eventually built up above the ocean floor. A simplified model for the emplacement of the Cole Mountain basalt is presented in figure 68.

Siliceous hydrothermal deposits are commonly associated with basaltic sills intruded into water-saturated marine sediments and with submarine flows (Einsele, 1982). Irregular siliceous pods and veins have been reported from andesitic sills intruded into marine sediments on Unalaska Island (Snyder and Fraser, 1963). The siliceous pods in this area commonly occur as fillings between pillows on sill margins. The Cole Mountain basalt also contains irregular pods of silica and veins of microcrystalline quartz. Siliceous masses are commonly found in modern soils developed on Cole Mountain basalt. It is possible that more of these basaltic rocks were originally pillowed.

The Cole Mountain basalt was erupted and emplaced in a forearc setting on the upper continental slope or outer shelf. This is evidenced by the following: 1) the presence of upper bathyal to outer neritic foraminifera and diatoms in the mudstones directly above, below, and between the volcanic rocks (appendix 2); 2) common vesicles in the Cole Mountain basalt (vesicular basaltic eruptions generally occur at water depths less than 500 m; McBirney, 1963); 3) the presence of Keasey Formation glauconitic sandstones directly above Cole Mountain basalt. Glauconite typically forms in marine environments with low sedimentation rates and is especially common on topographically elevated areas in the outer shelf (McRae, 1972; Kulm <u>et al</u>., 1975); and 4) the presence of the Cascade volcanic arc to the east of the thesis area during the late Eocene (Wells <u>et al</u>., 1984).

Therefore, it is possible that Cole Mountain basalt formed a small submarine topographic high in the slope basin. Einsele (1982), however, has suggested that fluidization of wet sediments during emplacement of shallow intrusions results in a lateral rather than a vertical displacement of surrounding sediments. The Cole Mountain basalt contains a few hyaloclastite and debris-flow deposits which suggest topographic relief. Therefore, the Cole Mountain basalt probably formed a small submarine "knoll" with relief less than what might be expected based on the total thickness of the unit. This knoll apparently formed a high in which terrigenous sediments were funneled around resulting in low sedimentation rates allowing glauconite to form and resulting in a high concentration of waterlaid tuff.

Recently, it has been shown that when subaerially erupted basaltic rocks flow into lacustrine or marine environments, the igneous rocks can form "invasive" dikes and sills (Schmincke, 1967; Byerly and Swanson, 1978; Beeson et al., 1979). Since the Cole Mountain basalt is chronologically, geochemically, petrographically, and stratigraphically correlative to the predominantly subaerially erupted type Goble Volcanics, it is possible that the Cole Mountain basalt is "invasive" in origin. There are, however, several lines of evidence that argue against an invasive origin. First, surface and subsurface mapping (e.g. Newton, 1976; Bruer et al., 1984) show that Cole Mountain basalt and type area Goble Volcanics are separated by a large area in Columbia County where middle Eocene to middle Miocene sedimentary rocks are present but where few or no age equivalent (latest Eocene) basaltic rocks are present (see previous section).

If the sills are invasive, they should show a distinct connection with the subaerial flows. Secondly, the Cole Mountain basalt include many dikes which reorient as sills upsection (Plate II). The dikes can be traced to within several meters of the Tillamook Volcanics, although not into the volcanics (Safley, in prep.; Timmons, 1981). Invasive basalts appear to consist of sills with dikes "sprouting" upward from the sill (Byerly and Swanson, 1978; this study, Columbia River Basalt Group section). Thirdly, the degassed invasive basaltic rocks appear to contain few or no vesicles (Byerly and Swanson, 1978). This may not, however, be an inherent feature of all invasive sills. Finally, a hyaloclastitic sediment gravity flow occurs on top of the Cole Mountain basalt at several localities. This hyaloclastite deposit appears to be localized in the thesis area and, therefore, probably was not sourced from the type area Goble Volcanics. It is unlikely that a subaqueous basaltic flow would remain at the sediment-sea water interface as it crossed the shelf. Poorly consolidated, water-saturated sediments on the sea floor would be indusive to invasion rather than surficial flow. In conclusion, there is considerable evidence that the Cole Mountain basalt was locally erupted. However, since Cole Mountain basalt dikes were not found intruding Tillamook Volcanics and since the late Eocene stratigraphic section to the east of the thesis area has been removed by late Miocene to Recent erosion, the possibility of invasion from correlative hypothetical late Eocene volcanic centers to the east of the thesis area cannot be completely eliminated.

It has been shown that the Cole Mountain basalt is chemically similar to volcanic rocks formed in compressional tectonic environments (e.g., island arcs). The Cole Mountain basalt was emplaced somewhere between 36 and 41 Ma. At this time the volcanic arc was sweeping westward in response to plate reorganization and the oldest rocks of the Cascade arc were being erupted (Wells et al., 1984). The Cole Mountain basalt may represent an early, more diffuse western expression of Cascade arc volcanism or could be the result of localized forearc volcanism (see Tillamook Volcanics, Geologic History and Tectonic Setting section).

KEASEY FORMATION

Nomenclature and Distribution

Schenck (1927) first used the term "Keasey shale" for a sequence of tuffaceous siltstone and mudstone along Rock Creek near Keasey Station, in Columbia County Oregon. Weaver (1937) expanded the definition of the unit by including an overlying sequence of tuffaceous siltstone. Warren and Norbisrath (1946) divided the Keasey Formation into three informal members: 1) a lower shale member of variable thickness; 2) a middle tuffaceous, concretionary siltstone member (approx. 520m thick); and 3) an upper sandy shale member (approx. 50 m thick). Van Atta (1971) noted that the members of Warren and Norbisrath (1946) can be defined only in a limited area in Columbia County.

Olbinski (1983) and Nelson (1985) informally proposed the Jewell member of the Keasey Formation for a sequence of laminated to thinly bedded, locally glauconitic mudstones near Jewell, in Clatsop County Oregon. The Jewell member is lithologically similar to the lower two members of the type section Keasey Formation, but tends to be thinner, better laminated, darker colored, and contains clastic dikes. Olbinski (1983) restricted the Keasey Formation in eastern Clatsop County to the Jewell member. Nelson (1985), however, divided the Keasey Formation adjacent to the thesis area into three informal members: 1) the basal Jewell member; 2) the Vesper Church member; and 3) an upper mudstone member. The Vesper Church member of Nelson (1985) is correlative to the Vesper Church formation of Olbinski

(1983) which has recently been redefined as the Sager Creek formation by Niem and Niem (in press). The upper mudstone member of Nelson (1985) has been included in the Oswald West mudstone (Smuggler Cove formation of this study) by Peterson (1984) and Niem and Niem (in press). The upper mudstone member of the Keasey Formation of Nelson (1985) is a lithologically distinct unit which is thought to be a facies equivalent of the Pittsburg Bluff Formation and, therefore, was not included in the Keasey Formation in this study.

Following the nomenclature of Olbinski (1983) the Keasey formation in the study area has been restricted to the Jewell member. Lower Smuggler Cove formation mudstones mapped in the thesis area are probably, in part, correlative to the upper Keasey Formation near the type area (see fig. 6). The Refugian lower Smuggler Cove formation mudstones were not included in the Keasey Formation because of their lithologic similarity to upper Smuggler Cove formation mudstones and because of the difficulty in distinguishing lateral bathyal facies of the Keasey Formation from overlying lateral bathyal facies of the Pittsburg Bluff Formation and the Vesper Church formation (Sager Creek formation of Niem and Niem, in press).

The Jewell member of the Keasey Formation crops out throughout the central part of the thesis area (fig. 4 and plate 1). The best exposures of the unit occur one kilometer northeast of Hamlet along logging roads (localities 179, 180, 181 and 191), in the North Fork of the Nehalem River (sec. 25, T4N, R9W), and near Rector Ridge (localities 377 and 382). The unit typically forms low hills and hummocky topography. It is commonly subject to landsliding. The thickness of the Jewell member in the thesis area is difficult to ascertain accurately, owing to moderately poor exposures. Field estimates indicate a thickness of 500 to 800 m. In the CZ 11-28 well, adjacent to the thesis area, the unit is approximately 750 m thick (plate 3). Olbinski (1983) reported that the Jewell member was 365 m thick in the Quintana Watzek 30-1 well.

Lithology

The Jewell member consists of thinly bedded to laminated tuffaceous mudstone and rare fine-grained arkosic sandstone beds and clastic dikes (fig. 69). Fresh mudstones are dark gray (N 3) but are typically weathered to a very pale orange (10YR 8/2). The mudstones commonly contain thin (<15cm) tuff beds (fig. 70) which are very light gray (N 8) to grayish yellow (5Y 8/4) and may contain faint laminations. Irregularly shaped small (1/2m dia.) calcareous concretions are rarely present (e.g., locality 535). Several fragmented and weathered molluskan fossils were collected from the unit, but they were too poorly preserved for further identification. Helmenthoida trace fossils are locally abundant and larger foraminifera such as Cyclamina can be seen in some hand samples. At the base of the unit, directly above the Cole Mountain basalts, the mudstones are very tuffaceous, light colored (N 8), and distinctly bedded to laminated (e.g. localities 162 and 659). Small (<1cm) pumice fragments are sparsely scattered in these highly tuffaceous mudstones. A debris flow deposit composed of well-laminated, highly tuffaceous siltstone blocks was found at one locality (162) at the

Fig. 69. Good exposure of dark colored, bedded, and laminated mudstones in the lower-middle part of the Jewell member. A glauconitic sandstone bed, which is difficult to see, occurs near the middle of the outcrop (locality 191, NE 1/4 SW 1/4 sec. 4, T4N, R8W).

Fig. 70. Thin tuff beds in the upper part of the Jewell member (locality 170). Note typical chippy weathering of the Jewell member.

base of the Jewell member (fig. 67).

Scattered thin (<1m) glauconitic sandstone and glauconitic siltstone beds are generally present in the basal part of the Jewell member above the highly tuffaceous mudstone. The glauconite in these sandstones and siltstones is generally coarse sand-sized and is in both grain and matrix support. These beds may be highly bioturbated, resulting in distribution of the glauconite into surrounding mudstones. The glauconitic sandstone and siltstone, which comprises less than 5% of the Jewell member, is best exposed at localities 191, 401, 377, and 827.

