has been recrystallized to a coarse pseudospar. The recrystallization was an aggrading, coalescive neomorphism (Folk, 1965). The evidence that the pseudospar was recrystallized from an original micrite includes the irregular boundaries of the grains and the considerable amount of tiny, included clay specks and bits of organic matter that are visible under high magnification.

One sample of this limestone was run for insoluble residue; a

weight percent of 1.49 was derived. This is surprisingly small in view of the widely dispersed organic matter and hematite visible in the slide; most of this must be an extremely thin coating widely dispersed throughout the rock. The insoluble residue was for the most part a medium gray clay, with some small grains of hematite and irregular silt-size pieces of silica present.

In Dunham's (1962) classification, this limestone would be a mudstone. It would classify as a micrite or a dismicrite in Folk's (1959) classification.

Age

As with all the other limestones in the area of this study, the platy limestone member has produced no fossils. These limestones are dissimilar enough to the Platy Limestone Member of the Facey Rock Limestone of known Ordovician age (Porter, 1974) so that the two cannot be correlated. As with the massive limestone member of the same group, a tentative pre-Silurian age is assigned to this unit.

Structural Relationships

Both the upper contact, with the massive limestone member, and the lower contact, with the basalt member, of the platy limestone appear to have originally been depositional contacts, but a great deal of slipping and faulting has occurred along these boundaries. The

degree of dislocation of the beds of platy limestone, particularly near the lower contact, indicate that a substantial amount of faulting has occurred both parallel and perpendicular to the bedding, and that the original depositional surfaces have served as slip surfaces. It is certainly possible that additional volumes of rocks that may have existed between the platy limestone and its bordering lithologies have since been brecciated by faulting and removed by erosion.

Thickness

The high degree of internal disruption in the platy limestone member renders a direct determination of the stratigraphic thickness by measurements perpendicular to the bedding impossible. If one accepts the distance between the upper and lower contacts as a close estimate of the amount of limestone deposited then the maximum exposed thickness is roughly 120 feet.

Environmental Interpretation

The platy limestone member has all the appearances of a deposit formed in a deep oceanic, quiet water basin that was mainly unaffected by any currents carrying terrigeneous detritus from distant land masses. There are no clay-rich laminations, no lenses of angular clastic material that would indicate periodic storms in a detritus source. The platy limestone member is characterized by an

abundance of micrite and a small amount of evenly dispersed red and brown clay. The even dispersal of the clay suggests small amounts were continuously settling to the ocean floor along with the volumes of micrite; this continuous, "light rain" of clay along with its red color suggests much of the clay may have been aeolian in origin, much like the wind-borne, red clays that reach the centers of modern ocean basins. The grains of hematite, presumably original scoria and volcanic rock fragments, and the devitrified shard-like grains are small enough so that they too were probably transported to this deep basin by the wind after a violent explosive eruption in some distant volcanic island.

While most limestones in the geologic record were deposited in relatively shallow marine water, deep-water limestones do occur and have been documented in the geologic literature. One well known and often discussed deep-water limestone is the Brushy Canyon Formation of the Permian reef complex of the Guadalupe Mountains in Texas, a deposit whose exact mode of formation has been intesnely debated (Jacka and others, 1968; Harms, 1974). While the platy limestone member was most probably deposited in deep water, and was certainly unaffected by terrigeneous detritus-bearing currents, it need not have been deposited in the middle of large oceanic basin, far removed from any land mass. A basin can be shielded from oceanic currents by tectonic silling or by the configuration of the

current patterns, as the modern Sargasso Sea is unaffected by currents that circle around it. Hence the interpretation that the platy limestone member was deposited in a deep, restricted basin is not contradictory with the later deposition of massive oomicrites which had their origin on a nearby carbonate bank. All that is required is a reorganization of current patterns or the erosion of a tectonic barrier to ooid-bearing tractive currents.

Basalt Member

Distribution and Type Locality

The Basalt Member of the Spring Branch Group underlies the massive and platy limestone members, extending from the NE 1/4 SE 1/4 NE 1/4 of section 2 through the SE 1/4 SE 1/4 NE 1/4 and into the NE 1/4 NE 1/4 SE 1/4 of section 2 (Plate 1). It is not a continuous band but rather is separated into two bodies, presumably due to movement along faults. The type locality is designated as the exposure immediately northwest of the road in the NE 1/4 NE 1/4 SE 1/4 of section 2, underlying the type locality of the platy limestone member.

Lithology

The basalt member is commonly a dark greenish gray (5GY 4/1) or brownish gray (5YR 4/1) color on a weathered surface, although

many stained surfaces also have a moderate brown (5YR 4/4) color. On a fresh surface the basalt has a medium bluish gray (5B 5/1) to dark greenish gray (5G 4/1) color. The basalt is very aphanitic and has a structureless appearance; there are no phenocrysts that are visible, even with the aid of a hand lens.

The only macroscopic features in the basalt are numerous fine fractures, all of which are less than one millimeter in thickness. The largest of these fractures are filled with an iron oxide-clay mixture that has a moderate reddish brown (10R 4/6) or dark reddish brown (10Y 3/4) color. In addition to these, several fractures are filled with clear white calcite.

In thin section, the largest crystals are opaque minerals, which make up as much as 20% of the rock. These opaques are mostly magnetite, much of which is in the form of cubic euhedra, but some of it is probably ilmenite. The rest of the original rock was composed primarily of plagioclase and probably pyroxene, but all of the mafic minerals have been altered, so that all that is left of them now is patches of a green mixture of chlorite, epidote, and serpentine minerals. The plagioclase minerals have also undergone severe alteration, and while their lath-like shapes are still discernible, no twins have been preserved. The principal alteration products derived from the feldspars are sericite and calcite; large patches of the slide have been altered to dirty calcite with indistinct borders. Numerous

fractures dissect the thin section and these are filled with a variety of minerals, including coarse calcite, chalcedony, serpentine, and a red clay-oxide mixture.

Al Potter (1976, pers. comm.) had a trace element analysis run on this basalt, and a Y/Nb ratio of 0. 98 was obtained. According to the classification of Pierce and Cann (1973), this ratio classified the basalt member as an alkali basalt, probably of oceanic origin. Its association with marine cherts and limestones indicate that this basalt may have been extruded in a medium depth, sea-mount type of setting, or possibly in a deep-water basin in the vicinity of a source of shallow-water carbonate sediments.

Contacts

The upper contact of the basalt, which separates it from the platy limestone member, may have been conformable originally, but it has experienced considerable shearing during the thrust faulting of the entire block. The platy limestone directly above the contact has been broken up into dislocated blocks that have been tilted at various attitudes. Numerous faults transect both the platy limestone and basalt both perpendicular and parallel to the contact; along the parallel faults, and along the contact itself, the rocks have been substantially sheared.

The lower contact of the basalt, which separates it from the

black chert member, is obscured along most of its length by regolith and vegetation. As has the upper contact, the lower contact of the basalt has experienced considerable shearing, and in the majority of localities where it is visible, the contact is actually a zone five to ten feet wide composed of slivers of both sheared basalt and chert. Whether the basalt ever lay conformably on top of the chert is uncertain; nowhere is a depositional contact preserved that isn't extensively sheared. At the southeastern end of the lower contact of the basalt member, slivers of red shale have been emplaced tectonically between the basalt and the black chert.

Age

No age can be assigned to the basalt member with any degree of certainty; none of the members above or below it contain any fossils. Presumably, since the entire Spring Branch Group has been thrust over the Silurian Moffett Creek Formation, all the members are older than this graywacke, perhaps Ordovician in age, perhaps older still. Should the platy limestone member yield any microfossils upon further examination of the samples collected, it would shed considerable light on the problem of dating the entire Spring Branch Group.

Black Chert Member

Distribution and Type Locality

The black chert member of the Spring Branch Group outcrops as a band running roughly north-south along the western margin of the northern block of the group (Plate 1). As the chert band curves to the southeast it becomes discontinuous, cropping out as small distinct blocks, presumably fault-bounded. The band of black chert runs from the NE 1/4 SE 1/4 NE 1/4 of section 2 south along the border of the SE 1/4 SE 1/4 NE 1/4 and the NE 1/4 NE 1/4 SE 1/4 of section 2. The type locality is designated as the widest outcrop of chert that lies in contact with the basalt member in the SW 1/4 SE 1/4 NE 1/4 of section 2.

