AN ABSTRACT OF THE THESIS OF

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Title: GEOLOGY OF THE NICOLAI MOUNTAIN-GNAT CREEK AREA, CLATSOP COUNTY, NORTHWESTERN OREGON

Abstract Approved: _____

Alan R. Niem

Twelve rock units, from upper Eocene to middle Miocene are exposed in the Nicolai Mountain-Gnat Creek area. They are, from oldest to youngest: Pittsburg Bluff Formation; Oswald West mudstone; Big Creek sandstone, upper Silver Point mudstone, and Pipeline mudstone members of the Astoria Formation; Depoe Bay Basalt; Grande Ronde Basalt; Cape Foulweather Basalt; and sandstone correlative to the Vantage Member of the Ellensburg Formation; Frenchman Springs Member of the Wanapum Basalt; Clifton Formation (defined by author); and Pomona Member of the Saddle Mountains Basalt.

Upper Eocene Pittsburg Bluff Formation is a fine-grained sandstone and laminated siltstone deposited in an inner to middle continental shelf environment. Concordant onlapping by the upper Oligocene, middle shelf to upper slope glauconitic Oswald West mudstone, establishes a tie between the northeastern and northwestern Oregon Coast Range stratigraphies.

The lower to middle Miocene Astoria Formation represents another marine onlap sequence, beginning with the fine-grained cross-bedded inner shelf Big Creek sandstone member. The overlying laminted upper Silver Point and Pipeline mudstone members were deposited in a deepening middle shelf to upper slope environment.

The middle Miocene tholeiitic basalts extruded from two sources. Grande Ronde, Frenchman Springs, and Pomona basalts of the Columbia River Group erupted from fissures east of the Cascades and flowed down an ancestral Columbia River valley entering the sea in the study area. Simultaneously, petrologically similar but less voluminous Depoe Bay and Cape Foulweather are correlative to the low MgO Grande Ronde and Frenchman Springs in the study area. Abundant dikes and sills and the bathyal mudstone interbeds suggest that the coastal pillow basalts extruded locally onto the sea floor. Subaerial plateau-derived flows are associated with cross-bedded, fluvial to shallow marine arkosic sandstone interbeds. Some of these subaerial lavas flowed into the sea, forming "lava deltas" of foreset pillow palagonite and possibly "invasive" sills. The basalt stratigraphy allows a nearly flow by flow correlation of Grande Ronde and Frenchman Springs units from the Clackamas River area (Western Cascades) into this study area. The presence of Pomona Member in this area is the first substantiated recognition of this petrologically and chemically distinctive subaerial flow in northwestern Oregon.

The 200-meter thick Clifton formation (previously called the Pliocene (?) sandstone at Clifton) is now dated by diatom assemblages and stratigraphic position as middle Miocene in age. Three lithofacies are recognized. Facies 1, at the base and top of the unit, is an arkosic, fine- to coarse- grained sandstone with cross-bedding, vertical Rosselia burrows, lignitic coal beds, and rare molluscan shells. It represents a river mouth and shallow marine offshore bar deposit. Facies 2, in the middle of the formation, consitst of well-laminated diatom-bearing carbonaceous and micaceous shelf/slope siltstones with thin fine-grained turbidite sandstones. Facies consists of channelized siltstone breccias, chaotic debris flows, thick amalgamated arkosic grain flows, structureless sandstones, and minor volcanic pebble conglomerates. These

lithologies suggest a canyon head and slump deposit formed as a submarine channel cut into the shelf/slope siltstones of Facies 2. Sandstone petrography and grain size analyses indicate that the detritus of the Clifton and other sandstones in the area was derived from acid igneous, metamorphic, and intermediate volcanic provenances similar to those drained by the Columbia River today.

Despite high measured permeabilities, breaching limits the reservoir potential of the sandstone. Mudstone units are immature to mature hydrocarbon source rocks. A potential for gas exists in the Cowlitz or its equivalent in the subsurface and in the Clifton channel sandstones projected into the offshore area.

Frontispiece: Nicolai Ridge rocks dipping to the northwest. Bend in the Columbia River is in part the result of normal faulting along the ridge. View is from the east.

THE GEOLOGY OF THE NICOLAI MOUNTAIN-GNAT CREEK AREA, CLATSOP COUNTY, NORTHWESTERN OREGON

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GEOLOGY OF THE NICOLAI MOUNTAIN - GNAT CREEK AREA, CLATSOP COUNTY, NORTHWESTERN OREGON

INTRODUCTION

Purpose of Investigation

The geology of northwest Oregon has, until recently, only been described on a reconnaissance scale. Since 1973 graduate students at Oregon State University, under the guidance of Dr. Alan Niem of the Geology Department, have been working on detailed geologic mapping, structural analysis, and paleoenvironmental interpretation of the Tertiary units in the northwest Oregon Coast Range. They have recognized a complex sequence of deltaic and deep marine sedimentary units which are interbedded with a series of extrusive and intrusive basalts in an arc-trench gap setting.

The purposes of this study in conjunction with the regional project are:

1) to map and describe the Tertiary sedimentary and volcanic rocks in the Nicolai Mountain-Gnat Creek area;

2) to interpret the paleoenvironment, stratigraphic relations, age, and provenances of each sedimentary unit defined;

3) to determine the stratigraphic relationship between the locally erupted and Columbia Plateau-derived basalts; and

4) to evaluate the economic potential of the area, in particular the petroleum potential of the sedimentary units.

The study of the Nicolai Mountain-Gnat Creek area proved interesting for four major reasons. First, a tentative relationship between the local and plateau-derived middle Miocene basalts is described. Secondly, a sandstone unit shown on the state geologic map of Oregon as a late Miocene-Pliocene (?), fluvial sequence (Wells and Peck, 1961) has been shown to be a middle Miocene fluvial to marine sandstone. Third, a tie between on the western and eastern flanks Tertiary stratigraphic sections of the northern Oregon Coast Range is developed. Finally, a basalt previously only postulated to be present in northwest Oregon was recognized along the northern border of the study area.

Location and Accessibility

The Nicolai Mountain-Gnat Creek area is 15 miles upstream from the mouth of the Columbia River in northeast Clatsop County. The villages of Svensen and Westport lie in the study area. The 70 square mile area is bordered by the Colum- bia River to the north and by the Clatsop-Columbia county line to the east (Fig. 1). Nicolai Mountain is situated in the extreme southern part and with an elevation of 929 meters above sea level it is the highest point in the study area.

Primary access into the thesis area is by U.S. Highway 30 and by a number of paved county roads (Plate I). Much of the area is forested and is owned by the state of Oregon and Boise Cascade Corporation. The logging industry has created an ex-

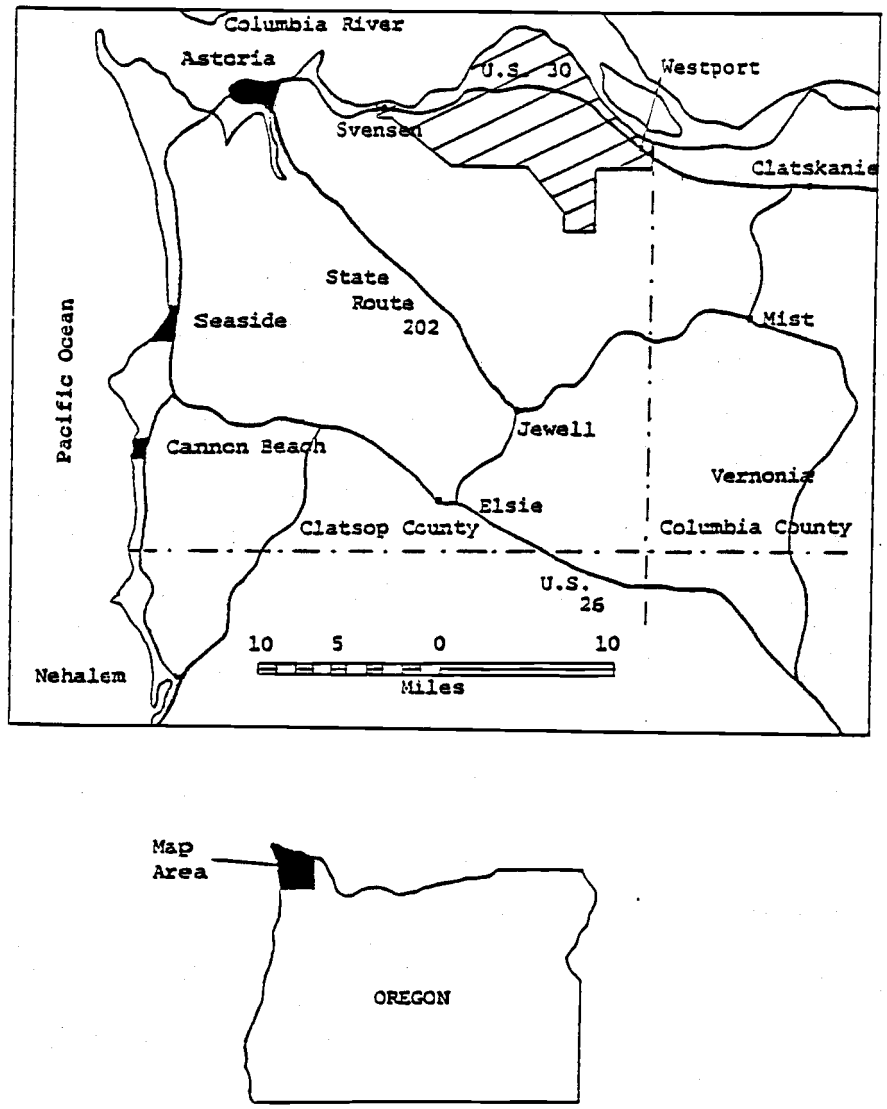


Figure 1 Index map showing location of thesis area.

tensive system of gravel roads which provide access into much of the forested land. Small streams provide access into the more remote regions of the thesis area.

The high rainfall (approximately 80 inches per year) and mild marine coastal climate are ideal for the dense coniferous forest which covers the area. As a result of this temperate climate and heavy undergrowth, the best exposures occur in road cuts, in stream cut valleys, on basaltic ridges, in gravel quarries, and along the scarps adjacent to the Columbia River. Big Creek Gorge with a vertical relief of 450 meters in a quarter mile provides natural cliff exposures of the basalt stratigraphy in the southwest part of the thesis area (Plate I). Nicolai Ridge near Bradley State Park along U.S. Highway 30 displays the plateau-derived basalt stratigraphy in the northeastern section of the study area.

Methods of Investigation

Field Methods

Four months were spent in the field at different times from October, 1978 to September, 1979. The majority of the time, two and one-half months, was during the summer of 1979. Field accomodations were provided through the generosity of Robert "Cupie" Ziak of Knappa, Oregon.

Mapping was to a scale of 1:18,000, utilizing enlarged parts of Svensen (1955) and Cathlamet (1953) U.S. Geo- logical Survey 15-minute quadrangle maps. Aerial photographs (scale 1:12000) taken in 1977 by the Oregon State Forest Ser- vice were used for location and structural interpretation.

Attitudes of rock strata, dikes, and directional features were obtained by the use of a Brunton compass. A Jacob's staff and Abney level were employed to measure five stratigraphic sections (Appendix I). A Geological Society of America Rock-Color Chart (1970), based on the Munsell system, was used in description of the units. Reineck and Singh's (1975) classification of bed thickness and cross-stratification was utilized. The basaltic units were described with the aid of the Fisher (1961) classification of volcaniclastic rocks and Swanson and Wright's (1978) description of basaltic flow charat-teristics. Field terminology of the basalt units was based on the work of Snavely and others (1973). This was later augmented by revision of the nomenclature of the Columbia River Basalt Group (Swanson and others, 1979).

Approximately 200 sedimentary rock samples were collected, from which critical examples were selected for further laboratory analysis. Included in these samples were 47 micro and macro fossil collections. Basalt samples were collected during the summer field season for chemical and paleomagnetic analyses in the laboratory.

Laboratory Methods

Laboratory studies were divided into three major phases: sediment analysis, preparation of fossils, and determination of the composition and magnetic polarity of the basalts.

Sediment analyses performed included mechanical sieve and hydrometer tests for grain size, thin section and grain mount petrography, and pebble counts. Work done by Tenneco Oil Company in conjunction with this study included porosity, permeability, and maturation tests of selected sam- ples from the sedimentary units (See Appendix IX).

A total of 30 mechanical sieve and 12 hydrometer analyses were performed, using standard statistical methods from Royse (1970), on 39 selected sandstone and mudstone samples from the sedimentary units in the study area (Appendix V). The statistical size parameters from these analyses were plotted on environmentally sensitive bivariant plots (see Grain Size Analysis Section).

The mineral composition of the sedimentary units was determined by studying 25 thin sections and eight heavy mineral grain mounts. Modal analyses were performed on 11 of the thin sec- tions. Tetrabromoethane (sp. gr. 2.94) was used to separate the heavy minerals from the 2.5 to 4.0 phi-size fractions for provenance determination. Six pebble counts of the con- glomeratic layers aided in the provenance study (see thin sec- tion Petrology Section). Paleocurrent azimuths obtained in the field were analyzed and plotted on rose diagrams using methods outlined by Royse (1970).

Warren Addicott and Ellen Moore of the U.S. Geological Survey identified and dated the molluscan assemblages collected during the field work. Eight foraminiferal assemblages were recovered from mudstone samples and sent to Weldon Rau of the Washington State Department of Natural Resources. The foram assemblages were extracted from the mudstones by soaking the samples in warm, dilute, hydrogen peroxide; by wet sieving through 1 phi intervals; by drying; and by hand picking the forams from the different size fractions under a binocular microscope.

Small samples (1-5 grams) of mudstone were broken down using warm, dilute hydrogen peroxide. Smear slides were made and studied at high magnification (450X) for diatom fragments. Samples with an abundance of siliceous material were sent to

John Barron of the U.S. Geological Survey for identification. Trace fossils found in the study area were sent to C. Kent Chamberlain of Cities Service for paleoenvironment interpretation. Fossil assemblages and locations are listed in Appendix IV.

Major oxide concentrations were determined for 37 basalt samples from the study area. Sample preparation followed Taylor's Cookbook for Standard Chemical Analysis (E.M. Taylor, Dept. of Geology, Oregon State Univ., in-house lab manual). The actual analyses were done by Ruth Lightfoot using X-ray fluorescence spectroscopy and atomic absorption spectorophotometry (See Appendix II).

Abundances of 17 trace elements were determined on 20 basalt samples by sequential instrumental neutron activiation analysis (INAA). Standard sample preparation utilizing an aluminum jaw crusher, and a ceramic pulverizer, and polyurethane vials decreased contamination of the samples. The actual activation and subsequent counting was done by Monte Smith under the guidance of Dr. Roman Schmitt at the Nuclear Radiation Laboratory on the campus of Oregon State University.

Remanant magnetism was determined for 82 oriented basalt samples obtained during the summer field season. A portable fluxgate magnetometer was used following the methods of Dr. E.M. Taylor (1979, personal communication) and Doell and Cox (1964).

Seventeen thin sections were studied to further verify the petrographic differences and similarities between the basalt units which were first recognized by Snavely and others (1965; 1973).

The earliest published studies on the Tertiary of the Pacific Northwest were conducted by Conrad and Dana in 1840 (Moore, 1963). Dana reported on the lithology and structure of the "Tertiary Formation" at the city of Astoria. Conrad described the molluscan fossils from the unit and assigned a Miocene age to them. The name Astoria Shale[1] was first applied to this unit in 1880 by Condon (Schenck, 1929). Diller embarked on a geologic reconnaissance of northwestern Oregon in 1896, but it appears from the traverses on his map that he did not enter the Nicolai Mountain-Gnat Creek area.

Early reconnaissance work within the Nicolai Mountain-Gnat Creek area was stimulated by regional studies that evaluated the petroleum potential of northwestern Oregon. In 1914, Washburne reported outcrops of "fresh water" sandstone near the town of Clifton. He postulated that these sandstones overlie both the Astoria Shale and the Columbia River Basalt. The first detailed stratigraphic investigation of the Astoria Formation was undertaken by Howe (1926) in the city of Astoria. He subdivided the formation into three members which he dated as late middle Miocene. The type section that he described unfortunately has been lost as a result of the to urbanization of the city. This has caused subsequent correlation and mapping problems for others working on Astoria-age strata elsewhere in Oregon (see Moore, 1963, for a compilation of the literature).

Warren, Norbisrath, and Grivetti (1945) published a regional geologic map of northwestern Oregon in which they were the first to differentiate the Tertiary sedimentary strata from

1 Technically this unit is not a shale, but is better defined as a mudstone.

the adjacent volcanics. Within this study area they defined four sedimentary units: Pittsburg Bluff Formation, "Beds of Blakely Age", the Astoria Formation and undifferentiated Pliocene sandstone. The geologic map of Oregon west of the 121st meridian was compiled by Wells and Peck in 1961. They essentially updated the geologic mapping of Warren and others (1945) in northwest Oregon. Engineering and environmental geology studies of the Tillamook and Clatsop counties were accomplished by Schlicker and others (1972) and Beaulieu (1973). These investigators discussed the regional geology, geologic hazards, and mineral resources of the area. Van Atta (1971) described the Tertiary units in the upper Nehalem River, in particular the Pittsburg Bluff and Cowlitz Formations.

Recently there have been a number of attempts to resolve the problems of correlation and mapping of the Astoria Formation based on the biostratigraphy of the molluscan and foraminiferal faunas (Lowry and Baldwin, 1952; Moore, 1963; Dodds, 1969). Many varied lithofacies of lower to middle Miocene strata have been mapped as the Astoria Formation in western Oregon. However, as Niem and Van Atta pointed out in 1973, there has been a lack of lithostratigraphic correlation to coincide with the fossil correlations previously developed. Since then, Dr. Niem and Oregon State University graduate students working under his supervision have been defining new type localities for these lithofacies within the Astoria Formation and have divided the unit into separate members. The most recent completed study is that of Cooper (1980). He has described the relationship of the lithofacies in the Astoria Formation within a deltaic-turbidite model. Other recent workers who have contributed to the understanding of the Astoria Formation and its relationship to the middle Miocene

basalts include: Cressy, (1974); Smith, (1974); Neel, (1976); Tolson (1976); Penoyer (1977); Nelson, (1978); and Coryell, (1978).

Wolfe and McKee (1972) mapped similar Tertiary eugeosynclinal sedimentary and basaltic units directly north of the thesis area across the Columbia River in the Grays River Quadrangle, Washington. They recognized an Oligocene open marine, sublittoral to bathyal mudstone unit (Lincoln Creek Formation) which is unconformably overlain by a middle Miocene nearshore sequence (Astoria Formation). Locally erupted middle Miocene to Pliocene(?) coastal basalts overlie and intrude both formations. Armentrout and others (1980) recently summarized the Cenozoic stratigraphy of southwest Washington. Mapping in southest Washington, directly north and northwest of this study area, is presently being undertaken by Ray Wells of the U.S. Geological Survey.

There has been substantial geologic debate as to the age, stratigraphic position, and depositional origin of the sandstone at Clifton and along Gnat Creek in the study area. It has been referred to as an upper unit of the Astoria Formation by Lowry and Baldwin (1952); as a late Miocene-Pliocene(?) fluviatile and lacustrine deposit by Wells and Peck (1961), by Bromery and Snavely (1964), and by Niem and Van Atta (1973); and as an estuarine deposit by Beaulieu (1973). This thesis is the first detailed study of this misunderstood unit.

Around Wickiup Mountain, directly southwest of the thesis area, Coryell (1978) mapped a feeder dike system with associated pillow lavas and submarine breccias of Cape Foulweather and Depoe Bay basalts. He suggested a local origin for these middle Miocene basalts and also reiterated the origin postulated by Snavely and Wagner (1963) and Snavely and others

(1973) of subaerial Columbia River flows from eastern Oregon reaching a marine embayment in this study area. Kienle (1970) measured a westerly flow direction for the subaerial columnar jointed Columbia River Basalt flows at Bradley State Park in the thesis area. This, along with the recent stratigraphic studies of the Columbia River Basalts in the western Oregon Cascades region (Anderson, 1978; Beeson and Moran, 1979), substantiates that at least some of the subaerial middle Miocene basalt originated from an eastern plateau source and flowed down an ancestral Columbia River Valley to the study area.

Swanson and others (1979) have proposed revisions in the stratigraphic nomenclature of the Columbia River Basalt Group. In that study and subsequent studies Swanson and Wright they outline an approach to differentiating the units within the Columbia River Basalts. Their method is adopted in this study. A more detailed discussion of the early work in the Columbia River Basalts will be found in the basalt stratigraphy sec- tion of this thesis.

BIOSTRATIGRAPHIC FRAMEWORK USED IN THIS STUDY

The geologic time table adopted for this thesis is from John Armentrout (1980, in preparation) (Fig. 2). Armentrout's time table will soon be published in a Geological Society of America Memoir under the title Cenozoic Geologic Timetable for Oregon and Washington. In that paper, Armentrout has revised the definition of the provincial biochronologies and has calibrated them by correlation to the world wide geologic time-scale. This provides a basis for time correlation in Oregon and Washington to the current worldwide time-scale.

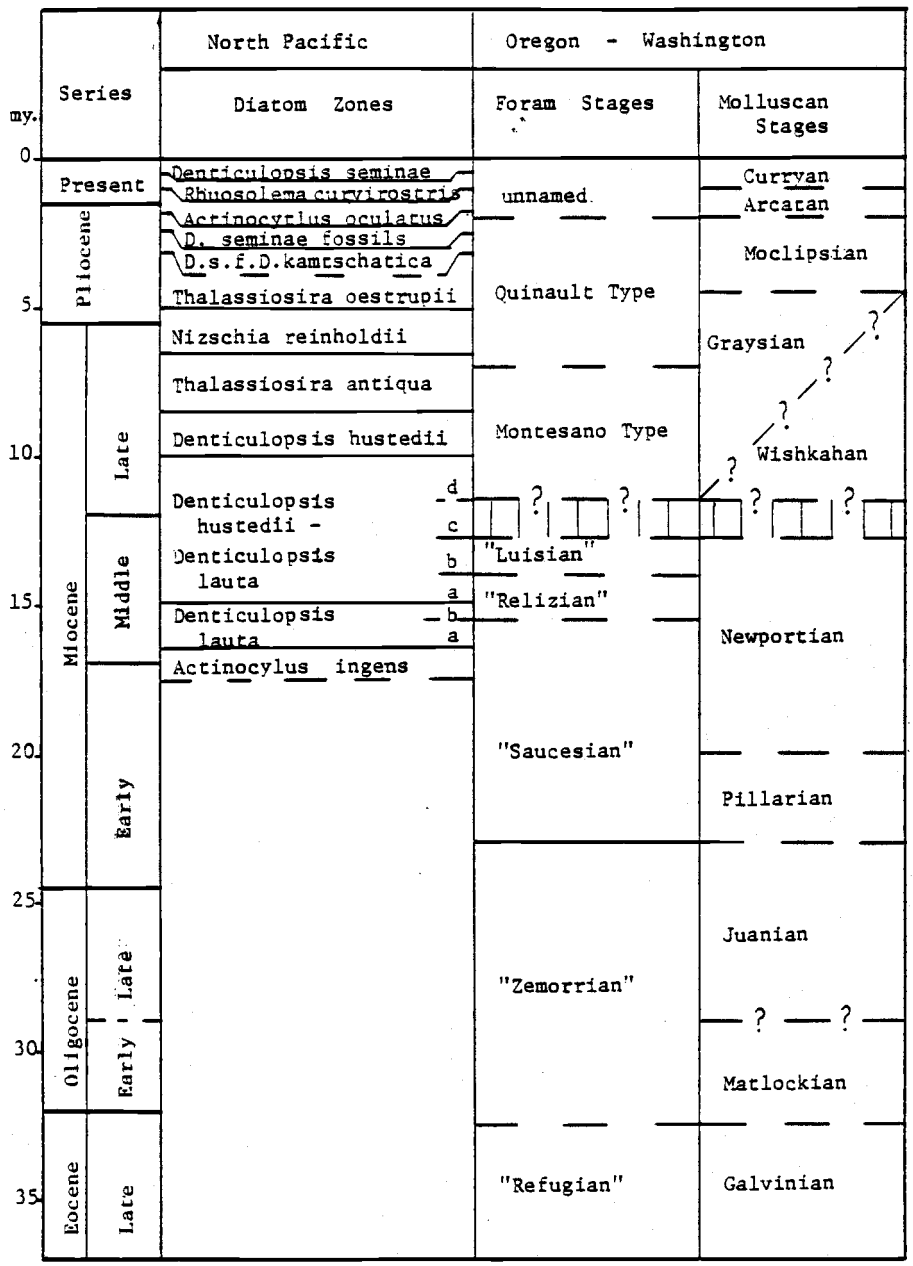


Figure 2 Correlation chart of biostratigraphic units (from Armentrout, 1980, in press).

Foraminiferal stages are those of Rau (1980). Molluscan stages follow Armentrout (1975, 1978), Addicott (1976), and Turner (1938). The diatom zones are those defined by Koisumi (1973) and modified to the planktonic foram stages by Barron (1980). K/Ar dates listed on Figures 4 and 5 have been recalibrated by Armentrout (1980) using the new decay and abundance constants of Steiger and Jager (1977). These are valid to within ± 0.5 m.y. in the Neogene and ± 1.0 m.y. in the Paleogene. The K/Ar dates quoted from other sources in the text have not been adjusted. Addition of approximately 0.5 m.y. to these dates will allow comparison with this new radiometric time scale.

REGIONAL GEOLOGY

In the Coast Range and on the adjacent inner continental shelf, over 7,000 meters of Cenozoic sedimentary and volcanic rocks accumulated in a marginal basin on a early to middle Eocene oceanic crust (Snavely and others, 1980). This sequence of strata is exposed in a northward plunging anticlinorium which comprises the northern Oregon Coast Range (Fig. 3). Uplift of the anticlinorium is thought to have occurred in late middle Miocene (Baldwin, 1976). Formation of the anticlinorium is related to periodic underthrusting by the eastward moving Farallon Plate (Snavely and others, 1980a), and the structure is outlined by a series of north and northwest trending open folds. These compressional folds are cut by northeast- and northwest trending normal and reverse faults. (Snavely and Wagner, 1964; Niem and Van Atta, 1973).

The central basaltic core of the structure is composed of lower to middle Eocene Tillamook Volcanics (Baldwin, 1976). An off lapping sequence of middle Eocene to middle Miocene sedi-

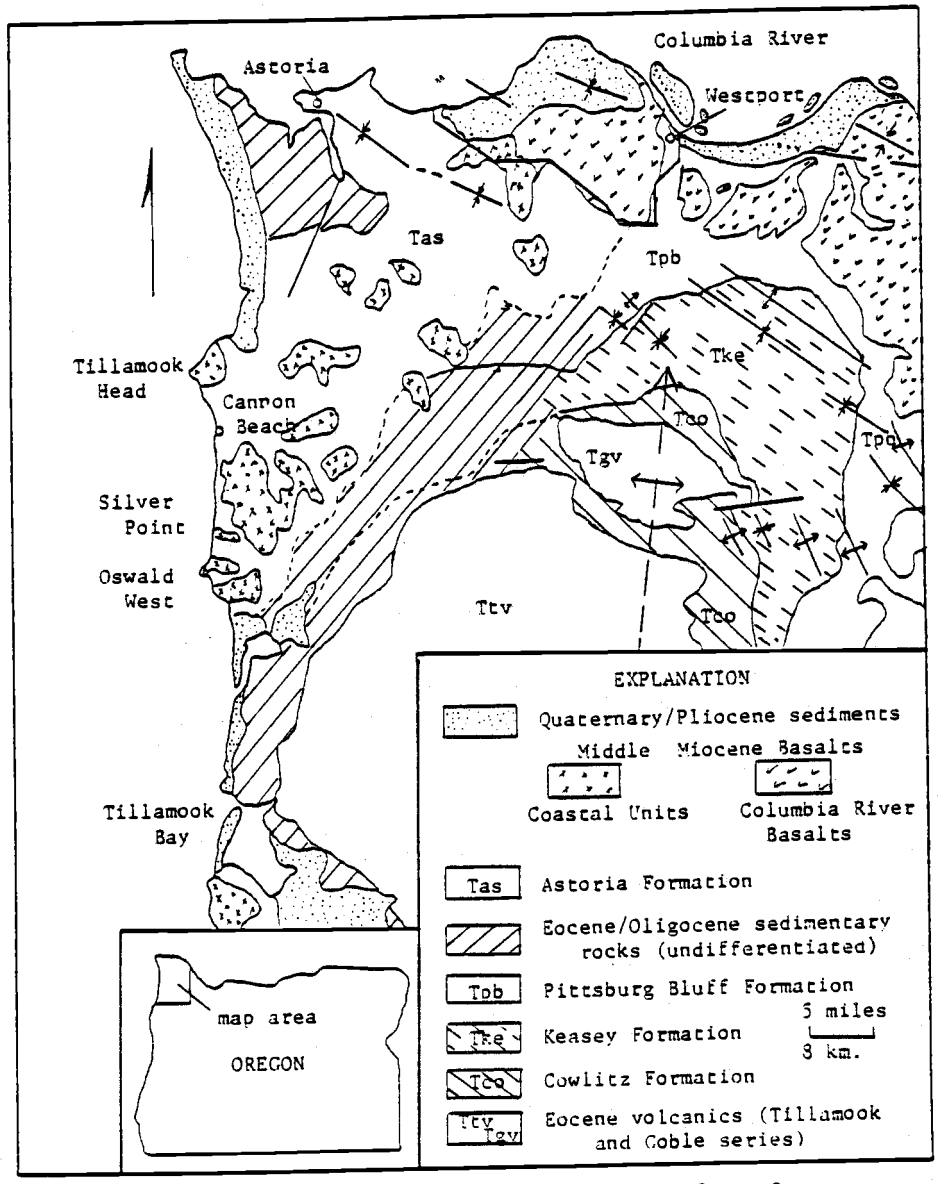
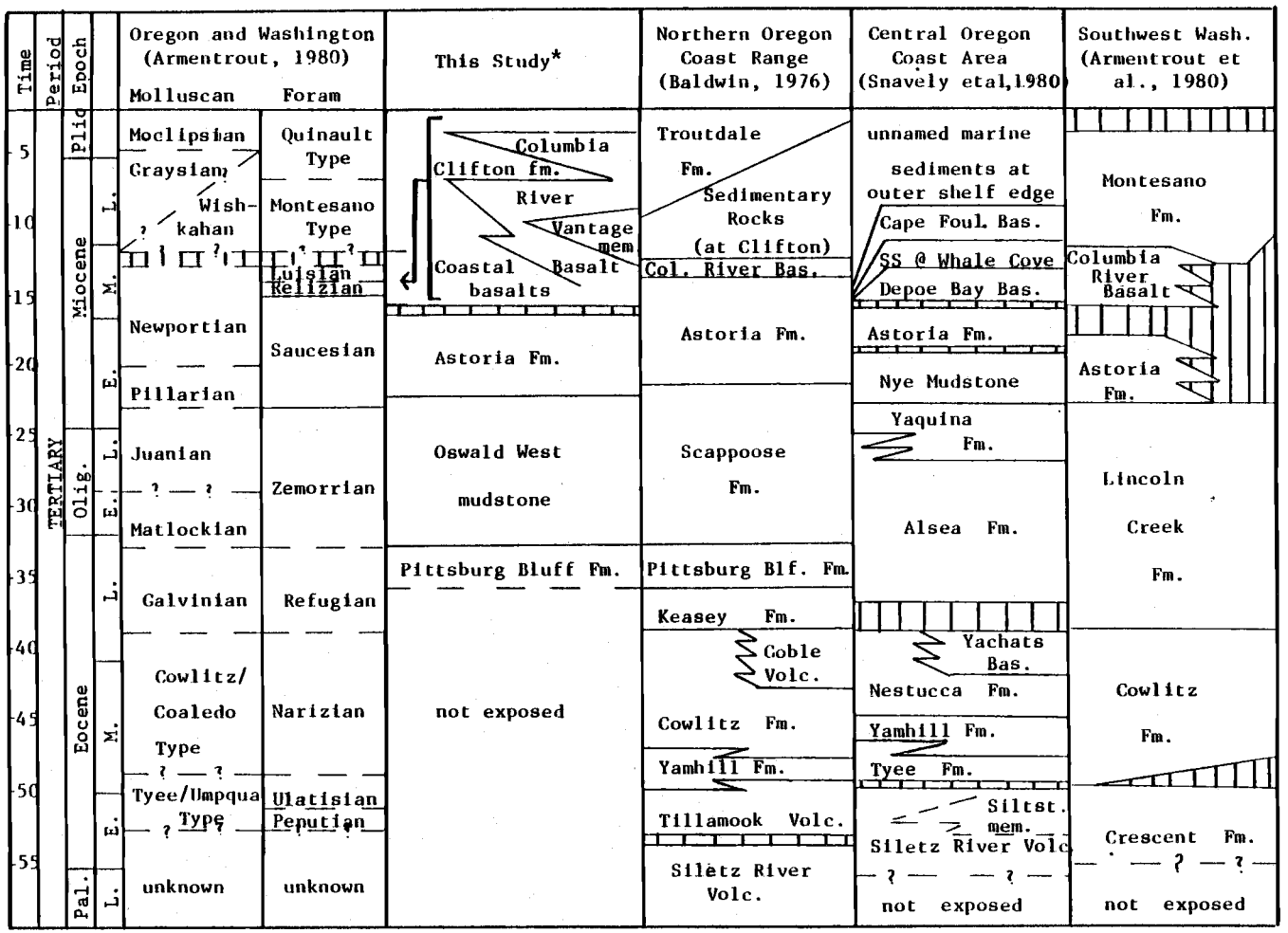


Figure 3 Regional geologic map of northwest Oregon Coast Range.
(modified from Wells and Peck, 1961)

mentary and volcanic strata form the homoclinal flanks of the anticlinorium. The stratigraphy of the northern and central Oregon Coast Range and of the related southwest Washington Coast Range is shown on Figure 4.

The Tillamook Volcanics are composed of oceanic tholeiitic pillow basalts, basaltic breccias, and minor tuffs. The estimated thickness of 6,000 meters for this unit includes interbedded undifferentiated tuffaceous siltstones and graded volcanic feldspathic wackes (Snavely and Wagner, 1964). The Tillamook Volcanics correlated with the Siletz River Volcanics which form the basement rocks of the central Coast Range (Snavely and Baldwin, 1948). The Crescent Formation of southwest Washington, composed primarily of tholeiitic pillow basalts, is the chemical and age equivalent of the Siletz River Volcanics (Wolfe and McKee, 1972) and of the early to middle Eocene Tillamook Volcanics (Fig. 4). Five pre-Columbia River Basalt sedimentary and volcanic units which overlie the Tillamook Volcanics have been defined by Warren and Norbisrath (1946), on the northeastern flank of the Coast Range. They are: the Cowlitz Formation, the Goble Volcanics, the Keasey Formation, the Pittsburg Bluff Formation and the Scappoose Formation (Fig. 4). Van Atta (1971) suggested enclosing these sedimentary units within the Nehalem Group.

The oldest sedimentary unit is the middle to upper Eocene Cowlitz Formation. The 300-meter thick unit is composed of arkosic sandstone, siltstone, and tuffaceous mudstone with a few beds of basaltic conglomerates (Niem and Van Atta, 1973). Recent discoveries of commercial quantities of natural gas from the Cowlitz Formation near Mist, Oregon have greatly increased interest in the aereal extent, environment of deposition and burial history of this unit. The Cowlitz Formation dips northward under younger Tertiary units from the central Tillamook



*see figure 5 for detailed stratigraphy

Figure 4 Correlation of Coast Range Stratigraphy for Oregon and southwest Washington.

Volcanic anticlinal core. This suggests that the unit is present in the subsurface of the study area. Across the Columbia River in southwest Washington the "Cowlitz Formation" is estimated to be 3,000 meters thick (Armentrout and others 1980). It consists of three distinct units mapped by Snavely and others, (1958), as the McIntosh, Skookumchuck and Northcraft Formations (Armentrout and others, 1980).

The Goble Volcanics which interfinger with the upper part of the Cowlitz Formation are composed of subaerial basalt flows, pillow basalts, flow breccias and interbedded sedimentary rocks. This unit crops out in the northeastern part of the Oregon Coast Range and in southwestern Washington (Niem and Van Atta, 1973). The overlying upper Eocene Keasey Formation is approximately 500 meters thick (Van Atta, 1971). The unit has been divided into three members: a basal pebbly volcanic sandstone with gray tuffaceous mudstone; a structureless, fossiliferous, tuffaceous siltstone; and a thinner concretionary tuffaceous siltstone (Warren and Norbisrath, 1946). The fossiliferous nature of the middle member characterizes the Keasey Formation. Rare Tertiary crinoids have been discovered within this member near Mist, Oregon (Baldwin, 1976).

The upper Eocene Pittsburg Bluff Formation conformably overlies the Keasey Formation in the Nehalem River Valley (Niem and Van Atta, 1973). The 260-meter thick unit is composed of arkosic, tuffaceous sandstone and siltstone. The Pittsburg Bluff Formation has been mapped into this thesis area (Warren and others, 1945; this study) where an estimated 60 meters occurs in outcrop. This unit is equivalent to the lower part of the Lincoln Creek Formation of southwest Washington (Fig. 4).

The next youngest unit is the 450-meter thick Scappoose Formation which disconformably overlies the Pittsburg Bluff

Formation (Niem and Van Atta, 1973). This Oligocene to lower Miocene unit is composed of fossiliferous, micaceous, arkosic, and gray tuffaceous mudstone (Warren and Norbisrath, 1946). The Scappoose Formation is, in part, equivalent to the Oswald West mudstone (Cressy, 1974), to the Lincoln Creek Formation, and to the Alsea Formation. The upper part of the unit may be a stratigraphic equivalent to the lower part of the Astoria Formation (Van Atta, 1971; this study).

The pre-coastal basalt sedimentary units in the northwestern Coast Range consist of undifferentiated Eocene to Oligocene sedimentary rocks and the Astoria Formation (Wells and Peck, 1961). More recently, Cressy (1974), Smith (1975) and Neel (1976) defined an upper Eocene to upper Oligocene Oswald West mudstone (informal) as a unit within the previously undifferentiated sedimentary rocks of Wells and Peck (1961). This unit is composed of deep marine interbedded silty claystone and tuffaceous siltstone (Niem and Van Atta, 1973). The Oswald West mudstone has been mapped from its type locality at Short Sands Beach in Oswald West State Park into this thesis area, where it unconformably overlies the Pittsburg Bluff Formation (see the Oswald West section in this study for references).

The lower to middle Miocene Astoria Formation, first named by Howe (1926) for sandstones and shales exposed in the town of Astoria, occurs in three structural embayments extending from Astoria to Newport, Oregon. Several informal members in this 1,300-meter thick unit have been named and described by graduate students from Oregon State University (see Astoria Fm., this study). Recognized facies in the Astoria Formation include: fluvial deltaic conglomeratic sandstone (Angora Peak member); deep-water turbidite sandstone and mudstone (lower Silver Point member); hemipelagic mudstone (upper Silver Point

member); deep-water channelized sandstone (Pipeline member); and beach and shelf sandstone (Big Creek member) (Nelson, 1978). In southwest Washington, the Astoria Formation is estimated to be 1,400 meters thick (Wolfe and McKee, 1972).

During the middle Miocene there was a great outpouring of tholeiitic basalt in the Pacific Northwest. These basaltic flows unconformably overlie the upper Eocene to middle Miocene sedimentary strata on both the east and west flanks of the northern Oregon Coast Range (Fig. 4). Middle Miocene flow basalts in the lower Columbia River basin and in the Willamette Valley are considered part of the plateau-derived Columbia River Basalt Group of eastern Oregon and Washington (Swanson and others, 1979). The rock units recognized in this study are part of the Yakima Basalt Subgroup. They consist of the Grande Ronde Basalt, the Frenchman Springs and Priest Rapids members of the Wanapum Basalt (Beeson and Moran, 1979), and the Pomona Member of the Saddle Mountains Basalt (Schmincke, 1967). These formations are primarily composed of subaerial flows which originated on the Columbia Plateau and travelled down an ancestral Columbia River Valley, entering the sea in the vicinity of Astoria and the Columbia River embayment (Snavely and others, 1973). Interbedded between the Grande Ronde and Wanapum Basalt Formations is the arkosic sandstone Vantage Member of the Ellensburg Formation of eastern Washington, which has been tentatively correlated to the Bradley State Park area in this study (See basalt interbeds section in this study).

Along the western flank of the northern Oregon and southern Washington Coast Ranges are three chemical-petrologic basalt units that are considered consanguineous with the Columbia River Basalts. These units are the Depoe Bay Basalt, the prophyritic Cape Foulweather Basalt, and the basalt at Pack Sack Lookout, Washington. Snavely and others (1973) are of the

opinion that these basalt flows, breccias, pillow flows, and associated intrusions were derived from local vents. Associated with these flows in northwest Oregon are deep-water mudstone interbeds (Neel, 1976; Penoyer, 1977; Coryell, 1978). Depoe Bay and Cape Foulweather units have been mapped very close to the area where the consanguineous subaerial Columbia River Basalts entered the sea (Fig. 12) (Coryell, 1978). See the Basalt Stratigraphy section of this thesis for further discussion of the relationship between the Columbia River Basalts and locally derived coastal basalts.

A sequence of arkosic sandstone and tuffaceous siltstone exposed along the Columbia River near Clifton, Oregon has been referred to as "sedimentary rock (at Clifton)" (Baldwin, 1976; Niema and Van Atta, 1973); as "post-Astoria sandstone" (Dodds, 1969), and originally as "Freshwater sands at Clifton" (Washburne, 1914). In this study the unit is informally called the Clifton formation. It was previously thought to be middle Miocene to Pliocene (Niem and Van Atta, 1973). Marine diatoms found during the course of this tudy in the siltstone facies indicate a late middle Miocene age for this unit.

The lower Pliocene Troutdale Formation is exposed along both sides of the Columbia River from Portland to Astoria (Lowry and Baldwin, 1952; Schlicker and others, 1972). This unit is composed of quartzite and basalt conglomerate with well rounded clasts of sandstone and mudstone that indicate a fluvial origin (Armentrout and others, 1980).

DESCRIPTIVE GEOLOGY

Introduction

This study area is situated in the northern Oregon Coast Range near the axis of a northward plunging anticlinorium. The area is unique because it is one of the few places where the interfingering relationship between the consanguineous Columbia River and coastal basalts can be studied. This area is also important because the sedimentary rock units bridge the northeastern and northwestern lithostratigraphies of the northern Oregon Coast Range.

There are twelve major rock units recognized in this study area. Figure 5 illustrates the stratigraphic relationships of these units. The Pittsburg Bluff Formation, first recognized in this area by Warren and Norbisrath (1945), is the oldest unit. Overlying the Pittsburg Bluff is the upper Oli- gocene Oswald West mudstone. Members recognized of the lower to middle Miocene Astoria Formation are the Big Creek sandstone, the upper Silver Point mudstone, and the Pipeline mudstone. These members and the Oswald West mudstone have been informally named and described to the west of this study by previous graduate students from Oregon State University (Cressy, 1974; Nelson, 1978; Coryell, 1978; and Cooper, 1980).

The basalt units described in this study include the plateau-derived Grande Ronde Basalt, the Frenchman Springs Member of the Wanapum Basalt, and the Pomona Member of the Saddle Mountains Basalt. These units are part of the Yakima Basalt Subgroup of the Columbia River Basalt Group. Also present are the locally derived Depoe Bay and Cape Foulweather basalts described by Snavely and others (1973).

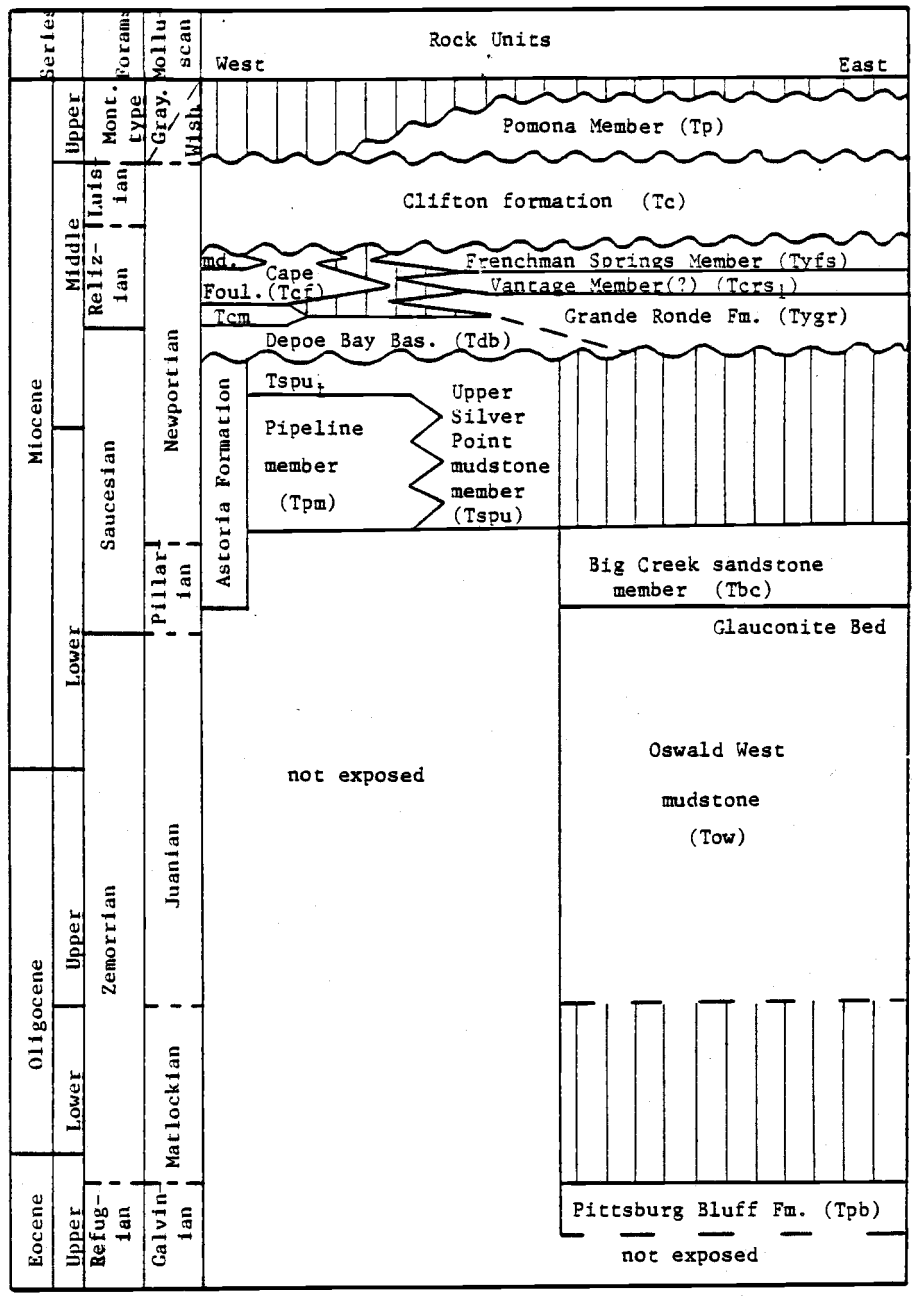


Figure 5 Tertiary stratigraphy of study area.

Major sedimentary units interbedded within the basalts that have been recognized in this study include a possible equivalent to the Vantage member of the Ellensburg Formation and the Clifton formation. The Vantage member, first described by Mackin (1961) in south-central Washington, lies stratigraphically between Grande Ronde and Wanapum Basalts. The Clifton formation, prior to this study, had been referred to as a non-marine sandstone (at Clifton) by Niem and Van Atta (1973) and Baldwin (1976). It overlies the Frenchman Springs Member of the Wanapum Basalt Formation and is overlain by the Pomona Member of the Saddle Mountains Basalt Formation.

Pittsburg Bluff Formation

Nomenclature and Distribution

The name Pittsburg Bluff Formation (Tpd) is adopted here for an approximately 100-meter thick sequence of upper Eocene tuffaceous sandstone exposed in the extreme eastern part of the study area (Plate I). This unit has been studied by various authors. Hertlein and Crickmay (1925) were the first to use the name "Pittsburg Bluff" for Oligocene strata of northwest Oregon. Weaver (1937) suggested the term "Pittsburg Bluff Formation" and named the type area as the bluffs along the Nehalem River between Pittsburg and Mist, Oregon. Warren and others (1945) mapped the aereal extent, divided the unit into two members, and estimated its total thickness to be 260 meters. They described molluscan fossils equivalent to the Pittsburg Bluff fauna along old U.S. Highway 30 at the eastern border of Clatsop County in this study area (M6, Sec. 1,T7N, R6W on Plate I). For an in-depth review of the literature pertaining to the early work in the Pittsburg Bluff Formation see Moore (1976).

Exposures of the Pittsburg Bluff are confined mainly to the upper Nehalem River Valley. Warren and Norbisrath (1946) also described the unit along State Highway 202 near Jewell, Oregon to the south and as mentioned, along the eastern border of this study area. A sandstone similar to the Pittsburg Bluff has also been recognized on the Washington side of the Columbia River north of the thesis area (Ray Wells, 1980, personal communication).

This writer measured and described a section along old U.S. Highway 30 where the unit is best exposed (see Appendix I). An old logging tunnel cut into the sandstone bluff defines the beginning of the measured section. The Pittsburg Bluff Formation is also exposed in stream cuts along Plympton Creek (OC 720 and 721, sec 36, T8N, R6W), up Westport School Road (OC 240, sec 36, T8N, R6W), and in West Creek (OC 51 & 52, sec 1 T7N, R6W).

Lithology and Sedimentary Structures

The Pittsburg Bluff Formation in the study area is characterized by a basal very fine-grained sandstone member and an upper silty sandstone to sandy siltstone member (informal). Associated with these units are rare thin, pebbly coarse-grained sandstone layers and tuffaceous mudstone interbeds. The lower sandstone member forms resistant bluffs.

The basal member, comprising up to 60% of the unit, is primarily composed of a finely laminated bluish gray (5B 7/1), very fine-grained, texturally immature sandstone. The member is well exposed along old U.S. Highway 30 east of Westport (Appendix I). It is moderately to well sorted and displays a positively skewed, and leptokurtic grain size distribution (see Size Analysis section). The 1.7-meter thick pebbly to very

coarse-grained sandstone interbedded in this unit is reddish brown (10R 4/6) and contains subangular to subrounded clasts. It is poorly sorted, very positively skewed and leptokurtic. This friable, extensively bioturbated sand- stone is cemented by hematite which produces the characteristic reddish color. This pebbly coarse-grained sandstone is a dis- tinctive marker bed in the Pittsburg Bluff Formation and can be traced discontinuously from the bluffs along old U.S. Highway 30 (Appendix I) several miles to the south (Jeff Goalen, 1980, personal communication).

The contact between the pebbly coarse-grained sandstone and the very fine-grained sandstones is gradational over 0.5 meters. It consists of a bimodal sandstone produced from the mixing of the two sandstones due to bioturbation. Incomplete mixing is evident from the presence of oval shaped burrows or pods (1-3 cm in diameter) of the pebbly very coarse-grained sandstone within the very fine-grained unit.

Sedimentary structures in this member consist principally of bioturbated layers, abundant laminations, and rare very faint low angle ($<10^{o}$) planar cross laminations. Lenses of molluscan fossils, including Perse pittsburgensis Durham, and Spicula pittsburgensis Clark, are common in the very fine-grained sandstone and rare in the pebbly coarse-grained beds. Both disarticulated shells and shell fragments are found within these lenses. Individual articulated pelecypod shells also occur within the main body of the sandstone. Calcareous concretions are present, usually localized along particular bedding planes, and may be related to the concentration of fossil lenses. Mollusks identified from this member are listed as OC-46, OC-49, and OC-51 in Appendix IV.

Compositionally, the Pittsburg Bluff sandstone is a volcanic arenite with the coarser grained pebbly layers consisting principally of basalt, andesite and pumice fragments. The very fine-grained sandstone contains these same rock types as well as an abundance of volcanic glass fragments quartz and feldspar (see Petrology Section).

The upper member of the Pittsburg Bluff Formation consists of light gray (N7) to pale yellow orange (10 YR 6/4), bioturbated, tuffaceous, silty sandstone. Interbedded within this sandstone are thin, very light gray (N8), well indurated, tuffaceous mudstone layers. This member is approximately 20 meters thick and is exposed west of the tunnel along old U.S. Highway 30, east of Westport (Appendix I). Bedding ranges from thick to thin in the silty sandstone whereas the interbedded mudstone is very thinly laminated and cross laminated. The unit is characterized by a blocky weathering pattern in comparison with the overlying Oswald West mudstone which forms "chippy" talus slopes. Bioturbation is the most conspicuous sedimentary structure in the upper silty sandstone member. The trace fossil, Scalarituba, was recognized in the mudstones. A Planolites (?) trace fossil also was identified by Kent Chamberlain (1979, written communication) within the silty sandstone. Mollusks, including Macoma pittsburgensis (Clark), were identified by Addicott (1979, written communi- cation) from this upper member and are listed as OC-50 in Appendix IV.

The contact between members in the Pittsburg Bluff Formation in this study is marked by a sharp lithologic change between the pebbly very coarse sandstone of the basal member and a laminated mudstone layer of the upper member. This break probably represents a diastem within the for- mation, but

molluscan fossil control on either side indicates the same late Eocene Galvinian age, suggesting that no major hiatus occurred at this lithologic break (Fig. 6).

Contact Relations

The lower contact of the Pittsburg Bluff Formation is not exposed in the study area. However, at the type locality, near Pittsburg, Oregon, the Pittsburg Bluff conformably overlies the lower upper Eocene Keasey Formation (Moore, 1976), and is conformably overlain by the Scappoose Formation (Van Atta, 1971).

An apparent disconformable upper contact of the Pittsburg Bluff with the overlying upper Oligocene to lower Miocene Oswald West mudstone can be inferred at three localities in this study area based on lithologic changes and age differences. A typical very fine-grained sandstone of the Pittsburg Bluff basal member is exposed on the west fork of the Westport School Road (OC 240, Sec. 36, T8N, R7W). Approximately 40 meters up section, mollusks collected from the overlying Oswald West mudstone were dated as late Oligocene to early Miocene (OC 241; see Appendix IV) (Moore, 1980, written communication). Early Oligocene fauna appear to be missing between the two formations. This same relationship occurs in Plympton Creek, with the Oswald West mudstone stratigraphically above the upper member of the Pittsburg Bluff (Plate I). In West Creek this relationship is seen between the upper member of the Pittsburg Bluff (OC-732) and the overlying Oswald West (OC 730 and 731). Late Eocene ages have been reported for the Oswald West mudstone to the southwest by Neel (1976), Penoyer (1977), and Nelson (1978). This would suggest that the Pittsburg Bluff and Oswald West are, in part, time equivalent and conformable to the west. The lateral facies relationship between the two units also suggests a gradational contact. The

apparent unconformity based on difference in fossil ages observed in this study may be a function of incomplete fossil control. The Pittsburg Bluff is also thought to be in fault contact with the Big Creek member of the Astoria Formation in Plympton Creek (OC 724, Sec. 2, T7N, R6W, Plate I).

Age and Correlation

The Pittsburg Bluff Formation is the oldest unit exposed in the Nicolai Mountain-Gnat Creek area. It is dated as late Eocene, equivalent to the upper part of the Galvinian Stage of Armentrout (1975) (Fig. 5). This equates to the Lincoln Stage of Weaver and others (1944). The basis for this date is a series of molluscan assemblages identified by Warren Addicott (1979) and Ellen Moore (1980) of the United States Geological Survey (see Appendix IV for complete listing and fossil localities).

The correlation of the sandstone from this study area to the type area near Pittsburg, Oregon (Weaver, 1937) is based on similarities in age, fossils and lithologies. Three independent sources have identified fossils from the Pittsburg Bluff Formation of this study area. Warren and others (1945) reported Pittsburg Bluff type fauna from along Highway 30 near the Clatsop County line. Addicott (1979, written communication) identified four collections, reaffirming the age and correlation of Warren and others (1945). Moore (1980, written communication) referred to the collection she reviewed as "typical of Pittsburg Bluff Formation". Plate I shows the close proximity of the sample localities of this study (OC 46, 49, 50) to those of Warren and others (1945) (M6, Sec. 36, T8N, R6W). Lithologically, the Pittsburg Bluff in both areas is characterized by a lower member of fine-grained, arkosic to lithic, tuffaceous marine sandstone with concretionary layers,

mollusk fossil lenses, and coarse-grained interbeds. The overlying member in both areas is a fine-grained tuffaceous mudstone. These members are referred to as the lower and middle members by Van Atta (1971). See Warren and Norbisrath (1946) for a description of the units at the type locality.

The age and stratigraphic position of this unit suggest that it is, in part, correlative to the deep water mudstone mem- ber of the Oswald West mudstone mapped to the west and south- west of this study by Penoyer (1977) and Nelson (1978). Apparently, the Pittsburg Bluff sandstone must lens out into deep-water upper Eocene mudstone to the west and southwest. Other units in Oregon thought to be correlative in part or entirely with the Pittsburg Bluff are the Tunnel Point Sandstone at Coos Bay, the upper part of the Eugene Formation in the Willamette Valley (Baldwin, 1976), and the Alsea Formation near Newport, Oregon (Snavely and others, 1969).The Pittsburg Bluff is also considered to be coeval with the lower part of the Lincoln Creek Formation of south- western Washington (Moore, 1976; Fig. 4).

Depositional Environment

The Pittsburg Bluff Formation in the study area was deposited at the inner to middle continental shelf depth in an open marine environment with a fluctuating energy regime. This is based on fossil ecology, textural aspects of the various lithologies, and the sedimentary structures present. A comparison can be made between the sediment facies described by Kulm and others (1975) on the present day Oregon continental shelf and the members of this unit.

The lower member consists of very fine-grained sandstone with parallel to low-angle cross-laminations and fossil shell lenses. The molluscan fossils indicate sublittoral water depths of 20 to 50 meters (Moore, 1980, written communication). The "CM" grain size graph of Passega (1957), Friedman's (1961, 1962) statistical grain size graphs and overall fine grain-size of the unit indicate deposition from low energy tractive currents (see Grain Size Analysis Section). The paucity of well developed cross-bedding, the water depth indicated by the fossil assemblages, and the low energy tractive current regime indicated by grain size analyses suggest that deposition occurred below wave base on the inner to middle continental shelf. The mixture of articulated mollusk shells and broken shell fragments within lenses, and the rare low-angle cross-laminations, may be the result of storm activity lowering the wave base and increasing the energy input into the depositional regime. Moore (1976) suggested a similar interpretation for the"mixed" shell lenses in the Pittsburg Bluff in the Nehalem River Valley.

The pebbly very coarse-grained sandstone layers within the lower member are interpreted as high energy storm deposits. The lower contact of this unit show increased bioturbation with incomplete mixing of the sand sizes. The burrows may reflect escape structures (Howard, 1975) of the molluscan or soft bodied infauna, and were preserved as a result of rapid deposition of the overlying very coarse-grained sandstone.

The basal member of the Pittsburg Bluff is interpreted to be an upper Eocene sublittoral, inner continental shelf sand lithologically similar to the inner continental shelf facies off the Oregon as described by Kulm and others (1975). The present inner shelf facies, like the Pitts- burg Bluff Formation, is characterized by fine-grained sand with parallel

laminations. It is formed in water depths ranging from the shoreline to 50-100 meters depending upon the proximity to the sediment supply source (e.g. river mouth). The molluscan fauna in the Pittsburg Bluff Formation suggest water depths of 20 to 50 meters (Moore, 1980, written communication). Rare coarse-grained to pebbly sand layers accompany these sands and are attributed by Kulm and others (1975) to storm deposits.

The upper member of the Pittsburg Bluff Formation is thought to have been deposited in quieter, somewhat deeper water. The silty constituents of the upper very fine-grained sandstone, the interbeds of laminated tuffaceous mudstone, the extensive bioturbation, and trace fossils support this conclusion. The influx of silt and clay deposition suggests diminishing current energy and turbulence in a low-energy environment. Very fine sand beds transported by weak tractive currents, alternating with the silt beds are deposited by low-density turbid currents (Kulm and others, 1975), indicates fluctuations in energy input within the low-energy environment. The thorough bioturbation, caused by the molluscan and/or soft bodied infauna, has destroyed the bedding and any primary sedimentary structures. Extensive bioturbation suggests a low sedimentation rate, allowing the infauna to homogenize the sediments (Howard, 1975). Scalarituba trace fossil recognized in the mudstone interbeds are indicative of middle to outer continental shelf environments (Chamberlain, 1978).

The mixed sand and mud facies described by Kulm and others (1975) on the continental shelf off Oregon is a possible model for the upper member of the Pittsburg Bluff. This bioturbated facies, are usually present seaward of the laminated inner shelf sand facies on the outer part of the inner to middle shelf but can be extended as far as the outer shelf region. It

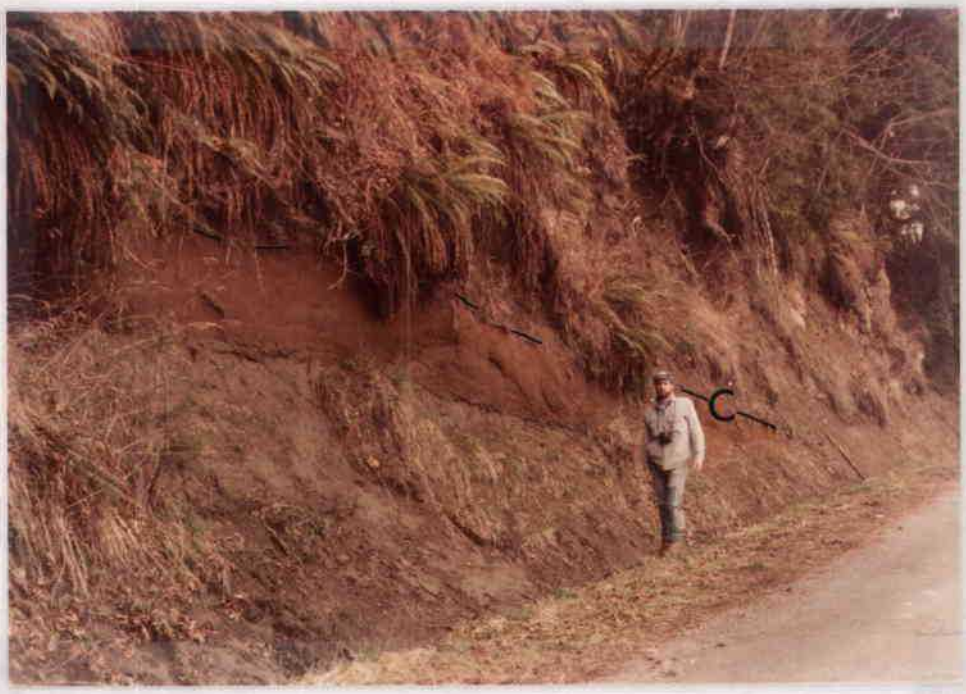


Figure 6 Contact between lower and upper member (informal) of Pittsburg Bluff Fm. It is marked by a reddish pebbly very coarse-grained sandstone layer and overlying mudstone. Exposure is along old Highway 30 within reference section a-a' (OC 49 and 50).

Figure 7 Typical exposure of Oswald West mudstone along logging road (OC 411, Sec. 20, T7N, R6W). Note hat for scale.

is characterized by highly bioturbated silty sands with rare disarticulated shell fragments.

The Pittsburg Bluff Formation in this study area shows a gradual increase in water depth. The high erosion rate of active volcanoes, indicated by the abundance of unaltered volcanic detritus in the fine-grained sandstone, coupled with this deepening depositional environment suggest local increased subsidence rate and/or sea level rise in late Eocene Pittsburg Bluff time. A eustatic sea level rise during the late Eocene is indicated by the global sea level cycles of Vail and Hardenbol (1979).

Oswald West Mudstone

Nomenclature and Distribution

The Oswald West mudstone (Tow) is named from a sequence of tuffaceous, clayey siltstones of Oligocene (Zemorrian) to early Miocene (Saucesian) age exposed in sea cliffs along Short Sands Beach in Oswald West State Park (Niem and Van Atta, 1973; Cressy, 1974). This unit has since been extended to include late Eocene (Refugian and Narizian) mudstone described to the west of this study area (Neel, 1976; Penoyer, 1977; Nelson, 1978). This name is adopted here for the clayey siltstones and interbedded sandy siltstone stratigraphically between the sandstone-rich underlying Pittsburg Bluff Formation and the overlying Astoria Formation (Figure 5).

The Oswald West mudstone has been mapped continously over 50 km. from the type locality to the southwestern border of this study area by Cressy (1974), Neel (1976), Tolson (1976), Penoyer (1977), Nelson (1978), and Coryell (1978). Two Oregon State University graduate students, Carolyn Peterson and Jeff

Goalen, have commenced thesis projects to the south of the study area with the goal of completing the mapping of this unit in Clatsop County. The Oswald West mudstone is present in the eastern part of this study area (Plate I).

Excellent exposures of Oswald West mudstone are present on steep stream banks in West Creek (OC 729 & 730, Sec. 1, T7N, R6W), but access to this area is difficult (Plate I). Typical exposures of the unit occur along the power line road (OC 60, Sec. 36, T8N, R6W); up Westport School Road (OC 241, 242 Sec. 36, T8N, R6W); and along Kerry Road, south of Nicolai Mt. (OC 298, Sec. 21 and OC 698, Sec. 28, T7N, R6W). It forms rounded forested hills of low to moderate relief. The hills are usually capped by the lower to middle Miocene Astoria Formation or the middle Miocene Grande Ronde Basalt.

The unit is estimated to be 125 to 175 meters thick in the study area. This is based on outcrop pattern and regional dip. Thickening to the west, the Oswald West mudstone is reported to be 370 meters in the Wickiup Mountain area (Coryell, 1978) and 600 to 900 meters near Astoria (Nelson, 1978). Tolson (1976) reported a maximum thickness for this unit of 1,500 meters based on information obtained from well logs of the Standard #1 Hoaglund well drilled near the coast (Sec. 11, T7N, R10W).

Lithology and Sedimentary Structures

The Oswald West mudstone is composed of clayey siltstone with interbedded sandy siltstone, glauconitic sandstone, and rare pebbly sandstone.

The dominant lithology is medium gray (N5) to olive gray (5Y 4/1), structureless, slightly micaceous, tuffaceous, clayey

siltstone (see Grain Size Analysis Section). Typically, the exposures of this unit are iron-stained and weathered to moderate yellowish brown (10YR 5/4). Conchoidal fractures are characteristic in exposures. This weathering pattern produces blocks and angular chips (0.5 to 2 cm) that accumluate at the base of logging road exposures. The mottled structureless apppearance in outcrop reflects the extensive bioturbated nature of the mudstone. Not observed in this study but commonly found in the Oswald West mudstone, are Scalarituba burrows (Nelson, 1978; Coryell, 1978). Forams, small, thin-shelled articulated mollusks, and fish scales are locally abundant in this unit. Fossils identified, including Nuculana sp. and Callistra cathcartensis (Weaver), are listed in Appendix IV.

X-ray diffraction studies by Penoyer (1977) and Nelson (1978) revealed the clay minerals smectite, glauconite, chlorite intergrade, and possibly kaolinite in the Oswald West mudstone. It is probable that these also are the clay minerals in the Oswald West in this study area. Clayey siltstone form approx- imately 75% of the Oswald West mudstone.

The interbedded sandy siltstone is olive gray (5Y 4/1) and weathers to pale yellowish orange (10YR 8/2). These beds are moderately bioturbated to structureless and contain larger molluscan fossils, including Bathybenbix washingtonensis (?) and Acila gettysburgensis (?), than in the clayey siltstone (see OC-241 Appendix IV).

Road cut and streamcut exposures also tend to be more resis- tive than the clayey siltstone. Contacts between beds of clayey siltstone and sandy siltstone are gradational over 2 to 3 meters.

A 1 to 3 meter thick pebbly sandstone lens is poorly exposed on Westport School Road (OC242). This deeply weathered, iron-stained, friable sandstone contains subrounded to rounded pebbles in a medium-to fine-grained sandstone matrix. In the sandstone classification of Williams and others (1954), the unit is a feldspathic volcanic wacke (see Petrography Section). The composition of some of the pebbles was identified in thin section or with a binocular scope. Compositions include: basalt, andesite, dacitic pumice, granodiorite, chert, and quartzite. Minor components include quartz and plagioclase clasts (see Appendix VII for a listing of the percentages of pebble components). Penoyer (1977) recognized similar pebbly sandstone lenses in the upper Oswald West near Saddle Mountain in southwest Clatsop County. He indicated that the pebbles were composed predominantly of basic volcanics with subordinate metamorphic quartzite and chert.

A grayish green (10G 4/2) to greenish black (5G 2/1), poorly to moderately indurated, fine- to coarse-grained, glauconitic sandstone defines the boundary between the Oswald West mudstone and the Astoria Formation. The unit is exposed in the upper reaches of West Creek (OC 248, Sec. 1, T7N, R6W), and up the powerline road (OC60, Sec. 36, T8N, R6W). The layer decreases in thickness to the west, being approximately 10 meters thick at West Creek and only 3 meters at the powerline road (Plate I). In the upper 1-2 meters, the greens and grades upward from 95% to 20% glauconite. This may be due to mixing by bioturbation of the glauconite sandstone with fine-grained sandstone from the overlying Astoria Formation.

Contact Relations

The Oswald West mudstone rests with apparent unconformity upon the upper Eocene Pittsburg Bluff Formation in the study

area. The contact is not exposed but the unconformity can be inferred from age differences, outcrop patterns, and regional dips (see Plate I). A close approximation (within 40 m) can be observed up Westport School Road (OC 241 & 240) and in Plympton Creek (OC 720, SW, 36, T8N, R6W). (See Contact Relations of Pittsburg Bluff Formation for more complete discussion). One of the difficulties with the Oswald West mudstone is that the base of the unit has never been defined or recognized to the west or southwest (Neel, 1976; Nelson, 1978; Smith, 1975; Cressy, 1974; Coryell, 1978). It is suggested that at least in this study the mudstone overlies the upper Eocene Pittsburg Bluff Formation. This should define its base for the first time. This contact also forms the basis for a link between the eastern and western stratigraphies of the northern Oregon Coast Range.

The upper contact of the Oswald West mudstone is placed at the top of the glauconitic sandstone layer (OC 60 & 248). This bed separates a thick underlying Oswald West clayey mudstone from an overlying thick, clean, fine-grained arkosic Astoria sandstone. Glauconite characteristically marks a period of slow sedimentation (Pettijohn, 1975), and its presence is suggestive of a submarine disconformity (diastem) (Krumbein and Sloss, 1963). These factors, coupled with the abrupt lithologic change, dictated placing the contact here. The fine-grained sandstone is assigned to the overlying Big Creek member of the Astoria Formation. The contact is gradational over two meters; the ratio of glauconitc sandstone to arkosic sandstone decreases from 9:1 to 1:9. This gradation is attributed to reworking by burrowing organisms or to current reworking of the glauco- nitic sandstone during intial deposition of the Big Creek member of the Astoria Formation.

At Nicolai Mountain, in the southeastern part of the study area (Plate I), the Oswald West mudstone is overlain unconformably by the middle Miocene Grande Ronde Basalt and the Astoria Formation. A thick fine-grained arkosic sandstone bed, tentatively assigned to the Big Creek member of the Astoria Formation, is exposed discontinuously at the base of the pillow basalt sequence (OC 301, Sec. 21, T7N, R6W). The basalt also unconformably overlies sandstone equivalent in age to the Oswald West mudstone on Plympton Ridge (OC 312, Sec. 15, T7N, R6W) southeast of the study area.

Age and Correlation

The age of the Oswald West mudstone in this study area is Oligocene to early Miocene based on foram and mollusk assemblages collected from two localities. The forams occur in a clayey mudstone five meters below the Oswald West-Astoria Formation contact (OC60; Appendix IV). Rau (1980, written communication) assigned the forams to the Zemorrian or Saucesian Stage. Molluscan fauna collected 40 meters above the Pittsburg Bluff Formation (OC 241; Appendix Iv) are assigned to the Juanian Stage of Addicott (1976) which is equivalent to the Zemorrian Foraminifera Stage (Moore, 1980, written communication) (Fig. 2). Molluscan fossils collected along Plympton Ridge to the south of this study were also assigned to the Juanian Stage by Moore (OC 312, Sec. 15, T7N, R6W).) A complete listing of fossils collected is presented in Appendix IV.

Elsewhere the Oswald West mudstone ranges from late Eocene to early Miocene in age based on mapping completed to the west and southwest by Neel (1976), Penoyer (1977), and Nelson (1978). Considering that there may be up to 40 meters of Oswald West mudstone which lie stratigraphically below the

Juanian age molluscan fossil locality (OC241) reported in this study the known extent of the unit elsewhere in northwest Oregon and the late Eocene age for the underlying Pittsburg Bluff the base of the Oswald West in this study area may be older than Oligocene (Zemorrian). The unit appears to thicken and is, in part, a lateral deep-water facies equivalent of the upper Eocene Pittsburg Bluff Formation.

The Oswald West in this study appears to be at least partially correlative with the type locality at Short Sands Beach (Cressy, 1974). This is based on continuous mapping from the type locality to the eastern edge of this study area (Coryell, 1978). Close similarities in age (late Oligocene-early Miocene), stratigraphic position (below the Astoria Formation), and lithology (bioturbated silty mudstone and glauconite beds) substantiate this correlation. Based on the fossil age and lithology, the Oswald West mudstone in this study would equate to the upper Oswald West mudstone of Penoyer (1977) and Nelson (1978).

The Oswald West is, in part, age equivalent to the Scappoose Formation of the Nehalem River valley (Baldwin, 1976), to the upper part of the Lincoln Creek Formation of southwest Washington (Wolfe and McKee, 1972), and to the Yaquina and Nye Formations of the Newport embayment (Niem and Van Atta, 1973) (Fig. 4).

Depositional Environment

The Oswald West mudstone was deposited in a middle continental shelf to upper slope basin in a relatively constant, low energy environment. This interpretaton is based on the benthic foram paleoecology, lack of primary sedimentary structures, fine-grain size, presence of unbroken molluscan fossils, exten-

sive bioturbation, and thick glauconitic sandstone layer. The depositional environment of the Oswald West is comparable to the mud facies described by Kulm and others, (1975) on the middle to outer continental shelf off Oregon today.

The benthic forams collected from mudstone near the glauconitic sandstone (OC60) suggest cold water, open sea conditions at water depths ranging from outer shelf to upper slope (100-400m) (Rau, 1980, written communication). The abundance of chlorite and chlorite intergrades clay reported by Nelson (1978) in the Oswald West supports the interpretation of mid-latitude, cold water (Jacobs, 1970) paleoenvironment of the Oswald West mudstone. Glauconite forms as a marine authigenic mineral. Porrenga (1967) reported glauconite at depths of 1850 meters, however, it is found more often in shelf and upper slope water depths of 18 to 730 meters where slow sedimentation and slightly reducing conditions prevail (Pettijohn, 1975).

The silt and clay size material reflect weak current activity in a low energy environment. The extensive bioturbation would destroy any primary sedimentary structures. The bottom currents are normally not strong enough to offset the biological activity (Kulm and others, 1975). The sandy siltstone interbeds reflect increased sediment supply and stronger current activity or a slight shallowing of the basin. The presence of unbroken mollusk shells within these interbeds indicate that the environment was still fairly low energy. The clayey siltstones probably reflect hemipelagic deposition. The sandy siltstones with concentrations of forams and sand in irregular layers may be referred to as muddy contourites or turbidites (Stow, 1978). These are difficult to differentiate from hemipelagites but the presence of the fine sand (up to 15%) and forams in the coarser sandy siltstones is a distinguishing feature of contourites (Stow, 1978). They are the

result of slightly higher energy turbid layers which winnow the very fine clay and silt particles and concentrate the coarser grained very fine sand and silt (Stow, 1978). The sedimentary structures distinctive in muddy turbidites were not recognized. They may have been destoyed by the pervasive bioturbation in the Oswald West mudstone.

The pebbly sandstone lens found in this unit may reflect a small channel or sea gully which originated from the coeval Scappoose Delta(?) to the east. The transport mechanism is unknown but may have been grain flow or turbidite (Middleton and Hampton, 1973).

Many features in the Oswald West mudstone occur on the modern continental shelf off Oregon (Kulm and others, 1975). A mud facies consisting of extensively bioturbated, structureless silts and clays extends from the mid-shelf to shelf edge. This fine material is transported by turbid layers associated with the thermoclines at the surface and mid water levels. A second transport mechanism for the mud is the very dilute turbidity currents moving along the bottom. Similar mechanisms are suggested for the Oswald West mudstone. Glauconite is presently forming on the shelf with the highest concentrations (up to 98%) occurring near the shelf edge. The close comparison of the features in the Oswald West mudstone and the mud facies of Kulm and other (1975) indicates the that Oswald West was deposited in a middle continental shelf to upper slope basin, similar to those found off the Oregon coast today.