Approximately 30 meters above the glauconite beds there are a few arkosic sandstone channels and clastic sandstone dikes. The sandstone channels are lenticular with dimensions approximately 2 1/2 m wide by 1/2 m deep (fig. 71). The sandstones in these channels are grayish orange (10YR 7/4), moderately well-sorted, fine-grained, and massive. Sandstone channels were found in only three exposures, within the thesis area (localities 179 and 180), but thin beds and laminae of arkosic sandstone occur at localities 137 and 145. These sandstone beds and laminae are bioturbated and lithologically similar to the arkosic sandstones. Clastic arkosic sandstone dikes are common throughout the "lower middle" part of the Jewell member (localities 181, 353, 377, 382, 532, and 623). The clastic dikes are typically thin (<1/2m), nearly vertical, generally oriented northwesterly, and are lithologically identical to arkosic channel sandstones, indicating that they are a result of rapid loading and liquefication of the channel sands.

The Jewell member to the east of the thesis area (Olbinski,

Fig. 71. Lenticular arkosic sandstone channel in the Jewell member (locality 179, SW 1/4 NE 1/4 sec. 4, T4N, R8W).

1983; Nelson, 1985; Mumford, in prep.) is lithologically similar to the Jewell member in the thesis area. The differences being that Olbinski (1983) found basaltic sandstones near the base of the unit and arkosic sandstone channels have been found only within the thesis area. The presence of arkosic clastic dikes in adjacent areas suggests that the sandstone channels are present but have not been observed. Olbinski (1983) reported the presence of thin arkosic sandstone beds in the Jewell member in the Quintana Watzek 30-1 well, located in eastern Clatsop County, and Martin et al. (in press) reported fine-grained basaltic conglomerates from near the base of the Jewell member in the Boise Cascade well, located in northeastern Clatsop County.

The Jewell member is more tuffaceous, less carbonaceous, and less micaceous than the Sweet Home Creek mudstone member of the Hamlet formation. In addition tuff beds, clastic dikes, glauconitic sandstones to siltstones, and small arkosic sandstone channels appear to be restricted to the Jewell member. Therefore, even though both the Jewell member and the Sweet Home Creek member are composed primarily of mudstone they can be distinguished in the field. It is, however, generally difficult to distinguish between upper Jewell member mudstones and the overlying Smuggler Cove formation mudstones. The upper part of the Jewell member is typically faintly bedded whereas the lower Smuggler Cove formation mudstones are more massive and bioturbated. In highly weathered exposures it is not always possible to distinguish between the two units.

Contact Relations

The Jewell member is unconformable upon the Cole Mountain basalts and the Sweet Home Creek member of the Hamlet formation. This contact was discussed in the Cole Mountain basalt section. The upper contact of the Jewell member is conformable and gradational with the overlying lower Smuggler Cove formation mudstones. Although this contact is not exposed in the thesis area the absence of an unconformity is evidenced by the similar ages, depositional environments, lithologies, and structural attitudes of the units.

Age

The Jewell member in the thesis area is lower Refugian (late Eocene) in age. Over fifty different benthic foraminifera species from fourteen surface and subsurface localities were collected from the Jewell member and identified by Kristin McDougall (U.S. Geological Survey) and Weldon Rau (Wash. Dept. of Natural Resources) (appendices 1 and 2). Three surface localites contain age-diagnostic foraminiferal faunas which were assigned to the lower Refugian stage (McDougall, pers. comm., 1983) (appendix 2). Two of these localities were from the base of the unit (659 and 535); the third from the top of the unit (528). Calcareous nannofossils collected from locality 529 indicate a late Eocene to early Oligocene age which is consistent with the foraminiferal data (Bukry, personal communication, 1983). The Jewell member foraminiferal fauna is similar to the fauna in the type section Keasey Fomation and is considered to be primarily "lower" Refugian (McDougall, pers. comm., 1983).

McDougall (1975) made the latest and most deatiled biostratigraphic study of the type Keasey Formation, located in western Columbia County along Rock Creek. She considered the very base of the formation to be late Narizian in age with the remainder of the unit being assigned to the Refugian stage. In adjacent eastern Clatsop County, Niem (personal communication, 1984) has collected a sample of foraminifers from the type locality of the basal Jewell member which Rau assigned to the late Narizian whereas McKeel and McDougall assigned the foraminifera to the lower Refugian. This suggests that the basal Keasey Formation straddles the Narizian-Refugian boundary. Samples from near the basal contact of the Jewell member in the thesis area contain a lower Refugian fauna. In conclusion, the Keasey Formation and the Jewell member of the Keasey Formation are primarily Refugian in age with the basal part of the unit straddling the Narizian-Refugian boundary.

Correlation

The Jewell member is correlative to the lower and middle part of the type Keasey Formation in Columbia County. The glauconite-rich "Nehalem formation" (mapped in southwestern Columbia County) of Deacon (1953) is probably correlative to the lower part of the Jewell member. The Jewell member may also be correlative, in part, to the Narizian upper mudstone member of the Cowlitz Formation of Bruer et al. (1984) in the subsurface of Clatsop and Columbia

Counties. In the Quintana Watzek 30-1 well a thin section of micaceous mudstones overlies the Cowlitz Formation sandstones and underlies the glauconitic sandstones of the basal Jewell member (fig. 68). I prefer to assign these micaceous mudstones to the Cowlitz Formation. Bruer et al. (1984), however, have depicted a very thick section of mudstone in the upper Cowlitz Formation in the subsurface of Clatsop and Columbia counties. These mudstones were assigned to the Cowlitz Formation, at least in part, because they contain a "Narizian" fauna. Since there appears to be some discrepancy between workers on the precise location of the Narizian-Refugian boundary, it is possible that some of these mudstones are correlative to the Jewell member. Martin et al. (in press) have recently mapped Jewell member mudstones in the subsurface of Clatsop County where Bruer et al. (1984) had mapped upper Cowlitz Formation mudstones.

South of the thesis area, Wells et al. (1983) have mapped undifferentiated Refugian mudstones that are, in part, correlative to the Jewell member. Regionally, the Keasey Formation has been correlated to the basal part of the Lincoln Creek Formation of southwest Washington, the basal part of the Alsea Formation of the central Oregon coast, the Bastendorff Formation of the southern Oregon coast, and the Townsend shale of the Olympic peninsula (Armentrout et al., 1983; Cushman and Schenck, 1928; Durham, 1945) (fig. 5).

Petrology

Modal analyses were performed on nine thin-sections and seven heavy mineral grain mounts of Jewell member arkosic sandstones (appendices). Glauconitic sandstones were not examined. A number of mudstone smear slides were also prepared and examined. Scanning electron microscopy and energy dispersive X-ray (EDX) analyses were performed on an arkosic sandstone channel sample (locality 179). The sandstones examined are matrix rich (15% matrix) and, therefore, are considered to be wackes (Williams et al., 1954). Sandstones in the Jewell member classify as arkosic wackes, feldspathic wackes, and subfeldspathic lithic wackes using the classification scheme of Williams et al. (1954) (fig. 72). Much of the matrix in the Jewell member sandstones appears to be authogenic and, therefore, these sandstones were probably arenites at the time of deposition. The sandstones classify as lithic arkoses, feldspathic litharenites, and subarkoses on Folks 1980 classification if matrix is excluded (fig. 72). The average composition of the sandstones is: 38% monocrystalline quartz (both strained and unstrained), 4% polycrystalline quartz, 4% plagioclase (An 15-55), 9% orthoclase, 1% microcline, 1% chert, 6% biotite, 3% muscovite, 1% horneblende, 8% metamorphic rock fragments, 4% andesitic volcanic rock fragments, 17% matrix, and approximately 5% porosity (appendix 9) (fig.73). Framework grains are subangular to subrounded with rare well-rounded clasts, fine-to very fine-grained, and moderately well-sorted.

Figure 72 and appendix 9 show that the arkosic sandstones in the Jewell member all have similar composition. Furthermore, there are

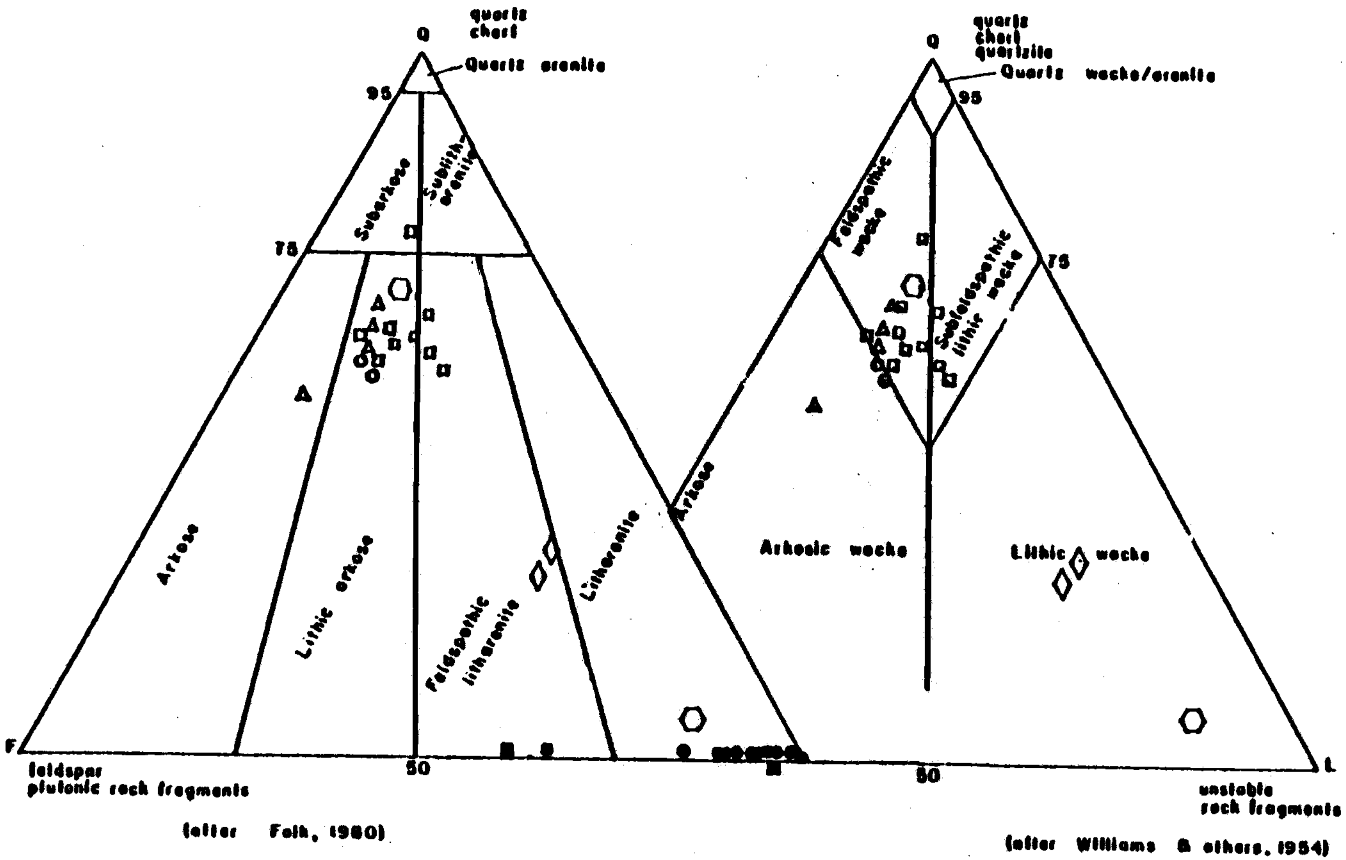


Fig. 72. Classification diagrams for sandstone samples from the thesis area. The left hand diagram is for samples with less than 15% matrix whereas the right hand diagram is for samples with more than 15% matrix. Samples with abundant diagenetic matrix were plotted on both diagrams. Roy Creek member (●), Sweet Home Creek member (■), Jewell member (□), Smuggler Cove formation (⬡), Pittsburg Bluff Formation (◇), ball park unit of the S.C. fm. (A), and Astoria Formation (O) samples have been plotted.