Lithology

The black chert member has a wide range of colors on a weathered surface. Commonly, the chert is stained by a thin coating of iron oxides of clay minerals and has a pale yellowish brown (10YR 6/2), moderate yellowish brown (10YR 5/4), or light olive gray (5Y 6/1) color. Many weathered surfaces, however, are not stained by oxides and have a medium dark to dark gray (N4-3) color, and rarely the chert has a medium bluish gray (5B 5/1) tint on the surface. On a fresh, slabbed surface the predominant color is dark gray (N3), but

patches of the chert are quite light in color, the lightest being light gray (N7). All shades of gray between light and dark gray (N7-3) occur, but the darker shades predominate.

There is no discernible bedding or laminations in the black chert member. The chert is, however, highly fractured; several sets of joints transect the chert both parallel and perpendicular to the upper and lower contacts. Many of these fractures are healed by white microcrystalline quartz, but along several fractures a linear porosity has developed. Most of these open fractures are a millimeter or less wide, but along some fractures considerably larger pores have opened up. The largest of these are 2.5 cm. long and 0.5 cm. wide.

In addition to the small fractures observed in hand specimen, there are a few sets of larger fractures and small scale faults that give the chert a blocky appearance in outcrop. The visible offset along the faults, most of which are oriented perpendicular to the contacts, is not more than 0.5 m. Towards the southern end of the exposure of the black chert member faulting has apparently broken up the chert into disconnected blocks that are 20 to 100 feet long (measured parallel to the lower contact).

Thin section examination reveals that the bulk of the black chert is composed of extremely fine-grained chalcedony. In many places this cryptocrystalline material is recrystallizing chalcedony. The

chert has a brecciated appearance in thin section, with areas of contrasting grain size being juxtaposed against each other without any gradational zones. This brecciated texture is apparently due to intense faulting on a small scale. Also observed in thin section are large patches of coarse granular quartz crystals of equant dimensions. The outlines of these patches are very similar to the shape of the cavities developed along fractures visible in hand specimen, and it is probable that these patches of quartz are, in fact, cavity fillings. In addition to the filled cavities, there are several irregular and circular pore spaces that have remained unfilled.

Thin section examination reveals that the chert is almost entirely microcrystalline quartz. The only other constituents are some irregular splotches of hematite and some silt-sized grains of magnetite that are evenly dispersed through portions of the thin section. Together these amount to only a fraction of a percent of the chert. One important feature observed in the black chert member is the occurrence of spherical bodies which have been dissolved and partially infilled with chert (Figure 16). These are probably the remains of radiolaria, and along with the absence of any carbonate in the thin section, indicate that the black chert member was a primary deep marine chert, and not the product of the replacement by silica of a limestone bed.

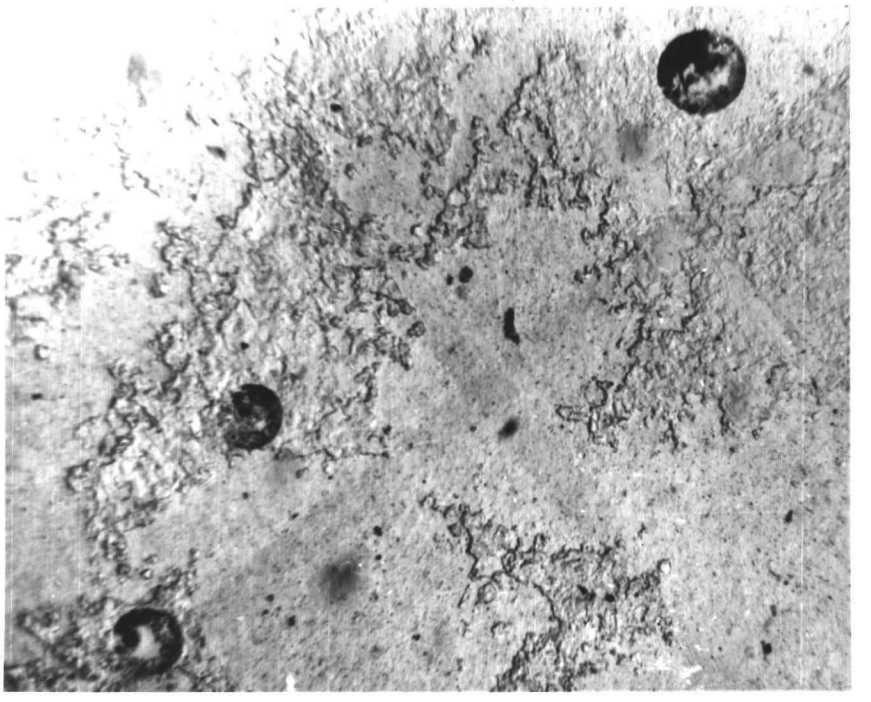

Figure 16. Spherical, partially infilled cavities in the Black Chert Member of the Spring Branch Group, believed to be the remnants of exsolved radiolaria. Note coarser grain size of silica within cavities. Magnification 100X.

Age

Since no fossils occur in the black chert member other than the spheres which are probably replaced radiolaria, no age can be assigned to the black chert member with any degree of certainty. The occurrence of radiolaria rules out the possibility of a Precambrian age. Tentatively, an early Paleozoic age can be ascribed to the black chert as was done with other members of the same group.

Structural Relationships

The lower contact of the black chert, with the unnamed red shale, is a thrust contact as evidenced by the shearing and fracturing of the lower part of the chert. The red shale is extensively sheared near the contact as well, as it is throughout its exposure. The upper contact of the chert, which separates it from the basalt member, may have originally been a depositional one but it since experienced considerable movement along its length. The northern half of the contact is a shear zone of basalt and chert slivers, and along the southern portion of the contact, red shale has been thrusted between the chert and basalt, while the chert itself has been faulted apart into separate blocks.

Thickness

The maximum thickness, from the upper to the lower contact, that the black chert member attains is 120 feet. This maximum exposure occurs in the vicinity of the "curve" of the outcrop, where the strike of the chert changes from north-south to northwest-southeast.

Environmental Interpretation

In my opinion, the black chert member of the Spring Branch Group represents a primary, deep-marine deposit of siliceous oozes. The occurrence of radiolaria, the lack of carbonate or clay minerals, and the association of the chert with subsequent deposits of oceanic alkali basalts and presumably deep-water limestone all support the the argument of a primary depositional origin.

The assumption that has been made in reconstructing the depositional environment for all the members of the Spring Branch Group is that the contacts between the four members of the group are the remnants of originally conformable, depositional contacts. If this assumption is valid, then the sequence within the group would appear to be right side up, with the black chert being the oldest member, and the massive limestone the youngest. The evidence for this conclusion is the occurrence of intraclasts of mottled micrite very

similar to the platy limestone member in the brecciated base of the massive limestone members. These clasts of platy limestone were probably carbonate mud rip-ups dislodged by the currents that carried the volumes of ooids to the basin where the massive limestone member was deposited.

Although shearing has taken place along the contact between the platy and massive limestone members and the contact between the platy limestone and the basalt members, these contacts appear to have been originally conformable. In a few localities, bedding within the limestone roughly parallels the orientation of the contacts; in these localities brecciation has been minimal. The original nature of the contact between the black chert and basalt members is much more uncertain. The contact has experienced extensive shearing all along its length, and it is unknown whether the basalt was actually extruded on a substrate of black chert; it is certainly conceivable that the two lithologies were formed in different portions of the same basin, or even separate basins, at different times and were later juxtaposed by one of the many events of thrusting that has affected the entire area.

Problem of Correlating the Duzel Rock and Spring Branch Groups

There are similarities between the association of rocks that comprises the Duzel Rock Group and those rocks that make up the

Spring Branch Broup. At first glance, these similarities might suggest that the two groups are correlative, and actually are fault-separated blocks of the same group. There are several structural and lithological differences, however, that tend to refute such an interpretation.

The most striking structural difference between the two groups is the nature of the contacts between the individual members. The chert, basalt, and two limestone members of the Spring Branch Group appear to conformably overlie each other, while the members of the Duzel Rock Group are a series of thrust slivers separated from each other by zones of sheared and brecciated rocks.