Astoria Formation

The lower to middle Miocene Astoria Formation was described in detail by Howe in 1926 from exposures in and around the town of Astoria, Oregon. He indicated a middle Miocene age for

these sandstones and "shales" (Moore, 1963). Since that time, various lithologies ranging from deltaic sandstone to deep-water mudstones have been mapped as the Astoria Formation in the Oregon Coast Range using age equivalence of strata as the principal correlation tool (Warren and others, 1945; Wells and Peck, 1961; Dodds, 1969; Beaulieu, 1973). The lack of litho- logic homogeniety and the areal extent of the formation has led to considerable confusion as to the stratigraphic boundaries and the depositional environment of the Astoria Formation in western Oregon.

Cressy (1974) working under the guidance of Dr. Alan Niem at Oregon State University suggested that the different mappable lithologic units within the Astoria Formation be desig- nated information members in accordance with the American Code of Stratigraphic Nomenclature (1961). Subsequent workers (Smith, 1976; Tolson, 1976; Neel, 1976; Penoyer, 1977; Nelson , 1978; Coryell, 1978; Cooper, 1980) have followed that suggestion, defining a series of facies related members. During this period, Addicott (1976) extended the base of the Astoria Formation into the lower Miocene based on a redefinition of molluscan assemblages which occur in the unit. Type sections, distribution, and facies relationships of these members have most recently been discussed by Cooper (1980) in a doctorate study of the Astoria Formation.

Three informal members of the Astoria Formation are recog- nized in this study area. These include the upper Silver Point mudstone member, the Pipeline mudstone member, and the Big Creek sandstone member (Fig. 5 and Plate I). The laterally equivalent upper Silver Point and Pipeline member mudstones are present in the western part of the study area. The underlying Big Creek member, exposed in the eastern section of the study area, forms the base of the Astoria Formation.

Big Creek Sandstone Member

Nomenclature and Distribution

The lower to middle Miocene Big Creek sandstone member was informally named by Cooper (1980) for a fine- to medium-grained marine sandstone exposed in stream banks along Big Creek two km. to the southwest of this study area. His reconnaissance study included a shallow-marine sandstone and a deep-marine sandstone facies in this member. Further detailed mapping by Coryell (1978) and Nelson (1978) revealed lithologic and stratigraphic differences between the shallow-marine and deep-marine facies. They restricted the Big Creek sandstone member to the shallow-marine facies and named the deep-marine facies the Pipeline member (see Pipeline member this study).

A thick fine-grained sandstone unit with subordinate interbedded siltstone and conglomerate is exposed in the eastern part of this study area (Plate I). The unit is included in the Big Creek sandstone member based on equivalent strati- graphic position, lithology, and age to the type area. This unit is exposed along the west bank of Plympton Creek (OC64, 252, and 692, Sec. 35, T8N, R6W) and capping the strike ridge east of Lost Lake (OC 246, Sec. 1, T7N, R6W). Its presence east of Nicolai Ridge extends the outcrop distribution of the Astoria Formation farther east than previously reported on the state geologic map of Wells and Peck (1961).

The Big Creek sandstone is estimated to be 50 to 100 meters thick in this study area based on the outcrop pattern and regional dip. At the type section the unit is 340 meters thick (Cooper, 1980). Coryell (1978) estimated the total thickness to be 460 meters in the type area.

Lithology and Sedimentary Structures

The Big Creek member is composed of coarse- to very fine-grained feldspathic sandstone, minor volcanogenic sandstone, siltstones, and a few volcanic pebble conglomerates.

The dominant lithology is a light bluish gray (5B 5/1) to light gray (N7), fine- to very fine-grained, bioturbated, struc- tureless to laminated sandstone. Commonly, exposures of this unit are weathered to very pale yellow orange (10YR 8/2). The sandstone ranges from very friable to well indurated and is an excellent bluff former, producing steep strike ridges along Plympton Creek (Fig. 8). Structures in this sandstone include: laminations formed by heavy mineral concentrations and micas rare poorly developed graded beds from medium-grained sand to coarse silt faint trough cross-bedding; and vertical burrows. These burrows are straight, cylindrical forms, approximately 6 cm long. The burrows are filled with fine-grained sand; the walls of the burrows are composed of mixed mud and sand. They appear similar to Rosselia burrows or to the straight tubed Skolithos. Both are indicative of shore-face or offshore bar deposition (Chamberlain, 1978).

Texturally, the sandstone is immature, based on Folk's (1951) definition. Composed of subangular grains, the sand-stone is poorly sorted, very positively skewed, and leptokurtic (see Grain Size Analysis Section). These sandstones are compositionally immature and classify as arkosic arenites (see Petrography Section).

A dark yellow orange (10YR 6/6), medium- to coarse-grained sandstone caps the fine-grained sandstone on a strike ridge east of Lost Lake (Plate I). This poorly sorted, arkosic sand-

Figure 8 Cliff formed by the Big Creek sandstone member. Unit is typically very fine-grained, with faint cross laminations. Note intraformational conglomerate in center of photograph. Exposure in Plympton Creek (OC 252, Sec. 2, T8N, R6W).

Figure 9 Conglomerate layer in Big Creek sandstone. Unit is composed of rounded volcanic cobbles and pebbles. White chips are abraded molluscan shells. See Appendix VII for composition of pebbles (OC 64, Sec. 35, T8N, R6W). Hammer for scale.

stone is characterized by medium-scale trough cross-beds. Individual troughs are up to 1 meter wide and 0.3 meter deep. They indicate a westward paleocurrent direction (S75^{o}W). Coryell (1978) also recognized a generally westward paleocurrent direction for the Big Creek sandstone in the eastern part of the type area. Further west, a northerly direction was recorded by Nelson (1978) and Coryell (1978).

A clayey siltstone, 10 meters thick, occurs in the middle of the Big Creek member interbedded with the fine-grained sandstone. This medium gray (N5) to dark greenish gray (5G 4/1), carbonaceous, micaceous siltstone is highly bioturbated and thinly laminated to structureless. The siltstone is well indurated, forming steep, smooth banks along Plympton Creek. Concentrations of mollusks in lenses and thin interbeds of fine-grained, silty sandstone are common.

Two 1 to 1.5 meter thick pebbly conglomerate layers are present in the fine-grained sandstone cliffs above Plympton Creek (OC 64 and 252). These lens-shaped units consist of subrounded to rounded, poorly sorted basaltic and andesitic cobbles to pebbles with minor amounts of chert, quartzite, and granodiorites(?) in framework support. Disseminated throughout the conglomerate are equant pieces, up to 2cm in diamter, of abraded molluscan shell fragments (Fig. 9).

Molluscan fossils are common in the fine-grained sandstone and siltstone of the Big Creek member. These fossils typically occur as unbroken, though rarely articulated, shells in lenses along bedding planes. This is in contrast to the fossil hash in the conglomerate layers. The molluscan fossils identified by Ellen Moore, (1980, written communication) including Andara devincta and Nuculana saccella calkinsi, are listed in Appendix IV. Diatoms, radiolarians and sponge spicules also were

observed in the siltstone. Most are abraded or dissolved beyond identification. The diatoms fragments, principally Thjalassiothrix, indicate an age of Miocene to recent (John Barron, 1980, written communication). A complete list is presented in Appendix IV.

Contact Relations

The Big Creek sandstone member disconformably overlies the Oswald West mudstone. The contact is placed at the top of a 5- to 10-meter thick glauconitic sandstone in the Oswald West mudstone (OC 60, Sec. 36, T8N, R6W; OC 248, Sec. 1, T7N, R6W). It is sharp with some reworking of the glauconitic sandstone into the overlying fine-grained sandstone of the Big Creek Member. This sharp lithologic change from outer shelf mudstone and glauconite to shallow-marine inner shelf sandstone suggests a possible submarine unconformity, equivalent to a diastem, between the units. The presence of the glauconitic sandstone indicates that the rate of deposition slowed or ceased and that subaerial exposure did not occur between the deposition of the Oswald West mudstone and the Big Creek sandstone as suggested by Coryell (1978). The high-energy beach to shallow marine environment described for the lower Big Creek by Coryell (1978) would rapidly erode the soft glauconite of the underlying Oswald West mudstone. In defense of Coryell, he never observed this contact in his study area.

The Big Creek sandstone is in normal fault contact with the Pittsburg Bluff Formation in Plympton Creek (Plate I, also see Contact Relations section for Pittsburg Bluff).

The middle Miocene Grande Ronde Basalt unconformably overlies the lower to middle Miocene sandstone member. This sharp contact is exposed in a quarry south of Nicolai Mountain

(OC 301, Sec. 21, T7N, R6W) where a fine-grained, micaceous, arkosic sand- stone is overlain by a closely-packed pillow basalt. The contact is considered unconformable because the Grande Ronde also overlies upper Oligocene sandstones on Plympton Ridge to the southeast (SE, Sec. 15, T7N, R6W).

The contact between the Big Creek sandstone and the overlying Pipeline and upper Silver Point mudstones is not exposed due to the cover by the basalt in the central part of the study area. Coryell (1978) and Nelson (1978) described a conformable contact or possible diastem between the Big Creek member and both the Pipeline and upper Silver Point members. The contact is at the top of a 1 meter thick glauconitic sandstone west of this study area.

Age and Correlation

An early to middle Miocene age is indicated for the Big Creek sandstone mapped in this study area. This age is based on molluscan fossil assemblages which range from the Pillarian to the Newportian Stages (Moore, 1980, written communication). Typical fauna present in this unit which range from Pillarian to Newportian include *Anadara* sp.cf. *A devincta* and *Nuculana* (*saccella*) cf.n. (s.) *calkinsc*. Only the pelecypod *Nuculana* (*saccella*) cf.n. (s.) anelga, which Moore tenatively identified from an assemblage collected east of Lost Lake (OC245, 2, T7N, R6W) is diagnostic exclusively of the Newportian stage. Therefore it can not conclusively be stated that the Big Creek sandstone in this study extends into the Pillarian or is exclusively Newportian in age.

Nelson (1978), Coryell (1978), and Cooper (1980) postulate a time transgressive model for the Big Creek sandstone, with the unit becoming progressively younger (i.e., Newportian) to

the east toward the type section. However, if the Big Creek member in this study, which is to the east of the type section, does extend down into the Pillarian then the time transgressive model does not work.

The minimum upper time boundary of the Big Creek sandstone member is defined by the age of the overlying Grande Ronde Basalt. The basalt is dated at 14 to 16 $\pm$ 0.5 m.y. (Swanson and Wright, 1978) (Fig. 13). This is equivalent to the upper part of the Newportian Stage (Fig. 2). Coryell (1978) defined the Big Creek as older than the upper Silver Point mudstone based on contact relations observed in his thesis area to the west.

The Big Creek sandstone in this study is equivalent to the Big Creek member in the Young's River area described by Nelson, (1978) and Coryell (1978). This is based on similar lithologies (fine-grained, shallow-marine sandstone and siltstone); age (Pillarian to Newportian); and stratigraphic position (above the Oswald West mudstone and below the middle Miocene basalts). This sandstone is also equivalent, in part, to the Angora Peak sandstone member of the Astoria Formation exposed to the southwest (Cooper, 1980). This is based on equivalent aged fauna and similar stratigraphic position.

The Scappoose Formation described by Van Atta (1971) in the Nehalem River valley is considered to be late Oligocene to early Miocene in age by Warren and Norbisrath (1946). This age is based on only a few fossil assemblages. Van Atta (1971) postulated that the Scappoose may be equivalent in part to the Astoria Formation based on the presence of the overlying Columbia River Basalts in the Nehalem River area. The age of the Scappoose, however, has not been fully documented as yet. Portland State graduate students are working on this formation

on the eastern flank of the Coast Range and will hopefully define the age more completely.

The Scappoose Formation is similar in lithology to the Big Creek sandstone member of this study. Both are dominated by fine-grained shallow marine sandstone with subordinate siltstone and a few volcanic conglomerates. The presence of the Pittsburg Bluff Formation in this study area, which is stratigraphically below the Scappoose Formation to the east, suggests that some of the Scappoose shallow marine sandstone could have extended this far west. The late Oligocene fine-grained sandstone on Plympton Ridge (312, Sec. 15, T7N, R6W) may be the westward expression of the Scappoose Formation. This would suggest that in its upper part it may be partially equivalent to the lower to middle Miocene Big Creek member. Alternatively, the Big Creek sandstone in this study may be better defined as an upper lateral facies of the Scappoose Formation.

At this time, the author would prefer to keep the Big Creek sandstone, in this study area, within the Astoria Formation. The relationship of the Oswald West mudstone, Scappoose Formation, and Big Creek member provides another link between the eastern and western stratigraphic columns of the northern Coast Range. A study in progress to the south by Jeff Goalen should substantiate this correlation.

Depositional Environment

An open marine sublittoral to middle shelf depositional environment is postulated for the Big Creek sandstone member. This environmental interpretation is based on the molluscan fossil assemblages, textural features, and sedimentary structures. Cooper (1980) has summarized the depositional environment of this unit to the west at the type section.

An inner to middle shelf, low-energy environment is indicated for the fine-grained sandstone and minor siltstone which comprise approximately 80% of the Big Creek member. The siltstone interbeds may represent slightly deeper, quieter water conditions or decreases in energy input. The molluscan assemblages present support this conclusion. Nuculana (saccella), common in this unit, denotes water depths of 10 to 80 meters (Moore, 1980, written communication). Anadara devincta found in the siltstone is an inner shelf indicator (Cooper, 1980). The diatoms present are dominated by a near-shore environment species (0-200m) (Barron, 1980, written communication).

The fine-grained size and moderate sorting of the Big Creek sandstone support an inner to middle shelf environment (see Grain Size Analysis Section). Statistical grain size analyses suggest that the fine-grained sandstone was transported by tractive current, possibly from a bottom turbid layer (Kulm and others, 1975), rather than surf activity (Friedman, 1962). The sorting and grain sizes of the sieved samples correspond to the shelf sand and mixed sand and mud facies characteristic of inner to middle shelf environments (Kulm and others, 1975).

The sedimentary structures in these shelf sandstones and siltstones consist of rare laminations, lens-shaped molluscan fossil accumulations, and rare graded beds(?). Laminations typically develop in shelf sands on the inner continental shelf off Oregon today (Kulm and others, 1975). They can also form in shoreface zones of modern beaches (Reineck and Singh, 1975) and in the outer planar facies of non-barred nearshore high energy environments (Clifton and others, 1971). The sand and mixed sand and mud facies of Kulm and others (1975) are commonly bioturbated destroying the primary laminations.

Fossils are found in lenses along bedding planes. They are usually unbroken but are rarely articulated, suggesting that weak currents concentrated them into lenses. Some may have been transported from shallower water. Graded beds may be the result of a waning concentrated turbid bottom layer developed during storm activity. The molluscan fossils present along with the lack of well developed stratification, poor sorting, and dominant silt size fraction are characteristic of middle shelf deposits (Reineck and Singh, 1975; Kulm and others, 1975). The rare framework supported pebbly conglomerates with rounded clasts and abraded molluscan fragments are interpreted as high-energy storm shelf or surf deposits.

A highly bioturbated fine-grained silty sandstone is found above the inner shelf Big Creek sandstone in Plympton Creek (OC64, Sec. 35, T8N, R6W). This carbonaceous siltstone is characterized by vertical burrows which are similar to Rosselia burrows or Skolithos tubes. Both are indicative of shallow-water shoreface or offshore bar environments. At the base of the member but predominantly overlying these units are trough cross-bedded fine- to coarse-grained clean sandstones. These thick sandstones are moderately sorted and lack significant argillaceous interbeds. They denote higher energy, nearshore, inner sublittoral deposits (Clifton and other, 1971). Alternatively, these sandstones could represent sand ribbons which form on the inner to middle shelf (Walker, 1979) by strong bottom currents related to storms that build the sand into elognate parallel bodies (Swift, 1976).

In summary, the Big Creek sandstone in this study are is composed predominantly of a slightly deepening then shallowing upward marine sequence. The middle units which make up 70% of the member consist of fine-grained inner to middle shelf sand-

stones and some siltstones. Overlying and underlying this are coarser grained higher energy shoreline sandstones. The rare pebbly conglomerates in the upper part of the sequence are interpreted as high energy storm shelf or surf deposits.

Upper Silver Point Mudstone Member

Nomenclature and Distribution

The Silver Point member was first proposed by Smith (1975) for a thick sequence of marine turbidite sandstones and finely laminated mudstones exposed in the sea cliffs at Silver Point near Cannon Beach, Oregon. This member was divided into an upper and lower unit based on the presence or absence of the interbedded turbidite sandstone layers (Smith, 1975; Neel, 1976). Coryell (1978) recognized only the upper Silver Point (laminated mudstone without turbidite sandstones) in his thesis area to the southwest of this study. He also mapped the structureless, mottled, grayish orange (10YR 7/4) mudstone associated with the overlying middle Miocene basalts into this unit. This study has attempted to separate the interbedded mudstone unit from the underlying Silver Point (see Sedimentary Interbeds in Basalt Section). By definition the Astoria Formation includes the sedimentary rocks which lie below the middle Miocene extrusive basalts (Fig. 5).

Coryell (1978) and Nelson (1978) have defined a unit within the upper Silver Point mudstone (Tspu). This unit, designated as $Tspu_1$, overlies the upper Silver Point mudstone exposed in the extreme northwest corner of the study area (Plate 1).

The total thickness of the upper Silver Point is estimated to be 120 meters, based on regional dip and outcrop pattern.

Lithology and Sedimentary Structures

The upper Silver Point mudstone (Tspu) is composed of medium to dark gray (N6-N3), laminated, highly carbonaceous, micaceous, clayey siltstone. Characteristically, exposures weather to a yellowish gray (5YR 8/1) to brownish gray (5YR 4/1) color with conchoidal fractures which produce "chippy" talus slopes at the base of road cuts. This mudstone produces low, slump prone hills (Carter, 1976). Variation in grain size from claystone to coarse siltstone and rarely very fine-grained sandstone produce the laminations. Mica flakes and carbonized plant fragments are concentrated along many bedding plane surfaces. These commonly cause a shaly or papery splitting character to the mudstone. Rare complete deciduous leaf imprints were observed along bedding planes (OC 200, Sec. 14, T8N, R8W).

Hydrometer analysis showed that the mudstone is a clayey siltstone consisting of 60% silt and 32% clay size fraction (see Grain Size Analysis Section). The remaining 8% is composed of carbonaceous material (3%) and sand (5%). Mineralogically the silt fraction is composed of quartz, mica, and feldspar. Nelson (1978) determined the clay mineralogy of the Silver Point mudstone. He identified montmorillonite, mica, kaolinite and possible chlorite within the mudstone in his thesis area.

Thin-shelled molluscan fossils (principally molds and casts) are scattered throughout the unit. These along with the benthic forams enable determination of both the age and environment of deposition for this unit. See Appendix IV for a complete list of the fossils identified.

The upper Silver Point is distinguished from the lower Silver Point by the absence of rhythmically bedded turbidite sandstones (Nelson, 1978). The upper part of this mudstone ($Tspu_1$) is lithologically identical to the lower part (Tspu) with the exception that it may have slightly more very fine-grained sandstone lamination present and that the Pipeline member lies between the two upper Silver Point mudstones (Fig. 5).

This unit is distinguished from the Pipeline member mudstone (Tpm) on the basis of the occurrence of numerous sandstone beds, lenses, and clastic dikes in the Pipeline member.

Contact Relations

The lower contact of the upper Silver Point mudstone (Tspu) is not exposed in the study area. To the southwest a concordant contact with the underlying Big Creek member was described by Coryell (1978). The contact is defined by a glauconitic sandstone at the top of the Big Creek member. Nelson (1978) and Smith (1975) showed that the contact between the upper and lower Silver Point turbidite member is gradational over a few hundred meters.

The upper Silver Point mudstone (Tspu) gradationally gives way to the Pipeline member mudstone (Tpm). This gradational contact is defined as between the first occurrence of the arkosic sandstone interbed in the Pipeline mudstone (OC 216, Sec. 24, T8N, R7W), and the last exposure of the upper Silver Point mudstone (Tspu) (OC 98, Sec. 25, T8N, R7W) along Tripp Road the contact between the Pipeline mudstone and the overlying Silver Point mudstone unit ($Tspu_1$) is again gradational, being placed above the disappearance of the arkosic sandstone lenses and clastic dikes. This relationship can be observed along Hillcrest Road (Sec. 26, T8N, R8W).

An unconformable relationship between the extrusive middle Miocene Depoe Bay Basalt and the upper Silver Point mudstone is inferred to exist in this study area. The contact is obscured by vegetation and alluvium but the change in slope designates the approximate contact (Plate I, Sec. 30, T8N, R7W). A discordance in strike and dip between the units and the regional map pattern of the Depoe Bay Basalt indicate that the contact is a slight angular unconformity. The Depoe Bay Basalt overlies the lower part of the upper Silver Point member (Tspu). Both the overlying Pipeline mudstone member and upper unit of Silver Point member ($Tspu_1$) are missing. Penoyer (1977) also depicted this contact as an angular unconformity to the southwest of this study. There, the Depoe Bay Basalt is in contact with both the upper Eocene to lower Oligocene Oswald West mudstone and the early Miocene Silver Point mudstone.

Both the Cape Foulweather and Frenchman Springs basalts and the Clifton Formation are considered younger than the Depoe Bay Basalt (Fig. 5). The contact between the Silver Point and these younger basalts and sedimentary units depicted on Plate I is therefore also considered unconformable.

The unconformable contact between the Silver Point and the overlying Clifton Formation is further discussed in the Contact Relation Section of the Clifton Formation.

Age and Correlation

The age of the upper Silver Point mudstone in this study area is considered to be late early Miocene to middle Miocene. This is based on the stratigraphic relationships and fossil collections. Two molluscan fossil assemblages (OC91 and 208, Appendix IV), which included *Anadara sp.* and *Delectopecten cf.*

D. peckhami, have ranges equivalent to the Newportian and Pillarian Stages. A poorly preserved specimen, comparable to Nuculana (S.) amelga, may restrict the age to the New- portian Stage (Moore, 1980, written communication). Forams collected, including Globigerina pacifica and Florilus incisum, from locality OC 208 range from the Zemorrian to the Saucesian foram stages (Rau, 1980, written communication), but the Pipe- line mudstone which is stratigraphically below fossil locality OC 208, has forams diagnostic of only the Saucesian Stage (see Pipeline - Age and Correlation Section). This suggests that the upper Silver Point (Tspu) is restricted to the Saucesian Stage. The overlying Depoe Bay Basalt is elsewhere dated at 14.0 to 16.0 $\pm$ 0.5 m.y. (Snavely and others, 1973). This radiometric date substantiates the middle Miocene age for the upper Silver Point strata.

The upper Silver Point in this study is equivalent to the upper Silver Point of Coryell (1978) and Nelson (1978). This correlation is based on similarity of lithology, stratigraphic position, and age. The unit can be continuously mapped from Coryell's thesis area into this study area.

Pipeline Member

Nomenclature and Distribution

The term "Pipeline Member" of the Astoria Formation was informally applied by Coryell (1978) and Nelson (1978) to deep-water submarine canyon-fill sandstones and subordinatwe mudstones exposed along the Pipeline mainline logging road immediately to the west of this study area. Cooper (1980) placed the upper sandstone of this member in the Big Creek sandstone member as a deep-water facies and included the mudstone in the upper Silver Point member. This study follows the

nomenclature of Coryell (1978) and Nelson (1978). They divided the Pipeline member into two units: a lower mudstone dominated interval (Tpm) and an upper arkosic sandstone dominated interval (Tps).

The Pipeline mudstone (Tpm) is exposed in the western section of this study area (Plate I) where it lies between the upper and the lower parts of the upper Silver Point mudstone (Tspu and $Tspu_1$; cross section B-B, Plate III). The unit is best exposed in a measured section along Tripp Road (OC 216, Sec. 24, T8N, R8W; Appendix I).

The Pipeline member thins considerably from the west into this area. Nelson (1978) estimated total thickness of the member to be 600 to 900 meters. The canyon-fill sandstone portion of the unit (Tps) of Coryell (1978) and Nelson (1978) is not present in this study area. The mudstone dominated member (Tpm) is estimated to be not more than 100 to 200 meters thick in the thesis area based on the regional dip and outcrop pattern.

Lithology and Sedimentary Structure

The Pipeline mudstone is composed of carbonaceous, micaceous, clayey siltstone with lenses and clastic dikes of arkosic sandstone. The ratio of siltstone to sandstone varies from 9:1 to 6:1.

The mudstone is lithologically identical to the upper Silver Point mudstone (see Lithology and Sedimentary Structures of the Silver Point detailed description).

Figure 10 Typical exposure of Pipeline mudstone. Gray is fresh mudstone, yellowish brown is weathered horizon. One meter thick arkosic sandstone interbed near top of exposure differentiates this unit from the upper Silver Point mudstone. Exposure along Tripp Road (OC 216, Sec. 24, T8N, R8W).

The interbedded sandstone is fine- to very fine-grained and structureless. It is iron-stained, yellow gray (5YR 8/1) to very light gray (N8) in weathered exposures. Texturally, the sandstone is moderately to poorly sorted, very positively skewed (e.g., enriched in clay matrix) and leptokuritic (see Size Analysis section). It is texturally immature (Folk, 1951). The fines may be a result of the diagenetic alteration of the feldspars. This sandstone is classified as a micaceous arkosic wacke (see Petrography section).

Sandstone beds which range from 0.5 to 1.5 meters are carbonaceous and micaceous, and display sharp contacts with the adjacent mudstones. Sandstone layers are lens-shaped. The mudstones are thickly bedded to laminated, carbonaceous, and micaceous. Rare clastic dikes are present in the mudstone. The clastic dikes are up to 30 cm wide and extend two to three meters into the mudstone from the source beds.

The presence of arkosic sandstone lenses and dikes is the main distinguishing feature between the Pipeline member and upper Silver Point member (Nelson, 1978; Coryell, 1978; Fig. 10). No fossils were present in the sandstone, but forams and rarely mollusks are found throughout the mudstone layers (OC 212, Sec. 26, T8N, R8W; OC 216, Sec. 24, T8N, R8W).

Contact Relations

The contacts between the Pipeline (Tpm) and upper Silver Point mudstones (Tspu and $Tspu_1$) are described in the Contact Relation Section of the Silver Point. See the Contact Relation Section of the Clifton Formation and the measured section b b' in Appendix I for a detailed description of the Pipeline mudstone-Clifton Formation contact.

Age and Correlation

The Pipeline mudstone lies stratigraphically between the upper and lower parts (Tspu & $Tspu_1$) of the upper Silver Point mudstone which is middle early Miocene to middle Miocene. Fossil assemblages collected from the Pipeline mudstone confirm the age equivalence. A Saucesian age was determined from forams, such as *Siphogenerina sp.* and *Valvulineria araucana,* collected at one locality (OC 212 Appendix IV). Forams, including *Globigerina spp.*, from near the upper contact with the Clifton formation (OC 216) indicated an age no older than Saucesian but the upper age boundary could not be determined due to a poorly preserved assemblage (Rau, 1980, written communication). Mollusks collected from the same site were placed in the Pillarian or Newportian Stages by Moore (1980, written communication).

The Pipeline mudstone can be directly correlated with the Pipeline mudstone member of Coryell (1978) and Nelson (1978), based on continuous mapping, age, and stratigraphic positions. The unit interfingers with the upper Silver Point member and is therefore, considered correlative to it.

Depositional Environment of the Upper Silver Point and Pipeline Mudstones

The upper Silver Point and Pipeline mudstones were deposited on the inner continental shelf to upper slope, in a quiet open marine environment. This is evident from the fossil paleoecology and lithologies present.

Foram assemblages indicate a middle to upper bathyal depth (220-800m) at the site of deposition (Rau, 1980, written

communication) for both units. The mollusk Delectopecten present in the Silver Point mudstone ($Tspu_1$), is commonly found in water depths greater than 200 meters (Moore, 1980, written communication). A shallower water environment was reported by Cooper (1980) for the lower part of upper Silver Point (Tspu), based on the mollusk Anadara devincta (OC 91, Sec. 29, T8N, R7W; this study). This inner shelf water depth is further supported by the presence of Nuculana (saccella) sp. found at the same locality. Living species of N. (saccella) exist in water depths of 10 to 80 meters (Moore, 1980, written communication).

There are two possible interpretations for the disparity between water depths indicated by the fossil evidence. First, there may have been a gradual deepening of the environment through time from the lower part of the Silver Point (Tspu) through the Pipeline (Tpm), and then a shallowing into upper Silver Point ($Tspu_1$) time. Alternatively, the shallower water fauna may have been transported into a deeper water environment. I favor the first explanation because the fossils were unbroken and rarely articulated. Coryell (1978) also suggested a deepening of the depositional environment from the lower part of the Silver Point (Tspu) to the Pipeline member.

The dominant silt and clay size fraction of these units indicate hemipelagic deposition in a quiet water environment. The mudstone of the upper Silver Point plots in the quiet water region of Passega's (1957) "C-M" diagram. The data also plot into Kulm and others (1975) mixed sand and mud facies, characteristic of the middle continental shelf (See Grain Size Analysis Section).

The alternating laminations of silt and clay suggest slight fluctuations of currents in a low energy environment. The

thin-shelled unbroken, rarely articulated bivalves again denote a low energy environment. The abundance of organic plant material and lack of bioturbation imply a partially restricted environment. The mica and carbonaceous material commonly associated with the coarse silt and very fine-grained sand probably are the result of low density turbid bottom currents which carried the hemipelagic material into the quiet water shelf to slope environment. Currents transporting this sediment are common on the continental shelf today being related to major river discharges and storm waves (Kulm and others, 1975).

To the west, associated with these units, is a 100-meter thick six-mile wide submarine canyon-fill facies (Pipeline sandstone member) consisting of a series of amalgamated grain flow deposits (Coryell, 1978). Nelson (1978) suggests that the fine- to very fine-grained sandstone lenses and interbeds associated with the underlying Pipeline member mudstone are the result of temporary turbidite or grain flow influxes of sand related to channel shifting in the submarine canyon system. The arkosic sandstone interbeds in the Pipeline mudstone (Tpm) in this study are probably related to these processes. The few clastic dikes occur due to spontaneous liquefaction of rapid loading of pore fluid-rich muds and sand.

In summary, the upper Silver Point and Pipeline mudstones were deposited in a gradually deepening marine environment from the inner or middle shelf to upper slope. These mudstones were deposited on the periphery of the northeast-southwest oriented submarine canyon facies described by Nelson (1978) and Coryell (1978), receiving only an occasional overbank sand deposit from the system. The low density turbid currents, which carried the continentally derived hemipelagic material into the quiet shelf or upper slope, may have originated from overbank turbidity

currents related to the submarine canyon system; from storm waves stirring up bottom currents, or by discharge into the ocean of suspended load from a major river system.

Basalts

Columbia River Basalt Group

In the middle Miocene a thick sequence of continentally derived tholeiitic flood basalts were erupted from northwest trending vent systems along the Oregon-Idaho border. These basalts covered some 200,000 sq. km. of eastern Oregon, eastern Washington, and western Idaho (Swanson and Wright, 1979, preprint). Some of these basalts flowed down an ancestral Columbia River valley and into a marine embayment near the mouth of the present day Columbia River (Snavely & Wagner, 1964; Figure 12). The radiometric ages of these basalts range from 16.5 to 6 m.y. (McKee and others, 1977; Watkins and Baksi, 1974) with the majority extruded between 14.0 and 16.0 $\pm$ 0.5 m.y. or middle Miocene (Swanson and Wright, 1978). During this same time interval a sequence of chemically and petrologically similar basalts were extruded on the sea floor along the present coast of northern Oregon and southwestern Washington (Snavely and others, 1973).

Until 1961, the Columbia River Basalts of eastern Oregon and Washington were considered a thick pile of monotonous flood basalts. At that time, Waters (1961) and Mackin (1961) through detailed mapping augmented with chemical analyses differentiated regional stratigraphic units. Since then, further various stratigraphic schemes have been developed using different combinations of lithologies, chemical analyses, age, remnant magnetism, and relative stratigraphic position in an attempt to differentiate individual cooling units on the

Figure 11: Gnat Creek Falls developed on high MgO Grande Ronde Basalt. Note typical entablature and lower colonnade joint sets. Exposure in Gnat Creek Gorge (NW, Sec. 6, T7N, R8W).

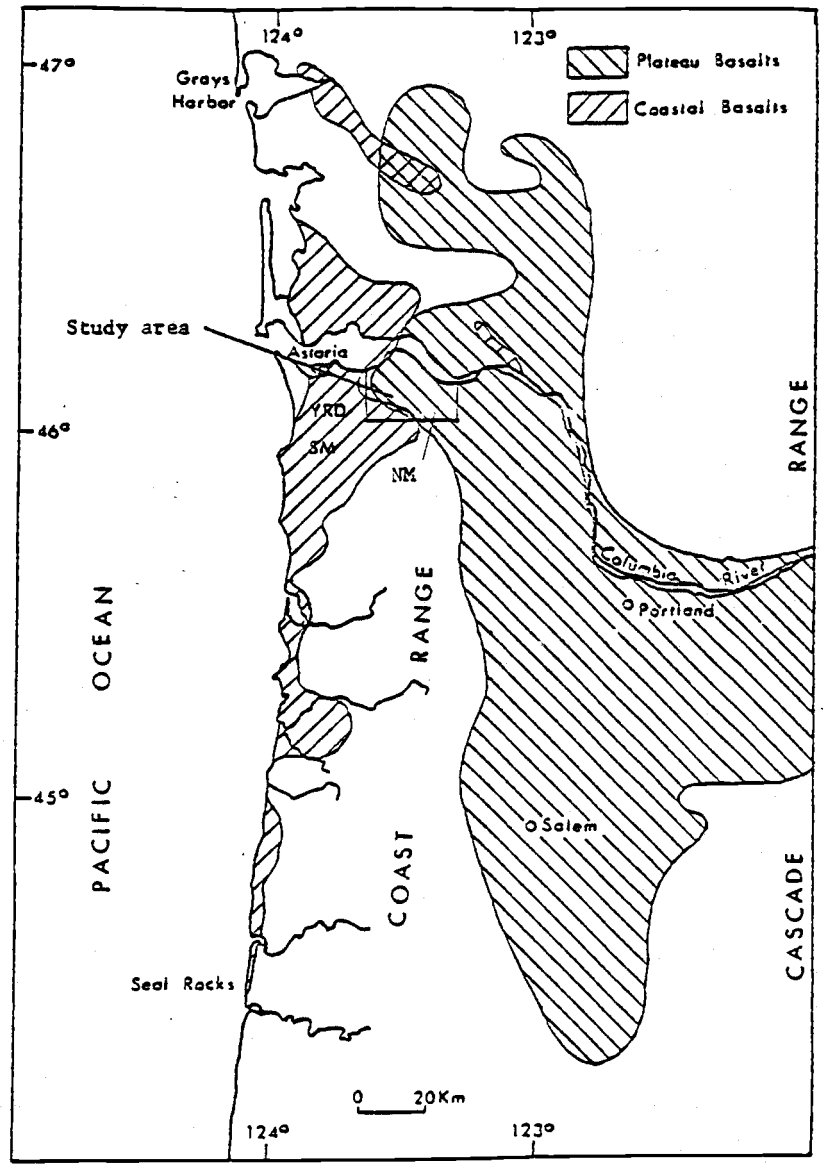


Figure 12 Areal distribution of plateau and coastal basalts in western Oregon and Washington (modified from Beeson and others, 1979) NM = Nicolai Mountain; SM = Saddle Mountain YRD = Youngs River dike.

Columbia Plateau (Bingham and Grolier, 1966; Wright and others, 1973). The latest and most complete stratigraphy of these plateau-derived basalts is that of Swanson and others (1979) (Fig. 13). They followed Grigg's (1976) suggestion and raised the Columbia River Basalt (CRB) to group status and the Yakima Basalts to subgroup level leaving the less differentiated Picture Gorge and Imnaha Basalts at the formation level. It is the Yakima Basalt Subgroup which is of interest to this study (Fig. 13).

Three formations have been defined within the Yakima Basalt Subgroup. They are from oldest to youngest, the Grande Ronde Basalt, Wanapum Basalt, and Saddle Mountains Basalt. Associated with these units but excluded from the Columbia River Basalt Group are a series of sedimentary interbeds belonging to the Ellensburg and Latah Formations (Swanson and others, 1979). The most useful unit stratigraphically is the Vantage Member of the Ellensburg Formation.

The Grande Ronde Basalts are the most voluminous, extensive, and oldest flows of the Yakima Basalt Subgroup. The name was proposed by Taubeneck (1969) and formalized by Swanson and others (1979). The formation ranges in thickness from over 1200 meters in the central plateau to 150 to 500 meters in the peripheral areas. The Grande Ronde Basalts are composed of non-porphyritic, finely crystalline tholeiitic basalts. Few flows are sufficiently distinctive in lithology in the field to be traced regionally, but Swanson and Wright (1978) have shown that the formation can be divided by four magnetostratigraphic units (Fig. 13). These along with the known geochemical divisions will be discussed in the basalt stratigraphy section. Physical criteria such as jointing characteristics and mineralogy can also be used on a local scale, however, intraflow variations cause uncertainty over large distances.

Series		Group	Sub-group	Formation	Member or Flow	K-Ar age (m.y.)	Magnet. Polarit.
MIOCENE	UPPER MIOCENE	Columbia River Basalt Group	Yakima Basalt Subgroup	Saddle Mountains Basalt	Lower Monumental Member	6	N
					erosional unconformity		
					Ice Harbor Member Basalt of Goose Island	8.5	N
					Basalt of Martindale	8.5	R
					Basalt of Basin City	8.5	N
					erosional unconformity		
					Buford Member		R
					Elphant Mountain Member	10.5	N,T
					erosional unconformity		
	MIDDLE MIOCENE				Mattawa Flow		N
					Pomona Member	12	R
					erosional unconformity		
					Esquatzel Member		N
					erosional unconformity		
					Weissenfels Ridge Member Basalt of Slippery Creek		N
					Basalt of Lewiston Orchards		N
					Asotin Member		N
					local erosional unconformity		
					Wilber Creek Member		N
					Umatilla Member		N
					local erosional unconformity		
				Wanapum Basalt	Priest Rapids Member		R_3
					Rosa Member		R_3/T
					Frenchman Springs Member	14.5	N_2
					Eckler Mountain Member		
					Basalt of Shumaker Creek		N_2
					Basalt of Dodge		N_2
					Basalt of Robinette Mountain		N_2
	LOWER MIOCENE			Grande Ronde Basalt		14.5-16.5	N_2
							R_2
							N_1
							R_1
				Picture Gorge Basalt			R_1
				Imnaha Basalt			T
							N_0
							R_0

Figure 13 Columbia River Basalt Group stratigraphy of eastern Oregon and Washington. Marked units are those which reached this study area (modified from Beeson and Moran, 1979)

The Vantage sandstone interbed has formally been assigned to the Ellensburg Formation by Swanson and others (1979). Although the lithology of the Vantage Member is variable, these strata comprise an important stratigraphic unit which separates the Grande Ronde Basalt from the Wanapum Basalt. The Vantage Member is the most distinctive stratigraphic marker in the western Columbia plateau and has been traced into the Western Cascade area (Anderson, 1978). This unit has been recognized within the study area and will be more fully discussed in the next section.

The Wanapum Basalts, first named by Mackin (1961), overlie unconformably the older Grande Ronde Basalts of the Plateau (Swanson and others, 1978). The flows are distinguished in the field by a generally medium crystalline, olivine-bearing basalt which normally contains a few percent of plagioclase phenocrysts up to 2 cm in length. Chemically most flows contain higher Fe and Ti and lower SiO_2, than the Grande Ronde. The formation has been divided into four members based on petrography and magnetic polarity (Fig. 13). The Frenchman Springs and Priest Rapids Members of the Wanapum Basalt have been recognized in Western Oregon (Beeson and Moran, 1979) but only the Frenchman Springs Member is present in the study area.

The youngest formation of the Yakima Basalt Subgroup is the heterogeneous Saddle Mountains Basalt. First named by Bingham and Golier (1966) and formalized by Swanson and others (1979), the formation is characterized by its diverse chemistry, petrography, age and paleomagnetic polarities. The age of the formation is 13.5 to 6$\pm$ 0.5 m.y. (McKee and others, 1977), and yet it comprises less than one percent of the total volume of the Columbia River Basalts and is by far the most chemically variable. The Pomona Member, dated at 12 $\pm$ 0.5 m.y. by McKee

and others (1977), is the only member thought to have flowed into western Washington and Oregon (Schmincke, 1967; Swanson and others, 1979). Originating in western Idaho, the Pomona flow moved down the ancestral Snake River Canyon, across the Columbia Plateau as a sheet flow and was funneled down the ancestral Columbia River valley to at least as far as Kelso, Washington (Kienle, 1971; Schmincke, 1967). This single cooling unit averages 30m in thickness and covers an estimated 600 km^2 (Choiniere and Swanson, 1979). The Pomona Member is characterized by reverse polarity, abundant small plagioclase phenocrysts (<5mm) and glomerophenocrysts of plagioclase and clinopyroxene. This unit has been located in the Nicolai Mountain-Gnat Creek area and will be described more fully in the next section.

Coastal Basalts

According to Snavely and others (1973), during the same time period that the Plateau-derived Columbia River Basalts were extruded in eastern Oregon and Washington, a similar but much less voluminous series of submarine tholeiitic basalts was produced in a narrow north-south trending belt of volcanic centers along the present Oregon and Washington coast (Fig. 12). These Miocene flows are the youngest of three periods of basaltic volcanism in the Tertiary sequence of the Coast Range (Snavely and Wanger, 1963). The unit consists of a thick sequence of hyaloclastite breccias, broken pillow and isolated pillow lavas along with sills, dikes, and irregular intrusive bodies. Rare subaerial flows have also been described (Snavely and others, 1973; Penoyer, 1977; Tolson, 1976; Coryell, 1978). The hyaloclastite breccias, up to 600 meters thick, and sills, up to 170m thick, have been described by Penoyer (1977), Neel (1976), Cressy, (1974), and Tolson, (1976). The dikes range in width from 1 to 15 meters but are generally only 1 to 4 meters.

Three major petrologic and chemical types have been differentiated within the middle Miocene basalts of the Coast Range by Snavely and others (1973). They are, from oldest to youngest: the Depoe Bay basalt, the Cape Foulweather basalt, and the basalt at Pack Sack Lookout, Washington.

The Depoe Bay petrologic type was named from outcrops in the town of Depoe Bay, Oregon by Snavely and others (1973). The unit is composed of undifferentiated aphyric lavas in the form of extrusive submarine hyaloclastite bedded breccias, isolated pillow lavas, intrusive dikes and sills, and rare subaerial flows. Hill (1974) using both trace elements and major oxides, recognized on a regional scale a high and low MgO division within the Depoe Bay basalts. Neel (1976), Penoyer (1977), and Coryell (1978), showed that deep-water foram-bearing mudstone interbeds are present within the basalts, suggesting that a stratigraphy could be worked out. Radiometric ages of Depoe Bay basalts range from 16.5 $\pm$ 0.65 m.y. to 14.0 $\pm$ 2.7 m.y. (Snavely and others, 1973).

Stratigraphically above the Depoe Bay basalts in the vicinity of the type locality are marine arkosic sandstones and siltstones, referred to as the sandstone at Whale Cove (Snavely and others, 1969 and 1973). This unit is commonly intruded by Cape Foulweather basalt which produced peperitic units and penecontemporanceous deformation.

The Cape Foulweather petrologic type is named from Cape Foulweather along the Oregon Coast where submarine breccias and a volcanic vent with associated ring dikes and radial dikes were described by Snavely and others (1973). This basalt is distinguished in the field from Depoe Bay Basalt by the presence of large scattered plagioclase phenocrysts (1-2 cm in

length) in a finely crystalline groundmass. Chemically, the Cape Foulweather Basalt displays a lower SiO_2 and higher FeO and TiO_2 content than the Depoe Bay Basalt. To the north of the type locality, a series of Master's theses at Oregon State University under the guidance of Dr. Alan Niem have continued to verify the stratigraphic, petrographic, and chemical differences between the Depoe Bay and Cape Foulweather basalts depicted by Snavely and others (1973). It appears from this mapping that the basalts were emplaced in progressively deeper water northward into Clatsop County, Oregon (Neel, 1976; Penoyer, 1977; Tolson, 1976; Coryell, 1978).

The basalt at Pack Sack Lookout, Washington is the youngest petrologic type in the coastal basalts and is dated by radiometric methods at 9 ± 1.4 m.y. (Snavely and others, 1973). The unit has limited areal extent and principally consists of a subaerial flow and two sills (Snavely and others, 1973; Wolfe and McKee, 1972). The Pack Sack basalt is medium gray with abundant small plagioclase phenocrysts (up to 5mm) which have resorption voids filled with glass and augite. Less common are glomerophenocrysts of plagioclase and pyroxene. The basalt is chemically distinctive from the other coastal basalts and has reverse polarity. The lack of feeder dikes and uncertainty as to the age of the overlying sediments has left open the possibility that this unit was derived from the Plateau via flowage down an ancestral Columbia River valley (Snavely and others, 1973). The absolute age of this basalt is also uncertain as it conflicts with that of the Pomona Member dated in eastern Oregon by McKee and others, (1977).

Snavely and others (1973) also showed the chemical, age, and petrologic similarities between the petrologic types of the coastal basalts and the equivalent Columbia River Basalts (Fig. 14). The Depoe Bay Basalt is equivalent to the Grande Ronde

Basalts (Yakima Basalt of Snavely and others, 1973) and the Cape Foulweather Basalt matches with the Frenchman Springs Member of the Wanapum Basalts (Late Yakima of Snavely and others, 1973). The Pack Sack basalt equates to the Pomona Member of the Saddle Mountains Basalt. Hill (1974) further confirmed the chemical similarity between the coastal and Yakima Basalt Subgroup using trace elements and major oxides. Bowman and others (1973) also showed that the Pack Sack Lookout flow was geochemically the same as the Pomona Member.

This thesis area is situated at a point where the coastal petrologic types and their Columbia River Basalt counterparts are thought to interfinger (Fig. 12). Dr. Niem and his students have mapped the coastal basalts up to the southwest edge of the study area. Kienle (1971), using major oxide chemistry and paleomagnetism, correlated the plateau-derived Yakima, Late Yakima (now equivalent to the Grande Ronde Basalt and Frenchman Springs Member of the Wanapum Basalt), and the Pomona flow to the eastern edge of the study area (Bradley State Park). Snavely and others (1973) also recognized Frenchman Springs and Grande Ronde Basalt petrologic types on top of Nicolai Mountain. More recently the detailed stratigraphy of the Columbia River Basalts in western Oregon, particularly in the Mount Hood-Clackamas River area, has been done by Dr. Marvin Beeson and his students at Portland State University (Beeson and others, unpublished).

Through detailed stratigraphic analysis of the different basalt units this study attempts to show the stratigraphic relationship between the coeval Columbia River Basalts and coastal basalts in the Nicolai Mountain-Gnat Creek area. This relationship becomes critical in light of ideas which have been put forth as to the diverse origins of the coastal basalts suggested by Snavely and others (1973) and more recently by

STRATIGRAPHIC RELATIONS OF MIOCENE BASALT UNITS
IN THE OREGON AND WASHINGTON COAST RANGES

COASTAL BASALT PETROLOGIC TYPES		COLUMBIA RIVER BASALTS
PACK SACK LOOKOUT	- - - - - - - - - - - - -	POMONA (CLIFTON SANDSTONE)
CAPE FOULWEATHER (WHALE COVE SANDSTONE)	- - - - - - - - - - -	FRENCHMAN SPRINGS (VANTAGE MEMBER EQUIVALENT)
DEPOE BAY	- - - - - - - - - - - - - -	GRANDE RONDE

(modified from Snavely and others, 1973)

Figure 14

Beeson and others (1979). These ideas will be discussed later in the section on the origin of the basalts.

Stratigraphy

The basalts in the Nicolai Mountain-Gnat Creek area are divided into five formations (Plate I). They are the Grande Ronde, Wanapum, and Saddle Mountains Basalts of the Yakima Basalt Subgroup within the Columbia River Basalt Group, and the Depoe Bay and Cape Foulweather petrologic types of the coastal basalts (Fig. 15). Nowhere in the study area are all basalt units seen in continuous vertical section. The composite section (Fig. 15) is based on stratigraphic relationships developed from field mapping, measured section, and laboratory analyses. Seven partial sections evolved from this study (Plate II). They are: Bradley State Park, Aldrich Point-Clifton Road, Gnat Creek, Gnat Creek Gorge, Nicolai Mountain, Big Creek and Big Creek (west).

The Grande Ronde Basalt can be further subdivided on the basis of geochemistry (Figs. 16 and 17) and magnetic polarity. Three units and a subunit are tentatively defined: a low MgO unit with reverse polarity ($Tygr_1$- R_2). Stratigraphically above the Grande Ronde is a 2-70 meter thick sandstone that is correlative to the Vantage Member of the Ellensburg Formation of eastern Oregon and Washington. This sandstone is overlain by the Frenchman Springs Member (Tyfs), which is the only member of the Wanapum Basalt recognized in the study area (Fig. 15). The younger Saddle Mountains Basalt occurs stratigraphically above the Clifton formation which overlies the Frenchman Springs basalt. The Saddle Mountains Basalt is represented by the Pomona Member (Typ).

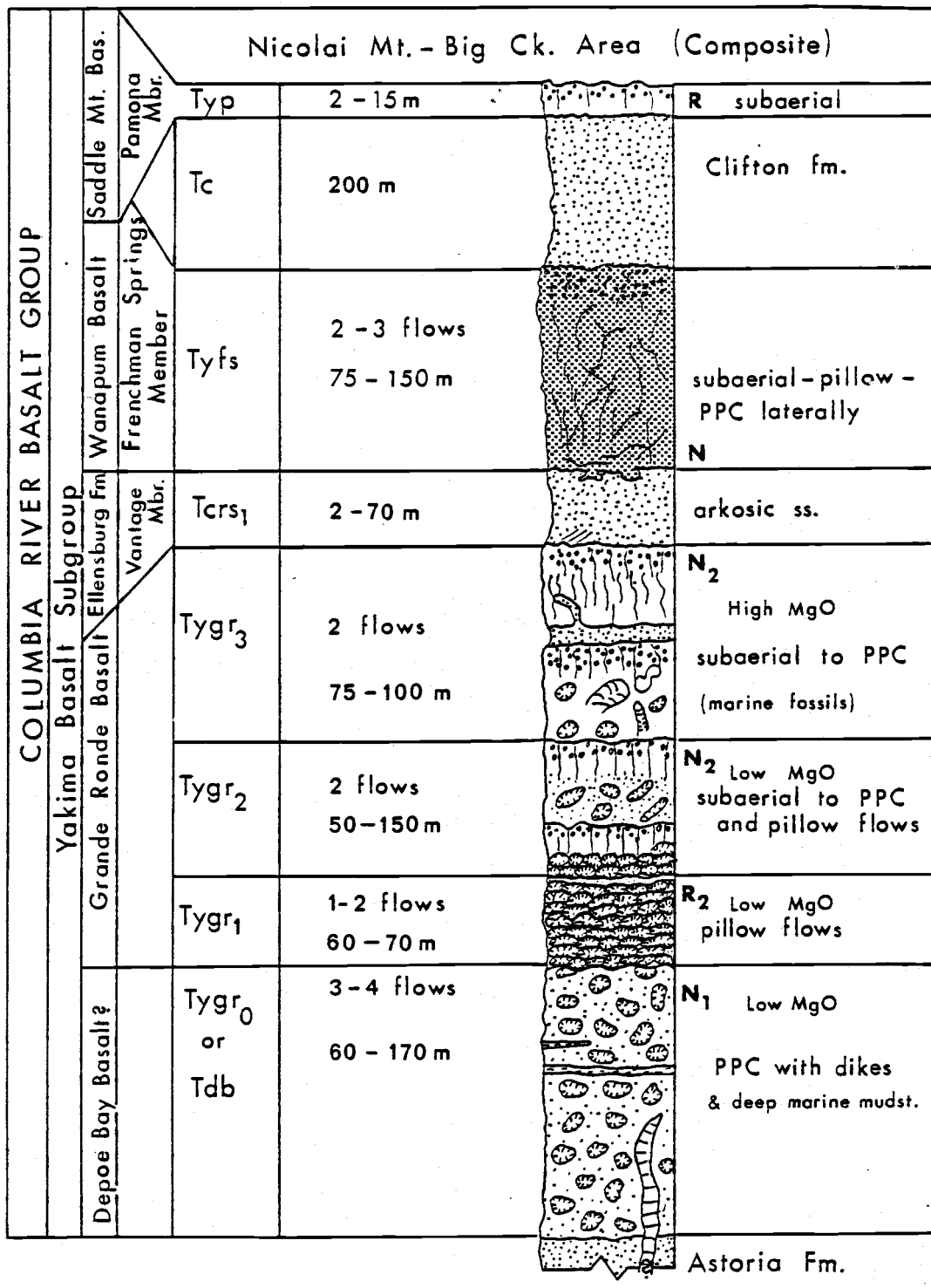


Figure 15 Composite section of Miocene basalt units and interbeds in this study with Cape Foulweather basalt not shown (N and R indicate normal and reverse polarity, respectively) (see Plate I for areal distribution).

The unit defined as Depoe Bay Basalt is confined to the lower reaches of Big Creek Gorge and is a low MgO unit with normal polarity (Tdb-N_1). A minimum of two and possibly four individual flows are recognized, separated by interbeds of deep marine mudstone (Fig. 15). These flows underlie the Grande Ronde Basalt. The Cape Foulweather Basalt (Tcf) is principally made up of sills and dikes with associated hyaloclastite breccias (Plate I). No individual flows of this unit were recognized. The stratigraphic position of the Cape Foulweather Basalt relative to its Columbia River Basalt equivalent, the Frenchman Springs Member, is uncertain. Depending on the geologic interpretation, the Cape Foulweather Basalt occurs below the Grande Ronde and Frenchman Springs Basalts but above the Depoe Bay basalt or the Cape Foulweather Basalt may be a lateral equivalent of the Frenchman Springs member.

Magnetostratigraphy

Remnant magnetism was measured from 82 basalt samples collected in the field (Appendix XI). The samples were oriented using a Brunton compass, and known geographic features. Actual measurement was performed in the laboratory with a portable fluxgate magnetometer using the method described by Dr. E.M. Taylor (personal communication, 1979) and Doell and Cox (1964). Reversed polarity events that were recorded were rechecked by running additional samples at the outcrop locality.

Two reversed and two normal polarity events are preserved in the flows in the Nicolai Mountain-Gnat Creek area (Fig. 15). The oldest reversed polarity event (R_2) is recorded in the low MgO flows of the Grande Ronde Basalts (Tygr$_1$); the other is associated with the Pomona Member of the Saddle Mountains Basalts (Typ) at the top of the section. These

polarity events divide the basalts into four magnetostratigraphic units. The lowest normally polarized unit is represented by low MgO Depoe Bay (Tdb) pillow flows and breccias in the Big Creek section. Stratigraphically above this is a reverse-polarized unit consisting of the low MgO pillow flow at the base of the Nicolai Mountain Section ($Tygr_1$, Plate II). This reverse-polarized unit is found nowhere else in the study area. A possible reverse-polarized pillow basalt was also noted on Porter Ridge to the southeast during reconnaissance along the eastern border of the study area. A thin arkosic sandstone interbed separates the reverse-polarized pillow basalt unit from the overlying normally-polarized basalt unit (N_2, Fig 15). The normally polarized Grande Ronde basalt sequence consists of low MgO flow ($Tygr_2$) and two overlying high MgO flows ($Tygr_3$, Fig 15). The normally polarized sequence occurs in the Bradley State Park section, in the Nicolai Mountain section, in the Big Creek section, and is thought to be present in the Gnat Creek section (Plate III).

The boundary between the reversed and normally polarized units ($Tygr_1$ & $_2$) of the Grande Ronde Basalt in this study is based on stratigraphic position, geochemistry, and polarity and is tentatively equated to the R_2/N_2 boundary outlined by Swanson and others (1979) for the Grande Ronde Formation (Fig 13). This magnetic polarity boundary is also present in the Clackamas River area in the Western Cascades in Oregon (Anderson, 1978).

Two units are distinguished in the thick sequence of low MgO basalt in Big Creek Gorge (section 6, Plate II); both units have normal polarity. The sequence was subdivided on the basis of physical characteristics and stratigraphic position into the underlying Depoe Bay Basalt (Tdb) and the overlying Grande

Ronde Basalt ($Tygr_2$) (see further discussion in Depoe Bay section). These two basalt units have been tentatively called two normal magnetic polarity units (N_1 and N_2) based on the probable projection of the intervening reverse-polarized units from Nicolai Mountain (section 5, Plate II). Unfortunately the stratigraphic relationship between the R_2 pillow basalt in the Nicolai Mountain section (Section 5, Plate II) and the N_1 Depoe Bay Basalt in Big Creek is not exposed. Therefore, alternative geologic interpretations of the magnetic stratigraphy are possible (see discussion of correlation of basalt units in the Nicolai Mountain section).

The Frenchman Springs Member is considered to be in a N_2 polarized unit (Fig. 13; Swanson and others, 1979). This is despite the presence of the thick Vantage sandstone equivalent between the lower N_2 Grande Ronde unit and the Frenchman Springs Member in this study (Fig 15).

The youngest reversed polarity magnetostratigraphic unit (R in Fig. 15) in the thesis area is the Pomona Member. It does not sequentially occur after the N_2 normally polarized unit because of the long period of time (represented by the Clifton Formation) in this study and the number of other polarity units recognized by Swanson and others (1979) between the N_2 and R magnetostratigraphic units on the Columbia Plateau.

Grande Ronde Basalts

Wright and others (1973) recognized two geochemical types within the Grande Ronde Basalt in eastern Oregon and Washington. These informal types are based on variations in major oxide abundances, principally MgO. The high MgO/low MgO division has since been recognized in the western part of the Columbia Plateau (Nathan and Fruchte, 1974) and more recently

in the Clackamas River area of the Western Cascades (Anderson, 1978). In western Oregon the high MgO type overlies the low MgO flows. This is not always the case in other parts of the Plateau, as high MgO flows can be found lower in the Grande Ronde Section. Anderson further defined the variation of major oxides and trace elements which occur between the two chemical types. This variation in chemistry, particularly the high MgO basalt overlying the low MgO basalt, has also been utilized to differentiate units in this study area.

Chemistry

The average chemical analysis of the major oxides of 18 samples of Grande Ronde Basalt from this study are listed in Table 1. Complete laboratory data and samples locations are tabulated in Appendix II. The summary data in Table 1 show that high MgO Grande Ronde basalts also contain more CaO and Al_2O_3 than the underlying low MgO basalts. However, the relative abundances of SiO_2 and K_2O are substantially less in the high MgO basalts. Smaller depletions were noted for FeO, TiO_2 and Na_2O.

Comparison of these values with the average values of Grande Ronde Basalts of eastern Oregon and Washington (Swanson and others, 1979) and from the Clackamas River area (Anderson, 1978) are also shown in Table 1. Some variation in chemistry should be expected between areas due to differences between laboratories that made the analyses, due to freshness of samples used, and due to the greater number of Grande Ronde flows that could be sampled on the plateau than in the study area. Inspection of the table reveals that the relative variation of major oxides between the high and low MgO basalts in the study area is generally consistent with the variations seen in the basalts of the other area . Therefore, geochemical

TABLE 1 Comparison of Major Oxides from Columbia Plateau, Clackamas River, Coast Range, and Nicolai Mountain-Gnat Creek Area for the Grande Ronde and Depoe Bay Chemical Types

	Average Grande Ronde[1]		Average Grande Ronde[2]		Average Depoe Bay[3]		Average** Grande Ronde[5]		Average** Depoe Bay[5]	Analytical Error %
	13 HMgO	8 LMgO	2 HMgO	4 LMgO	52	* LMgO[4]	11 HMgO	7 LMgO	6 LMgO	
SiO_2	54.00	56.20	53.85	55.60	56.00	N.D.	54.30	55.90	56.40	± 0.3
Al_2O_3	14.50	14.10	15.67	15.00	14.10	12.70	14.60	14.10	14.30	± 0.2
FeO	11.40	11.80	9.90	10.67	12.10	12.00	11.60	11.80	11.40	± 0.1
MgO	5.27	3.40	5.48	4.10	3.60	3.60	4.70	3.80	3.85	± 0.1
CaO	9.11	6.90	9.38	7.80	7.10	6.30	8.55	6.80	6.90	± 0.1
Na_2O	2.84	3.16	2.79	2.88	3.30	3.20	2.80	3.00	3.10	± 0.1
K_2O	1.05	2.00	1.03	1.65	1.40	1.50	1.20	1.90	1.70	± 0.05
TiO_2	1.78	2.28	1.90	2.28	2.00	1.90	1.93	2.15	2.04	± 0.05
Total	99.95	99.84	100.0	99.98	99.60	--	99.68	99.45	99.96	

* Not reported ⊥ number of samples

** Data listed in Appendix II

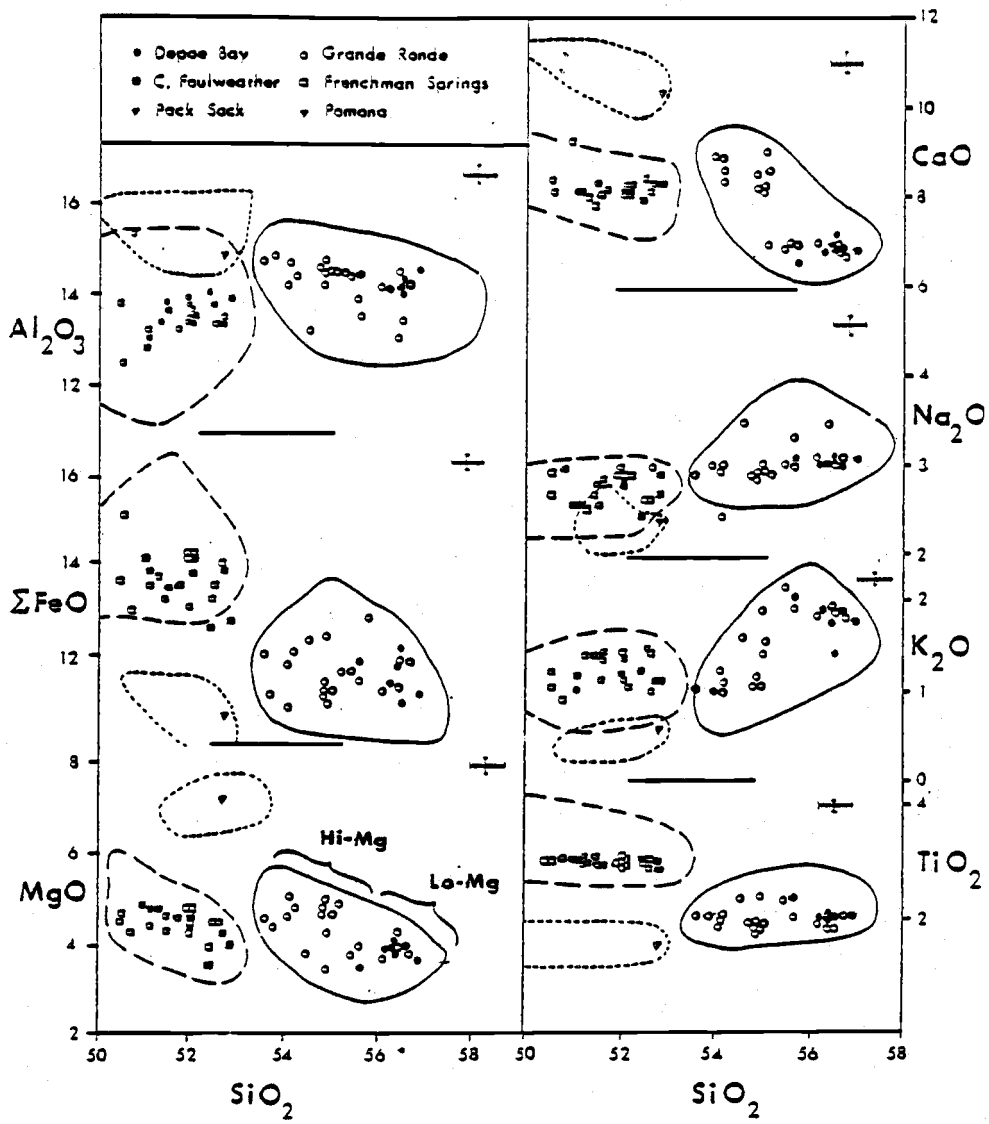
1. From Swanson and others (1979)
2. From the Clackamas River Area (Anderson, 1978)
3. From Snavely and others (1973). Combined both high and Low Mgo values.
4. From Hill (1974), used INAA methods
5. This study

correlation of units can be made. For some oxides most notably MgO and SiO_2, the amount of variation is less in basalts of the study area but still noticeable. When analytical error from table 1 is taken into account the actual abundances area also similar. Thus, because the chemical data agree so well, it is concluded that the high and low MgO Grande Ronde Basalt units are present in the Nicolai Mountain-Gnat Creek area.

Figure 16 is a silica variation diagram modified from Snavley and others (1973). It outlines the compositional fields of three major petrologic types in the Coast Range and their equivalent Yakima Basalt units. Inspection of the Depoe Bay-Grande Ronde field shows the separation of high and low MgO geochemical types depicted in Table 1. The segregation is best illustrated in the MgO, CaO, and K_2O versus SiO_2 graphs. This diagram also indicates that the two high MgO flows in the study area (Sections 1 & 2 on Plate II) may be distinguished from each other on the basis of percent SiO_2. This is suggested because data from two independent laboratories (O.S.U. and U.S.G.S.) show the same separation.

The variation of major oxides within a single flow can be seen by comparing results from samples 2, 3 and 4 or 5 and 6 of the Bradley State Park section (Section 1, Plate II; Appendix II). Samples 5 and 6 how variation which is within the experimental error of the method. Samples 2 and 4, taken from the top and bottom of a single flow ($Tygr_3$, Plate II), are also within the experimental error of the atomic absorption method. Sample 3, from the middle of the same flow, has a higher SiO_2 percentage than sample 2 and 4. This variation may be related to weathering or contamination during the analysis. The conclusion from these samples suggests that the variation within a flow is not very large.

Silica Variation Diagram - Miocene Coastal Basalts and Columbia River Basalts (Nicolai Mtn. area)



Compositional limits of Snavely & others (1973):

Depoe Bay

Cape Foulweather

Pack Sack

Error Bars

Figure 16

Relative abundances of trace elements are also diagnostic in defining the high MgO/low MgO boundary in the Grande Ronde Basalts. Anderson (1978) and Beeson and Moran (1979) have shown that minor elements Th, La, Sm, and Sc are useful for distinguishing these geochemical units in the Willamette Valley and Clackamas River drainage areas. The Th, La, and Sm decrease while Sc increases noticeably in high MgO flows relative to their abundances in low MgO flows. In the Nicolai Mountain-Gnat Creek area, this same trend is also seen; however, Sm values may overlap when experimental error is taken into account. Table 2 lists average values for selected trace elements from the Clackamas River area (Anderson, 1978), average values from the Columbia River Plateau (Swanson and Wright, 1979, preprint), and average values of five Grande Ronde samples from this study. A complete list of abundances for 17 trace elements for each sample analyzed in this study is presented in Appendix III. The relative trace element variations between high and low MgO types as well as actual absolute abundances appear remarkably close considering that three laboratories performed the analyses on different samples from different locations. The trace element data further confirm that the high and low MgO Grande Ronde Basalt units are present in the study area (as the major oxide chemistries indicated), and that the high MgO/low MgO horizon can be used as a stratigraphic boundary along with magnetic polarity and lithologic characteristics for correlation of these units over wide areas.

The high and low MgO boundary has been recognized in two stratigraphic sections within this study area. It is present low in the Bradley State Park section between a subaerial unit ($Tygr_2$) and a unit which consists of a pillow palagonite complex ($Tygr_3$) (See Plate II and Appendix I). The contact is a weathered oxidized zone suggesting subaerial exposure

TABLE 2

Comparison of Trace Element Averages from Columbia Plateau, Clackamas River, Coast Range and This Study for Grande Ronde and Depoe Bay Chemical Types.

	Grande Ronde[1]		Grande Ronde[2]		Depoe Bay[3]		Grande Ronde[4]		Depoe Bay[5]
	1 HMgO	1 LMgO	7 HMgO	17 LMgO	5 HMgO	3 LMgO	2 HMgO	3 LMgO	5 LMgO
Sc (ppm)	37.04	31.20	38.00	34.00	37.00	31.00	34.50	29.00	30.40
Co	41.30	36.40	43.00	39.00	38.00	35.00	35.60	35.00	32.60
Th	3.50	6.10	3.80	7.80	3.30	5.30	3.37	5.30	5.32
Ba	496.00	783.00	800.00	900.00	420.00	580.00	440.00	686.00	520.00
La	18.20	28.70	18.50	26.50	19.10	24.30	18.90	23.50	23.80
Sm	5.40	7.70	5.40	6.60	5.90	6.40	5.60	6.19	6.22
Eu	1.68	2.18	1.80	2.00	1.74	1.79	1.64	1.52	1.65
Ce	38.00	58.00	40.00	50.00	41.00	49.00	39.00	48.00	48.50

1. Swanson and Wright, (1979 preprint)
2. Anderson, 1978; Clackamas River Area
3. Hill, 1974
4. This study
5. This study

Analytical Error shown in Appendix III

INAA Methods used

Number of samples

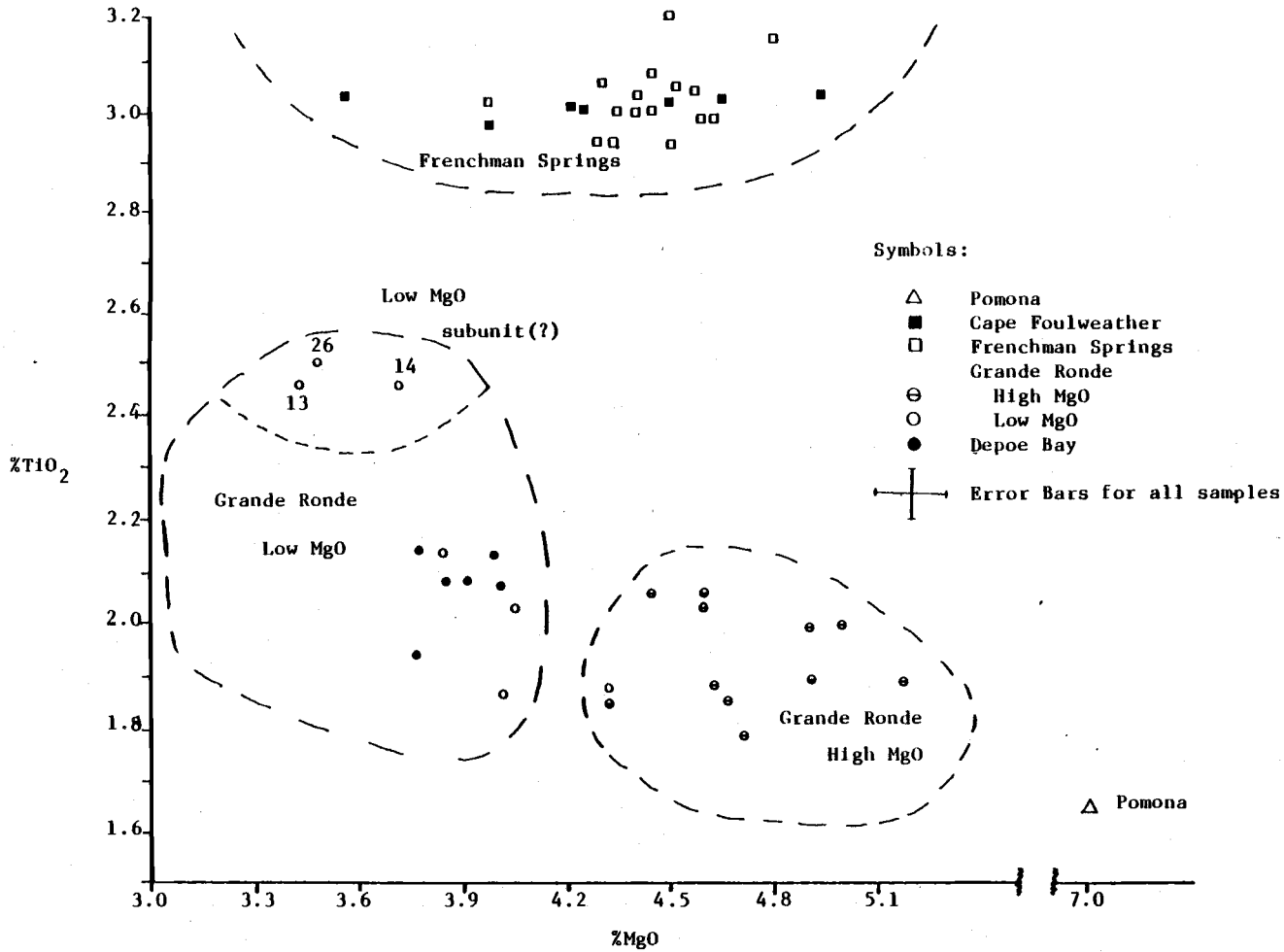


Figure 17 General fields for basalt types on a plot of TiO_2 vs. MgO.

prior to emplacement of the overlying high MgO flow. In the Nicolai Mountain section the horizon is not exposed but lies between the lower pillow flows in the quarry at OC 300 and the subaerial flows near the top of Nicolai Mountain (Plate II). A high MgO flow, similar to the basal high MgO flow in the Bradley State Park section, has been identified low in the Gnat Creek Gorge section (Plate II). Unfortunately, an analysis of the underlying subaerial flow in the latter section was not performed. However, physical characteristics and stratigraphic position suggest that the high MgO/low MgO horizon is present there. This analysis should be made at a later date for better correlation of this stratigraphic horizon. Since no high MgO flows were found in Big Creek Gorge, the horizon is not present.

Further variations within the low MgO basalts in the study area are suggested by differences in MgO, TiO_2, and SiO_2 values of the units in the Nicolai Mountain section. A plot of TiO_2 versus MgO values (Fig. 17) shows a wide separation of the pillow units in the Nicolai Mountain section from the subaerial low MgO flow in the Big Creek and Bradley State Park sections even though they were all emplaced during the same normal paleomagnetic interval, N_2. This separation tentatively suggests a geochemical subunit within the low MgO Grande Ronde unit in the study area. Long and others, (1980, unpublished) used a similar plot of TiO_2 versus MgO to define the Umtanum Subtype of the Grande Ronde Basalt in the Pasco Basin of south-central Washington. Further sampling and trace element data are needed to substantiate this separation.

Physical Characteristics of Grande Ronde Basalts

The physical features of the Columbia River Basalts in the Nicolai Mountain-Gnat Creek area can be divided into two main

categories: (1) those produced by emplacement under subaerial conditions and (2) those laid down in a subaqueous environment. Snavely and others (1973) postulated that the Columbia River Basalts entered a middle Miocene marine embayment with the strand line in or near this study area. This is evident because some flows can be traced chemically and by polarity from subaerial to subaqueous conditions laterally. This paleogeography limits the usefulness of the physical characteristics as correlation tools over regional distances, but it does work within the study area.

Swanson and Wright (1979, preprint) in their study of the Columbia River Basalts have concluded that physical characteristics are usually not reliabe for tracing individual flows on a regional scale. But most authors (Diery and McKee, 1969; Anderson, 1978; Swanson and Wright, 1979) will agree that the jointing character of pre-Vantage and post-Vantage subaerial flows differs significantly enough to be distinguishable and can be used on a local scale for correlation purposes.

In the Grande Ronde Formation, individual subaerial flows consist of well developed colonnade and entablature jointing, the latter comprising from 50 to 90 percent of the flow. The post-Vantage Frenchman Springs Member is characterized by cross-jointed columns and large irregular blocky joints (Anderson, 1978). These jointing styles are characteristic of the subaerial flows in this study (Figs. 18 and 19).

The Grande Ronde low MgO subaerial basalts are poorly exposed in the study area. Principal exposures occur at the base of the Bradley State Park section (section c-c', Appendix I), in the Nicolai Mountain section associated with the closely packed pillow basalts, and in the Big Creek section associated

with pillow breccias and colonnade basalt (Plates I and II). Large vesicles in the basalt and poorly developed large colonnade joints are typical of the low MgO basalt flows in these three areas. A red frothy oxidized zone, 1 to 2m thick, caps the flow in Bradley State Park and represents the boundary between the low and high MgO Grande Ronde Basalts (section e-e', Appendix I). The basal subaerial flow in the Gnat Creek Gorge section can be correlated to the low MgO flow in the Bradley State Park section based on similar jointing characteristics and the presence of the same high MgO pillow palagonite complex flow which overlies both units (Plate II).

The high MgO Grande Ronde flows are moderately exposed to well exposed in the Bradley State Park section and in Gnat Creek Gorge but poorly exposed above the quarry in the Nicolai Mountain section. High MgO flows also crop out in the precipitous cliffs on the east side of Nicolai Ridge above the landslide (QTls on Plate I). These basalts overlie the low MgO Grande Ronde flows and are overlain throughout the study area by the Vantage-equivalent sandstone and by the Frenchman Springs flows (Plates I, II, and III).