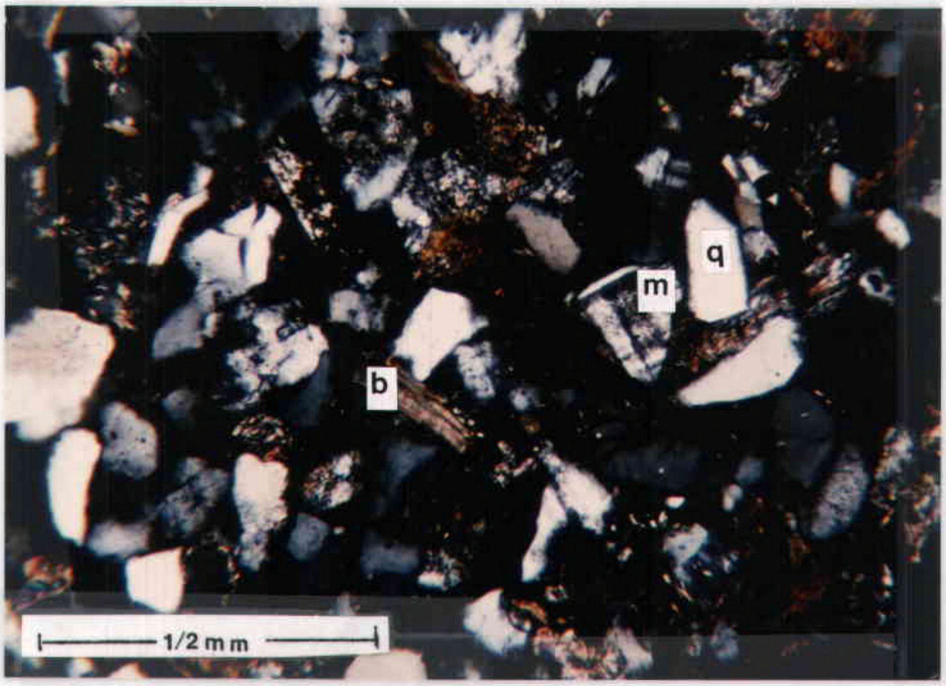


Fig. 73. Photomicrograph of arkosic sandstone sample 179 from the Jewell member (crossed nicols). Note the presence of biotite (b), angular microcline (m), and monocrystalline quartz (q).

no compositional differences between clastic dikes and channel sandstone. The composition of the Jewell member arkosic sandstones in the thesis area is very similar to the composition of Jewell member sandstone reported by Olbinski (1983) and Nelson (1985) and to Cowlitz Formation sandstone reported by Olbinski (1983) and Van Atta (1971). Abundant biotite (avg. 6%) in the Jewell member sandstones serves to distinguish them from other arkosic sandstones in the thesis area.

Common heavy minerals (specific gravity 2.92) in the Jewell member are, in decreasing abundance: Opaque minerals, epidote, biotite, muscovite, zircon, garnet, hornblende, tourmaline, hypersthene, augite, apatite, sphene, rutile, clinozoisite, and staurolite (appendix 10) (fig.74). Trace amounts of diopside, lambrobolite, tremolite, orange zircon, kyanite, monazite, andalusite, and sillimanite are present. Heavy minerals compose from 0.2 to 0.8% of the 3ϕ to 4ϕ size fraction and are typically angular to subangular, but rare grains of well-rounded abrasionally resistant zircon, garnet, and tourmaline are present suggesting some recycling from sedimentary sources. The heavy mineral assemblages from the 7 samples examined are all very similar and there is no apparent difference between clastic dike samples and channel sandstone samples suggesting a common origin. Olbinski (1983), Nelson (1985), and Van Atta (1971) have reported heavy mineral assemblages from the Cowlitz Formation that are similar to the Jewell member assemblage.

Smear slides show that the mudstones consist of clay minerals and less common silt-sized quartz, feldspar, mica, and volcanic glass. Rare marine diatoms (locality 533, appendix 1) and a few

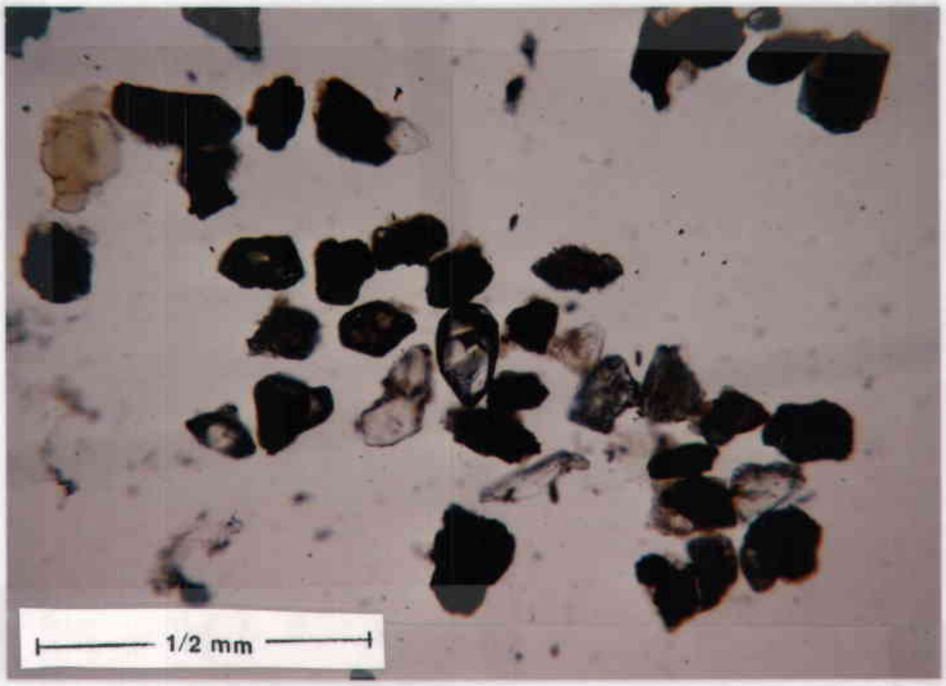


Fig. 74. Heavy mineral assemblage from arkosic sandstone sample 179 in the Jewell member (plane polarized light). Note the presence of euhedral zircon, biotite, and epidote.

calcareous nannofossils (localities 529, 533, 2116-10, and 2116-17) are present. A number of mudstone samples were disaggregated and the greater than 4ϕ size fraction was examined. Benthic foraminifera, rare plankonic foraminifera, rare fish teeth, rare geodites (fish ear bones), and rare very fine-grained silicate minerals comprise the sand-sized fraction. Benthic foraminifera comprise 0 to 2% of the rock with other constituents occurring in trace amounts. Heavy minerals from a mudstone sample (2116-21) consist of very rare biotite, very rare apatite, and abundant opaque minerals. Most opaque minerals are pyrite and pyritized microfossils (appendix 10).

Scanning electron microscopy (SEM) was performed on a channel sandstone (locality 179). The extensive clay matrix observed in thin-section can easily be seen in SEM photographs (fig. 75). Clay matrix has destroyed most primary porosity with clays being both detrital and diagenetic in origin. Some of the clay has a highly crenulated to honeycomb morphology indicative of authigenic formation whereas other clays are massive to irregular suggesting a detrital origin (see Welton, 1984) (fig. 75). Thin-section analysis shows that the channel sandstones have been slightly bioturbated and that injection of clastic dikes into mudstones resulted in incorporation of adjacent muds into the sand. Energy dispersive x-ray analysis (EDX) on the clay in an arkosic sandstone (sample 179) suggests that the clay is primarily either kaolinite or smectite or a combination of both. This is evidenced by EDX patterns showing abundant Si and Al with minor Fe and Ca. EDX analysis of a mudstone sample (locality 529) showed abundant Si and Al as well as fairly common Mg, Fe, K, Ca, and Ti. Smectite clays generally have a chemical composition

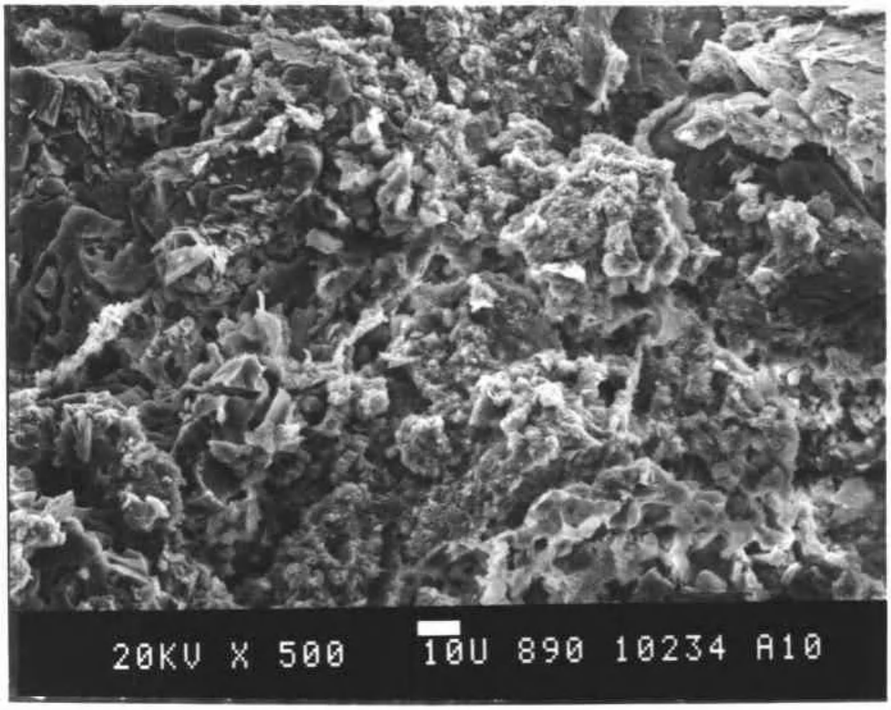


Fig. 75. SEM photograph of arkosic sandstone in the Jewell member (sample 179). Porosity and permeability are low because of extensive clay infilling.

similar to this (Welton, 1984). Mumford (in prep.) using X-ray diffraction analysis has found that Jewell member mudstones are composed primarily of smectite and characteristically contain the zeolite clinoptilite, which is a common alteration product of volcanic glass.