There are petrologic differences between each member of the Duzel Rock Group and its lithologic counterpart in the Spring Branch Group. The basalts of the Duzel Rock Group are vesicular and amydaloidal, contain occasional phenocrysts of plagioclase and exhibit well-defined pillow structures. The basalts of the Spring Branch Group contain no vesicles, even in thin section, are very aphanitic, and are structureless in outcrop. In addition, they contain a much higher percentage of opaque iron and titanium oxides than do the Duzel Rock basalts. The platy limestones of the two groups are perhaps the most dissimilar. The platy limestones of the Spring Branch Group are blockier in appearance and lack the flaggy fracture of their Duzel Rock counterparts. The Spring Branch platy limestones

have a lower clay content and lack the clay mineral concentrations along multiple laminations and stylolites that are so prevalent in the Duzel Rock platy limestones. The massive limestone member of the Spring Branch Group has been extensively recrystallized; as a result it is difficult to compare it to the oomicrites that cap Duzel Rock.

It is my opinion that these two groups can not be correlated and are in fact distinct lithologic associations. The limestone units of the Duzel Rock Group may be correlative with the Facey Rock Limestone to the south, but the Spring Branch Group is an assemblage that differs from those described to date from the eastern Klamath belt.

Unnamed Red Shale

Distribution and Type Locality

The red shale lies entirely in the eastern 1/4 of section 2. The unit crops out in a strip, oriented north-south, that occupies the center of the SE 1/4 NE 1/4 of section 2; the outcrop pattern then bends southeast to occupy much of the center of the NE 1/4 SE 1/4 of section 2. The type locality is designated as the exposure between the black chert member of the Spring Branch Group and the White's Gulch Limestone, in the SW 1/4 SE 1/4 NE 1/4 of section 2.

Lithology

Fresh surfaces of the unnamed red shale are consistently of a

moderate red (5R 4/6) color; weathered surfaces are also of a moderate red color usually, but in places attain a pale reddish brown (10R 5/4) or a moderate reddish brown (10R 4/6) color. The red shale is extremely fissile; no coherent samples thicker than 1.5 cm. can be collected due to the extreme friability of the strata. The fissility of the rock is in part due to original shale parting, but it is also the result of extensive faulting and shearing.

Slopes in the red shale are consistently gentle, due to its extreme friability and low resistance to weathering. Much of the exposure of the red shale lies in a small, flat valley between two slightly higher knobs.

The red shale is extremely fine-grained being composed entirely of clay minerals stained by iron oxides. No silt or sand layers have been located in the outcrop. Furthermore, no laminations or bedding features of any type are apparent, which makes an estimation of the amount of tectonic disruption that has occurred impossible.

Contacts

The eastern and western contacts of the red shale, with the Spring Branch Group and the White's Gulch Limestone respectively, have all the appearances of thrust faults. Both the bordering groups are sheared in the vicinity of the contacts, and in the southern portion of the exposure of the shale isolated blocks of black chert are

"floating" in a sheared matrix of red shale. The contacts and outcrop pattern of the red shale suggest that it is essentially a fault slice or wedge that was thrust in between the two large fault blocks occupied by the White's Gulch Limestone to the west and the Spring Branch Group to the east (Plate 2). All three have been thrust on top of the Moffett Creek Formation, either as a coherent mass or as separate blocks. The southern contact of the red shale, and presumably the southern contact as well (which is covered by deep regolith) are parts of this large thrust fault contact with the Moffett Creek Formation.

Thickness

The extreme faulting and shearing that has affected the red shale, coupled with the total lack of any bedding features, makes an estimate of the stratigraphic thickness of this lithology impossible. The maximum surface exposure, measured in an east-west direction over the ground surface, is roughly 200 feet wide, but this figure probably has little bearing on the true stratigraphic thickness of the red shale.

Age

No fossils were found in the red shale and as it is fault-bounded, none of the surrounding lithologies can shed light on the question of its age. Similar red shale has been observed at Bonnet Rock, eight

miles essentially due east of the Duzel Rock area. The red shale at Bonnet Rock is similarly fault bounded, and structurally underlies the Gazelle Formation of Silurian age (Potter, 1976, pers. comm.). The age of this red shale is uncertain, but it may be older than the Silurian formation it underlies. The possible correlation with the red shale at Bonnet Rock, together with its distribution as a fault wedge atop the Silurian calcareous siltstone of Moffett Creek, suggests an early Paleozoic age for the unnamed red shale.

Moffett Creek Formation

Distribution

The Moffett Creek Formation is the most areally extensive unit in the area of study; it completely encompasses the limestone fault blocks in the map area. It occupies almost all of section 35 and roughly 2/3 of section 2, extending beyond the map area on both sides of both of these sections. It similarly covers about 2/3 of section 36, extending beyond the northern edge of it. In section 1, the Moffett Creek Formation occupies a band in the northwest corner and along the western margin; it is bordered on the southeast both in this section and section 36 by the Duzel Phyllite.

Lithology

Weathered surfaces of the Moffett Creek Formation are often stained by iron oxides to a grayish orange (10YR 7/4), moderate yellowish brown (10YR 5/4) or dark yellowish brown (10YR 4/2) color. Many of the finer grained, more fissile outcrops are dusky yellow green (5GY 5/2), grayish green (5G 5/2), or light olive gray (5Y 6/1). Some outcrops have a pink tinge to them, assuming a light brownish gray (5YR 6/1) or pale red (5R 6/2) color. On a fresh surface, the Moffett Creek siltstones and sandstones are commonly light to medium light gray (N7-6).

The Moffett Creek Formation ranges from a silty mudstone through siltstones and sandstones to granule conglomerates, with siltstones and fine-grained sandstones being most prevalent. The conglomerates are virtually monomineralic, being composed of angular to sub-rounded fragments of gray, black and white chert in a siliceous matrix. Some exposures are extremely fine-grained and have shale partings.

Throughout its exposure, the Moffett Creek Formation has been complexly faulted and thrusted. Individual beds do not extend more than 3 or 4 m. before ending abruptly at a fault or shear zone. Attitudes of individual beds vary extremely over short distances. Commonly the more fine-grained outcrops have experienced the most

severe shearing. A. W. Potter (1976, pers. comm.) has termed this unit a "broken formation" based on work conducted south of the map area; certainly this applies to the siltstones around Duzel Rock.

Where not sheared, the siltstones and sandstones are extremely well indurated and have virtually no porosity, due to the high clay matrix content. Descriptions of this unit elsewhere cite the frequency of calcite as a cementing material, and previous names for this lithology include calcareous siltstone and calcareous wacke (Porter, 1974). Calcite is generally lacking in the samples collected in the Duzel Rock area; those samples that do contain appreciable amounts of calcite have been extensively sheared, the fractures being infilled by carbonate minerals.

As at Facey Rock, exposures that exhibit graded bedding occur in the Moffett Creek Formation, particularly in the coarser sandstone beds. A few load casts and features that may be flute casts were found in scattered pieces of float from this unit.

Visual estimates of the amount of matrix range from 15 to almost 50 percent. Generally the framework of the sandstones and siltstones is disrupted, but several samples, including the ones studied in thin section, show grain support. Calcite and silica cement are confined to vein fillings and healed fractures.

The major framework constituents are quartz and chert grains (Table 2). Other constituents, in decreasing abundance, are lithics,

plagioclase, green alteration products (celadonite, chlorite, and serpentine), microcline, biotite, opaques (chiefly magnetite), muscovite, and detrital calcite. The lithic fragments include, in decreasing abundance, granitics, mafic volcanics, mudstones and slates, chlorite schists, and wackes.

The matrix is composed of silt-size fragments of quartz and chert suspended in clay. A small amount of silt-size, presumably diagenic chlorite is also discernible in the matrix under high magnification.

Texturally, the rock is very immature, suggesting rapid transport and burial with very little reworking by currents. The high clay content and very poor sorting are indicative of this. There is a variation in the degree of rounding, but this is likely to be a reflection on the sources of sediment rather than the mode of transport. Mineralogically, the siltstones and sandstones are submature to mature; again, this is probably due to the nature of the source areas. Individual grains range from angular to well rounded; the average grain is subangular (following Pettijohn's (1957) rounding scale). Following the classification of Williams, Turner, and Gilbert (1954), the sample studied in thin section would be termed a lithic wacke; several of the samples collected would classify as quartz wackes. Pettijohn (1957) would classify this rock as a lithic graywacke; Krynine's (1948) term would be a low rank graywacke.