The field appearance of high MgO Grande Ronde flows in the study area is more characteristic of the general flow characteristics described by Diery and McKee (1969) for Yakima flows on the Columbia River Plateau. There are two high MgO subaerial flows which are 15 to 35m thick. Each displays well developed lower colonnade columns with middle undulatory to straight entablature columns that grade abruptly into vesicular upper colonnade columns (Fig. 18). This jointing pattern is best developed in the Bradley State Park section in the flow just below the Vantage sandstone Member (Plate II). The zone of entablature columns comprises approximately 70% of the flow. The well developed pattern of columnar joints suggests

that these units were ponded flows because these jointing characteristics normally form under static cooling conditions (Swanson and Wright, 1978). Also present in the base of a high MgO flow are oriented zeolite-filled pipe vesicles (2 cm. in diameter and 8 to 15 cm long), suggesting a flow direction to the southeast. A possible 10m high spiracle extends up into the basalt from a thin underlying sandstone interbed. Colonnade and entablature columns are also present in the flows in the Gnat Creek Gorge section where these flows form high water falls (Plate II). These physical characteristics along with the thin arkosic sandstone interbed, stratigraphic position and polarities are the basis for correlation of flows from the Bradley State Park section to the Gnat Creek Gorge section.

In the landslide area east of Nicolai Ridge (QTls Plate I), a 30m slide block from Nicolai Ridge provides a rare exposure of the top of a Grande Ronde flow with well developed pahoehoe flow texture (OC 75; Fig. 20). Preservation of this surface feature is testimony of the rapid burial by arkosic sandstone before erosion of the flow surface.

In comparison to the orderly colonnade and entablature columns of the high MgO Grande Ronde flows, the Frenchman Springs subaerial flows, above the Vantage sandstone Member, are generally characterized by disorderly block, hackly jointed, thick flows (75m) (Fig. 18 and 19). Also common in the Frenchman Springs are platy, equidimensional columns (0.5 to 1 m in diameter). These overlying flows are easily distinguishable in the field by the presence of large plagioclase phenocrysts within the finely crystalline groundmass.

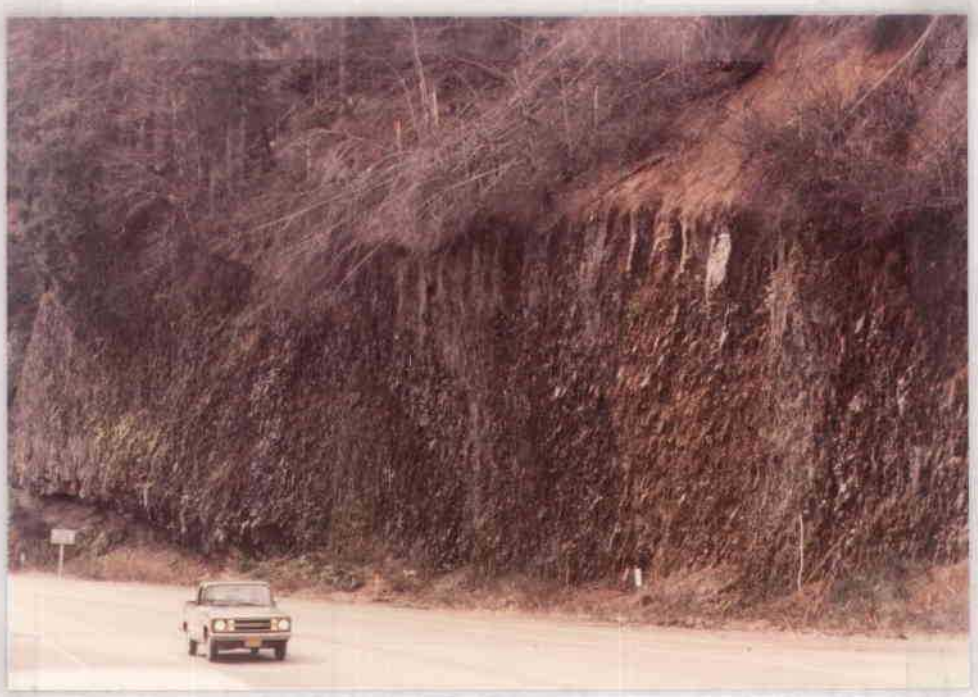


Figure 18 Pre-Vantage high MgO Grande Ronde Basalt flow. Note lower colonnade, middle hackly entablature, and well developed upper or pseudo colonnade joint sets. Exposure along Highway 30 (reference section e-e', Appendix I).

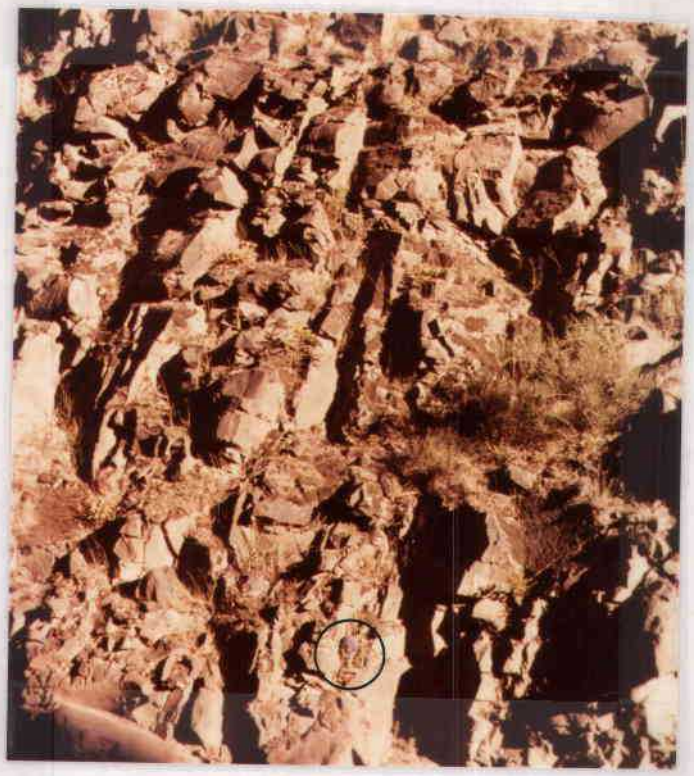


Figure 19 Typical blocky jointing of Post-Vantage Frenchman Springs Basalt. Exposure is along reference section e-e' on U.S. Highway 30.

Figure 20 Bob "Cupie" Ziak resting on a rarely exposed top of a subaerial Grande Ronde flow preserved in a landslide block. View is looking down onto top of flow. Note pahoehoe flow texture and spatter ridges. Exposure in quarry of basaltic colluvium (OC 75, Sec. 27, T8N, R6W).

The subaerial Pomona flow differs from other plateau-derived flows by its light gray color, abundant tiny phenocrysts, and highly weathered appearance. Pomona outcrops typically consist of spheroidal weathered blocks, up to 1 m in diameter, with orange weathering rinds (Fig. 20). One locality (OC 592) displays poorly developed irregular columnar jointing. Vesicles observed are large, irregular and widely spaced.

Subaqueous Grande Ronde Flows

Associated with the subaerial units are subaqueously deposited flows. These subaqueous flows consist of thick pillow palagonite (hyaloclastite) complexes (abbreviated ppc) (Swanson, 1967) and closely packed pillow lavas. Carlisle (1963) defined a series of volcanic terms which can further subdivide a subaqueous flow based on the dominating features of the flow. These terms include pillow breccias, isolated pillow breccias, broken pillow breccias, and bedded hyaloclastite breccias. The variety of breccia types produced as lava enters water is dependent upon the slope of the bottom, the viscosity of the lava, the flow rate of the material, and the amount of suspended sediment in the water. For example, closely-packed pillow basalts are thought to form on steeper slopes (Williams and McBirney, 1979) in relatively quiet, clear water while bedded hyaloclastite breccias form on shallow slopes in turbulent, vapor charged water. Isolated pillow and broken pillow flows are gradational between these extremes. Parameters such as slope, flow rate, and viscosity, may also vary as a lava flow moves through water. This suggests that a flow could change appearance laterally as it is being emplaced in a subaqueous environment. In Big Creek Gorge, variation within a subaqueous flow has been recognized vertically as well as laterally along the north cliff face. A general lateral

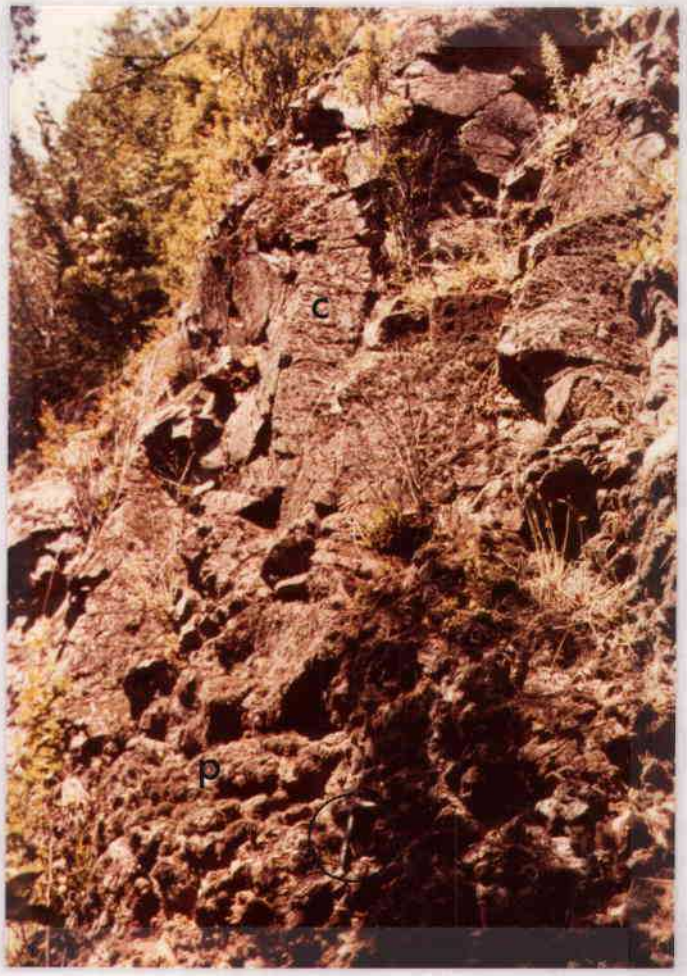


Figure 21 Transition zone between subaqueous flow-foot breccia and subaerial sheet flow in the Frenchman Springs Basalt. Note colonnade above and pillow palagonite below. The flow-foot is estimated to be 85 meters thick. Hammer in shade for scale. Big Creek Gorge (OC 449, Sec. 35, T8N, R7W).

change in the Grande Ronde from an isolated pillow breccia to a broken pillow breccia flow can be seen from east to west along the cliff face.

Pillow palagonite complexes are common at the margins of the Columbia River Plateau (Swanson and Wright, 1978). They formed where subaerial flows entered lakes, ponded behind older flows that dammed rivers. This relationship was noted as far back as 1902 when Russell described two flows on the Snake River Plain that advanced into shallow bodies of water (Fuller, 1931). Since then, Fuller (1931), Waters (1960), and more recently Jones (1968), and Moore and others (1973) have described sequences in which lava has flowed into water creating "lava deltas" (see Origin of Basalt section). In some cases , the trailing subaerial part of the flow may override the slowly prograding "deltaic" front breccia of the flow which has built up to the water surface. Therefore, a single flow can create a subaqueous-subaerial flow couplet upon entering a body of water. Jones and Nelson (1970) divided such a flow into three units: a flow-foot breccia which consists of inclined pillows and hyaloclastite material with depositional dips as high as 30°; a passage zone or transition zone with both subaqueous and subaerial characteristics; and a sheet lava which is the horizontal columnar jointed subaerial part of the flow. They point out that the contact between the subaqueous and subaerial parts of the flow could be misinterpreted as an unconformity between two separate flows.

Snavely and Wagner (1963) postulated that the subaerial flows originating from the Columbia Plateau entered a Miocene marine embayment in the vicinity of Big Creek (Plate I). This study documents that this is the case. Both the Grande Ronde and Frenchman Springs flows contain ppc and pillow lavas underlying subaerial flows (Fig. 21). These are interpreted as

the flow-foot breccias and the overlying sheet lavas of single flows. In the cliffs along Big Creek a low MgO Grande Ronde hyaloclastic breccia displays oriented isolated pillows (with dips of 30-35° to the west) in a palagonite breccia (Fig. 22). This unit is overlain by a nearly horizontal 20m thick columnar jointed part of the same flow. The passage zone of James and Nelson (1970) was not recognized in this flow. The subaerial portion of the flow thins out to the west suggesting the flow moved into the basin from the east.

The Grande Ronde pillow palagonite complexes in Big Creek are composed of vesicular to dense, mostly vesicular fragments of dark tachylyte glass, which are pebble to fist size, within a yellow to bleached orange palagonite groundmass. Pillows, 0.3 m to 1 m in diameter, are common. Rarely bedding (1 - 10m thick) is observed, marked by oriented pillows or rarely by a differentiation of fragment sizes within the hyaloclastite lavas.

Over 100 meters of low MgO Grande Ronde closely packed pillows are exposed in the lower half of the Nicolai Mountain section (Plate II). The pillows are irregular ellipsoidal (1 to 1.5 m in diameter) in cross section with well developed chill margins (1-2 cm thick) and small ridges of tachylyte glass on the exterior rind (Fig. 23). Moore (1975) suggested that these ridges are a ubiquitous feature of fresh pillows on the ocean floor and that they would rarely be preserved due to their fragile nature and composition. Thin sandstone interbeds separate at least two or three pillow flows in this section. The upper sandstone represents the R_2/N_2 boundary described in the magnetostratigraphic section. The upper pillow flow is overlain by an associated vesicular colonnade subaerial basalt, suggesting possibly another lava "delta" with emersion and build up above sea level.

Figure 22 Longitudinal view of oriented pillow in a pillow palagonite complex (flow-foot breccia) of low MgO Grande Ronde Basalt. Angle of dip is 35° to the west (OC 652, NW, Sec. 32, T8N, R7W).

Figure 23 Close-up of closely packed pillow basalt in low MgO reverse-polarity Grande Ronde Basalt. Reddish brown is altered pillow rinds (OC 300, NW, Sec. 21, T8N, R7W).

The high MgO Grande Ronde basalt flow in Bradley State Park and in Gnat Creek Gorge is a pillow palagonite complex which again built up above sea level. The ppc part of the flow is characterized by irregular, massive basalt-filled tubes (?) up to 3 m wide and 7 m in length, isolated pillows, and broken pillows in a yellow orange matrix of hyaloclastite breccia. The irregularity of the massive basalt may suggest that a large volume of lava flowed into water rapidly with little time to develop any coherent pattern of emplacement as "delta" lavas. The presence of an echinoid coquina (See Sedimentary Interbeds Section), described as "definitely marine" by Ellen Moore (written communication, 1980) between pillows within this unit in Gnat Creek Gorge, indicate that the lava flowed into the sea.

Petrography of Grande Ronde and Depoe Bay Basalts

Regionally, the mineralogy of the Plateau-derived Grande Ronde Basalt has been shown to be the same as the Depoe Bay coastal basalts by Snavely and others (1973). There appears to be no exception to this generalization in the study area. Therefore, these equivalent units will be discussed together.

Generally, the subaerial Grande Ronde and intrusive Depoe Bay basalts are dark gray (N3) to grayish black (N2), aphanitic to fine-grained crystalline rocks. These basalts weather to a dark reddish brown (10R 3/4); vesicles, forming from 1 to 30% of the rock, occur in a variety of shapes and sizes.

In thin section, the Grande Ronde subaerial and pillow basalt flows and the Depoe Bay sills and dikes are characterized by hypocrystalline, intersertal to hyalopilitic textures. The more crystalline basalts are composed of 40 to 55% plagioclase, 25-30% clinopyroxene, 5-10% opaques, and 0-1%

olivine with varying amounts of glass and mineraloids. Clinopyroxenes are principally represented by augite. Pigeonite in the groundmass was recognized by Snavely and others (1973). Some glomeromicrophenocrysts (< 3mm) of plagioclase with pyroxene are present. Light brown sideromelane and dark brown to black tachylyte along with yellowish palagonite glass are common constituents of the Depoe Bay and Grande Ronde hyaloclastite breccias. The glass forms 20 - 70% of the rock, increasing in abundance in the more rapidly chilled parts of the flow. This characteristic sets the Yakima flows (Grande Ronde) apart from the underlying nearly holocrystalline Picture Gorge Basalt (Waters, 1961). Alteration minerals include chlorophaeite, celadonite, and nontronite (?). Calcite is common as a pseudomorph of pyroxene, filling vesicles and replacing the groundmass. Zeolites and iron oxides were noted cementing the palagonite breccia fragments. Also observed were inclusions of apatite in the plagioclase microphenocrysts.

In general, the Grande Ronde Basalt flows have a finer groundmass than the Frenchman Springs which is still finer crystalline than the Pomona flow in the thesis area (see Frenchman Springs and Pomona members sections). Glomeromicrophenocrysts were recognized in thin section in the Grande Ronde Basalt-Depoe Bay petrologic type and in the Pomona Member but not in the Frenchman Springs-Cape Foulweather petrologic units. Large phenocrysts and glomerophenocrysts of plagioclase are ubiquitous only in the Frenchman Springs and Cape Foulweather Basalts.

In summary, the petrography of the Grande Ronde and Depoe Bay basalts is useful for separating the two units from the other Columbia River Basalts and coastal basalts petrologic types only in a broad sense. There appears to be little

petrologic distinction which can be made between the equivalent units of coastal basalts and the plateau-derived units when comparing flows of similar crystalline character.

For a more complete discussion of the thin section petrography of the Columbia River Basalt units see Waters (1961), Swanson (1967), Wright and others (1973). The coastal basalts have been described by Snavely and others (1973), Tolson (1976), Penoyer (1977), and Coryell (1978).

The Frenchman Springs Member of the Wanapum Basalt

The Frenchman Springs Member (Tyfs) is the most widely exposed basalt unit in the study area, covering the majority of the Nicolai Mountain cuesta (over 20 sq. miles) (Plate I). This unit can be correlated with the Frenchman Springs in the Columbia Plateau region by geochemistry, magnetic polarity, and physical characteristics. Table 3 illustrates the similarities of the average values of major oxides between the Frenchman Springs flows in the Plateau from Swanson and others (1979) and the Frenchman Springs flows in this study area. Trace element abundances, shown in Table 4, from the Plateau (Swanson and Wright, 1979, preprint), from the Clackamas River area (Anderson, 1978), and from this study further confirm this geochemical similarity. Snavely and others (1973) suggested that the flows which cap Nicolai Mountain originated from vents in the Columbia Plateau and flowed down an ancestral Columbia River Valley at least as far as this thesis area. The geochemistry, paleomagnetism, and physical characteristics of the Frenchman Springs flows in this study area substanitates that hypothesis. Coryell (1978) described a subaerial Cape

Foulweather unit on top of Wickiup Mountain. This may be the westward extension of the Frenchman Springs Basalt, but further work is needed for verification of this correlation.

Chemistry

The Frenchman Springs can be distinguished from the Grande Ronde Basalts in the study area by the lower percentage of SiO_2, and higher percentage of FeO and TiO_2 (compare Tables 1 and 3; Fig. 16). The higher TiO_2 values for the Frenchman Springs relative to the Grande Ronde has been recognized by numerous workers of the Columbia Plateau (Seims and others, 1974). Similar discontinuities in TiO_2 content between Frenchman Springs and Pomona flows have also been reported (Schmincke, 1967). The variation in TiO_2 content are graphically presented in Fig. 17.

Beeson and Moran (1979) suggest that the variation in abundances of four trace elements (Sm, La, Sc, and Eu) is useful for distinguishing between chemical types of the Columbia River Basalts in estern Oregon. They indicated that values of Sm (7ppm), La (between 25 and 30ppm), Sc (between 35 and 40ppm), and Eu (2.5ppm) were characteristic for the Frenchman Springs Member. The average values listed in Table 4 for Frenchman Springs from this study are within these limits with the exception of La. The low values for La may be attributed to variation between laboratories, as all chemical types analyzed in the study shows low abundances of La (Appendix III).

No chemical distinction between Frenchman Springs and Cape Foulweather basalts was recognized in this study (Figs. 16 and

TABLE 3

Comparison of Average Major Oxides with other Average Concentrations of Frenchman Springs and Cape Foulweather Chemical Types

	21 Frenchman Springs$_1$	8 Frenchman Springs$_2$	21 Cape Foul-weather$_3$	15 Frenchman Springs$_4$***	7 Cape Foul-weather$_5$
SiO_2 (%)	51.5	52.78	52.46	51.65	52.0
Al_2O_3	14.15	13.33	14.00	13.66	13.48
FeO	14.20	14.50	14.14	13.70	13.60
MgO	4.60	4.08	4.14	4.78	4.34
CaO	8.64	8.00	7.98	8.26	8.16
Na_2O	2.66	2.70	3.03	2.80	2.66
K_2O	1.20	1.42	1.01	1.27	1.32
TiO_2	2.95	3.20	3.03	3.02	3.01
Total	99.90	100.01	99.79	99.14	98.57

1. Swanson and others, 1979
2. Wright and others, 1973
3. Snavely and others, 1973
4. This report
5. This report

Number of samples

Analytical error shown on Table 1

*** Includes 5 samples from Snavely and others (1973), Frenchman Springs samples within study area, and one sample from Coryell (1978).

TABLE 4

Comparison of Na, Fe and Selected Trace Element Abundances of this study with other Average Values for Frenchman Springs and Cape Foulweather Chemical Types

	[**] Frenchman Springs$_1$	[6] Frenchman Springs$_2$***	[7] Cape Foul-weather$_3$	[5] Frenchman Springs$_4$	[3] Cape Foul-weather$_5$
Na%	--	2.03	2.20	2.05	2.10
Fe	--	10.80	10.70	10.20	10.80
La(ppm)	26.50	25.60	24.30	22.80	22.80
Sm	7.20	7.40	7.70	7.10	7.13
Ce	52.50	53.00	52.00	51.00	49.00
Eu	2.27	2.20	2.28	2.07	2.09
Lu	0.63	0.64	0.53	0.53	0.53
Th	3.70	5.50	3.90	3.82	3.83
Hf	4.35	5.40	4.80	4.50	4.50
Co	39.40	42.80	39.00	37.80	38.60
Sc	36.40	37.20	36.00	35.00	35.60
Ba	564.00	770.00	480.00	488.00	491.60

** Not reported
*** From Anderson, (1978)

1. Swanson and Wright, preprint
2. Clackamas River Area (Three Lynx Section)
3. Hill, 1974
4. This study
5. This study

[ ] Number of samples

Analytical error listed in Appendix III

17; Tables 3 and 4). Average rare earth values standardized to condrites also shows the similarity between these basalt units (Appendix III).

Physical Characteristics

The main physical characteristic of the Frenchman Springs Member is the presence of large yellow plagioclase phenocrysts scattered throughout the unit. Anderson (1978) was able to further subdivide the Frenchman Springs flows in the Clackamas River area and attempted to correlate them to flows in the Plateau based on variation in abundance of plagioclase phenocrysts. The results appear somewhat inconclusive and therefore this method was not attempted here. However, the subaerial Frenchman Springs flows have roughly 1 - 6% phenocrysts in the study area. Another reason for not using this method stems from the difficulty in determining the abundance of phenocrysts in pillow palagonite breccia flows.

In the study area, the Frenchman Springs member which is 75 - 110m thick is composed of both subaerial and subaqueous basalt. The subaerial flows are typically moderately crystalline, irregular, and blocky with colonnade and rarely platy jointing. Individual columns average 1 - 2m in diameter (Fig. 19). The subaqueous part of the member is a pillow palagonite complex composed of isolated pillow lavas, broken pillow lavas, and hyaloclastite breccias. There are two or possibly three flows of Frenchman Springs present in the thesis area. These flows are geochemically similar, display normal polarity and as suggested previously, are not divisible on the basis of percent plagioclase phenocrysts. The basis for separation of flow units is the presence of thin sandstone interbeds between them. The flows are exposed in the steep cliffs along the length of Nicolai Ridge from the Columbia

River to the top of Nicolai Mountain. Access to these exposures is best at the top of Nicolai Mountain and where U.S. Highway 30 cuts across the ridge (Bradley State Park Section, Plate II). In the southwest part of the study area, the Frenchman Springs is well exposed in the precipitous cliffs along the north side of Big Creek Gorge. Individual flows are up to 75m thick in Bradley State Park area whereas the total thickness of the member is approximately 110m in Big Creek Gorge.

The Frenchman Springs unconformably overlies the Vantage Member and is overlain by the Clifton formation. The contact with the Vantage sandstone is sharp to irregular, displaying evidence that the lava splayed into the unconsolidated fluvial sandstone with maximum penetration of 1 meter. This contact can be traced discontinuously along Nicolai Ridge to the top of Nicolai Mountain and as far west as Big Creek Gorge (Plate I). A basalt conglomerate of the Clifton formation unconformably overlies the Frenchman Springs. This relationship is best exposed in the woods and cliffs along U.S. Highway 30 at the Bradley State Park section (OC 355) and along Clifton Road (OC18, Plate I).

In the Bradley State Park section (Plate II), only one Frenchman Springs basalt is recognized. The massive basalt is characterized by a blocky to columnar jointed, porphyritic subaerial flow. This basalt fractures easily when struck with a hammer and shows an increase in irregular shaped vesicles toward the top of the flow. In the Nicolai Mountain section, the member displays a platy columnar jointed character. A few arkosic sandstone interbeds on the Nicolai cuesta suggest the presence of more than one flow in the Frenchman Springs Member. A 15 to 20cm thick basaltic sandstone interbed occurs between two subaerial flows west of Bradley State Park in the

Gnat Creek section (Plate II). This sequence is exposed in a stretch of Gnat Creek north of the fish hatchery (OC 614, 19, T8N, R6W). To the south and west, Frenchman Springs subaerial flows are poorly exposed over much of the Nicolai homocline. The member forms basaltic red soil on the relatively flat forested topography and is principally exposed in small gravel quarries and road cuts. In Big Creek Gorge a thick subaerial Frenchman Springs basalt is exposed along the rim of the gorge (Plates I, II, and III). The western exposure is directly above the Tillusqua Fish Hatchery. The vesicular subaerial basalt in this section is characterized by well developed colonnade and platy joints. The subaerial part of the flow ranges from 10 to 50 meters in thickness and is underlain by a subaqueous part.

Subaqueous Frenchman Springs Flows

The subaqueous part of the Frenchman Springs creates a bench in the nearly vertical precipitous cliffs along the north side of Big Creek Gorge. The basalt is a pillow palagonite complex, approximately 75-100m thick, and composed of isolated pillow breccias, broken pillow breccias, rare bedded hyaloclastite breccias, and closely packed pillow basalts. These very poorly sorted breccias are made up of angular, highly vesicular to dense fragments (1mm to 15cm) within a palagonite matrix.

Characteristic of the subaqueous flows are a series of elongate pillows which dip as much as 37°, suggesting a general flow direction to the west. Based on 28 readings of pillow longitudinal sections from a closely packed pillow basalt, an average flow direction of S 51° W is indicated for the subaqueous portion of the Frenchman Springs Member in the Big Creek Gorge.

This pillow palagonite complex unit is underlain by a thin arkosic sandstone (OC 169) which has been correlated with the Vantage Member (Plate II). The contact between the subaerial and subaqueous portions of the Frenchman Springs flow varies from sharp to gradational with the subaerial columns extending down into the palagonitized breccia for 1 - 3 meters. The contact can be traced the length of the gorge, rising in elevation from 1310 feet (403m) in the west to 1775 feet (546m) in the east. There is no break or oxidized zone between the subaerial and subaqueous parts. This relationship implies that Frenchman Springs lava from the Plateau poured into a body of water in the vicinity of Big Creek producing a "lava delta" which was eventually covered by subaerially emplaced lava from the same flow (Fig. 21). This is the same process which formed the subaqueous breccia-subaerial basalt flow-couplet of Grande Ronde affinity exposed stratigraphically lower in a bench in Big Creek Gorge. Waters (1960) described a similar subaqueous-subaerial origin for a single Columbia River flow exposed in the Columbia River Gorge near The Dalles, Oregon. Fuller (1931) noted dips of 30° for foreset breccias that were overlain by the subaerial part of the flow near Moses Coulee along the Columbia River.

The subaqueous/subaerial contact is also inferred to be 3-4 km to the north of Big Creek Gorge in Rock Creek (Secs. 22, and 27, T7N, R8W) where closely packed pillow lavas are overlain by the subaerial columnar jointed part of the basalt flow. Correlation of the massive, thick subaerial flow at Bradley State Park and the subaerial-pillow palagonite breccia flow couplet in Big Creek Gorge suggests that a Frenchman Springs lava entered a Miocene marine embayment in the Big Creek area (Plate II).

A second pillow palagonite complex flow caps the 1865 foot (574m) hill along the north side of Big Creek Gorge (Sec. 35, T8N, R7W, Plate I and A-A' Plate III). A thin arkosic sandstone interbed separates this flow from the underlying subaerial-ppc flow couplet discussed above. This is the only place where this overlying subaqueous flow has been recognized. This flow is characterized by well developed radially jointed closely packed pillows up to 6m in length and 2m in diameter. The flow may correlate with the upper subaerial Frenchman Springs flow recognized in the Gnat Creek section.

Petrography

The Frenchman Springs and Cape Foulweather basalts are differentiated from the older Grande Ronde and Depoe Bay basalts by the presence of amber tinted plagioclase phenocrysts, up to 2cm in length. These dark gray (N2) to grayish black (N1) basalts have vitrophyric, hyalopilitic, or intersertal textures depending on the relative abundance of crystals versus glass. They are composed of the same mineral constituents as the Grande Ronde and Depoe Bay basalts with the major exceptions being a slightly coarser crystallinity in the subaerial or intrusive portions and the presence of the plagioclase phenocrysts and olivine. The Frenchman Springs and Cape Foulweather lavas in the study area consists of: 35 - 40% plagioclase, 20 - 30% pyroxene, 1 - 5% olivine, 5 - 10% opaques, glass and mineraloids. The glass component in the samples studied ranges for 30 - 80% of the rock; glass is most abundant in the hyaloclastite breccias. Euhedral plagioclase phenocrysts, comprising 1-5% of the rock, are labradorite (An_{61-65}). The plagioclase microlites in the groundmass form from 10 - 40% of the rock and are composed of andesine (An_{48-55}). The pyroxene is principally augite. It is mainly

an interstitial component in the groundmass although a few microphenocrysts are present. The glass is mainly sideromelane which is commonly altered to yellowish isotropic palagonite in the hyaloclastite pillow breccias of which it is the dominant component. Tachylyte is present in the rapidly chilled margins of dikes and sills of Cape Foulweather Basalt. Olivine was recognized in trace amounts commonly associated with magnetite the principal opaque mineral. Alteration products include chlorophaeite which commonly ranges up to 10% of the basalts, calcite, celadonite, and zeolites.

For more detailed discussion on the thin section petrography of these basalt units see Wright and others (1973) and Snavely and others (1973).

In summary, the Frenchman Springs Member of this study appears to be similar in all respects to its counterparts in the Columbia Plateau and in the Clackamas River area of the western Cascades. This conclusion is based on major and trace element chemistries, magnetic polarity, and physical characteristics, in particular the presence of large plagioclase phenocrysts. No distinction could be made between Frenchman Springs and Cape Foulweather basalts in this study area based on chemistry, polarity, or petrography. The separation is based on method of emplacement (intrusive and subaqueous vs. subaerial and subaqueous) and association of sedimentary interbeds (deep marine mudstone vs. fluvial to shallow marine). It appears that the Frenchman Springs did flow into a Miocene marine embayment in this study area.

Depoe Bay Petrologic Type

Exposures of the Depoe Bay petrologic type in the Nicolai Mountain-Gnat Creek area are restricted to the lower levels of

Big Creek Gorge and upper part of Tripp Creek (Plate I). This unit, approximately 160m thick, consists of pillow palagonite complexes (Tdbe) similar to those present in the subaqueous part of the Grande Ronde and Frenchman Springs basalts. In addition, Depoe Bay basalts also occur as dikes (Tdb_d), which cut through the breccia piles.

Overlying the Depoe Bay units in Big Creek Gorge is a ppc-subaerial flow couplet assigned to the Grande Ronde Basalt. Differentiation of the Grande Ronde Basalt from the Depoe Bay is based on the presence of thin, fossiliferous deep-water mudstone interbeds and "feeder" dikes within the stratigraphically lower Depoe Bay basalt breccias. The Grande Ronde (ppc) in Big Creek has westward oriented pillows which are overlain by subaerially emplaced basalt capped by the Vantage sandstone. These features as well as the lack of extensive subaerial flows in the Depoe Bay units to the southwest of this study area (Coryell, 1978; Penoyer, 1977; Neel, 1976; and Tolson,1976) suggest that the lower unit is a Depoe Bay equivalent.

Chemistry

Averages of major oxides for six Depoe Bay basalt samples from Big Creek Gorge are listed in Table 1. These values show that the Depoe Bay in this study are all low MgO types. Comparison of these values with regional averages for the Grande Ronde and Depoe Bay shows a close similarity between the Depoe Bay basalt in this study area and its Plateau and coastal equivalents with one exception; slightly higher MgO values are noted in the Depoe Bay of this study. The chemical differences between the Depoe Bay and high MgO Grande Ronde basalts is more noticeable. In addition to MgO being lower in the Depoe Bay, CaO is lower, and SiO_2 and K_2O are higher (Table 1 and Fig.

17). Inspection of the data suggest that the differences in the major oxides are insufficient to separate the Depoe Bay from the low MgO Grande Ronde basalts (Figs. 16 and 17).

In Table 2, the trace elements in the Depoe Bay basalts in this study can be compared to the trace element data of the low MgO Depoe Bay basalts of the Oregon Coast Range (Hill, 1974), the low MgO Grande Ronde basalts of the Columbia Plateau (Swanson and others, 1979), and the low MgO Grande Ronde basalts in the Clackamas River area (Anderson, 1978). There is very little variation evident between these geographically distinct basalts. A similar conclusion is reached when comparisons are made between the Grande Ronde low MgO units and the Depoe Bay within this study, however barium (Ba) appears low in all the Depoe Bay units when compared with Grande Ronde. The trace elements of the Depoe Bay basalt, with possibly the exception of Eu, are distinctly different when compared with the high MgO Grande Ronde flows in the study area and regionally.

Magnetic Polarity

The Depoe Bay basalt, breccias and dikes in the study area, display normal polarity (Appendix IX). Reconnaissance of selected Depoe Bay dikes, sills, and breccia units to the southwest of the study area in Penoyer's (1977) and Coryell's (1978) thesis areas by Dr. Niem and myself showed normal polarities. Choiniere and Swanson (1979) reported two reversed samples in the Depoe Bay flows at Maxwell Point, north of Oceanside, Oregon and four normal samples at Cape Lookout, Oregon. They suggested that the reversed units lie stratigraphically below the normally polarized basalts at Cape Lookout based on relative position above sea level. The

stratigraphic position of the Depoe Bay Basalt in this study relative to the reversed unit at Maxwell Point is unknown at this time.

Tentatively, the Depoe Bay petrologic type is placed in the N_1 magnetostratigraphic unit below the R_2 in the Nicolai Mountain section (Fig. 15). This is based on the presence of bathyal foram, diatom and mollusk bearing mudstone within the Depoe Bay basalt at Big Creek and the unconformable contact. Thin, irregular bifurcating dikes, 1 to 2m wide, with horizontal jointing are also well exposed parallel to the Big Creek logging road at the east end of the gorge (Fig. 24).

Physical Characteristics

The main feature that distinguishes the Depoe Bay Basalt from Grande Ronde in this study is the presence of dikes cutting through the Depoe Bay pillow palagonite breccias. These dikes are composed of massive hotizontally jointed basalt with thin (1 to 2 cm) cooling selvages of glass. They range from 1 to 11 m thick although most are within the 1-4m range (Fig. 24). A large 10 m thick vertical dike is exposed at the bridge crossing Big Creek in the gorge (OC-131, Sec. 33, T8N, R7W). It consists of massive, blocky jointed basalt that is surrounded by vesicular, frothy, basaltic breccias. The dike extends upward from the stream bed, cutting apparent crude bedding planes in the breccias and in the vertical walls on both sides of the canyon for at least 40m before disappearing into the breccia. The vesicularity of the breccia associated with this dike may be suggestive of a "feeder" system in which molten material expelled gas very rapidly as it cooled in an aqueous solution. If this basalt had originated east of the Cascades, one would expect that more of the gas would have already been expelled before emplacement.

A second, large (11 m thick), vertical dike occurs in Tripp Creek (OC360, Sec. 31, T8N, R7W). It is not as well exposed as the dike in Big Creek, but it can be traced in contact with the broken pillow breccia for 0.8 km. This dike is similar to the one in Big Creek, but the surrounding Depoe Bay breccia does not show the vesicularity.

Dikes are not as well exposed in the upper part of the section in Big Creek Gorge. There, the Depoe Bay basalts consist of less vesicular, isolated pillow and broken pillow breccias as well as coarse, thick, structureless breccias (OC 659, Sec. 33, T8N, R7W). Rare mudstone interbeds (0.3 - 2 m thick) occur within these breccia piles such as in the quarry in the west end of Big Creek Gorge (OC 125 and 646, Sec. 29, T8N,R7W). These mudstones were deposited in bathyal water depths (250-1000m) based on Foraminifera identified by Weldon Rau (written communication, 1980) (see Appendix IV). The mudstone is baked only along the contact with the overlying breccias, suggesting that the mudstone was laid down after the underlying closely packed pillow flow. The presence of two mudstone layers suggests that there are three or more Depoe Bay flows within the study area. The lack of dikes in the upper Depoe Bay units of the gorge may be the result of poor exposure due to the forest cover. The breccia units tend to form exposed vertical cliffs while the more massive blocky basalt has a greater tendency to mass waste, producing the vegetation covered slopes.

The thin section petrography of the Depoe Bay Basalt is discussed in the section on petrography of the Grande Ronde Basalt.

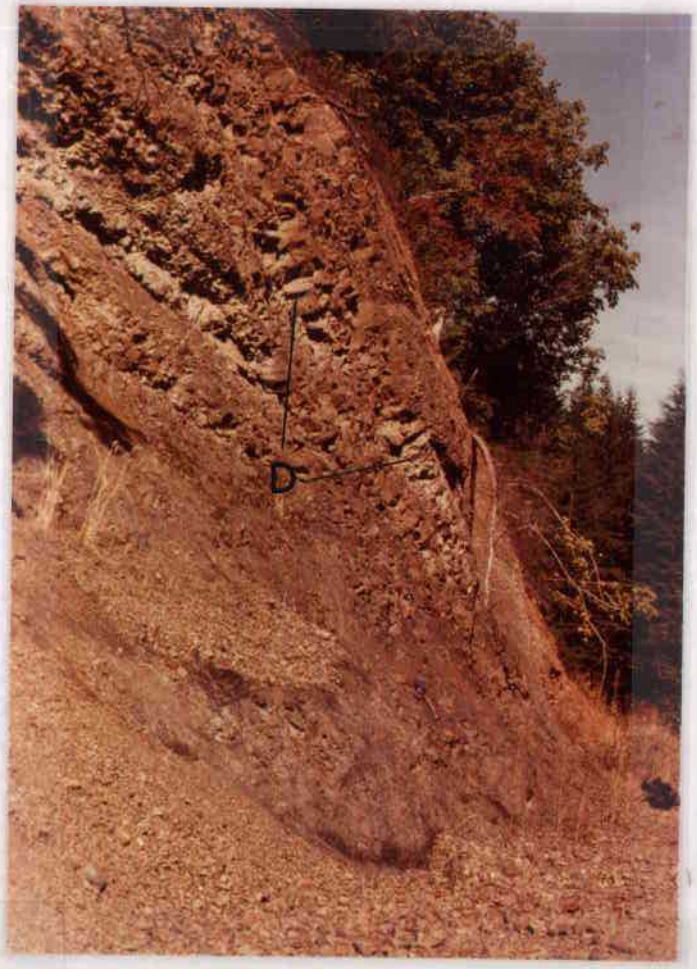


Figure 24 Depoe Bay dike intruding extrusive Depoe Bay pillow palagonite complex in Big Creek Gorge (OC 144, NE, Sec. 33, T8N, R7W).

Figure 25 Cape Foulweather sill intruding mudstone in Big Creek Gorge. Note irregular baked contact of mudstone. Mudstone in center is 2 meters thick (OC 127, SE, Sec. 29, T8N, R7W).

In summary, the major criteria for differentiation of the Depoe Bay Basalt from the low MgO normal Grande Ronde Basalt is the presence of dikes, the lack of oriented pillow units, and the interbeds of deep water mudstones in the Depoe Bay basalt. Geochemically, the basalts vary little in major oxides trace elements from either the Grande Ronde of this study or the average values of either Grande Ronde or Depoe Bay determined regionally by others. The mudston interbeds suggest that the Depoe Bay flows were emplaced at bathyal depths. This is in contrast to the fluvial to very shallow marine(?) interbeds between flows in the Grande Ronde Basalt higher in the Big Gorge Section and Bradley State Park area. The remnant magnetism of the Depoe Bay unit indicates normal polarities within the study area. There is insufficient magnetostratigraphic control on the Depoe Bay Basalt along the coast to determine the exact position of the Depoe Bay on the Astoria Formation in this study area and nearby Wickiup Mt. (Coryell, 1978). An alternative hypothesis will be discussed in the internal stratigraphy section.

Cape Foulweather Petrologic Type

Cape Foulweather Basalt , like Depoe Bay Basalt is a petrologic field term. It is applied to sparsely porphyritic middle Miocene dikes (Tcfd), sills (Tcfs), and extrusive subaqueous breccias (Tcfe) which are petrologically and chemically similar to the basalts at Cape Foulweather, Oregon (Snavely and others, 1973). These basalts display normal polarities (Appendix XI) and are petrologically similar to the Plateau-derived Frenchman Springs member (Snavely and others, 1973). Both are distinguished by the presence of scattered plagioclase phenocrysts in a finely crystalline groundmass.

Chemistry

The Cape Foulweather Basalt in the study area can be distinguished from the Depoe Bay chemical type on silica variation diagrams of the major oxides (Fig. 16). Cape Foulweather Basalts have higher total FeO and TiO_2 values and lower SiO_2 content that the Depoe Bay Basalt. Comparison of trace element data (Tables 2 and 4) also indicates the differences between Depoe Bay and Cape Foulweather chemical types. In particular Sm, Eu, Sc, and Th differ significantly enough to be useful for distinguishing between the two basalt types in the study area. Sc and Th are not as useful when comparing the high MgO Depoe Bay basalts of Hill (1974) with the Cape Foulweather in this study, but Ce appears to be another distinguishing character between these chemical types.

As Tables 3 and 4 show, there is little chemical difference between average values of major oxides and trace elements of the Cape Foulweather and Frenchman Springs basalts in this study area. This is also reflected in the average rare earth values plotted in Appendix III. In the west end of the Big Creek Gorge, a Cape Foulweather sill (OC 127 and 638, Sec. 29, T8N, R7W) has been mapped in close proximity to the plateau-derived Frenchman Springs subaerial-subaqueous flow couplet (OC 662, Sec. 33, T8N, R7W; OC 449, Sec. 34, T8N,R7W; and 442, Sec. 2, T7N, R7W). Comparison of the chemical analyses of these units (Appendix II and III) reflects the average values although the FeO values of the Cape Foulweather sill are slightly higher than those of the Frenchman Springs flow couplets. Average values of the Cape Foulweather and Frenchman Springs basalts from other areas in western Oregon and the Columbia Plateau also match closely with these from this study (Tables 3 and 4).

Magnetic Polarity

The Cape Foulweather Basalts are normally polarized as are the Frenchman Springs Basalts in this study area (Appendix IX and Plate II). Choiniere and Swanson (1979) have suggested that the Cape Foulweather may have been erupted during a geomagnetic polarity transition (gradual shift from normal to reverse polarity transition) which they have tentatively correlated to a similar geomagnetic transition that they noted in the flows of Frenchman Springs in south-central Washington. This transitional field excursion was not recognized in this study which used the fluxgate magnetometer.

This study area may be an ideal location to test their hypothesis since the Frenchman Springs and Cape Foulweather units are in such close proximity.

Physical Characteristics and Distribution

The physical characteristics and the depositional environment into which the Cape Foulweather basalts were emplaced distinguish it from the largely subaerial Frenchman Springs Member basalts and associated sandstone interbeds. The abundance of Cape Foulweather dikes, sills, and breccias as well as the associated deep-water mudstone (see sedimentary interbed section) which they intrude and with which they are interbedded is the basis for this separation. Cape Foulweather dikes are 3 - 10 m wide and show local brecciation and peperitic contacts with the adjacent baked mudstones. Baked contacts are bleached mudstones that are 0.3 - 5 m thick. Splaying and increased brecciation of the dikes into the mudstone and into the overlying Cape Foulweather breccias suggest that the dikes may be feeders for the overlying subaqueous breccias similar to those at Haystack Rock near

Cannon Beach, Oregon (Neel, 1976). One example of this phenomenon can be observed in the quarry southwest of the Tillusqua Fish Hatchery (OC 367, Sec. 29, T8N, R7W) where a 6 m thick Cape Foulweather dike intrudes and alters mudstone and breccia. Dikes are generally finely crystalline and display horizontal jointing. They commonly have a north-south to northwest-southeast orientation in the southwestern corner of the map area (Plate I). A thick vertical Cape Foulweather dike that intrudes lower to middle Miocene fossiliferous mudstone is exposed in SOC 366 and 367a (Sec. 32, T8N,R7W).

Cape Foulweather sills principally intrude the deep-water mudstone interbeds or the Silver Point mudstones (Plate I). The sills range in thickness from 10 - 50 m and are characterized by massive, blocky, vertical columnar joints. The basalt is medium gray, dense, finely to medium crystalline with 1 to 5% plagioclase phenocrysts. Sill contacts are generally concordant with the bedding although locally the mudstone is deformed by the intrusion. A large sill, exposed in an old quarry next to Big Creek close to the Tillusqua Fish Hatchery, is estimated to be 30 to 50 m thick (OC 638, Sec. 29, T8N, R7W). It trends east for 1.2 km and is found in another quarry near the edge of the topographically higher westward extent of the Frenchman Springs basalt (OC 127, Sec. 29, T8N-R7W) (Fig. 25). The sill is overlain by deep-water mudstone and partially underlain by mudstone and by the Depoe Bay breccias (OC 127Br). The sill has been tentatively correlated on the basis of similar chemistry, polarity, and stratigraphic position across the gorge to the south side (OC 714, Sec. 33, T8N,R7W) but could not be traced any further. Other thick sills occur in quarries at OC 277, (Sec. 36, T8N,R8W); OC 228, (Sec. 31, T8N,R7W); and OC 373, (Sec. 32, T8N,R7W). There appear to be more Cape Foulweather sills than Cape Foulweather dikes in the study area. This is the general

rule throughout the coastal basalts in the Oregon Coast Range. It is not too surprising that this is the case, since a dense magma would tend not to reach the surface through the soft, less dense, semi-consolidated water-saturated muds but rather would intrude horizontally along equal isobars.

The Cape Foulweather breccia is poorly exposed in the southwestern part of the study area. It is better exposed on, around, and west of Wickiup Mountain (Coryell, 1978). The breccias in the study area consists of non -vesicular to vesicular, coarse- to fine-grained hyaloclastites. Breccas clasts are poorly sorted, angular pieces of tachylyte with sparse phenocrysts of plagioclase. The thick structureless breccia is usually associated with dikes and sills which help distinguish it from he Frenchman Springs breccias and subaerial flow on the the north side of the gorge. Most outcrops are weathered to a dark reddish brown (10R 3/4) lateritic clay. Thick dark gray foram and mollusk-bearing mudstone interbeds (Tcm) are associated with the breccia and sills in the southwestern corner of the study area (Plate I).

Thin section petrography of the Cape Foulweather basalts was discussed in the section on the petrography of the Frenchman Springs Member.

In summary, the Cape Foulweather basalts are normally polarized and show very little geochemical distinction from the Frenchman Springs basalts in the study area. The presence of dikes and sills associated with the breccias is the main distinguishing feature between these units. The bathyal mudstone interbeds present in the Cape Foulweather breccia and intrusions in contrast to the arkosic, cross-bedded sandstone interbeds in the Cape Foulweather basalt also help to distinguish these units in the field. The stratigraphic

position of the Cape Foulweather relative to the Frenchman Springs is still uncertain at this time, but topographically the Cape Foulweather occurs below the Frenchman Springs breccia and subaerial flows.

Pomona Member of the Saddle Mountains Basalt

The Pomona Member, first named by Schmincke in 1967 and formalized by Swanson and others (1979) on the Columbia Plateau, is exposed in the extreme northern part of the study area (Plate I). This basalt, dated at 12 $\pm$ 0.5 m.y. on the Columbia Plateau by McKee and others (1977), is the youngest Miocene unit present. The Pomona basalt in the study area consists of deeply weathered spheroidal blocks with orange weathering rinds exposed in road cuts and cliff faces along the Columbia River near Aldrich Point (Sec. 31, T9N, R6W) (Fig. 26). The unit is composed of a single, sparsely porphyritic vesicular subaerial basalt flow, 10 - 15 m thick. It has both a distinctive chemistry and reversed polarity like the Pomona basalt on the Columbia Plateau described by Swanson and others, (1979). Fresh samples of the basalt occur in a rock quarry (OC 582, Sec. 31, T9N, R6W) and in a landslide scarp where the contact with the underlying Clifton formation is also exposed (OC 585; Sec. 31, T9N, R6W). This unconformable contact is a sharp, straight boundary with little brecciation or splaying of the basalt flow into the underlying sandstone, suggesting a subaerial emplacement (see Contact Relations- Clifton formation). A 6 m thick arkosic sandstone overlies the basalt in the quarry near Aldrich Point(OC 582).

Chemistry

Table 5 presents data on the major oxides for the Pomona flow from the Columbia Plateau by Swanson and others (1979),

Table 5

Comparison of Average Major Oxide Concentrations for the Pomona and Pack Sack Chemical Types

	30 Pomona Flow ** (Swanson et al, 1979)	2 Pomona Flow* (Hill 1974)	1 Pomona Flow ** (This Study)	Basalt at Pack Sack Lookout 10 (Snavely et al 1973) **	4 (Hill 1974) *
SiO_2 (%)	52.1	N.D.	52.7	52.3	N.D.
Al_2O_3	14.9	13.4	14.9	15.1	13.8
FeO	10.6	11.5	10.6	10.3	10.6
MgO	7.0	6.7	7.2	6.9	6.5
CaO	10.7	10.0	10.3	10.6	10.5
Na_2O	2.4	2.5	2.3	2.3	2.4
K_2O	0.64	0.8	0.65	0.56	0.5
TiO_2	1.62	1.7	1.65	1.6	1.65
Total	99.96	---	100.30	99.66	---

Number of analyses Analytical error listed in Table 1

* INAA Methods

** AA Methods

Table 6

Comparison of Average Trace Elements for the Pomona and Pack Sack Chemical Types

	*] Pomona Member (Swanson and others, 1979)	2] Pomona Member (Hill, 1974)	1] Pomona Member (This study)	4] Pack Sack basalt (Hill, 1974)
Ba(ppm)	235.0	220.0	250.0	220.0
Co	43.4	42.0	68.0	41.0
Cr	112.0	84.0	91.0	95.0
Hf	3.2	3.9	3.5	3.4
Ta	0.77	0.80	1.02	0.68
Th	2.5	2.7	2.5	2.5
Zr	--	140.0	170.0	100.0
Sc	34.4	37.0	35.0	35.0
La	17.0	19.0	18.0	17.0
Ce	34.5	40.0	36.0	35.0
Sm	4.7	5.6	4.8	5.1
Eu	1.43	1.68	1.51	1.49
Yb	2.7	3.0	2.4	2.6
Lu	0.39	0.43	0.41	0.40

] Number of analyses Analytical error listed in Appendix III

* Not reported

using AA methods, and Hill (1974) using INAA methods, and from the Pomona flow in this study. Inspection of the tabulated data shows that the chemistry of the Pomona in this area is nearly identical to the chemistry of the Pomona flow on the plateau. Also reported on Table 5 are the average values for the basalt at Pack Sack Lookout, the coastal equivalent of the plateau-derived Pomona Member (Snavely and others, 1979; Hill, 1974). The major oxide values of the Pack Sack basalt are very simlar to the values of the Pomona flow in this study and from the plateau.

In comparison to the other basalts in the study area, the Pomona is higher in MgO and CaO and lower in total FeO, TiO_2, and K_2O than the Frenchman Springs, Cape Foulweather, Depoe Bay or Grande Ronde (compare Table 5 with Tables 1 and 3; Fig. 16 and 17).

Trace element data are presented in Table 6. Comparison of these values confirms the observations from the major oxide values. The Pomona is distinctly different from the other basalts in the study area in that it contains more Cr, and less Ba, Hf, and rare earths (Appendix III).

Magnetometer readings of several samples indicate that the Pomona has reversed polarity.

Petrography

The Pomona Member in the study area, unlike the other darker basalt units, is characterized by a medium (N4) to light gray (N7), finely crystalline, inequigranular basalt. It is crudely jointed, contains a highly vesicular, reddish oxidized top, and is the most weathered of the basalt flows in the study area. The basalt is sparely porphyritic with zoned phenocrysts

Figure 26 Typical exposure of the Pomona Member with spheroidal weathering. Blue hat for scale. Aldrich Point (OC 609, SW, Sec. 31, T9N, R6W).

Figure 27: Photomicrograph of Pomona Basalt. Note glomeroporphyritic texture consisting of labradorite (An_{60-65}) and clinopyroxene in an intersertal groundmass. Crossed nicols, field of view is 4mm across.

of plagioclase (<5mm) (10-20%) and microphenocrysts of augite and olivine. Schmincke (1967) noted the same petrography of the type Pomona flow on the Columbia Plateau. The intersetal groundmass is composed of laths of plagioclase, pyroxene, magnetite, glass, and olivine. Also distincitve of the Pomona flow are glomerophenocrysts of plagioclase with augite (Fig. 27) and plagioclase phenocrysts riddled with clinopryoxene and tachylyte glass filling resorption voids. The composition of the Pomona Member in this study consists of 45% labradorite (An 60-65), 40% augite, 10% tachylyte glass, 3% olivine, and 1-2% magnetite. Glomerophenocrysts and poikilitic plagioclase phenocrysts are also present. Alteration minerals include celadonite, calcite, and zeolites.

Previous workers (Kienle, 1971; Schmincke, 1967) had traced the Pomona flow, which was erupted in western Idaho, down the Columbia River as far west as Kelso-Cathlamet, Washington. Kienle postulated that the Pomona flow occurred above the "Post Astoria" sandstone of Dodds (1969) (renamed the Clifton formation in this thesis) in the Bradley State Park area but did not specify whether he had located any outcrop in his thesis. The chemical, magnetic, and petrographic data presented in this study substantiate that indeed a Pomona type basalt did reach the Nicolai Mountain-Gnat Creek area.

Snavely and others (1973) showed that the Pomona flow was chemically, petrologically, and magnetically the same as the coastal basalt at Pack Sack Lookout, in southwest Washington, Bowman and others (1973) and Hill (1974) verified Snavely and others (1973) chemical correlation using trace elements. Therefore, the flow in this study could have originated as a locally produced coastal basalt. Snavely and others (1973) correlated the Pack Sack with the subaerial flow (Pomona of Kienle) at Kelso, Washington. Due to the lack of feeder dikes,

they left open the possibility that the Pack Sack basalt had originated in eastern Idaho as the Pomona flow. Wells (1979) noted a 500+ foot sill of Pack Sack basalt in southwest Washington to the northwest of this study area.

This study does not answer the question as to whether the Pack Sack is coastal in origin or derived from the Columbia Plateau. However, it is this writer's opinion that the basalt in the Aldrich Point area correlates best with the plateau-derived Pomona Member of the Saddle Mountains Basalt based on:

1) The geochemcial and magnetostratigraphic correlations of the Pomona flow down the lower Columbia River by Schmincke (1967) and Kienle (1971),

2) The lack of Pomona intrusives in the study area,

3) The age and stratigraphic position of this basalt relative to the underlying Clifton formation and to the subaerial Frenchman Springs Member (Fig. 15), and

4) The subaerial nature of this flow and its similarity in geochemistry and paleomagnetism to the Pomona flow of the plateau (Tables 5 and 6).

Sedimentary Interbeds

Associated with the middle Miocene basalts are sedimentary interbeds which represent time between volcanic events. These beds are useful for determining the paleoenvironment into which the basalts were emplaced. These sedimentary units are also used for stratigraphic correlation within the study area and for regional correlation. Fossils in some beds help to bracket

the relative age of the basalts. Interbeds consist of fluvial sandstone and deep-marine mudstone. The most useful interbed for stratigraphic correlation is the sandstone equivalent to the Vantage Member of the Ellensburg Formation of eastern Washington which lies between the Grande Ronde and Wanapum Basalts. This member along with five minor sandstones (.5 - 5m thick) are interbedded in the Columbia River Basalts of this study area (Plate I and II). Another major interbed is the 200m thick fluvial to marine Clifton formation which is an informal formation proposed in this study. This unit is stratigraphically bounded by the Wanapum and Saddle Mountains Basalts in the study area (Fig. 15). Due to its widespread exposure and variety of depositional facies the Clifton formation is discussed separately (see Clifton formation).

In contrast to the sandstone interbeds in the Columbia River Basalts, the interbeds in the coastal basalts in the study area are typically deep-marine mudstone. There are at least 2 or 3 mudstone interbeds in the area, ranging from 1 to 5m thick. These fossiliferous units are most useful for age and paloenvironmental interpretation. At this time individual mudstone interbeds are not sufficiently distinctive to be used in regional correlation. To the south, in the central Oregon Coast Range, near the type localities of the Depoe Bay and Cape Foulweather petrologic units, the interbeds consist more of shallow-marine sandstones and siltstones, in particular the sandstone at Whale Cove which separates these two petrologic types (Snavely and other, 1973). In contrast, in the northwestern part of the Coast Range the lithology of the interbeds is predominantly deep-water mudstone (Neel, 1976; Penoyer, 1977; Coryell, 1978). The sandstone interbeds associated with the Columbia River Basalts will be discussed first, and then the mudstones that are interbedded with the coastal basalts will be discussed.

Vantage Member ($Tcrs_1$)

The Vantage sandstone at the type locality in south-central Washington is composed of sandstone and siltstone which overlies the Grande Ronde Basalt and is overlain by the Wanapum Basalt (Bingham and Grolier, 1966). To the west on the plateau, the Vantage sandstone merges laterally with and cannot be distinguished from the Ellensburg Formation. Therefore, Schmincke (1964) placed it within the Ellensburg Formation and Swanson and others (1979) changed the name to the Vantage Member of the Ellensburg Formation ($Tcrs_1$). This unit forms a distincitve horizon recognized throughout the western Columbia Plateau. Anderson (1978) has recognized this member in the Western Cascades in the Clackamas River area.

In this study, the sandstone unit present between the Grande Ronde and Frenchman Springs is equated to the Vantage Member on the plateau based on it similar stratigraphic position and lithology. The unit is best exposed along U.S. Highway 30 in the Bradley State Park section where it is approximately 70 meters thick (Fig. 28; Plate II). This sandstone can be traced almost continuously to the top of Nicolai Mountain along Nicolai Ridge and the Shingle Mill logging road (OC 38, Sec. 8, T7N,R6W), where it has a thickness of 20 - 25 meters. A complete exposure of the unit in the inaccessible cliffs of Frenchman Springs and Grande Ronde basalts on the northeast side of Nicolai Ridge can be observed from the powerline road to the east. The unit continues to thin to the west being only 3 - 10 meters thick in Big Creek Gorge (OC 169).

In Bradley State Park section, the basal contact of the sandstone with the Grande Ronde Formation is sharp and

irregular with some scouring of the underlying basalt. The upper contact is irregular with the overlying Frenchman Springs Basalt splaying into the arkosic sandstone. The most westward contact of this sandstone with the basalts is exposed in Big Creek (OC 169, Sec. 34, T8N, R7W), where the Frenchman Springs breccias are in sharp planar contact with the underlying sandstone.

The Vantage Member in the study area is composed of carbonaceous and micaceous arkosic arenite with a thin basal volcanic arenite. It is texturally submature, structureless to faintly trough cross-bedded and is characterized by fine-grained, angular to subrounded, moderately to poorly sorted clasts. The sandstone is very positively skewed and leptokurtic (see Grain Size Analysis Section). The induration of this sandstone varies from very friable to locally well cemented with calcite. Carbonaceous layers (1-3cm thick) are present throughout the unit. A single trace fossil, possibly a Rosselia burrow was found in Big Creek Gorge (OC 169) (Chamberlain, 1979, written communication). This is the only fossil encountered in the Vantage Member of this study. The major mineralogic constituents of the arkosic arenite are listed in Table 10 (See Sedimentary Units petrography section). They are quartz (25%), plagioclase (15%), potassium feldspar (13%), volcanic rock fragments (14%), and mica (3%). Calcite cement forms as much as 20% of the indurated sandstone in Big Creek Gorge (OC 169). The volcanic arenite is composed of subrounded to rounded andesite and basalt rock fragments, volcanic chert, and altered glass shards, totaling 75% of this sandstone. Associated with the volcanic detritus is muscovite, polycrystalline quartz, plagioclase, micrographic intergrowths of quartz and plagioclase, and quartz grains totalling 8% of

the rock. Carbonaceous plant debris (1-2%) and calcite cement (15-20%) comprise the rest of the basal volcanic sandstone in the Bradley State Park section

A trough cross-bedded basaltic sandstone breccia in Big Creek Gorge (670, Sec. 34, T8N,R7W) consists of alternating thin layers (5cm thick) of coarse- and fine-grained material. Trough cross sets are 8 - 15cm in amplitude and suggest a flow direction to the southwest. The grains are subangular to subrounded fragments of basalt. Brown basaltic glass and its alteration products of smectite clays make up the bulk of the sandstone giving the rock the appearance of a hyaloclastite breccia. This unit overlies a Grande Ronde subaerial flow and is below the arkosic sandstone (OC 169), suggesting that it is a local correlative to the basal volcanic sandstone in the Bradley State Park section. Analyses of the major and heavy minerals in the arkosic arenite (Sample 169, Appendix IV) indicate source terrains of granites and metamorphics, probably from eastern Oregon and Washington. Similar source terrains are indicated for the fluvial part of the Clifton formation (See Provenance Section-Clifton formation). The basal basaltic unit is thought to be derived from erosion of the underlying Grande Ronde Basalt.

Depositional Environment

The depositional environment of the Vantage member is principally fluvial, reflecting deposition between the subaerial Grande Ronde and Frenchman Springs flows. The trough cross-bedded laminations, moderate sorting, roundness of the basaltic grains, carbonaceous lenses and lens-like geometry of the deposit are suggestive of a fluvial environment. Plots of the grain size statistics of this sandstone on Passega (1957) "C-M" diagrams and Friedman (1962) diagrams indicate tractive

Figure 28 Vantage Member equivalent above high MgO Grande Ronde (figure 18) exposed along Highway 30. Sandstone is approximately 60 meters thick. Overlying it is the Frenchman Springs basalt shown as the horizon in this photograph (figure 19).

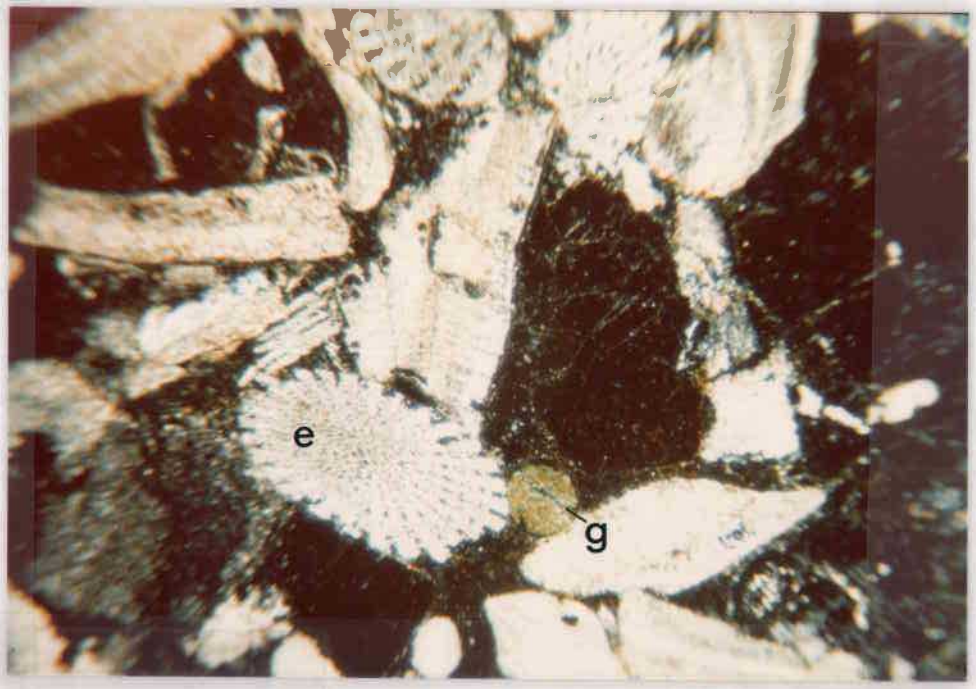


Figure 29 Photomicrograph of an echinoid, gastropod coquina between pillows in a high MgO Grande Ronde pillow palagonite complex flow in Gnat Creek Gorge. Note glauconite grain (pale green in center) and isotropic basaltic glass fragment. Crossed nicols, field of view is 4mm across (OC 2gc, NW, Sec. 6, T7N, R6W).

current processes as a mechanism of transport (See Grain Size Analysis Section). These factors and the presence of subaerial basalt above and below the sandstone suggest that a major part of this unit was deposited in a fluvial environment. These channels and stream bed deposits represent the avenues along which the lava could flow because they define the local topographic low points in the river system. A possible exception to the fluvial interpretation for the Vantage Member is the sandstone in Big Creek Gorge. This unit is overlain by the Frenchman Springs subaerial-ppc unit and underlain by a Grande Ronde subaerial-ppc unit which are thought to have flowed into a marine embayment. The stratigraphic position of the Vantage Member between these units along with the possible Rosselia (?) burrow which typically are found on very shallow sand bars (Chamberlain, 1979, personnal communication) suggest that a part of the sandstone may be shallow marine.

Minor Sandstone Interbeds

There are at least five minor interbeds (Tcrs) associated with the Columbia River Basalt Group in the study area that are primarily composed of fine-grained arkosic and basaltic sandstone. Also there is a local bed of marine coquina. These units do not have the lateral extent of the Vantage Member but are useful for local correlation and environmental interpretation of the basalts.

The lowest interbed is a 5 - 10-meter thick, fine-grained, arkosic sandstone located at the R_2/N_2 horizon between pillow lavas in the Nicolai Mountain section (Fig. 15). This unit is not exposed anywhere else in the study area. It does aid in dividing what initially appears to be a thick continuous exposure of pillow basalts.

The next sedimentary interbed upsection is a thin (1 m) coquina found in Gnat Creek Gorge (OC 2gc, Sec. 5, T7N,R6W) (Fig. 29). The outcrop is along the bed of Gnat Creek in an isolated pillow palagonite basalt sequence. The lateral extent of the unit is impossible to determine due to the poor exposure of the outcrop. According to Ellen Moore (1980, written communication), the fossils in the coquina are small mollusk fragments, primarily pelecypods and gastropods, with an abundance of echinoid spines (sea urchins). She suggests that both the molluska and echinoids are marine but could not confirm a water depth. Also present in the coquina are smectite clays, chlorophaeite, and rare glauconite grains. The presence of this unit within an isolated pillow basalt of high MgO Grande Ronde support the hypothesis of Snavely and others, (1973) that these basalts flowed into the sea.

The third pre-Vantage sandstone interbed is exposed in the Bradley State Park section along U.S. Highway 30 and beneath Gnat Creek Falls in the Gnat Creek Gorge section (Plate II). The interbed is approximately 2-5 meters thick. It separates two high MgO Grande Ronde flows in the Bradley State Park section. The arkosic sandstone is dark orange-brown (10YR 6/6) medium- to fine-grained, and composed of quartz, feldspar, muscovite, weathered basaltic grains, and carbonaceous material. This interbed along with the similar physical, geochemical, and paleomagnetic characteristics of the overlying basalt flow is the basis for correlation of the high MgO Grande Ronde flows from Bradley State Park to the Gnat Creek Gorge section (see Internal Stratigraphy section).

There are two post-Vantage sandstone interbeds, other than the Clifton formation, identified in the study area. One is only 0.5 - 1.5 meters thick and is exposed in the stream bed of Gnat Creek south of the fish hatchery. It occurs between two

Frenchman Springs subaerial flows (OC-614). The thin-bedded unit is a grayish olive (10YR 4/2), coarse- to fine-grained, poorly sorted basaltic sandstone composed of altered basalt fragments in a clay matrix. Many grain boundaries appear obliterated by weathering and forming the clay matrix. The sandstone was presumably derived from erosion of the underlying Frenchman Springs basalt.

The second sandstone interbed occurs in Big Creek Gorge between the lower Frenchman Springs subaerial flow and an overlying Frenchman Springs closely-packed pillow flow. The sandstone is exposed along the cliff on the north side of the gorge near the top of the 1869 foot mountain (OC 442, Sec. 2, T7N, R7W). This sandstone is medium- to fine-grained and cross-bedded with an arkosic composition. The trough cross-bedding suggests transport directions of southwest ($S60^{o}W$) for this sandstone. A fluvial origin is suggested for the basaltic sandstone in Gnat Creek and for the cross-bedded arkosic unit based on the close association of the subaerial flows of Frenchman Springs and lithologic similarities to other fluvial sandstone interbeds. This, however, is speculative because insufficient data were available.

Scattered throughout the central part of the thesis area are outcrops of arkosic sandstone (Plate I). These units are primarily underlain by Frenchman Springs with the exception of OC 426 (Sec. 7, T7N,R6W) which is underlain by Grande Ronde Basalt. These were not assigned to any particular interbed.

Interbeds in the Coastal Basalts

The sedimentary interbeds (Tcm) associated with the Depoe Bay and Cape Foulweather petrologic types in this study are

principally deep-marine mudstones. This differs from the fluvial and shallow marine(?) sandstone which is the typical lithology of interbeds in the plateau-derived Columbia River Basalts. The mudstones are not laterally continuous due to the extensive disruption caused by emplacement of the basaltic material as intrusions and submarine breccias. The mudstone cannot be easily divided into individual units due to lithologic homogeneity and disruption of the basalts. This limits the usefulness of the mudstones to interpreting the environment in which the coastal basalts were emplaced and for differentiating the coastal basalts from the Columbia River Basalts. These mudstones are principally found in the southwest part of the study area (Plate I). Most are associated with Cape Foulweather sills and dikes.

These clayey siltstones are orange (10YR 7/4) to grayish black (N2), rarely mottled, laminated to structureless, and have minute mica flakes dispersed throughout. Exposures rapidly weather into equant pieces producing chippy talus. A distinctive yellow to white weathering covering the grayish black mudstone is characteristic of a baked zone adjacent to basalt intrusions. Peperite zones (1 - 5m thick) composed of altered basalt and mudstone fragments are common near contacts. The peperites indicate a degree of mixing between the mudstones and basalt due to effects of steam blasting as hot magma intruded the water-saturated sediment.

At least one mudstone interbed separates two Depoe Bay subaqueous flows in a rock quarry in Big Creek Gorge (OC 125, Sec. 29, T8N, R7W). The gray black mudstone is 3 meters thick. It overlies a closely packed pillow flow of low MgO Depoe Bay Basalt and is overlain by a pillow breccia which contains a second 1 meter thick penecontemporaneously deformed mudstone near the top of the quarry. Thus, there are two and

possibly three Depoe Bay subaqueous flows exposed in the quarry. Fossils identified in the lower interbed include pelecypods, pyritized forams, and diatoms (Appendix IV). Also present are glauconite and quartz grains. The forams equate to the Saucesian Siphogenerian kleinpelli stage (early to middle Miocene) (Rau, 1980, written communication) while the diatoms indicate an age no older than late early Miocene (Barron, 1980, written communication). Rau (1980, written communication) suggested that the forams indicate upper to middle bathyal depths of 250 to 1000 meters. A mudstone unit (OC 646) on the north side of Big Creek near the fish hatchery has similar pelecypods to those in OC 125 (Appendix IV). It may be the lateral continuation of the mudstone at OC 125 or is another interbed stratigraphically above the Depoe Bay basalt and intruded by Cape Foulweather Sill. There is little problem in separating these interbeds from the older, lithologically similar Silver Point mudstone which underlies the Depoe Bay basalt because they are bounded above and below by Depoe Bay pillow basalts.

The mudstones associated with the Cape Foulweather Basalts cannot be differentiated as easily into separate stratigraphic interbeds because of the predominant intrusive nature of the basalts in the study area. For example, the interbedded mudstones north of the Tillusqua Fish Hatchery (OC 127), which are at least 20m thick are intruded by a thick Cape Foulweather sill (Fig. 25). This leads to problems as to whether these interbedded mudstones are actually equivalent to the upper Silver Point mudstone which lies stratigraphically below the Depoe Bay Basalts, or were deposited after the Depoe Bay basalts. To complicate the situation neither the age nor the depositional environment, based on forams and mollusks, is sufficiently different to distinguish Silver Point from "interbedded" mudstones.

Rau (1980, written communication) reported that forams from both mudstone units indicate a Saucesian age and depositional environment of middle to upper bathyal depths (400-800m) (Appendix IV). However, the general lack of laminations, abundant carbonaceous material, or large mica flakes in the"interbedded" mudstones was used to distinguish interbeds from Silver Point mudstone in the field. This distinciton is somewhat arbitrary; therefore, the contact between Silver Point mudstones and younger interbeds is denoted with question marks where there is no intervening basalt (Plate I). Another aid to differentiation is the associated basalt outcrop pattern and the topography. Outcrops of the interbedded mudstone are topographically higher than the predominantly low-lying Silver Point mudstone and are associated with breccias or intrusions (Plate I).

Reconnaissance in the Wickiup Mountain area suggests that similar mudstone interbeds between breccias may exist there also. Alternatively, these mudstones may be upthrown blocks of Silver Point intruded by Depoe Bay and Cape Foulweather basalts or erosional windows below piles of basalt breccias, as suggested by Coryell (1978). Neel (1976) noted a 5-meter thick foram-bearing mudstone interbed between Depoe Bay breccias near the town of Arch Cape, and Penoyer (1977) mapped a 100-meter thick massive mudstone interbed in the Depoe Bay basalts in the Humbug Mountain area. Coryell (1978) collected deep-water pelecypods from a 1 meter thick mudstone in subaqueous Cape Foulweather breccias west of Wickiup.

In summary, the sedimentary interbeds are useful for correlation of the basaltic units on a local scale as well as on a regional scale. They are also useful as paleoenvironmental indicators.

The Vantage Member is a key unit in defining the break between the Grande Ronde and Frenchman Springs Basalts. This unit and the associated minor micaceous sandstone interbeds were derived from east of the Cascades. The presence of the arkosic sandstone interbeds within the sequence of plateau-derived basalts is testimony for the existence of a fluvial system (e.g., ancestral Columbia River) which drained the plutonic and low-grade metamorphic terrains east of the Cascades. The sandstone also indicates that the basalt flows on the plateau did not block the downwarp through the Cascades which would have cut off the supply of plutonic detritus to the area. Alternatively, these sandstone interbeds are recycled Astoria or Scappoose Formation material from west of the Cascades. The echinoid coquina lens in the Grande Ronde pillow flow substantiates the hypothesis that the Columbia River Basalt did enter a shallow marine embayment in the area.

The mudstone interbeds aid in separating the Depoe Bay and Cape Foulweather petrologic types from their Columbia River Basalt coeval equivalents which have sandstone beds. The mudstone interbeds are distinguished from the lithologically similar upper Silver Point mudstone on the basis of the lack of mica flakes, abundant carbonaceous material and laminations. The peperities and contorted baked mudstone contacts indicate that the mudstone was soft when the basalts were emplaced. Fossils in the mudstone indicate that the coastal basalts were emplaced in bathyal water depths (250 - 1000m). The geographic (spatial) relationship of the bathyal mudstone interbeds and the nearly coeval shallow-marine to fluvial sandstone interbeds require a special set of circumstances which will be discussed in the section on the origin of the basalts.

Correlation of the Basalt Units within the Nicolai Mountain-Gnat Creek Area

A fence diagram presented as Plate II depicts the stratigraphic and geographic distribution of the middle Miocene basalt units in this study area. The stratigraphic position and chemical, paleomagnetic, and physical characteristics of each of the units shown in the diagram have been described (see Stratigraphy Section). Of the factors used for identifying the different basalt units, the major and trace element geochemistry and petrography are most useful for correlation purposes because of the rapid variations in physical characteristics noted over relatively short distances (e.g., subaerial flows to subaqueous pillow and breccia flows.) The paleomagnetic data also are useful in determining the relative stratigraphic positions of basalts of similar geochemical types. Within the study area both local subaqueous coastal basalts and plateau-derived subaerial-subaqueous couplets of Columbia River Basalts are exposed. Because these coeval basalts are identical in chemistry and petrology (Snavely and others, 1973) the depositional environment into which each unit was emplaced becomes critical to differentiating the two groups.