The following diagenetic sequence was observed in the Jewell member sandstones: 1) mechanical crushing of volcanic rock fragments and micas; 2) development of authigenic smectite clay minerals as evidenced be scanning electron microscopy and partial dissolution of feldspars; and 3) minor teleogenetic oxidation of iron-bearing minerals such as augite and plagioclase. No pore-filling cements, silica overgrowth, and authogenic zeolites were observed. Mechanical crushing is minor, probably a result of abundant detrital clay, and evidenced by slightly deformed mica. The diagenetic sequence observed in the Jewell member sandstones is similar to that of shallow buried Tertiary sandstones in Wyoming (Stanley and Benson, 1979), arc derived Tertiary sandstones in western Washington (Galloway, 1979), and the Miocene Astoria Formation in northwest Oregon (Cooper, 1981; Peterson, 1984). These sandstones are similar to the Jewell member sandstones in that they lack significant quartz, carbonate, and zeolite cements; are cemented by "smectite" clays; and are moderately friable.

Provenance

Source terranes for the Jewell member included igneous, metamorphic, and sedimentary rocks. An andesitic source is indicated by the presence of andesitic rock fragments, andesine, hypersthene, green hornblende, lamprobolite, and tuff beds. The developing Cascade arc to the east was probably the source for much of the andesitic detritus. A minor basaltic source area is indicated by the presence of basaltic rock fragments, augite, magnetite, and leucoxene. The western Cascade arc may also have been the source of the basaltic detritus. It is unlikely that the Tillamook Volcanics were subaerialy exposed during deposition of the Jewell member and, therefore, were probably not a source of the basaltic detritus. It is, however, possible that some of the volcanic detritus at the base of the unit was derived from the Cole Mountain-Goble volcanics.

An important low to high rank metamorphic source is indicated by the presence of abundant biotite, epidote, and garnet as well as by a few grains of sillimanite, andalusite, kyanite, clinozoisite, diopside, tremolite, and staurolite. Rare phyllitic, schistose, and gneissic rock fragments were observed in the Jewell member sandstones. The presence of microcline, orthoclase, euhedral zircon, and tourmaline (schorlite) suggest a minor acid igneous source for Jewell member. The Northern Cascades of Washington (Misch, 1966), the Klamath terrane of southern Oregon and northern California (Scheidegger et al., 1971), and the Mesozoic Idaho and Wallowa batholith are probable source areas for the metamorphic and acid igneous detritus.

Well rounded zircon and very well-rounded quartz were found in several samples. This indicates that they were derived from sedimentary rocks as several cycles of deposition are needed to round these minerals (Pettijohn, 1974). Since well-rounded clasts are rare in the Jewell member, recycled sedimentary rocks probably contributed little detritus. The Cowlitz Formation sandstones, as previously noted, are lithologically similar to the Jewell member sandstones. Therefore, it is possible that much or all of the Jewell member sandstones were derived by reworking of Cowlitz Formation "sands" at the unconformity between the units. The Cowlitz Formation sandstones would have been poorly lithified or non-lithified at this time and transportation would have resulted in little textural change.

In conclusion, the Jewell member arkosic sandstones were derived primarily from metamorphic and acid igneous source areas with lesser contributions from intermediate volcanic and sandstone sources. Jewell member mudstones are tuffaceous and contain common tuff beds indicating that they were primarily derived from explosive intermediate to silicic volcanic sources such as the western Cascade arc (e.g. Little Butte Volcanics of Wells and Peck, 1961). Van Atta (1971) noted that the Cowlitz Formation sandstones, which are lithologically similar to Jewell member sandstones, were derived from plutonic and metamorphic terrains in Oregon, Washington, and Idaho and were transported via an ancestral Columbia River to northwest Oregon. A similar source is likely for much of the Jewell member sandstones. In other words, the ancestral Columbia River system that supplied detritus to the Narizian Cowlitz Formation continued into the Refugian supplying minor sand size detritus to the Jewell member. Finer-grained tuffaceous detritus from the developing Cascade arc did, however, become more abundant during the Refugian partially masking the Columbia River source. As previously mentioned, it is possible that the arkosic sands in the Jewell member were derived from reworking of Cowlitz Formation "sands" but, unfortunately, this hypothesis is difficult to prove.

Dickinson and Suczek (1979) devised a classification scheme of detrital framework modes for sandstones that can be used to help determine plate tectonic setting. Jewell member sandstones plot within the recycled orogen provenance field on the quartz-feldspar-lithic triangular diagram of Dickinson and Suczek (1979) (fig. 38). The recycled orogen provenance is defined as having sources from uplifted subduction zones, along collision orogenies, or within foreland fold-thrust belts. Data from this study and from Olbinski (1983) show that the Jewell member was deposited in a forearc basin. These conflicting conclusions support the hypothesis that an ancestral Columbia River system which recieved detritus from a "recycled orogen provenance" (e.g., Rocky Mountain, North Cascades, Klamath terrain) and bypassed the western Cascade arc depositing some coarse detritus in a forearc basin.

Grain Size Analysis

Nine samples of Jewell member sandstone were disaggregated and sieved. Cumulative frequency curves were constructed and the statistical parameters of Folk and Ward (1957) were calculated (appendix 11). The statistical parameters were calculated both with matrix and matrix-free. The matrix free calculation eliminates most post-depositional textural changes (e.g., authigenic clay and mechanical emplacement of clay but do not take into account detrital clay.). The grain size characteristics of the sand at the time of deposition were somewhere between the matrix and matrix-free calculations. In most cases, the difference between the two calculations is minor (appendix 11).

Jewell member sandstones are very fine-to fine-grained (mean 2.5ø to 3.8ø), usually moderately sorted (std. dev. 0.42 to 1.37) and have highly variable skewness and kurtosis values. There are no apparent differences in grain size distribution between the channel sandstone (179) and the clastic dike samples suggesting a common origin. The only trend noted is that the sandstones in the southern and eastern part of the thesis area are usually coarser than those in the northwest part of the thesis area.

Jewell member sandstones were plotted on the coarsest 1% versus median diameter plot of Passega (1957) as an aid in determining depositional environment (fig. 47). The samples plot as a fairly tight group with most samples plotting in or near the tractive current field. The Jewell member sandstones have grain size distributions similar to the upper slope and outer shelf sands off Oregon reported by Kulm et al. (1975) and Kulm and Scheidegger (1979).

Depositional Environment

Environmentally diagnostic benthic foraminiferal assemblages were collected from three Jewell member samples in the thesis area (appendices 1 & 2). These foraminifera were interpreted to represent upper to "lower" bathyal deposition (200-2,000 m). Foraminifera collected from the CZ 11-28 well adjacent to the thesis area were interpreted as bathyal by both Weldon Rau (Wash. Div. of Natural Res.) and Kristin McDougall (U.S. Geological Survery). The two workers examined the same samples, but arrived at slightly different bathyal interpretations. For example, Rau considered the fauna in sample 528 to be middle bathyal whereas McDougall considered it to be lower bathyal. The discrepancy is probably due to the difficulty in subdividing the bathyal biofacies (as discussed in Hamlet formation section). Nelson (1985) and Mumford (in prep.) have collected upper bathyal foraminifera from the Jewell member to the east of the thesis area suggesting a slight shallowing trend towards the east. Diatoms collected from locality 529 indicate deposition near the self.edge (approx. 200-300m) (Barron, personal communication, 1984). The presence of common Helmenthoida trace fossils is consistant with a bathyal depositional environment (Chamberlain, pers. comm. to Nelson, 1985).

The composition of Jewell member mudstones is similar to recent upper slope muds (Kulm and Scheidegger, 1979) and to ancient upper slope mudstones (Dott and Bird, 1979; McDougall, 1980). The Jewell member is primarily composed of clay minerals and less common silt-sized silicate minerals. Microfossils compose about 2% of the mudstones with benthic foraminifera being the most abundant. A few calcareous nannofossils, rare planktic foraminifera, and rare marine diatoms are also present. Jewell member mudstones have a moderately high species diversity (8-20 species per sample). Kulm and Scheidegger (1979), Dott and Bird (1979), and McDougall (1980) considered these characterisitics to be indicative of upper slope deposition.

Both glauconitic and arkosic sandstones are present in the Jewell member. Gauconite is currently forming on the outer continental shelf and upper slope of Oregon under slow or negative sedimentation rates (Kulm et al., 1975) McRae (1972) considered glauconite to be "characteristic" of outer shelf depositional environments. The arkosic sandstones have a grain size distribution similar to outer shelf and upper slope sands (see previous section). The rare channel sandstones probably represent very small "sea gullies" crossing the outer shelf and upper slope. Dott and Bird (1979) reported larger shelf-slope channel sandstones from the middle Eocene strata of southwestern Oregon. They suggested a model where an array of "sea gullies" served to transport sand from a deltaic complex across the shelf and slope to feed deep sea fans. The rare channel sandstones in the Jewell member may have had a similar origin but at a smaller scale. It is also possible that the Jewell member channels were sourced from relict outer shelf sands that moved downslope as sediment gravity flows. Downslope the sands may have overflowed the channels resulting in the thin overbank sandstone beds observed at localities 137 and 145. Rapid deposition and compaction of Jewell member muds resulted in overpressuring the water-saturated channel sands and subsequent injection of clastic dikes into muds. Clastic dikes are more common than sandstone channels, suggesting that most channels were destroyed by overpressuring.