Table 2. Modal analysis (700 points) of fine-grained sandstone of the Moffett Creek Formation.

Constituent	Percent of thin section DZ15
Quartz	34.2
Chert	14.2
Lithics	10.8
Plagioclase	7.6
Celadonite, Chlorite, and Serpentine	5.8
Microcline	2.0
Biotite	1.6
Opaques	0.6
Muscovite	0.4
Detrital Calcite	0.2
Calcite Cement	0.4
Silica Vein-fillings	1.6
Matrix	21.0
Totals	100.0

Locality DZ15: NW 1/4 NE 1/4 SW 1/4 section 36

The mineralogy of the framework constituents of this rock suggests that there were multiple sources of sediments. The variety of lithic fragments evoke acid plutonic, basic volcanic, and metamorphic terranes. Some rare shard-like grains altered to celadonite in thin section are further indications of explosive volcanic activity. Some of the quartz grains have a high sphericity and a few have quartz overgrowths indicative of recycled sedimentary rocks, probably arenites, which are commonly associated with cratonic strata (Folk, 1968). Most of the quartz and chert grains, however, are very angular, indicating a short history of transport. A study of the quartz grains shows that virtually all of them have irregular (dust) inclusives, more than half contain acicular inclusions (needles), only a few have regular inclusions (equant crystals) and only one or two percent have vacuoles of gas or liquid. The dominance of irregular and acicular inclusions is indicative of a plutonic source (Keller and Littlefield, 1950). While much of the monocrystalline quartz in these wackes is probably plutonic in origin, many polycrystalline grains are clearly derived from a high-grade metamorphic terrain, as they contain crenulate boundaries between individual crystals in each grain and a few have mica flakes aligned in a preferred orientation, reflecting a foliation.

Age

No fossils occur in the Moffett Creek Formation within the map area. Clasts of this formation have been identified in conglomerates of the Gazelle Formation, which has been dated as late Llandovery or earliest Wenlock to Devonian and occupies a belt roughly six miles to the east. One of the clasts in the Gazelle conglomerate contains Silurian brachiopods (Potter, 1976, pers. comm.). The Gazelle Formation structurally overlies the Moffett Creek siltstones; if the former has been faulted on top of the latter, the latter is Silurian (and older?). If the Gazelle was deposited on top of the Moffett Creek Formation, the latter is Llandoverian (and older?).

Structural Relationships

The Moffett Creek Formation extends beyond the map area along the entire northern and western margins. The lower contact is exposed in the southeastern portion of the thesis area. Due to the steepness of the slope on which the contact is located, and the high degree of internal fracturing in the overlying units, most of the thrust contact of the Moffett Creek siltstones and sandstones with the underlying Duzel Phyllite is obscured by a thick talus cover. In addition, most of the entire east flank of the ridge is covered by talus, with

isolated outcrops of siltstone and sandstone visible above an elevation of roughly 4800 feet and similar disjointed fault-bound outcrops of phyllite below 4800 feet. In many localities the contact can only be approximated, as no outcrops appear for zones that encompass differences in elevation of over 160 feet. The locality where the contact is best preserved is roughly 70 meters northwest of the corner of section 1, where a small basaltic intrusion has been injected along the thrust contact (see section on Intrusive Rocks). Outcrops of Moffett Creek siltstones and silty shales occur a few meters above the basalt; blocks of phyllite compressed into tight chevron folds one to five centimeters in amplitude crop out immediately below the basalt intrusion. The intrusion itself has been separated into two bodies offset from each other, indicating that thrusting occurred after the basalt was injected along the contact. The contact in the vicinity of this basalt intrusion appears to be nearly vertical, angling slightly to the northwest.

In addition to the thrusting along its contacts, the Moffett Creek Formation has experienced extreme internal disruption in the Duzel Rock area. Where exposures are intensive, particularly along the crest and upper flanks of ridges, it is readily apparent that there have been several phases of tectonic activity. Groups of beds are consistently disjointed into fault band blocks and many have been compressed into broad folds up to a meter in amplitude. In addition, several of

these folds have experienced subsequent faulting, as the axis of one fold may dip in very different directions over a short distance.

The upper contacts of the Moffett Creek sandstones and siltstones, with the overlying fault blocks of the Duzel Rock group and the Spring Branch Group with the White's Gulch Limestone, are also thrust contacts. A small lens of tectonic breccia is exposed under the northeastern corner of the massive limestone of Duzel Rock. The massive limestone member of the Spring Branch Group is extremely brecciated along its eastern contact, and slivers of granule conglomerates of the Moffett Creek Formation have been emplaced in this brecciated zone. A larger block of a similar granule conglomerate has been thrust into the complex series of fault wedges that comprise the southern portion of the Duzel Rock Group. Most of the upper thrust contacts of the Moffett Creek rocks are obscured by talus from overlying units, but localities where contacts can be accurately determined include the southern edge of the White's Gulch Limestone and the southeastern edge of the Duzel Rock Group.

Thickness

There is no stratigraphic section of the Moffett Creek Formation preserved within the map area; all outcrops are disjointed and fault bounded. This may well be true of the entire formation throughout its exposure in the eastern Paleozoic belt of the Klamath

Mountains; A. W. Potter (1976, pers. comm.) envisions these sandstones and siltstones as comprising a "broken formation" in which the original thickness of strata may have been considerably telescoped by thrusting.

No reliable estimate of the stratigraphic thickness of this unit can be made in the map area. There is roughly 2000 feet of elevation difference between the highest exposure of the Moffett Creek formation and its exposure at the northwest corner of the area of study, but this is probably an inaccurate indication of the original depositional thickness represented here.

Environmental Interpretation

The lack of stratigraphic continuity in the Moffett Creek Formation makes a detailed paleoenvironmental interpretation difficult. Nevertheless, several features of the sandstones and siltstones indicate that these rocks were deposited in a deep marine setting, probably in an environment similar to a proximal turbidite. Kuenen (1964) listed features that are characteristic of ancient turbidite formations. Among these are graded bedding, load casts and sole markings, features identified in the Moffett Creek Formation at Duzel Rock. Other features that can be used to identify turbidite deposits are described by Bouma and Hollister (1973); those included in their list that occur in the Moffett Creek Sandstones are moderate to poor

sorting, sharp bottom contacts (where visible), lack of bedding, little or no preferred grain orientation in graded portions, and classification as a graywacke or subgraywacke (following Pettijohn's (1957) classification). Kuenen (1964) states further that the scarcity of fossil material and a size distribution ranging from fine silt to medium pebbles are common although not diagnostic features of turbidite sandstones. The Moffett Creek clastics range from fine silt to granules, and no fossil remains were found in this formation in the Duzel Rock area.

Kuenen also states that the absence of shallow water features such as mudcracks, oscillation ripple marks, channels, winnowed sands, etc., is an important requirement for the interpretation of a turbidite. None of these shallow water depositional features were found in the Moffett Creek Formation.

Over most of its exposure in the eastern Paleozoic belt of the Klamath Mountains, the Moffett Creek Formation is highly sheared and disturbed. There is a locality south of the town of Callahan, however, where relatively undisturbed, alternating, rhythmic beds of sandstone and shale have been observed (D. M. Rohr, 1976, pers. comm.). Such rhythmic bedding is characteristic of flysch deposits, which most authors feel are deposited in part by turbidity currents (Dzulynski and Walton, 1965).

Many turbidity current deposits are presently being interpreted

by using a deep sea fan model (Nelson and Kulm, 1973; Nelson and Nilsen, 1974). In terms of such a model, the portions of the Moffett Creek Formation exposed in the Duzel Rock area would classify as upper fan or middle fan deposits. Evidence for such a classification includes the predominance of sand and silt over shale, the occurrence of coarse granule conglomerates, and the poor sorting.