This section is an attempt to tie the different middle Miocene basalt units into a coherent sequence of events. The interpretation of the stratigraphic sequence of basaltic units and their origins presented here is not unique and other alternatives will also be suggested. There are also a number of unanswered questions which possibly could be resolved with further field and laboratory work but which could not be accomplished in the time available for this study.

The fence diagram (Plate II) was constructed from seven composite stratigraphic sections. Only the Bradley State Park

Section was measured with a Jacob's staff and Abney level; it is described in detail in Appendix I (Section e-e'). The other sections were drawn from estimation of thicknesses based on topography and observations made during traverses over the field area and from outcrop distributions and attitudes (Plate I). The positions of the numbers along the side of each section relate to the relative position of the samples used for chemical analysis. The samples along with the remanent polarities are listed in Appendices II & IX. The location of the sections are shown both on the index map on Plate II and on the geologic map (Plate I).

The Nicolai Mountain Section (No. 5) is a composite of reconnaissance sampling by Snavely and MacLeod in 1969 and field mapping in this study. The section starts in a large quarry 1/2 mile south of Nicolai Mountain and continues north to the Nicolai Mountain lookout. The Big Creek sections (Nos. 6 and 7) were developed from 15 traverses up the north side of the gorge. The Gnat Creek Gorge section (No. 4) resulted from a traverse up the six waterfalls of Gnat Creek. Only one sample from this section was submitted for chemical analysis but it established a correlation line, together with the physical and paleomagnetic characteristics and stratigraphic position of these flows to units in other sections. The section along lower Gnat Creek (No. 3) is easily accessible, located just south of where U.S. Highway 30 crosses Gnat Creek. The Aldrich Point section (No. 2) is a composite drawn from a traverse made by Dr. Niem and myself in February, 1980 and a projection of the Frenchman Springs basalt along Clifton Road (Sec 8, T8N, R6W) where the units dip to the northwest under the Columbia River.

A discussion of the different rock units will follow the generalized stratigraphic column illustrated in Figure 15. The

lowest basalt unit in the study area is the N_1 low MgO Depoe Bay petrologic type of the Big Creek area (Tdb). The unit is underlain by Silver Point mudstone of the Astoria Formation and overlain by a bathyal mudstone intruded by a Cape Foulweather sill in the westernmost part of Big Creek Gorge (Section 7, Plate II). It is overlain by a thick sequence of subaerial and subaqueous basalts along the north side of Big Creek Gorge (Section 6, Plate IV). The Depoe Bay pillow palagonite complex (ppc) consists of at least two, and possibly four, subaqueous isolated pillow lavas and breccias, and associated "feeder" dikes with interbedded thin mudstone layers. The presence of these dikes and of bathyal mudstone within and overlying the Depoe Bay Basalts when contrasted with the fluvial arkosic sandstone associated with the geochemically similar, but stratigraphically and topographically higher, Grande Ronde flows are the main reasons for placing these flows at the base of the section.

Alternatively, the Depoe Bay Basalt unit may be equivalent to the N_2 low MgO Grande Ronde closely-packed sequence ($Tygr_{2a}$ Section No. 6, Plate II) which overlies the Depoe Bay in Big Creek. Stratigraphic equivalence to the N_2 low MgO unit in the Nicolai Mountain section is unlikely, however, on the basis of differnt associated lithologies and physical characteristics and on the basis of the MgO versus TiO_2 plot. The N_2 low MgO flow is postulated to be a thick, individual, closely-packed pillow flow with a unique amount of TiO_2. The percentage of TiO_2 is significantly higher (2.5% TiO_2) than in the Depoe Bay flows at Big Creek (2.o To 2-2% TiO_2) (Fig. 17). Also interbeds in the Nicolai Mountain section are shallow-marine or fluvial sandstone rather than bathyal mudstone which are interbedded with the Depoe Bay breccias and dikes. Unfortunately, no trace element data are available for any of the units in the Nicolai Mountain Section

to substantiate the TiO_2 variation noted on Fig. 17. Further study should include determination of trace element abundances of the low MgO units in the Nicolai Mountain section.

The second alternative stratigraphic interpretation is to call all the basalts in Big Creek N_2 Grande Ronde Basalts, equivalent to the N_2 low MgO Grande Ronde in Bradley State Park ($Tygr_{2b}$). This alternative explanation has more merit than the first alternative interpretation for three reasons. One, is that the major and trace elements geochemistries of the low MgO subaerial flow in the Bradley State Park section (No. 1) and in Big Creek (17,20 and 21) completely overlap the geochemistries of the Depoe Bay breccia in Gib Creek (Nos. 24, 25, 23, 22 and 35) (Appendix II; Figs. 16 and 17). Secondly, the mapped contact between the Grande Ronde and Depoe Bay on the north face of Big Creek Gorge is poorly defined. It is based primarily on the lack of mudstone and dikes and the presence of oriented "flow-foot pillow breccias" and associated subaerial flows in the upper Grande Ronde unit in Big Creek Gorge. Lastly, the low MgO Grande Ronde basalt is incompletely exposed in the Bradley State Park and Gnat Creek Gorge sections. Therefore, underlying low MgO flows may exist in the subsurface below the described sections in these two areas which would better equate to the 3 or 4 flows present in Big Creek. That is, the Depoe Bay unit may actually be unexposed, low MgO Grande Ronde flows.

Problems with this second alternative interpretation are: 1) the presence of "feeder" dikes splaying upward and cutting the pillow palagonite complex which forms the Depoe Bay unit, and 2) the rapid changes in sea level needed to first deposit bathyal mudstone (400-800 meters of water depth based on forams) over the Depoe Bay Basalt and then emplace a subaerial flow exposed in Big Creek, less than 50 meters

stratigraphically above the bathyal mudstone. This requires a very rapid minimum change in water depth of 400 meters during the N_2 magnetic period. Although this relationship does not answer the question as to the chemical equivalence of the two units, it does suggest that the Depoe Bay Basalts are slightly older than the low MgO ($Tygr_{2b}$) Grande Ronde Basalt.

On-going field and geochemical investigations by Jeff Goalen, geology graduate student at Oregon State University, in the Porter Ridge and Elk Mountain area, by R.E. Wells, U.S. Geological Survey, in southwestern Washington, and by Dr. M. Beeson, Portland State University, in the Willamette Valley will hopefully shed some light on the stratigraphic relationship of these units. Further studies may also reveal how many low MgO, normal Grande Ronde flows reached this area.

The next overlying unit in the study area is a reverse polarity (R_2), low MgO Grande Ronde closely-packed pillow basalt ($Tygr_1$). This 60 m thick unit overlies the Big Creek sandstone (?) of the Astoria Formation and is overlain by a N_2 low MgO pillow flow ($Tygr_{2a}$) in the quarry in the Nicolai Mountain Section (Plate II). It is not exposed anywhere else in the study area but may equate to a reversed pillow flow located on Porter Ridge to the east. Associated with this unit are overlying and underlying, thin, arkosic sandstone interbeds that suggest shallow marine or fluvial water depths for emplacement of these pillows. The interbeds also indicate that there may be 1 or 2 flows of pillow lavas that comprise this unit. This unit is tentatively assigned to a chemical subunit defined by higher TiO_2 values within the low MgO Grande Ronde Basalt (Fig. 17). The absence of these low MgO closely packed reversed polarity pillow flows elsewhere in the study area may be due to the lack of exposure. But from

geochemical and paleomagnetic data, and stratigraphic position it would appear that this flow never reached the Big Creek area where only normal polarity flows were encountered (Plate II). A 20 m thick basalt dike (chemical sample 26) intrudes the underlying Oswald West mudstone 3 km southwest of Nicolai Mountain. This dike has similar chemistry and reversed magnetic polarity to the pillow lava flow. The relationship of the dike to the pillow lavas is not clear. It may be that the dike is a "feeder" for the pillow lavas. Alternatively, the dike may have an invasive relationship with the overlying pillow flow; that is, these "plateau-derived" pillow lavas intruded the underlying "late Oligocene soft" marine sediments to form a dike in the manner envisioned by Beeson and others (1979) for Columbia River Basalts that reached the coast. The relationship of this dike to the reversed polarity pillow lavas in the Nicolai Mountain area and in Porter Ridge is now being studied by Jeff Goalen of Oregon State University.

A normally polarized (N_2), low MgO Grande Ronde pillow flow ($Tygr_{2a}$) overlies the reversed pillow lava in the Nicolai Mountain section. The basal contact is a thin (1-2 m thick), arkosic sandstone interbed which marks the N_2/R_2 magnetic boundary in the study area. These 60 m thick pillow lavas appear to be overlain by a corresponding normally polarized subaerial flow, suggesting that it caps the pillow flow which built above the surface of the water. This pillow lava-subaerial flow tentatively corresponds to the TiO_2-rich chemical subunit in the low MgO Grande Ronde (Sample No. 13) (Fig. 17) and forms the basal flow of the N_2 magnetic polarity unit in this study (Plate II). Due to its distinctive chemical fingerprint this couplet cannot be correlated with any other normal low MgO Grande Ronde unit in the study area. It stands alone but is similar in chemistry to the underlying reverse unit (Sample 14, Fig. 17). This suggests that these

flows ($Tygr_1 + {}_{2a}$), if from the plateau, were topographically restricted to a low in the Nicolai Mountain area or, if they were erupted locally, were restricted geographically in distribution.

Within the N_2 magnetic unit, a second low MgO subaerial Grande Ronde flow has been designated ($Tygr_{2b}$). This unit, the lowest unit exposed in the Bradley State Park Section (Unit 1 Apendix I), is correlated on the basis of similar chemistry, polarity, and stratigraphic position to the 50-75 m thick subaerial ppc unit low MgO Grande Ronde unit in Big Creek Gorge and is tentatively correlated to the basal subaerial flow in the Gnat Creek Gorge section (sections 1,4, and 6; Plate II). This unit placed above the N_2 closely packed, pillow lava in the Nicolai Mountain section because the pillow flow lies on a reversely polarized pillow lava and has slightly different chemistry (i.e., high TiO_2 values, Fig. 17 and Appendix II). The areal distribution of the N_2 low MgO Grande Ronde flow indicates that the lava may have followed a topographic low around the north side of the high created by the earlier pillow flows in the Nicolai Mountain area and entered the sea in the Big Creek Gorge area.

Two high MgO Grande Ronde flows are described in this study area as unit $Tygr_3$. The flows are within the N_2 magnetic polarity unit and overlie both of the N_2 low MgO Grande Ronde flows (Plate II). A 1 to 4 m thick arkosic sandstone separates the two flows in Bradley State Park and in Gnat Creek Gorge. Correlation of these two flows from the Bradley State Park section to the Gnat Creek and Nicolai Mountain sections is based on chemistry, polarity, and stratigraphic position (Sections 1,4, and 5; Plate II). They are not found in the Big Creek section (Section 6; Plate II) where the Frenchman Springs ppc-subaerial flow couplet overlies the low MgO N_2 Grande

Ronde flows. This leads to further speculation that topography controlled emplacement of the basalts in the study area. Alternatively, the two high MgO basalt flows may have reached Big Creek, but were removed by erosion prior to emplacement of the Frenchman Springs Basalt.

The $Tygr_3$ unit consists of an older 50-60m thick ppc-m subaerial flow couplet and a younger 40-50m thick subaerial flow. A local echinoid coquina occurs within isolated sediments within the pillow breccia part of the lower flow in Gnat Creek Gorge, indicating that the flow entered a marine environment in that area. The overlying subaerial part of this flow suggests that the water was relatively shallow because the basal ppc unit, which is less than 30m thick, developed a subaerial surface. The younger high MgO subaerial flow has no ppc base and overlies a fluvial sandstone, suggesting that the younger basalt flowed over a new land surface created by the older flow.

The Vantage Member is almost 70 meters thick in the Bradley State Park section (Unit 7, Section e-e', Appendix I). It thins to the south and southwest (sections 1,5,and 6, Plate II). It is interpreted to be fluvial in the Bradley State Park area and possibly shallow marine (based on a Rosselia burrow (?)) in the Big Creek area. The unit is an excellent marker bed as it separates the Grande Ronde and Frenchman Springs basalts.

The Frenchman Springs Member (Tyfs) overlies the Cape Foulweather (Tcf) intrusive basalts in Big Creek although this stratigraphic relationship is not fully understood. The Cape Foulweather sills, alternatively, may be related to invasion of the sediment by a ppc-lava delta flow of Frenchman Springs basalt. The reverse-polarized Pomona Member (Typ) is the

youngest subaerial flow preserved in the study area. It overlies the fluvial to middle shelf Clifton formation and the Frenchman Springs basalt member near Aldrich Point.

This study area is unique in that plateau-derived subaerial flows of the Columbia River Basalt Group can be traced into the marine environment and that the paleotopography must have played an important role in the emplacement of these units relative to each other. Also deep marine locally derived middle Miocene coastal basalt ppc units appear to have formed before much of the subaerial Columbia River basalt flows reached the area. Finally, this study shows the importance of determining the paleoenvironment at the time of emplacement of the basalt units.

Correlation of the Yakima Subgroup from the Clackamas River Area in Western Cascades to the Nicolai Mountain-Gnat Creek Area

Throughout the discussion of the basalt units in this study, reference has been made to the Columbia River Basalts in the Clackamas River area described by Anderson (1978) and Hammond and others (1980). Figure 30 is a preliminary correlation chart of the composite basalt section described by Anderson and the section described in this study. The Clackamas River area, approximately 161 km east of the study area, represents a location where plateau-derived Columbia River Basalt flowed into the Willamette Valley via an ancestral Columbia River valley cut through the Cascades (Anderson, 1978). The correlations between the two areas are based on magnetic polarity signatures, chemistry, physical and petrographic characteristics of the rocks, as well as the stratigraphic position of these basalts. Tables 1, 2, 3, and 4 show the close correspondence of the average major oxides and trace element concentrations for equivalent basalt units between the

PRELIMINARY CORRELATION CHART OF COLUMBIA RIVER BASALT GROUP
FROM
CLACKAMAS RIVER AREA TO NICOLAI MT. - BIG CREEK AREA, NW OREGON

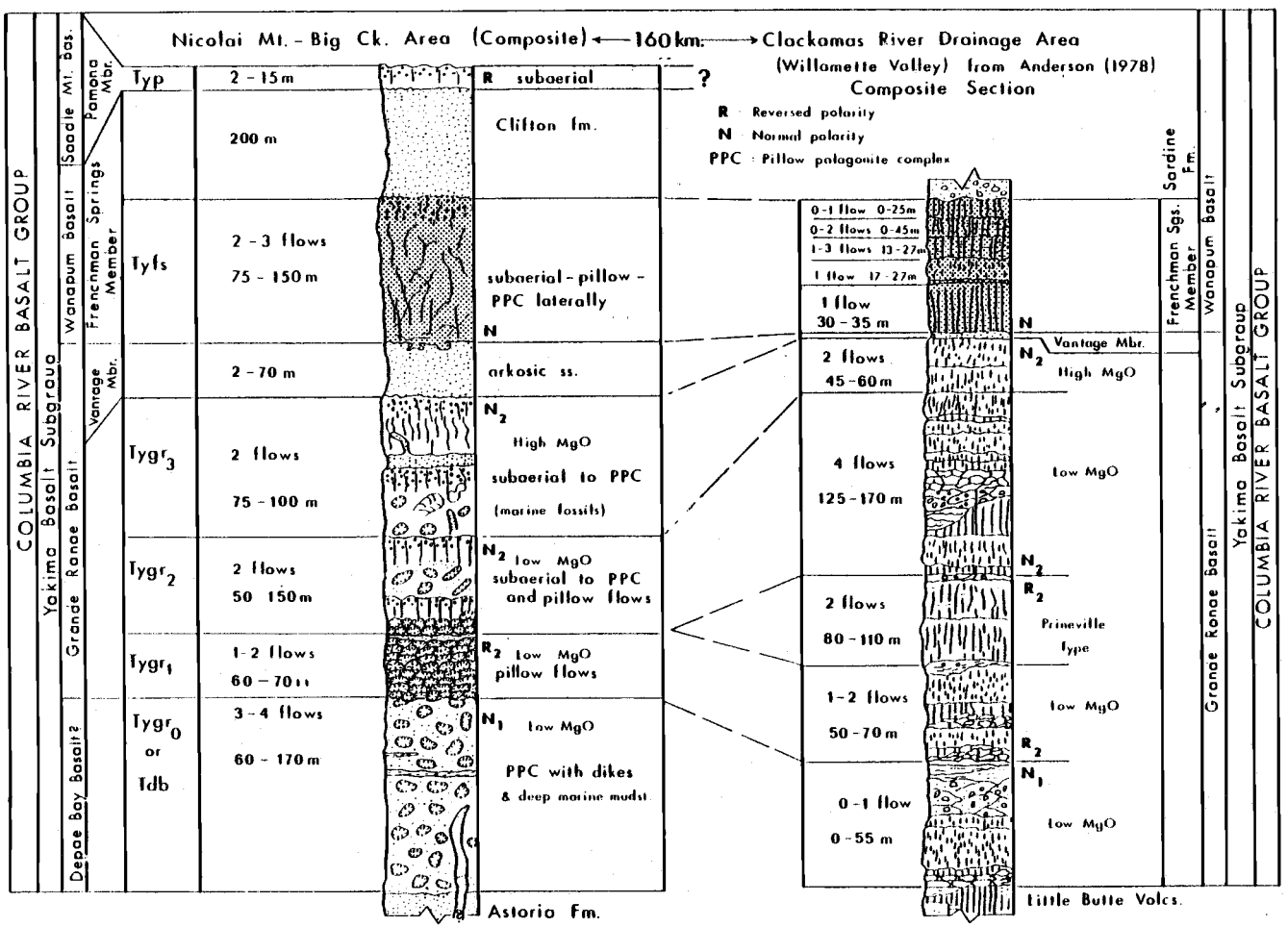


Figure 30

two areas. All the paleomagnetic divisions recognized by Anderson (1978) in the Clakamas River area are present in the Nicolai Mountain-Gnat Creek area.

The best correlation between the two areas is the sandstone unit equivalent to the Vantage Member of the Ellensburg Formation. The sandstone lies at a major stratigraphic division in the Yakima Basalt Subgroup. In both areas it separates the sparsely porphyritic Frenchman Springs Member of the Wanapum Basalt from the underlying high MgO flows of the Grande Ronde Basalt. Compositionally the Vantage Member in the Clackamas drainage area consists of a basaltic sandstone. Although the base of the member in this study area is basaltic, much of the unit is arkosic in composition (See Vantage Member equivalent in the interbed section). Total thickness and composition of the unit varies considerably from the Clackamas River area (1-2 m) to this study area (up to 60m). This may be attributed to the Nicolai Mountain-Gnat Creek area being closer to the depositional axis of the ancestral Columbia River which carried the arkosic sands that are more typical of the type Vantage Member of eastern Washington through the ancestral Columbia valley.

The Frenchman Springs Member in the Nicolai Mountain-Gnat Creek area is not divided into individual flow units based on abundance of plagioclase phenocrysts as Anderson (1978) has done in the Clackamas region where he recognized 8 separate flows. Therefore, none of the individual flows within the Clackamas composite section has been carried into this study area. Distinctive variations in plagioclase abundance in the Frenchman Springs Basalt were not systematically noted in the study area, but at least two flows occur based on interbeds in the Gnat Creek Gorge section (Plate II). The resulting correlation is thus based on the total thickness of Frenchman Springs Basalt (150 m thick) in both area (Fig. 30).

Another tie between the two areas is the high MgO/low MgO geochemical horizon in the Grande Ronde Basalts. Two high MgO flows are found above this horizon in both areas. The older high MgO in this study area is thicker, the result of entering the shallow Miocene sea and developing a prograding lava delta. There are four MgO flows (a maximum of 170m thick) below the high MgO/low MgO geochemical boundary and above the N_2/R_2 magnetic division in the Clackamas area according to Anderson (1978). Only two low MgO flows (a maximum of 150m thick) have been ascribed to this unit in the Nicolai Mountain-Gnat Creek area. This suggests that the other flows did not reach this area. An alternative to this correlation is that the two or three deep marine pillow palagonite complexes in Big Creek, referred to as Depoe Bay, are instead the missing plateau-derived flows. The reasons why this correlation was not made are discussed in the internal stratigraphy section.

The N_2/R_2 magnetic polarity boundary in the Clackamas River area separates the chemically unique Prineville basalt type, with distinctive apatite crystals from the overlying low MgO Grande Ronde units. Chemical analyses suggest that the Prineville basalt type is not present in the Nicolai Mountain-Gnat Creek area, but the N_2/R_2 magnetic polarity boundary is present in the Nicolai Mountain section between the low Mgo Grande Ronde closely-packed pillow flows (Plate II). Within the lower reverse polarized basalt one to two low MgO Grande Ronde flows (50 to 70m thick) are described by Anderson (1978). Similarly, 1 to 2 low MgO pillow flows are present in this same reverse polarized unit in the Nicolai Mountain-Gnat Creek area. These are characterized by higher than normal TiO_2 values and comprise part of a subunit in the low MgO Grande Ronde field (Fig. 17). Anderson (1978) did not show this chemical variation in his study. This may suggest a local

origin for pillow flows in the Nicolai Mountain area, but they are tentatively correlated to the R_2 low MgO Grande Ronde flows in the Clackamas area because the data are not conclusive enough. The time period involved for emplacement of the Prineville type flows in the Clackamas River area presumably is represented in part by an arkosic sandstone interbed in the Nicolai Mountain section between the R_2/N_2 flows (Fig. 30).

The underlying R_2/N_2 magnetic polarity boundary appears in both generalized stratigraphic columns. In the Clackamas River area this line separates a reverse-polarized low MgO Grande Ronde flow from an underlying normally polarized low MgO flow. The older normally polarized flow is 55m thick and unconformably overlies the Little Butte Volcanics of the Western Cascades. In this study area the R_2/N_2 boundary is inferred to occur between the low MgO normally polarized Depoe Bay flows (Tdb) and the reverse-polarized low MgO Grande Ronde flows (Tygr) in the Nicolai Mountain section (see internal stratigraphy section). The basal low MgO flow in the Clackamas River areas is poorly exposed and was not included in Anderson's (1978) measured sections but is present on his (1978) thesis map as Tgr_o. This flow is also listed on Figure 12 in Hammond and others (1980) as Tgn_1 (Anderson, written communication, 1980).

The N_1 low MgO Depoe Bay unit in this study area consists of three or four ppc flows with thin bathyal mudstone interbeds and dikes. Tentatively these units (a total of 170 m thick) are thought to represent the locally derived Depoe Bay Basalts. Alternatively, one of these flows could be related to the 55 m thick low MgO flow of the Clackamas River area or all could be submarine extensions of the subaerial N_2 MgO flows. If, however, the initial correlation described here is correct it would imply that there are more N_1 low MgO flows below the

R_2/N_1 polarity boundary in the study area than recognized in the Clackamas River area by Anderson (1978). That would support the contention that the Depoe Bay ppc flows are indeed local in origin. But the composite section in the Clackamas River area may not include all the N_1 low MgO Grande Ronde flows that came down the ancestral Columbia River Valley which has an axis north of the Clackamas River area where older flows could be buried.

The Pomona member of the Saddle Mountains Basalt caps the section in the Nicolai Mountain-Gnat Creek area. This flow has not been recognized in the Clackamas River area, but Anderson (1980) has recently recognized the Pomona member as an intracanyon flow in the Columbia River Gorge. In addition, the Waverly and Priest Rapids flows reported by Beeson and Moran (1979) in the Mount Hood, Oregon area are not present in the Nicolai Mountain-Gnat Creek area.

In conclusion, the presence of so many correlative units between this study area and the Clackamas River area suggest that many of the flows reached the middle Miocene shoreline via a wide ancestral Columbia River valley. The variations between the two composite stratigraphic columns, in particular, the lack of the Prineville type, Waverly and Priest Rapids flows and the presence of the intracanyon Pomona Member in this study area suggest that the plateau-derived flows may have been channelized even on a regional scale. The apparent over abundance of flows in the N_1 low MgO unit in the Nicolai Mountain-Gnat Creek area as compared to the same unit in the Clackamas River may be attributed to the local origin for these flows. For the present the correlations are tentative; more work needs to be done in key areas between the Clackamas

drainage and this study area (e.g., lower Columbia River to the east and along the west Coast Range) to fully establish the relationship between the local and plateau-derived basalts.

Comments on the Origin of the Coastal Basalts

The origin of the consanguineous coastal and Yakima Basalts with source vents 500 km apart has been an enigma since their chemical, age, and petrologic equivalence was revealed by Snavely and others in 1973. They speculated on three models of petrogenesis to resolve this problem but also recognized that each model had serious drawbacks. Sketches of these models from the descriptions of Snavely and others (1973) were drawn by Dr. Alan Niem and are shown in Figure 31.

The first model (Figure 31a) calls for partial melting of the Juan de Fuca plate, as it is subducted beneath western North America. They suggest that the basaltic magma originated from partial melting of eclogite in the mantle and rose following major north trending fracture zones in the plateau region and along the coast. Most of the magma was extruded in the Columbia River Plateau region. Some of the basalt magma may have migrated up the underthrust zone, intersecting and ascending along a second major fracture system parallel to the present coastline. The main problems which they recognized with this model are: 1) it fails to explain generation of calc-alkaline magma which was erupted in the Cascade Range at this time; 2) it is difficult to imagine how a magma could migrate several hundred kilometers up a subduction zone without significant contamination; 3) the magma must be generated deep within the mantle; and 4) the other tholeiitic flood basalts are not related to subduction zones.

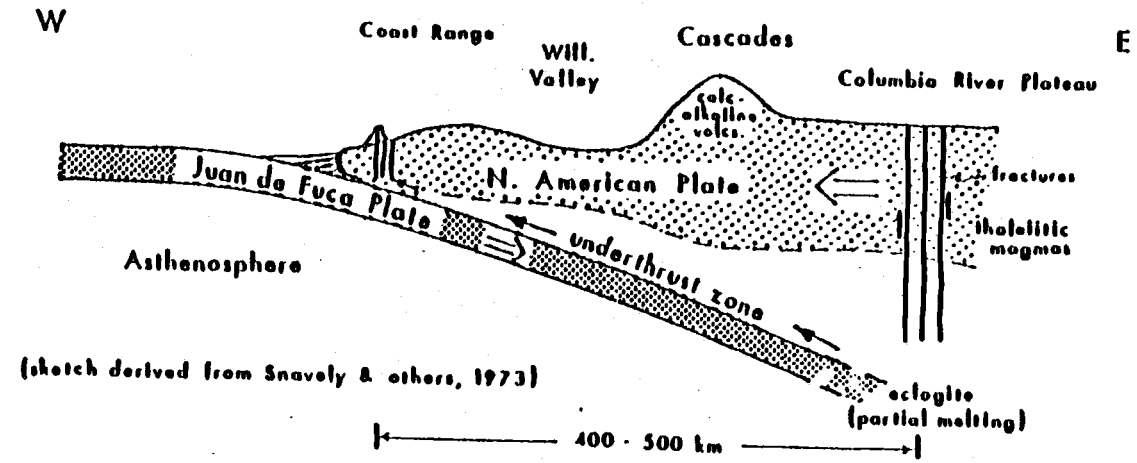
The second model (Fig. 31b) proposes that the tholeiitic magma results from partial melting in a broad weakened shear zone at the top of the asthenosphere. The shearing is the result of westward drift of the American plate over the asthenosphere. This would create an elongate, thin magma chamber which intersected the fracture systems along the coast and in the Columbia Plateau. The problems with this hypothesis, according to Snavely and others, (1973) is that it again fails to resolve the question of the origin of the calc-alkaline volcanics in the Cascade Range. In addition, it requires a uniform composition over a broad region with unique conditions to produce three different periods of magma generation (e.g., Grande Ronde-Depoe Bay, Frenchman Springs-Cape Foulweather, Pomona-Pack Sack).

The third model (Fig. 31c) suggested by Snavely and others (1973) involves partial melting well below the lithosphere with fractionation of the magma as it rises. Upon reaching the base of the lithosphere the magma may then spread laterally and vent along fracture zones at the coast and in the plateau. Again they point out that the difficulty of three magmas retaining compositional uniformity while migrating along the base of the lithosphere.

Recently, a fourth hypothesis (Fig. 31d) to explain this petrogenetic enigma has been proposed by Beeson and others (1979) and Swanson and Wright (preprint). Beeson and others suggest that the coastal basalts are not the result of local venting as Snavely and others (1973) surmised but rather are the western fringe of the Columbia River Basalts. In their opinion, the "coastal" basalts originated from the same vents and fissures as the plateau basalts and flowed through topographic lows in the Cascade Range onto a broad flat coastal plain (now occupied by the Williamette Valley and Coast Ranges

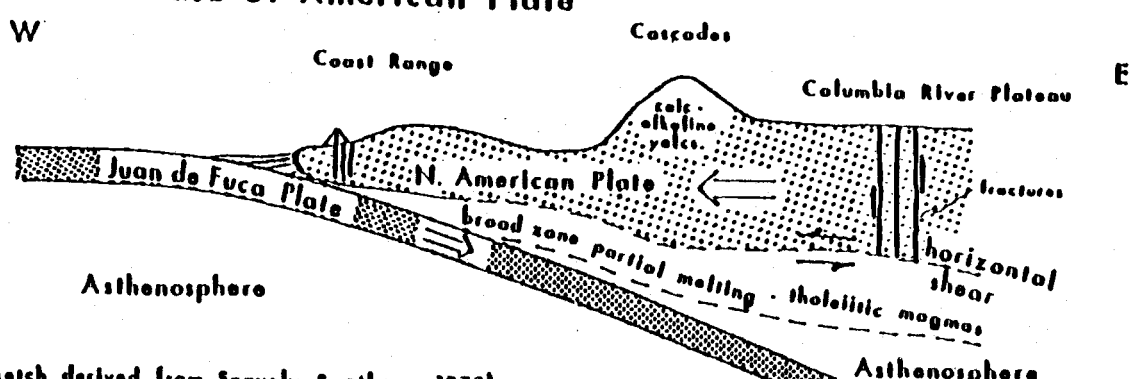
Models for Origin of Miocene Coastal Basalts and Coeval Basalts of the Columbia Plateau

MODEL 1 Partial melting of Juan de Fuca Plate



(sketch derived from Snavely & others, 1973)

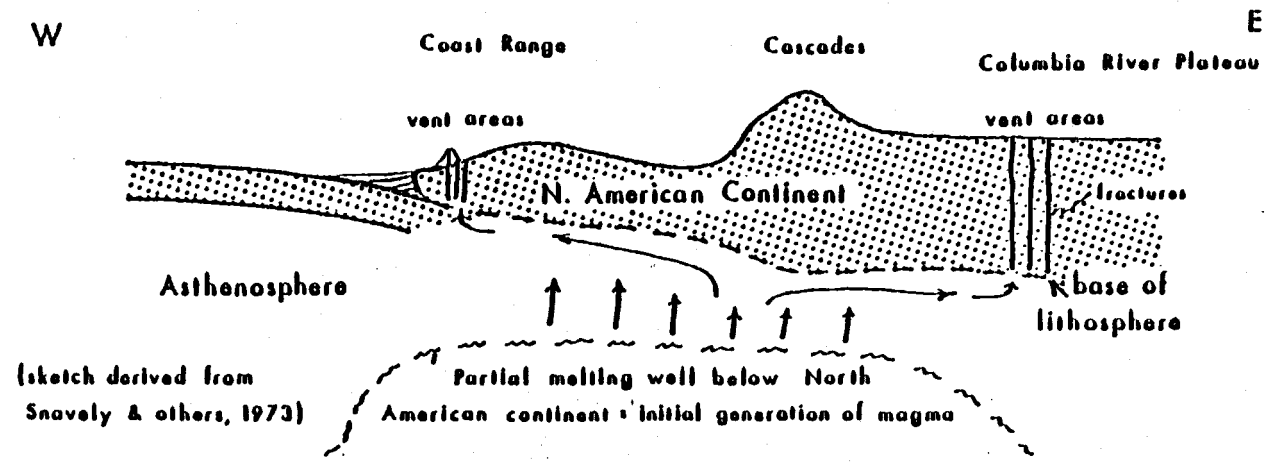
MODEL 2 Zone of horizontal shear & partial melting at base of American Plate



(sketch derived from Snavely & others, 1973)

Figure 31

MODEL 3 Partial melting below base of lithosphere



MODEL 4 Plateau-derived invasive flow

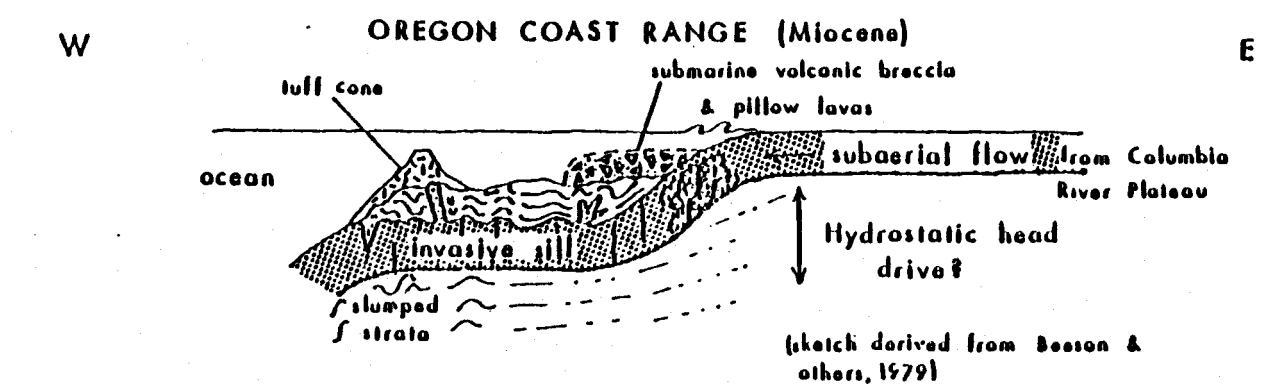


Figure 31 (continued)

of Oregon and Washington). Upon entering the sea, the very fluid basaltic lava invaded the relatively soft, water-saturated shelf sediments, producing invasive sills and dikes. In cases where the lava flowed over the irregular sea floor, hyalo- clastite breccias accumulated in topographic lows.

The one vent hypothesis is appealing because it simplifies the petrogenetic and chemical enigma outlined by Snavely and others (1973) and it suggests an explanation for why the geographic distribution of the Columbia River Basalts in the Willamette Valley and southwest Washington is nearly identical to the coastal basalts distribution. Swanson and Wright (preprint) noted examples of subaerial Columbia River basalt forming invasive dikes and sills in thin Miocene lake sediments on the plateau. Although the one vent hypothesis is appealing from a geochemical viewpoint, there are a number of field relationships of the coastal basalts in the Oregon Coast Range that are very difficult to explain applying the invasive hypothesis. Some of these relationships are:

1. The cross-cutting relationships of the Depoe Bay sills and submarine breccias by Cape Foulweather dikes (Penoyer, 1977 and Neel, 1976) and Depoe Bay dikes intruding Eocene camptonite breccia and sediment (Snavely and others, 1980);

2. Many middle Miocene dikes (up to 30 m wide and 100 m high) intrude from below and bow up the host Oligocene and Miocene strata such as the Young's River dike (Tolson, 1976). Where are the invasive feeder lavas for these dikes?

3. Many dikes also splay and brecciate upward into the overlying breccias as feeder dikes, such as at Haystack Rock (Neel, 1976), in the Saddle Mountain area (Penoyer, 1977), and in the Wickiup Mountain area (Coryell, 1978 and this study);

4. The presence of concentric ring dike systems and plugs, feeding overlying subaqueous hyaloclastite breccias at Cape Foulweather (Snavely and others, 1980);

5. Elongate northeast-southwest middle Miocene dikes (up to 30 m wide and 10-15 km long) in relatively undeformed older consolidated Oligocene-upper Eocene strata and within the Miocene basaltic breccias (Beaulieu, 1973; Penoyer, 1977; Goalen, in prep.); and

6. A Depoe Bay petrologic type sill 3000 m below a middle Miocene Depoe Bay flow, in upper Eocene-Oligocene strata in an offshore petroleum exploration well 19 km off the central Oregon coast (Snavely and others, 1980b).

Another aspect of the single vent hypothesis which has not been satisfactorily explained is the actual mechanics of invasion. The physical parameters would include the bulk density variation between the basaltic lava, water-saturated sediment and consolidated sedimentary rocks; the frictional forces of the sediment; the thickness of the flow; the cooling rates in an aqueous environment; and the gradient of the surface over which the flow is moving.

The density differential between the molten basaltic lava ($2.7 g/cm^2$) and the water-saturated sediment ($1.5-2\ gm/cm^2$)

may initiate the penetration of the lava into the sediment. Once started the hydrostatic head along with the density differential would push the magma as a wedge into the sediment in the same fashion as a dike intrudes from below. Once these forces were equalized by the resisting forces (increasing density of sediment and frictional forces) the lava would move hori- zontally continuing to wedge into the sediment along an equla force isobar. At this point the weight of the overlying sedi- ment would be added to the resisting forces. Dikes could sprout from the sill and move upwards into the sediment but since the major driving force is the hydrostatic head the dike could not intrude vertically higher than the initial point of entry of the lava.

This mechanism has been recognized on a small scale on the Columbia River Plateau by Swanson and Wright (preprint). Lake sediments up to 50 m thick have been gently raised and left relatively undisturbed by an advancing invading sill. Swanson and Wright have also recognized 15m high dikes which have intruded the overlying sediment.

If the invasion mechanism is a viable alternative to a local origin for the middle Miocene basalt along the Oregon Coast the hydrostatic head produced must be sufficient to penetrate over 3000 m of sediment as is suggested by the Depoe Bay sill discovered in the offshore well. Ponding of the lava in confined topographic lows (e.g., as in submarine can- yons) may increase the hydrostatic head, but rates of cooling due to brecciation of the lava would be greatly increased in a submarine environment. Theoretical and experimental work needs be done to better quantify the mechanics required for an "invasive flow". Any explanation must remain within the geo- logic field constraints just mentioned.

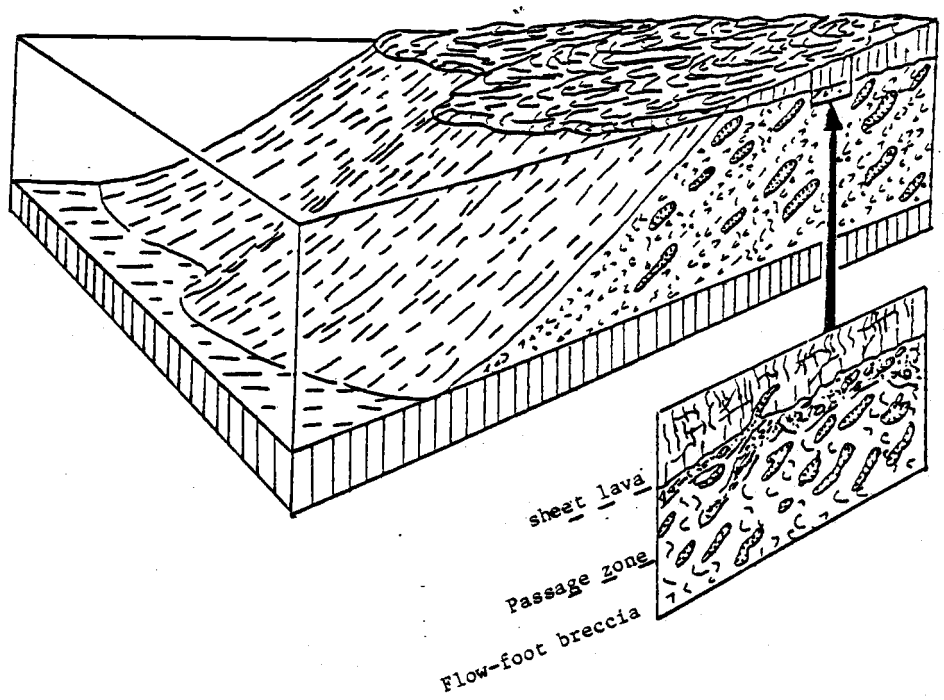
In conclusion, it would appear that none of the four hypotheses adequately explains all the geochemical and field relationships between the coastal and plateau-derived middle Miocene basalts. In the writer's opinion, the one vent hypothesis has serious difficulties from the viewpoint of a field geologist. The two vent hypothesis has difficulties from a geochemical viewpoint as Snavely and others (1973) pointed out. In order to prove or disprove a one or two vent hypothesis an important approach would be the detailed investigation of coastal areas where the plateau-derived and coastal basalts are thought to interfinger.

The Nicolai Mountain-Gnat Creek area is such an area because it is situated between the known Columbia River Basalts and mapped coastal basalts (Fig. 12). This detailed study does offer some insight into the understanding of the field relationships between these coeval basalts. The stratigraphic relationships described in the internal basalt stratigraphy section of this thesis do not conclusively support either a one vent or two vent hypothesis, rather it describes mappable differences between the consanguineous units. Although the two vent hypothesis is more strongly supported by this study based on the field relations. The Depoe Bay petrologic type is distinguished from its Grande Ronde equivalents on the basis of presence of dikes and bathyal foram-bearing mudstones associated with Depoe Bay Basalts; whereas subaerial basalt units and fluvial arkosic sandstone interbeds are associated with the Grande Ronde units. The relationship between the interbeds and the Grande Ronde basalts is best exposed in Big Creek where the Grande Ronde ppc-subaerial flow couplet formed a lava delta and is overlain by the fluvial or shallow-marine (?) Vantage Member (Cross Section B-B, Plate III). The total depth of water into which this lava flowed is estimated to be 50 meters, based on the thickness of the lava delta. The Depoe

Bay breccia is interbedded with mudstones that contain foram assemblages indicative of bathyal water depths (ranging from 400 to 800 meters). This disparity of water depth as well as the presence of "feeder" dikes in the lower Depoe Bay breccia leads to a hypothesis that these basalts are more likely local in origin and stratigraphically underlie their Grande Ronde chemical equivalents (see internal stratigraphy section).

The field relationship between the Cape Foulweather and Frenchman Springs units is not as straight forward. The Cape Foulweather petrologic type is defined as a series of sills, dikes, and hyaloclastite breccias associated with bathyal mudstones. The close proximity of one sill (OC-127) to the Cape Foulweather ppc-subaerial flow couplet unit in Big Creek Gorge creates some question about the "local" intrusive origin of this sill. This sill could be an "invasive sill" where the Frenchman Springs pillow breccia penetrated the mudstone which overlies it. It has also been correlated to a sill on the south side of the gorge (OC-714) suggesting it continues to the south. More geophysical work is needed in this poorly exposed area to clarify the field relationship of this topographically lower sill to the nearby Frenchman Springs flows.

This study shows that the Columbia River Basalts did flow into the middle Miocene sea in the Big Creek-Nicolai Mountain area as postulated by Snavely and others (1973). The field relationships suggest that the plateau-derived flows built "lava deltas" composed of thick flow-foot foreset breccias (and/or pillow lavas) and subaerial topset sheet lava. This is recorded no less than three and possibly four times within the study area (Plate II). When lava entered the water, it created a lava delta front of pillow palagonite complex material with primary dip slope angles of 30 to 40°. This material builds up until it equals the depth of water into which it is flowing



(Modified from Jones and Nelson (1970)

Figure 32 Idealized model for emplacement of the Columbia River Basalt into the Miocene sea in the Nicolai Mountain-Gnat Creek Area.

(Figs. 21 and 22). Additional ponded lava moves over the ppc lava delta front as a subaerial flow. Progradation occurs as lava cascades down the front of the delta, producing the inclined bedding commonly observed within the ppc flow units. Figure 32 is a sketch of this process modified from Jones and Nelson (1970).

Three important implications result from the above hypothesis for the emplacement of the Columbia River Basalt flows into the Miocene sea. One is that these sequences may record changes in sea level from emplacement of one flow couplet to the next. Jones and Nelson (1970) discussed a similar situation from their study of basaltic volcanoes in Iceland and Western Antartica. Second, thickening of the flow could decrease the slope or gradient in which the lava is moving, possibly resulting in backing the flow or ponding it within the ancestral Columbia River valley. This could account for the anomalous thickness (400 meters) mentioned by Beeson and others (1979) of subaerial Columbia River Basalt preserved along the present Columbia River Gorge.

The third implication concerns the point of entry of the lava into the sea. Since the subaerial part of a flow travels over the "build up" of the subaqueous lava delta part of a flow, progradation of the entire flow couplet occurs at the farthest seaward point where subaerial lava comes into contact with seawater. This flow mechanism suggests that a carapace is not likely to form over the liquid lava resulting in a residual liquid core which could be injected into the soft sediment or into the underlying breccias, as suggested by Beeson and others (1979). How far lava delta-front progradation could extend into the ocean is uncertain. Whether there would be sufficient

residual liquid lava present in cores of pillows within the delta breccias to inject down into the sediment to create 100 to 300 m thick sills and 10-15 km long dikes is doubtful.

Throughout this discussion, I have alluded to the problem of solving the questions about the origin of the coastal basalts and their relation to the plateau-derived basalts. The hypothesis put forth in this section on the emplacement of the Columbia River Basalts in Nicolai Mountain-Gnat Creek area as well as the stratigraphic relationships described in this study suggests many field difficulties with the one vent "invasive" hypothesis. However, I believe that more work must be done in order to fully evaluate the merits of these two hypotheses. Listed below are suggestions which may be useful to aid in resolving this perplexing problem. They include:

1) continued detailed mapping and stratigraphic study of the basalts and their association with the surrounding sedimentary units;

2) determination of the paleoenvironment of the emplacement of the basalts;

3) geophysical studies (e.g. gravity, magnetic, seismic) to see if the coastal dikes have roots at depth and whether there is a deep-seated structural system available as a conduit (Virginia Platt, Portland State University and Jim Olbinski, Oregon State University are starting projects to do this);

4) development of criteria to distinguish an invasive unit from an intrusive one, such as geometry of the sills, percentage of microlites in the chill margins as more should form in a flow travelling over a great

distance compared to a sill which is quenched rapidly (MacLeod, personal communication, 1980) and flow direction features within the sills (e.g. eastern plateau source ofr"local" source directions);

5) oxygen isotopic and total sulfur analyses of dikes in both regions to see if there is any variation, and;

6) theoretical considerations of the mechanics of invasion.

Critical areas within the Nicolai Mountain-Gnat Creek area which need further work for better understanding of the interfingering relationship include:

a) the south side of Big Creek Gorge and the relationship of the basalt flows present to the "known" coastal basalts around Wickiup Mountain described by Coryell, (1978);

b) the western end of Big Creek where the Frenchman Springs ppc and Cape Foulweather sill are found (OC-124 and 131) and;

c) the relationship of the R_2 low MgO Grande Ronde Basalt ($Tygr_1$) at the base of Nicolai Mountain to the underlying elongated reverse-polarized dike and to the reverse-polarized pillow basalts on Porter Ridge to the east.

On-going studies by graduate students under the guidance of Dr. Alan Niem at Oregon State University are now being conducted on these problems and hopefully will answer the question as to which hypothesis is valid.

In summary, this study aids in solving the problem of the local relationship between the Columbia River and coastal basalts by :

1) establishing the stratigraphy of various basalt units in a key area where these basalts interfinger;

2) mapping the extent of each unit;

3) differentiating and interpreting the paleoenvironment and sequence in which the basalt units were emplaced;

4) postulating the field relationships of the equivalent basalts, particularly the low Mgo Grande Ronde and Depoe Bay petrologic types;

5) establishing the close relationship between the Columbia River Basalt members of this study area with those present in the Western Cascades unit by unit based on geochemistry, magnetic, polarity, and stratigraphic position;

6) offering a prograding delta model for the emplacment of the Columbia River Basalts into the shallow marine environment, and;

7) delineating areas and possible methods which may be useful for future studies to solve the problem of one vent versus two vent hypothesis.

This problem is important to solve because it has implications as to the plate tectonic setting and basaltic volcanism of the middle Miocene in western Oregon and Washington (Fig.

31); the stratigraphic definition of whether there should be just one set of stratigraphic units (e.g., Columbia River Basalt Group) defined in Oregon and Washington (one vent hypothesis) or the present local coastal and plateau nomenclatures (two vent hypothesis); the ramifications as to how basalt dikes are viewed in other regions of the world (i.e., are there many more instances of "invasive" relationships not recognized); and finally a solution may shed some light on the current research revelations of counterclockwise rotation of the Oregon Coast Range during the Tertiary.

Clifton Formation

The Clifton formation is a 200-meter sequence of middle Miocene sandstones and siltstones exposed in the northen half of the thesis area. There has been considerable confusion as to the stratigraphic position, age, and depositional environment of this unit. Most authors have incorrectly referred to the unit by its possible age, stratigraphic position, or depositional environment. Previously, the Clifton formation has been described as: a "freshwater" sandstone by Washburne (1914); a Pliocene (?) sandstone by Warren and others (1945); an upper unit of the Astoria Formation by Lowry and Baldwin (1952); and late Miocene-Pliocene (?) fluvial and lacustrine deposits by Wells and Peck 9161) and Bromery and Snavely (1964). Niem and Van Atta (1973) and Baldwin (1976) described the unit as non-marine sedimentary rocks (at Clifton) while Schlicker and others (1972) and Beauliu (1973) described it as an upper Miocene sandstone.

In this study I propose Clifton formation as an informal name for this sequence of middle Miocene strata. The name was chosen for the geographical location of the Town of Clifton, Oregon (Sec 5, T8N, R6W on Plate I), where the unit is exposed

and was first described by Washburne in 1914. A type section is exposed in road cuts along the Clifton Road 1.5 km southeast of Clifton (Secs. 8 and 9, T8N, R6W; see measured section c-c' Appendix I). The basal unconformable contact is east of the county road where the Clifton overlies the Frenchman Springs basalt. Reference sections are also exposed along Tripp Road (section b-b') and Aldrich Point Road (section d-d', Appendix I). The top contact of the composite or upper part of the Clifton formation is found in a composite section with the overlying Pomona flow is located in a slump scarps below Aldrich Point (OC 585, Sec 31, T9N, R6W). The unit includes all the sedimentary rocks which lie above the Frenchman Springs Member of the Wanapum Basalt and the Silver Point member of the Astoria Formation preserved in the thesis area, and below the Pomona Member of the Saddle Mountains Basalt (Murphy and Niem, 1980). It does not include any sandstone strata which are minor interbeds within the other basalt formations of the Columbia River Basalt Group. It also excludes the 3- to 70-meter thick sandstone layer (Vantage Member) which is stratigraphically between the Grande Ronde and Frenchman Springs Basalts (see stratigraphic column Plate I). This definition restricts the stratigraphic position of the Clifton formation and precludes ambiguities related to interfingering relationships between formations.

The Clifton formation can be subdivided into three mappable members based on facies relations: 1) a sandstone facies (facies 1 on Plate I and Figure 33), 2) a middle siltstone facies (facies 2 and 3) a channel sandstone facies (facies 3). Facies 1 is further subdivided into two subfacies: 1a) a cross-bedded, burrowed sandstone facies and 1b) a cross-bedded, rippled sandstone facies. Throughout this study these informal members are referred to as facies to better describe the relationships between them.

Distribution

The outcrop distribution of the middle Miocene Clifton formation is concentrated north of U.S. Highway 30 and covers more than 32 square kilometers of the study area (Plate I and Fig. 33). The unit is fairly well exposed in road cuts, stream cuts, slump scarps, quarries, and in natural cliffs along the Columbia River. The sandstone facies (1) is best exposed along Ivy Station Road (OC 116, Sec.19, T8N, R7W), Tripp Road (OC 215, Sec. 19, T8N, R6W), Ziak-Gnat Creek Road (OC 177, Sec. 8, T8N, R7W), along a slump scarp near Aldrich Point (OC 585, Sec. 31, T9N, R6W) and in a large sandstone quarry along U.S. Highway 30 (OC 9, Sec. 18, T8N, R6W). The siltstone and channel facies (2 and 3) are exposed in road cuts along Clifton Road (Secs. 5,8,and 9, T8N, R6W). Many additional exposures of the siltstone facies occur along Aldrich Point Road (Sec. 2, T8N, R7W) and the logging raods above Gnat Creek State Park (Secs. 7, 8,and 13, T8N, R6W). Spectacular exposures of the channelized facies (3) are also found in Gnat Creek State Park (OC 484, Sec 13, T8N, R7W).

Due to the poorly consolidated nature of the sandstone strata of the Clifton Formation, the unit is very susceptible to slumping and stream erosion. These processes form the hummocky and intensely stream dissected hills characteristic of the northern parts of the study area (Plate I).

Lithology and Sedimentary Structures

The 200-meter thick Clifton formation is primarily composed of cross-bedded, micaceous, arkosic sandstone and clayey siltstone with minor lithic conglomerate, intraformational breccias, and rare lignite seams.

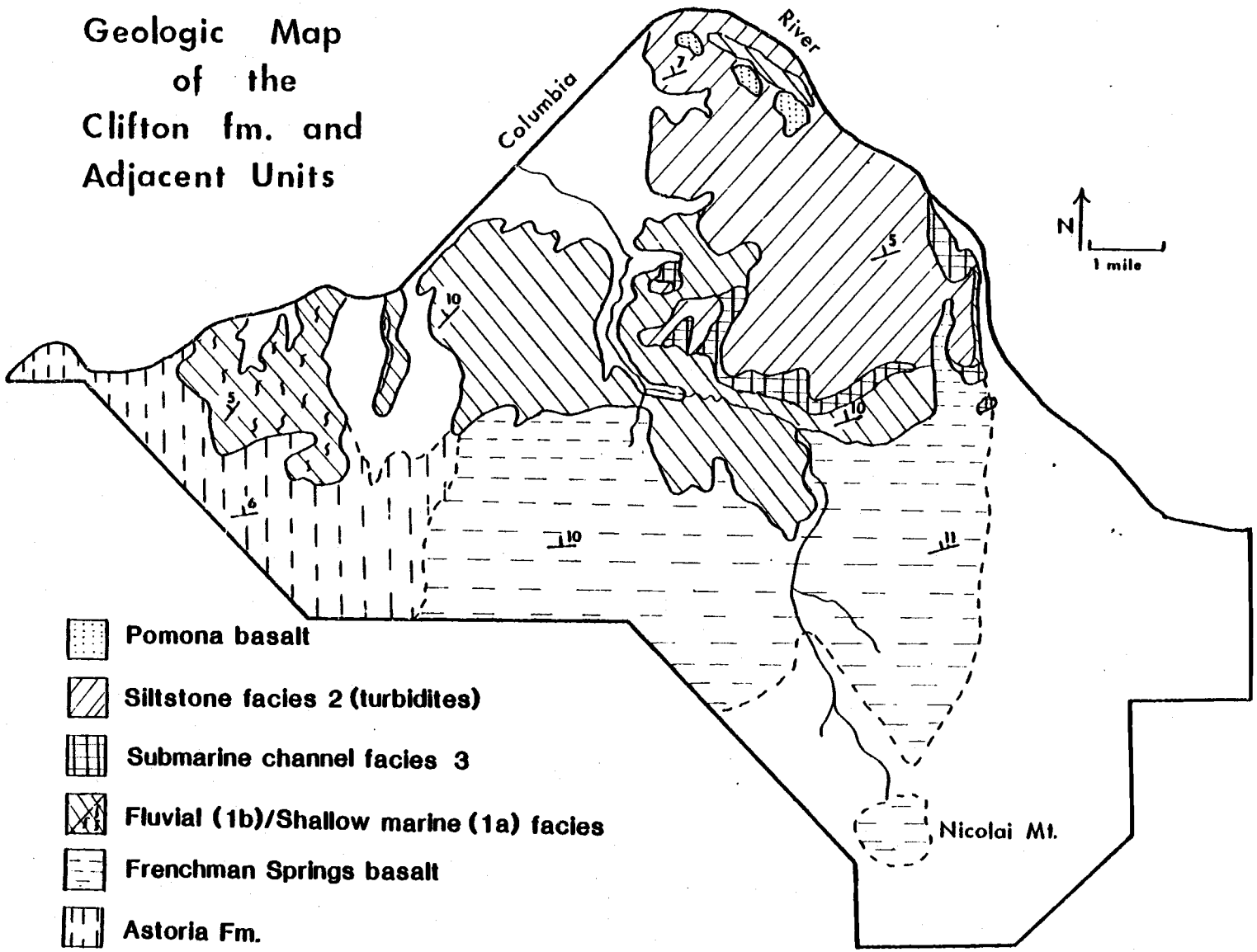

Geologic Map
of the
Clifton fm. and
Adjacent Units
Columbia
River
N
1 mile
Nicolai Mt.
Pomona basalt
Siltstone facies 2 (turbidites)
Submarine channel facies 3
Fluvial (1b)/Shallow marine (1a) facies
Frenchman Springs basalt
Astoria Fm.

Figure 33

As discussed in the nomenclature section, the Clifton formation has been divided into three major lithofacies based on field relations and laboratory analysis (see Fig. 33 and Plate I). Criteria used to distinguish these facies are summarized in Table 7, which forms the basis of discussion for the rest of this section.

Facies One

The stratigraphically lowest and highest unit of the Clifton formation is the sandstone facies (facies 1, Figs. 33 and 42). This lithofacies is estimated to be 120 meters thick and has been further sub- divided into two subfacies (1a and 1b) which are outlined in Table 7.

The most widespread unit of facies 1 is the cross-bedded sandstone subfacies (1b) which is estimated to be 90 meters thick calculated from outcrop and dip measurements (Fig. 33). This unit is composed of arkosic, micaceous sandstone interspersed with thinner mudstone beds and rare lignite seams. The ratio of sandstone to mudstone is estimated to be 9:1. The sandstone is texturally submature, cross-bedded, and rippled. The coarse to fine, mostly medium sand-sized grains are subangular. The sandstone is sorted to well-sorted and very friable, apparently never being cemented. Large-scale trough cross-beds (0.3 to 1 meter) are the dominant sedimentary structure although in-phase asymmetrical ripple marks (length 5-10 cm; height 2-3 cm) and rare symmetrical ripple marks were also observed in the large sandstone quarry near U.S. Highway 30 (OC 9, Sec. 18, T8N, R6W). The ripple sets usually are draped with very thin carbonaceous mudstone and are almost exclusively overlying the large-scale trough cross-beds (Fig. 34). Cylindrical iron concretions (17 cm) petrified wood

TABLE 7 Characteristics of the lithofacies in the Clifton Formation

Unit	Facies 1		Facies 2	Facies 3
	1a (marine)	1b (non-marine)		
Lithology	sandstone; silty sandstone	sandstone; minor mudstone; rare lignite	clayey siltstone; sandstone	sandstone; intraformational mudst. breccia; conglomerates
SS: Mudst. Ratio	75:25	90:10	30:70 range 10:90 to 80:20	75:25
Total thickness	20m	90m	140m	10-20m
Geometry	sheet	sheet	sheet	channels
Bed thickness	laminated to medium-bedded	thick to very thick mudst.- thin-bedded	thin-bedded to thinly laminated	thick to very thick amalgamated sandst. lens
Bedding character	discontinuous; parallel truncations; disrupted to bioturbated	even; parallel truncations of cross sets; sharp with mudst.	even; parallel; sharp; laterally persistant; alternating ss/ siltst.	discontinuous; curved; non-parallel
Relation to other facies	below and laterally equivalent to 1b; at base of formation over Astoria Fm.	overlies 1a and the Col. R. basalts; underlies and overlies facies 2; cut by facies 3	between upper and lower portions of facies 1; overlies and cut by facies 3	channels into facies 1b and 2; forms basal conglomerates on basalt
Composition	micaceous and carbonaceous arkosic wacke	micaceous arkosic arenite	volcanic wacke and volcanic arenite	volcanic wacke; micaceous arkosic arenite; volcanic-chert congl.
Color	med. gray (N5); wx. to grayish orange (10YR 7/21)	lt. gray (N7) to yellow orange (10YR 6/6)	med. gray (N5); wx. to very lt. gray (N8)	lt. gray to dark yellowish orange (10YR 6/6)
Grain size	fine to very fine sand	coarse to fine, mostly medium sand	silt with clay; very fine sand	coarse gravel to fine sand ang. boulders to pebble breccia of facies 2

TABLE 7 continued

Sorting	poor	mod. to mod. well	sandst. mod.	very poor
Rounding	angular	subang. to ang.	subang.	subang. to ang.
Textural maturity	immature	submature	sandst.- immature	immature to submature
Weathering character	friable to mod. indurated	very friable; bluff former; liesegang rings	siltst.- mod. to well indurated; sandst. - friable; prone to slumping	friable
Sedimentary structures	small scale trough cross-lam.; heavy mineral concentrations in micro cross-lam.; rare parallel lam.; plant mat.	large-scale trough cross-bedding; asymmetrical ripples; mud drapes; point bar sequences; possible grading; planar trunc- of cross- bed sets	graded sandst. to siltst.; Bouma T_{de}, T_{cde}, T_{ab}, and rare T_{bce} intervals; convoluted beds; flame structures; starved ripples; clastic dikes; sharp boundary contacts of sandst.	anastomosing channels disorganized and organ- congl.; graded pebbly sandst.; Bouma T_{ab}; flame structures; gravel lags; chaotic slump deposits; mudst. ripups
Fossils	Rosselia burrows; very rare gastropods and pelec rare foram tests in calcareous concretions	carbonaceous plant fragments; petrified wood	dominantly marine diatoms with mixed freshwater diatoms	none observed except in mudst. breccia clasts
Concretions	calcareous- up to 1m in diameter	cylindrical iron up to 0.8m in length	none observed	none observed
Distinguishing features	Rosselia burrows; large mica flakes; disseminated wood fragments; ripple trough cross-beds; high % fines in sandst. matrix	large scale cross-beds; planar truncation; point bar sequence; rare lignite seams; petrified wood fragments; Iron concretions	laterally persistent laminated sandst. and siltst.; graded beds with Bouma sequences; convoluted beds; marine and freshwater diatoms	anastomosing channels; amalgamated sandst.; graded beds; Bouma sequences; fragments of facies 2 incorporated as blocks up to 6m
Depositional environment	shallow marine offshore bar or lower shore face; river mouth bar	fluvial; mouth of a large river; point bars; minor marsh deposits	levee or overbank deposits (turbidite facies E) interchannel (turbidite facies D) and slope deposits (turbidite facies G)	submarine canyon head deposits; organized congl. (turbidite facies 1A); structureless amalgamated units (turb. facies B); chaotic blocks (turb. facies F)

fragments (up to 1.3m long), and iron stained liesegang banding are common in these thick-to very thick-bedded sandstones (up to 4m). Subfacies 1b is exposed in the lower part of the Clifton formation along Ziak-Gnat Creek Road (OC 177, Sec. 8, T8N, R6W) and in a commercial quarry along U.S. Highway 30 where it is mined by the Portland Cement Company. Another excellent exposure occurs in a 100-meter long slump scarp at Aldrich Point at the top of the Clifton formation (OC 585, Sec. 31, T9N, R6W). Chemical analyses of this sandstone made by the Portland Cement Company are listed in Appendix VII (compliments of Stewart and Les Lahti). It is high in SiO_2 (67 - 77%), Al_2O_3 (11.5-17.6%), and K_2O (1.8-2.0%), reflecting the abundance of quartz and potassium feldspar in this sandstone.

The mudstone interbeds are thin-bedded (5-10cm), light gray (N7), highly micaceous, carbonaceous (1-10%), clayey siltstones. Many fragments of fossil leaves and twigs occur along the bedding planes. The contacts between mudstone and sandstone beds are sharp, usually parallel, and even in the classification of Reineck and Singh (1975, p. 82), although mudstone commonly forms uniformly thick (1-10 cm), undulatory contacts (i.e., mud drapes) outlining the underlying sandstone ripple sets. The contact between thick individual sandstone beds is usually indistinguishable but commonly it occurs as planar truncations between adjacent trough cross-bedding sets.

Warren and others (1945) mentioned the presence of lignite in their "Pliocene" sandstone unit (facies 1b in this paper). Two coaly seams, (0.5 to 1.0 meter thick) were discovered in the thesis area. One is near the Ziak-Gnat Creek Road stratigraphically low in this subfacies (OC 178, Sec. 8, T8N, R7W). This seam is present 80 meters behind Stewart and Les Lathi's house in a heavily vegetated area along a bull dozer cut. The other is approximately 25 meters below the upper contact with

Figure 34 Point bar sequence in facies 1b of the Clifton formation as depicted in figure 45. Parallel laminations absent. Sandstone Quarry off U.S. Highway 30 (OC 9, NW, Sec. 19, T8N, R7W).

Figure 35 Coal bed (lignite) and associated underclay in fluvial facies 1a of Clifton formation. Exposure is 25 meters below contact with the Pomona Member. Aldrich Point (OC 585, NW, Sec. 31, T9N, R6W).

the overlying Pomona flow on a slump scarp near Aldrich Point (OC 585, Sec. 31, T9N, R6W) (Fig. 35). Lignite is a low-rank, brownish black coal which cracks badly upon drying and retains its original wood structures (Pettijohn, 1975). All these features are characteristic of the "coal" found in the thesis area. The coaly beds appear black in fresh exposures but oxidize to brown upon prolonged exposure to air. The presence of a well developed, 6 cm, light brown, carbonaceous underclay suggests that the lignite originated as an autochthonous coal in marsh areas along the fringes of the sandstone deposit. The two coal seams were traced laterally for distances of 10 and 80 meters, respectively.

A cross-bedded, burrowed sandstone typifies subfacies 1a of facies 1. This sandstone is arkosic, micaceous, and highly carbonaceous. It is poorly sorted, fine- to very fine-grained (up to 25% clay and silt). The sandstone is medium gray (N5) when fresh but usually is light gray (N7) to light brown (5Y 6/6) in exposures. Discontinuous bedding and non-parallel bedding contacts (Reineck and Singh, 1975) characterize this unit due to the homogenizing effect of bioturbation on the laminated to medium-thick sandstone beds (up to 0.3 meters). The abundant trace fossil burrows which consist of vertical back-filled tubes 3-10 cm long and 3 cm in diameter were identified as Rosselia burrows by C.K. Chamberlain, (1980, written communication). Trough cross-bedding (5-10 cm amplitude) (Fig. 36) is the principal sedimentary structure in the unit. Carbonized wood fragments up to 3 cm long show no preferred orientation within the sandstone. Calcareous concretions up to 1 meter in diameter are characteristic of this subfacies. These are of interest because they represent the only source of megafossils in the Clifton Formation. Fossils from the concretions included a high-spiraled gastropod, pelecypod shells, and foraminifera test (OC 111, Sec. 24, T8N,

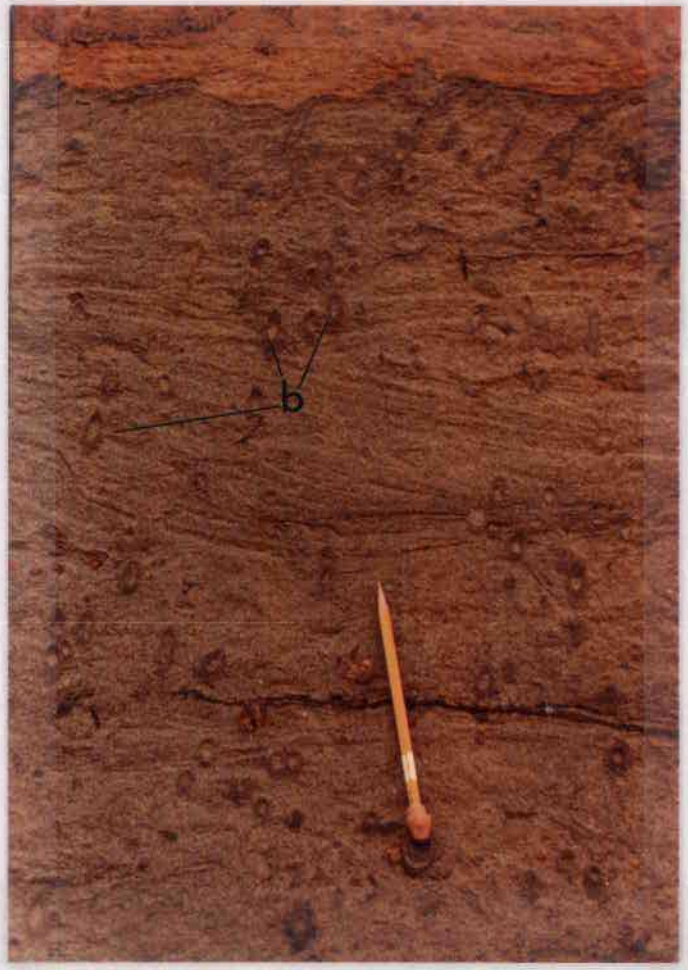


Figure 36 Shallow marine sandstone facies 1a of the Clifton formation characterized by Rosselia burrows and trough cross-bedding. Note concentration of heavy minerals and carbonaceous plant fragments along bedding planes. Ivy Station Road (OC 116, SW, Sec. 13, T8N, R7W).

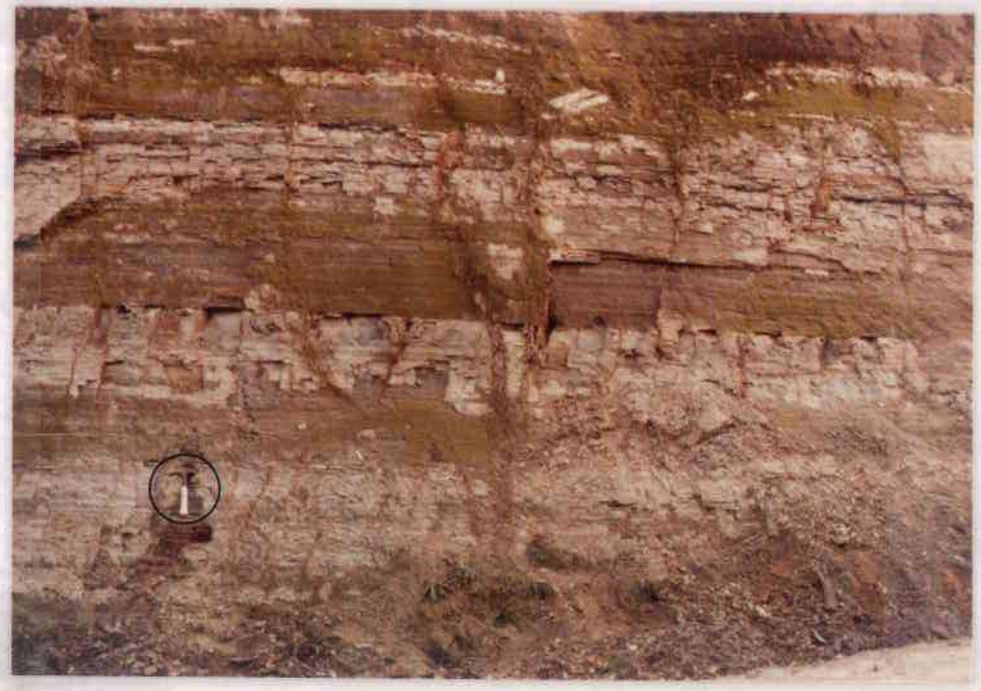


Figure 37 Typical exposure of siltstone and sandstone of facies 2 in the Clifton formation. This unit represents interchannel slope facies D and G of Mutti and Ricci-Lucchi Model (Table 8). Hammer for scale (OC 511, SW, Sec. 7, T8N, R7W).

R8W). This 30-meter thick sandstone and silty sandstone is exposed in the northwestern part of the study area (Plate I and Fig. 33). This subfacies is described in Table 7 and in a measured section along Tripp Road (Sec. 19, T8N, R7W) where it unconformably overlies the Astoria Formation (reference section b-b', Appendix I).

Facies 2

The siltstone lithofacies which makes up 60 percent of the Clifton formation is estimated to be 140 meters thick and to lie stratigraphically between the upper and lower parts of facies 1 (Fig. 42). Lithofacies 2 is composed of medium gray (N5), highly carbonaceous, micaceous, clayey siltstone interbedded with very fine-grained arkosic sandstone. The siltstones are well laminated and moderately indurated whereas the sandstones are moderately sorted and very friable. The sandstone to siltstone ratio varies from outcrop to outcrop, ranging from 10:90 to as high as 80:20 with an average ratio of 30:70. As the ratio of sandstone to siltstone decreases, the chance of mass movement caused by slumping appears to increase (Beaulieu, 1973). Bed thickness varies from thin laminations to medium thick beds (3 mm to 40 cm). The siltstone beds are thinly to thickly laminated, whereas the thickness of the sandstone layers ranges from 3 mm to 40 cm. Contacts between individual beds are parallel and even (Reineck and Singh, 1975) although some thick-bedded sandstones are lenticular in shape. The contacts are predominantly sharp, linear, and continuous in outcrop except where sandstone lenses, flaser bedding, flame structures, sandstone dikes, intrabed slump folds, and convoluted beds disrupt the plane (Fig. 37).

In facies 2 the thicker sandstone layers (normally associated with the channels of facies 3) consist of normally

graded sandstone (medium- to very fine-grained) overlain by parallel laminations, rare microcross-lamination, and thin mudstone layers (Bouma sequence T_{abe} and T_{abce}). Thin pebbly layers are locally present along the sharp base of these beds. The upper contact is gradational with the overlying siltstones. Moving away from the channel walls of facies 3 sandstone beds are finer grained, more thinly bedded, and contain concentrations of mica flakes and carbonaceous material in laminae. These very thin beds grade upward into the siltstones. This vertical change suggests a density grading within individual layers and is interpreted as Bouma T_{de} intervals. Locally, starved ripples associated with these beds expand the sequence to T_{cde} intervals. The intervening even-bedded siltstone and clayey siltstones are interpreted as periods of quiet hemipelagic deposition between turbidite events. These turbidite features suggest rapid deposition of sediment in an environment of fluctuating energy regimes.

The siliceous marine and freshwater diatoms used for the age determination of the Clifton formation were found in the laminated clayey siltstone within this facies at the type section along Clifton Road (OC 14,24, 25, Secs. 8 and 9, T8N, R6W; Appendix IV). As is characteristic of most of the Clifton Formation, no calcareous mega or micro fossils are preserved.

Facies 2 is exposed over much of the northern part of the thesis area (Fig. 33 and Plate I). This lithofacies is well exposed along Clifton Road (Sec. 8, T8N, R6W) where it is cut by channel facies 3. The unit is best exposed along logging roads, in stream cuts and in slump-prone hills north of Gnat Creek State Park (Secs. 7,8 and 18, T8N, R6W). It is also exposed in a creek bed below the upper unit of sandstone facies 1 below the slump scarp near Aldrich Point (OC 585). Reference section c-c' along Clifton Road describes the relationship

between facies 2 and 3 (see Internal Contact Section and Appendix I).

Facies 3

The third lithofacies of the Clifton formation is composed of channelized, arkosic, micaceous sandstone. This semi-consolidated, thick-bedded sandstone is channeled into siltsone facies 2. In some cases, the channelized facies cuts into the underlying sandstone facies 1b and has even eroded down to the underlying Frenchman Springs basalt at the base of the Clifton formation (Clifton Road, Sec. 9, T8N, R6W) and noth of U.S. Highway 30, OC 355, Sec. 21, T8N, R6W). This yellow-orange (10 YR 6/6) sandstone is poorly sorted and immature to submature both texturally and compositionally. It consists generally of coarse angular grains of quartz and feldspar and very-coarse flakes of muscovite and biotite. The sandstone can be very thick-bedded, structureless, and iron-stained in outcrop as at Gnat Creek State Park. Gravel lag deposits are present at the base of most channels.

The most spectacular feature of the channelized facies 3 is the abundant angular to subrounded clasts of well-laminated siltstone of facies 2 in channels of this unit (Fig. 38). This feature is well exposed along Clifton Road and in Gnat Creek State Park (OC 484, Sec. 13, T8N, R7W). The siltstone clasts are elongate parallel to the bedding and range in length from 2 cm to 8 meters (Fig. 39). They commonly form a disorganized, framework-supported, very poorly sorted chaotic breccia in a coarse, arkosic sandstone matrix. Undercutting the facies 2 siltstone channel walls by turbidity and tide-induced currents with subsequent slumping of semi-consolidated siltstone walls into the channels is postulated as a mechanism for emplacement of some of the clasts. A few large blocks up to 2 meters long

Figure 38 Submarine sandstone channel facies 3 of the Clifton formation composed of disorganized, chaotic breccias of siltstone facies 2 in a coarse sandstone matrix. This lithofacies represents facies F of Mutti and Ricci-Lucchi (1973) turbidite model (Table 8) and probably represents channelized debris flow deposition. Hammer for scale (OC 25, Sec. 8, T8N, R6W).