In conclusion, the Jewell member was deposited in an "upper" slope (upper bathyal) setting. Rare small channelized arkosic sand sediment gravity flows interrupted the normal hemipelagic sedimentation. Explosive andesitic to silicic Volcanism in the

Cascade arc resulted in deposition of thin tuff beds. The tuff beds may have been emplaced by sediment gravity flows as well as by settling of fine ash through the water column. Nelson (1985) reported a few shallow marine foraminifera in the deep marine mudstones of the Jewell member, indicating that deposition by silty turbidity flows occurred occasionally.

There is some suggestion of a deepening upward trend in the Jewell member indicating that basin subsidence was greater than the sedimentation rate. Foraminifera at the base of the unit indicate an "upper" bathyal environment whereas foraminifera from near the top of the unit indicate a "middle to lower" bathyal environment (appendix 2). In addition glauconite, which occurs at the base of the unit, is most common in modern outer shelf, slightly reducing depositional settings. The presence of fine laminations and lack of extensive bioturbation in the lower part of the unit are also suggestive of a slightly reducing depositional environment. Benthic foraminifera show that the environment was not strongly reducing. The type Keasey Formation, located some 15 km east of the thesis area, was deposited in an environment similar to but slightly shallower (outer shelf to upper slope) than the Jewell member (McDougall, 1975).

SMUGGLER COVE FORMATION

Nomenclature and Distribution

Niem and Niem (in press) have recently proposed that the name Smuggler Cove formation replace the informal name Oswald West mudstone. This was done so that terminology would be in accordance with the Code of Stratigraphic Nomenclature and would meet the guidelines of the Geologic Names Committee of the U.S. Geological Survey. The name Smuggler Cove will be used in this report.

The Oswald West mudstone was proposed by Cressy (1974) and Niem and Van Atta (1973) for a sequence of bathyal, Oligocene to lower Miocene, thick bedded, tuffaceous silty mudstone and minor sandstone exposed at Short Sands Beach in Oswald West State Park. The definition of the unit was subsequently expanded to include lithologically similar, upper Eocene (Refugian to Narizian?) mudstones below the type section strata in adjacent areas (Smith, 1975; Neel, 1976; Tolson, 1976; Penoyer, 1977; M. Nelson, 1978; Peterson, 1984; D. Nelson, 1985). Murphy (1981) considered the base of the Oswald West mudstone to be the top of the Pittsburg Bluff Formation in northeastern Clatsop County. Peterson (1984), however, found that this definition was inadequate as the Pittsburg Bluff Formation sandstones pinch-out into mudstones west of the area mapped by Murphy (1981) (fig. 6). Peterson (1984) included mudstones that are laterally correlative to the Pittsburg Bluff Formation, as well as older mudstones correlative to the Keasey Formation, in the Oswald West mudstone.

Smith (1975) and Neel (1976) divided the Oswald West mudstone into three informal "parts" which, because of subtle lithologic differences between the parts, they were not able to map. The "upper part" of Smith (1975) and Neel (1976) is Zemorrian in age and is thought to be correlative to the to the type section of the Oswald West mudstone at Short Sands Beach. The "upper part" consists of thick-bedded, moderately well-indurated, bioturbated, tuffaceous siltstone, mudstone, and minor sandstone. The "middle part" (uppermost Refugian) is composed of thick-bedded to massive, moderately-indurated, bioturbated tuffaceous siltstone and glauconitic sandstone. The "lower part" consists of poorly-indurated, bioturbated, tuffaceous siltstone and mudstone. Peterson (1984) was able to recognize the above units to the northeast of the area mapped by Smith (1975) and Neel (1976) (see fig. 1).

In the present study area, the Smuggler Cove formation (Oswald West mudstone of previous workers) has been divided into and mapped as four informal units: the lower member of the Smuggler Cove formation; the glauconitic sandstone member; the upper member of the Smuggler Cove formation; and the "ball park unit" (plate 1, figs. 4 and 6). These subdivisions are similar to the subdivisions of Peterson (1984), Neel (1976), and Smith (1975) except that lower Refugian mudstones have been included in the Jewell member of the Keasey Formation rather than the Smuggler Cove formation and that the ball park unit, which is thought to be correlative to the lower Silver Point member of the Astoria Formation of Peterson (1984) (Northrup Creek formation of Niem and Niem, in press), has been included in the upper part of the Smuggler Cove formation. The lower Refugian mudstones form a lithologically distinct unit, which can be traced from the type area of the Keasey Formation, in Columbia County, to the thesis area and, therefore, in this writer's opinion these mudstones should not be included in the Smuggler Cove formation. Niem and Niem (in press) have suggested that the ball park unit (Tscm) be included in the Smuggler Cove formation.

The Smuggler Cove formation (Oswald West mudstone) has traditionally been a difficult unit to define. To the east, in Columbia County, several lithologically distinct units are present in the in the upper Eocene to lower Miocene stratigraphic section (e.g., Keasey, Pittsburg Bluff, and Scappoose formations of Warren and Norbisrath, 1946) but these units pinch-out into thick, deeper marine, lithologically similar mudstones to the west making it impractical to to retain the nomenclature of Columbia County (fig. 6). It is felt that use of the term Smuggler Cove formation for the thick section of of uppermost Eocene to lower Miocene mudstones and siltstones in Clatsop County is the best solution to the stratigraphic problem. The subdivision of the Smuggler Cove formation into informal units serves to help correlate the stratigraphy in Clatsop County to the stratigraphy in Columbia County and to better interpret depositional environments.

The distribution of the Smuggler Cove formation is restricted to the northern and western parts of the thesis area (fig. 4 and plate I). The best exposures of the lower member of the Smuggler Cove formation occur at localities 124, 142, 329, and 766 (plate I). The glauconitic sandstone member is best exposed at localities 8a, 40a, 46, 93, and 99 (locality 99 is along Hamlet Road at the western boundary of the thesis area) and the upper member of the Smuggler Cove formation is exceptionally well exposed at localities 584 and 584a. The ball park unit is well exposed at locality 930 along an unnamed logging road in the southwestern part of the thesis area. The Smuggler Cove formation typically forms low relief, hummocky topography with the glauconitic sandstone member locally forming small ridges. The formation is approximately 900 meters thick in the thesis area. The thickness is difficult to estimate because of moderately poor exposures and the unconformity at the top of the unit with the overlying Astoria Formation. Tolson (1976) reported a maximum thickness of 1,500 m for the unit in the Standard Hoagland #1 exploration well, and Peterson (1984) reported a thickness of at least 1,663 m from the Diamond Shamrock 30-1 exploration well. Tolson (1976) and Peterson (1984), however, included some strata that have been assigned to the Keasey Formation in this report and did not include strata correlative to the ball park unit. Over 100 km2 of Smuggler Cove formation have been mapped in the surface and subsurface of Clatsop County (Niem and Niem, in press).

The lower three members (lower, glauconitic sandstone, and upper) in the Smuggler Cove formation will be discussed separately from from the ball park unit which, in the thesis area, overlies the other members. The proper formational assignment of the ball park unit is uncertain and in some respects the ball park unit is lithologically distinct from the lower three units.

Lower Member, Glauconitic Sandstone Member, and Upper Member

Lithology

The lower member of the Smuggler Cove formation in the thesis area consists of massive, bioturbated, poorly indurated, tuffaceous silty mudstone and some waterlaid tuff beds. The mudstones are medium gray (N 4) when fresh and weather to grayish orange (10YR 7/4). Outcrops of this unit are generally poor, small, and are characterized by abundant chippy (1 cm) mudstone talus (fig. 76).

The glauconitic sandstone member consists of a basal sequence (approx. 5 m thick) of very well-laminated, slightly micaceous, silty mudstone which is best exposed at localities 29, 48, and 117 in the northern part of the thesis area (fig. 77). Overlying the basal mudstone is a sequence (approx. 25 m thick) of glauconitic sandstone beds, glauconitic mudstone, thin tuff beds, and tuffaceous siltstone. The glauconite in this unit is generally fine sand-sized and rounded in contrast to the coarser grained, more angular glauconite aggregates found dispersed in mudstones and thin sandstone beds at the base of the Jewell member. Smuggler Cove formation glauconitic sandstone beds in the thesis area are relatively thin (<2m), commonly bioturbated, some are calcite-cemented, and are dusky yellow green (5GY 5/2) to light olive gray (5Y 5/2) in outcrop. A few miles west of the thesis area correlative glauconitic sandstone beds in the Smuggler Cove formation are several meters thick and have been mapped over a large area (Smith, 1975; Neel, 1976, Niem and Niem, in press). Directly overlying a glauconitic

Fig. 76. Typical exposures of the lower and middle parts of the Smuggler Cove formation. A) is a mudstone and tuff bed exposure from the lower Smuggler Cove formation (locality 99b, SW 1/4 NW 1/4 sec. 36, T5N, R8W). B) shows typical chippy mudstone talus below a glauconite bed at locality 99 (SW 1/4 Nw 1/4 sec. 36, T5N, R8W).

Fig. 77. Typical exposure of thinly-laminated, shaly siltstone and mudstone at the base of the glauconitic sandstone unit (locality 48, SE 1/4 NE 1/4 sec. 20, T5N, R8W).

sandstone bed at locality 99 along Hamlet Road is a sequence of 9 thin (20 cm), buff colored tuff beds. The tuff beds are approximately 1/2 meter apart with massive tuffaceous mudstone occurring between the tuff beds. Above the mudstones and tuff beds is a sequence of bioturbated glauconitic mudstones which range from 5 to 20 m in thickness. The top of the glauconitic sandstone member is defined as the last occurrence of glauconitic mudstone.

The base of the upper member of the Smuggler Cove formation consists of fossiliferous, tuffaceous, bioturbated, concretionary siltstones and silty mudstones. Large benthic foraminifera (e.g., Cyclamina) are very abundant and can be seen easily in hand samples. Bivalves, gastropods, and scaphopods are fairly common throughout the strata with the bivalve Delectopecten being especially common (appendix 1). Large pieces (3-4 cm) of carbonized wood and fern fronds are present locally in the northern part of the thesis area (e.g., localities 13, 22, and 41) (fig. 78). Spherical to ellipsoidal calcareous concretions (15 cm to 2 m) are locally abundant (fig. 79). Some concretions contain a significant amount of reddish iron oxide and others may contain molluscan fossils at the nucleus. Mudrocks in these upper Smuggler Cove formation strata consist of bioturbated, medium gray (N 4) to grayish orange (10YR 7/4) siltstone to silty mudstone. They tend to be slightly coarser-grained than mudrocks at other stratigraphic horizons in the Smuggler Cove formation.

The middle and upper parts of the upper member consist of moderately well-indurated, resistant, thick-bedded, bioturbated, tuffaceous silty mudstone and rare arkosic to lithic sandstone beds

Fig. 78. Carbonaceous debris in lower part of the upper Smuggler Cove formation.