In classifying the Moffett Creek Formation as a turbidite, an implication of continental origin is not necessarily intended. A few recycled sedimentary grains suggest a cratonic source, but the large majority of the clastic material could have been derived from an island-arc source. Such a source would fit in with Hsü's (1970) conceptualization of flysch settings. Hsü envisions flysch sediments being deposited in the deep marine trench that is formed seaward of an island-arc by plate convergence; most if not all of the sediments that would be rapidly eroded and transported to such a trench would be derived from the nearby island arc. Such a model is consistent with the sedimentary features of the Moffett Creek Formation.

Duzel Phyllite

Distribution

The Duzel Phyllite lies in the southeastern corner of the map area. It occupies roughly half of section 1, extending beyond the

southern and eastern borders of this section. It occupies approximately 1/8 of section 36, the extreme southeast corner.

Lithology

Weathered surfaces of the phyllite are light olive gray (5Y 6/1) or greenish gray (5GY 6/1 and 5G 6/1) and commonly have a micaceous sheen along the foliation planes. Fresh surfaces that transect the foliation are usually medium light gray (N6) and sometimes have a light brownish gray (5YR 6/1) tinge to them. The most conspicuous feature of the Duzel Phyllite is the high degree of folding evident over most of its exposure. The foliation is often crumpled into tight isoclinal folds which have amplitudes ranging from a few mm. to 6 cm. (Figure 17). Many of these folds are composite, having several smaller crenulations within the crest of each fold. A sawed surface reveals that the rock is composed of thin, alternating layers, up to 1.5 mm. thick, of reddish brown phyllosilicates and gray, granular tectosilicates, but the rock is too fine grained in most samples to discern individual minerals.

The entire exposure of the phyllite within the area of study has experienced extreme internal disruption. Isolated outcrops are often suspended in a sheared matrix, and most of the unit is only visible as float. Much of the exposure is further obscured by talus derived from the overlying units to the west. Individual outcrops are

Figure 17. Sample of float from Duzel Phyllite. Note intense crenulation of laminae and tight chevron folds. Penny for scale. SE 1/4 SE 1/4 NW 1/4 section 1.

often very friable, having a tendency to split into plates parallel to the foliation. In addition, there are several small veins of calcite and of a serpentine-chalcedony mixture up to 2 mm. wide. Most of these run parallel to the foliation and have been folded with the other layers of the rock, but a later generation of calcite veins has filled cavities along the axes of the larger folds.

In thin section, the phyllite can be seen to have experienced a mineral segregation into alternating bands of phyllosilicates and tectosilicates. The phyllosilicate layers consist primarily of bent and kinked aggregates of chlorite. Incorporated between flakes of chlorite are patches of muscovite and a few elongate actinolite crystals. The tectosilicate layers, which are roughly four times as wide as the chlorite bands, are characterized by large, angular quartz grains imbedded in a matrix of granulated, silt-size quartz and chert. Except for the larger, roughly equidimensional quartz grains, all the minerals in the tectosilicate bands have been elongated by shear and are oriented so they closely parallel all the folds and kinks of the phyllosilicate layers. The constituents of the tectosilicate bands are, in decreasing order, quartz (most of which exhibits undulatory extinction), chert (most of which has experienced at least partial recrystallization to coarser-crystalline quartz), opaques (chiefly hematite, but some elongate magnetite grains), plagioclase, epidote, microcline, sphene, lithics, and actinolite (see Table 3).

Table 3. Modal analysis (700 points) of the Duzel Phyllite.

Constituent	Percent of thin section DP1
Quartz	41.4
Chlorite	20.1
Chert	10.2
Muscovite	9.7
Opaques	5.0
Plagioclase	2.3
Epidote	2.1
Microcline	1.3
Sphene	1.3
Lithics	0.3
Actinolite	Tr.
Calcite	4.3
Serpentine-chalcedony vein fillings	2.0
Total	100.0

Locality DP1: SE 1/4 SE 1/4 NW 1/4 section 1

The intense folding that has affected the phyllite created elongate cavities, parallel to the foliation, along the crests of folds. These cavities were later infilled, most of them with a serpentine-chalcedony mixture, a few with sparry calcite. A later set of calcite veins are superimposed over the deformed rock fabric; these veins transect the foliation and parallel the axes of folds. In addition, irregular patches of calcite have replaced portions of the phyllite. Patches of the phyllite have also been altered to, or replaced by, irregular grains of hematite. Some of the hematite replacement material can be seen to be pseudomorphs after sphene or epidote crystals; other patches of hematite transect crystal boundaries and have obliterated the original minerals.

The mineral assemblage of muscovite, chlorite, albite and epidote indicate that the phyllite has been metamorphosed to the greenschist facies (Williams, Turner, and Gilbert, 1954, Chapter 12). The mineral assemblage further indicates that the original sedimentary rocks from which the phyllite was derived were originally either quartz-feldspathic or pelitic detritus, and probably a combination of the two. It is the author's impression that the Duzel Phyllite may represent a thick segment of the Moffett Creek Formation that was subjected to regional metamorphism. The high percentage of quartz and chert, the rare grains of lithic fragments, plagioclase and microcline, and the high percentate of phyllosilicates indicate that the

phyllites could have had their source in the wackes and silty mudstones of the Moffett Creek Formation.

Structural Relationships

As discussed in the descriptions of the Moffett Creek Formation, the upper contact of the Duzel Phyllite, although largely obscured by talus, has all the appearances of a thrust fault. This contact is best exposed near the center of section 1, where a basalt intrusion has been injected along the fault. In addition to the Moffett Creek Formation that has been thrust over the phyllite, a block of the massive limestone member of the Duzel Rock group lies on top of the phyllite. The lower contact of the limestone block is also a thrust fault as evidenced by the sheared limestone that is visible along the block's eastern margin. This block of limestone is, to the author's knowledge, the only case of a carbonate thrust block sitting on top of the Duzel Phyllite rather than the Moffett Creek sandstones and siltstones. This would suggest the thrusting of the limestock blocks of Duzel Rock occurred after the Moffett Creek Formation was superimposed by thrusting over the Duzel Phyllite.

The lower contact of the Duzel Phyllite is not exposed in the area of this study.

Thickness

The extreme amount of internal dislocation within the Duzel Phyllite renders any estimate of the thickness of this unit meaningless. Furthermore, the lower contact of the phyllite lies outside the map area. The difference in elevation between the lowest exposure of the phyllite in the map area and its upper contact is approximately 1000 feet, but this is not likely to be any indication of the original thickness represented here.

Age

The author is of the opinion that the Duzel Phyllite may be a metamorphosed equivalent of the Moffett Creek Formation. This implies that original age of the sediments is Silurian. The metamorphism may have occurred in either of the two major tectonic events that have affected the Klamath Mountains. The Duzel Phyllite may have undergone metamorphism in the middle of the Devonian; strontium evolution diagrams indicate an age of primary metamorphism for the Abrams mica schist of the Central Metamorphic Belt of 380 M. Y. (Lanphere and others, 1968). Evidence for Middle Devonian tectonism exists in other localities in western North America, including the southern Sierra Nevada, northwest Washington, and southeastern Alaska (Boucot and others, 1974). Alternatively, the Duzel

Phyllite may have been metamorphosed in the widespread Nevadan Orogeny of Jurassic and Cretaceous age. The Stuart Fork Formation, a sequence of phyllites, schists and micaceous quartzites lithologically similar to the Duzel Phyllite (Zdanowitz, 1971), experienced metamorphism during the Nevadan Orogeny (Lanphere and others, 1968). It was also during this orogeny that most of the large plutons in the Klamath Mountains were emplaced.

INTRUSIVE ROCKS

Distribution

Several small dikes and pods of andesite occur in various localities within the area of study. Dikes transect various members of the Duzel Rock Group in the NW 1/4 of section 1; the intrusions in and around the Spring Branch Group, in the eastern edge of section 2, are more lenticular in shape.

One outcrop of intrusive basalt was located. It has intruded along a major thrust fault, and lies in the SE 1/4 SE 1/4 NW 1/4 of section 1.

Lithology

The one basalt intrusion straddles a large thrust fault between the Moffett Creek Formation and the phyllite, and has been severely sheared by movement along this fault. The contacts on either side of the fault do not parallel each other; the outcrop of the basalt itself has been offset several feet. The basalt has experienced extreme chemical alteration--streaks of green weathering products and veins of oxide-stained calcite pervade the rock. No phenocrysts are visible; the basalt was originally very fine-crystalline, and it has been so extensively altered and fractured that it is almost impossible to determine its original composition.