Figure 39 Close-up of chaotic blocks of well laminated siltstone pictured in fig. 38. Note irregularity and angularity of the blocks. Matrix is a pebbly sandstone (OC 25).

of the cross-bedded, carbonaceous sandstone from the underlying facies 1 were also found in some channel-fills. The packing density of these blocks, which range from 1 to 90% of an outcrop, is a function of the proximity of the outcrop to the original slump site. Some of the disorganized siltstone clast conglomerates were transported away from the slump site by debris flows (Fig. 38). Poorly sorted volcanic quartzitic, pebble conglomerates occur in the channel-fills as lenticular irregular, discontinuous beds. These 0.5 to 1.5 meter thick conglomerates are locally graded. The pebbles are subrounded to well-rounded and are composed of andesite, basalt and tuffs (volcanics), quartzite and schist (metamorphics), granitic and rare sedimentary clasts. A detailed list of lithologies from pebble counts is provided in Appendix VII.

Individual channels range in size from 5 to 25 meters across and less than 2 to 10 meters in depth. These channels anastomose, thereby forming deeper and wider cuts into the underlying siltstone; some of which are estimated to be 40-50 meters across (Fig. 10). The larger channels are filled with amalgamated, thick, structureless, coarse-grained sandstone. Bedding in the channels generally conforms to the channel geometry, being curved, unconformable to bedding in the channel walls and discontinuous (Reineck and Singh, 1975). Iron-stained liesegang rings may mask the form of the channel, but orientation of siltstone chips and pebble conglomerate lag deposits, where present, outline the channel configuration (Fig. 41).

This facies distribution is outlined in Plate I and on Fig. 33. Characteristic exposures occur in Gnat Creek State Park and along Clifton Road where a measured section was described (reference section c-c'; Appendix I).

Figure 40 Anastomosing channels in Clifton formation (facies 3). Note the channel with large siltstone blocks and underlying siltstone levee deposits at facies 2. Dots outline the contact of two amalgamated channels. Outcrop along Clifton Road (OC 25, NE, Sec. 8, T8N, R7W).

Figure 41 Al Niem standing on a structureless arkosic sandstone with imbricated siltstone chips (up to 30 cm.) which outline amalgamated channels of facies 3. This unit probably was deposited downchannel from a channel wall slump site by grain flow mechanisms. This lithology corresponds to facies B of Mutti and Ricci-Lucchi Model (Table 8). Turbidite in Gnat Creek State Park (OC 484, Sec. 13, T8N, R7W).

Contact Relations

The Clifton formation unconformably overlies the Frenchman Springs Member of the Wanapum Basalt in the eastern part of the thesis area, and the Astoria Formation in the western part of the area. Overlying the Clifton formation in the north part of the study area is the subaerial Pomona Member of the Saddle Mountains Basalt.

Lower Contact

The unconformable relationship between the Frenchman Springs Member and the Clifton formation is best exposed along Clifton Road (Oc 550, Sec. 17, T8N, R6W) in Hunts Creek (OC 18, Secs. 17 and 19, T8N, R6W on Plate I), and along an overgrown section of old U.S. Highway 30 at the crest of Nicolai Ridge (OC 355, Sec. 21, T8N, R6W). The contact is marked by a 3-meter thick basal volcanic conglomerate of facies 3 which rests upon an irregular erosional surface of vesicular, subaerial basalt. The contact is also exposed along Gnat Creek below the U.S. Highway 30 bridge (OC 616, Sec. 4, T8N, R7W), where a fine-grained, cross-bedded, fluvial sandstone of facies 1 overlies the basalt.

The relationship between the Clifton formation and the Astoria Formation is exposed along Tripp Road (reference section b-b', Appendix I) and in railroad cuts which parallel the Columbia River (Sec. 14, T8N, R8W). In these areas, a local unconformable relationship between the Astoria Formation and the Clifton formation is suggested by: 1) difference in age, 2) sharp change in lithology and depositional environments, 3) slight angular unconformity, and 4) the presence of hundreds of meters of stratigraphically intervening subaerial basalt.

Assemblages of the Foraminifera from fossil localities 208, 212, and 216 (Plate I) indicate an early to middle Miocene Saucesian age (Rau, 1980, written communication) for the Astoria Formation in this area. Molluscan fossils of Pillarian to lower Newportian age (Moore, 1980, written communication) from the Astoria Foramation at localities 91 and 208 support the date indicated by the Forminifera (Fig. 2). In contrast, diatoms from the Clifton formation indicate an upper middle Miocene age. The diatom zones, subzone b of Denticula hustedtii-D. lauta and subzone of D. lauta equate to the uppermost Relizian, lower Luisian forminiferal stages (Barron, 1980, written communication). The fossil data indicate that the lower part of the Relizian stage may be missing between the two units and that the Clifton formation is younger than the Astoria Formation.

The sharp change in depositional environments and lithology between the underlying Silver Point and Pipeline members of the Astoria Formation and facies 1 of the Clifton formation suggests a hiatus between these two units. Foraminifera of the Astoria Formation lived at outer shelf to middle bathyal depths (100 to 400 m) (Rau, 1980 written communications). The mollusks inhabited depths of middle to outer shelf (Moore, 1980, written communication). These water depths are much greater than the very shallow-marine lower shoreface depth (3-10 m) indicated for facies 1 of the Clifton formation by vertical Rosselia burrows in the cross-bedded sandstone (Chamberlain, 1980, written communication) which immediately overlies the deep-marine Astoria. Rosselia burrows are abundant in the Clifton in road cuts along Tripp Road (OC 215) and Ivy Station Road (OC 116, Secs. 24 and 14, T8N, R8W) immediately above the contact. The mixed assemblage of freshwater and shallow marine diatoms within the siltstone

facies 2 of the Clifton formation also indicates a nearshore shallow marine shelf environment (0 to 200 meters) (Barron, 1979, written communication).

A slight angular unconformity between the Silver Point mudstone and the overlying sandstone facies 1 of the Clifton formation is postulated from strike and dip measurements taken along the railroad tracks north of Svensen, Oregon (Plate I). From this relationship, a disconformable contact of a few degrees in dip is inferred for the contact between the Pipeline member of the Astoria Formation and sandstone facies of the Clifton formation. This disconformable relationship is best exposed along Tripp Road although the actual contact is covered (Sec. 19, T8N, R6W; see section b-b' Appendix I).

The most convincing evidence that there is a local unconformity between the Astoria Formation and overlying Clifton formation is the presence of almost 230 meters of Columbia River subaerial basalt flows stratigraphically between the two units in the eastern part of the study area (Fig. 5). The Astoria Formation has been defined in northwest Oregon as the mappable sedimentary unit which is older than and intruded by the coastal basalts and subaerial Columbia River Basalt (Niem, 1980, personal communication; Coryell, 1978; Penoyer, 1977). The presence of Pillarian and Newportian molluscan fossils (Moore, 1980, written communication) in fine-grained Big Creek sandstone and siltstone in the West Creek in the eastern part of the study area (OC 245, and 246 Sec. 1-T7N-R6W) suggest that the Astoria Formation lies stratigraphically below the subaerial Columbia River Basalt that forms Nicolai Ridge. This thesis has defined the Clifton formation as the sedimentary unit above the Frenchman Springs member of the Wanapum Basalt (see Nomenclature Section). Therefore, the time represented by these hundreds of meters of flows and

intervening sandstone interbeds in the study area is not represented at the Astoria Formation-Clifton formation contact in the western part of the study area. In conclusion, a time lapse of sufficient length to account for the sharp change in depositional environment between the Astoria Formation and the Clifton formation, approximately equal to the emplacement of subaerial basalt, is essential.

Upper Contact

Unconformably overlying the Clifton formation is the subaerial flow of the Pomona member of the Saddle Mountains Basalt. The contact is exposed in a slump scarp high on the cliffs facing the Columbia River in the extreme northern part of the thesis area near Aldrich Point (OC 585, Sec. 31, T9N, R7W; on Plate I). The contact consists of highly weathered iron-stained exfoliated fractured basalt on fine-grained, cross-bedded, non-marine sandstone of facies 1. A coal bed occurs 25 meters below the contact. The contact is very sharp with no peperitization or pillowing of the lava into the sandstone, suggesting that non-marine sandstone was in the process of being eroded immediately prior to the time of emplacement of the lava flow. Anderson (1980) has shown that the subaerial Pomona flow is commonly confined to canyons and unconformably overlies the fluvial basaltic conglomerate in the Columbia River Gorge near The Dalles, Oregon.

Intraformational Contacts of the Clifton Formation

The vertical and lateral association of facies 1-3 in the Clifton formation is depicted in Fig. 42. The geographic distribution of the facies is shown in Fig. 33 and Plate I.

Schematic Cross Section of the Facies within the Clifton fm.

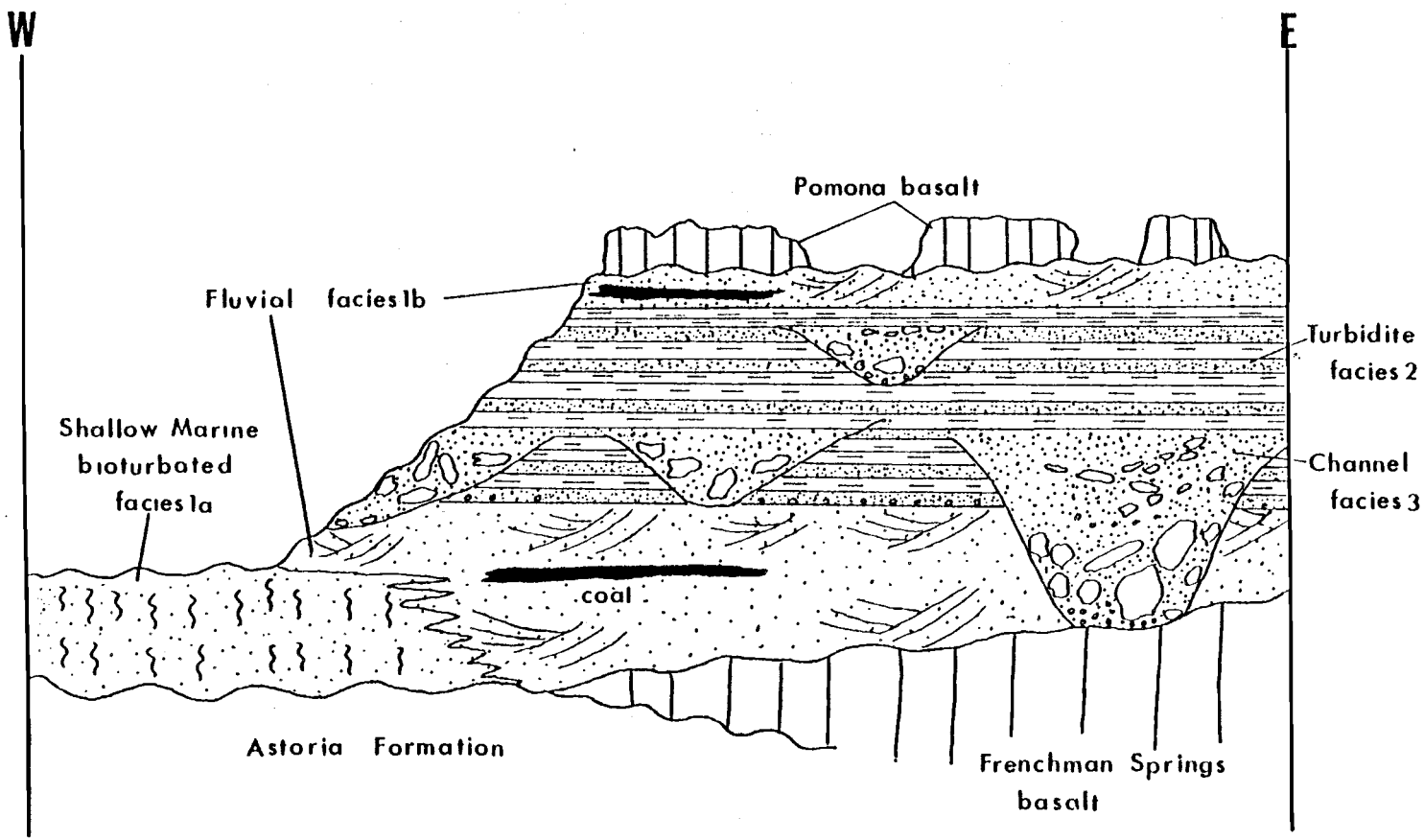


Figure 42

The cross-bedded, fluvial-marsh sandstone subfacies 1b occurs both at the base and top of the Clifton Formation. The lower subfacies laterally interfingers with and overlies the shallow-marine bioturbated sandstone subfacies 1a in the western part of the outcop area. This associaton is inferred from outcrop patterns near Knappa High School (OC 111, 214 and 215) and in the lower Big Creek area (OC 79 and 84). The shallow-marine subfacies (1a) is exposed only in the western part of the area and is never found stratigraphically above the fluvial-marsh sandstone facies (1) and the thin-bedded, turbidite siltstone facies (2). The contact is not exposed in the area. It is inferred from the outcrop pattern north of Bradley State Park and along Davis Bottom Road (Secs. 11 and 12, T8N, R7W). The upper contact between the turbidite-siltstone (facies 2) and the overlying cross-bedded fluvial-marsh sandstone (facies 1) near the top of the Clifton formation is well exposed in a creek bed a few hundred meters below a large slump scarp that exposes the Pomona Basalt-Clifton formation contact near Aldrich Point (OC 585). It is an erosive contact that is sharp and planar between consolidated well-laminated, siltstone (facies 2) and the overlying, very friable fluvial-marsh sandstone (facies 1).

The submarine channel sandstone and breccia (facies 3) is laterally equivalent to and cuts into the thin-bedded turbidite-siltstone (facies 2) that forms most of the middle part of the formation into the underlying lower fluvial-marsh sandstone (subfacies 1b) and at places have eroded down to the subaerial basalt which underlies the Clifton formation. This latter relationship can be seen near the waterfalls along Clifton Road (OC 18, Sec. 9, T8N, R6W) and along a section of old U.S. Highway 30 (OC 355, Sec. 21, T8N, R6W). The sharp lower contact is erosive with channel-fills containing basal pebble conglomerate lag deposits. Numerous road cuts along

Clifton Road expose a series of channels which cut the flat-lying, thick-bedded turbidite sandstone and laminated siltstone beds of facies 2. The anastomosing channels result in a complex interfingering relationship between facies 2 and 3.

The contact between the lower cross-bedded fluvial-marsh sandstone (facies 1) and the submarine channel sandstone (facies 3) is exposed along Davis Bottom logging road near the intersection with Gnat Creek Road (OC 525 and 527, Sec. 12, T8N, R7W). In this area, a channel filled with thin-bedded turbidites has cut into the cross-bedded fluvial sandstone, forming a sharp, irregular erosive contact.

In summary, the contacts of the deep marine thin-bedded turbidite siltstone (facies 2) and channel sandstone (facies 3) with the non-marine cross-bedded fluvial sandstones (facies 1b) are sharp, planar or irregular, erosive contacts. Those between the thin-bedded turbidite siltstone facies and the submarine channels sandstone are both erosional and depositional. The channel breccias cut into underlying siltstone beds and are depositional where spillover flow from channels formed the overbank (levee) and interchannel turbidite-siltstone deposits of facies 2. There is a lateral interfingering relationship between the shallow marine bioturbated subfacies (1a) and the fluvial-marsh subfacies (1b) in the lower part of the Clifton Formation.

Age and Correlation

The Clifton formation was previously undated but suspected to be upper Miocene-Pliocene (?) by Lowry and Baldwin (1952) and by Wells and Peck (1961) based on a tentative stratigraphic position above the middle Miocene Columbia River Basalt. The formation has been dated in this study as middle Miocene on the

Age and Stratigraphic Position of Clifton formation

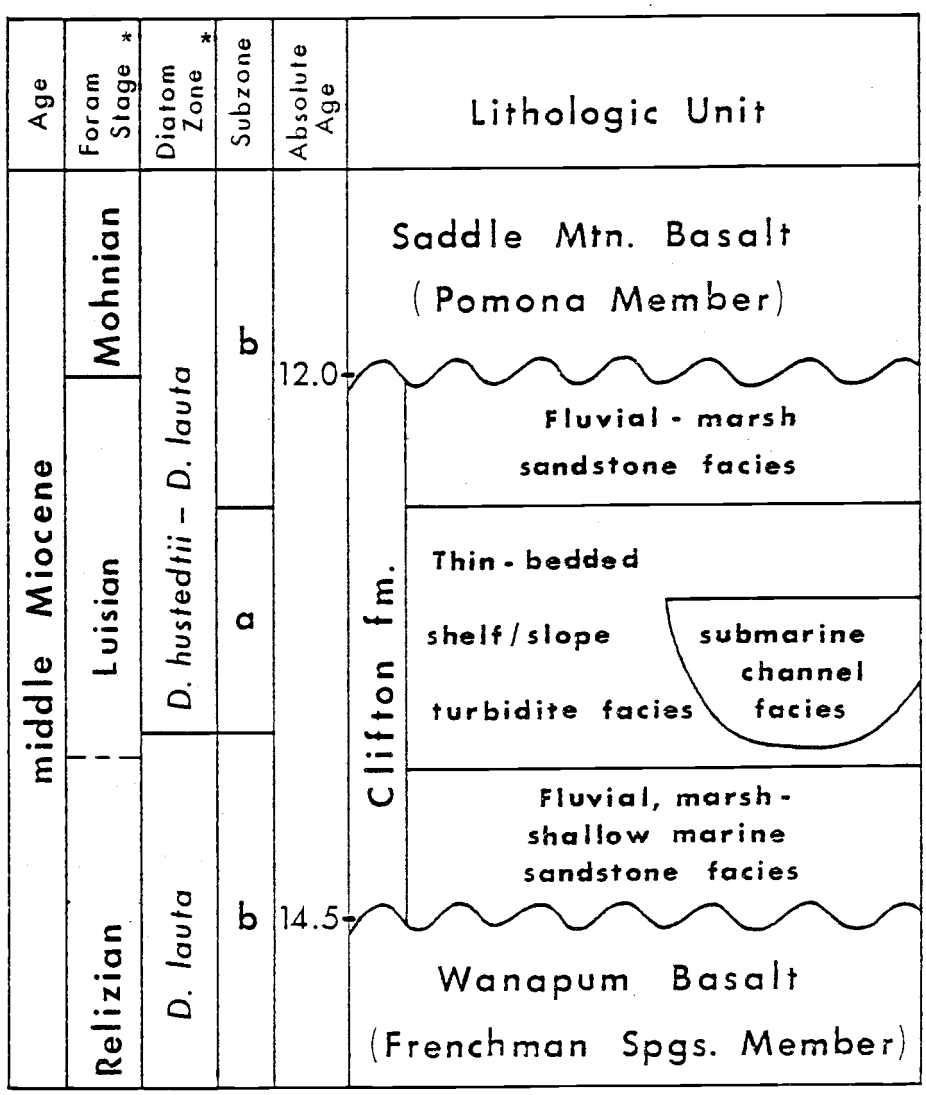


Figure 43 California time scale (*) from Barron (1980, written communication). See figure 2 for Oregon-Washington time scale from Armentrout (1980, written communication.

basis of diatom fossils and stratigraphic position between the Columbia River Basalt units (Murphy and Niem, 1980) (Fig. 43). Several benthic and neritic diatom assemblages from the siltstone facies 2 placed much of the unit in subzone b of the Denticula lauta zone to low in subzone a of the D. hustedtii-D. lauta diatom zone of southern California (John Barron, 1980, written communication; see fossil list in Appendix IV and Fig. 43). In southern California, the D. lauta zone equates to the Luisian Stage and possibly the upper Relizian foraminiferal Stage. Subzone a of the D. hustedtii-D. lauta zone falls immediately below the middle Miocene Mohnian-Luisian foraminferal Stage boundary of southern California (Barron, 1979 and 1980, written communication). Recently, Armentrout (1980) reevaluated the correlation between the diatom zones of the North Pacific and the foraminiferal stages of western Oregon and Washington with the worldwide time correlation chart. He suggests (personal communication) that the Relizian/Luisian foraminiferal boundary is time-transgressive between southern California and western Oregon and Washington. That is, what has been defined as Relizian in the Pacific Northwest would include portion of the Luisian Stage in southern California.

In Oregon subzone b of D. lauta and subzone a of D. hustedtii-D. lauta equate according to Armentrout (1980, personal communication) to the Relizian foraminiferal stage rather than to the Luisian Stage. Therefore, diatoms of subzone b of the D. lauta zone and subzone a of the D. hustedtii-D. lauta zone in the Clifton formation indicate that the unit is Relizian in age (middle Miocene). This correlation between southern California and Oregon by Armentrout is based, however, upon Relizian forams in the sandstone at Whale Cove near Newport, Oregon (Armentrout, 1980, personal communication). These sandstones are overlain by Cape Foulweather Basalt which

is a petrologic, chemical, and age equivalent of the Frenchman Springs basalt (see Basalt Stratigraphy Section). In this study area the Frenchman Springs basalt underlies the Clifton formation. This stratigraphic relationship suggests that the Clifton formation is younger than the Relizian sandstone at Whale Cove and may be actually Luisian or upper Relizian in age as Barron suggested on the basis of diatoms.

The Clifton formation is stratigraphically overlain and underlain by lavas of the Columbia River Basalt Group. This affords the chance to bracket the formation between radiometric ages established for these flows in eastern Oregon. The age of the overlying subaerial Pomona Member of the Saddle Mountains Basalt, dated by K/Ar methods at 12.0 m.y. in eastern Oregon (McKee and others, 1977) (12.3 m.y. new scale) places an upper limit on the absolute age of the sedimentary unit. The absolute age of the underlying Frenchman Springs Member is 14.5$\pm$ 0.5 m.y. (14.9 m.y. new scale) based on K/Ar dating of this basalt in eastern Oregon (Don Swanson, 1980, personal communication). The radiometric age of the Pomona flow falls within the Montesano type foraminiferal stage which is younger than the Luisian stage. The Luisian spans from 14.0 to 12.9 m.y. (Armentrout, 1980) (Fig. 2). Therefore, although the diatom-bearing middle siltstone facies of the Clifton formation could possibly be only Relizian in age as suggested by Armentrout's (1980) timetable, the upper fluvial facies of the Clifton probably is Luisian in age. Compositionally and texturally similar fluvial sandstone above the Pomona flow indicates that there is not a large hiatus between the Clifton formation and overlying Pomona flow.

The radiometric dates allow crude estimation of the sedimentation rate for the Clifton formation. Using an approximate thickness for the formation of 200 meters and a

time interval of 2.5 m.y. between the Pomona and Frenchman Springs flows the rate is 8 cm/10^3 years.

On the Washington side of the Columbia River directly across from Aldrich Point, Oregon a sequence of sandstone and siltstone beds with lithologic and facies characteristics similar to the Clifton formation has been identified by Ray Wells (personal communication, 1980). This sequence appears to be overlain by the Pomona flow and underlain by the Frenchman Springs basalt and would be correlative to the Clifton formation. Warren and others (1945) and Wells and Peck (1961) included the sandstone at Clifton (now the Clifton formation) with an undefined sandstone exposed near Walluski River 16 km. west of the study area. More recently, Nelson (1978) correlated the sandstone near the Walluski River with the slightly older Pipeline member of the lower to middle Miocene Astoria Formation based on Saucesian foraminiferal assemblages.

The Clifton has no other known correlative units in western Oregon or Washington at this time. The Clifton formation may correlate in part with the upper part of the Ellensburg or Latah Formations of the western part of the Columbia Plateau which are also stratigraphically interbedded with the Columbia River Basalt Group (Swanson and others, 1979). The Ellensburg Formation, like the fluvial facies of the Clifton Formation, consists of cross-bedded volcanic, micaceous, arkosic sandstone (Schmincke, 1969). This correlation is tentative due to the distance between the two areas and the number of other basalt flow units which are present between the Frenchman Springs member and the Pomona member in the Plateau area, but not seen in the thesis area. The Clifton formation is not equivalent to the Vantage Member of the Ellensburg Formation, which has been correlated as far west as the Clackamas River area. The Vantage Member is stratigraphically bounded by the Grande Ronde

and Wanapum Basalts; therefore, the Vantage Member is older than the Clifton formation. A Vantage equivalent occurs along U.S. Highway 30 in this study (see Basalt Stratigraphy Section).

Paleocurrent Analysis of the Clifton Formation

Statistical analyses of 85 measurements of the trend of the axes of trough cross-beds in facies 1 of the Clifton formation from eight localities show an overall paleocurrent disperal direction to the west-southwest (mean 252°; S.D. $\pm$ 36°) (Fig. 44). Paleocurrent directions measured at three outcrops of the shallow-marine, burrowed bar subfacies 1a diverged from the general trend. The divergent directions suggest strong southward and north-westward components in the dispersal of this subunit which reflects longshore current reworking of fluvial sands at the mouth of a river or jet flow dispersion of sand as river water flowed into the shallow marine environment. The orientation of flame structures in sandstone beds of siltstone facies 2 and channelized facies 3 indicate that these facies were deposited on a paleoslope inclined toward the west. Although difficult to measure, the orientation of channel axes in facies 3 is generally to the west-southwest (Fig. 44). In summary, all the paleocurrent indicators show a general transport direction from the east to the west-southwest and a westward paleoslope for the clastic sediment of the Clifton Formation.

The method used to evaluate the measeurements of the trough cross-beds in facies 1 is based on Royse's (1970) statistical method for evaluation of paleocurrent data. It was used to construct the rose diagram and to calculate the mean and standard deviaiton of the cross-beddings readings from facies 1 (Fig. 44). No correction for bedding attitude was necessary due to the generally low regional dip (20°) of the Clifton

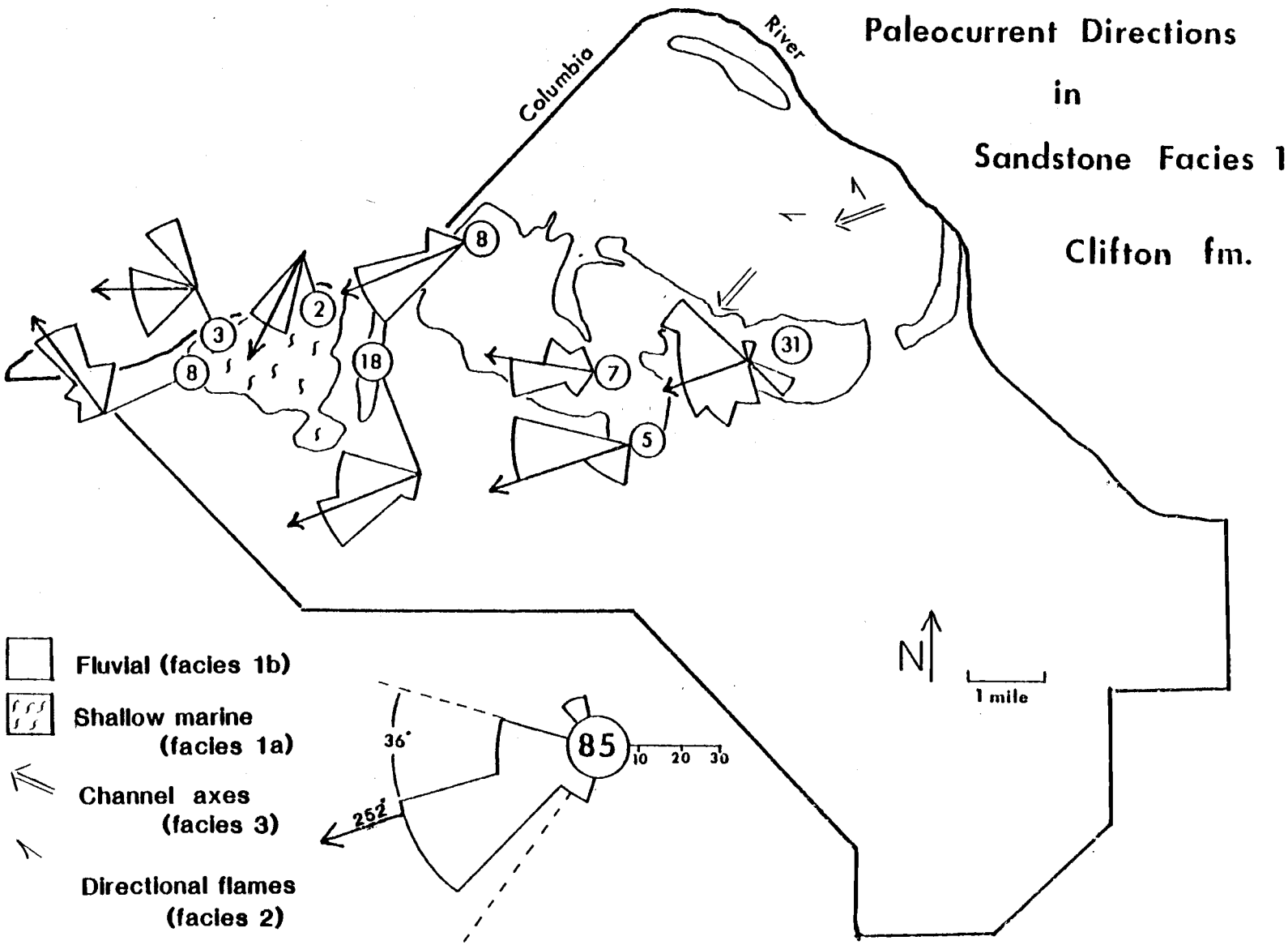

Paleocurrent Directions
in
Sandstone Facies 1
Clifton fm.
Columbia
River
8
2
3
8
18
7
31
5
N
1 mile
Fluvial (facies 1b)
Shallow marine
(facies 1a)
Channel axes
(facies 3)
Directional flames
(facies 2)
36°
252°
85
10
20
30

Figure 44

formation; Potter and Pettijohn (1977) demonstrated graphically that rotation of directional features can be neglected if structural dips are under 25 degrees. The rotational error incurred amounts to less than ± 3 degrees, which is within the degree of uncertainty of the field measurements (± 5 degrees) and well within the standard deviation about the grand mean (± 36 degrees; Fig. 44). According to Potter and Pettijohn (1977) flame structures that generally point in one direction may be the result of current drag along a density gradient boundary which may indicate the direction that sediment slumped down a paleoslope. Multidirectional flame structures can form as a result of dewatering during compaction or due to the density variation between two rapidly deposited sediment types (i.e., sand and soft mud) and not related to current drag or paleoslope (Dott, 1966). However, the unidirectional orientation measured at any one location in the field for the flame structures suggest that these structures indicate a paleoslope or current direction. The northwest reading on Figure 44 may be the result of channel wall slumpinng reorienting the features.

Flame structures in siltstone facies 2 consists of fine- to very fine-grained sandstone and siltstone which intrude slightly coarser grained graded turbidite sandstone. The bulk density contrast required between beds to produce flame structures may have resulted from a higher percentage of silt and clay size material in the underlying bed and/or greater volume of initial trapped pore water in the upper part of the underlying bed. Spontaneous liquefaction occurs as a response to rapid increase in the upward buoyancy forces and pore pressures due to rapid loading by the overlying sandstone (Dott, 1966). If this occurred before the drag forces of the turbidity current which was depositing the sand were completely

dissipated, a unidirectional orientation of the flame structures would result, according to Allen (1970).

Depositional Environment

In this section, the depositional environment and mechanisms of each facies of the Clifton formation are discussed. The facies include: the fluvial-marsh and shallow-marine (facies 1b and 1a), the turbidite-siltstone (facies 2), and the submarine cahnnel (facies 3). The interpretation is then combined into an overall depositional model using analogies to both modern sedimentary environments and the rock record.

Facies 1

Subfacies 1b

Large-scale trough cross-beds in coarse lenticular beds of arkosic sandstone, laminated carbonaceous mudstone, fragments of petrified wood, and minor coal seams suggest that this subfacies was deposited by a large, low-gradient, braided river with marshes. Paleocurrent directions and statistical analyses as well as provenance studies further support this conclusion.

Large-scale trough cross-bedding is volumentrically the dominant sedimentary structure in the fluvial sandstone subfacies (1b). Harms and Fahnestock (1965) suggested that this feature forms in migrating subaqueous dunes or sand waves in the upper part of the lower flow regime. They reported that sand waves are the most common structure in the channel bed of the Rio Grande River near El Paso, Texas. Whetten and Fuller (1967) determined that large-scale sand waves (0.9 to 3.4 m cover 86% of the Columbia River channel between Longview, Washington and

Astoria, Oregon. This represents the final 80 km of the river before it debauches into the Pacific Ocean. Downstream migration of these lingoid or lunate dunes is thought to be the dominant mechanism for producing large-scale trough cross-bedding preserved in river deposits according to Allen (1965). Harms and Fahnestock (1965) suggested that scouring of the river bed by vortices from the turbulent flow of the river current, not directly linked to dune migration, create the troughs. The troughs, which are then filled by the migrating dune, form to create the trough cross-bedding.

Visher (1965) first postulated an "ideal" accretionary point bar sequence in fluvial deposits and related it to the lateral migration of river meanders. This sequence consists of trough cross-bedded sandstone overlain by parallel laminated sandstone which is, in turn, overlain by microtrough, cross-bedded sandstone and a mudstone drape. He suggested that trough cross-beds may be the only part of an accretion unit preserved after erosion and reworking by subsequent migration of other channels over the older point bar. Abrupt planar truncation of some trough cross-bed sets by other cross-bed sets in the Clifton formation may be the result of channel reworking and may be a partially preserved point bar sequence. Most of the trough cross-bedding, however, probably formed within the channel proper by migration of subaqueous dunes or large sand waves over scours in the river bed.

The vertical sequence of sedimentary structures in the fluvial subfacies (OC-9) of the Clifton formation appears to be that of a point bar deposit (Fig. 34). These deposits are laid down primarily during flood stages which produce varying flow regimes and associated sedimentary structures in the channel bed across a meander. Once a meander bend is produced, it is propagated by continuing, but reduced differential helicoidal

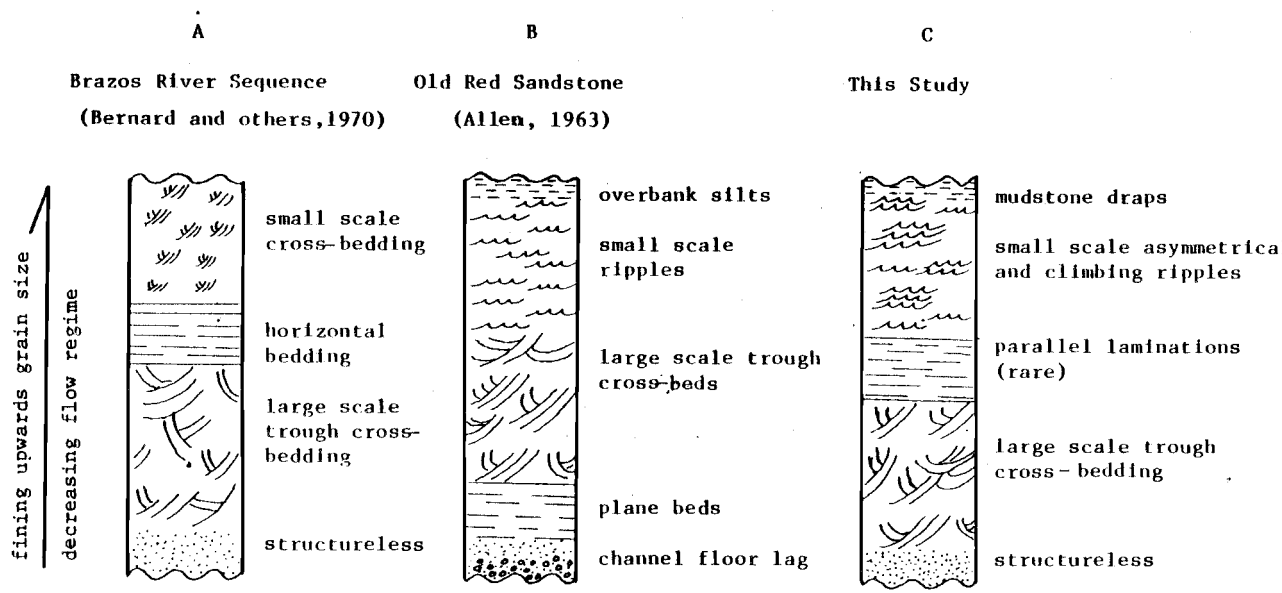


Figure 45 Comparison of modern and ancient point bar deposits with generalized vertical sequence from the fluvial facies (1b) of the Clifton formation.

flow. Helicoidal flow erodes the outside channel wall and deposits bed lado detritus on the inside of the bend, thereby forming a point bar sequence (Allen, 1977). Bernard and others (1970) study of the Brazos River in Texas describes a modern example of a point bar sequence (Fig. 45A). Allen (1963) reported a similar vertical point bar sequence which he interpreted as an abandoned meandering fluvial channel in the Old Red Sandstone of England (Fig. 45B). He explained development of point bar deposits in terms of a decreasing flow regime. The major differnces between the Brazos River deposit and the Old Red Sandstone sequence is the position of the parallel laminations relative to the trough cross-beds. In flume experiments, laminations form in the upper flow regime and in the lower flow regime between the current energies necessary to create sandwaves and trough cross-beds according to Harms and Fahnestock (1965). In a natural river setting, development of laminations requires a particular combination of depth and velocity which can be reproduced at different times and positions of a point bar within the flood stage of a river (Walker and Cant, 1979). In some cases, laminations may be entirely absent. Allen (1977) suggested that in keeping with the decreasing flow regime concept, ripples always occur above the large-scale trough sets.

The more complete point bar deposits in the Clifton formation consists of a 2to 3-meter thick vertical sequence of a basal, thin, structureless, coarse-grained sandstone overlain by a large-scale trough cross-bedded sandstone which is overlain by a thin layer of fine-grained sandstone with asymmetrical and/or climbing ripples (microcross-laminiations). The ripples are capped with a very thin mudstone drape (Fig. 45C). This sequence is indicative of a decreasing flow regime, from the upper to lower part of the lower flow regime to deposition from suspension during

relatively low flow periods. Parallel laminations are absent in most of the vertical sequences observed; only one sequence displayed parallel laminations between large-scale cross-beds and ripples similar to that pictured in the modern Brazos River by Bernard and others (1970).

Additional supportive evidence for a fluvial depositional model for this subfacies is the presence of large chunks of petrified wood up to 1.3 meters long and 12 cm in diameters (some with possible beetle boring) scattered throughout the sandstone. The lignite coal seams (1 m thick) and underclays, indicative of non-marine depositional (Pettijohn, 1975), may represent marshy floodplain deposits along the margins of the main channel. The well-laminated, micaceous siltstone may also be a floodplain or levee deposit.

The unidirectional paleocurrent direction inferred from trough cross-bedding implies a general flow to the southwest (Fig. 44). Divergent directions, especially in the point bar deposits (OC-9), are common in fluvial channel systems (Potter and Pettijohn, 1977).

Grain size statistics from sieve analyses plotted on the binary graphs by Friedman (1961) and Passega (1957) also support the conclusion that the Clifton sandstones were deposited by tractive current (e.g. river) (See Size Analysis Section). These sandstones are, for the most part, texturally submature (Folk, 1951), with a low percentage of silt and clay, again suggesting tractive deposition and some winnowing of the fines.

The relatively confined areal extent of the point bar sequences (OC-9) and the marsh deposits (OC-585 and 177) coupled with the widespread trough cross-bedding channel sand

bar or sand wave deposits suggest a wide channel depositional environment. At least 70% of this fluvial subfacies consists of large-scale trough cross-beds which compares favorable with Whetten and Fuller's (1967) estimation that 86% of the bed of the lower Columbia River is covered with large sand waves. The lack of extensive floodplain siltstone again may indicate a relatively wide channel and a narrow floodplain, unlike most meandering rivers (e.g. like the Willamette River in western Oregon). The close proximity of the laterally equivalent shallow-marine bar subfacies to the west, along with westward paleocurrent directions, the percentage of trough cross-bedding, and the presence of only a few laterally confined point bar sequences all suggest that this fluvial unit may have been deposited at a broad river mouth similar to the mouth of the present lower Columbia River. The lower Columbia River today is a drowned estuary that fills much of the valley and is characterized by a very broad channel bifurcated into smaller diverging and converging channels by sand bars, much like that of a braided stream.

Subfacies 1a

This subfacies was deposited on a lower shoreface or an offshore shallow marine bar west of the river mouth environment in which facies 1b was deposited. Distinguishing features of this shallow-marine subfacies include a fine-grained, dirty sandstone abundant trough cross-bedding, disrupted in places by 1-3 cm vertical *Rosselia* burrows (Fig. 36). Parallel laminations and abundant mica and carbonaceous debris in cross-laminations also occur. A few foraminifera tests, small high-spiraled gastropods, and pelecypods in calcareous concretions in the cross-bedded sandstone also suggest a shallow-marine origin.

Vertical Rosselia burrows are found in lower shoreface or offshore bar environmenmts according to C. K. Chamberlain (written communication, 1979). He mentioned that these burrow forms are similar to Asterosoma which Howard (1972) recognized in the Cretaceous Black Hawk Formation of Utah. Howard described these lower shoreface Cretaceous sandstones as being characterized by Asterosoma burrows, fine-grain sizes, and parallel laminations. These characteristics are similar to the laminated Clifton sandstones in facies 1a that are cut by Rosselia burrows. The small-scale trough cross-bedding associated with the burrows reflects tractive currents which may be related to dispersion of flood water from a river mouth, or alternatively reworking of river mouth sand (facies 1b) by longshore or tractive currents (Fig. 44). Small sand waves and offshore bars with trough cross-bedding, formed by tractive currents, are not uncommon on the continental shelf (Walker, 1979). Reinson (1979) suggests that most lower shoreface deposits in the rock record are storm deposits rather than normal day-to-day material delivered into the area. This is due to the large influx of sediment which occurs during a storm. For example, storm waves may generate small sand waves and shoreline bars on the shelf.

The size and high concentration of carbonized wood fragments in the trough cross-laminations also suggest proximity to a shoreline or river mouth and may have been the result of storm deposition. Although trough cross-bedding and sand waves can be generated by waves in the beach-surf zone, the shallow-marine cross-bedded Clifton sandstones are too poorly sorted and dirty to be beach-surf zone deposits. The Clifton sandstones also lack the typical vertical sequence of sedimentary structures and grain sizes of modern high-energy beach-surf zones described by H.E. Clifton and others (1971). In addition, trough cross-beds in modern beach sands are

inclined landward whereas Clifton sandstone trough cross-beds are oriented seaward.

Only minor study has been done on the type and dispersal of modern shallow-marine sediment near the mouth of the Columbia River. Mancus (1972) showed that the predominant sediment is sand and silty sand. Kulm and others (1975) are of the opinion that much of the sand on the inner shelf off Oregon is relict sand deposited by longshore currents, storm waves, and tractive currents at a lower sea level stand. They plotted grain size statistics of various sediments from the Oregon shelf on binary graphs. Grain size statistics of the shallow-marine silty sandstone of subfacies la plot within the mixed mud and sand field on the binary graphs of Kulm and others (1975) and within the tractive environment fields on the binary graphs of Friedman (1961) and Passega (1957) (See Analysis Section and Figs. 55 and 56).

The similariy in mineralogy (See section on mineralogy), similar stratigraphic position in the lower part of the Clifton formation (Fig. 42), and the interfingering relationship of the shallow-marine facies (la) and the fluvial facies (lb) suggest that these two facies are genetically related. This facies relationship of a river mouth, shallow-marine environment (facies lb and la) was present in the Nicolai Mountain-Gnat Creek area during initial deposition of the Clifton formation in the middle Miocene time.

Facies 2

The rhythmic, laterally persistent laminated siltstone and thin-bedded, fine-grained, graded sandstone of facies 2 of the Clifton formation are interpreted as interchannel or continental slope deposits. Associated with these lithologies

TABLE 8

Basic Classification of Turbidite and Other Resedimented Facies from Walker and Mutti (1973)

BOUMA SEQUENCE NOT APPLICABLE	FACIES A -- Coarse-grained sandstones and conglomerates A1 Disorganized conglomerates A2 Organized conglomerates A3 Disorganized pebbly sandstones A4 Organized pebbly sandstones. FACIES B -- Medium-fine to coarse sandstones B1 Massive sandstones with "dish" structure B2 Massive sandstones without "dish" structure.
BEDS CAN REASONABLY BE DESCRIBED USING THE BOUMA SEQUENCE	FACIES C -- Medium to fine sandstones -- classical proximal turbidites beginning with Bouma's division A. FACIES D -- Fine and very fine sandstones, siltstones -- classical distal turbidites beginning with Bouma's division B or C. C-D FACIES SPECTRUM -- can be described using the ABC index of Walker (1967). FACIES E -- Similar to D, but higher sand/shale ratios and thinner more irregular beds.
BOUMA SEQUENCE NOT APPLICABLE	FACIES F -- Chaotic deposits formed by downslope mass movements, e.g. slumps. FACIES G -- Pelagic and hemipelagic shales and marls -- deposits of very dilute suspensions.

are thicker (5 cm to 3 m) sandstone layers characterized by Bouma sequences, convolute bedding, flame structures, and clastic dikes which may represent overbank or levee deposits along submarine channels of facies 3. These lithologies best fit the slope depositional facies of Mutti and Ricci Lucchi's (1972) and Walker and Mutti's (1973) turbidite fan-canyon model based on their work in the northern Appennines of Italy.

The characteristic feature of facies 2 is the continuous, evenly bedded, well-laminated, alternating, diatom-bearing siltstones which comprise the middle part of the Clifton formation (Fig. 37). These beds, which account for approximately 75% of facies 2 are interpreted as hemipelagic sediments equivalent to facies D and G of the Mutti and Ricci Lucchi's (1972) model (Table 8). They are the result of deposition from a nepheloid layer suspension, from a rain of fine sediments from surface oceanic waters, and from deposition of fine sand by dilute turbidity currents (Walker and Mutti, 1973). These facies G and D are the most characteristic deposits of the slope or outer submarine fan and basin depositional environment (Rucci Lucchi, 1975; Bouma, 1979). The overall fine grain size, well-laminated character, abundance of carbonaceous plant matter, and lack of bioturbation or benthonic fauna in facies 2 are indicative of deposition in low turbulence and anoxic or low oxygen conditions (perhaps where the oxygen minimum zone impinges on the continental slope sediment/water interface). Small intraformational slumps and pemecontemporaneous folds of 1-3 meters in thickness in this facies are interpreated as upper slope sediment slides similar to those described by Cook (1979) in the Upper Cambrian carbonate slope deposits of Central Nevada.

The benthic marine diatom assemblages in this facies aid in

the reconstruction of the depositional environment. A maximum water depth of 50 to 200 meters, (middle to outer shelf) is indicated by these flora (H. Schrader, 1980, personal communication). The assemblages from the sample localities are either a pure marine flora or are mixed non-marine and marine flora. The mixed assemblages are dominated by a higher percentage of marine flora (80%) versus fresh water flora (20%). One sample (OC-602, Appendix IV) contained a high percentage of freshwater diatom flora (80%). However, two genera in OC-547 of the siltstone facies, Nitzschia c.f. curta? and Delphineis spp., are high productivity phytoplankton. Along with the abundance of brissles and spores present, these genera indicate a cold water coastal upwelling environment similar to that along the Oregon coast today (Schrader, 1980, personal communication). Actinocyclus ingens, the most common species in most samples, occurs only in marine environments. Most samples contain a higher proportion of terrigenous sediment than sample 547, suggesting slightly shallower marine conditions or a proximity to shoreline (OC 513 and 519).

The freshwater diatoms, in particular Melsoira islandica sp. and Phyroliths, which are the dominant types of non-marine flora present, are easily transported by eolian processes, and are commonly mixed with marine flora (Schader, 1980, personal communication). Three diatom species found in sample 602, Cymbella sp., Pinnalaria sp., and Synedra spp., are restricted to freshwater environments. Paralia salcata is a brackish water form also present in this sample. These non-marine genera dominate the assemblage in sample 602 from facies 2 with the marine flora poorly preserved. This is opposite of the more marine-rich flora present in the rest of the samples (Appendix IV).

A possible explanation for the mixture of freshwater and brackish water diatoms with normal marine flora is proximity of siltstone facies 2 to the fluvial deposits of facies 1. The freshwater flora could have been carried into the marine environment by an enlarged freshwater wedge caused by a flood discharge from the river mouth. A second explanation could be reworking of non-marine diatoms from the underlying fluvial sediment (facies 1b) by downcutting of the submarine channel system (i.e., facies 3). Hans Schrader (1980, personal communication) does not favor transport of diatoms over a long distance through a liquid medium. He also doubts that diatom tests could survive reworking. Furthermore, no freshwater diatoms were found in the fluvial strata of facies 1b. Thus, evidence is lacking to support the second hypothesis over the first. The first explanation, therefore, seems more plausible and also shows a mechanism to achieve the "mixed" flora characteristics of this slope facies.

Between the slope deposits of facies 2 and the channels of facies 3 are thicker (1-3 m), lenticular, graded to structureless beds of sandstone and thinner intervening beds of siltstone (ss/sh ratio = 70/30) which corresponds to facies E of Mutti and Ricci Lucchi turbidite faceis model (Table 8). This facies is interpreted by those authors as overbank or levee deposits. These strata in facies 2 are characterzied by Bouma sequences T_{abe}, T_{ae}, and rarely T_{ace}, clastic dikes, flame structures, load casts, and flaser-like bedding, which suggest rapid deposition and loading of sands by high density turbidity currents. The sandstone layers pinch out laterally or grade into thinner bedded finer grained sandstone/siltstone units which display normal density grading with concentrations of large but less dense flakes and carbonaceous material at the top of individual beds. These are similar to "distal" turbidites of outer fans and basins but are

considered interchannel deposits here. These strata are interpreted as facies D in the Mutti and Ricci Lucchi (1972) model and are commonly interbedded with hemipelagic slope deposits (facies G) in the Clifton formation (Table 8).

Submarine levee and interchannel sediments are deposited from overbank discharge of high and low density channelized turbidity flows (Ricci Lucchi, 1975). Sediment gravity flows too large to be contained in active channels would spill out and overflow onto the continental slope (Nelson and Nilsen, 1974) rapidly dumping the coarse fraction to build levee deposits (facies E in the Clifton Formation). Thus diluted, the turbidity current may continue to flow down slope away from the channel at an acute angle to the channel. Gradually losing the fine-grained tractive material, the flow would fill abandoned submarine channels and interchannel lows to form the interchannel deposits (facies D in the Clifton Formation). In the interim period between turbidity flows, the diatomaceous hemipelagic muds and silts would blanket the sediment deposited from the tails of dilute turbidity currents and nepheloid layers and from the continuous rain of fine clay sediment settling out of the water column.

Facies 3

Disorganized conglomerates, massive arenaceous sandstones, and breccia deposits of Mutti and Ricci Lucchi's (1972) and Walker and Mutti's (1973) turbidite model (facies A_1, A_2, B and F; Table 8) are recognized in the submarine channel facies 3 of the Clifton Formation. The occurrence and distribution of these facies as well as the overall channel geometry, internal sedimentary structures, sandstone/mudstone ratio, and the association with the adjacent levee and slope deposits of

siltstone facies 2 suggest that facies 3 is a filled upper part or bifurcating head of a submarine channel system.

The outcrop pattern shows a relatively narrow (less than one mile wide), elongate, shoestring distribution for facies 3 in the study area (Fig. 33 and Plate I). This channel facies cuts into, is laterally equivalent, and overlies siltstone facies 2 which as discussed above, is interpreted as a sequence of thin-bedded, overbank, and interchannel turbidites and hemipelagic continental slope deposits.

The spectacular exposures of sedimentary breccia composed of disorganized large blocks of laminated siltstone incorporated into a pebbly, coarse-grained sandstone matrix are similar to the chaotic facies F of the Mutti and Ricci Lucchi (1972) model (Fig. 38 and 39). These slump deposits, which occur sporadically throughout channel facies 3, are charaterized by allocthonous blocks of sediment which was originally deposited elsewhere in the basin. This remobilization was originally deposited elsewhere in the basin. This remobiliziation is a charcteristic feature of the chaotic deposits (facies F) of Mutti and Ricci Lucchi turbidite model. The siltstone in these blocks was originally deposited as interchannel or overbank levee deposits of facies 2 adjacent to a channel (e.g. channel walls). Undercutting of the channel walls by turbidity and other tractive currents created the blocks of laminated siltstone that form the chaotic breccias. In outcrop, these chaotic deposits grade from numerous blocks in clast-support to fewer blocks in matrix-support.

The more concentrated, clast-supported parts of the deposits are interpreted as rock fall and slide deposits; that is, they were transported by submarine avalanching of individual blocks or clasts of siltstone, en masse, down steep

slopes with little matrix support (Nardin and others, 1979). Individual blocks may have fallen into the sand that flooded the channel at different times. This differs from the debris flow mechanism which Nardin and others (1979) define as transport of the entire slump mass by matrix support downslope after the plastic limit has been reached by incorporation of sufficient water and sand matrix to produce a mass flow. Debris flows are thought to be the transport mechanism for the matrix-supported "chaotic" siltstone breccia deposits in some of the channel facies (Fig. 40).

The arenaceous-conglomerate facies A of the Mutti and Ricci Lucchi (1972) turbidite fan model is represented by disorganized pebble conglomerates and thick structureless sandstone in facies 3 of the Clifton Formation. These conglomerates form lenticular or irregular, non-graded layers within the sedimentary breccia channels and probably formed from high density grain flows on high density turbidite currents. Mutti and Ricci Lucchi (1972) pictured that the entire sediment load of high density grain flows is deposited from the upper flow regime all at once or with little or no sorting of the clasts. Hence, there is no grading. The main support mechanism is grain to grain collisions during movement down steep slopes that create an upward impact force to support the grains during transport. In Nardin and others (1979), classification of gravity flows, this type of flow is an inertial (high concentration) grain flow of cohesionless sediment (i.e., sand or gravel) bordering the liquid limit where high density turbidity currents can form.

Also associated with the chaotic sedimentary breccia facies are thick, structureless, medium- to coarse- grained arkosic sandstones with and without mudstone chips (0.4 m or less; Fig. 41). These structureless sandstones fit facies B and B_1 of

the Mutti and Ricci Lucchi model (Table 8). The sandstones appear to be a series of amalgamated beds deposited down channel from chaotic sediment slump breccia deposits. The transport mechanism is thought to have been inertial and viscous grain flows wich support the cohesionless sediment by upward dispersive pressure (Nardin and others, 1979). The relatively steep slope (18°) required to propagate this kind of flow (Middleton and Hampton, 1973) even with high concentrations of material, suggests that it is a local deposit within the marine channel. Alternatively, thick strucureless amalgamated sandstones may be a series of nongraded channelized deposits formed by repeated high density turbidity flows. Such thick, structureless sandstones, "fluxoturbidites" of the older literature, are thought to be common in upper submarine fan and canyon deposists (Mutti and Walker, 1973).

Four types of sediment gravity flow (turbidity, fluidized, grain or debris flow) occur in the subaqueous environment and reflect the laws of fluid mechanics. These four types are end members (Nardin and others, 1979). According to Middleton and Hampton (1973), a sediment gravity flow in nature probably is a hybrid of various types of flows. For example, the forces which support the particles in a gravity flow, grain flow, turbidity flow, or fluidized flow may operate for only a brief time and, thereby, create a variety of sedimentary structures. The sedimentary features of a deposit that results from a sediment gravity flow are related to a number of parameters such as: the amount and type of sediment support, the avenue of transport, the slope, the triggering event, and the amount of interstitial fluid present between grains. Therefore, a coarse-grained, deep marine deposit can be a result of a mixture of these gravity flow transport processes. The characteristics of the deposit may reflect one dominant gravity flow or may reflect a gradation between processes. The

relative abundance and widespread nature of turbidite deposits in submarine fan facies attests to the uniformity in parameters listed above. Uniformity of conditions results in a flow that maintains its integrity over a long distance such as a turbidity current. In contrast, in a submarine canyon facies many parameters that control the type of gravity flow change radically over short distances, resulting in a wide spectrum of of sediment gravity flow processes. These processes are reflected in the variety of turbidite, grain flow or chaotic slump deposits in the levee and channel deposits of the Clifton Formation. A submarine fan also is not favored as the depositional environment of facies 2 and facies 3 of the Clifton because of the close proximity of the fluvial and shallow-marine facies, the shallow water depths indicated by the diatoms (50 to 200 m), and the thinness of the unit (200 m). Most fans are thousands of feet thick. In addition, the Clifton does not show the upper, middle, and outer facies typical of most fans (Mutti and Walker, 1973).

Depositional Model

The Clifton formation represents a continental margin deposit in which the transitional shallow-marine environments between deep marine and non-marine are virtually absent. For example, in the eastern part of the study area, the fluvial-marsh facies (1b) passes directly into the overlying siltstone shelf/slope facies 2 and submarine channel deposits of facies 3 and back into fluvial deposits (facies 1b) at the top of the section. There is little or no intervening transitional beach or shallow shelf deposits present. This "bypassing" of the transitional shallow marine-beach environment probably is a function of the erosional channelzied nature of both the deeper marine and fluvial facies. A head of a submarine canyon on a narrow continental shelf directly off the river mouth would funnel fluvial detritus directly into deeper water, thereby by passing the shelf/beach environments.

Schematic Model for Deposition of the Clifton formation

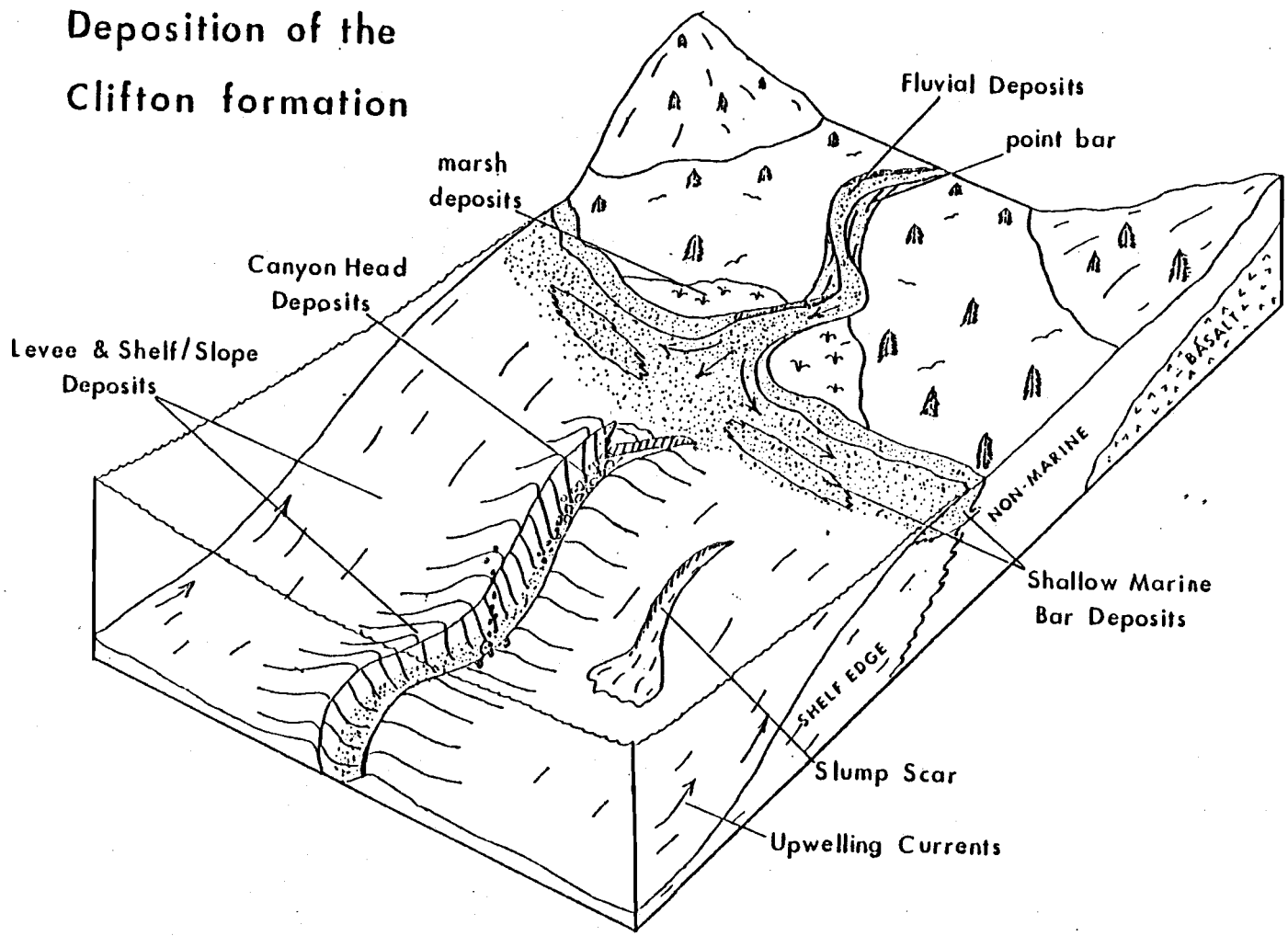


Figure 46

Local rapid sea level fluctuations also could have played a part in reworking and removing shelf/beach sediments that were deposited. Some shallow marine bioturbated sandstone (facies 1a) was recognized at the base of the Clifton formation in the western part of the study area.

The cross-bedded, carbonaceous sandstone (facies 1b) in the lower and upper part of the unit and associated coal beds represent fluvial sand waves, point bar, and marsh deposits in or near a river mouth with dispersal patterns from east to west. The Rosselia bioturbated cross-bedded sandstone in the western part of the outcrop area (facies 1a) interfingers with the lower fluvial subfacies (1b) probably represents shallow-marine shelf or offshore bars produced as the river mouth sands were dispersed by longshore and tidal currents acting along the coast line.

Further offshore well-bedded, shelf/slope, hemipelogic siltstones and very fine-grained sandstones were rapidly deposited and locally slumped to create penecontemporaneous folds. Off the river mouth, cutting into the shelf/slope deposits was a series of submarine channels or bifurcating canyon heads. Associated with the channels were overbank levee turbidite deposits and rock falls or slides of siltstone breccias slumped from levee channel walls. Also acting in the channels were debris flows and local grain flows that formed thick structureless sandstones. The coarse sand was funneled directly from the river mouth (Fig. 46). Reworking of slumps by grain flows and high density turbidity currents formed aligned or imbricated mudstone clasts within the channels and turbid overbank flows constructed the levees present.

Lithologically similar deposits of submarine canyon head and seagullies cutting into non-marine or shallow marine tidal

flat facies have been recently described by Lohmar and others (1979) and by Howell and Link (1979) in the Eocene of the San Diego embayment area; in the Oligocene Ballemy Group of New Zealand by Carter and Linquist (1975, 1977); and in the Elkton Siltstone of the Tyee Formation in southwest Oregon by Dott and Bird (1979). Each example suggests that a very shallow marine to deep marine slope deposition.

The Eocene units of the San Diego Embayment area are analogous to the Clifton formation in that the submarine canyon head facies cut back into a tidal flat facies (Delmar Formation) and a shallow marine cross-bedded sandstone (Tonney Sandstone). This is similar to the channelized facies 3 cutting into the underlying fluvial facies of the Clifton although in the Eocene example, the canyon head had eroded in a landward horizontal sense (Lohmar and others, 1979); whereas the Clifton also cut into the underlying fluvial facies. The Poway conglomerate of this southern California Eocene sequence can be traced laterally from the non-marine alluvial fan deposits cutting through a coastal plain tidal flat unit, and into a thick, submarine channel-subsea fan conglomerate. It also cuts the adjacent deep-water, thin-bedded, well laminated mudstone and turbidite sandstones. This is again analogous to the Clifton formation in that the cross-bedded arkosic sandstone of the fluival facies 1b is compositionally and texturally similar to the thick, structureless, submarine channel sandstone of facies 2. It was also probably the source of the coarse sands of the submarine channel (Fig. 46). The Clifton formation displays a vertical relationship with the fluvial sandstones both stratigraphically above and below the submarine channel sandstone (Fig. 42). Whereas, in the Eocene of the San Diego area the non-marine and marine units can be traced laterally as well as vertically (Howard and Link, 1979).

An alternative model for the Clifton formation is one in which the submarine channels represent small sea gullies cut into delta slope mudstone off an active subaerial delta system with coaly marsh deposits. Such a depositional model is pictured by Dott and Bird (1979) for the Eocene Elkton Siltstone (sea gullies facies) and the associated Coaledo Formation (deltaic facies). This model is not favored for the Clifton because, unlike the Coaledo, the Clifton lacks the many coal beds and other typical coarsening upward facies of deltaic sequence.

Modern analogies can also be made for the model depicted in the Clifton Formation. Many submarine canyons cut narrow continental shelves (less than 20 km) and are located off large river valleys which funnel sand directly down a canyon, bypassing shelf areas. Shepard and Dill (1966) suggested that the heads of approximately 50% of the canyons which they surveyed were within or immediately seaward of a river mouth source. An extreme example is the Congo Submarine Canyon which extends alomst 36 km into the mouth of the Congo River (Heezen and others, 1976). A geographially closer analogy is the present Columbia River mouth-upper part of the Astoria Canyon system (Kulm and others, 1975). For this to be a viable model of the Clifton Formation the river must have been closer to the canyon head in middle Miocene time than the 45 km that separates the two topographic features today.

Another example which describes the relationship between the fluvial and canyon head deposits of the Clifton formation is the Rio Balsas Submarine Canyon system approximately 10 km off the Rio Balsas River, along the west coast of Central Mexico. Cutting into the very narrow shelf near the mouth of the Rio Balsas River, the largest in Mexico, is a series of canyon heads which merge down slope to form a large submarine

canyon that terminates in the Middle American Trench. Reimnitz and Gatierrez-Estrada (1970) have demonstrated that with the shifting position of the main distributary river channel either activation of a pre-existing, partially filled, canyon head or erosion of a new one occurs near the new river mouth. This also results in deactiviation and filling of the abandoned canyon near the abandoned river mouth. Rapid erosion on the sacle of 20-30 meters/100 years was documented by C_{14} dates on wood fragments collected from the walls of the new canyon head.

This final analogy is not exact, because the river system during Clifton time did not appear to have an associated delta complex at the mouth, but rather had various incised fluvial channels which may have shifted position from time to time. The Rio Balsas example demonstrates the effect a river system has on the creation, abandonment, and filling of canyon heads and suggests a mechanism for producing the different submarine channel deposits described in the Clifton Formation.

Finally, the presence of quartzite in the channelized conglomerate and arkosic sandstone (see Provenance Section), along with the plateau-derived subaerial Columbia River Basalt above and below the Clifton formation as well as the east to west paleocurrent dispersal pattern suggest that there was an ancestral Columbia River and canyon head system in this geographic vicinity during the middle Miocene.

Quaternary Deposits

For mapping purposes the Quarternary deposits are divided into four groups: Holocene deposits (Qal), recent stream gravels (Qal_2), older tidal flat and estuarine deposits (Qal_3), and colluvium deposits (QTls).

Holocene deposits include the tidal flats, confined mainly to the Columbia River estuary near Westport and Brownmead and active stream deposits. The tidal flats material is composed of laminated medium- to fine-grained sand and silt. These tidal deposits are differentiated from the older tidal flat deposits in that they are normally covered at high tide and floods. During low tide or slack water, they are exposed as bars and tidal creeks. Active stream deposits consists primarily of basaltic boulders to sand-size detritus being transported in Big Creek and Gnat Creek. West Creek and Plympton Creek also contain a component of mudstone clasts originating from the underlying Oswald West mudstone. These thin stream deposits were not outlined on the geologic map (Plate I) because they would cover more important stratigraphic relationships of the bedrock.

The terrace gravels (Qal_2) are composed of rounded to subangular, poorly sorted boulders and pebbles, simliar in composition and texture to the present stream deposits. Associated with these are minor lenses of quartzo-feldspathic sand and silt. The deposits form flat terraces 9 to 10 meters above the entrenched drainage of Big Creek and Gnat Creek. Widespread stream dissected terraces occur in the Knappa Junction-Swensen area (Plate I). They are up to 25 meters thick at the mouth of Big Creek Gorge. The lens-shaped beds (1-3 m thick) are defined by variation in clast sizes. Large-scale planar cross-bedding is common in these deposits (gravel quarry near the Tillusqua fish hatchery, OC 122, Sec. 29, T8N, R7W). Two terrace levels along Big Creek possibly indicate continued uplift and local temporary base level developent and reentrenchment of streams (Beaulieu, 1973).

The older tidal flat and estuarine deposits (Qal_3) are composed of structureless, medium light gray (N6) sand and silt. This unit is found in the low broad, grassy flat areas between bedrock and Holocene deposits along the Columbia River and at the mouth of rivers that drain into the Columbia. Associated with these deposits are areas of organic rich soils and peat development (Secs. 33 and 34, T9N, R7W) (Beaulieu, 1973). These deposits are at least several feet thick.

Thick accumulation of basaltic colluvium (QTls) occurs along the base of Nicolai Ridge (Plate I). These deposits are also found in the Lost Lake area and in Big Creek Gorge. No attempt is made to outline all areas in which colluvium is present because it would obscure other more important bedrock relationships. Only where the colluvium is greater than 10 meters thick was it mapped.

The colluvium at the base of Nicolai Ridge is composed of rockfall and rockslide debris of Grande Ronde and Frenchman Springs Basalts mixed with arkosic sandstone. The clasts of basalt range in size from pebbles to blocks up to 20 m in diameter. Characteristically, the reddish iron-stained exposures are composed of angular boulders which display a variety of volcanic textures. The chaotic masses of rubble are unconsolidated, very poorly sorted, and lack stratigication (Fig. 48). The sand is unconsolidated and found structureless piles and disrupted layers which originated from the sandstone interbeds between the basalt flows on Nicolai Ridge (i.e., Vantage Member).

The deposit along Nicolai Ridge, is wedge shaped, thinning away from the ridge. It extends at least as far as the Columbia River, near Wauna (Plate I). Large blocks of basalt (up to 5 meters in diameter) from dredging operations of the

Figure 47 Landslide debris composed of basalt colluvium. Note poor sorting and angularity of the blocks. Hammer in center for scale. East of Nicolai Ridge (OC 255, SW, Sec. 35, T8N, R6W).

shipping lanes in the Columbia River near Wauna suggest that these landslide deposits may extend under the river. The colluvium produces very hummocky topography. In some cases it caps strike ridges of the older Astoria and Oswald West units, making them less susceptible to erosion.

The colluvium deposits are the result of undercutting of the older Tertiary sandstone and mudstones by the Columbia River with subsequent slumping and rockfalls of the overlying subaerial Columbia River Basalt (Beaulieu, 1973). Inland from the river, near the top of Nicolai Mountain, colluvium covers an area between Nicolai and Porter Ridge (Secs. 9 and 16, T7N, R6W). This material is postulated to be the result of slumping after normal faulting (see Structure Section). Extrapolation of this relatinship northward suggests that all the colluvium initiated along a fault trace trending northeast. Susequent undercutting by the Columbia River cut back the scarp to its present position.

The age of the colluvium deposit is uncertain. It must be younger than the Pomona member, dated at 12 m.y. (McKee and others, 1977). If related to the faulting, it may be pencontemporaneous with formation of the Nicolai questa or began soon after it developed and maybe as old as late Miocene or Pliocene in age. Most likely it is Quarternary in age since it is reltively fresh in appearance.

Sedimentary Rock Petrography and Provenance

A total of 25 thin sections and 8 heavy mineral grain mounts were examined from the different sedimentary units in the study area. Modal (point count) analysis (Tables 9 and 10) was performed on 11 of the thin sections to aid in classification. Approximately 600 counts were done on each

sample. Six pebble counts from different units using both thin sections and the binocular microscope were done on disaggregated samples from the conglomeratic layers in the units (Appendix VII). The pebbles were broken or slabbed.

In this section, the petrography of each sedimentary unit is dicussed separately with the exception of the members in the Astoria Formation which are considered together. Another exception to this format is that the petrography of the sandstone and mudstone interbeds between the middle Miocene basalt flows is discussed in the section "Sedimentary Interbeds". Since the main thrust of this study relates to the previously undefined Clifton Formation, the petrology of that unit is dicussed first. The provenance of each sedimentary unit is discussed in at the end of each section. The sandstone classification used in this investigation is based on Williams and others (1954) (Fig. 48). Finer grained units are classified as mudstone (greater than 75% silt and clay) using Dott's (1964) classification.

The very friable nature of most Tertiary sandstones dictated the use of fiber glass resin (laminac) and epoxy 154 to impregnate the samples. All but three of the thin sections were essentially grain mounts, which once impregnated, could be made into thin sections (3 mu thick) for easier identification of the minerals. This technique limits the usefulness of the thin sections for visual determination of porosity in the sandstones. Therefore, the most indurated sandstone samples from the study area were sent to Tenneco Oil Company of Bakersfield, California for porosity and permeability tests (see Economic Geology Section).

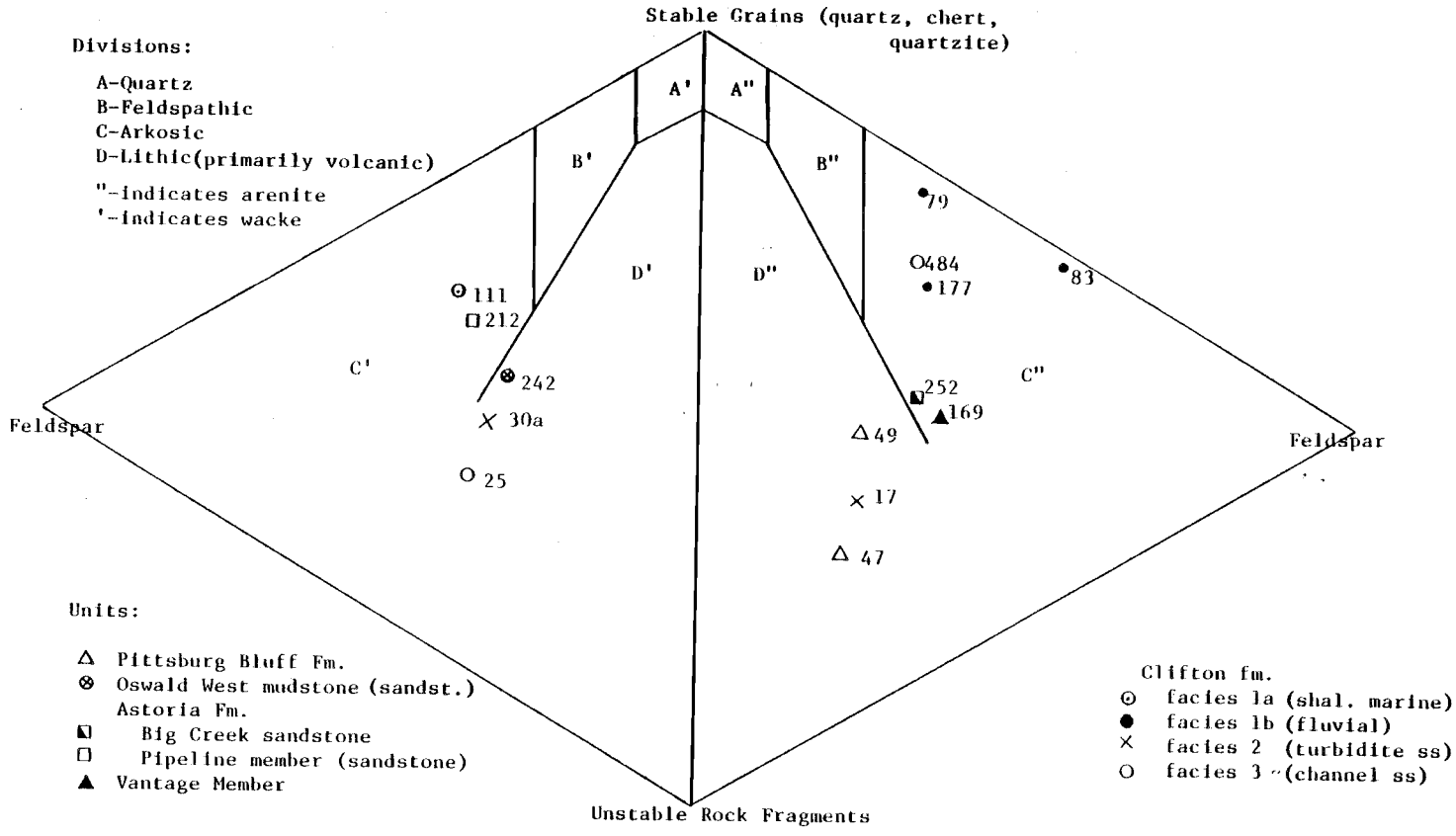


Figure 48 Classification of sandstones from the Nicolai Mountain-Gnat Creek Area. (modified from Williams and others, 1954)

Clifton Formation

Fifteen whole rock thin sections of impregnated sandstone and four heavy mineral grain mounts were studied and modal analysis performed on six representative sandstone samples from the three facies of the Clifton formation (Table 9). The thin sections show that the sandstones are primarily arkosic arenites (facies 1b and 3) and wackes (facies 1a) along with volcanic arenites and wackes (facies 2 and 3) (Fig. 48). Grain size analyses reveal that the mudstones of facies 2 are siltstone, clayey siltstone, and sandy siltstone (see Grain Size Analysis section). The mineralogic composition of the siltstones is in general similar to that of the arkosic sandstones in the Clifton where grain size permits identification of the minerals.