Fig. 79. Large calcareous concretion in the lower part of the Upper Smuggler Cove formation (locality 151).

(fig. 80). The mudstone is medium dark gray (N 4) when fresh and weathers to grayish orange (10YR 7/4). Helmenthoida fecal traces are fairly abundant. Mudstones are very well-indurated near thick intrusions of Columbia River Basalt. This baking may extend some 75 meters above and below the intrusions. Adjacent to the intrusions the mudstones develop a porphyroblastic texture due to contact metamorphism. Rare 1/2 meter thick lithic sandstone beds (locality 451) and rare arkosic sandstone dikes (locality 580) occur within the mudstones. The thick-bedding, good induration, and abundance of Helmenthoida trace fossils serve to distinguish the upper part of the upper member from other mudstone units in the thesis area.

Contact Relations

The lower Smuggler Cove formation is conformable upon the underlying lower Refugian Jewell member of the Keasey Formation as is evidenced by the gradational contact between the units and by the similar ages, depositional environments, and structural attitudes of the units. The contact between these units is not exposed in the thesis area.

Previous workers (e.g. Cressy, 1974; Neel, 1976; Peterson, 1984) have noted that the Astoria Formation is unconformable upon the Smuggler Cove formation. The abrupt change in lithologies and depositional environments (bathyal to fluvial), differing structural attitudes, a sharp contact between the units, and the fact that Astoria Formation overlies Smuggler Cove formation mudstones of differing ages demonstrates that the contact is an angular

Fig. 80. Exposure of well-indurated, thick-bedded mudstones of the upper Smuggler Cove formation. Waterfall is approximately 10 meters high (locality 584, NW 1/4 NW 1/4 sec. 32, T4N, R9W).

unconformity (Cressy, 1974; Murphy, 1981; Peterson, 1984).

The Astoria Formation crops out in only a small part of the thesis area where the basal contact is not exposed, however, the contact is assumed to be unconformable. The contact occurs in a covered zone in the southwestern corner of the thesis area where mudstones of the Smuggler Cove formation are overlain by cross-bedded sandstones of the Angora Peak member of the Astoria Formation. The ball park unit interfingers with and may locally channel into the upper Smuggler Cove formation (Niem and Niem, in press).

The glauconitic sandstone unit within the Smuggler Cove formation may represent a brief hiatus in sedimentation (i.e., diastem) but probably not a significant unconformity. Glauconite is often deposited during slow or negative sedimentation (Kulm et al., 1975).

Age

The lower three members of the Smuggler Cove formation within the thesis area range from upper Refugian (upper Eocene) to Zemorrian (Oligocene) in age. Foraminifera, calcareous nannofossils, and mollusks were collected from ten localities (appendix 1). Four of these localities contain age-diagnostic fossils (appendix 2). A sample (R-83-141) was collected from the area mapped by Smith (1975) just to the west of the thesis area (fig. 1). This sample is from 8 m above the glauconitic sandstone member and contains a very diverse foraminiferal fauna which Rau (pers. comm., 1983) assigned to the Refugian stage (appendix 2). Foraminifera collected from directly above the glauconitic sandstone unit at locality 116 were assigned to the late Eocene-early Oligocene (McDougall, pers. comm., 1983). Molluskan fossils from the basal part of the upper member of the Smuggler Cove formation were considered to be "probably Matlockian" (locality 98) and "possibly Galvinian" (locality 22) in age (Moore, pers. comm., 1983). The Galvinian stage is correlative to the Refugian foraminiferal stage and the Matlockian stage is correlative to the Zemorrian foraminiferal stage (fig. 5).

Two microfossil samples were collected from the lower member in the CZ 11-28 exploration well adjacent to the thesis area (plate III). These samples contain an upper Refugian foraminiferal fauna that is distinct from the lower Refugian fauna of the underlying Keasey formation (McDougall, pers. comm., 1983). Calcareous nannofossils from these samples were assigned to subzone CP 15b (Bukry, pers. comm., 1983) which is correlative to part of the Refugian foraminiferal stage (figure 5).

No age-diagnostic fossils were collected from the upper part of the upper member in the thesis area. Where the unit is exposed, it has been extensively baked by intrusions of Columbia River Basalt, making it difficult to obtain microfossil assemblages. Adjacent to the thesis area at the Nehalem Fish Hatchery lithologically similar, and presumably chronologically correlative, mudstones have been dated as Zemorrian (Cressy, 1974). Dr. Alan Niem (Oregon State Univ.) and this author collected Zemorrian foraminifera from the upper part of the upper member 1/2 km west of the thesis area (Rau, pers. comm. to Niem, 1985). Cressy (1974) collected early Miocene (lower Saucesian) fossils from the uppermost part of the type section, but it is likely that this part of the section is not present in the thesis area.

In summary, the lower three members of the Smuggler Cove formation in the thesis area are "upper" Refugian to Zemorrian (upper Eocene to Oligocene) in age. The Refugian-Zemorrian boundary occurs within the upper member approximately 40 meters above the glauconitic sandstone member.

Correlation

Previous workers have described vertical lithologic changes within the lower three members of the Smuggler Cove Formation in different areas (e.g. Oswald West Mudstones) Cressy, 1974; Smith, 1975; Neel, 1976; Penoyer, 1977; M. Nelson, 1978; Murphy, 1981). Peterson (1984) constructed a fence diagram and correlated the subunits recognized by previous workers throughout western Clatsop County. The correlations were based on biostratigraphy and lithology with a glauconitic sandstone-siltstone unit being the primary lithologic correlation tool. Peterson (1984) concluded that the glauconitic sandstone in the middle portion of the Smuggler Cove formation reported by Smith (1975), Neel (1976), Tolson (1976), Coryell (1978), M. Nelson (1978), and Peterson (1984) was probably Refugian in age. Data from the present study shows that the glauconitic sandstone is definitely "upper" Refugian in age (appendix 2).

The glauconitic sandstone member in the thesis area can be directly traced several miles to the west into the thesis map areas of Smith (1975) and Neel (1976). Locality 99 in the thesis area and locality R-83-140 in the map area of Smith (1975) contain an identical stratigraphic sequence of glauconite, mudstone, and tuff beds. The foraminiferal fauna from both localities is similar (appendix 1). Therefore, the two localities are probably precisely correlative.

As part of this study the Refugian glauconitic sandstone member has been correlated to the base of the Pittsburg Bluff Formation of eastern Clatsop County. Within the thesis area the glauconitic sandstone member can be traced and mapped into the base of the David Douglas tongue of the Pittsburg Bluff Formation (plate I). Olbinski (1983) and Goalen (in prep.) have described thick glauconitic sandstone beds at the base of the Pittsburg Bluff Formation in eastern Clatsop County. These basal beds are Refugian in age and are, almost certainly, correlative to the glauconitic sandstone member of the Smuggler Cove formation.

Beneath the glauconitic sandstone member is the Refugian lower member. This unit is correlative to the upper mudstone member of the Keasey Formation of Nelson (1985) and the Sager Creek (Vesper Church) formation of Niem and Niem (in press). This correlation is based on the presence of similar microfaunas, similar ages, similar lithologies, and stratigraphic position.

Uppermost Refugian to lower Zemorrian siltstones of the basal part of the upper member overlie the glauconitic sandstone member. These silty mudstones and siltstones have some lithologic characteristics similar to the Pittsburg Bluff formation in that they are fossiliferous, tuffaceous, concretionary, and contain scattered carbonized wood debris. The finer grain size and the presence of a deeper marine fauna suggests that these mudrocks are are a deeper marine (upper slope to outer shelf) facies equivalent of the nearshore to shelf sandstones of the Pittsburg Bluff Formation.

The type section of the Smuggler Cove formation at Short Sands Beach in Oswald West State Park is Zemorrian to early Saucesian in age and correlates to the upper part of the upper member in the thesis area. Cressy (1974) considered the type section to be a deep-marine correlative of the Scappoose Formation in Columbia County. Warren and Norbisrath (1946) proposed the term Scappoose Formation for a sequence of shallow marine sandstone and siltstone mainly in Columbia County that they considered to be Oligocene to early Miocene in age. Van Atta and Kelty (1985) and Kelty (1981), however, recently concluded that the Scappoose Formation was entirely middle Miocene in age based on the very local occurrence of channelized basaltic conglomerates containing middle Miocene Columbia River Basalt Group clasts in the unit. Moore (1976) collected abundant middle Oligocene mollusks, typical of the "Pittsburg Bluff fauna", from within and above strata that Van Atta and Kelty (1985) considered to be middle Miocene in age and part of the Scappoose Formation (e.g. Moore, 1976 localities M-3858 and 15516). Therefore, it is felt that much of the Scappoose Formation as mapped and described by Warren and Norbisrath (1946) and Van Atta and Kelty (1985) is Oligocene to early Miocene in age and correlative to the to the upper portion of the Smuggler Cove formation. Middle Miocene rocks, as indicated by the presence of Columbia River Basalt Group clasts, probably comprise a small part of the Scappoose Formation and should be renamed a seperate younger fluvial unit associated with the

Columbia River basalt flows.

Figure 6 summarizes the correlation of Smuggler Cove formation in the thesis area to the formations in eastern Clatsop County and Columbia County. This figure shows that the Smuggler Cove formation consists of deep-marine facies equivalents of the Keasey, Sager Creek, and Scappoose formations (as defined by Warren and Norbisrath, 1946).

Petrology

Four sandstone and siltstone samples from the lower three members of the Smuggler Cove formation were thin-sectioned and analyzed with a petrographic microscope (appendix 9). Grain size analysis and heavy mineral analysis were performed on sample 451 from the southwestern part of the thesis area (appendices 10 and 11). Smear slides of mudstones from a number of localities were also examined.