The andesite intrusions are usually dark greenish gray (5G 4/1) on a weathered surface; some weathered surfaces are stained olive gray (5Y 4/1) by clay minerals and iron oxides. On a fresh surface it has a greenish black (5G 2/1), or a medium dark to dark gray (N 5-4) color. It is highly resistant to weathering and often stands out in relief from the host rock. The thickest dikes are roughly 45 feet wide, most are between 10 and 20 feet wide. The longest exposure along strike is roughly 150 feet.

The andesites have a fresh appearance; they are not cut by the numerous fine fractures and calcite-filled veins that transect the surrounding host rocks. Most of the intrusions are offset in at least one place by near-vertical faults. Blocks of andesite, bordered by widely separated fractures, commonly exhibit spheroidal weathering; rounded knobs protrude out from the host rock.

The andesites have a porphyritic appearance in hand specimen; elongate, black minerals stand out against a fine-crystalline, gray-green groundmass. In thin section, these black phenocrysts are, in some samples, exclusively hornblende crystals, and in other samples the phenocrysts include hornblende, a brown, pleochroic orthopyroxene (bronzite) and colorless clinopyroxene. All these mafic minerals have experienced considerable alteration to chlorite. The groundmass consists of a mixture of plagioclase laths and small microlites of clinopyroxene with a diabasic texture. A few small

interstices between plagioclase crystals are filled with brown glass. The plagioclase subhedra have experienced substantial alteration, the principal weathering products being sericite and kaolinite.

Contacts

All of the contacts of the andesites are intrusive. These intrusions have experienced virtually none of the shearing and folding that has affected all of the surrounding host rocks, and they transect all bedding and tectonic features except the last episode of near-vertical faulting. Evidence for contact metamorphism exists in one locality, where the massive limestone member of the Duzel Rock Group in the NE 1/4 NW 1/4 of section 1 has been recrystallized extensively along the edge of the andesite dike.

The contacts of the basalt intrusion were probably originally intrusive, but as the basalt intruded along a thrust fault, it has experienced considerable shearing since its emplacement. All of the rocks in the vicinity of this major thrust plane are highly disturbed, and often the original contacts have been obliterated.

Age

The andesites transect the youngest unit in the area, the Moffett Creek Formation of Silurian age, and hence must be younger than Silurian. Further, as the dikes have intruded all the thrust blocks

that lie on top of the Moffett Creek Formation, they must have been emplaced after the principal tectonic episodes, which probably took place in the Devonian and/or Jurassic to Cretaceous and may have occurred sporadically for a long time afterwards. It is hence likely that the intrusions are, at the oldest, late Mesozoic in age. They may be as young as Tertiary, and could be offshoots of the intense Cascade volcanism that created lofty Mt. Shasta to the east.

CENOZOIC BRECCIA

Distribution

Surficial deposits of carbonate-cemented breccia, composed almost entirely of limestone clasts, exist in two portions of the map area (Plate 1). One is a continuous, stratified deposit covering over half of the small block of the massive limestone member of the Duzel Rock Group, in the NE 1/4 of section 1. The other deposit occurs as scattered outcrops mantling massive limestones of the White's Gulch Limestone in the SE 1/4 of section 35.

Lithology and Morphology

The two deposits, while they are both composed of cemented limestone debris, are sufficiently distinct in their lithology and outcrop pattern as to warrant separate descriptions (see Figure 18). The more easterly, stratified breccia is composed primarily of clasts of massive and platy limestone with rare fragments of wacke. The clasts range from sand-size to 1.0 m. across, but most clasts measure 1.0 to 2.5 cm. along the longest dimension. The clasts range in shape from roughly equidimensional to tabular, the latter being exclusively derived from platy limestones. All the clasts are loosely bound by a friable, porous mixture of iron oxides, clay minerals and carbonate cement, presumably aragonite. This matrix is of

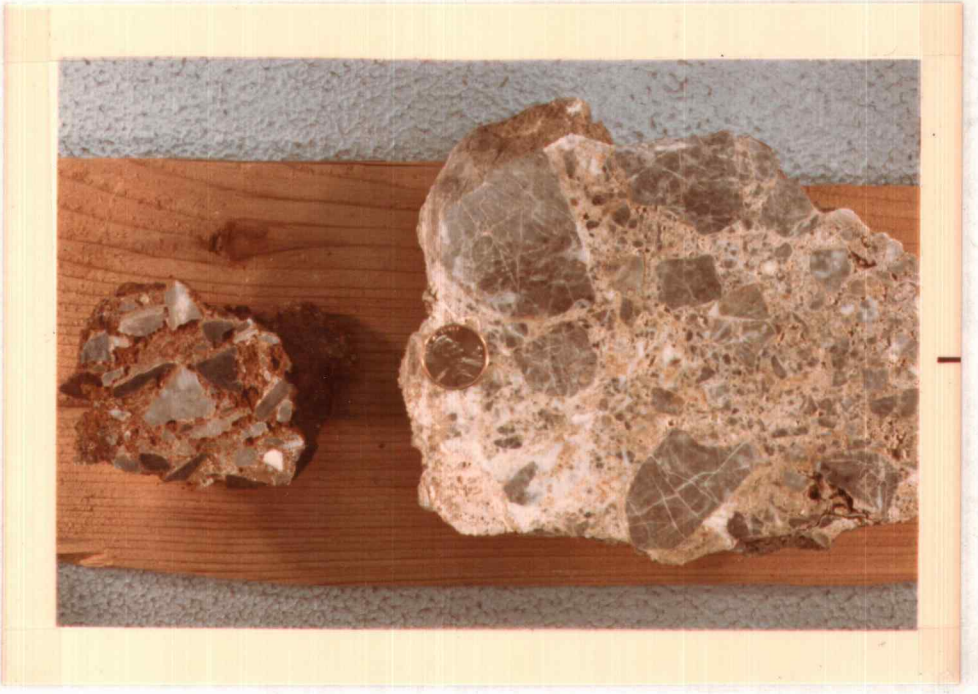

Figure 18. Samples of Cenozoic breccias from the Duzel Rock area. Sample on left is characteristic of thick deposits to the east of Duzel Rock. Sample on right was collected from thin deposits mantling the White's Gulch Limestone to the west of Duzel Rock. Penny for scale.

a light brown (5YR 5/6) color due to the abundance of ferric oxides. This deposit of breccia mantles much of the underlying block of massive limestone and drapes over a segment of its northern edge. On this northern edge the breccia achieves its maximum thickness of almost 10 m. At this locality the breccia is crudely stratified (Figure 19). The bedding is controlled by variations in the degree of cementation, and dips into the hill (to the southwest) at an angle of approximately 30°.

The breccia to the west of Duzel Rock occurs as isolated patches ranging from 20 to 1200 feet long. These deposits generally form a thin cover over the massive limestones underneath; the maximum thickness observed in these breccias was 0.8 m. This breccia is composed exclusively of clasts of massive limestone and fragments of white calcite. The clasts range in size from sand to pebbles 10 cm. long. While most of the clasts are angular, several of the largest clasts have rounded corners. These breccias are better cemented, less porous and less friable than the breccia to the east, in section 1. The cement has a very pale orange (10YR 8/2) color and is composed almost entirely of calcium carbonate with a small amount of clay and other impurities. It is somewhat porous, but in general the pores are small and unconnected.

In his study of Facey Rock, R. W. Porter (1974) identified several patches of similar limestone breccia, and he attributed their

Figure 19. Stratification in Cenozoic breccia, composed of limestone talus. View looking west, Duzel Rock in background. NE 1/4 NW 1/4 NE 1/4 section 1.

formation to "cementation of Quaternary landslide and talus deposits by groundwater supersaturated with calcium carbonate." This author favors a similar interpretation for the breccias on the flanks of Duzel Rock. There are several springs on the west side of Duzel Rock that flow from elevations above the patches of breccia on the White's Gulch limestone. No springs were observed on the east flank of Duzel Rock, but the stratified breccia lies nearly adjacent to a major thrust fault. It is possible that at times of more abundant precipitation in the past, springs arose from the east side of Duzel Rock and transported enough calcium carbonate in solution to cement the thick deposits of breccia. Studies of the groundwater draining from Duzel Rock and Facey Rock show that the water is very hard, with a high calcium carbonate content (Mack, 1958). It is the author's opinion that much of the cementation of these breccias could have occurred quite recently, although no incorporated plant or animal remains were found within the rock.