The sandstones of the Clifton formation are compositionally immature and texturally submature to immature, based on Folk's (1951 and 1974) classification. This probably reflects fairly rapid erosion from highlands, rapid transport, and rapid burial with only some winnowing of fines. Sorting is moderate to poor and framework grains are subangular (quartz and feldspar) to subrounded (volcanic rock fragments).

The sandstone of the fluvial facies (1b) consists of clean arkoses and arkosic arenite whereas the sandstone in the lateral correlative shallow marine subfacies (1a), due to the greater percentage of silt and clay sizes, is an arkosic wacke. The wackes in this facies are a result of mixing of grain sizes by bioturbation. The turbidite sandstones of siltstone facies 2 are volcanic arenites and wacke. The overbank turbidite deposits, closer to the submarine channels of facies 3, are the volcanic arenites. They grade into the volcanic wackes away from the channel axis. The amount of

volcanic detritus decreases upsection. The submarine channel facies (3) contains volcanic and quartzite conglomerates which have a volcanic wacke matrix. These conglomerates also grade upward in the section to arkosic arenites (Sample 484). Both the volcanic wacke and arkosic arenite rock types contain abundant mudstone chips from facies 2 (see lithology section).

The framework constituents of the Clifton formation sandstone are: quartz (5-19%), plagioclase An_{11-37} (12-23%), potassium feldspar (5-23%), rock fragments (10-37%) principally volcanic, micas (0.5-7%), iron oxides (1-4%) and non-opaque heavy minerals (1%). Some carbonaceous material (trace-7%), volcanic glass (trace-7%), glauconite, and fossil shell fragments also constitute portions of the framework material. An average composition of the Clifton formation is not meaningful due to the variety of depositional environments, grain sizes, and apparent dual sources of the framework constituents.

Quartz occurs as angular to subangular, coarse to fine sand-sized grains. Both monocrystalline strained and unstrained quartz is present, ranging from 9.5 to 35 percent. Unstrained quartz is uncommon in the coarser grained sandstones but increases in abundance with decreasing grain size. Polycrystalline quartz, both polygonized and sutured, also shows a fairly wide range of abundance (5.2 to 21.4%) in the unit. The percentage of polycrystalline quartz grains decreases with smaller grain size. Folk (1974) suggested that both strained and polycrystalline quartz, in particular polygonized forms, follow this general trend because of the increased tendency for polycrystalline grains to disaggregate more rapidly than monocrystalline grains. Inclusions in the quartz grains included zircon, apatite, and rutile (?). Dirty light brown, subrounded chert grains are present in minor

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Table 9

Modal Analyses of Sandstone Samples in the Clifton Formation

Mineral Species	facies 1a	facies 1b	facies 2		facies 3	
Sample No	111 **	177	17	30A	484	25*
Quartz						
Mono. qtz.	11.7	17.4	9.5	13.0	35.8	19.0
Poly. qtz.	10.6	21.3	8.5	5.2	15.2	16.0
Chert	.9	1.7	1.2	2.5	1.7	-
Quartzite	-	Tr.	Tr.	-	1.7	-
Feldspar						
Plagioclase	12.2	11.7	22.7	21.5	11.8	16.0
K- Feldspar	14.0	22.0	7.2	4.8	12.7	5.0
Rock Fragments						
Volcanic	8.1	13.2	28.0	22.8	8.4	35.0
Granitic	Tr.	3.8	0.8	-	0.7	Tr.
Metamorphic	0.9	2.3	1.2	0.9	2.0	1.0
Sedimentary	Tr.	-	-	-	2.9	2.0
Mica	7.4	0.5	2.5	8.7	5.4	2.1
Mafics	3.1	3.9	1.4	0.9	Tr.	1.7
Fe -oxides	2.3	1.9	2.5	0.7	1.7	.7
Carbonaceous fragments	7.2	-	-	Tr.	-	-
Fossil fragments	.9	-	Tr.	-	-	-
Alteration Minerals	-	-	2.5	1.5	Tr.	1.5
Volcanic Glass	-	-	3.5	6.7	-	Tr.
Glauconite	Tr.	-	Tr.	-	-	Tr.
Matrix	20.1	-		11.9	-	-
Cement	-	-	-	-	-	-

* Matrix of channel conglomerate

Tr. = $< 0.5\%$

** Calcareous concretion, matrix is composed of fines and calcite cement

amounts. Silicic volcanic chert composed of microcrystalline grains with rare feldspar (?) is common in some samples. These clasts were counted in the volcanic rock fragment category (Table 9). Quartzite fragments, which appear in trace amounts in the Clifton Formation, consist of polycrystalline quartz grains with highly sutured intercrystalline boundaries and strained extinction. They are normally better rounded than the sutured polycrystalline quartz and contain aligend mircocrystalline mica (sericite) flakes which define a foliation within the grains. Also present are rare subrounded chalcedony grains.

Table 9 illustrates the differences in amount and type of quartz in the facies of the Clifton Formation. The fluvial subfacies 1b contains a higher percentage of monocrystalline (17 versus 12%) and polycrystalline (21 versus 11%) quartz grains than the finer grained, shallow marine subfacies 1a. The interchannel turbidite sandstones of facies 2 have the lowest percentage of both monocrystalline and polycrystalline quarz grains of any of the facies. This may be a function of the fine grain size, but the high percentage of unstable volcanic rock fragments suggests that comminution is not the major reason for this discrepancy. It may be the result of a fluctuation in the source of volcanic and granite material.

The submarine channel facies can be divided into two compositional groups. A highly quartzone sandstone (51%; primarily strained monocrystalline quartz 25.8%) is similar to the sandstone of the fluvial subfacies (1b). The quartz-rich submarine channel sandstone occurs stratigraphically higher in the Clifton formation than the volcanic-rich sandstone and appears to have been derived from a similar source as the fluvial subfacies (1b). The quartz-rich sandstone is typical of most of the sandstones in the channelized facies. The

second compositional group is a fine pebble conglomerate from low in the Clifton section (Fig. 49). This conglomerate displays charcteristics similar to the facies 2 volcanic sandstone interbed; that is, there is less quartz (35% due to the increased amount of volcanic detritus.

Plagioclase is the dominant feldspar in the sandstones of the Clifton Formation. It comprises from 11.2 to 22.7% of the sandstone. The composition, as determined by the Michel-Levy method (Heinrich, 1965), ranges from An_{11} (oligoclase) to An_{46} (andesine) with an average of An_{35} (andesine). The grains are angular to subrounded, fresh to some dusted with sericite and clay minerals. They range in size from coarse- to very fine-grained sand. Normally zoned, untwinned and twinned andesine occur in most units and increase in abundance as the percentage of volcanic clasts increases. The grains commonly display albite twins, Carlsbad twins, and rare pericline twins. Myrmekitic intergrowths of quartz and oligoclase are also present in some samples (sample 25).

Potassium feldspar comprises from 5 to 21% of the sandstones. Orthoclase appears as the dominant potash feldspar with much lesser amounts of microcline (1-5%) and perthite (1%). Micrographic intergrowths of K-feldspar with quartz form some grains. Alteration of the potassium feldspar to clay, possibly kaolinite, and sericite is seen in various stages of development. In some cases, untwinned plagioclase may have been mistaken for orthoclase in thin section. Rough estimates of the relative percentage of plagioclase versus potassium feldspar were visually estimated with the aid of staining methods.

Variation of feldspar content within the Clifton formation reflects the increase in volcanic detritus. The sandstones in

the fluvial and shallow marine facies contain a higher percentage of potassium feldspar (22 and 14%) compared to plagioclase (12 and 11%), which may indicate a more "granitic" source. The 4:1 ratio of plagioclase to potassium feldspar in the facies 2 interchannel and overbank turbidite sandstones reflect an increase in volcanic material. The abundance of oscillatory zoned andesine, common for volcanic plagioclase, also increases from fluvial facies 1 to turbidite sandstone facies 2.

The feldspar content of the sandstones of channel facies 3 likewise reflect an influx of volcanic material. The almost 1:1 ratio of plagioclase to potassium feldspar in the sandstone from stratigraphically higher in facies 3 correlates well with the fluvial facies. Sandstones in the stratigraphically lower part of the channel facies have a ratio of plagioclase to K-feldspar (3:1) similar to the turbidite sandstones in facies 2, suggesting a genetic link between them.

Unstable rock fragments include principally volcanic grains (20%) with subordinate granitic (1%) types. Volcanic rock fragments range from 8 to 35% of the sample and consist principally of porphyritic andesites. Phenocrysts in the clasts include hornblende, clinopyroxene, and plagioclase (andesine). The groundmass of the andesite clasts shows a hyalopilitic to intersertal texture that is composed of brown glass, magnetite, andesine microlaths and pyroxene. The hornblende phenocrysts commonly have well developed magnetite reaction rims. The andesine phenocrysts usually display normal zoning (more calcic centers). Included in the volcanic rock fragments category are silicic volcanic chert, rare dacite fragments, and altered intermediate to basic volcanic fragments. The alteration products include glauconite as well as greenish yellow clays, possibly chlorophaeite and nontronite.

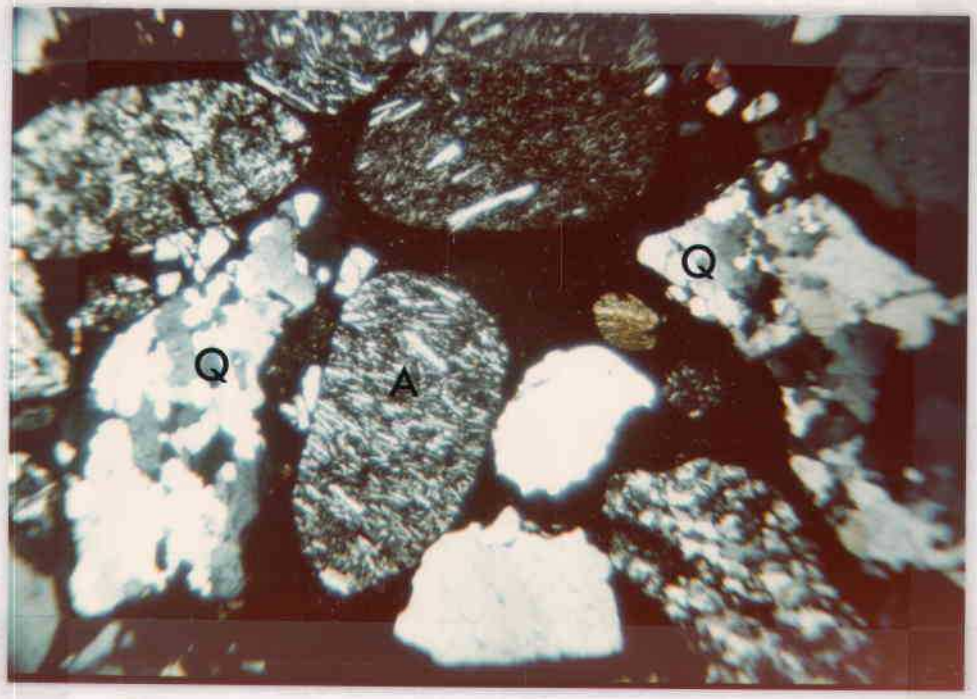


Figure 49 Photomicrograph of sandstone in the Clifton formation (facies 3). Note sutured polycrystalline quartz grains and pilotaxitic andesite(?) rock fragment. Thin section, crossed nicols, field of view is 4mm across (OC 25, NE, Sec. 8, T8N, R6W).

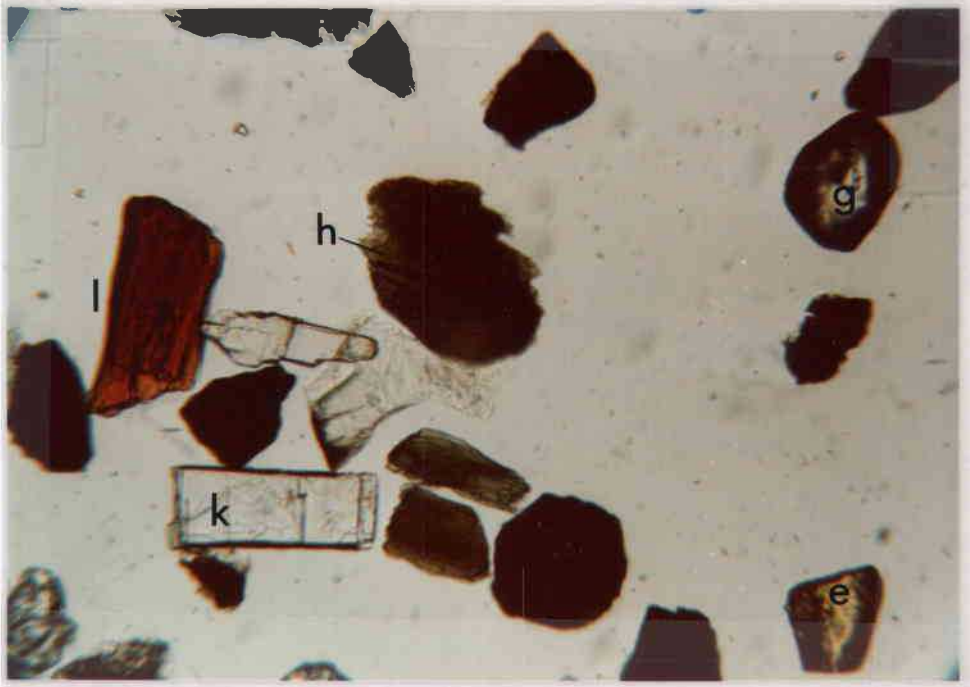


Figure 50 Photomicrograph of heavy mineral suite from the Clifton formation. Minerals include: Green hornblende (center), reddish brown lamprobolite, clear kyanite, pistachio green epidote (lower right), and clear, with high relief, garnet (upper right). Grain mount, plane light, field of view is 1.3mm across.

No Columbia River Basalt fragments were identified in the the thin sections but Columbia River (?) basaltic pebbles have been identified from the coarse-grained channel conglomerates (see Appendix VII for composition of pebble count). Brown and clear unaltered sickle-shaped bubble wall glass shards, and rarer pumic fragments comprise up to 5% of the sandstones in facies 2 and 3.

The distribution of volcanic rock fragments within the Clifton formation verifies the trends indicated by the feldspar ratios. As Table 9 illustrates, the greatest percentage of volcanic fragments, including pumice and glass shards, occurs in the turbidite sandstones of facies 2 and in the assoicated channel deposit of facies 3 (sample 25). This indicates that a pulse of volcanic detritus was introduced into the depositional environment of primarily arkosic sandstone near the beginning of facies 2 and 3. The upper part of channel facies 3 (sample 484), petrologically associated with the fluvial facies, represents a return to the original arkosic source.

Granitic rock fragments consist of grains composed of interlocking crystals of quartz, potassium feldspar, and/or plagioclase. Rarely hornblende crystals and biotite flakes are assoicated with the above minerals. This assemblage, as well as the myrmekitic texture, suggest granite which has assimliated mafic material or granodiorite (Williams and others, 1954). Pelitic phyllites and quartz mica schist fragments composed of strained quartz with crenulated boundaries and varying percentages of aligned mica flakes and carbonaceous material form the metamorphic rock fragments in the Clifton Formation. The granitic and metamorphic constituents are concentrated in the fluvial facies (1b) and associated submarine channel facies (facies 3).

Sedimentary rock fragments are principally confined to the channel sandstone of facies 3. Blocks of siltstone from facies 2 were incorporated into the sandstone by massive gravity flows and channel slumps as discussed in the Lithology and Depositional Environment sections. Microscopic siltstone ripups were also observed in the sandstones of facies 2 and 3.

Micas are minor constituents of the sandstones of the Clifton Formation. They range from 0.5 to 1.0% in the average sandstone (Table 9). Clear large muscovite flakes dominate this category; both green and brown biotite flakes also are present. Green chlorite anc clear penninite, with its distinctive "berlin blue" interference color, as well as sericite are other micas in the sandstones.

The greatest percentage of micas is concentrated in the finer grained sandstones such as in the shallow marine subfacies (1a) and the well laminated siltstone/very fine-grained sandstone overbank and slope deposits of facies 2. The shallow marine sandstones have large mica flakes (up to 10mm in diameter) which show no preferred orientation due to the bioturbated character of the unit. Micas are concentrated along the tops of individual laminae in the overbank turbidite sandstones of facies 2. This concentration is interpretated to represent density grading and is ubiquitous throughout this facies. Sample 79 (Ziak's Ranch, Sec. 34, T8N, R7W) from the fluvial facies contains up to 10% mica flakes, which is again the result of concentration by specific gravity (Rubey, 1933).

Opaque minerals include magnetite, ilmenite, leucoxene, hematite, and pyrite. The magnetite occurs as individual detrital grains and as rims around hornblende and pyroxene grains. Ilmenite grains are commonly altered to leucoxene.

magnetitie alters to hematite. Diagenetic yellow and red iron oxides, in particular hematite and limnite, coat the quartz and feldspar grains and act, in part, as cement for these poorly consolidated sandstones. This is common in the more porous sandstones which are also characterized by liesegang rings and a yellow-red outcrop color. Rare authigenic pyrite crystals are present in calcareous concretions in the shallow marine sandstone.

Nonopaque heavy minerals (>2.94 sp.gr.) comprise up to 4% of the sandstones of the Clifton formation by volume. Appendix VI lists four representative heavy mineral assemblages and their relative abundances in the Clifton. Sandstones in the fluvial subfacies (1b) show the greatest diversity of heavy minerals. Blue green hornblends, lamprobolite (brown hornblende), augite, and enstatite are the most common ferro magnesian heavy mineral constituents. The pyroxenes display angular saw tooth terminations suggesting that they have not been transported a great distance and are first cycle in origin. Pistachio green epidote is the most common heavy mineral species in all the sandstones. Zircon grains are abundant in the submarine channel facies (3) and in the fluvial facies (1b). Kyanite, staurolite, garnet, and rare sillimanite are present in all samples, suggesting some contribution from an intermediate to high grade metamophic source terrain (Fig. 30).

Clay matrix in the sandstone of the Clifton formation is negligible; since the grain size analysis has shown that most of the sandstone, with the exception of the wackes, have under 10% silt and clay. The friable nature of the sandstones prevented detailed analysis of the relationship of the clay matrix and the framework grains. The small amount of matrix present probably originated, in part, from diagenetic

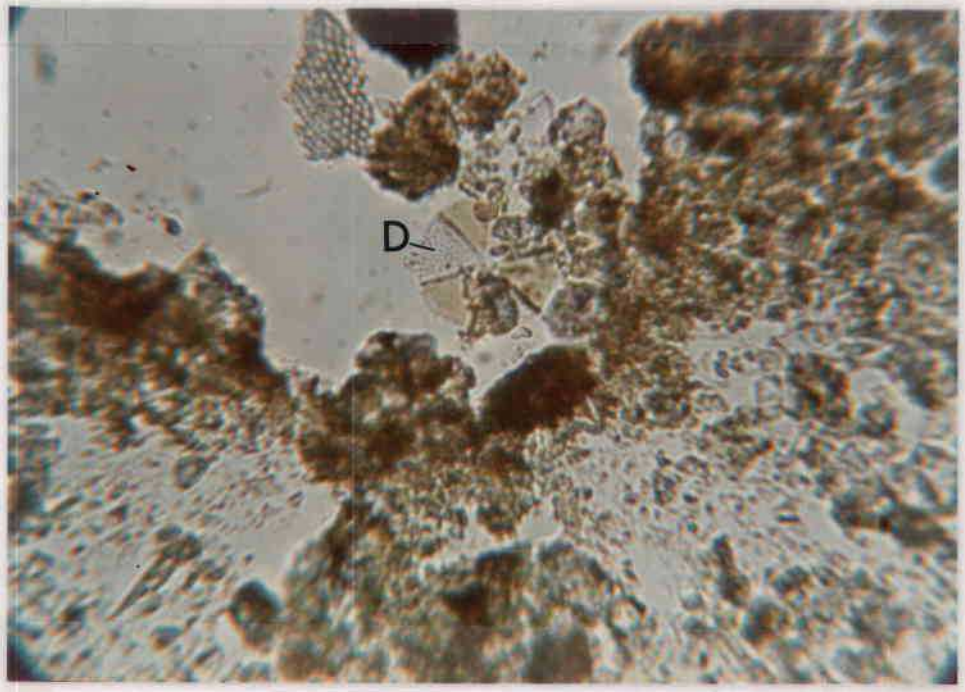


Figure 51 Photomicrograph of diatom frustules in siltstone facies 2 of Clifton formation. Plane light, field of view is 0.25mm across (OC 547, NW, Sec. 17, T8N, R6W).

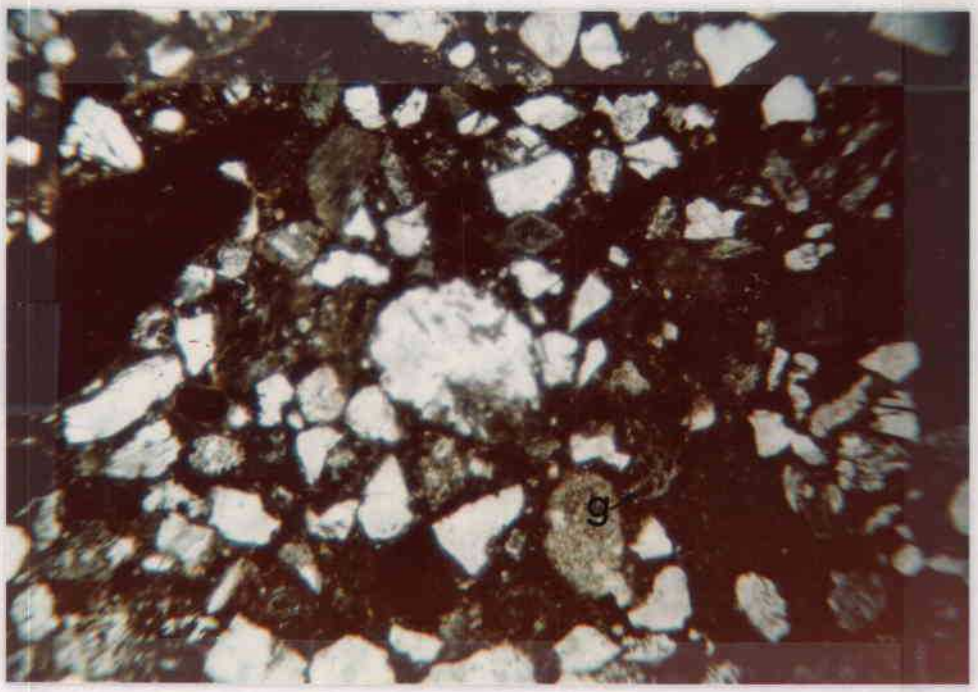


Figure 52 Photomicrograph of concretion in the shallow marine facies 1a of the Clifton formation. Note carbonized wood fragment in upper left and small gastropod in right center. Plane light, field of view is 4mm across (OC 111, Sec. 24, T8N, R8W).

alteration of the feldspars and unstable volcanic fragments to clays. However, in the shallow marine sandstone facies (1a), some matrix may be detrital due to mixing of clay layers with sandstone layers by bioturbation. The medium-grained sandstones associated with he blocks of siltstones and pebbly conglomerates in the channel facies 3 are considered a coarse-grained sandstone "matrix".

The fossil material in the Clifton formation consists principally of opaline diatom frustales (facies 2) and rare small gastropods (facies 1a). These were observed in smear slides and thin sections (Fig. 51 and 52).

Provenance

The Clifton formation represents a deposit at the mouth of a fairly large river and offshore submarine canyon system. The character and composition of the mineral assemblages suggest that the river drained granitic, metamorphic, and intermediate volcanic terrains. The paleocurrent directions indicate transport was generally from east to west. The compositional similarity between the middle Miocene Clifton formation and the present Columbia River sediment suggests both drained the same or similar source rock terrains.

The fluvial and shallow marine sandstone facies (1b and 1a) are dominated by granitic and metamorphic mineral assemblages. The submarine canyon head and associated slope deposits are also characterized by this same assemblage. The lower part of these facies has an intermediate volcanic material assemblage which is dominant. The major mineral constituents, angular to subangular potassium feldspar (orthoclase, microcline, oligoclase), polycrystalline and monocrystalline quartz, as well as very coarse-grained muscovite flakes, biotite, and

hornblende suggest a first cycle origin from a granitic or granodiorite source terrain. The subrounded to rounded heavy minerals such as zircon, magnetite, tourmaline, and garnet also indicate a plutonic source terrain (Williams and others, 1954). A low to high rank metamorphic provenance is indicated by the heavy mineral assemblage of kyanite, staurolite, sillimanite, actinolite, and epidote. Staurolite and kyanite indicate that the metamorphic rock was originally argillaceous (Deer and others, 1972); actinolite and muscovite may suggest contact metamorphism of green schist. The phyllites, quartzite, and quartz-mica schist rock fragments present in trace amounts support this interpretation.

There are no metamorphic or acid plutonic rocks exposed in the northern Coast Range of Oregon or Washington (Snavely and Wagner, 1963). Nelson (1978) has suggested a series of pre-Neogene metamorphic and acid igneous terrains including the North Cascades of Washington, the Precambrian Belt Series, the Eocene and Cretaceous batholiths of northeast Washington and southeastern British Columbia, the Blue Mountains of eastern Oregon and the Cretaceous Idaho batholith for the origin of similar metamorphic and acid plutonic heavy and major mineral assemblages in the older Astoria Formation. These same source areas may have supplied detritus to the Clifton Formation, particularly the Idaho Batholith which is principally composed of granodiorite. A geographically closer source for the granodiorite clasts could be the upper Eocene to upper Miocene intrusives of the Western Cascades described by Peck and others (1964). The subrounded to well rounded garnets, zircon, tourmaline, and quartz also indicate some recycling of older sedimentary units in the Oregon and Washington coast ranges (e.g. Astoria, Pittsburg Bluff, Scappoose, and Cowlitz Formations) or in eastern Washington (e.g.. Eocene Swank Formation). Because these minerals are so hard, they abrade

very slowly. Therefore, their well rounded form indicates a recycled origin (Pettijohn, 1975).

Volcanic detritus in the Clifton formation undoubtedly reflects erosion of older extrusive basaltic and andesitic rocks characteristic of the Teritary Western Cascades and eastern Oregon and Washington. Some Eocene and middle Miocene basalts of the Oregon and Washington also could have acted as sources. The hornblende and pyroxene andesites and dacite clasts, along with the abundance of zoned andesine, subangular green hornblende, hypersthene, and augite all suggest an extrusive intermediate to silicic volcanic source terrain. The more mafic ferromagnesian minerals, lamprobolite and augite, generally occur both as subrounded and very angular "hacksaw" grains, suggesting possible different sources or varying shorter distances of transportation. Hacksaw or sawtooth terminations on augite are common among augite grains in rivers that drain the Eocene Tillamook Volcanics of the Oregon Coast Range (Niem and Glenn, 1980). The origin of ther intermeidate to silicic detritus may be the Oligocene Sardine Formation and Little Butte Volanics or the Eocene to Oligocene Ohanapecosh Formation of the Western Cascades of Oregon (Peck and others, 1964) and Washington (Hammond, 1979). Nelson (1978) postulated these same sources for the older Miocene and Oligocene marine units of northwest Oregon which have similar heavy and major mineral assemblages as the Clifton Formation. Another possible intermediate source are the Tertiary volcaniclastics and flows of eastern Oregon described by Oles and Enlows (1971) (e.g. Eocene-Oligocene Clarno Group and Oligocene John Day Formation).

The lower beds of the submarine canyon and associated slope facies are characterized by an abundance of hornblende and pyroxene andesites and dacites, pumice fragments, and unaltered clear and brown volcanic glass shards. This influx of unstable

volcanic material may reflect an increase in penecontemporaneous volcanic activity possibly associated with an event related to the middle to upper Miocene Rhododendron Formation of the Western Cascades (Hammond and others, 1980). The return of the acid plutonic and metamorphic mineral assemblages dominance in the upper parts of turbidite sandstone facies 2 and submarine channel facies 3 and in the overlying fluvial facies (1b) indicates that the contribution from the volcanic events was not continuous during deposition of the Clifton.

The lack of abundant Columbia River basaltic clasts in the thick arkosic sandstones of the Clifton formation is surprising because the Clifton is bounded by the Columbia River Basalt. There is a thin (1 m) conglomeratic layer at the base of the Clifton in which cobbles of the underlying Frenchman Springs are found (OC 18, Sec. 9, T8N, R7W). This would be analogous to the sudden imput of volcanic ash and pumice into the Columbia River drainage system by the modern eruption of Mt. St. Helens. The great volume of this material temporarily overwhelmed the contribution of non-volcanic detritus. This paucity may be the result of topography; basalt flows tended to fill lows and did not create extensive, erodable highs. Another possible explanation is that the resistant nature of the finely crystalline, blocky basalt prevents extensive weathering and production of abundant sand-sized basalt fragments.

The composition of the modern Columbia River sands has mineral assemblages similar to the Clifton Formation. Whetten and others, (1968) showed that sands of the upper Columbia River (above Bonneville dam) are characterized by plutonic and metamorphic mineral assemblages with a proportionately low

percentage of volcanic detritus. These sands reflect the rocks of the North Cascades of Washington.

The Snake River which empties into the Columbia River and drains the Idaho batholiths also carries a similar acid igneous mineral assemblage. Whetten and others (1968) suggested a general granodiorite composition for these sediments and also expressed surprise in the lack of basaltic detritus from the surrounding Columbia River Basalts that cover much of eastern Washington and northeast Oregon. Lower Columbia River sands are characterized by an increasing amount of andesitic volcanic detritus downstream. Whetten and others (1968) suggested that the origin of this intermediate detritus is the high Cascade andesitic volcanoes, the older Tertiary volcaniclastics of eastern Oregon, particularly the Clarno and John Day Formations, and the Oilgocene to Pliocene rocks of the Western Cascades. Erosion is rapid due to the poorly consolidated nature of these tuffaceous rocks, the relatively high relief and more rapid weathering processes in western Oregon and Washington (Baldwin, 1976).

In summary, there appears to be close correlation of similar source rocks between present Columbia River sands and the Columbia Formation. The source terrains for both rivers appear to be many of the same volcanic, sedimentary, metamorphic, and acid plutonic units, suggesting that the drainage basin for the Clifton formation was similar to that of the present Columbia River although the High Cascades were not present. The Western Cascades may have been high rugged peaks, as the High Cascades are now, during Clifton time and was a major source of intermediate volcanic material.

In addition, lithologic and textural characterisitics and depositional environment of the Clifton formation (fluvial

facies 1b and shallow marine facies 1a) are very similar to the modern mouth and estuary of the Columbia River. The Clifton formation is sandwiched between two subaerial flow units (Pomona and Frenchman Springs) which are thought to have flowed down an ancestral Columbia River valley through a gap in the Cascade Range from eastern Oregon and Washington plateau sources (Snavely and others, 1973). Both the Clifton formation and the Columbia River also show physical relations to a submarine canyon system that cuts the slope and shelf (e.g. Astoria Canyon of the modern Columbia River and facies 2 and facies 3 of Clifton). These independent factors along with the present geographic position of the unit next to the Columbia River suggest that the middle Miocene Clifton formation is a partial sedimentary record of an ancestral Columbia River river mouth - canyon system.

Pittsburg Bluff

Modal analysis was performed on two thin sections and one heavy mineral grain mount of the Pittsburg Bluff sandstones (Table 10 and Appendix VI). The point counted sandstones are classified as volcanic arenite (Fig. 48). They are compositionally immature and texturally submature to immature suggesting rapid erosion and deposition of detritus with little reworking. Table 10 lists the major framework constituents. One very fine-grained sandstone (sample 49), typical of the Pittsburg Bluff Formation, and one very coarse sandstone (sample 47) were studied.

The quartz fraction consists of monocyrstalline and polycrystalline strained and unstrained grains. The relative abundance of each appears to be related to grain size; the more stable monocrystalline quartz is almost three times more common in the very fine-grained than in the coarse-grained sandstone.

Table 10

Modal Analyses of Sandstone Samples from Units other than Clifton

Mineral Species	Vantage Member	Pittsburg Bluff		Astoria Fm. Pipeline SS	Big Ck.	Oswald West
Sample No.	169	49	47	212	252	242
Quartz						
Mono.	12.7	17.1	7.4	22.8	22.5	13.2
Poly.	12.7	4.7	11.5	12.0	4.0	6.9
Chert	tr.	tr.	tr.	5.8	6.6	6.0
Quartzite	-	-	-	-	-	-
Feldspar						
Plagioclase	14.7	12.7	7.8	20.2	25.0	17.5
Potassium Feldspar	13.4	7.5	6.0	9.3	4.3	4.9
Rock Fragments						
Volcanic	13.9	20.3	31.2	12.0	27.2	19.2
Granitic	2.4	tr.	7.7	tr.	tr.	3.2
Metamorphic	2.0	0.7	1.2	0.7	-	.6
Sedimentary	-	-	tr.	-	-	-
Mica	2.9	3.0	1.1	7.1	0.8	3.7
Mafics	2.4	5.7	2.4	2.3	5.0	tr.
Fe-oxides	2.3	2.1		3.5	1.3	1.4
Carbonaceous fragments	-	-	-	-	-	-
Fossil fragments	-	5.2	-	-	tr.	-
Alteration Minerals	tr.	0.7	-	tr.	1.3	2.9
Volcanic Glass	-	16.1	7.6	-	0.8	-
Glauconite	-	-	-	-	tr.	-
Matrix	-	-	-	3.3	-	19.4
Cement	20.0	-	16.5	-	-	-

The opposite is true for polycrystalline grains which are more abundant in the coarse-grained sandstone. A possible explanation for this variation has been discussed in the Clifton formation petrography section.

Feldspar includes both fresh and altered plagioclase (andesine to labradorite An_{37-53}) and potassium varieties. The high percentage of intermediate to calcic plagioclase (13%) reflects the breakdown of porphyritic andesitic and basaltic volcanic rock fragments. Albite, pericline and Carlsbad twins are common in the plagioclase. Also observed were resorption voids in some grains, normally and oscillatory zoned grains, and glomeroporphyritic grains typical of volcanic sources. Orthoclase, microcline, and perthite are the common K-feldspars. Sanidine is also present in minor amounts.

Volcanic rock fragments are the single most abundant constituents in these sandstones (20-31%). This constituent with the addition of volcanic glass shards and pumice (8-16%) dominates the composition of the Pittsburg Bluff sandstone. Volcanic rock fragments include both porphyritic and non-porphyritic basalts, andesites, rhyolites (?) and tuffs. The glass includes bubble wall shards, glass fragments, and long tube pumice. The glass fragments are predominantly unaltered and, except for pumice, are more common in the finer grained sandstone (Fig. 53).

The very coarse-grained sandstone also contains a relatively high percentage of acid plutonic rock fragments. These hypidiomorphic granular granodioritic grains consist of an intergrowth of quartz, plagioclase, K-feldspar, and rarely brown hornblende. Grains with micrographic textures are also present.

Figure 53 Photomicrograph of very fine-grained volcanic-feldspathic sandstone in the Pittsburg Bluff Formation. Note sickle shaped glass shard to the right, and abundance of pumice (isotropic), and volcanic rock fragments. Elongate grain (right center) is muscovite. Crossed nicols, field of view is 1.3mm across (OC 46, NW, Sec. 6, T7N, R8W)

The presence of well preserved calcareous shell fragments is distinctive in the Pittsburg Bluff. The original internal shell structure is commonly preserved; pelecypods display both two and three layer prismatic and foliated calcite microstructures. The gastropods show the common carbonate cross lamellae shell structure described by Majewske (1969).

Heavy minerals (>2.94 sp. gr.), comprising from 3-6% of the sandstones are listed in Appendix VI for sample 49. They appear to be similar to those found in the Clifton formation although in slightly different proportions. The ferromagnesian heavy mineral constituents cosist primarily of green hornblende, orthopyroxene (hypersthene?) and clinopyroxene (augite?). Also present are epidote, brown tourmaline, garnet, kyanite, monazite, mica, and zircon.

Hematite cement accounts for up to 16% of the very coarse-grained sandstone. Its diagenetic orgin is due, in part, to percolation of ground water through this very porous sandstone causing oxidation of the iron rich minerals and rock fragments.

The silty sandstone and mudstone of the upper member were examined under a binocular microscope. They are composed principally of tuffaceous material and clay minerals with minor amounts of quartz, muscovite, and biotite.

Provenance

The shallow marine Pittsburg Bluff Formation was deposited during late Eocene time. The nature of the intermediate and basaltic volcanic detritus and marine character of the unit indicate that the source of the sand was older basalt highs in the Coast Range. The sediment received little reworking.

Alternatively, sand was rapidly transported from penecontemporaneous eruptive volcanic sources to the east, possibly the Western Cascades. Van Atta (1971) suggested local volcanic islands of older Goble and Tillamook Basalt to the south as a source for the basaltic material in the Pittsburg Bluff in the Nehalem River valley. The upper Eocene Colestin and Ohanapecosh Formations of the Western Cascades, which are composed of principally of pumice lapilli vitric tuff along with andesite and dacite flows (Peck and others, 1964; Baldwin, 1976), may be a possible source for the intermediate and silicic volcanic detritus of the Pittsburg Bluff. Although today the Colestin is primarily exposed to the south, and east of Eugene, Oregon equivalent units, now covered by younger Western Cascades rocks, may have been exposed during the late Eocene. A second more distant source for the volcanic material may be the Eocene Lower Clarno Group of central Oregon, which is composed of a series of andesites and basalt flows and tuffs (Oles and Enlows, 1971). The volcanic sediment may have travelled down an ancestral Columbia River drainage system through the Western Cascades and into the late Eocene sea.

The plutonic and metamorphic sources for the granitic rocks and aluminum silicate heavy mineral suite (Appendix VI) present in this unit are postulated to have originated from the same pre-Tertiary and early Tertiary units discussed for the Clifton Formation. The presence of the plutonic and metomorphic material, derived from as far east as the Idaho batholith, substantiates the existence of an ancestral Columbia River drainage system during the late Eocene.

Oswald West Mudstone

The very fine sand and coarse silt components of the Oswald West mudstone are composed of angular quartz and feldspar.

These framework minerals form between 10 to 50% of the clayey siltstone. The clay material matrix is composed of montmorillonite and chlorite intergrades with minor amounts of chlorite, mica and kaolinite (Nelson, 1978). There are rare sandstone interbeds associated with this mudstone. A glauconitic sandstone from the top of the unit and a pebbly sandstone were examined in this thin section.

The glauconitic sandstone is composed almost entirely of glauconite (85-95%) with minor amounts of quartz and plagioclase (An_{30-40}). Microcline, green hornblende, biotite, volcanic chert, and altered volcanic rock fragments as well as granitic rock fragments occur in trace amounts. The glauconite grains are well-rounded, elliptical green pellets which are, in part, oxidized to a moderate yellowish brown (10YR 5/4). The glauconite appears to have been formed from transformation of volcanic material and biotite under reducing conditions (Pettijohn, 1975).

The pebbly sandstone classifies as a feldspathic volcanic wacke (Fig. 48). The rounded to well rounded, medium to fine pebbles comprise from 46 to 55% of the rock. The remainder is made of fine sand (30%) and clay matrix (20%). A modal analysis of the sandstone (Table 10) shows the framework constituents to be: quartz (20%), feldspar (22%), volcanic rock fragments (19%), chert (6%), granitic rock fragments (3%), mica (3-6%) and iron-oxides (1%). A hematite stained clayey matrix (19.4%) forms the rest of the sandstone.

The angular, fine-grained quartz grains consist of monocrystalline (14%) and polycrystalline (7%) varieties. The feldspar is predominately plagioclase (oligoclase and andesine An_{18-41}; 18%). Microcline and orthoclase make up the small amount of potassium feldspar present (5%). Chert, with rare

plagioclase phenocrysts, appears to be principally of silicic volcanic origin.

Volcanic rock fragments include basalt, andesites, and dacites (?) with minor siliceous tuff and pumice fragments. These are the dominant pebble compositions (63%) (See Appendix VII for compilation of pebble count). In thin section the fragments of tuff and pumice are extensively altered to clay minerals. The mafic to inetremediate volcanic fragments disply porphyritic, intersertal to hyalopilitic textures. Associated with the volcanic pebbles are minor metamorphic quartzite and granodiorite (?) pebbles. The quartizte has both sutured and polygonized grains. The hypidiomorphic granular granodiorite is composed of interlocking crystals of quartz, plagioclase, and potassium feldspar.

The hematite-rich matrix (19%) is composed of clays which may have originated, in part, from diagenetic alteration of the volcanic rock fragments. Hematite is the result of weathering of the mafic volcanic and ferromagnesian constituent formed by ground water percolation and oxidizing pore water.

Provenance

A tuffaceous nature of the Oswald West mudstone is suggested by minor altered pumice and tuff clasts in the conglomerate and by the abundance of detrital montmorillonite clay which commonly forms from alteration of silicic ash (Nelson, 1978). The contemporaneous Oligocene Little Butte Volcanic tuffs of the Western Cascades (Peck and others, 1964), Fisher and Eugene Formations (Baldwin, 1976) and the John Day Formation are likely sources for the altered volcanic ash in the Oswald West mudstone. Local basalt highlands in the Coast Range (Tillamook and Goble Volcanics) of Eocene age may still

have been supplying material during deposition of the Oswald West mudstone as they were during deposition of the Pittsburg Bluff Formation. More distant source terrians for the intermediate and basaltic volcanic clasts would be the Eocene to Oligocene Clarno Group of Central Oregon (Oles and Enlows, 1979), and the Oligocene Ohanapecosh Formation of south central Washington (Fiske and others, 1963). The deltaic deposits of the Scappoose Formation which are in part equivalent to the Oswald West, also are very tuffaceous in nature (Van Atta, 1971) and support the tuffaceous source for much of the Oswald West mdustone. In addition, the scarce quartz, microcline, and granitic clasts indicate some contribution of older pre-Tertiary and Eocene acid plutonic and metamorphic sources from eastern Oregon, eastern Washington and Idaho discussed in the section on petrography of the Clifton Formation.

Astoria Formation

Four thin sections and two heavy mineral grain mounts were examined from the Astoria Formation. These slides were studied to help clarify the correlations of the members from Coryell's (1978) thesis area (Wickiup area - Big Creek area) into this study area and to contrast the compositions of the Big Creek member with the upper Silver Point and Pipeline mudstones in this study area. Pebble counts were also performed on two conglomerate beds in the Big Creek member.

A turbidite sandstone (sample 212) in the Pipeline is classified an arkosic wacke (Fig. 48). A Big Creek shallow marine sandstone (sample 252) is an arkosic arenite. The greater percentage of fines in the Pipeline sandstone may be related to diagenetic alteration of the feldspars. If that is the case, then the original composition of the arkosic sandstone in these two members was quite similar. Point count

analysis, however, shows that there are distinct differences between these two members of the Astoria Formation (Table 10). The major framework constituents of these two sandstones are quartz, fedspar, rock fragments, mica, heavy minerals, and chert. Minor components include volcanic glass, alteration minerals, and glauconite.

Monocrystalline strained and unstrained quartz comprise approximately 23% of each sample. Polycrystalline quartz grains are three times more abundant in the Pipeline member than in the Big Creek member (12% vs. 4%). Chert, the other stable mineral component, occurs in equal amounts in both samples. It is primarily of silicic volcanic origin although not exclusively.

The Pipeline and Big Creek samples contain similar abundancces of feldspar, but there is a significant difference in the type of feldspar present. The Pipeline sandstone is composed of plagioclase oligoclase - andesine (An_{25-35}) (2%) and potassium feldspar (9%). The K-feldspar consists of microcline and orthoclase and rarely perthite. Big Creek has 25% plagioclase, oligoclase - labradorite (An_{20-61}) but only 4% potassium feldspar. Coryell (1978) also noted this variation in feldspar content between the two members.

Volcanic ropck fragments form 27% of the Big Creek sample. They include hornblende and orthopyroxene andesites and glassy basalts. There are only 12% volcanic rock fragments in the Pipeline member. Trace amounts of granitic and metamorphic rock fragments occur in both samples.

The Pipeline sample has 7.3% mica, whereas the Big Creek sample has less than 1%. Coryell (1978) determined that both

members contain 8% mica in the Big Creek - Wickiup Mountain area.

Heavy mineral in the deep-marine Pipeline (sample 212) and Big Creek shallow-marine sandstone (Sample 246) are listed in Appendix VI. This second Big Creek sample (no. 246) was collected from a fine-grained arkose that is lower in the section than the silty sandstone (Sample 252). The major differences in heavy mineral suites is the higher abundance of biotite, kyanite, staurolite, and zircon in the Pipeline sample. There is also a greater proportion of mafic minerals in the Big Creek sample. In particular, green hornblende, lamprobolite, hyhpersthene, and enstatite dominate.

The comparison of these few sandstone samples shows a larger amount of volcanic rock fragments and volcanic-derived material in the Big Creek unit and a greater proportion of micas, polycrystalline quartz, K-feldspar and metamorphic and acid plutonic rock fragments in the Pipeline member. The Pipeline sample appears to be compositionally more mature. These differences, however, were not reported by Coryell (1978) in his comparison of the same units to the west.

The coarse silt and very fine sand of the Silver Point mudstone are composed of quartz, feldspar, and muscovite with minor amounts of magnetite and chlorite. The clay mineral montmorillonite, mica, and kaolinite comprise the abundant fine silt and clay size fraction according to X-ray diffraction studies of these units in the Young's River area reported by Nelson (1978). The mudstone is characterized in this study area by an abundance of carbonaceous plant material (5-6%), by the arkosic fine sand and silt composition, and by alignment of mica flakes and elongate carbonaceous plant stems parallel to bedding. This is in close agreement with the description of

the upper Silver Point mudsonte to the west (Coryell, 1978; Nelson, 1978).

In contrast, the coarse silt and fine sand size fraction in the subordinate mudstone layers in the Big Creek member is composed of quartz and feldspar but lacks abundant mica. Also present is an abundance of volcanic detritus including pumice, clear shards and ferromagnesian minerals (hornblende and clinopyroxene). Siliceous (opal) planktonic fossils also occur in the very fine-grained sandstone, indicating the lack of extensive diagenetic alteration. Diatoms, radiolarians, and sponge spicules were recognized in the finer fractions (OC 692, Sec. 2, T7N, R6W). The abundance of volcanic detritus, siliceous biogenetic fossil debris, lack of mica, and the non-alignment of the silt and fine sand grains distinguishes the subordinate mudstone from the upper Silver Point and Pipeline mudstones to the west.

Pebble counts were performed on two pebble conglomerates within the Big Creek member. The pebbles are predominantely volcanic (53 to 74%), consisting of basalt, andesite, and undifferentiated siliceous volcanics and tuffs. Other minor rock types are metamorphic quartzite and quartz mica schists, chert, and granodiorite (?) (see Appendix VII).

In summary, the Big Creek member in this study area appears to have a greater amount of volcanic material than the Big Creek member at the type locality to the west (Coryell, 1978). The upper Silver Point and Pipeline members are compositionally more mature and show more diagenetic alteration than the facies member in this study area. They are compositionally similar to the upper Silver Point and Pipeline members of the Astoria Formation described by Coryell (1978) and Nelson (1978) to the west.

Provenance

The source of the sediment in the Astoria Formation is very similar to that of the previously discussed formations. The principal source of the basaltic volcanic detritus is the Eocene aged Coast Range volcanics of Washington and Oregon (Tillamook, Goble, and Crescent Volcanics). An ancestral Western Cascades source (Little Butte Volcanics, Sardine Formation, and Ohanapecosh Formation) is postulated for the andesites and silicic tuffaceous debris. The variability of the volcanogenic material in the different members may reflect varying amounts of eruptive and erosional activity in the Western Cascades during this time. As in the younger Clifton Formation, the Eocene-Oligocene Clarno Group and Oligocene John Day Formation of central Oregon are other possible sources for the volcanic material.

The non-volcanic detritus must have been derived from east of the Cascades. As suggested for the Clifton formation (see Provenance Section), there is no source for the metamorphic and K-feldspar rich granitic clasts in the Oregon and Washington Coast Range or in the ancestral Cascades although some Eocene-Oligocene granitic plutons occur in the Western Cascades. This material is postulated to have come from the early Tertiary and pre-Tertiary plutons and metamorphics of eastern Oregon, Washington and Idaho. The material was transported by an ancestral Columbia River drainage system. Previous workers in the Astoria Formation have also postulated the existence of this river system (Cressy, 1974; Coryell, 1978; Nelson, 1978; Penoyer, 1977; Tolson, 1976; and Cooper, 1980).

Grain Size Analysis

Thirty sandstones samples were sieved and 12 mudstone samples underwent hydrometer analysis, using Royce's (1970) statistical methods, in order to obtain grain size statistics for the various sedimentary units within the study area. The analyses were done to quantify field descriptions of the units and to show textural differences between and variations within units. These size statistics, when plotted on binary graphs produced by Friedman (1961) and Passega (1957), provide an additional indication of the paleoenvironment of these Tertiary units.

Appendix V lists the raw data from these analyses. Folk and Ward's (1957) statistical formulas for median size and values of sorting, skewness, and kurtosis were used whenever possible but Inman's (1952) formulas and Trask's (1932) quartile measures were necessary in some cases where the cumulative frequency curves were insufficiently complete. Incomplete curves resulted from abbreviated hydrometer analyses which were primarily designed to determine percentage of silt and clay in the mudstones.

Table 11 is a verbal summary of the quantitative size analysis listed in Appendix V. The intention of this table is to facilitate description of the units and to aid in showing differences between and variations within formations. The percent clay is left out because it can be obtained by subtracting the sum of the sand and silt fromm 100. The lack of a silt value indicates that no hydrometer analysis was performed for that sedimentary unit. In that case, the percent silt and clay is obtained by subtracting the percent sand from 100. Kurtosis values are not listed for the siltstones and mudstones due to the incomplete frequency curves.

Table 11

Summary of Statistical Parameters from the Grain Size Analysis of the Tertiary Units in this Study Area

Unit	% Sand	% Silt	Grain Size	Sorting	Skewness	Kurtosis	Folks (1951) Textural Maturity Index
Pittsburg Bluff							
Fine ss (1)	90	9	v.f	Mod-well	Pos.	Lept.	Im
Coarse ss (1)	99	-	v.c	Poor	V. Pos.	Lept.	Sub
Oswald West							
Mudstone (1)	1	70	silt-clay	Poor	V. Pos.	-	-
Astoria Formation							
Pipeline Mudstone (ss beds) (2)	83	15	f-v.f	Mod-poor	V. Pos.	Lept.	Im
Silver Point Mudstone (1)	5	60	silt-clay	Poor	V. Pos.	-	-
Big Creek Sandstone (3)	9-80	18-75	v.fs-silt	Poor	V. Pos.	Lept.	Im
Basalt Interbeds							
Columbia River Sandstone (4)	85-95	-	f (med)	Mod-poor	V. Pos. (Pos)	Lept. (v. Lept.)	Sub
Coastal Bas. Mudst (1)	5	65	silt-clay	Poor	V. Pos.	-	-
Clifton Formation							
Facies 1a (2)	75	15	f-v.f	Poor	V. Pos.	Plat	Im
Facies 1b (10)	90	-	f (m&c)	Mod (Poor)	Pos. (V.Pos.)	Lept. (v. Lept.)	Sub (Im)
Facies 2 Siltstone (2)	10-20	50-60	silt	V. poor	Sym-pos.	-	-
Sandstone (2)	50-90	10-25	f-v.f	Mod (V. poor)	Pos. (V.Pos.)	Meso.	Im
Facies 3 (3)	85-90	1	c-f	Poor	Pos.	Lept. (ex. lept.)	Im-sub

Data from Appendix V ; values from Royse (1970). Values in parentheses are present but less common; dashes between values indicate range. Numbers in parentheses indicate number of samples analyzed. Textural maturity: Im - Immature; Sub - submature, after Folk, (1951). Kurtosis: Lept.- Leptokurtic; Plat - Platykurtic; Meso - Mesokurtic.

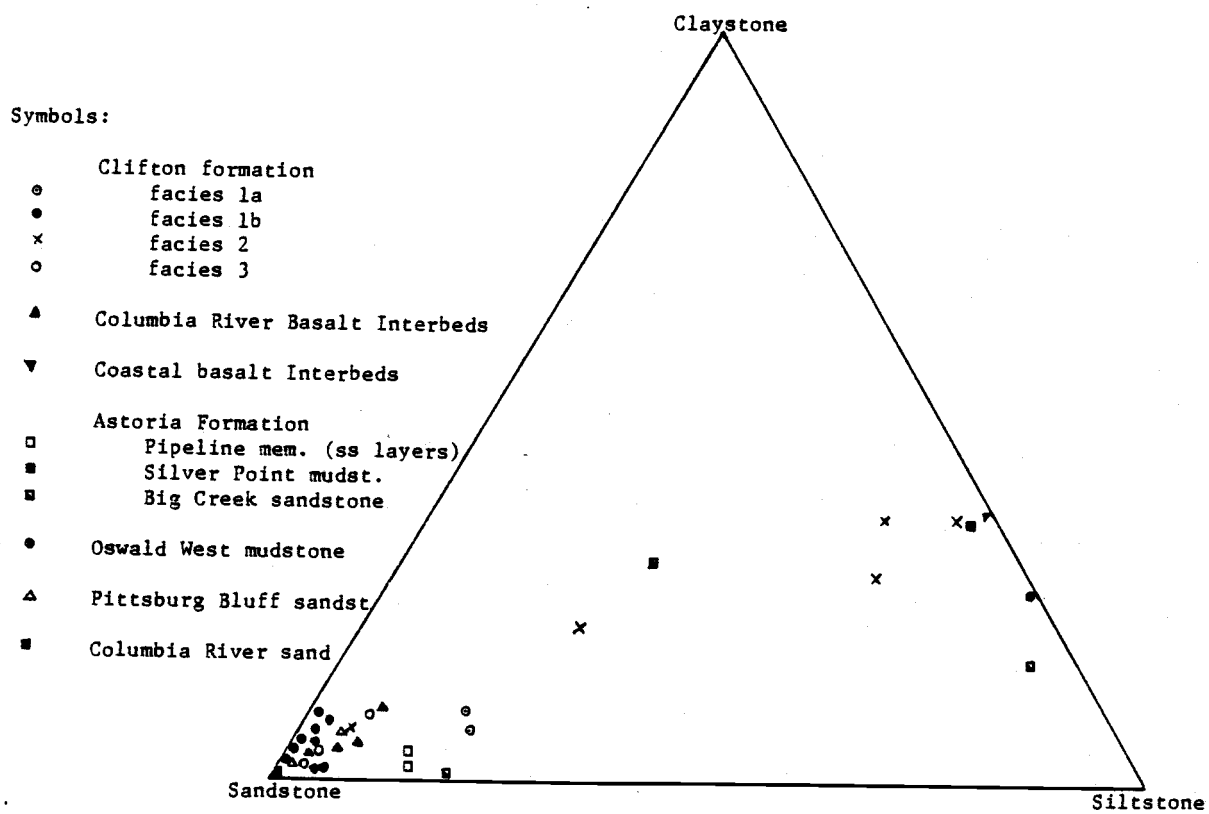


Figure 54 Grain size plot of sedimentary units from the study area. Data listed in Appendix V. Diagram modified from Shephard, 1954.

Figure 54 is a ternary diagram modified from Shephard (1954) which divides clastic rocks into size classes based on the percent of sand, silt, and clay present.

Some observations from the summary of the statistical textural parameters listed in Table 11 and on the ternary grain size classification diagram (Fig. 54) show that:

1. The coarse-grained samples plot in the sandstone class whereas the finer-grained mudstones classify as clayey siltstone. Only two samples plot in the sandy portion of the clayey silty sandstone class in Figure 54.

2. The very-fine grained sandstone of the Pittsburg Bluff Formation appears to have a grain size distribution slightly less skewed and less peaked than the other sandstone units. The coarse-grained sandstone is distinctive from other samples by its coarse-grained size.

3. The Oswald West and upper Silver Point mudstones are dominated by the silt size fraction. There appeats to be no distinctive textural differences between the two units.

4. The Clifton siltstone (facies 2) when compared with the Oswald West and Silver Point mudstones has a greater percentage of sand, slightly less silt, is more poorly sorted and less skewed.

5. None of the samples contained more than 38% clay size material.

6. The turbidite sandstone interbeds of the Pipeline mudstone member display no unique textural characteristics except a higher percentage of silt, thus are finer grained. This may be the result of a diagenetic decomposition of feldspars to clays and/or turbidite deposition with no further winnowing of detrital fines.

7. The shallow marine Big Creek sandstone is finer grained, has poorer sorting, and is more positively skewed (e.g. more finer grain sizes) then the fine-grained shallow marine Pittsburg Bluff Formation.

8. The arkosic sandstone interbeds within the Columbia River Basalts are texturally very similar to the fluvial facies 1b of the Clifton formation although they appear to be slightly finer grained.

9. A comparison of the facies within the Clifton Formation shows:

   a) facies 1a (shallow marine bar) is deficiently peaked (platykurtic), possibly due to mixing by bioturbation after deposition, and has a higher percentage of silt associated with it;

   b) facies 1b (fluvial) is slightly finer grained and better sorted than facies 3 (channel facies), and

   c) the turbidite sandstone layers in facies 2 (overbank) tend to have more silt size material and are less skewed symmetrically then the other facies in the Clifton.

There are a number of qualifications which must be placed on the interpretative value of these statistical grain size parameters in order to understand what they indicate and how valid they are.

Theoretically, Folk and Ward's statistical grain size measures, which emcompass the largest portion of the cumulative frequency curves, are the best approximation of the statistical parameters. Freidman (1962) has graphically shown that Folk and Ward's sorting coefficient and Inman's sorting measure correlate reasonably well and that these two along with Trask's quartile measures can be related to the standard deviation. This is not the case for shewness and kurtosis values and therefore direct comparison of these parameters using Folk and Ward's measures and Trask's and Inman's formulas is of less value.

Grain size analyses are standardized so that results can be easily reproduced for meaningful comparisons. The analyst may introduce small variations which, if compensated for by having the "same" operator variation throughout the study, will not adversely affect the results. Larger operator errors may be introduced if more than one analysist is involved in a comparison, even though the same sieve and hydrometer method are used. In comparison of the Pipeline sandstone of the Astoria Formation, analyzed by Nelson (1978) and Coryell (1978), with the Clifton formation, some of the variation seen could be due, in part, to operator error.

A final qualification which should be mentioned in this grain size analysis is related to the methods employed. The mechanical sieving technique measure grain size distribution by determining the shape from the diameter of the b-c axis of the grains (e.g. minimum cross-section diameter that falls through

Symbols:

Clifton formation
- facies 1a
- facies 1b
- facies 2
- facies 3

Columbia River Basalt Interbeds

Coastal basalt Interbeds

Astoria Formation
- Pipeline mem. (ss layers)
- Silver Point mudst.
- Big Creek sandstone

Oswald West mudstone

Pittsburg Bluff sandst

Columbia River sand

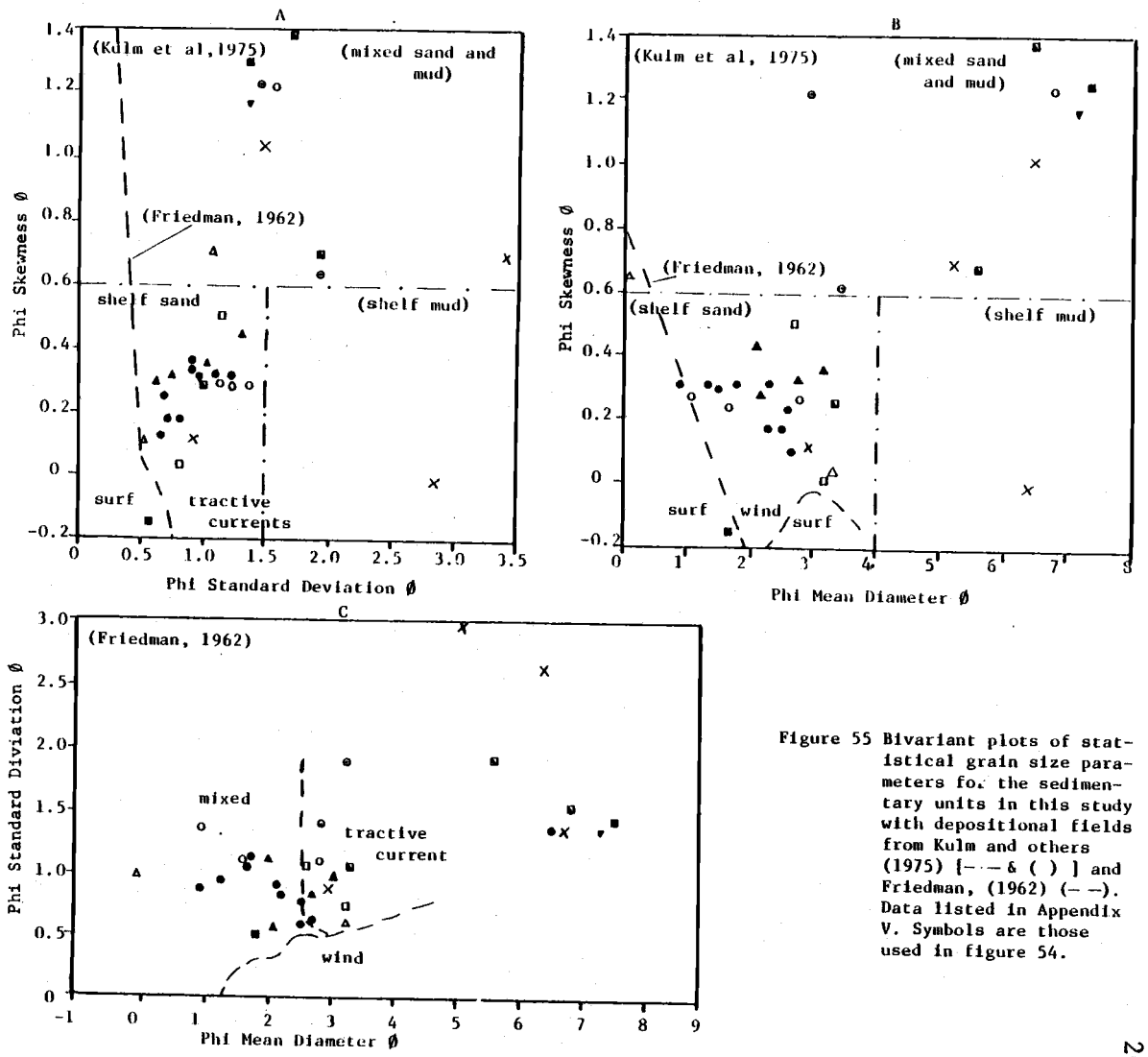


Figure 55 Bivariant plots of statistical grain size parameters for the sedimentary units in this study with depositional fields from Kulm and others (1975) [— — & ( )] and Friedman, (1962) (— —). Data listed in Appendix V. Symbols are those used in figure 54.

a square sieve opening), while the hydrometer method measures the grain size distribution by relating the hydraulic setting velocity of the grains to the specific density, shape, and size (Royse, 1970). Therefore, when both methods are employed to describe a frequency curve, different parameters are measured and are equated to size. Ideally, the settling velocity would be best to reflect the depositional conditions for the range of ragin sizes in the sandstones and silstones of the study area, but a settling tube for the sand grains was not available. It has been shown, however, that the actual difference between the two methods is small if the settling velocity is used for sizes less than .062 mm (40) and mechanical sieving techniques for larger grain sizes (Folk, 1974; Royse, 1970).

The statistical grain size parameters from the sieve and hydrometer analyses were plotted on environmentally sensitive binary graphs of Friedman (1961, 1962), Kulm and others (1975), and Passega (1957) in order to show size variations between the formations defined and to aid in interpretation of the depositional environment of these units. Binary graphs of standard deviation (sorting) versus phi skewness, skewness versus mean diameter, and phi standard deviaton versus phi mean diameter are shown in Figures 55a, 55b, and 55c. Disregarding any environmental implications of these graphs, it appears that Figure 55b, phi mean diameter versus skewness, is the best for comparative analysis of the sedimentary units in the study area, although Figure 55a is also of some value.

A few further observations, not as obvious from Table 11 or from the ternary classification diagram (Fig. 54) which can be seen in these binary graphs are:

1) the fluvial sandstone interbeds between the Columbia River basalt flows tend to be slightly more positively

skewed and finer grained than the Clifton formation fluvial sandstone (facies 1b);

2) the shallow marine subfacies (facies 1a) of the Clifton formation segregates from the coarser, cleaner fluvial and channel sandstone facies (facies 1 and 3) and from the much finer grained older mudstone units; and

3) there is little statistical grain distinction between the fluvial sandstone and submarine channel sandstone (facies 3) of the Clifton formation though facies 1b may be better sorted than facies 3 (Fig. 55a).

The use of these diagrams as a tool for paleoenvironmental interpretation of ancient sandstones is based on statistical size analysis of sediments from modern sedimentary environments which have been extensively sampled and the results plotted with best fit boundaries separating differenct known enviornments drawn. Friedman (1961) was one of the first sedimentologists to do this and created three environmentally sensitive graphs which must be compared together to determine which environment or process is the dominant one for creating the statistical size distribution observed in the sample (Fig. 55). Recently, Tucker and Vocher (1980) have questioned the effectiveness of these fields and boundaries in determining whether a definitive paloenvironment interpretation can really be made. They suggest that the probability of error can be up to 35 $\pm$ 15% if one of three basic transport mechanisms (surf, tractive, wind) on Friedman's graphs cannot be eliminated. This error drops to 25 $\pm$ 15% if one of the three processes can be eliminated. In addition, on Firedman's graphs the environmental fields were based principally on submature quartzo-fedspathic sands on the East Coast of the United

States, not on the immature volcanic sands of the West Coast which may produce a different grain size distribution during transport. Friedman (1967) recognized problems with these bivariant plots and indicated that they should be used along with other environmentally sensitive field and lab data in interpretation of paleoenvironment. He also suggested that a possible problem may be that the grain size distribution of a sandstone could reflect a previous environment or depositional process(es) if rapidly transported then buried before sufficient reworking could be achieved in the final depositional environment (e.g. river to beach). That is, the grain size distribution may not reflect the grain size segregation of the processes in the last depositional environment if there was insufficient time for the grain size distribution to come into equilibrium with those processes before burial.

The sedimentary units in the study area invariably plot into a single environmental field on Freidman's graphs, usually tractive currents, although some plot close to the boundary line between surf or wind and tractive currents. Wind deposition for these sands can be eliminated on the basis of comparison of all three graphs. The fluvial sandstone facies of the Clifton formation with the large-scale trough cross-bedding, point bar and coal deposits of a subaerial river deposit fits this tractive interpretation from grain size analysis. The sandstones interbedded with the Columbia River Basalts, which are thought to reflect a river environment from field evidence, plot in the tractive current field (Fig. 55). The submarine channel facies (3) associated with the river deposits may have been depositedly rapidly enough to retain the characteristics of fluvial sediment. It must be mentioned that these graphs do not take sedimentary gravity flows into account; and since this mechanism is closest to the tractive

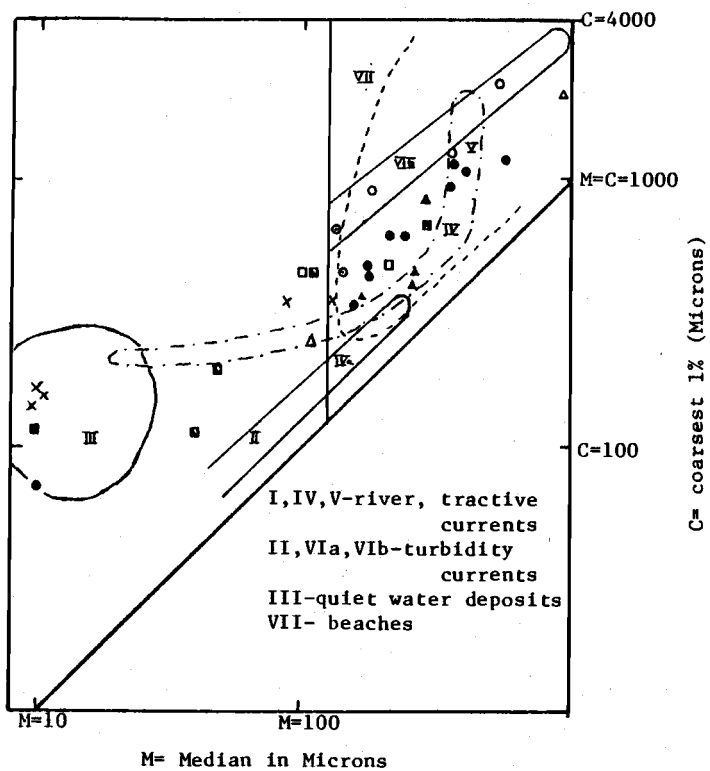
current transport mechanism, it is not surprising that these sediments plot in this field.

The other sandstone units in the study area, the Pittsburg Bluff Formation, sandstone interbeds of the Pipeline mudstone, and the Big Creek member are thought to be marine in origin. Molluscan fossil data indicates an inner shelf marine environment for the Pittsburg Bluff sandstones, in correspondence, the grain size statistics of these sandstones plots in both the tractive and surf fields of Friedman's graph. The Pipeline sandstones have been interpreted as deep marine turbidites and the fossiliferous Big Creek member as an inner to middle shelf sandstone deposited by tractive (e.g. longshore current) and wave processes (Nelson 1978; Coryell, 1978). The Big Creek samples plot in the tractive fields which may be related to bottom currents along the continental shelf. Friedman's (1961, 1062) graphs are not useful in analyzing the depositional environment of the mudstone units in the study are because these graphs were developed from sands collected in modern environments.

Kulm and others (1975) used the same binary statistical plots to separate modern shelf sands from mixed bioturbated mudson the continental shelf off the Oregon Coast. The fields and boundaries between these fields or environments defined that study are also drawn on Figures 55a and 55b. The shallow marine sandstone of the Clifton formation (facies 1a) plots in the mixed sand and mud field of the middle to outer continental shelf reflecting the mixed grain size due to bioturbation. This Clifton sandstone, however, plots a good distance away from the finer grained Oswald West and Silver Point siltstones which also plot within or above the middle to outer shelf mixed sand and mud field. The glauconite, molluscan fossils, and micro fossil assemblages in these siltstones also indicate that

deposition occurred in low energy outer shelf or upper slope depths (see depositional environment sections for Oswald West and Silver Point). The fossiliferous deep-marine mudstone interbed between submarine coastal basalts also plots close to the Silver Point outer shelf-slope mudstone. The Pittsburg Bluff fine-grained sandstone sample plots in the inner shelf sand region but the poorly sorted coarse sandstone sample plots just above the inner shelf line in the zone of mixed sand and mud. The coarse sandstone is rather anomalous and may be the result of mixing of coarser sediment by higher energy storm waves on the shelf. Alternatively, these mat be coarse-grained surf deposits based on Friedman's graphs. Big Creek shallow marine sandstones plot among inner shelf sands whereas the Big Creek clayey siltstones plot in the mixed middle to outer shelf facies. This interpretation agrees with environmental conclusions drawn from sedimentary structures (mostly parallel laminations) and fossils from these units (see depositional environment of Big Creek member). The graphs of Kulm and others (1975) are not useful for environmental interpretation of the fluvial sandstones of the Clifton formation (1b) or of the fluvial sandstone interbeds between the Columbia River Basalt flows because grain size statistics for modern Oregon fluvial sands were not used in constructing these graphs.

Diagenetic breakdown of unstable sand sized detrital grains may reduce the original grain sizes and decreases the usefulness of these plots as environmental indicators. This may be the case for the turbidite and fluvial sandstone interbeds of the Pipeline mudstone and Columbia River Basalts which appear to have higher percentatges of fines present. Alternatively, the abundant fines may be a result of rapid deposition and burial such that detrital fines were not winnowed away by later reworking.



Symbols:

Clifton formation
- facies 1a
- facies 1b
- facies 2
- facies 3

Columbia River Basalt Interbeds

Coastal basalt Interbeds

Astoria Formation
- Pipeline mem. (ss layers)
- Silver Point mudst.
- Big Creek sandstone

Oswald West mudstone

Pittsburg Bluff sandst

Columbia River sand

Figure 56 Plot of sedimentary units from study area on "C-M" diagram modified from Passega,(1957). Symbols used same as those used in figure 54 .

The binary "C-M" diagram of Passega (1957), uses the coarsest 1% and median % to define different depositional environments (Fig. 55b). This graph is also of some interpretive value in indicating possible environments for the various units in the study area. The Oswald West and Silver Point mudstones as well as the siltstone slope facies (facies 2) of the Clifton formation plot in or near the quiet water field (III). This is in agreement with the fossil data available for these units.

The Clifton fluvial sandstones generally plot in or parallel to the middle and upper tractive current field (IV and V) on Figure 55b, while the submarine channel deposits fall in the upper turbidity current field (VIb). This agrees with the other depositional information available for this unit (see Depositional Environment Section Clifton Formation).

The very fine-grained shallow marine sandstone of the Pittsburg Bluff Formation plots in the lower tractive current field (I). Again, this may indicate transport by bottom currents below wave base. This graph must be used with caution because, by itself, it is not definitive of a depositional environment.

In summary, the grain size statistics of the sedimentary units in the study area were plotted on the evironmentally sensitive graphs of Friedman (1961, 1962), Passega (1957), and Kulm and others (1975). The depositional environments determined by this analytical method generally agree with the environments indicated by other independent data derived from the field and lab and by the paleontological data. That is, the Big Creek and Pittsburg Bluff sandstones were deposited in an inner shelf environment by tractive current processes. The various Clifton sandstone and siltstone facies formed in

tractive, fluvial, and inner to middle shelf and outer shelf/slope environments and by turbidity currents. The Silver Point mudstones and Oswald West mudstones accumulated in quiet water in a middle to outer shelf environment.

Comparison of Statistical Size Parameters of the Clifton Formation and the Pipeline sandstone of the Astoria Formation

Table 12 is a verbal summary of quantitative grain size data of the Clifton formation and of the Pipeline sandstone. The data for the Clifton formation are from fluvial sandstone (facies 1b), shallow marine facies (facies 1a), and submarine channel sandstone (facies 3). The Pipeline data are based on 33 samples from Nelson (1978) and Coryell (1978) and two samples from this thesis. As this table shows, the Clifton formation is generally finer grained (mean), slightly better sorted (standard deviation), less skewed, and less peaked size distribution (leptokurtic) than the Pipeline sandstone. Nelson (1978) suggested that the grain size distribution of the Pipeline sandstone has been affected by diagenetic alteration of the unstable volcanic and mineral constituents. It appears, from petrographic studies that the Clifton formation is not as the Pipeline sandstone of the Nicolai Mountain-Gnat Creek area. This could account for some of the statistical differences and is reflected in the slightly greater percentage of fines in the Pipeline sandstones. Thus, it likely that the overall grain size distribution of these two units reflect different depositional environments and processes.

A binary plot of skewness and standard deviation versus mean diameter (Fig. 57) is also a way to visually differentitate the two units. The separation is not obvious, but as Table 12 shows, the Clifton plots slightly lower (less skewed) and finer grained than the Pipeline sandstone.

TABLE 12

Comparison of Statistical Grain Size Analyses for the Clifton formation Sandstones and the Pipeline Sandstone of the Astoria Formation

	Clifton Formation (13) (data from Table 11)	Pipeline Sandstone (34) (From Nelson, 1978, and Coryell, 1978)
% Sand	>90 (80-90)	80-90 (> 90)
Grain Size	fm (c-m)	med (fn-v.f)
Sorting	mod-Poor	Poor (mod)
Skewness	pos (V. pos)	V. Pos (pos)
Kurtosis	lept (v. - ex lept.)	V. Lept. (Lept)

Parentheses indicate it occurs but is not common.