Sandstones in the lower three members of the Smuggler Cove formation have highly variable lithologies. For example, sample 758, from the glauconitic sandstone member is a fine-grained calcite-cemented glauconitic sandstone. The sandstone is composed of of 47% glauconite, 29% clay-silt matrix, 9% sparry calcite cement, 5.5% albite twinned plagioclase (An 38-55), 4% andesitic rock fragments, and minor amounts of monocrystalline quartz, mica, and hornblende (fig. 81). The glauconite clasts have a rounded fractured shape indicating a fecal origin (McRae, 1972). Plagioclase clasts are angular to subangular and the volcanic rock fragments are

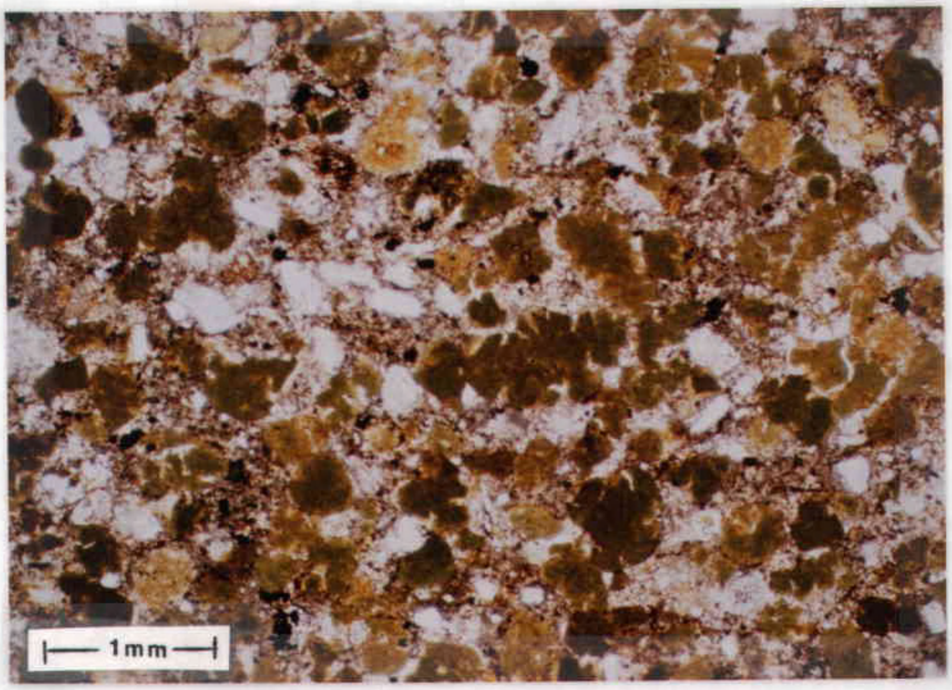


Fig. 81. Thin-section of glauconitic sandstone in the glauconitic sandstone member (sample 758). Note the albite twinned plagioclase, rounded green glauconite pellets and the basaltic rock fragments cemented by highly birefringent calcite.

subangular to subrounded. This sample appears to be lithologically similar to other glauconitic sandstones in the thesis area except that it is cemented by calcite rather than clay. Neel (1976) and Smith (1975) reported similar compositions for the glauconitic sandstone unit to the west of the thesis area.

In comparison, sample 451, from a thin (5 cm) fine-grained sandstone bed in the upper member in the southwestern part of the thesis area, consists of 60% andesitic rock fragments, 17% clay matrix, 8% plagioclase, and minor amounts of monocrystalline quartz, glauconite, and siltstone ripup clasts. This sandstone classifies as a lithic wacke using the classification scheme of Williams et al. (1954) (fig. 72). Heavy minerals from this sample include in decreasing abundance: opaque minerals, epidote, euhedral zircon, hornblende, biotite, and garnet (appendix 10). The abundant zircon and epidote in this volcanic clast-rich sandstone may be the result of selective post-depositional alteration of unstable heavy minerals such as pyroxene and hornblende. The mineralogy of this sample indicates a mixing of volcanic, metamorphic, and plutonic provenances. Grain size analysis shows the sandstone to be very fine-grained, moderately sorted, and positively skewed (appendix 11).

Furthermore, sample 580, from a 12 cm thick clastic dike in the upper member in the southwest part of the thesis area, is fine-grained and composed of 39% monocrystalline quartz, 18% clay matrix, 7% plagioclase and lesser amounts of microcline, metamorphic rock fragments, andesitic rock fragments, chert, biotite, and muscovite (fig. 82). Framework grains are subangular to subrounded. The sandstone classifies as a feldspathic wacke using the

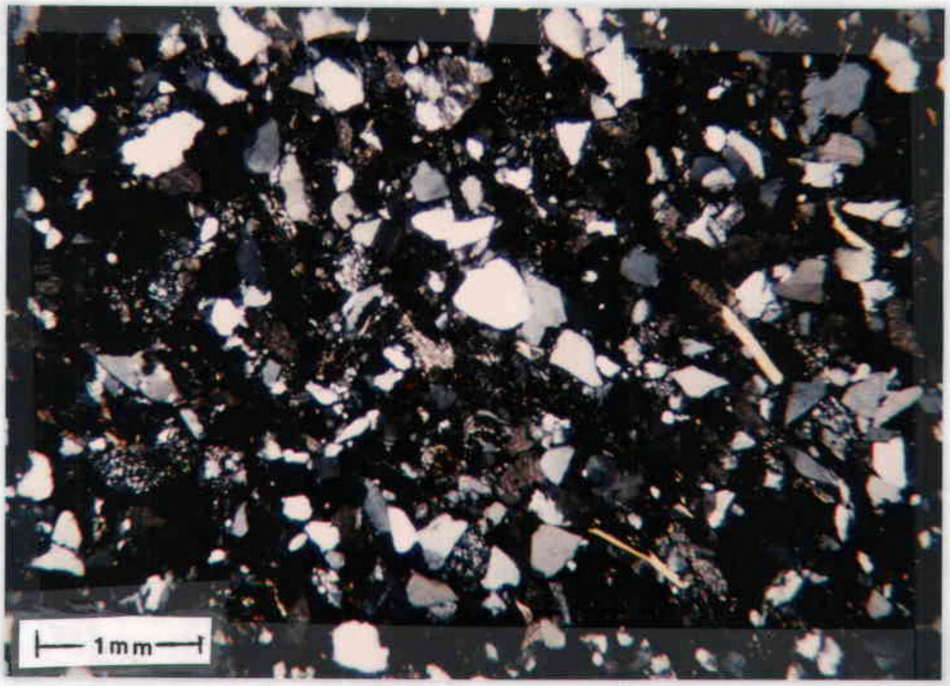


Fig. 82. Photomicrograph (crossed nicols) of fine-grained sandstone in the upper Smuggler Cove formation (sample 580). note chert, mica and quartz.

classification scheme of Williams et al. (1954). Much of the matrix does, however, appear to be authogenic in origin.

Smear slides show that the mudrocks are primarily mudstone with true siltstones being subordinate in abundance. Silt clasts include quartz, plagioclase, orthoclase and abundant volcanic glass shards. Foraminifera are locally abundant, comprising up to 2% of the rock and calcareous nannofossils were observed in several samples. Sand laminae consisting of andesitic rock fragments with a felted texture are locally present in the siltstones of the upper member (e.g., thin section sample 872). Clay size detritus was not X-rayed for clay mineralogy but Mike Nelson (1978), Coryell (1978), and Peterson (1984) reported that smectite and mixed layered smectite clays are the most abundant clay minerals in the Smuggler Cove formation.

The abundant tuffaceous detritus and rare volcanic sandstones in the mudrocks of the lower three members of the Smuggler Cove formation indicate derivation from primarily an andesitic to rhyolitic source, most probably the Cascade arc. The presence of rare micaceous arkosic sandstones indicates a minor acid plutonic-metamorphic source. An ancestral Columbia River system, similar to that mentioned for Jewell member sandstones, is a likely source of the scattered sand-sized arkosic detritus.

Depositional Environment

The lower three members of the Smuggler Cove formation were deposited in a forearc basin, middle slope to outermost shelf environment. Deposition of hemipelagic muds predominated with turbidity current flows being infrequent. Tuffaceous slope and outer shelf muds may have been derived from the muddy plume of a major river to the east and from ash falls settling through the water column. Lithologies and faunas suggest that a maximum shallowing occurred during deposition of the glauconitic sandstone member in the middle part of the formation. The glauconitic sandstone member and the strata directly above it were probably deposited in an outermost shelf to upper slope environment. The glauconite consists of fecal pellets which were formed by mud-ingesting organisms during reduced or negative sedimentation under slightly reducing conditions (McRae, 1972). McRae (1972) and Kulm et al. (1975) have noted that pelletal glauconite presently occurs on the outer slope or upper shelf. The occurrence of sand size plagioclase and basaltic rock fragments in the glauconitic sandstone suggest some current winnowing of the muds and concentration of the glauconitic.

Foraminifera collected above and below the glauconitic sandstone member indicate middle to upper bathyal deposition (McDougall, pers. comm., 1983) (appendix 2), whereas the sample collected from directly above the glauconitic sandstone (R-83-141), contains foraminifera indicative of upper bathyal deposition (Rau, pers. comm., 1983). Mollusks from directly above the glauconitic sandstone along Hamlet Road (locality 13), indicate middle shelf deposition (Moore, pers. comm., 1983). The disarticulated mollusks, however, are slightly abraded and may have been transported from the middle shelf to the outer shelf during storm events. The presence of abundant Helmenthoida trace fossils and scattered Delectopecten bivalves throughout the Smuggler Cove formation support a middle bathyal to outer shelf depositional environment for the unit (Moore pers. comm., 1983).

The Pittsburg Bluff Formation at the type locality in Columbia County was deposited in an inner neritic (inner shelf) setting (Moore, 1976) and represents a marked shallowing from the underlying type Keasey Formation. The glauconitic sandstone member and the upper member of the Smuggler Cove formation strata directly above it are correlative to the basal Pittsburg Bluff Formation and represent the same regional shallowing trend. The thesis area is located basinward of the Pittsburg Bluff Formation and, therefore, the shallowing trend is represented by a change from middle slope to outermost shelf deposition.

Rare sandstone beds and clastic dikes occur in the upper member. The sandstone beds are thin, have sharp basal and gradational upper contacts, and contain intraformational siltstone rip-up clasts. These beds were probably deposited by sediment gravity flows in an upper to middle slope environment.
Rapid deposition of water saturated muds and silts by normal hemipelagic processes and silty turbidites followed by rapid compaction resulted in overpressuring of the sand beds and upward injection of sands as clastic dikes into surrounding mudstones.

Ball park unit

Nomenclature and Distribution

Niem and Niem (in press) have mapped an informal 100 to 200 feet thick unit (Tscm), which they did not name, in the upper part of the Smuggler Cove Formation. Niem (pers. comm., 1985) has suggested that this unit be informally named the "ball park" unit. This name was suggested because the unit is best exposed at the Woodland County park baseball diamond one kilometer west of the town of Nehalem (SE1/4 NW1/4 sec. 28 T3N, R10W). This location will serve as an informal type section of the unit. The type section strata can be mapped into the thesis area (Niem and Niem, in press).