Romey (1962) interpreted the limestone breccias at Facey Rock as thrust breccias, and Porter (1974) rejected this theory for several reasons. Those objections that also apply at Duzel Rock are as follows: 1) The distribution of the lithology, in many places, has no apparent relation to a thrust surface. 2) The clasts do not have regular fractures or show slickenslides. 3) The outcrops have a surficial appearance, often conforming to slope. Furthermore no

exotic fragments or fragments of the surrounding units have been incorporated into the breccia.

STRUCTURE

Introduction

Five large thrust faults dominate the structure of the Duzel Rock area. The largest of these extends beyond the map area and is considered to be of a magnitude comparable to the regional Mallethead Thrust of Churkin and Langenheim (1960), which it roughly parallels. All the other thrusts within the map area are believed to be subsidiary to the largest one. In addition to the large thrusts, three smaller thrust faults have been identified. Several high angle faults with relatively small displacement occur in the area. The internal structure of the sedimentary units is generally disrupted and often extremely complex.

Faults

General

Figure 20 depicts the pattern of faulting in the Duzel Rock area with individual faults labeled for the purpose of discussion. These are only the most significant faults in the area. There are many of lesser magnitude which are not shown. Most of the units are severely fractured by several generations of small-scale faults, making it impractical to trace the exposure of these small faults.

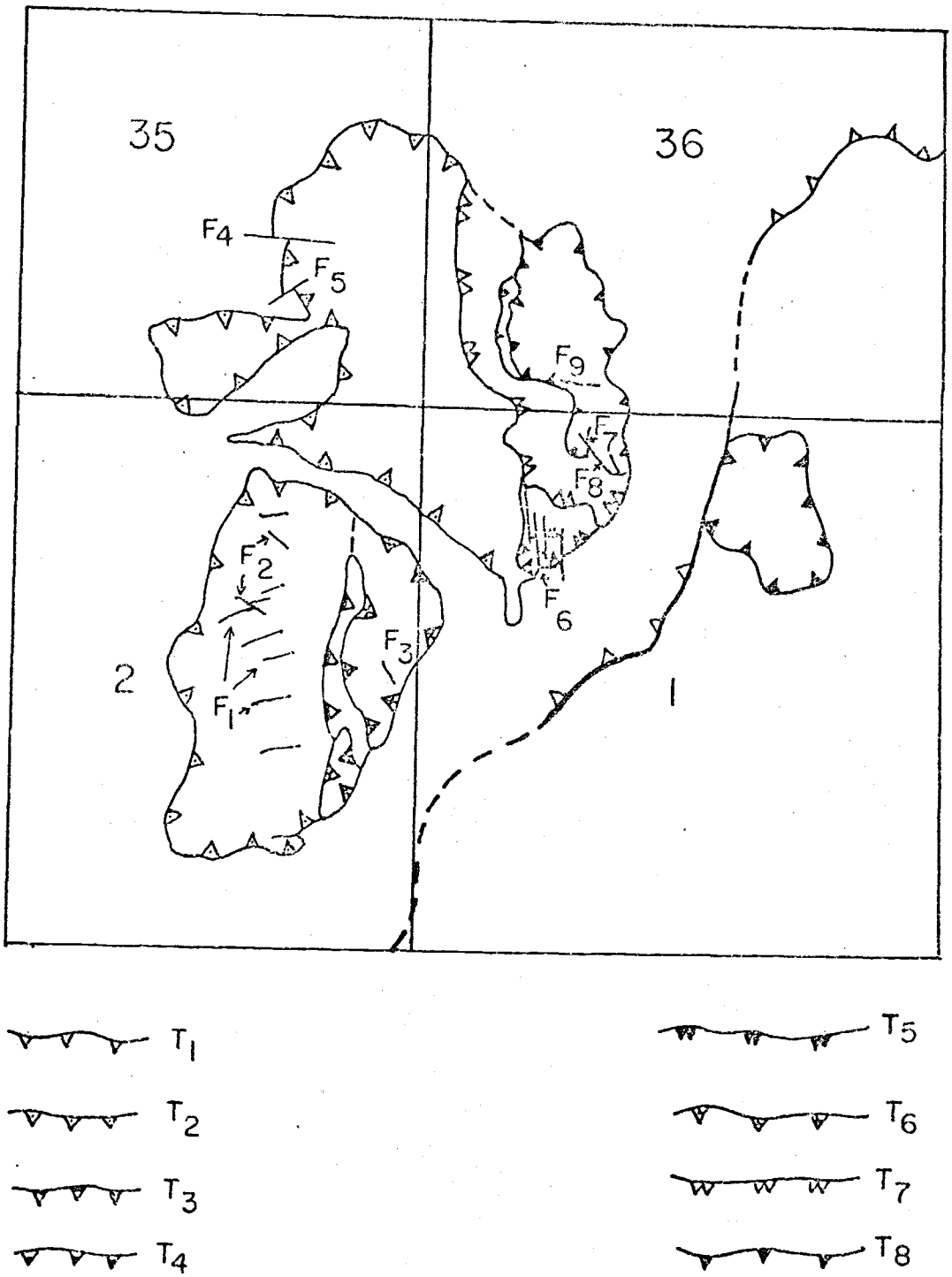

Figure 20. Major faults in the Duzel Rock area.

The smallest thrust fault, labeled T_4 in Figure 20, in actuality encompasses a complex series of fault wedges and slices, too small to be delineated in a diagram of this particular scale.

The high angle faults depicted in Figure 20 are only those which transect a marker bed or dike and along which the amount of offset can be measured. Several other faults can be inferred, but in most cases their extent, amount of offset and often the direction of dip can not be determined due to the jumbled nature of most units. Where a set of parallel or nearly parallel faults occur these are labeled with one number on Figure 20.

Types of Faults and Evidence

The faults in the area of study can be divided into several groups. In Figure 20, each thrust fault has been labeled with a T and each high angle fault with an F. Thrust fault T_1, the contact between the Duzel Phyllite and the Moffett Creek Formation, is obscured along most of its length by talus. This fault is best exposed near the center of section 1, as described earlier in the discussion of the structural relations of the Moffett Creek Formation. The contact of the White's Gulch Limestone with the Moffett Creek Formation along T_2 is exposed in a few places; a lens of sheared breccia along the western edge of the southern limestone block, and the brecciated nature of the limestone in the vicinity of high-angle faults F_4 and F_5

provide evidence for this thrust. The lower contact of the Unnamed Red Shale along thrust T_3 is only exposed along its northern end. Here the White's Gulch Limestone abuts directly against the shale; both units are highly fractured and sheared near the fault. The upper contact of the red shale along thrust T_6 is well exposed. The black chert member of the Spring Branch Group has apparently been thrust over the red shale. Both lithologies are extremely sheared along the contact, and at the contact's southern end blocks of chert have been broken off and are now suspended within the sheared shale. The eastern edge of thrust T_6 is quite shallow along its surface expression. Well exposed in a road cut, the contact is composed of interfingering slivers of Moffett Creek sandstones and Spring Branch massive limestones which are extremely brecciated. The thrust fault designated as T_4 actually encompasses a complex zone of faulted wedges and slices (see Plate 2). Blocks of limestone, chert and conglomerate are surrounded by sheared basalt breccias and Moffett Creek mudstones. The pillow basalts that are encompassed by thrust T_5 are bordered on the north and south by thick outcrops of thrust-generated basalt breccia, which was described previously as a separate member of the Duzel Rock Group. Most of the lower contact of the platy limestone member of the Duzel Rock Group along T_7 is covered by talus and regolith. The high degree of internal faulting and small-scale folding within the platy limestone member may be

due in large part to the stress generated by the overthrusting of the massive limestone along thrust T_8. A wide brecciated and faulted thrust zone is exposed by road-cuts at the base of the massive limestone cliffs. This zone contains a sheared mudstone which encompasses blocks of platy limestone and altered porphyries; many high-angle faults truncate the contacts and layering within the zone. The majority of the remaining portions of T_8 are covered by talus derived from the massive limestone cliffs.