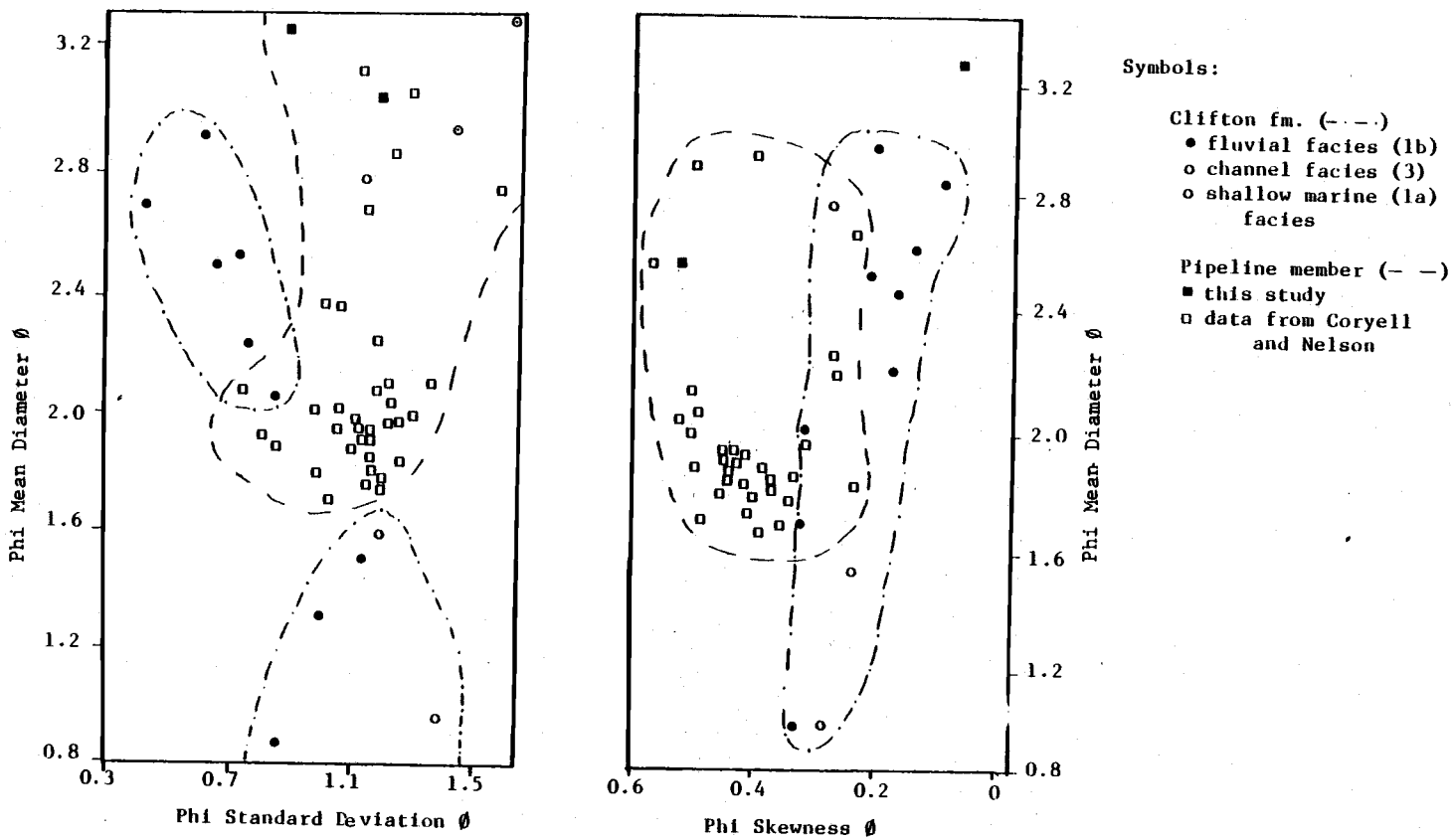


Figure 57 Comparison of textural size parameters for the Clifton formation and Pipeline member of the Astoria Formation. Includes data from Coryell (1978) and Nelson (1978). Shallow marine facies does not plot on skewness vs. mean diameter diagram.

Both Nelson (1978) and Coryell (1978) based on petrography and heavy minerals, suggested that the source of the Pipeline sandstone submarine channel was a river system which drained eastern Oregon, Washington and Idaho similar to the Columbia River today. I have suggested a similar source for the Clifton formation (see mineralogy section of Clifton). The lack of significant differences in grain sizes, the similar source rocks and mineralogy, and the close geographic position of the units suggest that the Clifton fluvial facies was deposited by a younger river system but very similar to the one which feeds the Pipeline sandstone.

Coryell (1978) postulated that the Clifton formation might be the fluvial feeder system or channel head facies for the Pipeline sandstone channel which cuts into the older Silver Point mudstone of the Astoria Formation. However, this study shows that the Clifton formation is finer grained, contains a lower percentage of polycrystalline quartz and K-feldspar, and is younger (Luisian based on diatoms) than the Pipeline sandstone (Saucesian based on forams). Therefore, it is unlikely that the Clifton was the fluvial feeder system or channel head facies for the Pipeline sandstone. Geologic mapping in the western part of this study area, furthermore, shows that the shallow marine-fluvial facies (1a and 1b) and the submarine channel facies (3) of the Clifton stratigraphically overlie the Pipeline sandstone (e.g. measured section at Tripp Road).

STRUCTURAL GEOLOGY

Regional Geology

A pattern of major vertical crustal movement in the Pacific Northwest appears to be related to periodic underthrusting of the eastward moving Farallon Plate under the North American Plate. This compressional regime is manifested by: uplift and imbricate thrusting along the continental slope and outer continental shelf; subsidence in the marginal basin along the inner shelf; and uplift in the Coast Range (Snavely and others, 1980a).

The structure of the northern Oregon Coast Range reflects this generally compressional field. The major structural features is a northward plunging anticlinorium. It consists of north and northwest striking gentle open folds (Wells and Peck, 1961) which suggest a normal compressional regime (Braislin and others, 1971). Northwest- and northeast-trending high-angle normal and reverse faults cut these folds (Snavely and Wagner, 1974; Niem and Van Atta, 1973). The core of the anticlinorium is composed of lower to middle Eocene tholeiitic volcanics (Tilliamook Volcanics) extruded from a sea floor ridge and/or sea mounts prior to the shift westward of the underthrust boundary between the Farallon and North American Plates (Snavely and others, 1980b). The upper Eocene to middle Miocene sedimentary strata which flank the Coast Range anticlinorium were deposited in an arc-trench gap marginal basin (Niem, 1976). This area is represented by Western Cascades and a trench facies presumably occurs offshore and in the Olympic Mountains (Hoh melange of Rau, 1976). Uplift of the Coast Range began in middle Oligocene time in response to renewed compressions (Snavely and others, 1980a). The major uplift faulting, and folding in the northern Oregon Coast Range

did not occur until late middle Miocene (Baldwin, 1976). Regional unconformities in the Coast Range and on the continental shelf reflect fluctuations in the compressional forces produced by the underthrusting (Snavely and others, 1980b). Mafic igneous activity during the late Eocene (Goble Volcanics) and middle Miocene (coastal basalts) used conduits formed by deep tensional rifting in the northern Oregon and Washington Coast Ranges Snavely and others, 1980b).

Recently, paleomagnetic investigations of the Eocene and Miocene basalts in the Oregon Coast Range by Magill and Cox (1981) and by Simpson and Cox (1977) have suggested that the whole Oregon Coast Range has undergone two-phase counterclockwise rotation (75° since late Eocene and 27° since middle Miocene) during this subduction. The pivot point is assumed to lie within the Klamath Mountains. In contrast, the Washington Coast Range has undergone much less counterclockwise rotation (25°) since the Eocene, based on studies of paleomagnetism by Wells and Coe (1979) and by Beck and Burr (1979). The structural discontinutiy between these two Coast Range blocks remains to be found; it may exist in the Nehalem valley or along the Columbia River (Niem, 1981, personal communication).

Local Structure

The structure of the thesis area is relatively simple. It consists of a gently northwestward dipping homoclinal sequence of upper Eocene to middle Miocene sedimentary rocks and basalt flows that is cut by a few northeast-southwest and northwest-southeast high-angle faults (Plate I). Erosion of the basalts has formed the distinctive gently northward dipping dip-slope of the Nicolai Mountain cuesta (Plate III, cross section A-A').

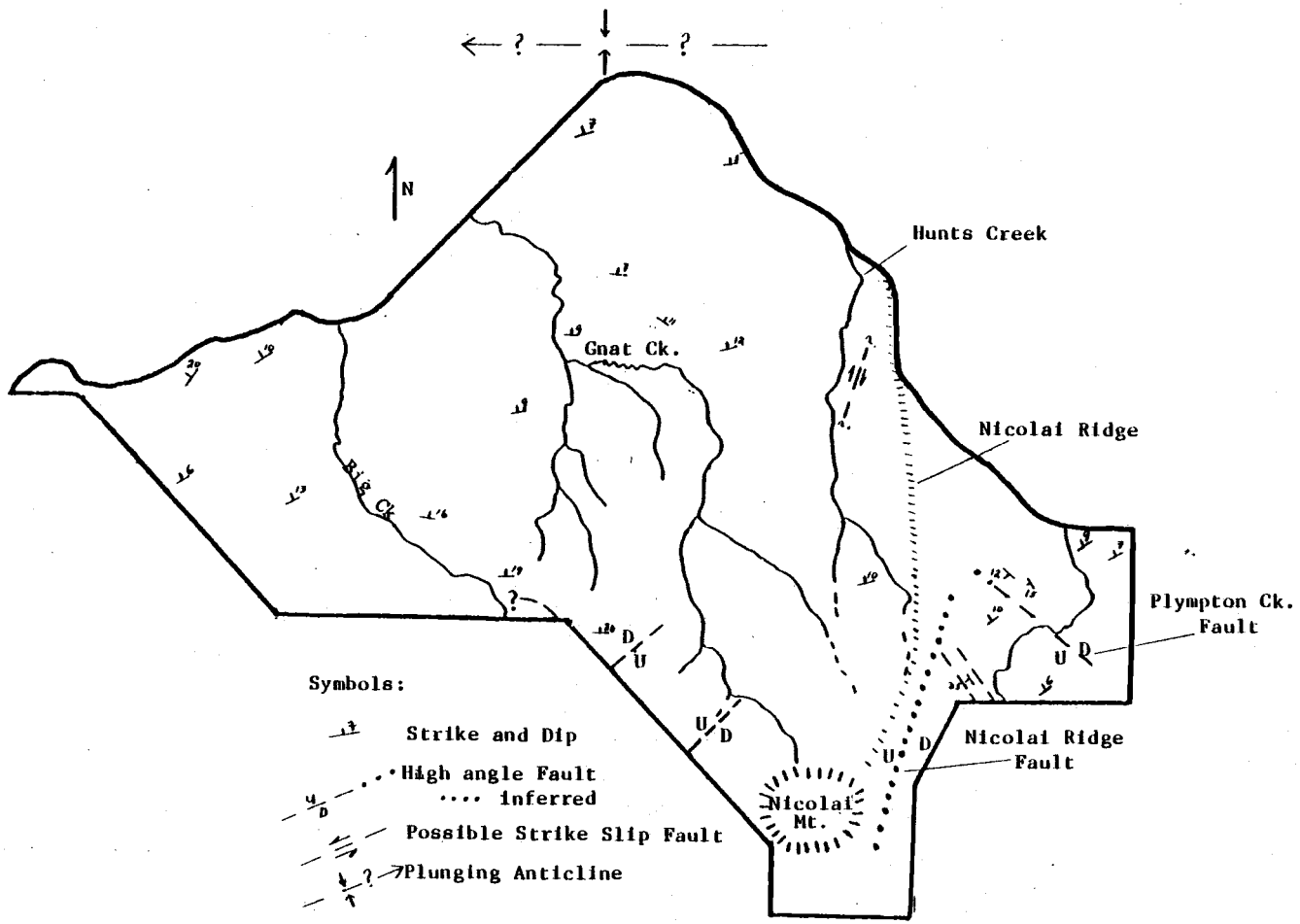


Figure 58 Generalized Structure Map of Study Area
Also see Plate I

The study area is situated near the nose of the northward plunging anticlinorium (Fig. 3). The 10 to 15° northwest regional dip of the Nicolai cuesta conforms well with the Coast Range regional structure forming part of the northwest flank of the anticlinorium (Fig. 58). A resistant middle Miocene basalt cap has preserved this feature (Plate I). From discussions with Ray Wells of U.S.G.S., who is doing geologic mapping in Washington on the north side of the Columbia River, it appears that the Nicolai cuesta also forms the southern limb of an east-west trending syncline, the axis of which is placed down the center of the Columbia River (Fig. 58).

The steep eastern side of the cuesta is called Nicolai Ridge. The southern part of this feature coincides with a northeast-trending high-angle fault (Fig. 58 and Plate I). Existence of this fault is based on the presence of Frenchman Springs Basalt on Plympton and Porter Ridges approximately 300 meters below the same basalt flow units on top of Nicolai Ridge. This indicates at least a 300-meter displacement on the fault. The down-dropped Plympton and Porter Ridges are located 2.0 and 4.0 kilometers respectively east of the 2,648 foot elevation marker on top of Nicolai Mountain (Sec. 10 & 11, T7N, R6W). The trend of the fault is lost in the colluvium cover to the north. On the Washington side of the Columbia River, the linear Elochoman River valley is on strike with the southern half of Nicolai Ridge. This topography may be an expression of the continuation of the Nicolai Ridge fault. Four to ten kilometers to the south, linear middle Miocene dikes trend to the northeast, nearly paralleling the trend of the Nicolai Ridge fault (Wells and Peck, 1961; Beaulieu, 1973). The change in strike of the northern part of Nicolai Ridge to a more northerly direction is probably the result of undercutting by the Columbia River (Beaulieu, 1973) and subsequent slumping and

westward retreat from this initial fault scarp. There are numerous basalt slump and slide blocks preserved that are related to this undercutting process south of Wauna (Fig. 20 and QTls, Plate I).

To the east of Nicolai Ridge, a N55^{o}W trending normal fault (Plympton Creek fault) juxtaposes southwest dipping middle Miocene Astoria Formation (Big Creek member) against northward dipping upper Eocene Pittsburg Bluff Formation (cross-section C-C', Plate III). This suggests a total displacement of at least the thickness of the Oswald West mudstone (125 meters). The topographic expression of the Plympton Creek fault is defined by a sharp offset in the northeast trending strike ridge of Big Creek sandstone to a northwest striking ridge at the same sandstone. The course of Plympton Creek also reflects this abrupt change in strike. The downthrown block dips 15 to 20° to the southwest into the fault plane approximately 90° to the regional northwest strike (Plate I). A fault with the southern block downthrown is shown to the east on the state geologic map of western Oregon (Wells and Peck, 1961). This feature, however, does not correspond to the normal Plympton fault with the northern block downthrown mapped in this study.

Two northwest-trending faults outline a high block of Grande Ronde basalt to the south of Lost Lake (Sec. 2, T7N, R6W, Plate I). Existance of this fault-bound block is based on its dramatic difference from the regional northeast strike and gentle regional northwest dip. It strikes N45^{o}W with a dip of 36° to the southwest (Plate I). Besides the resistant subaerial Grande Ronde basalt flows, the Big Creek sandstone member of the Astoria Formation is exposed to the south side of the fault block (Plate I). The origin of this block may be related to readjustment of the down-dropped block of the larger

northeast trending Nicolai Ridge.

Two northeast-southwest striking high-angle normal faults are postulated to offset the Columbia River Basalts in the southeastern part of Big Creek Gorge (see cross-section B-B', Plate III and Plate I). The existence of these faults is based on the abrupt change in elevation of the subaerial Frenchman Springs flow at the top of Nicolai Mountain (elev. 3010') to its much lower position at 1869 Hill, and the apparent juxtaposition of the Frenchman Springs (Tyfs) flow and Grande Ronde flow (Tygr), in the vicinity.

A slump block or a downthrown fault block exists near the east end of the Big Creek Gorge (SW, Sec. 34, T8N, R7W). The inferred fault is placed in a topographic low between the downthrown slump block (Knob Hill) and the north face of the gorge. Most likely this is a large slump or landslide block of basalt. There are similar slump blocks along the north face of the Wickiup Mountain to the southwest which Coryell (1978) also mapped as inferred fault blocks.

The final fault feature recognized in the study are is a 1-2 meter thick gouge zone in the Frenchman Springs Basalt along the north side of U.S. Highway 30 (NE, Sec. 20, T8N, R7W). There is no apparent vertical displacement because the blocky character of the flow is continuous across the gouge zone. Stike-slip movement is postulated, but no relative motion could be determined on this fault. The strike of the fault gouge zone projects into the linear Hunts Creek. This may be a tensional stress related fracture with little or no movement. Alternatively, the gouge zone may not be fault related. It may be a basalt spiracle that cut through the entire flow. Such features from by flowage of hot lava over wet sediment, resulting in an upward steam blast that fragments

over the basalt. Similar, but smaller, spiracles occur in the Grande Ronde basalt in the measured section along U.S. Highway 30.

Nicolai Mountain cuesta formed during the post late middle Miocene uplift of the Oregon Coast Range. Uplift must have occurred after emplacement of the Pomona Member of the Saddle Mountains Basalt dated at 12.0 $\pm$ 0.5 m.y. (McKee and others, 1977). The northeast-trending Nicolai Ridge may have developed penecontemporaneously with development of the cuesta. The youngest unit truncated by the Plympton Creek fault is the lower Miocene Big Creek sandstone member. The relationship between this fault and the Nicolai Ridge fault is covered by colluvium. It could not be determined due to lack of exposure if the Plympton Creek fault cuts the Columbia River Basalts. The fault block of Grande Ronde basalt south of Lost Lake formed during or after formation of the cuesta. The faults in Big Creek Gorge also formed sometime after or during the uplift of the cuesta.

The long linear deep nature of Big Creek gorge on aerial photos or topographic maps empirically implies structural control. However, geologic mapping in this study shows that it is more likely an entrenched stream formed during the post middle Miocene uplift of the Coast Range. The stream valley is very narrow where Big Creek cuts through the more resistant basalt. The valley widens to the northwest and southeast where the stream cuts through the sedimentary units of the Astoria Formation. Evidence for the lack of structural control is found at the narrowest point in the gorge (OC 131, Sec. 23, T8N, R7W). At that point a dike or filled lava tube can be traced across the gorge and up both sides of the gorge with no relative displacement evident. In addition, a Cape Foulweather sill on the northeast side of the gorge (sample locality 127)

appears to be the same sill exposed on the south side of the gorge at sample locality 714. There is a slight difference in elevation between these two sites, but that can be accounted for by the dip. The apparent sharp differences in rock units between the south side of the gorge (Cape Foulweather) and the north side of the gorge (Grande Ronde and Frenchman Springs) can be attributed to the controversy over the origin of the basalts (coastal versus plateau) and mapping of these units (see section on origin of the basalts). There is no need to infer a fault down the middle of the gorge. Depoe Bay Basalt occurs on both sides of the gorge at the same elevation, taking dip into account.

The state geologic map shows a small syncline in the northern part of the study area (Wells and Peck, 1961). The syncline was placed within the slump-prone Clifton formation on the cuesta. The consistant northwest dips of the in-place homoclineal Clifton formation suggest that the syncline does not exist (Plate I). The axis of the syncline is now postulated to be farther north, approximately along the axis of the Columbia River in this area (Fig. 58) based upon discussions with Ray Wells of the U.S.G.S. He noted that the Columbia River Basalts on the north shore of the Columbia River opposite the town of Clifton dip southward. These same basalts in the Nicolai cuesta dip northward. Ray Wells (1980, personal communication) postulates that the basalts are upturned steeply by a large fault on the northern edge of this syncline.

GEOLOGIC HISTORY

An arc-trench gap marginal basin formed during the middle Eocene on a lower to middle Eocene oceanic tholeiitic basaltic crust (Siletz River and Tillamook volcanics). The basin formed as a result of a westward shift in the locus of underthrusting

of the Farallon plate below the North American plate (Snavely and others, 1980a). The magmatic arc of the arc-trench system was located near the axis of the Oregon Cascades (Dickinson, 1974). The trench and subduction zone were at the base of the present Oregon continental shelf (Niem, 1976; Snavely and others, 1980a).

In the northern Oregon Coast Range, the arc-trench basin began filling soon after its inception with sediment shed off of volcanic arc (Western Cascades) to the east (Snavely and Wagner, 1963). It is postulated that the middle shelf mudstone and siltstone facies of the deltaic and shallow marine Cowlitz and Keasey Formations exposed on the northeast flank of the Coast Range are present in the subsurface of this study area. These units were deposited in a bathyal continental slope environment (Van Atta, 1971). In the Standard "Hoagland" oil exploration well which was drilled near the Lewis and Clark River area, west of this study, company paleontologists noted the tops of the Tillamook Volcanics (2,164m), the Tyee Formation (1,112m), the Cowlitz Formation (525m), and the Keasey Formation (33.07) (Tolson, 1976).

The upper Eocene (Galvinian) Pittsburg Bluff Formation is the oldest unit exposed in the study area. The unit is typically a cross-bedded shallow marine sandstone and pebbly sandstone overlain by thick bioturbated thinly laminated, dirty fine-grained molluscan-rich sandstone. These sedimentary features and fossils are indicative of an open marine, inner to middle continental shelf environment (Kulm and others, 1975; Reineck and Singh, 1975; Walker, 1979). The Pittsburg Bluff represents an early period of shoaling followed by gradual deepening near the end of the late Eocene. Nelson (1978) recognized a similar rapid shoaling in the laterally correlative upper Eocene part of the Oswald West mudstone. A

high sedimentaton rate related to explosive volcanism is suggested by the abundance of unaltered volcanic detritus present in the fine-grained sandstone of the Pittsburg Bluff.

Local upwarping of the Eocene volcanic basement, possible related to middle late Eocene underthrusting (Snavely and others, 1980b), may have resulted in a westward shift of the depocenter from the east. Gradual deepening of the depositional environment in the lower to upper Pittsburg Bluff may also reflect worldwide eustatic rise in sea level during the late Eocene (Vail and Hardenbol, 1979).

The early Oligocene is unknown in this study area but is represented by the bathyal Oswald West mudstone in western parts of the basin (Coryell, 1978; Nelson, 1978; Penroyer, 1977; and others).

The upper Oligocene to lower Miocene (Zemorrian to Saucesian) Oswald West mudstone which conformably(?) overlies the late Eocene Pittsburg Bluff Formation reflects hemipelagic deposition in a low energy, deep marine, middle shelf to upper slope environment. The top of the Oswald West mudstone is defined by a thick (10m) glauconitic sandstone. This represents slow sedimentation or non-deposition in slightly reducing conditions on the middle to outer continental shelf (Pettijohn, 1975). To the east, the Scappoose Formation reflects shallow water, inner shelf to deltaic sedimentation at this time (Van Atta, 1971). The lower contact between the upper Oligocene Oswald West mudstone and the lower Miocene Big Creek sandstone of the Astoria Formation, west of this study area, is marked by a slight angular unconformity produced by marine regression and subsequent subaerial exposure (Coryell, 1978; Nelson, 1978; Penroyer, 1977; and Cooper, 1980). This angular unconformity may have developed with the commencement

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of deposition of the Astoria Formation. This vertical uplift may have been in response to imbricate thrusting of the Farallon plate beneath the North American plate postulated by Kulm and Fowler (1974) and by Snavely and others (1980a). No evidence of subaerial exposure is seen in this study area during this time interval. Regression by seaward progradation of the Scappoose delta and/or local shoaling due to uplift may have resulted in deposition of the shallow marine inner Big Creek sandstone member over deep-marine glauconitic sandstone of the Oswald West mudstone.

The Big Creek sandstone member and overlying Silver Point mudstone member of the Astoria Formation reflect a subsequent marine transgression (onlap) during the early to middle Miocene. The Big Creek member in this area is characterized by inner to middle shelf, fine-grained arkosic sandstones and minor siltstones with rare volcanic-quartzitic conglomerates. The cross-bedded inner shelf sandstones show paleocurrent directions indicating sediment dispersal from east to west. Coryell (1978) described a similar easterly source for the Big Creek member in the Big Creek area.

Further to the southwest, at this time, the fluvial-deltaic-shallow marine Angora Peak sandstone member was being deposited (Cressy, 1974; Smith, 1975; Penoyer, 1977; Cooper, 1980). The majority of the 350m thick Big Creek member exposed in the Young's River-Big Creek type section is interpreted to be a lateral beach, barrier bar, and inner shelf sand facies of the Angora Peak delta by Nelson (1978), Coryell (1978), and Cooper (1980). A similar environment is postulated for the Big Creek member in the Plympton Creek area in this study, based on the presence of shallow water trace fossils and bioturbated cross-bedded sandstone (see depositional environment, Big Creek section).

Concurrent with this Big Creek member transgression is a eustatic sea level rise postulated from early Miocene to middle Miocene (Vail and Hardenbol, 1979).

Continued sedimentary onlap or transgression during the middle Miocene is represented by the upper Silver Point and Pipeline members of the Astoria Formation, exposed in the western part of the thesis area. These units are dominated by quiet water, hemipelagic, laminated micaceous and carbonaceous mudstones. They were depostied in a gradually deepening environment from the middle shelf to upper continental slope based on microfossil paleoecology. To the west, a thick submarine canyon sandstone facies of the Pipeline member cuts the upper Silver Point mudstone (Coryell, 1978; Nelson, 1978). A few of these channelized turbidite arkosic sandstones were deposited in the thesis area. These beds along with the sandstone dikes form the rare fine-grained beds in the mudstone-dominated Pipeline member.

Since the late Oligocene, after deposition of the Scappoose delta system in the eastern part of the Oregon Coast Range (Van Atta, 1971), the major depocenter of this prograding delta shifted to the south and west to form the Angora Peak member in the early to middle Miocene (Cooper, 1980). With continued transgression, slower sedimentation rate, or delta lobe switching and abandonment, the Silver Point outer slope muds covered the abandoned Angora Peak delta (Cooper, 1980). The northeast to southwest orientation of the Pipeline member submarine channel system (Nelson, 1978) indicates the river feeding it had shifted to the north from the Angora Peak delta. The shift may be related to continued downwarping of the Astoria empayment. This study area records only rare thin

overbank, turbidite sandstones from the Pipeline member canyon system (Pipeline mudstone).

During the middle to late Miocene there was a tremendous outpouring of continentally derived tholeiitic basalt in the Pacific Northwest. The majority of these flood basalts were emplaced east of the Cascades, between 14 and 16 m.y. ago, although they continued until 6 m.y. ago (Swanson and others, 1979; McKee and others, 1977). The basalts were extruded from northwest trending dike swarms on the Columbia Plateau, possibly related to backarc extension (Coryell, 1978; Karig, 1971). The volume and rate of extrustion of lava was so great that these flows "spilled" through topographic lows in the ancestral Western Cascades onto the coastal plain and entered the middle Miocene sea in this study area. During the same period, a petrologically similar but less voluminous series of tholeiitic basalts were extruded from a narrow belt of north-south trending, dominantely marine, volcanic centers along the present Oregon and Washington coastline according to Snavely and others, (1973). These centers may reflect a deep extensional rift located along the present Oregon coast (Snavely and others, 1980a).

Middle Miocene basalts unconformably overlie the Astoria Formation in this study area. These basalts include both the locally extruded Depoe Bay and Cape Foulweather basalts and the plateau-derived Columbia River Basalts. The emplacement of the basalts records a period of rapidly changing environments from deep-water, outer shelf to fluvial-shallow marine.

The Depoe Bay Basalt is the oldest basaltic unit in the study area. The unit is characterized by pillow palagonite complexes with "feeder" dikes and rare deep-water mudstone interbeds. Microfossils from the mudstones indicate bathyal

depths (250-1,000m). These mudstones represent the zenith of the marine transgression which began in early Miocene.

A regression of the sea had begun by the time the Grande Ronde Basalt flows were emplaced. The unit consists of basal low MgO closely packed pillow lavas, a low MgO basalt composed of a pillow palagonite-subaerial flow couplet, and two high MgO pillow palagonite-subaerial flow couplets. These basalts flowed down an ancestral Columbia River valley from eruptive fissures in eastern Oregon and Washington and entered a shallow Miocene sea. Shallow marine arkosic sandstone interbeds and a shell coquina between pillows verfiy the shallow marine environment. Two or possibly three lava deltas with pillow foresets and subaerial upper portions developed as these flows entered the water. The thickness of the pillow palagonite complex foreset part of the flow couplet indicates that lava entered water which was approximately 75 meters deep (see Plate II and the section on origin of basalts). The low MgO flows produced local topographic highs around which subsequent high MgO flows were diverted.

The next unit that entered the area was the Cape Foulweather Basalt. This unit intruded bathyal mudstones producing dikes and thick sills. The lava also was exturded from local feeder dikes onto the deep sea floor forming hyaloclastite breccias. These intrustions and breccias are concentrated in the southwest corner of the study area. This area may have been a bathyal shelf basin during a time of general transgression and subsidence based on molluscan and foram fossil paleoecology in the intruded mudstones and interbeds (Coryell, 1978). Alternatively, the dikes and sills may be subsurface continuations of the nearby Frenchman Springs subaerial-pillow palagonite complex unit. In this model the subaerial basalt flows and breccia entered the marine

environment and formed invasive relationships with the soft semiconsolidated deep-water mudstone by burrowing under it. Field evidence more strongly supports the local eruptive model but it is not conclusive.

The Vantage member seperates the Grande Ronde Basalt from the overlying Wanapum Basalt along the western edge of the Columbia Plateau. It represents the longest single interflow period within either of these formations (Mackin, 1961). The Vantage equivalent in this study area is a cross-bedded fluvial to shallow-marine (?) arkosic sandstone, indicating that strandline conditions were present in the central part of the study area. The Frenchman Springs member of the Wanapum Basalt is predominantly a subaerial plateau-derived basalt, which upon entering the sea in the study area produced a pillowed lava delta in the lower part of the flow. Water depths are estimated to have been 75 to 100 meters based on the thickness of the lower pillow palagonite part of the flow. It is the only member of the Wanapum Basalt in this study area. It ends the period of major basalt extrusion and emplacement in the Oregon Coast Range.

In late middle Miocene time (Relizian-Luisian), after emplacement of the Frenchman Springs basalt, an onlap-offlap sedimentary sequence is recorded by the 200-meter thick Clifton formation. Large-scale trough cross-bedded carbonaceous sandstones, point bar sequences, and associated coaly beds represent river mouth deposits (facies 1b) with paleodispersal patterns from east to west. The shallow marine Rosselia bioturbated cross-bedded carbonaceous sandstone (facies 1a) in the western part of the outcrop area interfingers with the fluvial subfacies. This foreshore or offshore bar deposit was produced as river mouth sands were dispersed by longshore and tidal currents acting along a high energy wave dominated

coast. As transgression continued, well-bedded hemipelagic diatom-bearing siltstones (facies 2) were rapidly deposited on the continental shelf. Graded bedding, Bouma sequences, flame structures, and rhythmic alterations of siltstone and sandstone all suggest overbank and interchannel deposition by turbidity currents for this slope facies. Cutting this shelf/slope facies and the underlying fluvial facies are a series of bifurcating submarine canyon head or channel deposits. This canyon head facies is characterized by anastomosing thick channel sandstones with slump blocks (up to 6 meters in diameter) of the shelf/slope facies (Murphy and Niem, 1980). The blocks were formed by submarine slides of the channel walls from undercutting by currents. Amalgamated sand grain flows, chaotic debris flows, siltstone breccias, graded pebbly sandstones and disorganized volcanic conglomerates, represent this canyon head facies (Mutti and Ricci Lucchi, 1972; Walker and Mutti, 1973). The marine onlap reached its zenith of a maximum of 50 to 200 meters (based on diatom assemblages) with continued deposition of the shelf/slope facies. A rapid regression is recorded by a return to fluvial and marsh conditions represented in the upper Clifton formation cross-bedded sandstones and coal beds. Termination of deposition of the unit was by emplacement of the plateau-derived subaerial Pomona member of the Saddle Mountains Basalt in late middle Miocene time.

Figure 59 shows a comparison of the middle Miocene global eustatic cycles of Vail and Hardenbol (1979) based on worldwide seismic stratigraphcy with the same radiometric age and bathymetry as the facies relations in the Clifton formation and the overlying and underlying basalts, based on field relationships discovered in this study. The transgressive curve for the Clifton appears to be twice a large and more extreme than expected from global eustatic changes and may

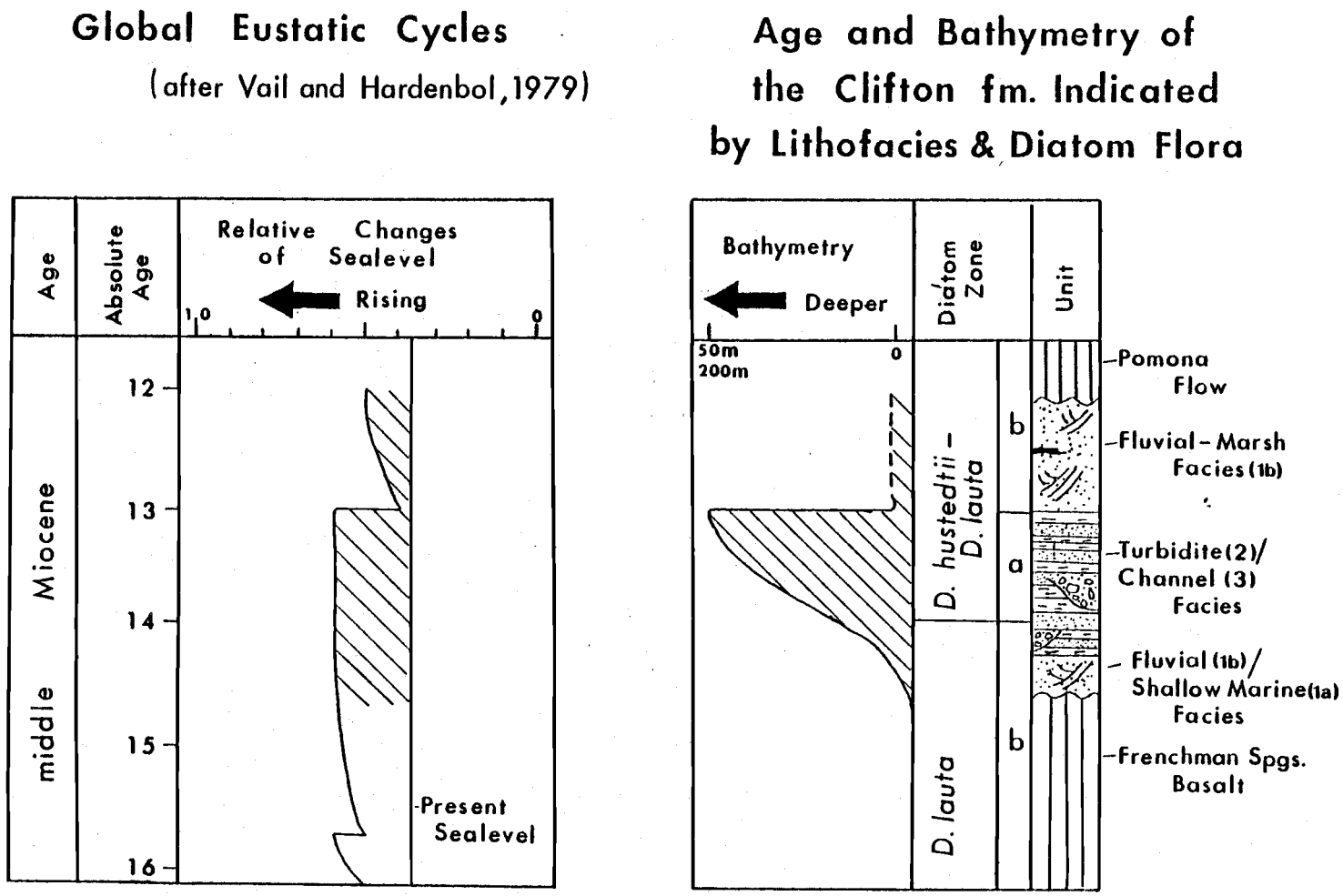


Figure 59 Comparison of global eustatic cycles determined by Vail and Hardenbol (1979) and the bathymetry of the Clifton formation

reflect rapid local tectonic subsidence in the study area. The abrupt regressive phase may parallel the worldwide eustatic offlap during the Miocene recognized by Vail and Hardenbol (1979).

Pebble counts and sandstone petrography indicate that the Pittsburg Bluff, Big Creek and Pipeline members of the Astoria Formation, and the Clifton formation were derived from acid plutonic and metamorphic terrains in eastern Oregon and Washington, Idaho and British Columbia via a Columbia River drainage system. Explosive eruptions in the Western Cascades and eastern Oregon contributed volcanic ash and intermediate to basaltic rock fragments to the sedimentary detritus. Local Coast Range Eocene and middle Miocene basalts as well as older Teritary sandstones also contributed some detritus to these units.

Regional uplift and erosion of the Oregon Coast Range, possibly related to late middle Miocene underthrusting (Snavely and others, 1980a), terminated further preservation of the rock record in the study area. Continued deformation in the late Miocene to Pleistocene produced the Nicolai cuesta and related high-angle faults in the study area. The Plympton Creek fault may be related to a pre-basalt event. Coryell (1978), Nelson (1978) and Penoyer (1977) have also reported pre-middle Miocene basalt faulting in the Astoria embayment.

Formation of the Nicolai Ridge fault changed the course of the Columbia River in the study area, forcing it to the north where, today, it flows in a synclinal axis between Nicolai Ridge and the Washington shore. Subsequent undercutting by the river along the northern part of Nicolai Ridge changed the strike of the ridge and produced the basaltic landslide colluvium present along the base of the ridge. Continued

uplift and local base level adjustments into the Pliocene-Quarternay are indicated by the two levels of river terrace gravels along the river valleys in the study area.

ECONOMIC GEOLOGY

Crushed Rock and Gravel

The extensive subaerial basalt flows and intrusives in the study area (Plate I) provide an abundant source for road gravel, riprap, and jetty material. As of 1973, the economics of transport costs dictated that quarry operations be located within 20 miles of the intended market (Beaulieu, 1973). Therefore, local logging companies and state and county highway departments are the primary markets for the crushed rock. Rock is quarried by blasting; the fragmented blocks are placed into portable jaw crushers to achieve the desired size. The thick columnar jointed subaerial flows and sills are useful for riprap and for jetty construction. The finely fractured (i.e. entablature) subaerial flows and closely packed pillow flows are more easily crushed for road gravel. The submarine palagonitized isolated pillow complexes yield poor grade road metal in terms of its resistance to abrasion and weathering.

Quarries located in the study area are marked on Plate I. Active quarry operations are scattered throughout the study area. The Clatsop County Highway Department is producing road aggregate from a Frenchman Springs subaerial flow along Gnat Creek (OC 430, Sec. 1, T7N, R7W). A permanent quarry located in the landslide deposits (QTls) along the base of Nicolai Ridge was established in 1980 (OC 75, Sec. 27, T7N, R6W). A third large quarry operation is present on the south side of Nicolai Mountain (OC 300, Sec. 21, T7N, R6W). The quarry was terraced into Grande Ronde closely packed pillow units. Well

indurated dikes and sills of Cape Foulweather Basalt in the southwest corner of the study area are presently or have been used for riprap. The pillow palagonite complex units and dikes in Big Creek Gorge are quarried by Boise Cascade Company for logging road gravel. The gentle slope of the Nicolai cuesta has an enormous potential as a source of quarry rock because of the thin overburden and coarsely jointed Frenchman Springs subaerial basalt which forms the cap.

The stream gravel terraces, at the mouth of Big Creek Gorge (OC 122, Sec. 29, R8N, R7W) are actively being quarried and used for road fill and construction. These gravel terraces are prehaps the best source for road construction because they are easily mined (unconsolidated) and have been partly sorted by stream working. A good future supply remains at the mouth of Big Creek (Plate I) for use as sand aggregate.

Cement Production

The Oregon Portland Cement Company of Portland, Oregon is actively mining the arkosic arenite of the Clifton Formation. The quarry is located along U.S. Highway 30 in the fluvial facies (1b) of the Clifton (OC 9, Sec. 19, T8N, R7W). The sandstone is useful because of its low alkali content and high SiO_2, Al_2O_3, and Fe_2O_3 values (Appendix VIII). The silica, alumnium, and iron oxides are mixed with calcium carbonate at a ratio of 1:3 to produce a series of calcium silicates which act as binders in the cement (Harry Winslow, Portland Cement Company). The amount of alkali, reported as percent K_2O and Na_2O, is critical because the aggreate component will react with the alkalies and break down the binding strength of the cement.

Appendix VII lists a series of chemical analyses performed by the Oregon Portland Cement Company on the fluvial sandstone facies from along the Columbia River (OC 177, Sec. 8, T8N, R7W). These sands on Stewart and Les Lahti land, as well as that exposed along Aldrich Point Road, have potential for use in cement production.

Coal

Two 0.3 m thick coal beds were discovered in the Clifton formation. The coal beds are located in Lahti boys land (OC 178, Sec. 8, T8N, R7W) and in a slump scarp at Aldrich Point (OC 585, Sec. 31, T9N, R6W). The beds are traceable for 10 to 25 meters laterally. These coals are probably not of economic quality (i.e. lignitic). The dark coal quickly oxidizes to brown lignite upon prolonged exposure to air. The limited thickness of the coal, the overburden, and the isolated location of these beds severly limit the economic potential of these beds.

Petroleum

The first discovery of commercial quantities of natural gas in the state of Oregon in April, 1979, near the town of Mist has dramatically increased interest in the oil and gas potential of the surrounding area. Mist is approximately 20 km to the southeast of this study area. Since the initial discovery, seven wells have been completed in four separate pools producing a total of over 19 MMcfd of gas (Olmstead, 1979). The reservoir is situated on a northwest-trending highly faulted anticline. The producing unit is from middle to upper Eocene upper Cowlitz Formation at a depth of 925m to 1108m (Newton, 1979). The lower Cowlitz sandstone had salt

water shows, indicating updip production in it may be possible (Newton, 1979).

No gas production has been reported in the Nicolai Mountain-Gnat Creek area, although leasing in Clatsop County has brought as much as $150.00 per acre south of this study area (Newton, 1979). The age equivalent strata of the Cowlitz Formation is postualted to extend from the Mist area into the subsurface of this study. This is suggested by the presence of similar aged siltstones in the Standard Hoagland No. 1 well to the west (see Geologic History section) and the presence of the Pittsburg Bluff Formation in the eastern part of the study area. Potential gas reserves may occur in stratigraphic pinchouts or in structural traps if the Cowlitz sandstones or fractured siltstones are present in the subsurface associated with the normal faults mapped on the surface. Without these stratigraphic or structural traps, any gas generated would have migrated updip to outcrop and would have been lost.

Surface samples of the sedimentary units in the study area were sent to Tenneco Oil Company for reservoir and source rock evaluation. Vitrinite reflectance measurements and visual keorgen analysis were performed on the fine-grained carbonaceous mudstones to determine the organic maturation of the units. Reservoir potenential was determined for the Clifton formation and the Pittsburg Bluff Formation using porosity and permeability tests. The results are listed in Appendix IX. Other sandstone samples from these units and other formations (e.g. Big Creek) were too friable for permeability and porosity analysis.

The samples tested for hydrocarbon maturation contain mainly woody and cuticular organics which suggest a source rock potential for gas rather than oil. The vitrinite reflectance

(V.R.) results indicate that the samples from the Astoria and Clifton Formations contain immature organic material in terms of oil and gas maturation. The thermal alteration index (TAI) shows that the samples are transitional between the immature and mature zones for hydrocabon generation. One mudstone sample from the Oswald West mudstone (Tow-411) displayed a slightly higher source rock potential, with a mature thermal history. The maximum paleotemperature implied by these tests is less than 60° C for all samples but the Oswald West mudstone which has a maximum of 80° C (Kamster and others, 1978). This suggests a maximum burial depth of less than 1.5 km for the Astoria and Clifton units and as much as 2.5 km. for the Oswald West. The pervasive volcanic flows and some dikes in this study area may have locally affected the thermal index of these units, altering the temperature-depth relationship, although the samples analyzed were not collected near flows or intrusions.

These results do not appear encouraging at first glance. They suggest that the units have not been subjected to sufficient maturing temperatures for generation of hydrocarbons. But the methane gas can be generated in immature maturation environments via the bacterial-fungal breakdown of organic matter (Ruffin, 1979). Another factor which can produce poor test results is that rock exposed to weathering (as these samples were) commonly show marked reductions in kergoen yields compared to unweathered subsurface rocks. It should be pointed out that in outcrop the 200 m thick Clifton (particularly the shallow marine and slope facies), Pipeline and Silver Point units contain an abundance of carbonaceous plant material in laminae. Given proper burial, these material could generate methane and other combustible gases.

Porosity and permeability tests indicate that the Clifton and Pittsburg Bluff sandstone units could act as potential gas reservoirs if buried. One sample from the fluvial facies of the Clifton had a permeability of 1.26 facies. Other samples tested below 100 millidarcies, still adequate for storage and passage of gas. The sandstone porosities are high, ranging from 31.8% to 39.7% (Appendix IX).

Like the Big Creek sandstone, both the Pittsburg Bluff and Clifton are well exposed and breached by erosion in the study area; thus, these units have little potential as gas reservoirs in the study area. Burial by younger units and structural traps would be needed for hydrocarbon accumulation in the subsurface. These units project into the subsurface below the Columbia River (see structural section) and are exposed on the Washington shore (Wells, 1980, personal communication). For the Pittsburg Bluff Formation, the overlying upper Oligocene to lower Miocene Oswald West mudstone could act as a cap rock. The Silver Point mudstone could overlie and act as a cap or seal for the Big Creek sandstone in the western part of the study area.

An oil seep has been reported near Olney, Oregon, 16 km west of the Nicolai Mountain-Gnat Creek (Nelson, 1978). Recent organic chemical analysis of the oil revealed it to be unrefined biodegraded crude (Niem, 1978, personal communication) and the chemical analysis reported by Nelson (1978) was incorrect (i.e. wrong samples analyzed by oil company). The proximity of the seep to an old logging company railroad grade and the chemical similarity of the oil to that found in southern California suggests that this crude may have been dumped by earlier logging activity.

The best potential for hydrocarbon production from the Tertiary units described in this study area lies offshore on the continental shelf. Braislin and others (1971) and Snavely and others (1977) indicated that the orgianic-rich siltstones and mudstones may have good source rock potential and that seismic studies indicate that sufficient structural traps exist. The few offshore wells drilled in the 1960's did not penetrate permeable reservoir sandstones. Cooper (1980) in his detailed analysis of the Astoria Formation postulated that this was the result of insufficient facies analysis of onshore geology to determine the location and geometry of the sandstones in the subsuface offshore. He showed with a fence diagram of onshore sections and offshore drillholes of an area of 500 sq. km. that the Astoria Formation between Cannon Beach and Cape Kiwanda has the best potential for reservoir rocks and cap rocks (Angora Peak sandstone and Silver Point mudstone). He suggested that the offshore wells were too far from the coastline due to facies changes to penetrate the thick permeable Angora Peak sandstone or to the Pipeline channel sandstone members of the Astoria Formation.

Braislin and others (1971) delineated an area off the mouth of the Columbia River as containing up to 7,700 meters of Tertiary sedimentary rocks. Nelson (1978) and Coryell (1978) postulated the presence of a middle Miocene submarine fan-canyon complex in this area based on the existence of a submarine canyon facies mapped onshore to the west of this study between Wickiup Mountain and Astoria, Oregon.

This study shows that excellent thick reservoir sandstones exist in the Clifton formation (facies 1a, 1b and 3) and that potential cap and source rocks (for gas) exist in the siltstone/slope facies (2). It is likely that the submarine channelzied sandstone (facies 3) which cuts into and is

overlain by the siltstone-turbidite slope facies (2) projects offshoe (somewhere between Astoria and Cannon Beach). This channel system could be a promising future target for offshore drilling. If the analogy between the Clifton and the Columbia River - upper Astoria canyon head is carried further, one is led to inquire "What became of the Clifton-age sandstone detritus that was funneled down the ancient canyon?". Today, the Astoria canyon feeds the Astoria submarine fan off the Oregon coast (Nelson and Kulm, 1973). If the interpretations of the Clifton facies, the older Astoria Formation, and the analogy with the modern Astoria system are correct, then it can be expected that there is an ancestral Astoria fan or Clifton subsea fan beneath the continental shelf/slope off Oregon. Only future offshore drilling and geophsical study in the 1980's can test this unproven but promising depositional model.

In summary, commercial quantities of gas in sandstones of the upper Eocene Cowlitz Formation, 16 km to the southwest of the study area and similar aged marine siltstones in a dry well to the west suggest that the Cowlitz Formation or age equivalents strata lies in the subsurface of the Nicolai Mountain-Gnat Creek area. The unit may exist at sufficient depth for generation of gas. Fault traps and/or updip facies pinchout of the Cowlitz sandstone to the west may also be present. Permeability, porosity, vitrinite reflectance and the thermal alteration index indicate that the Tertiary strata in the study area could act as source rocks for generation of gas and as reservoir rocks if properly buried and sealed. The potential for the other exposed upper Eocene to middle Miocene strata of this study area is offshore on the continental shelf/slope where a possible submarine fan and canyon complex was deposited during the middle to late Miocene.

CONCLUSIONS

Twelve major sedimentary and volcanic units of late Eocene to middle Miocene age occur in the Nicolai Mountain-Gnat Creek area. They are, in approximate stratigraphic order: the Pittsburg Bluff Formation; the Oswald West mudstone; the Big Creek sandstone; Pipeline mudstone; and upper Silver Point mudstone members (informal) of the Astoria Formation; the Depoe Bay Basalt; the Grande Ronde Basalt Formation; the Cape Foulweather Basalt; a sandstone equivalent to the Vantage Member of the Ellensburg Formation; the Frenchman Springs Member of the Wanapum Basalt Formation; the Clifton formation (informal); and the Pomona Member of the Saddle Mountains Basalt Formation.

The presence of the upper Eocene Pittsburg Bluff Formation stratigraphically below the Oswald West mudstone provides a link between the stratigraphic units of the eastern and western flanks of the northern Oregon Coast Range. Identification of the lower to middle Miocene Big Creek sandstone member east of Nicolai Ridge extends the distribution of the Astoria Formation farther east than previously known. This mapping study also determined the north-eastern outcrop extent of the upper Silver Point mudstone and the Pipeline members of the Astoria Formation in northwestern Oregon.

The detailed stratigraphic relationship between the Columbia River Basalt Group and the coastal basalts has been described for the first time in northwestern Oregon (Fig. 5). Based on the geochemistry and magnetic polarity of the flows, the Depoe Bay and Cape Foulweather are correlative with the low MgO Grande Ronde and Frenchman Springs basalts, respectively. These units are best distinguished by their physical characteristics and environment of emplacement. The presence

of deep-water mudstone interveds, dikes and sills differentiates the Depoe Bay and Cape Foulweather basalts from their respective subaerial correlatives in northwester Oregon, the Grande Ronde and Frenchman Springs Basalts. The Columbia River Basalts are associated with fluvial to shallow marine arkosic sandstone interbeds and are principally subaerial flows with foreset pillow-palagonite complexes.

At least two and possibly three submarine flows are recognized in the Depoe Bay basalt. Thin bathyal, fossiliferous mudstone interbeds are the basis for separation of the individual flows. Two normally polarized high MgO subaerial to ppc flows overlying a low MgO normally polarized subaerial to ppc flow and a reverse low MgO closely packed pillow flow have been differentiated in the Grande Ronde Basalt. Cape Foulweather basalt consists of sparsely porphyritic local sills, dikes, and a hyaloclastite breccia. Individual stratigraphic units or flows are not defined within this basalt. The Frenchman Springs is divided into two subaerial flows based on the presence of a sandstone interbed. The thickest flow contains a foreset pillow-palagonite complex in the western part of the study area.

The Depoe Bay and Cape Foulweather basalts intruded deep-water mudstones forming dikes and sills. The material was also extruded onto the sea floor and formed thick pillow palagonite units, some of which are cut by dikes. The Grande Ronde and Frenchman Springs Basalts flowed into the middle Miocene in this study area. These basalts formed "lava deltas" which prograded into water depths of up to 100 meters. The position of the flows was controlled in part by the topographic expression of the previous basalt units.

A 5-70 meter fluvial to shallow marine (?) sandstone, equivalent to the Vantage Member of the Ellensburg Formation, is present between the Grande Ronde Basalt and Frenchman Springs Member. This units is useful for both local and regional stratigraphic correlation.

The overlying Clifton Formation, previously referred to as the sandstone at Clifton by Niem and Van Atta (1973), is now dated as middle Miocene (Luisian) based on marine diatom assemblages in the siltstone facies and its stratigraphic position between the underlying middle Miocene Frenchman Springs basalt and overlying Pomona basalt. This changes the state geologic map of Wells and Peck (1961), which defined the unit as a late Miocene to Pliocene (?) fluvial sandstone. The facies recognized by field mapping and grain size analysis in this 200-meter thick unit include a basal and upper river mouth - coal marsh facies, a shallow marine offshore bioturbated bar facies, a submarine canyon head facies, and an overbank turbidite sandstone-siltstone facies. The stratigraphic relationship of these facies defines a short-lived marine onlap-offlap sequence betwen the subaerial Frenchman Springs Member and the subaerial Pomona basalt Member related to local subsidence. Sandstone and conglomerate petrography and heavy mineral study suggest that the Clifton formation is a middle Miocene deposit of an ancestral Columbia River - Astoria canyon system. Other sandstone units, such as the Pittsburg Bluff and Big Creek member, also were derived from source areas in eastern Oregon and Washington, Idaho, British Columbia, and the Western Cascades.

The presence of the Pomona Member of the Saddle Mountains Basalt in this study area is the first recognition of this unit in northwest Oregon. Identification is based on a unique

element geochemistry, its reverse remanant polarity, and its stratigraphic position above the Clifton Formation.

Nicolai Ridge, a homoclinal sequence of these Miocene basalts and sedimentary units, was formed by high angle faulting related to broad uplift of the northern Oregon Coast Range anticlinorium and subsequent erosional undercutting by the Columbia River. The river was forced to alter its course northward as a result of formation of the Nicolai fault and associated landslides along the scarp. Big Creek Gorge formed as a result of entrenchment of a stream, during uplift of the Oregon Coast Range in late Miocene to Holcene time.

Permeability, porosity, and maturation studies indicate a strong potential for gas accumulation in the upper Eocene to middle Miocene sandstone and mudstone units exposed in this study area. The best potential lies to the west beneath the continental shelf and slope. The distribution of the canyon head facies of the Clifton formation suggest the possibility of a middle Miocene submarine canyon/fan complex in the offshore area. The Cowlitz Formation or its siltstone lateral equivalent is postulated to be present in the subsurface of this study area. There is a possibility that reservoir sandstones and fault traps exist within this unit.

REFERENCES CITED

Addicott, W. O., Paleontologist, U.S. Geological Survey, written communication, 12-19-79.

Addicott, W. O., 1976, Neogene molluscan stages of Oregon and Washington, in Wornardt, W. W. (ed.), Symposium on the Neogene of the Pacific Coast: Soc. Econ. Paleontologists and Mineralogists, Pacific Section, San Francisco, California, April, 1976, p. 95-115.

Allen, J. R. L., 1963, Henry Clifton Sorby and the sedimentary structures of sand and sandstone in relation to flow conditions: Geol. en Mijnbouw, v. 42, p. 223-228.

Allen, J. R. L., 1965, A review of the origin and characteristics of recent alluvial sediments: Sedimentology, v. 5. p. 89-191.

Allen, J. R., 1977, Physical processes of sedimentation - an introduction; London, George Allan and Unwin, Ltd., Third Printing, p. 248.

American Commission of Stratigraphic Nomenclature, 1961, Code of stratigraphic nomenclature: Amer. Assoc. Petrol. Geol. Bull., v. 45, p. 645-660.

Anderson, J. L., 1978, The stratigraphy and structure of the Columbia River basalt in the Clackamas River drainage: unpublished M.S. thesis at Portland State University, Portland, Oregon, p. 136.

Anderson, J. L., 1980, Pomona Member of the Columbia River Basalt Group: an intercanyon flow in the Columbia River Gorge, Oregon: Oregon Geology, v. 42, no. 12, p. 199.

Anderson, J. L., 1980, written communication, June, 1980.

Armentrout, J. M., 1980, Paleontologist, Mobil Explo. and Prod. Serv., Research Lab, Dallas, Texas, personal communication.

Armentrout, J. M., 1975, Molluscan biostratigraphy of the Lincoln Creek formation, southwest Washington: in Weaver, D. E., ed., Future energy horizons of the Pacific coast; Paleogene symposium and selected technical papers: Annual meeting of the pacific sections Amer. Assoc. Petrol. Geol., Soc. Econ. Paleont. and Mineral., Soc. of Econ. Geol., Long Beach, California, April, 1975, p. 14-48.

Armentrout, J. M., 1978, Progress report of the Cenozoic correlations, west coast North America: Geol. Soc. Amer., Abstract with Programs, v. 10, no. 7, p. 360.

Armentrout J. M., 1980, Cenozoic geologic time table for Oregon and Washington: in Armentrout J. M., ed., Pacific northwest Cenozoic biostratigraphy: Geol. Soc. Amer., Memoir 184, in press.

Armentrout, J. M., McDougall, Kristin, Jefferis, P. T., and Nesbitt, Elizabeth, 1980, Geologic field trip guide for the Cenozoic stratigraphy and Late Eocene paleoecology of southwestern Washington; in Oles, K. F., Johnson, J. G., Niem, A. R., and Niem, W. A., (eds.) Geologic field trips in western Oregon and southwestern Washington; Oregon Dept. of Geology and Min. Ind., Bull. 101, p. 79-102.

Baldwin, E. M., 1976, Geology of Oregon, Revised Edition: Kendall/Hunt, Dubuque, Iowa, p. 147.

Barron, J. A., 1979 and 1980, Micropaleontologist, U.S. Geol. Survey, Menlo Park, Calif., written communications.

Barron, J. A., 1980, Late Cenozoic diatom biostratigraphy and paleooceanography of the middle latitude eastern north Pacific Deep sea drilling project leg 63, in Initial reports of the deep sea drilling project, v. 63; Washington, D.C., U.S. Gov't printing office (in press).

Beaulieu, J. D., 1973, Environmental geology in inland Tillamook and Clatsop Counties, Oregon: Oregon Dept. Geol. and Min. Ind., Bull. 79, p. 65.

Beck, M. E., Jr., and Burr, C. D., 1979, Paleomagnetism and tectonic significance of the Goble Volcanic Series, southwestern Washington: Geology, v. 7, p. 175-179.

Beeson, M. H., Moran, M. R., Anderson, J. L., and Susan Timm, unpublished, Stratigraphy and structure of the Columbia River Basalt Group in Cascade Range, Oregon: Earth Science Dept., Portland State University, p. 87.

Beeson, M. H., Perttu, R. and Perttu, J., 1979, The origin of the Miocene basalts of coastal Oregon and Washington: An alternative hypothesis: Oregon Geology, v. 41, No. 10, p. 159-166.

Beeson, M. H., and Moran, M. R., 1979, Columbia River Basalt Group stratigraphy in western Oregon: Oregon Geology, v. 41, no. 1, p. 11-14.

Bernard, H. A., Major, C. F., Jr., Parrott, B. S., and LeBlanc, R. J., Sr., 1970, Recent sediments of southeast Texas - a field guide to the Brazos River alluvial and deltaic plains and the Galveston barrier island complex - Guidebook #11, U. of Texas, Austin, - Bureau of Economic Geology.

Bingham, J. W., and Grolier, M. J., 1966, The Yakima Basalt and Ellensburg Formation of South-Central Washington: U.S. Geol. Survey Bull. 1224-G, p. 15.

Blatt, H., Middleton, G. V., and Murry, R. C., 1972, Origin of sedimentary rocks: Englewood Cliffs, N. J., Prentice Hall; p. 634; used Allen (1963) diagram, p. 122.

Bouma, A. H., 1962, Sedimentology of some flysch deposits: Elsevier, Amsterdam, p. 168.

Bouma, A. M., 1979, Continental slopes, in Doyle, L. J., and Pilkey, O. M., Jr., (eds.), Geology in continental slopes: SEPM Sp. Pub. #27, p. 1-17.

Bowmann, M. R., Schmincke, M., and Hebert, A., 1974, On the homogeneity of the Columbia River Plateau basalt and its relation to coastal basalt flow, in Nuclear Chemistry Annual Report 1973: Lawrence Berkeley Laboratory LBL2366, p. 381-383.

Braislin, D. B., Hastings, D. D., and Snavely, P. D., Jr., 1971, Petroleum potential of western Oregon and Washington and adjacent continental margin: Am. Assoc. Petrol. Geologists, Mem. 15, p. 229-238.

Bromery, R. W., and Snavely, P. D., Jr., 1964, Geological interpretation of reconnaissance gravity and aeromagnetic surveys in northwestern Oregon: U.S. Geol. Survey Bull. 1181-N, p. 1-53.

Buckovic, W. A., 1979, The Eocene deltaic system of west-central Washington, in Armentrout, J. M., Cole, M. R., TerBest, Harry, Jr., eds., Cenozoic paleogeography of the western United States: Pacific Section, Society of Economic Paleontologists and Mineralogists, p. 147-163.

Busch, D. A., 1974, Stratigraphic traps in sandstones--exploration techniques: AAPG Mem. 21, p. 108.

Carlisle, Donald, 1963, Pillow breccias and their aquagene tufts, Quadra Island, British Columbia: Journal of Geology, v. 71, no. 1, p. 48-71.

Carter, J. W., 1976, The environmental and engineering geology of the Astoria Peninsula area, Clatsop County, Oregon: unpublished M.S. thesis, Oregon State Univ., p. 138.

Carter, R. M., and Linquist, J. K., 1975, Sealers Bay submarine fan complex, Oligocene southern New Zealand: Sedimentology, v. 22, p. 465-483.

Carter, R. M., and Linquist, J. K., 1977, Balleny Group, Chalky Island, southern New Zealand, an inferred Oligocene submarine canyon and fan complex: Pacific Geology, v. 12, p. 1-46.

Chamberlain, C. K., 1978, Recognition of trace fossils in cores: in P. B. Basan (ed.), Trace fossil concepts: Soc. Econ. Paleontologist and Mineralogist, short course, no. 5, Oklahoma City, p. 125-184.

Chamberlain, C. K., 1979, Paleontologist, Cities Services, Denver, Colorado, written communication, October 31, 1979.

Choiniere, S. R., and Swanson, D. A., 1979, Magnetostratigraphy and correlation of Miocene basalts of the northern Oregon coast and Columbia Plateau, southeast, Washington: Amer. Jour. of Sci., v. 279, p. 755-777.

Clifton, M. E., Hunter, R. E., and Phillips, R. L., 1971, Depositional structures and processes in the non-barred high energy nearshore: Jour. Sed. Petrology, v. 41, p. 651-670.

Cook, M. E., 1979, Ancient continental slope sequences and their value in understanding modern slope development: in Doyle, L. J., and Pilkey, O. M., Jr., eds., Geology of continental slopes, Soc. Econ. Paleo. Mineral., Sp. Pub., no. 27, p. 287-306.

Cooper, M. D., 1980, Sedimentation, stratigraphy, and facies variation within the early to middle Miocene Astoria Formation in Oregon; unpublished Ph.D. thesis, Oregon State University, Corvallis, p. 443.

Coryell, G. F., 1978, Stratigraphy, sedimentation, and petrology of the Tertiary rocks in the Bear Creek - Wickiup Mountain - Big Creek area, Clatsop County, Oregon: unpublished M.S. thesis, Oregon State University, Corvallis, p. 178.

Cressy, F. B., Jr., 1974, Stratigraphy and sedimentation of the Neakanie Mountain - Angora Peak area, Tillamook and Clatsop Counties, Oregon: Unpublished M.S. thesis, Oregon State University, Corvallis, p. 148.

Deer, W. A., Howie, R. A., and Zussman, J., 1972, An introduction to rock forming minerals: Longman, London, p. 528.

Dickinson, W. R., 1974, Sedimentation within and beside ancient and modern magmatic arcs: in Dott, R. H., Jr., and Shavers, R. H., eds., Modern and ancient geosynclinal sedimentation: Soc. Econ. Paleon. and Mineral., Sp. Pub. 19. p. 230-239.

Diery, M. D., and McKee, Bates, 1969, Stratigraphy of the Yakima Basalt in the type area: Northwest Sci., v. 43, p. 47-64.

Dodds, R. K., 1969, The age of the "Columbia River Basalts" near Astoria, Oregon; in Proceedings of second Columbia River Basalt Symposium, Cheney, Washington: Cheney, Eastern Washington State College, p. 239-270.

Doell, R. R., and Cox, Allan, 1964, Determination of the magnetic polarity of rock samples in the field: U.S.G.S. Prof. Paper, 450 D, p. 105-108.

Dott, R. H., Jr., 1964, Wacke, graywacke and matrix - What approach to immature sandstone classification?: Jour. Sed. Pet., v. 34, p. 625-632.

Dott, R. H., Jr., 1966, Eocene deltaic sedimentation at Coos Bay, Oregon: Jour. Geology, v. 74, p. 373-420.

Dott, R. H., Jr., and Bird, K. J., 1979, Sand transport through channels across an Eocene shelf and slope in southeast Oregon: in Doyle, L. J., and Pilkey, O. M., Jr., eds., Geology of continental slopes, Soc. Econ. Paleo. and Mineral., Sp. Pub., no. 27, p. 327-343.

Fiske, R. S., Hopson, C. A., and Waters, A. C., 1963, Geology of Mount Rainier National Park, Washington: U.S.G.S. Prof. Paper No. 444.

Folk, R. L., 1951, Stages of textural maturity in sedimentary rocks: J. of Sed. Pet., v. 21, no. 3, p. 127-130.

Folk, R. L., 1974, Petrology of sedimentary rocks: Hemphill Publishing Co., Austin, Texas, p. 182.

Folk, R. L., and Ward, W. C., 1957, Brazos River bar: A study in the significance of grain size parameters: Jour. Sed. Pet., v. 27, p. 3-26.

Friedman, G. M., 1961, Distinction between dune, beach and river sand from their textural characteristics: Jour. of Sed. Pet., v. 31, no. 4, p. 514-529.

Friedman, G. M., 1962, On sorting, sorting coefficients, and the log normality of the grain-size distribution of sandstones: Jour. Geology, v. 70, p. 737-753.

Friedman, G. M., 1967, Dynamic processes and statistical parameters compared for size frequency distribution of beach and river sands: Jour. of Sed. Pet., v. 37, no. 2, p. 327-354.

Fuller, R. E., 1931, The aqueous chilling of basaltic lavas on the Columbia River Plateau: Amer. Jour. of Sci., 5th series, v. 21, no. 124, p. 281-300.

Griggs, A. B., 1976, The Columbia River Basalt Group in the Spokane quadrangle, Washington, Idaho, and Montana: U.S.G.S. Bull. 1413, p. 39.

Hammond, P. E., 1979, A tectonic model for evolution of the Cascade Range: Soc. of Econ. Paleon. and Min., Pacific Coast Paleogeography symposium #3, Cenozoic paleogeography of the western U.S., p. 219-238.

Hammond, P. E., Anderson, J. L., and Manning, K. J., 1980, Guide to the geology of the upper Clackamas and North Santiam River areas, northern Oregon Cascade Range, in Oles, K. F., Johnson, J. G., Niem, A. R., and Niem, W. A., eds., Geologic field trips in western Oregon and southwest Washington, Oregon State Dept. of Geol. and Min. Ind., Bull. 101, p. 133-169.

Harms, J. C. and Fahnstock, R. K., 1965, Stratification, Bed forms, and flow phenomena (with an example from the Rio Grande) in Middleton, G. V., ed., Primary sedimentary structures and their hydrodynamic interpretation: SEPM, Sp. Pub. No. 12, p. 84-115.

Heezen, B. C., Menzies, R. J., Schneider, E. O., Ewing, W. N., and Grannelli, N. C. L., 1976, Congo submarine canyon: Reprinted in Submarine canyons and deep-sea fans, Whitaker, J. M., McD. Dowden, Hutchison & Ross, Penn., p. 43-60 from A.A.P.G. v. 48, p. 1126-1143.

Heinrich, E. W., 1965, Microscopic identification of minerals, McGraw-Hill, p. 414.

Hertlein, L. G., and Crickmay, D. H., 1925, A summary of the nomenclature and stratigraphy of the marine Tertiary of Oregon and Washington: Amer. Philos. Soc. Proc., v. 64, no. 2, p. 224-282.

Hill, D. W., 1974, Chemical composition of Oregon and Washington coastal basalts: Oregon State University M.S. thesis, Corvallis, p. 99.

Howard, J. D., 1972, Trace fossils as criteria for recognizing shorelines in stratigraphic record, in Rigby, J. K., and Hamblin, W. K., eds., Recognition of ancient sedimentary environments: Soc. Econ. Paleon. Mineral., Sp. Pub. 16, p. 215-225.

Howard, J. D., 1975, The sedimentologic significance of trace fossils: in Frey, Robert, ed., The study of trace fossils, Springer-Verlag, N.Y., p. 131-146.

Howe, H. V., 1926, Mid-Tertic type of Pacific Coast: Pan American Geologist, v. 45, p. 295-306.

Howell, D. G., and Link, M. H., 1979, Eocene conglomerate sedimentology and basin analysis, San Diego and Southern California Borderland, J. of Sed. Pet., v. 49, no. 2, p. 517-540.

Inman, D. L., 1952, Measures for describing size of sediments: Jour. of Sed. Pet., v. 22, p. 125-145.

Jacobs, 1970, Clay mineral investigation of Cretaceous and Quaternary deep sea sediments of North American basins: Jour. Sed. Pet., v. 40, p. 864-868.

Jones, J. G., 1968, Pillow lava and pahoehoe: Jour. of Geology, v. 76, p. 485-488.

Jones, J. G., and Nelson, P. H., 1970, The flow of basalt lava from air into water - its structural expression and stratigraphic significance: Geol. Mag., v. 107, no. 1, p. 13-19.

Kamster, A. J., Cook, A. C., and Smith, G. C., 1978, Rank variation, calculated paleotemps in understanding oil, gas occurrence: Oil and Gas Journal, Nov. 20, p. 196-205.

Karig, D. E., 1971, Origin and development of marginal basins in the western Pacific: Jour. Geophys. Res., v. 76, p. 2542-2561.

Kienle, C. F., Jr., 1971, The Yakima Basalt in western Oregon and Washington: unpub. Ph.D. thesis, University of California at Santa Barbara.

Koisumi, I., 1973, The late Cenozoic diatoms of sites 183-193, Leg 19, in Initial reports of the deep sea drilling project, v. 19: Washington, D.C., U.S. Gov't Printing Office, v. 19, p. 805-855.

Krumbein, W. C., and Sloss, L. L., 1963, Stratigraphy and sedimentation (2nd edition): San Francisco, Freeman, p. 660.

Kulm, L. D., and Fowler, G. A., 1974, Oregon continental margin structure and stratigraphy: A test of the imbricate thrust model: in Burk, C. A., and Drake, C. L., eds., The geology of continental margins: Springer-Verlag, New York, p. 261-283.

Kulm, L. D., Roush, R. C., Harlett, J. C., Neudeck, R. M., Chambers, D. M., and Runge, E. J., 1975, Oregon continental shelf sedimentation: Inter-relationships of facies distribution and sedimentary processes; Jour. of Geology, v. 83, no. 2, p. 145-175.

Lohmar, J. M., May, J. A., Boyer, J. E., and Warme, J. E., 1979, Shelf edge deposits of the San Diego embayment, in Abbott, P. L., ed., Eocene depositional systems, San Diego, SEPM Pacific Section, Field Guidebook, Los Angeles, California, p. 15-34.

Long, P. E., Ledgerwood, R. K., Myers, C. W., Reidel, S. P., and Landon, R. D., 1980, Chemical stratigraphy of Grande Ronde Basalt, Pasco Basin, southcentral Washington: Rockwell Hanford Operations, Contract No. DE-AC06-77RL01030; RHO-BWI-SA-32, Richland, Washington.

Lowry, W. D., and Baldwin, E. W., 1952, Late Cenozoic geology of the lower Columbia River valley, Oregon and Washington: Geol. Soc. Amer. Bull., v. 63, p. 1-24.

Mackin, J. H., 1961, A stratigraphic section in the Yakima Basalt and Ellensburg Formation in south-central Washington: Washington Div. Mines and Geology Rept. Inv. 19, p. 45.

MacLeod, N. S., 1980, U.S.G.S., personal communication, May, 1980.

Magill, James and Cox, Allan, 1981, Post-Oligocene tectonic rotation of the Oregon Western Cascade Range and the Klamath Mountains: Geology, v. 9, p. 127-131.

Majewske, O. P., 1969, Recognition of invertebrate fossil fragments in rocks and thin sections: Leiden, The Netherlands, E. J., Brill, p. 101 + 106 plates.

Mancus, D. A., 1972, Bottom topography and sediment texture near the Columbia River mouth, in Pruter, A. T., and Alverson, D. L., eds., Columbia River Estuary and adjacent ocean waters bioenvironments studies: University of Washington Press, p. 241-254.

McKee, E. H., Swanson, D. A., and Wright, T. L., 1977, Duration and volume of Columbia River basalt volcanism, Washington, Oregon and Idaho: Geol. Soc. Amer. Abs. with Programs, v. 9, no. 4, p. 463-464.

Middleton, G. V., and Hampton, M. A., 1973, Sediment gravity flows: mechanics of flow and deposition in Middleton, G. V. and Bouma, A. M., eds., Turbidites and deep water sedimentation, SEPM short course, Anaheim, California, p. 1-38.

Moore, E. J., 1980, Paleontologist, U.S.G.S., written communication, February 13, 1980.

Moore, E. J., 1963, Miocene marine mollusks from the Astoria Formation in Oregon: U.S.G.S. Prof. Paper 419, p. 109.

Moore, E. J., 1976, Oligocene marine mollusks from the Pittsburg Bluff Formation in Oregon, U.S.G.S. Prof. Paper 922, p. 66.

Moore, J. G., 1975, Mechanism of formation of pillow lavas: Amer. Sci., v. 63, no. 3, p. 269-277.

Moore, J. G., Phillips, R. L., Grigg, R. W., Peterson, D. W., and Swanson, D. A., 1973, Flow of lava into the sea, 1969-1971, Kilauea volcano, Hawaii: Geol. Soc. Am. Bull., v. 84, p. 537-546.

Mutti, E., and Ricci Lucci, F., 1972, Turbidites of the northern Apennines: Introduction to facies analysis translated by Nilsen, T. H., in International Geology Review, 1978, v. 20, no. 2, p. 125-166.

Nardin, T. R., Hein, F. J., Gorsline, D. S., and Edwards, B. D., 1979, A Reveiw of mass movement processes, sediment and acoustic characteristics, and contrasts in slope and base-of-slope systems versus canyon-fan-basin floor systems, in Doyle, L. J., and Pilkey, O. M., eds., Geology of Continental slopes: SEPM Sp. Pub. no. 27, p. 7584.

Nathan, Simon, and Fruchter, J. S., 1974, Geochemical and paleomagnetic stratigraphy of the Picture Gorge and Yakima Basalts (Columbia River Group) in central Oregon: Geol. Soc. Amer. Bull., v. 85, p. 63-76.

Neel, R. H., 1976, Geology of the Tillamook Head - Necanicum Junction area, Clatsop County, northwest Oregon: unpub. M.S. thesis, Oregon State University, Corvallis, p. 204.

Nelson, C. M., and Nilsen, T. H., 1974, Depositional trends of modern and ancient deep-sea fans, in Dott, R. H., Jr., and Shaver, R. M., eds., Modern and ancient geosynclinal sedimentation, Soc. Econ. Paleon, Mineral. Sp. Pub. 19, p. 69-91.

Nelson, C. H., and Kulm, L. D., 1973, Submarine fans and channels, in Middleton, G. V., and Bouma, A. H., eds., Turbidites and deep-water sedimentation: Pacific Section, Society of Economic Paleontologists and Mineralogists, Short Course Notes, p. 39-78.

Nelson, M. P., 1977, Tertiary stratigraphy and sedimentation in the Youngs River - Lewis and Clark River area, Clatsop County, northwestern Oregon: unpub. M.S. thesis, Oregon State University, Corvallis, p. 242.

Newton, V. C., Jr., 1979a, Oregon's first gas wells completed: Oregon Geology, v. 41, no. 6, p. 87-90.

Newton, V. C., Jr., 1979b, Subsurface correlation in the Mist area, Columbia County: Oregon Geology, v. 41, no. 7, p. 94-97.

Niem, A. R., 1980, Associate Professor of Geology, Oregon State University, personal communication and written communication.

Niem, A. R., 1976, Tertiary volcaniclastic deltas in an arc-trench gap, Oregon Coast Range (abs.): Geol. Soc. Amer. Abstracts with Programs, v. 8, no. 3, p. 400.

Niem, A. R., and Van Atta, R. D., 1973, Cenozoic stratigraphy of northwestern Oregon and adjacent southwestern Washington, in Beaulieu, J. D., (ed.), Geologic field trips in northern Oregon and southern Washington: Oregon Dept. Geol. and Min. Ind., Bull. 77, p. 75-132.

Niem, W. A. and Glenn, J. L., 1980, Results of heavy-mineral analyses of surface sediments in Tillamook Bay, Oregon: Geol. Soc. America Abstracts with Programs, p. 144-145.

Oles, K. F., and Enlows, H. E., 1971, Bedrock geology of the Mitchell quadrangle, Wheeler County, Oregon: Oregon Dept. Geol. and Mineral Indus., Bull. 72, p. 62.

Olmstead, D. L., 1980, Oil and gas exploration and development in Oregon: Oregon Geology, v. 42, no. 3, p. 47-53.

Passega, R., 1957, Texture as characteristic of clastic deposition, A.A.P.G. Bull., v. 41, no. 9, p. 1952-1984.

Peck, D. L., Griggs, A. B., Schlicker, H. G., Wells, F. G., and Dole, H. M., 1964, Geology of the central and northern parts of the Western Cascade range in Oregon: U.S.G.S. Prof. Paper 449, p. 56.

Penoyer, P. E., 1977, Geology of the Saddle and Humbug Mountain area, Clatsop County, northwestern Oregon: unpub. M.S. thesis, Oregon State University, Corvallis, p. 232.

Pettijohn, F. J., 1975, Sedimentary rocks: Harper and Row Pub., New York, p. 628.

Porrenga, D. M., 1967, Clauconite and chamosite as depth indicators in the marine environment, in Hallcum A., (ed.), Depth indicators in marine sedimentary rocks: Marine Geology, spec. issue, v. 5, p. 495-502.

Potter, P. E., and Pettijohn, F. J., 1977, Paleocurrents and basin analysis; second, corrected and updated editions; Springer-Verlag, New York, p. 425.

Rau, W. W., 1980, Paleontologist, Washington State Department of Natural Resources, Geology and Earth Resources Divisions, written communication, January 11, 1980.

Rau, W. W., 1980, Pacific northwest Tertiary benthonic foraminiferal biostratigraphic framework - an overview: in Armentrout, J. M., ed., Pacific northwest Cenozoic biostratigraphy: Geol. Soc., Amer., Memoir 184, in press.

Reimnitz, E., and Gutierrez-Estrada, M., 1970, Rapid changes in the head of the Rio Balsas submarine canyon system, Mexico: Marine Geology, v. 8, p. 245-258.

Reineck, M. E., and Singh, I. B., 1975, Depositional sedimentary environments: Springer-Verlag, New York, p. 439.

Reison, G. E., 1979, Barrier island systems, in Walker, R. G., (ed.), Facies Models, Geoscience Canada reprint series 1, p. 57-74.

Ricci Lucchi, F., 1975, Depositional cycles in two turbidite formations of northern Apennines (Italy): Jour. Sed. Petrology, v. 45, p. 3-43.

Rock-color Chart Committee, 1970, Rock-color chart: Boulder, Colorado, Geol. Soc. of America.

Royse, C. F., Jr., 1970, An introduction to sediment analysis: Arizona State University, p. 180.

Rubey, W. W., 1933, Size Distribution on heavy minerals within a water lain sandstone, Jour. Sed., Pet., v. 3, p. 3-29.

Ruffin, J. H., 1979, Hydrocarbon source-rock evaluation: Tenneco Oil Co., Research and development, Houston, Texas, unpublished.

Schlicker, H. G., Deacon, R. J., Beaulieu, J. D., and Olcott, G. W., 1972, Environmental geology of the coastal region of Tillamook and Clatsop Counties, Oregon: Oregon Dept. Geol. and Mineral Indus., Bull. 74, p. 164.

Schmincke, H. U., 1964, Petrology, paleocurrent, and stratigraphy of the Ellensburg Formation and interbedded Yakima basalt flows, southcentral Washington: Johns Hopkins University, Baltimore, Ph.D. dissertation, p. 426.

Schmincke, H. U., 1967, Stratigraphy and petrology of four upper Yakima Basalt flows in south-central Washington: Geol. Soc. Amer. Bull., v. 78, p. 1385-1422.

Schenck, M. G., 1929, Stratigraphic relation of western Oregon Oligocene formations: Univ. California Publications in Geological sciences, p. 1-50.

Schrader, Hans, 1980, Micropaleontologist, Oregon State University, personal communication, 2-19-80.

Shepard, F. P., 1954, Nomenclature based on sand-silt-clay ratios: Jour. Sed. Petrology, v. 35, p. 100-108.

Shepard, F. P., and Dill, R. F., 1966, Submarine canyons and other sea valleys: Rand McNally, Chicago, Appendix.

Siems, G. A., Bush, J. M., and Crosby, J. W., III, 1974, TiO_2 and geophysical logging criteria for the Yakima Basalt correlation, Columbia Plateau, Geol. Soc. Amer. Bull., v. 85, p. 1061-1068.

Simpson, R. W., and Cox, A. V., 1977, Paleomagnetic evidence for tectonic rotation of the Oregon Coast Range: Geology, v. 5, p. 585-589.

Smith, T. N., 1975, Stratigraphy and sedimentation of the Onion Peak area, Clatsop County, Oregon: unpublished M.S. thesis, Oregon State Univ., Corvallis, p. 190.

Snavely, P. D., Jr., and Baldwin, E. M., 1948, Siletz River volcanic series, northwestern Oregon: Am. Assoc. Petrol. Geol. Bull., vo. 32, p. 806-812.

Snavely, P. D., Jr., Brown, R. D., Jr., Roberts, A. E., and Rau, W. W., 1958, Geology and coal resources of the Centralia-Chehalis district, Washington: U.S.G.S. Bull, 1053, p. 154.

Snavely, P. D., Jr., and Wagner, H. C., 1963, Tertiary geologic history of western Oregon and Washington: Washington (State) Div. of Mines and Geology, Report of Invest. No. 22, p. 25.

Snavely, P. D., Jr., and Wagner, H. C., 1964, Geologic sketch of northwestern Oregon: U.S.G.S. Bull. 1181-M, p. 17.

Snavely, P. D., Jr., Wagner, H. C., and MacLeod, N. S., 1965, Preliminary data on compositional variations of Tertiary volcanic rocks in the central part of the Oregon Coast Range: The Ore Bin (now Oregon Geology), v. 27, no. 6, p. 101-120.

Snavely, P. D., Jr., MacLeod, N. S., and Wagner, H. C., 1973, Miocene tholeitic basalts of coastal Oregon and Washington and their relations to coeval basalts of the Columbia Plateau: Geol. Soc. of Amer., v. 84, p. 387-424.

Snavely, P. D., Jr., Pearl, J. E., and Lander, D. L., 1977, Interim report on petroleum resource potential and geologic hazards in the outer continental shelf-Oregon and Washington Tertiary province: U.S. Geol. Survey. open file report 77-282, p. 64.

Snavely, P. D., Jr., MacLeod, N. S., Wagner, H. C., and Lander, D. L., 1980a, Geology of the west-central part of the Oregon Coast Range, in Oles, K. F., Johnson, J. G., Niem, A. R., and Niem, W. A., (eds.), Geologic field trips in western Oregon and southwestern Washington, Oregon Dept. of Geol. and Mineral Industries, Bull. 101, p. 39-76.

Snavely, P. D., Jr., Wagner, H. C., and Lander, D. L., 1980b, Interpretation of the Cenozoic geologic history, central Oregon continental margin: Cross-section summary: Geol. Soc. of Amer. Bull., part I, v. 91, p. 143-146.

Steiger, R. H. and Jager, E., 1977, Subcommission of geochronology: Convention on the use of decay constants in geo- and cosmochronology: Earth and Planetary Sci. Letters, v. 36, p. 359-363.

Stow, D. V., 1978, Deep sea, fine-grained sedimentation - a discussion paper; Penrose conference on submarine slope, fan and trench sedimentation-new concepts and problem solving, unpublished, p. 16.

Swanson, D. A., 1980, U.S.G.S., Menlo Park, personal communication, 2-80.

Swanson, D. A., 1967, Yakima Basalt of the Treton River area, south-central Washington: Geol. Soc. Amer. Bull., v. 78, p. 1077-1110.

Swanson, D. A., and Wright, T. L., 1978, Bedrock geology of the northern Columbia Plateau and adjacent areas: in Baker, V. R., and Nummedal, Dag, ed., The Channeled Scabland: [N.A.S.A., Planetary Geology Program,], p. 186.

Swanson, D. A., and Wright, T. L., 1979, preprint, A regional approach to studying the Columbia River Basalt Group, submitted to Geol. Soc. of India, November, 1979.

Swanson, D. A., Wright, T. L., Hooper, P. R., and Bentley, R. P., 1979, Revisions in stratigraphic nomenclature of the Columbia River Basalt Group: U.S. Geol. Survey Bull. 1457-G, p. 59.

Swift, D. J., 1976, Continental shelf sedimentaiton, in Stanley, D. J., and Swift, D. J., (eds.), Sediment transport and environmental management: John Wiley and Sons, p. 311-350.

Taubeneck, W. H., 1969, Dikes of Columbia River basalt in northeastern Oregon, western Idaho, and southeastern Washington, in Gilmour, E. M., and Stradling, Dale, eds., Proceedings Second Columbia River Basalt Symposium: Cheney, Eastern Washington State Coll. Press, p. 73-96.

Taylor, E. M., 1979, Associate Professor, Oregon State University, Corvallis, personal communication, 3-80.

Tolson, P. M., 1976, Geology of the Seaside-Young's River Falls Areas, Clatsop County, Oregon: unpub. M.S. thesis, Oregon State University, Corvallis, p. 191.

Trask, P. D., 1932, Origin and environment of source sediment of petroleum, Gulf Publ. Co., Houston, p. 323.

Tucker, R. W., and Vocker, M. L., 1980, Effectiveness of discriminating beach, dune, and river sands by movements and cumulative weight percentages: Jour. Sed. Pet., v. 50, no. 1, p. 165-172.

Turner, F. E., 1938, Stratigraphy and mollusca of the Eocene of western Oregon: Geol. Soc. Amer. sp. paper 10, p. 130.

Vail, P. R., and Hardenbol, Jan, 1979, Sea level changes during the Tertiary: Oceanus, v. 22, no. 3, p. 71-74.

Van Atta, 1971, Sedimentary petrology of some Tertiary formations, upper Nehalem River basin, Oregon: Ph.D. thesis, Oregon State University, Corvallis, p. 245.

Visher, G. S., 1965, Fluvial processes as interpreted from ancient and recent fluvial deposits, in Rigby, J. K. and Hamblin, W. K., eds., Recognition of ancient sedimentary environments: SEPM Sp. Pub., No. 16, p. 116-132.

Walker, R. G., 1979, Shallow marine sands, in Walker, R. G., ed., Facies Models, Geoscience Canada reprint series 1, p. 75-90.

Walker, R. G., and Cant, P. J., 1979, Sandy fluvial systems, in Walker, R. G., (ed.), Facies Models, Geoscience Canada reprint series 1, p. 23-32.

Walker, R. G., and Mutti, E., 1973, Turbidite facies and facies associations in Middleton, G. V., and Bouma, A. M., eds., Turbidites and deep water sedimentation, SEPM short course, Anaheim, p. 119-158.

Warren, W. C., Norbisrath, Hans, and Grivetti, R. M., 1945, Geology of northwestern Oregon west of the Willamette River and north of Latitude 45°15': U.S.G.S. Oil and Gas Invest., Prelim. Map 42.

Warren, W. C., and Norbisrath, Hans, 1946, Stratigraphy of upper Nehalem River Basin, northwest Oregon: Am. Assoc. Pet. Geol. Bull., v. 30, no. 2, p. 213-237.

Washburne, C. W., 1914, Reconnaissance of the geology and oil prospects of northwestern Oregon: U.S.G.S. Bull. 590, p. 111.

Waters, A. C., 1960, Determining direction of flow in basalts: Am. Jour. of Sci., Brakley Volume, v. 258-A, p. 350-366.

Waters, A. C., 1961, Stratigraphic and lithologic variations in the Columbia River basalt: Am. Jour. Sci., v. 259, p. 583-611.

Watkins, N. D., and Baksi, A. K., 1974, Magnetostratigraphy and oroclinal folding of the Columbia River, Steens, and Owyhee basalts in Oregon, Washington and Idaho: Am. Jour. Sci., v. 274, p. 148-189.

Weaver, C. E., 1937, Tertiary stratigraphy of western Washington and northwestern Oregon: Washington University Pub. in Geology, v. 4.

Weaver, C. E. and others, 1944, Correlation of the marine Cenozoic formations of western North America: Geol. Soc. Amer. Bull., v. 55, p. 569-598.

Wells, F. G., and Peck, D. L., 1961, Geologic map of Oregon west of the 121st meridian: U.S.G.S. Invest. Map I-325, scale 1:500,000.

Wells, Ray, 1980, U.S.G.S., personal communication, 3-80.

Wells, R. E., 1979, Geologic map of the Cape Disappointment-Naselle River area, Pacific County, Washington: U.S. Geol. Survey open-file report 79-389, scale 1:48,000.

Wells, R. E. and Coe, R. S., 1979, Paleomagnetism and tectonic significance of the Eocene Crescent Formation, southwestern Washington: Geol. Soc. America Abstracts with Programs, v. 11, p. 537-538,

Whetten, J. T., and Fuller, T. J., 1967, Columbia River bed forms: International Assoc. for hydraulic research contribution #426 of the Department of Oceanography, University of Washington.

Whetten, J. T., Kelley, J. C., and Manson, L., 1968, Characteristics of Columbia River sediment and sediment transport: Jour. Sed. Pet., v. 39, no. 3, p. 1149-1166.

Williams, H., Turner, F. J., and Gilbert, C. M., 1954, Petrography, an introduction to the study of rocks in thin section: Freeman & Co., San Francisco, p. 406.

Williams, Howel, and McBirney, A. R., 1979, Volcanology: Freeman, Cooper, & Co., San Francisco, p. 397.

Wolfe, E. W., and McKee, E. M., 1972, Sedimentary and igneous rocks of the Grays River quadrangle, Washington: U.S.G.S. Bull. 1335, p. 70.

Wright, T. L., Grolier, M. J., and Swanson, D. A., 1973, Chemical variation related to the stratigraphy of the Columbia River Basalt: Geol. Soc. of Amer. Bull., v. 84, p. 371-386.

APPENDICES

Appendix I

Reference section a-a' in the Pittsburg Bluff Formation

Initial Point: East of Westport, along old U.S. Highway 30, at entrance of 10 meter high (OC49) in the NW, NW corner of section 1, T7N, R6W.

Terminal Point: 10 feet past west side of bridge (OC-50) in the SW, SE corner of section 36, T8N, R6W. along old U.S. Highway 30

Unit	Description	Unit Thickness	Total (Meters)
5	Silty Sandstone: light gray (N7) weathers to a pale yellow orange (10 YR 6/4); very fine-to fine-grained; thick to thinly bedded; tuffaceous, micaceous, quartzwacke; highly bioturbated; molluscs fossil molds and Sclarituba trace fossil present; blocky weathering pattern. Mollusk fossil collection no. OC-50 collected at 24.4m. contact: gradational over 0.6 meters	8.1	27.4
4	Mudstone: light gray (N7) alternating with very light gray (N8); both planar and cross very thin laminations observed; tuffaceous with fine mica; small thin molluscs molds; Scalarituba burrows; chippy weathering character. contact: sharp; planar possibly a disconformity.	3.5	19.8
3	Sandstone: reddish brown (10 YR 4/6) to light brown (5 YR 5/6); very coarse grained with a fine grained matrix; subangular to subrounded; structureless; compositionally a quartzose volcanic arenite with hematite cement; rare molluscan fossil molds ; very friable. contact: gradational; bioturbation causes mixing.	1.7	15.8
2	Sandstone: similar to unit 1 but with a very coarse grained component due to mixing of the upper unit 3 by bioturbation; homogenization of the two grain sizes occurs along with discrete circular to oval pods (filled burrows) of very coarse-grained material up to 2cm in diameter within the very fine-grained sandstone; moderately indurated. contact: gradational	2.5	14.1

1	Sandstone: bluish gray (5B 7/1) weathers to a greenish gray (5 Gy 6/1); very fine-grained; subangular; appears structureless but shows faint planar cross laminations; composed principally of quartzose, feldspathic, volcanic arenite with rare very coarse tuffaceous siltstone fragments and lenses of disarticulated molluscan shell fragments; moderately indurated; weathers in flakes creating a smooth surface; bluff former; calcareous concretionary layer present from 8.7 to 9.8 meters. Mollusk fossil collection no. OC-49 at 4.5 meters.	11.6	11.6

Reference Section b to b' along Tripp Road

Contact Between Pipeline Mudstone and Clifton Formation

Initial Point: OC216, SW, SW, sec 19, T8N, R7W; 26 meters north of mile 1 marker at bend on Tripp Road. Section parallels north on Tripp Road.

Terminal Point: OC215; same section; 132 meters south of first driveway on west side of Tripp Road

Unit	Description	Thickness (Meters) Unit	Total
8	Clifton formation, facies 1a: same as unit 7; but mottled medium light gray (N6) to light gray (N7) with moderate reddish orange (10YR 6/6), mottling character commonly follows the burrow structures; highly bioturbated due to increased concentration of Rosselia burrows; sandstone is very firm, cliff former. contact: gradational	11.7	102.4
7	Clifton formation, facies 1a: similar to unit 6; light gray (N7) with Fe staining; fine-grained; arkosic, micacous, carbonaceous; a few Rosselia burrows; medium thick bedding (0.2 to 0.3m); concretions up to 1m in diameter with foram tests and mollusk shells preserved within them. contact: gradational	5.2	90.7
6	Clifton formation, facies 1a; light brown (5 YR 5/6), very weathered; medium-to fine-grained clayey; structureless; arkosic; poorly exposed contact: covered	17.5	85.5

5	Covered - not exposed in readcut	50.0	68.0
4	Pipeline mudstone: sandstone; same as unit 2; greater amount of Mica and carbonaceous material present; same laminations contact: covered	4.4	18.0
3	Pipeline mudstone; mudstone; same as unit 1; a moderate yellowish brown (10YR 5/4); weathered with characteristic "chippy" talus contact: gradational with unit 4 over 0.2m	6.0	13.6
2	Pipeline mudstone: sandstone grayish orange (10YR 7/4); fine to very fine-grained; very friable silty structureless; arkosic. contact: sharp planar	0.8	7.6
1	Pipeline mudstone: mudstone; grayish black (N2) to grayish brown (5 YR 3/2) weathers to a moderate yellow brown (10 YR 5/4); clayey siltstone;micaceous; highly carbonaceous, laminated; thin shelled pelecypod shells and Foraminifera tests contact: covered with Quaternary river gravels	6.8	6.8

Appendix I

Reference Section c-c' of Clifton Formation (facies 2 and 3) along Clifton Road

Initial Point: OC 17; NW, NW, Sec 9, T8N, R6W; Along Clifton Road at turn in road; Hunts Creek Waterfalls is off the east side of road.

Terminal Point: OC24, NE, NE, Sec 8, T8N, R6W; 2 meters south of sign for Bradwood road cutoff.

Unit	Description	Thickness (Meters) Unit	Total
11	Mudstone/sandstone: facies 2; ratio of mudstone to sandstone is 1:1 ; a few medium-grained sandstone channels (10 m cross-section) into alternating ss/mudst. units; few mudstone chips present in channels; channels are overlain by alternating layers of graded fine-grained sandstone and siltstone. contact: gradational -	9.8	48.3
10	Mudstone/sandstone: facies 2; same as unit 8; ratio of mudst/ss is 70:30; sandstone lamine become thicker (up to 5 cm) and more numerous toward the top; sandstone is same as unit 2 matrix but finer grained; mud-stone same as mudstone blocks contact: gradational	8.9	38.5
9	Mudstone/sandstone: same as 8 but slumped contact: sharp;covered with vegetation and talus.	1.8	29.6
8	Mudstone/sandstone: facies 2; mudstone is a clayey siltstone to siltstone, same as mudstone blocks in unit 3, dark gray (N3) to grayish brown (5 3/2); sand-stone is fine-grained, present as lamina-tions to thin beds in mudstone; ratio of mudst/ss is 80:20; good cliff former; chippy weathering character; breaks along conc. of mica flakes contact: gradational but covered with vegetation	5.8	27.8

Appendix I

Section c-c'

7	Mudstone/sandstone: very poorly exposed, covered with vegetation; undeterminable if large blocks or in place layers, suspect in place. contact: sharp; irregular	3.4	22.0
6	Sandstone: facies 3; sandstone matrix similar to unit 4; thickly bedded with layers defined by the abundance of mudstone/sandstone slump blocks incorporated into channels; percent blocks range from 10 to 95% of a layer contact: gradational with lower unit; sharp with facies 2	6.3	18.6
5	Sandstone: facies 3; similar to unit 2; medium-to fine-grained; thick-bedded; few mudstone blocks (10%). contact: sharp; irregular	1.5	12.3
4	Pebble conglomerate: facies 3; same as unit 1 contact: erosive; irregular concave upward.	0.6	10.8
3	Sedimentary breccia: facies 3; series of anastomosing channels composed of mudstone blocks within a pebbly sandstone matrix; pebbly sandstone is similar to unit 1 and 2, percentage of pebbles varies from 10-60%; mudstone blocks are up to 5m, grayish red (10R 4/2) to brownish gray (5 YR 4/1), have very fine-to fine-grained sandstone laminations with mica partings; carbonaceous; most blocks are subangular; mudstone/sandstone derived from facies 2; concentrations of mudstone blocks varies from 40 to 90%, same blocks show compaction features of pebbles pushed into mudstone layer; other structures include: Normally graded beds, Bouma Tae, Tade, Tcde, intervals, and flame structures. contact: erosive; concave upward	7.7	10.2
2	Sandstone: facies 3; light gray (N7) weathers to a dark yellowish range (10 YR 6/6); coarse to fine-	1.1	2.5

	grained mostly medium-grained; very friable; Fe-oxide only cement; thick-bedded with some thin carbonaceous mudstone layers; sandstone composed of quartz, feldspar, mica and volcanic lithics contact: sharp, erosive		
1	Pebble conglomerate: facies 3; light brown (5 YR 6/4) weathers to a grayish orange (10 YR 7/4); poorly sorted, possible hint of normal grading; lens-shaped unit; medium-grained sandstone matrix; very friable; subrounded pebbles composed of basalt, andesite, dacite and quartzite; few mudstone chips contact: covered with vegetation	1.4	1.4

APPENDIX T

Reference Section d - d' at OC 553

Contact Between Facies 3 and Facies 1 of Clifton Formation

Initial Point: West edge of Abu. Aldrich Point Road. Road cut (OC 553) in NW, SE, Sec 10, T8N, R7W

Terminal Point: North of Initial Point at top of gully, where the soil covers the outcrop.

UNIT	DESCRIPTION	UNIT Thickness	TOTAL (meters)
5	Conglomerate: grayish orange (10 YR 7/4) to very pale orange (10 YR 8/2); coarse-to fine-grained. Very poorly sorted sandstone matrix; cobble size mudstone clasts up to 0.6 m long. Orientation of chips may suggest small and anastomosing channels (facies 3) contact: erosive, irregular into lower sandstone facies 1.	2.6	20.5
4	Sandstone: same as unit 3; may be slightly finer grained; laminated to thinly laminated contact: gradational from unit 3.	7.2	17.9
3	Sandstone: Medium gray (N5), weathers to a grayish orange (10 YR 7/4); fine-grained, arkosic; carbonaceous and mica laminations; few mudstone layers similar to that in unit 2.	4.1	10.7
2	Sandstone: light gray (N7); medium-to fine-grained; clean, very friable, subangular, arkosic sandstone; mafic mineral laminations; a single 1-2cm thick light brownish gray (5 YR 6/1); medium scale trough crossbeds in upper part (facies 1) Contact: sharp; bedding plane with coarse sandstone (unit 1) below and finer-grained sandstone (unit 2) above.	2.8	6.6
1	Sandstone: very light gray (N8); medium to coarse-grained; subangular to subrounded; clean, very friable, arkosic sandstone, thickly bedded. Contact: basal, not exposed along cut. (facies 1)	3.8	3.8

Appendix I

Bradley State Park Reference Section e - e'
for Columbia River Basalts

Initial Point (b): 1150 eet west of east line, 800 feet north of south line in Section 21, T8N, R6W. Section is along new U.S. Highway 30 0.8 mile east of Bradley State Park, 750 feet east of sign reading "Yield center lane to uphill traffic" (Near BM 383). Section proceeds along U.S. Highway 30 up hill to west.

Terminal Point (b): 1100 feet east of west line, on north line in Section 21,T8N, R6W. End point along U.S. Highway 30 at entrance to Bradley State Park wayside.

Unit	Description	thickness (meters)	total (meters)
9	Clifton formation: conglomerate sandstone with mudstone ripups; grayish orange (10 YR7/4); friable; sandstone is coarse-to fine-grained with pebbles of andesite, quartzite and basalt; poorly exposed along road. contact: irregular, unconformity over oxidized zone of basalt.	3.4	235
8	Frenchman Springs Member of Wanapum Basalt (Tyfs): normal polarity; subaerial flow; finely crystalline groundmass with 1-2 cm long plagioclase phenocrysts; jointing is primarly irregular, hackly and blocky with poorly developed colonnade joints at the base and top; vesicles increase toward the top; basalt breaks along unseen weathering horizons easily when hit with hammer. Contact: irregular - basalt splays into underlying sandstone.	71.1	231.6

Appendix I

Unit	Description	thickness (meters)	total (meters)
7	Sandstone: equivalent to Vantage member of Ellensburg Formation; dark yellowish orange (10YR 6/6); medium to fine grained; carbonaceous lenses up to 5cm thick; arkosic with a basaltic sandstone base; Fe stained; friable nature produces negative profile; covered with vegetation along road therefore thickness is an estimate. contact: sharp; erosional.	81.5 (?)	160.5
6	Grande Ronde Basalt ($Tygr_3$): High MgO; normal polarity; subaerial flow; aphanitic to finely crystalline; well developed jointing patterns of lower colonade, straight and wavy entablature and upper colonnade, entablature comprise 70% of unit; vesicles increase upward Pipe vesicles with westward orientations and a possible spiracle found along base contact: sharp, undulatory	23.4	79.0
5	Sandstone: grayish olive (10Y 4/2); medium to fine grained: carbonaceous; quartz, mica, basalt grains abundant contact: sharp, erosional over basalt flow.	2.1	55.6
4	Grande Ronde Basalt ($Tygr_3$): same as unit 2; increase in vesicles which are concentrated in circular regions (1m in diameter); more massive aphanitic basalt, less glassy hyaloclastite fragments; poorly developed entablature which grade upward into colonnade joint sets. contact: covered.	17.2	53.
3	Grande Ronde Basalt ($Tygr_3$); same as unit 2; breccia clasts larger, angular, more vesicular; area mostly covered with vegetation. contact: sharp, erosional.	12.3	36.3

Appendix I

Unit	Description	Thickness (meters)	Total (Meters)
2	Grande Ronde Basalt ($Tygr_3$): High MgO; normal polarity; PPC unit: nonvesicular hyaloclastite fragments with pillows and irregular shaped, massive aphanitic features which may represent the collection and movement of fluid lava within a solid-liquid basalt pile; material becomes finer grained with less irregular features upward; very poor colonnade joints at base of unit. contact: undulatory over oxidized erosional surface.	20.0	24.0
1	Grande Ronde Basalt ($Tygr_2$): Low MgO; normal polarity; subaerial basalt; aphanitic groundmass with microphenocrysts of plagioclase, very vesicular; poorly developed colonnade joints; top of flow with red oxidized zone which has been baked white by overlying flow; zeolite filled vesicles; calcite replacing plagioclase.		

APPENDIX II

MAJOR OXIDE VALUES FOR BASALTS IN NICOLAI MOUNTAIN-GNAT CREEK AREA

Plot No.	FeO	TiO_2	CaO	K_2O	SiO_2	Al_2O_3	NaO_2	MgO	
22	11.2	2.1	6.9	1.75	56.8	14.8	3.1	3.7	Depoe Bay
23	11.7	2.05	6.9	1.75	56.4	14.8	3.1	4.0	
24	10.9	2.10	6.9	1.85	56.4	14.4	3.0	4.0	
25	11.4	2.15	6.8	1.9	56.1	14.1	3.0	3.8	
27	12.1	2.05	6.9	1.45	56.4	14.2	3.1	3.9	
35	11.3	1.9	6.8	1.7	56.1	14.2	3.2	3.7	
1	11.6	2.0	6.9	1.85	55.7	13.8	3.0	4.0	Grande Ronde Low MgO
13	12.4	2.4	6.9	1.85	54.9	14.2	3.0	3.4	
14	11.6	2.4	6.8	2.15	55.5	14.4	2.9	3.7	
17	11.8	2.1	6.6	1.85	56.7	14.3	3.1	3.8	
20	11.9	1.85	7.0	1.9	56.4	13.4	2.9	4.3	
21	11.3	1.85	6.8	1.8	56.5	14.5	3.0	4.0	
26	11.8	2.45	6.6	2.05	55.7	14.4	3.1	3.5	
2	10.8	2.05	8.7	1.25	54.1	14.7	2.8	4.6	Grande Ronde High MgO
3	11.2	2.0	9.1	1.10	54.9	14.6	2.9	5.0	
4	11.2	2.05	8.9	1.05	53.9	14.8	3.0	4.4	
5	10.9	1.85	8.1	1.45	54.9	14.9	2.8	4.3	
6	11.3	1.85	8.3	1.25	54.8	14.7	2.8	4.7	
12	12	2.05	8.4	1.10	53.7	14.6	2.8	4.6	
36	11.4	1.8	8.2	1.3	54.8	14.8	2.8	4.7	
41	11.7	1.9	8.95	1.0	54.1	14.2	2.9	5.2	
42	11.2	1.9	8.5	1.5	55.0	14.5	2.8	4.6	
43	12.1	2.0	8.5	1.13	54.2	14.3	3.0	4.9	
46	11.6	1.9	8.6	1.1	54.7	14.5	2.7	4.9	

APPENDIX II

Page 2

Plot No.	FeO	TiO_2	CaO	K_2O	SiO_2	Al_2O_3	Na_2O	MgO	
7	13.5	3.20	8.1	1.35	51.2	13.3	2.6	4.5	Frenchman Springs
8	13.7	3.15	7.9	1.35	51.4	13.4	2.7	4.8	
9	13.6	3.05	8.4	1.25	50.5	13.7	2.7	4.5	
10	13.3	2.95	8.3	1.15	51.5	13.9	2.8	4.3	
11	13.2	2.95	8.2	1.15	52.0	13.9	2.9	4.3	
15	13.5	3.05	8.1	1.40	51.5	13.7	2.6	4.6	
16	13.5	2.95	8.1	1.40	51.7	13.3	2.7	4.6	
18	13.5	2.95	8.1	1.45	52.6	13.4	2.7	4.5	
19	13.3	3.00	8.4	1.50	52.5	13.5	2.7	3.9	
39	14.1	3.00	8.2	1.40	52.0	13.8	2.8	4.5	
40	14.1	3.0	8.2	1.20	52.1	13.7	2.8	4.7	
44	14.2	3.0	8.0	1.40	52.0	13.5	3.1	4.7	
45	13.0	3.1	9.5	0.90	50.7	15.5	2.9	4.3	
47	14.0	3.0	8.3	1.00	52.6	13.5	3.0	4.5	
38	15.1	3.0	8.2	1.15	50.5	12.5	3.0	4.6	
28	12.6	3.10	8.0	1.50	52.4	14.1	2.4	3.6	Cape Foulweather
29	14.1	3.05	8.1	1.15	51.0	12.9	2.6	4.9	
30	13.9	3.00	8.3	1.30	52.7	13.4	2.7	4.3	
31	13.9	3.05	8.1	1.30	51.1	13.1	2.6	4.7	
32	12.8	2.95	8.3	1.30	52.8	14.0	2.8	4.0	
33	14.2	2.95	8.2	1.30	52.0	13.4	2.7	4.6	
34	13.8	2.95	8.1	1.40	52.0	13.5	2.8	4.3	
585	10.6	1.65	10.3	0.65	52.7	14.7	2.3	7.2	Pomona

Appendix II

PLOT NO.	FIELD NO.	FIELD LOCATION
1	06-I	**Bradley State Park Section - Unit 1
2	75-I	Bradley State Park Section - Unit 2
3	119-I	Bradley State Park Section - Unit 3
4	155-I	Bradley State Park Section - Unit 4
5	180-I	Bradley State Park Section - Unit 6
6	202-I	Bradley State Park Section - Unit 6
7	530-I	Bradley State Park Section - Unit 8
8	642-I	Bradley State Park Section - Unit 8
9	735-I	Bradley State Park Section - Unit 8
10	613	Subaerial; NW, 19, T8N, R6W
11	615	Subaerial; NW, 19, T8N, R6W
12	2GCG	Pillow; SW, 6, T7N, R6W
13	300-A	Pillow; SW, 21, T7N, R6W
14	300-B	Pillow; SW, 21, T7N, R6W
15	442	Pillow flow; NW, 2, T7N, R7W
16	457	Subaerial; NE, 33, T8N, R7W
17	633	Subaerial; NW, 2, T7N, R7W
18	449	ppc; C, 34, T8N, R7W
19	662	ppc; NE, 33, T8N, R7W
20	661	Subaerial; NE, 33, T8N, R7W
21	652	ppc; NW, 33, T8N, R7W
22	125	ppc; SE, 29, T8N, R7W
23	659	ppc; C, 33, T8N, R7W
24	131	Dike; C, 33, T8N, R7W
25	144	Dike; SE, 33, T8N, R7W
26	698	Dike; Kerry Rd.; NW, 29, T7N, R6W
27	360	Dike; NW, 31, T8N, R7W
28	100	ppc; SW, 25, T8N, R8W
29	127	Sill; SE, 29, T8N, R7W
30	638	Sill; C, 29, T8N, R7W
31	714	Sill; NW, 29, T8N, R7W
32	367	Dike; SW, 29, T8N, R7W
33	227	Sill; NE, 36, T8N, R8W
34	373	Sill; SE, 32, T8N, R7W
35	127Br	ppc; SE, 29, T8N, R7W
36	399	Subaerial; C, 17, T7N, R6W
*38	401	*** SW, 3, T7N, R6W
*39	MR69-215	NE, 31, T8N, R6W
*40	MR69-216	SE, 6, T7N, R6W
*41	MR69-218	OC 425, NW, 17, T7N, R6W
*42	MR69-220	NW, 18, T7N, R6W
*43	MR69-221	NW, 16, T7N, R6W
*44	MR69-222	SE, 17, T7N, R6W
*45	MR69-223	SW, 17, T7N, R6W
*46	MR69-225	SE, 17, T7N, R6W
*47	SR59-18	SE, 3, T7N, R6W
585	585	NW, 31, T9N, R6W

* Located on field map (Plate 1) by *plot No. (Snavely et al., 1973)

** See Appendix I

*** 118 of Coryell, 1978

APPENDIX III

Trace Element Concentrations for Basalts in Nicolai Mt.-Gnat Creek Area (Including %FeO and Na_2O)

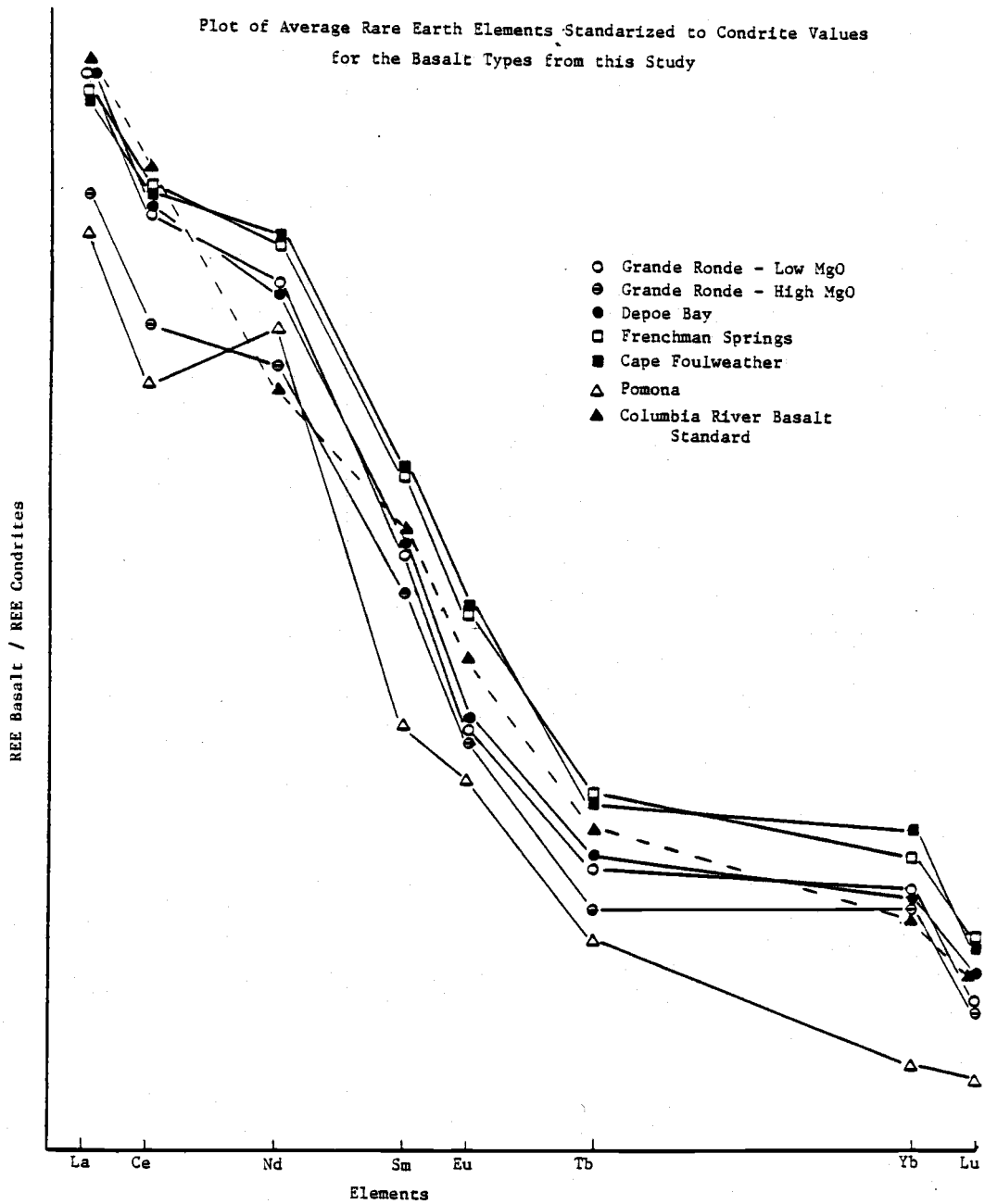
	DEPOE BAY					GRANDE RONDE						FRENCHMAN SPRINGS					CAPE FOULWEATHER			POMONA	STANDARDS		ERROR
	22	23	24	25	27	1	3	6	20	21	26	8	9	15	16	19	29	31	33	585	CRB	BHVO	% RANGE
FeO%	10.8	11.1	10.5	10.7	11.0	11.5	10.4	11.1	11.5	10.7	11.4	13.2	13.3	13.3	13.5	13.4	14.4	13.8	14.0	10.2	12.3	11.3	1-5
Na_2O	3.27	3.20	3.16	3.12	3.25	3.14	2.96	2.97	3.18	3.19	3.30	2.95	2.68	2.74	2.76	2.82	2.87	2.88	2.96	2.45	3.27	2.29	1-5
Cr*	14	13	13	13	13	14	32	41	9.9	10	9	43	41	42	43	42	40	41	42	91	11	26	5-15
Sc	30	29	29	30	34	30	35	34	29	29	30	36	34	35	35	35	36	35	35	35	32	31	1-5
Co	32	33	34	35	29	35	36	35	33	32	35	38	37	38	38	38	39	39	38	68	36	44	1-5
Sr	290	290	230	270	245	320	270	250	280	310	260	320	260	230	330	300	280	280	300	180	330	--	10-20
Zr	180	170	110	160	160	170	175	112	144	190	250	150	190	160	170	200	200	150	177	170	180	180	10-20
Ba	720	680	690	670	610	710	440	440	670	590	650	500	480	520	490	450	500	485	490	250	670	--	10-15
La	24.1	23.7	23.4	23.9	23.5	23.2	18.6	19.2	23.8	23.8	24.9	23.2	21.9	22.7	23	23.2	23.0	22.5	22.9	17.7	24.2	14.7	1-5
Ce	49	49	48	49	47	48	38	40	47	48	52	49	48	49	60	49	50	48	49	36	52	37	5-10
Nd	34	26	30	33	26	26	26	26	31	34	29	36	28	31	40	29	32	30	37	28	25	28	10-30
Sm	6.40	6.37	6.16	6.13	6.24	6.04	5.67	5.54	6.23	6.20	6.78	7.26	6.83	7.02	7.13	7.26	7.06	7.12	7.22	4.79	6.50	6.06	1-5
Eu	1.70	1.74	1.68	1.73	1.67	1.68	1.69	1.60	1.66	1.64	1.85	2.07	2.02	2.07	2.10	2.11	2.10	2.08	2.10	1.51	1.89	1.93	1-5
Tb	.942	1.01	.958	.949	.887	.980	.880	.844	.913	.890	1.03	1.08	1.00	1.03	1.10	1.04	1.10	1.04	1.09	.793	.99	.93	5-10
Yb	3.44	3.22	3.54	3.26	3.26	3.17	3.17	3.43	3.49	3.47	3.22	3.75	3.34	3.73	3.66	3.74	3.82	3.58	4.02	2.45	3.35	2.06	1-5
Lu	.506	.489	.496	.531	.513	.462	.521	.400	.474	.478	.485	.531	.511	.535	.527	.553	.534	.520	.524	.409	.500	.305	5-10
Hf	4.60	4.68	4.56	4.58	4.46	4.58	3.87	3.91	4.59	4.38	5.05	4.60	4.33	4.53	4.56	4.47	4.60	4.48	4.46	3.46	4.70	4.24	1-5
Ta	1.10	1.04	1.10	1.07	1.02	1.14	.877	.889	1.06	1.06	1.09	1.23	1.24	1.26	1.24	1.29	1.23	1.16	1.23	1.02	1.00	1.65	5-10
Th	5.61	5.17	5.18	5.40	5.18	5.17	3.06	3.68	5.17	5.48	5.17	3.96	3.62	3.83	3.90	3.80	3.96	3.73	3.80	2.48	5.82	1.01	1-10
Rb	40	38	41	41	27	39	21	32	37	36	47	30	25	26	25	25	24	25	31	8	46.6	--	--
Ni	--	45	--	23	12	19	20	<40	10	10	33	37	33	24	20	38	32	30	33	130	<40	120	20-60

*ppm (parts per million)

Location of Samples listed in Appendix II.

Appendix III continued

Plot of Average Rare Earth Elements Standarized to Condrite Values for the Basalt Types from this Study



Appendix IV

Fossil Checklist for Sedimentary Units in the Nicolai Mt. - Gnat Ck. Area

Pittsburg Bluff Formation

Fossil	Sample Location					
	OC46 (FS-1)	OC46 (FS-2)	OC46 (FS-3a)	OC50 (FS-3)	OC51 (FS-4)	OC49 (cave)
Mollusks identified by Moore (1980) and Addicott (1979)						
Gastropods						
Molopophorus cf. M Gabbi Dall	-	X	-	-	X	X
M. Aff. M. Newcombei (Merriam) ...	-	-	-	-	X	-
Perse Pittsburgensis Durham	-	X	X	X	X	-
Perse? sp. cf. P. Pittsburgensis Durham	-	-	-	-	-	X
Acteon Chehalisensis (Weaver)	-	-	-	-	X	-
Naticid	X	X	-	-	X	-
Scaphander Stewarti Durham	-	-	-	-	X	-
Scaphander sp. cf. s. stewarti Durham	-	-	-	-	-	X
Polinices cf. P/ Washingtonensis (Weaver)	-	-	-	X	-	-
Bivalves						
Spicula Pittsburgensis Clark	-	-	-	-	X	-
S. cf. s. Pittsburgensis Clark ..	-	-	X	-	-	-
S. sp.	X	-	-	-	-	-
Macoma Pittsburgensis (Clark)	-	-	-	X	-	-
Pitar sp.	-	-	-	X	-	-
Solen Townsendensis Clark	-	-	-	-	X	-
Solen sp.	-	-	X	-	-	-
Tellina Aduncanasa Hickman	-	-	-	-	X	-
Yoldia Oregona (Shumard)	-	-	-	-	X	-
Mytilus sp. - juvenile specimens .	-	-	-	-	X	-
Acila (Truncacila) Shumardi (Dall)	-	-	-	-	-	X
Callista (Macrocallista) Pittsburgensis Dall	-	-	-	-	-	X

Appendix IV

Fossil Checklist for Sedimentary Units in the Nicolai Mt. - Gnat Ck. Area
(continued)

Pittsburg Bluff Formation

Fossil	Sample Location					
	OC46 (FS-1)	OC46 (FS-2)	OC46 (FS-3a)	OC50 (FS-3)	OC51 (FS-4)	OC49 (cave)
Scaphopods						
Dentalium	-	-	-	-	-	X
Trace Fossils identified by Kent Chamberlain (1980)						
Planolites ?				-		

Sample	U.S.G.S. Cenozoic No.	Field Location
OC46	M7460 M7461 M7462	Roadcut, NW, 6, T7N, R5W
OC49	M7714	Roadcut, NE, 1, T7N, R6W
OC50	M7463	Streamcut, SE, 36, T8N, R6W
OC51	M7464	Streamcut, NE, 1, T7N, R6W

Appendix IV(continued)

Oswald West Mudstone

Fossil	Sample Location 241	312	60	285
Mollusks identified by Moore (1980)				
Pelecypods:				
Acila sp. cf. A. (A.) Gettysburgensis (Reagan)	X	-	-	-
Cyclocardia sp. cf. c. Castor (Dall)	-	-	-	X
Cyclocardia ? sp.	X	-	-	-
Callista (Macocallista) cathcartensis (Weaver)	X	-	-	-
Nuculana (saccella?) sp. cf. N. (s.) Washingtonensis (Weaver)	-	X	-	-
Nuculana (saccella?) sp.	X	-	-	X
Lucinoma sp. cf. L. Hannibali (Clark)		X	-	-
Panopea Generosa (Gould)	-	X	-	-
Anadara sp.	-	-	-	X
Gastropods:				
Bathybenbix sp. cf. B. Washingtonensis (Dall)	X	-	-	-
Forams identified by Rau (1980)				
Globobulimina Pacifica Cushman	-	-	R	-
Pseudo gland‿ulima cf. P. inflata (Bornemann)	-	-	R	-
Cassidulina sp.	-	-	R	-
Lenticulina sp.	-	-	R	-
Elphidium sp. ?	-	-	R	-
Epistominella sp. ?	-	-	R	-
Globigernia spp.	-	-	R	-
Lagena sp.	-	-	R	-

Appendix IV (continued)

Oswald West Mudstone (continued)

Fossil	241	312	60	285
Forams (continued)				
Gyroidina sp.	-	-	R	-
Fursenkoina sp.	-	-	?	-
Cyclammina sp.	-	-	C	-

Key:

C indicates Common.
F indicates Few.
R indicates Rare.

Sample	U.S.G.S Cenozoic No.	Location
60	-	Roadcut, SW, 35, T8N, R6W
241	M7700	Roadcut, SW, 35, T8N, R6W
312	M7708	Roadcut, SE, 15, T7N, R6W
285	M7706	Streamcut, NW, 2, T7N, R6W

Appendix IV (continued)

	Astoria Formation												
	Upper Silver Point Mudstone					Pipeline Member		Big Creek Sandstone Member					
Fossil	91	106a	200	208	210	212	216	245	246	257	259	283	692
Mollusks identified by Moore (1980)													
Pelecypods													
Acila? sp.	-	-	-	X	-	-	-	-	X	-	-	X	-
Anadara (Anadara) Devincta (Conrad)	-	-	-	-	-	-	-	-	-	-	-	-	X
A. (A.) sp. cf. A. (A.) Devincta (Conrad)	-	-	-	-	-	-	-	-	X	-	-	-	-
Anadara? sp.	X	-	-	-	-	-	-	X	-	-	-	-	-
Arca n. sp.?	-	-	-	-	-	-	-	-	-	X	-	-	-
Cardita? sp. cf. c. subtena (Conrad)	-	-	-	-	-	-	-	X	-	-	-	-	-
Delectopecten sp. cf. D. Peckhami (Gabb) of Arnold	-	-	-	X	-	-	-	-	-	-	-	-	-
Delectopecten? sp.	-	X	X	-	-	-	-	-	-	-	-	-	-
Katherinella (Katherinella) Augustifrons (Conrad)	-	-	-	-	-	-	-	-	-	-	X	-	-
K. (K.) sp. cf. K. (K.) Augustifrons (Conrad)	-	-	-	-	-	-	-	-	X	-	-	-	-
Macoma? sp.	-	-	X	-	-	-	-	-	-	X	-	-	-
Nuculana (saccella) sp.	-	-	-	-	-	-	-	-	-	-	-	X	-
N. (s.) sp. cf. N. (s.) Calkinsi (Moore)	X	-	-	X	-	-	X	X	-	-	X	-	-
N. (s.) sp. cf. N. (s.) Amelga (Moore)	X	-	-	-	-	-	-	X	-	-	-	-	-
	91	106a	200	208	210	212	216	245	246	257	259	283	692

Appendix IV(continued)

Astoria Formation (continued)

Fossil	Upper Silver Point Mudstone					Pipeline Member		Big Creek Sandstone Member					
	91	106a	200	208	210	212	216	245	246	257	259	283	692
Pelecypods (continued)													
Nuculana (saccella?)	-	-	-	-	-	-	-	-	X	-	-	-	-
Nuculana sp.	-	-	-	-	X	-	-	-	-	-	-	-	-
Unidentified venerid?	-	-	-	-	-	-	-	X	-	-	-	-	-
Gastropods:													
Ficus? sp.	-	-	-	-	-	-	-	-	-	-	-	X	-
Unidentified naticid?	-	-	X	-	-	-	-	-	-	-	-	-	-
Foraminifera identified by Rau (1980)													
Globobulimina Pacifica Cushman				R	-	-	-						
Globigerina spp.				F	-	-	R						
Fusenkoina sp.				F	-	-	?						
Praeglobobulimina cf. P. ovata (d'Orbigny)	NO FORAMS			C	-	C	R			NO FORAMS			
Valvulineria Araucana (d'Orbigny)				-	-	F	F						
Siphogenerina sp.				-	-	C	-						
Cyclammina sp.				C	-	-	-						
Pyrgo sp.				R	-	-	-						
Quinqueloculina sp.				C	-	-	-						
Florilus incisum (Cushman)				R	-	-	-						
	91	106a	200	208	210	212	216	245	246	257	259	283	692

Appendix IV (continued)

Astoria Formation (continued)

Fossil	Upper Silver Point Mudstone					Pipeline Member			Big Creek Sandstone Member				
	91	106a	200	208	210	212	216	245	246	257	259	283	692
Foraminifera (continued)													
Buliminella cf. B. subfusiformis Cushman				-	-	-	?						
Diatoms:									64	252			
identified by Barron (1980) from two samples in Big Creek Sandstone													
Coscinodiscus sp.									-	X			
Stephanopyxis sp.									-	X			
Thalassionema nitzschioides (Grunow)									X	-			
Thalassiothrix Fragments									X	-			

Sample	U.S.G.S. Cenozoic No.	Location
Silver Point Mudstone:		
91	M7690	Roadcut, NW, 29, T8N, R7W
106a	M7694	Roadcut, NE, 26, T8N, R8W
200	M7692	Railroadcut, SW, 14, T8N, R8W
208	M7693	Roadcut, NE, 26, T8N, R8W
210	M7695	Roadcut, SW, 23, T8N, R8W
Pipeline Mudstone:		
212		Roadcut, NE, 26, T8N, R8W
216	M7696	Roadcut, SE, 24, T8N, R8W
Big Creek Sandstone:		
245	M7701	Roadcut, SE, 2, T7N, R6W
246	M7702	Roadcut, SW, 1, T7N, R6W
257	M7703	Roadcut, SE, 35, T8N, R6W
259	M7704	Roadcut, SE, 34, T8N, R6W
283	M7705	Streamcut, NE, 2, T7N, R6W
692	M7713	Streamcut, SE, 35, T8N, R6W
64	-	Streamcut, SW, 35, T8N, R6W
252	-	Streamcut, SE, 35, T8N, R6W

Appendix IV (continued)

Basalt Interbeds

Fossil	Depoe Bay		Cape Foulweather							Columbia River	
	125	646	232	233	236	366	367a	704	227	169	2GC
Mollusks identified by Moore (1980)											
Pelecypods:											
Nuculana (saccella) sp. cf. N. (s.) Calkinsi (Moore)	X	X	X	X	X	-	-	-	-	-	-
Macoma? sp.	X	X	-	-	-	-	-	-	-	-	-
Acila sp. cf. A. (A.) Gettysburgensis (Reagan)	-	-	-	-	X	-	-	-	-	-	-
Acila? sp.	-	-	-	-	-	-	X	-	-	-	-
Lucinoma sp. cf. L. Hannibali (Clark)	-	-	-	-	X	-	-	-	-	-	-
Unidentified venerid?	-	-	-	-	X	-	-	-	-	-	-
Anadara sp.	-	-	-	-	X	-	-	-	-	-	-
Tellina? sp.	-	-	-	-	-	-	X	-	-	-	-
Gastropods:											
Vitrinella? sp.	-	-	-	-	-	-	-	-	-	-	X
Unidentified sm. frags.	-	-	-	-	-	-	-	-	-	-	X
Echinoids:											
Cidaroid echinoid (sea urchin) spines	-	-	-	-	-	-	-	-	-	-	X
Foraminifera identified by Rau (1980)											
Globobulimina Pacifica Cushman	R	-	NO FORAMS			-	-	-	-	NO FORAMS	
	125	646	232	233	236	366	367a	704	227	169	2GC

Appendix IV (continued)

Basalt Interbeds (continued)

Fossil	Depoe Bay		Cape Foulweather							Columbia River	
	125	646	232	233	236	366	367a	704	227	169	2GC
Foraminifera (continued)											
Cassidulina sp.	–	–				?	–	R	–		
Lenticulina sp.	R	–				R	–	–	–		
Globigerina spp.	–	–				F	F	F	–		
Praeglobobulimina cf. P. ovata (d'Orbigny)	F	–				R	C	F	–		
Florilus costiferum Cushman	?	–				F	C	F	–		
Valvulineria arencana (d'Orbigny)	F	–	NO FORAMS			C	C	C	–	NO FORAMS	
Siphogenerina sp.	R	–				–	–	C	–		
Cassidulinoides sp.	?	–				–	–	–	–		
Cyclammina sp.	–	–				R	–	–	–		
Bolivina cf. B. advena Cushman	–	–				R	R	–	–		
Buliminella cf. B. subfusiformis Cushman	–	–				R	?	–	–		
Siphogenerina Kleinpelli Cushman	–	–				F	F	–	–		
Saracenaria(?) cf. s. acutauricularis (Fichtel and Moll)	–	–				R	R	–	–		
Nonionella Miocenica Cushman	–	–				–	F	–	–		
Diatoms: identified by Barron (1979, 1980)											
Actinocyclus ingens (Rattray)	X	–	–	–	X	–	X	–	–	NO DIATOMS	
Coscinodiscus marginatus (Ehrenberg)	X	–	–	–	X	–	–	–	–		
	125	646	232	233	236	366	367a	704	227	169	2GC

Appendix IV (continued)

Basalt Interbeds (continued)

Fossil	Depoe Bay		Cape Foulweather							Columbia River	
	125	646	232	233	236	366	367a	704	227	169	2GC
Diatoms (continued)											
Melosirasulcata (Ehrenberg) Kutzing .	X	-	-	-	-	-	-	-	-		
Melosira sp.	X	-	-	-	-	-	-	-	-		
Thalassiothrix longissima (Cleve and Grunow)	X	-	-	-	X	-	-	-	-	NO DIATOMS	
Thalassionema nitzschioides (Grunow)	-	-	-	-	X	-	-	-	-		
Coscinodiscus sp.	-	-	-	-	-	X	X	-	X		
Stephanopyxis sp.	-	-	-	-	-	-	X	-	-		
Trace Fossils: by Kent Chamberlain (1979)											
Rosselia tube?	-	-	-	-	-	-	-	-	-	X?	-

Outcrop	U.S.G.S. Cenozoic No.	Location
125	M7691	Quarry, SE, 29, T8N, R7W
646	M7711	Streamcut, C, 29, T8N, R7W
232	M7697	Roadcut, NW, 32, T8N, R7W
233	M7698	Roadcut, NW, 32, T8N, R7W
236	M7699	Roadcut, SE, 29, T8N, R7W
366	-	Roadcut, NW, 32, T8N, R7W
367a	M7710	Roadcut, NW, 32, T8N, R7W
704	-	Cliff, SE, 29, T8N, R7W
227	-	Quarry, SE, 36, T8N, R8W
169	-	Cliff, C, 34, T8N, R7W
2GC	M7727	Streamcut, NW, 6, T7N, R6W

Appendix IV (continued)

Checklist of Diatom Flora from the Clifton Formation

Fossil	Sample Location 17	24	25	29[1]	513[1]	519[1]	547	595[1]	602[1]	628
Diatoms: identified by Barron (1979, 1980) and Schrader (1980)										
Actinocyclus ingens (Rattray)	X	-	-	X	X	X	X	-	X	-
Denticula lauta (Bailey)	-	-	-	X	-	X	X	-	-	-
D. hyalina (Schrader)	-	-	-	X	X	X	X	X	X	-
D. hustedtii (Simonsen and Kenaya)	-	-	-	-	-	-	X	-	-	-
Actinoptychus spp.	X	X	X	-	-	-	-	-	-	-
Aulacodiscus sp.	X	X	-	-	-	-	-	-	-	-
Rhaphoneis spp.	X	X	X	-	-	-	-	-	-	-
Cocconeis spp.	X	X	X	-	-	-	-	-	-	-
Thalassionema nitzschioidus (Grunow)	-	X	X	-	-	-	-	-	X	X
Stephanopyxis sp.	X	X	X	-	-	-	-	-	-	-
Pinnularia sp.[2]	X	X	-	-	-	-	-	-	X	-
Eunotia sp.[2]	-	X	X	-	-	-	-	-	-	-
Rouxia naviculoides (Schrader)	-	-	-	-	-	X	-	-	-	-
Synedra jouseana (Sheshukova-Porotzkaya)	X	-	-	X	-	-	-	-	-	-
Thalassiothrix longissima (Cleve & Grunow)	-	-	-	-	-	-	-	-	-	-
T. fragments	-	-	-	-	-	-	-	-	-	-
	17	24	25	29	513	519	547	595	602	628

Appendix IV(continued)

Checklist of Diatom Flora from the Clifton Formation (continued)

Fossil	Sample Location									
	17	24	25	29[1]	513[1]	519[1]	547	595[1]	602[1]	628
Diatoms (continued)										
Melosira sulcata (Ehrenberg) Kutzing	X	X	X	-	-	-	-	-	-	-
M. granulata[2]	-	X	X	-	-	-	-	-	-	-
M. islandica[2]	-	-	-	-	-	-	X	-	-	-
M. sp.	-	-	-	-	-	-	-	-	X	-
Coscinodiscus c.f. intersectus	X	-	X	-	-	-	-	-	-	-
C. plicatus (Grunow)	-	-	-	-	X	-	X	-	-	-
C. praeyabei (Schrader)	-	-	-	-	X	-	-	-	-	-
C. ssp.	X	X	X	-	-	-	-	-	-	-
Cymbella sp.[2]	-	-	-	-	-	-	-	-	X	-
Synedra spp.[2]	-	-	-	-	-	-	-	-	X	-
Paralia sulcata[4]	-	-	-	-	-	-	-	-	X	-
Phyroliths[2]	-	-	-	-	-	-	X	-	-	-
Delphineis spp.[3]	-	-	-	-	-	-	X	-	-	-
Nitzschia c.f. curta?[3].....	-	-	-	-	-	-	X	-	-	-
Rhizosolenia miocenica	-	-	-	-	X	-	-	-	-	-
Silica flagellates	-	X	X	-	-	-	-	-	-	-
Sponge spicules	X	X	X	-	-	-	-	-	-	-
Chaetoceros (Diatom resting spores)	X	X	X	-	X	-	-	-	-	-
Satae (brissles)	-	-	-	-	X	-	-	-	-	-
	17	24	25	29	513	519	547	595	602	628

Appendix IV(continued)

Checklist of Diatom Flora from the Clifton Formation (continued)

Fossil	Sample Location
Trace fossils: identified by Kent Chamberlain (1980)	
Rosselia Burrows	Present in facies 1a at outcrops 84, 116, 119, and 215.

Sample	U.S.G.S. Cenozoic No.	Location
17		Roadcut, NW, 9, T8N, R6W
24		Roadcut, NE, 8, T8N, R6W
25	None	Roadcut, NE, 8, T8N, R6W
29		Roadcut, SE, 5, T8N, R6W
513	Reported	Roadcut, C, 7, T8N, R6W
519		Roadcut, NE, 7, T8N, R6W
547		Roadcut, NW, 17, T8N, R6W
595		Roadcut, SW, 26, T9N, R7W
602		Roadcut, SE, 1, T8N, R7W
628		Roadcut, NW, 7, T8N, R6W

Notes:
1 - only age diagonistic species
2 - freshwater fauna
3 - indicates cold-water upwelling areas
4 - brackish water fauna

Appendix V Size Analysis Data of Sedimentary Units in the Thesis Area

Sample	Sieve	Hydro Meter	Median phi	Mean phi	Sorting phi	Skewness phi	Kurtosis phi	Coarsest 1% mm	Median (mm)	Sand	Silt %	Clay %	Name
Pittsburg Bluff Formation													
47 (coarse ss)	X	-	-0.47	-.06	1.09	0.69	1.45	2.00	1.40	98.63	-	-	ss
49	X	-	3.25	3.28	0.55	0.12	1.04	.25	0.11	89.27	-	-	ss
Oswald West Mudstone													
60**	-	X	5.84	6.66	1.47	1.23	-	0.06	0.01	1.04	71.03	28.97	c - silt
Astoria Fm.													
Pipeline Mudstone (sandstone beds)													
212	X	X	2.25	2.61	1.17	0.52	1.13	0.48	0.21	84.48	15.02	1.50	ss
106_a	X	-	3.26	3.30	0.77	0.01	1.01	0.45	0.10	83.36	-	-	ss
Silver Point Mudstone													
91**	-	X	6.05	7.47	1.49	1.30	-	0.11	0.01	5.16	59.79	32.68	c - silt
Big Creek Sandstone													
64	X	X	3.14	3.28	1.00	0.28	1.41	0.45	0.11	80.39	18.57	0.23	silty - ss
246**	X	X	4.41	6.51	1.60	1.76	-	0.20	0.05	41.10	29.96	29.44	c - silt-ss
692	X	X	4.60	5.55	1.81	0.70	1.02	0.11	.04	9.09	74.96	15.95	c - siltstone
CR & Coastal B Sand- & Mudstone Interbeds													
38 (ss)	X	-	1.86	2.06	1.28	0.44	1.80	0.85	0.27	88.28	-	-	ss
169 (Vantage M.)	X	-	2.60	2.70	0.74	0.33	1.38	0.34	0.16	90.43	-	-	ss
174_T (Vantage M.)	X	-	2.13	2.14	0.59	0.28	1.40	0.40	0.26	95.27	-	-	ss
174_B (Vantage M.)	X	-	2.78	3.02	1.07	0.37	1.36	0.45	0.26	82.56	-	-	ss
367_a** (Mudst.)	-	X	6.06	7.13	1.48	1.19	-	-	-	Trace	63.51	36.40	c - silt

Appendix V (continued)

Sample	Sieve	Hydro Meter	Median phi	Mean phi	Sorting phi	Skewness phi	Kurtosis phi	Coarsest 1% mm	Median (mm)	Sand %	Silt %	Clay %	Name
Clifton formation													
Sandstone facies 1													
Bar Unit													
116	X	X	2.78	3.34	1.87	0.63	0.77	0.45	0.14	75.34	15.44	8.97	ss
215**	X	-	2.47	2.95	1.45	(1.25)	-	0.66	0.13	75.27	-	-	Silt ss-ss
"fluvial" unit													
9	X	-	2.22	2.24	0.77	0.19	1.41	0.56	0.22	93.80	-	-	ss
26	X	-	2.43	2.48	0.62	0.24	1.33	0.42	0.18	94.92	-	-	ss
79	X	-	1.41	1.51	1.12	0.29	1.33	1.25	0.37	92.51	-	-	ss
177	X	-	1.23	1.35	0.95	0.31	1.44	1.10	0.41	96.11	-	-	ss
186	X	-	2.02	2.19	0.86	0.33	1.43	0.60	0.24	93.02	-	-	ss
379	X	-	2.49	2.50	0.73	0.19	1.60	0.44	0.17	92.77	-	-	ss
480	X	-	2.69	2.67	0.56	0.11	1.46	0.33	0.15	94.44	-	-	ss
524	X	-	1.50	1.68	1.22	0.32	1.17	0.96	0.35	92.83	-	-	ss
553 (coarse ss)	X	-	0.83	0.89	0.84	0.33	1.63	1.25	0.57	96.59	-	-	ss
Channel facies													
25*	X	-	2.46	2.78	1.16	0.28	-	0.90	0.18	84.98	-	-	ss
484	X	-	0.86	0.99	1.39	0.28	-4.54	2.50	0.55	93.64	-	-	ss
620	X	-	1.48	1.60	1.07	0.24	1.34	1.24	0.36	95.19	-	-	ss
Siltstone facies													
16	X	X	6.38	-	-	-	-	0.16	0.01	13.00	50.52	36.30	c - silt
20*	X	X	3.44	5.07	3.44	0.71	-	0.36	0.09	56.76	21.87	21.36	c - silt-ss
511*	-	X	6.42	6.42	2.58	0.00	-	0.17	.01	18.76	53.71	26.80	c - silt
513 (sandstone unit)	X	-	2.87	2.94	0.85	0.13	0.99	0.36	0.13	87.70	-	-	ss
628**	X	X	6.06	6.65	1.42	1.08	-	0.16	.01	8.45	58.96	32.54	c - silt
Columbia River Sand	X	-	1.74	1.76	0.55	-1.76	2.39	0.80	0.30	100.00	-	-	ss

Statistical parameters from Folk and Ward (1957) except those marked (*) are from Inman (1952) and those marked (**) are from Trask (1932).

Subscript T is Top of sandstone interbed

Subscript B is Bottom of sandstone interbed

Appendix VI

Heavy Minerals of Selected Sandstone in the Nicolai Mt. - Gnat Ck. Area

Mineral Species	Clifton Formation				Vantage Member	Pittsburg Bluff	Astoria Fm. Pipe-line ss	Big Ck ss
	9	177	484	116	169	49	212	246
Amphiboles								
Hornblende								
Green	C	A	R	R	VA	VA	R	VA
Brown	C-R	-	-	R	C	R	-	R
Lamprobolite	R	C	VR	-	R	R	R	C
Actinolite	VR	R	-	-	?	-	-	-
Pyroxene								
Hypersthene	C	R	C-R	VR	R	A-C	R	C
Enstatite	A-C	C	C-R	R	R	R	R	C
Augite	A	R	C-R	-	R	C	C	C
Micas								
Biotite								
Brown	R	C	R	V A	A	R	V A	C
Green	-	-	-	V A	R	-	A	-
Muscovite	-	-	-	A	-	-	-	-
Chlorite (Penninite)	VR	VR?	VR	VR	-	-	-	-
Epidote	A-C	A	A	C	C	C-R	R	C
Tourmaline								
Brown	R	VR	R	-	R	R	R	R
Blue (schorlite)	-	-	-	-	-	-	-	R
Garnet								
Clear	R	C	VR	C	C	R	C	C-A
Pink	VR	VR	-	VR	R	VR	-	-
Kyanite	R	R	C	R	R	R	A	VR
Staurolite	C-R	VR	C-R	R	VR	-	C	R
Sillaminite	VR	-	VR?	-	VR	-	-	-
Monazite	-	-	R	-	-	R	R	VR
Glaucophane	-	-	VR?	-	-	-	-	-
Rutile	VR	VR	-	-	R	VR?	-	VR
Zircon	C	R	A	R	R	R	C	R
Sphene	-	-	-	-	-	VR?	R	R
Opaques								
Magnetite	A	C	C	C	C	C	A	R
Hematite	C	R	R	R	-	C	C	R
Ilmenite	A	C	C	C	C	R	C	C
Leucoxene	R	C	R	R	C	R	VA	C
Pyrite	-	-	-	VR	-	-	-	-

Number of grains

V.A = 25
A = 15
C = 10 - 15
R = 2 - 10
VR = 1 - 2

Appendix VII

Pebble Count from Conglomerate Units in Study Area (in percent)

Rock Type	Clifton Formation			Oswald West Mudstone	Astoria Fm. Big Ck.	
Sam. No.	OC-18	595	355	242	64	287
Volcanic						
Andesite ?	23.0	10.5	12.5	9.5	9.5	8.5
Basalt	23.0	26.3	6.3	14.3	14.2	14.8
Siliceous Volc.	-	2.6		18.0	4.7	12.6
Undifferentiated Mafic	8.0	10.5	43.7	21.4	26.0	38.0
Plutonic						
Granitic	26.0	13.1	5.9	4.8	7.1	4.2
Metamorphic						
Schist	3.2	7.9	-	-	7.1	2.1
Quartzite	-	28.7	18.7	16.8	11.9	8.5
Sedimentary						
Sandstone	6.3	-	6.2	-	2.3	-
Siltstone	6.3	-	6.2	-	-	-
Chert	4.0	-	-	15.0	17.1	10.0

Appendix VIII

Chemical analyses of Clifton sandstone (facies 1b)

Compliments of Steward and Les Lahti

Analysis performed by Oregon Portland Cement Company on 1/22/72 and 3/20/75

	11/22/72		(3/20/75)				
	Lahti #12 (OC 177)	#18 (OC 178)	1A	1B (locations unknown)	10A	11A	16A
SiO_2	66.76	77.52	74.6	77.3	73.7	72.2	77.2
Fe_2O_3	6.20	2.82	3.1	2.6	4.0	3.7	3.9
Al_2O_3	17.64	11.88	12.7	11.8	12.5	15.5	11.5
CaO	0.86	1.30	-	-	-	-	-
K_2O	1.79	1.72	2.1	1.9	2.0	2.0	2.2
Na_2O	1.00	1.00	1.1	1.0	1.1	1.0	1.0
MgO	0.46	?	-	-	-	-	-
SO_3	0.05	0.01	-	-	-	-	-
S	0.02	0.00	-	-	-	-	-
Loss	5.70	3.03	-	-	-	-	-

Appendix IX

Test results of reservoir and source rock potential of the units in the Nicolai Mountain-Gnat Creek Area. Compliments of Tenneco Oil Company, Bakersfield, California.

Porosity and Permeability Test Results

Sample No.	Location	Permeability	Porosity
25	NW, 9, T8N, R6W	98	38.5
7	SE, 13, T8N, R6W	1262	36.0
84	NW, 42, T8N, R7W	27	31.8
312	NE, 22, T7N, R6W	67	36.2
49	NE, 1, T7N, R6W	42	39.7

Permeability measured in millidarcies
Porosity measured in percent volume

Hydrocarbon Maturation Test Results

Sample No.	Location	TAI	Organic Constituents	V.R.(%Ro)
9	NW, 9, T8N, R6W	2.3	50%W, 50%C	0.3
595	SW,19, T9N, R7W	2.3	90%W, 10%S-P	0.3
602	SE, 26, T8N, R7W	2.3	60%C, 40%W	0.3
538	NW, 25, T8N, R7W	2.3-2.5	60%C, 30%W, 10%S-P	0.3
98	NE, 25, T8N, R7W	2.3	60%W, 30%C, 10%S-P	0.3
216	SE, 24, T8N, R8W	2.3	60%W, 40%C, Trace S-P	0.3
411	NW, 20, T7N, R6W	2.3-2.5	90%W, 10%C, Trace S-P	0.7
91	NW, 29, T8N, R7W	2.3	50%W, 50%C, Trace S-P	0.3
692	SE, 35, T8N, R6W	2.3	70%C, 20%W, Trace S-P	0.3

Appendix X

Location of Remanant Polarity Basalt Samples in this Study Area

Sample Number (outcrop locations)	Remanant Polarity	Location * (Rk. type; ¼, Sec., Twn., Rng.)
Bradley State Park Section		
6	N	Listed on Plate II
15	N	
75	N	
119	N	
155	N	
180	N	
202	N	
530	N	
642	N	
735	N	
2	N	subaerial; NE, 20, T8N, R6W
13	N	subaerial; SE, 17, T8N, R6W
127	N	sill; chemical analysis 29
144	N	dike; chemical analysis 25
144b	N	pillow; same as 144
169	N	subaerial; SW, 34, T8N, R7W
131	N	dike; chemical analysis 24
125	N	pillow; chemical analysis 22
227	N?	sill; chemical analysis 33
300A	R	pillow; chemical analysis 13
300B	N	pillow; chemical analysis 14
323	N	subaerial; NW, 11, T7N, R6W
337	N	subaerial; NW, 14, T7N, R6W
351	N	subaerial; SE, 21, T8N, R6W
353	N	subaerial; NW, 21, T8N, R6W
356	N	subaerial; NW, 21, T8N, R6W
360	N	dike; chemical analysis 27
367	N	dike; chemical analysis 32
373	N	dike; chemical analysis 34
376	N	dike; SW, 32, T8N, R6W
384	N	subaerial;SW, 2, T7N, R6W

Appendix X continued

Sample Number	Remanant Polarity	Location
390	N	subaerial; NE, 25, T8N, R7W
397	N	subaerial; SE, 17, T7N, R6W
400	N	subaerial; NW, 16, T7N, R6W
401	N	subaerial; chemical analysis 38
430	N	subaerial; NW, 1, T7N, R6W
440	N	subaerial; C, 36, T7N, R7W
442	N	pillow; chemical analysis 15
448	N	subaerial; C, 34, T8N, R7W
449	N	pillow; chemical analysis 18
457	N	subaerial; chemical analysis 16
460	N	pillow; NE, 33, T8N, R7W
462	N	subaerial; NW, 34, T8N, R7W
552	N	subaerial; NE, 17, T8N, R6W
582	R	subaerial; NW, 31, T9N, R6W
585	R	subaerial; chemical analysis 58[illegible]
613	N	subaerial; chemical analysis 10
615	N	subaerial; chemical analysis 11
633	N	subaerial; chemical analysis 17
636	N	breccia; SE, 2, T7N, R7W
638	N	sill; chemical analysis 30
645	N	breccia; SW, 29, T8N, R7W
652	N	breccia; chemical analysis 21
658	N	dike; C, 33, T8N,R7W
659	N	pillow; C, 33, T8N, R7W
662	N	breccia; chemical analysis 19
664	N	subaerial; NE, 33, T8N, R7W
672	N	breccia; C, 34, T8N, R7W
673	N	dike; SE, 34, T8N, R7W
675	N	breccia; SE, 34, T8N, R7W
677	N	subaerial; SE, 34, T8N, R7W
698	R	dike; Kerry Rd., chemical analysis 26

* chemical analyses listed in Appendix II and on Plate II

Appendix XI continued

Basalt Interbeds

2gc, 125, 169, 227, 232, 233, 236, 366, 367a, 646, 704 - listed in Appendix IV

outcrop	location
38	rc, NW, 8, T7N, R6W
174	rc, Section e-e', Appendix I
350	rc, SW, 21, T8N, R6W
614	sc, NW, 19, T8N, R6W

Clifton Formation

17, 24, 25, 29, 513, 519, 547, 595, 602, 628 - listed in Appendix IV

9	q, NW, 19, T8N, R6W
16	rc, SE, 8, T8N, R6W
20	rc, NW, 17, T8N, R6W
26	sc, SE, 9, T8N, R6W
79	rc, C, 37, T8N, R7W
84	rc, C, 42, T8N, R7W
111	rc, NW, 24, T8N, R8W
116	rc, NW, 24, T8N, R8W
119	rc, SE, 14, T8N, R8W
177	rc, NE, 8, T8N, R8W
178	rc, NE, 8, T8N, R8W
214	rc, C, 24, T8N, R8W
215	rc, SW, 19, T8N, R8W
355	rc, NW, 21, T8N, R6W
379	rc, C, 24, T8N, R7W
384	rc, SE, 24, T8N, R7W
480	sc, SE, 13, T8N, R7W
484	rc, C, 13, T8N, R7W
511	rc, SW, 7, T8N, R7W
524	rc, SE, 12, T8N, R7W
525	rc, SE, 12, T8N, R7W
527	rc, SW, 12, T8N, R7W
553	rc, SW, 12, T8N, R7W
616	rc, NW, 19, T8N, R7W
620	sc, NW, 14, T8N, R7W

rc = roadcut
sc = streamcut
q = quarry

APPENDIX XI

Location of Outcrops Discussed in Text

Pittsburg Bluff Formation

46, 49, 50, 51 - listed in Appendix IV

outcrop	location
47	rc, NE,1, T7N, R6W
240	rc, SW,36, T8N, R6W
720	sc, SW,36, T8N, R6W
721	sc, SW,36, T8N, R6W
724	sc, NW, 2, T7N, R6W
732	sc, NW, 2, T7N, R6W

Oswald West Mudstone

60, 241, 285 - listed in Appendix IV

242	rc, SW, 36, T8N, R6W
248	rc, SW, 1, T7N, R6W
298	rc, SW, 21, T7N, R6W
411	rc, NW, 20, T7N, R6W
729	sc, SW, 1, T7N, R6W
730	sc, C, 1, T7N, R6W
731	sc, C, 1, T7N, R6W

Astoria Formation

91, 106a, 200, 208, 210, 212, 216, 245, 246, 257, 259,

283, 692 - listed in Appendix IV

98	rc, NE, 25, T8N, R7W
287	sc, NE, 2, T7N, R6W
301	q, NW, 21, T7N, R6W

Basalt Units

Sample locations of chemical analyses listed in Appendix II.

18	sc, NW, 9, T8N, R6W
75	q, C, 27, T8N, R6W
430	q, NW, 1, T7N, R6W
582	q, NW, 35, T9N, R6W
592	rc, SW, 25, T9N, R6W