Fault groups F_1 and F_2 are nearly vertical normal faults. The motion along these faults was probably oblique, with both a dip-slip and a strike-dip component. The total amount of offset of the chert marker bed, where these faults are best exposed, does not exceed 150 feet. Fault F_3 was inferred from the truncation and offset of the outcrop pattern of an andesite dike.

Faults F_4 and F_5 were inferred to explain the offset in the thrust contact between the Moffett Creek Formation and the White's Gulch Limestone. The motion on these faults was probably dip-slip, as a chert bed is truncated by both of these faults and does not appear outside the faults.

The nearly parallel set of faults designated F_6 are largely strike-slip faults inferred to explain the large offsets in the outcrop pattern of the Duzel Rock pillow basalts. The maximum amount of offset along these faults is approximately 450 feet. Fault F_7 and the

two parallel faults designated as F_8 are normal faults with dips between 60 and 70 degrees. The offset along F_7 does not exceed 50 feet and can be measured by the displacement in the bed of blue chert below the massive limestone. By the same criterion, the larger of the two faults grouped as F_8 has an offset of approximately 250 feet. Fault F_9 was inferred to explain the location of a large valley in the massive limestone. Jumbled piles of large blocks of limestone in this valley are indicative of past movement along this zone.

Age of Faults

Relative ages for most of the faults in the map area can be established on the basis of cross cutting relationships. As a group, the thrust faults are all older than the high-angle faults. The largest thrust T_1 is certainly as old as, and may be older than, the other thrust faults in the area. T_1 is probably contemporaneous with the Mallethead Thrust of Devonian to Cretaceous age. The other thrusts, T_2 through T_8, may be contemporaneous with T_1 or may have been overthrusted during a later tectonic episode. Assuming each thrust plate represents a different episode of thrusting T_1 would be the oldest and T_8 the youngest. The relative of ages of thrusts T_3 and T_4 is impossible to determine.

Fault sets F_1 and F_2 may or may not predate the thrusting along T_2. They are completely contained within the thrust plate and may be

a result of the tension created by the thrusting. F_3 is a relatively young fault, as are fault sets F_6 and F_8, as all these offset andesite dikes which may be as young as Tertiary. Faults F_4 and F_5 post-date the main episode of thrusting along T_2 as they have produced offset in this thrust contact.

There is evidence that the Duzel Rock area has experienced some seismic activity in the very recent past. Stratified deposits of carbonate-cemented talus, believed to be late Cenozoic in age, were described in the previous section. Presumably, the stratification in these breccias was originally horizontal, or nearly so. At present the bedding in the limestone breccia east of Duzel Rock dips into a slope at an angle of 30°. The author believes this rotation has occurred quite recently in the long, and probably ongoing seismic history of this region.

GEOLOGIC HISTORY

The paleodepositional environments of the Moffett Creek Formation and the various members of the Duzel Rock Group, the White's Gulch Limestone, and the Spring Branch Group were discussed in the sections describing these units. In summary, the Silurian Moffett Creek Formation is interpreted as a proximal turbidite fan deposit. The massive limestone member of the Duzel Rock Group is believed to have been deposited at the foot of an escarpment of a nearby carbonate bank; the platy limestone member is interpreted as a deep, open-marine carbonate mud, and the pillow basalt member is viewed as a segment of oceanic crust extruded upon an oceanic plate. The White's Gulch Limestone is considered to have been deposited in a medium-depth inter-arc basin which periodically received terrigeneous sediment from an emergent volcanic source and which was also periodically cut off from detrital sources by tectonic silling. The sequence of radiolarian chert, basalt, platy limestone, and massive limestone represented by the Spring Branch Group is interpreted as a sequence of deep-water deposits formed in a marine basin that periodically received abundant oolitic and carbonate mud detritus from a nearby shallow-marine carbonate bank.

All of the limestone and basalt units are allochthonous, but it is likely that they were not transported more than a few tens of miles.

The tectonic rearrangement of these presumably Ordovician or older rocks makes paleogeographical reconstruction extremely difficult. Collectively, the limestones and the Moffett Creek Formation may be indicative of an island-arc environment in an eugeosynclinal setting. The author envisions the various limestone members as being deposited in a complex series of inter-arc basins, landward (to the east) of the principal axis of the island-arc. Such a reconstruction was envisioned on a large scale by Churkin and McKee (1974) and was named the Klamath volcanic arc. Inside of such a volcanic arc, tectonic activity (generated, perhaps, by the convergence of major plates under the island-arc) would create a series of tectonically silled basins and emergent or nearly emergent carbonate platforms where high-energy currents, necessary for the formation of ooids, would prevail. The author would place the depositional environment of the Silurian Moffett Creek turbidites seaward (to the west) of the volcanic island arc, where a trench could accommodate large volumes of terrigeneous clastics. The island arc would be one with a complex history of plutonic and metamorphism, to account for the abundance of plutonic quartz and the variety of lithic fragments that constitute the Moffett Creek Formation.

The structural history of the area was discussed in the previous section. To summarize, a series of imbricate thrusts has presumably created an inverted stratigraphic sequence with Silurian

turbidites overlain by Ordovician (?) or older (?) limestones. While tectonism affected much of the rock of the Central Metamorphic belt of the Klamath Mountains in the Devonian, many authors are of the opinion that the severe thrusting evident throughout the Klamath Mountains province took place in Late Jurassic time (Irwin, 1964; Davis, 1968). G. A. Davis (1968) states that the thrusting in the Klamath Mountains has been to the west with a minimum displacement of each major belt relative to each other of 20 miles. Westward thrusting would account for the relationship observed at Duzel Rock, the superposition of inter-arc basin deposits over turbidites originally deposited to the west of the island arc. Davis also believes that the Late Jurassic episode of thrusting was accompanied by low-grade, regional metamorphism; it is probable, then, that this is the time when the Duzel Phyllite was recrystallized from its parent sedimentary rocks.

Andesite intrusions were subsequently injected into the rocks of the Duzel Rock area; presumably these intrusions are related to the Cenozoic volcanic activity of the Cascades Province to the east. The intrusive episode was followed by high-angle faulting with relatively little displacement.

The area has been severely eroded since the episode of thrusting, leaving Duzel Rock as a klippe. R. W. Porter (1974) suggested the thrusts underlying the massive and platy limestone members of

the Duzel Rock Group may correlate with similar thrusts at Facey Rock. The situation at Duzel Rock is complicated by the occurrence of limestone types, not observed at Facey Rock, as thrust plates between the Duzel Rock Group and the Moffett Creek Formation. Further efforts to determine a reliable age for any or all of the carbonate rocks at Duzel Rock would provide a great deal of insight into this still unsolved riddle.

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APPENDIX

Insoluble Residue Results

Appendix Table 1. Insoluble-Residue Percentages from Limestones of the Duzel Rock Group.

Collected from	Locality	Insoluble-residue percentage
Massive Limestone Member of the Duzel Rock Group	DQ 4	0.04
	DZ 78	0.28
	DZ 20	15.30
Platy Limestone Member of the Duzel Rock Group	DQ 7	4.11
	DZ 34	18.38
	DZ 73	25.53
	DZ 74	6.20
	DQ 11	25.59

Limestones were dissolved in concentrated hydrochloric acid

Appendix Table 2. Insoluble-Residue Percentages from Limestones of the White's Gulch Limestone.

Collected from	Locality	Insoluble-residue percentage
lower member of the White's Gulch Limestone	DS 1A	21.33
	DS 7	3.17
	DS 7A	32.37
	DS 8	8.12
	DS 9B	3.93
	DS 10	38.90
upper member of the White's Gulch Limestone	DS 16A	1.72
	DS 18	1.35
	DS 22	1.78
	DQ 2	7.01
	DZ 76	0.23

Appendix Table 3. Insoluble-Residue Percentages from Limestones of the Spring Branch Group.

Collected from	Locality	Insoluble-residue percentages
Massive Limestone Member of the Spring Branch Group	DZ 44	0.42
	DZ 45	7.30
	DQ 17	0.42
Platy Limestone Member of the Spring Branch Group	DZ 16	1.04
	DZ 46	7.38

Limestones were dissolved in concentrated hydrochloric acid.