The ball park unit is lithologically similar to the Northrup Creek formation (lower Silver Point member of the Astoria Formation of Peterson, 1984) and occupies a similar stratigraphic position. Niem and Niem (in press), however, have shown that the Northrup Creek formation cannot be continuously traced into the southwestern corner of Clatsop County (including the thesis area). In addition, Niem and Niem (in press) have shown that Smuggler Cove formation mudstones locally overlie the ball park unit whereas the Astoria Formation overlies the Northrup Creek formation. Therefore, Niem and Niem, in press, have included the ball park unit (Tscm) in the Smuggler Cove Formation. The same has been done in this report. Cressy (1974) included the unit mapped as Tscm (ball park unit) by Niem and Niem (in press) in the Oswald West mudstone.

The ball park unit crops out in only a small part of the thesis area at the top of three hills and near the North Fork of the Nehalem River (plate I). The best exposures of the unit in the thesis area occur southeast of Necanicum Junction (locality 66) and in the southwestern corner of the thesis area along an unnamed logging road (localities 927 and 930). Most exposures are small and commonly are baked by middle Miocene Columbia River Basalt Group intrusions. The ball park unit is approximately 70 meters thick in the thesis area.

Lithology

In the thesis area the unit consists of well bedded to laminated, carbonaceous and micaceous, dark gray (N 3) to greenish gray silty mudstone and subordinate thin fine-to medium-grained arkosic sandstone. The mudstones are characterized by abundant carbonized leaf imprints (fig. 83) and muscovite on parting surfaces.

Extensive baking of the mudstone by Columbia River Basalt intrusions results in a porphyroblastic texture in some of the contact metamorphosed mudstones. At locality 66, a thin (2 cm) "coaly" bed is present. Sandstone beds are 10-40 cm thick and have sharp basal and gradational upper contacts except at locality 311 where thin beds to laminae of sandstone are present. Thin, even bedded sandstones are structureless to plane laminated, may contain abundant carbonized plant debris, and are variable in color (fig. 83). The abundance of sandstone in the unit increases from north to south in the thesis area. No sandstone beds were found in the northern part of the thesis area, but sandstone comprises approximately 5% of the unit in the southern part.

Age and Correlation

Because the ball park unit is exposed in only a very small part of the thesis area, where baking by middle Miocene intrusive rocks

Fig. 83. Hand sample of sandstone bed in the ball park unit of the Smuggler Cove fm. Note the distinct laminations. Darker colored laminae contain abundant plant debris (from locality 926, SW 1/4 NW 1/4 sec. 34, T4N, R9W).

has made microfossil recovery difficult, the age of the unit is poorly constrained in the thesis area. Zemorrian foraminifera have been collected from the upper member of the Smuggler Cove formation beneath the ball park unit and lower Miocene sandstones of the Astoria Formation unconformably overlie the unit in the thesis area. To the west of the thesis area Zemorrian to possibly early Saucesian mudstones assigned to the Smuggler Cove formation and Oswald West mudstone locally overlie the ball park unit (Cressy, 1973; Smith, 1975; Niem and Niem (in press). Therefore, the age of the unit is confined to the Oligocene or lower Miocene. Regionally Niem and Niem (in press) consider the ball park unit (Tscm) to be of Oligocene? or early Miocene age.

Accurate correlation of the ball park unit to other units in northwest Oregon is difficult because of age, stratigraphic, and structural uncertainties. The ball park unit in the thesis area can be lithologically and stratigraphically correlated to the lower Silver Point member of the Astoria Formation as mapped by Peterson (1984) and Nelson (1985). These strata have been remapped as Oligocene to lower Miocene Northrup Creek formation by Niem and Niem (in prep).

The ball park unit is probably correlative to the Scappoose Formation of Columbia County as defined by Warren and Norbisrath (1946). Regionally, the unit is probably correlative to part of the Oligocene Yaquina Formation or the lower Miocene Nye Mudstone of the central Oregon coast.

Contact Relations

Peterson (1984) and Nelson (1985) considered the lower Silver Point member (Northrup Creek formation of Niem and Niem, in prep.), which is probably in part correlative to the ball park unit in southwestern Clatsop County, to be disconformable upon the Oswald West mudstone (Smuggler Cove formation). The apparent age disparity between the units, the fact that the lower Silver Point member appears to overlie units of different ages, and the presence of conglomeratic channels near the Oswald West mudstone-lower Silver Point member (Northrup Creek formation) contact were cited as evidence for the unconformity by Peterson (1984). Within the thesis area the ball park unit overlies Zemorrian upper member of the Smuggler Cove formation in the south whereas in the north it overlies possible Refugian upper member of the Smuggler Cove formation. It is likely that the apparent unconformity at the base of the Ball park unit is a result of submarine channeling and that there was not a significant time gap or water depth change between deposition of the underlying members of the bathyal Smuggler Cove formation and deposition of the bathyal ball park unit. The change from hemipelagic slope deposition of the upper Smuggler Cove formation to the sandy turbidite deposition of the ball park unit supports this conclusion.

With the exception of a small outcrop of Angora Peak member of the Astoria Formation north of Alderdale (Plate 1), the ball park unit is the youngest sedimentary rock unit in the thesis area, therefore the upper contact is not available for study. North of

Alderdale, the Angora Peak member of the Astoria Formation overlies the ball park unit but the contact between the units is not exposed. West of the thesis area, Niem and Niem (in press) have mapped the upper member of the Smuggler Cove formation above the ball park unit. This suggests that the upper member is conformable upon the ball park unit and that the Angora Peak member of the Astoria Formation is unconformable upon the upper member and locally upon the ball park unit. More regional biostratigraphic and stratigraphic work needs to be done before before the nature of the ball park unit and its contacts are clearly understood.

Petrology

Four ball park unit sandstone samples and one baked mudstone sample were thin-sectioned and analyzed petrographically (appendix 9). Grain size analysis and heavy mineral analysis were performed on four sandstone samples (appendices 10, 11). Scanning electron microscopy was used to examine porphyroblasts in baked mudstone (sample 929). Porosity measurements of a ball park unit sandstone (sample 928) were made by AMOCO Laboratories, and the organic content of a carbonaceous sandstone (sample 925) was estimated using a hydrogen peroxide treatment.

The sandstone samples are arkosic wacke, feldspathic wacke, and arkosic arenite using the classification scheme of Williams et al. (1954) and Folk (1976) (fig. 72). Concretionary sample 311 has an early calcite cement and classifies as an arkosic arenite. The calcite cement prevented alteration of unstable rock fragments and feldspar to clay. The other sandstone samples lack significant early calcite cement and probably contain abundant diagenetic clay in the matrix. Therefore, at the time of deposition these sandstones may have been arenites.

The sandstones consist primarily of monocrystalline quartz (25-30%), polycrystalline quartz (3-8%), orthoclase (5-18%), plagioclase (An 15-40)(4-6%), volcanic rock fragments (4-6%), siltstone clasts (1-3%), and carbonaceous debris (0-3%). In addition, there are minor amounts of microcline, chert, biotite, myremekite, pyrite, granitic rock fragments, perthite, and heavy minerals (fig. 84). Quartz consists of roughly equal amounts of strained and unstrained varities. Volcanic rock fragments are primarily andesitic in composition and metamorphic rock fragments include phyllite, schist, and gneissic? clasts. Micas, opaque minerals, garnet, and zircon are the most abundant heavy minerals but hornblende, lamprobolite, hypersthene, epidote, tourmaline, staurolite, monazite, rutile, sphene, and apatite are also present (appendix 11).

Matrix comprises from 5 to 35% of the sandstone samples analyzed. Except in the sandstone with early calcite cement (sample 311), much of the matrix appears to be authigenic. Unstable rock fragments and feldspars are generally partially altered to clay; in some samples alteration is sufficiently extensive to form pseudomatrix. Thin section estimates of porosity in sandstones range from 2-7% but a porosity determination by AMOCO laboratories of shows an effective porosity of 16.4% (sample 925). This discrepancy is probably due to the very small size of the pore spaces and pore

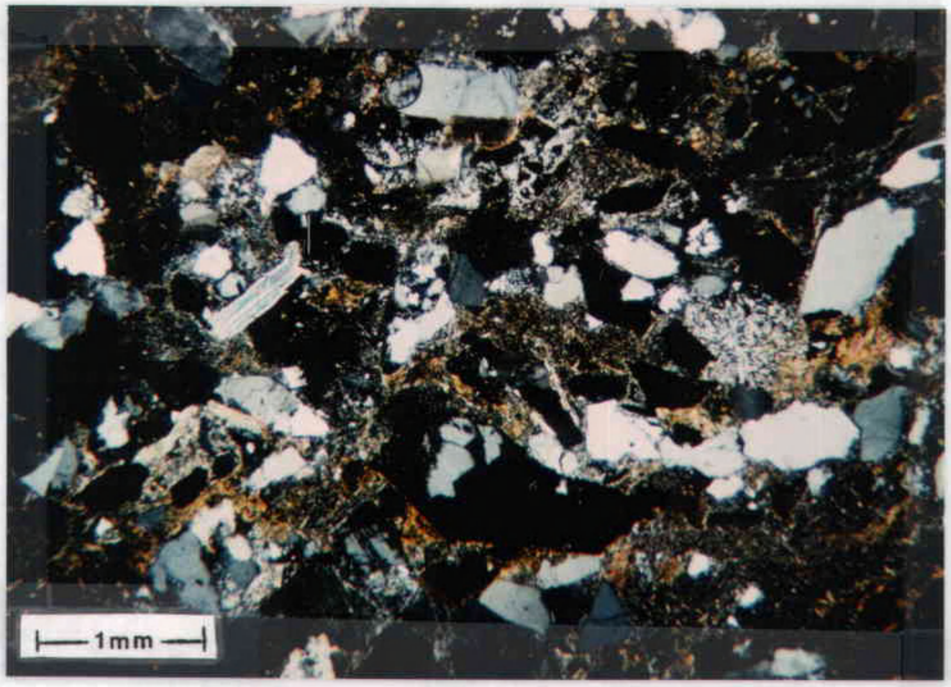


Fig. 84. Photomicrograph (crossed nicols) of medium-grained sandstone in the ball park unit of the Smuggler Cove fm. Note quartzite and muscovite flakes as well as the abundant yelowish-green birefringent clay-silt matrix (locality 930, SE 1/4 NE 1/4 sec. 33, T4N, R9W).

throats between grain boundaries. Pore spaces appear to be limited to very small areas between clay plates.

The carbonaceous sandstone (sample 925) was treated with hydrogen peroxide and contained approximately 7% organic debris. This sample appears to contain more carbonized plant debris than any other sandstones in the ball park unit.

A baked ball park mudstone (sample 929) from approximately 4 m above a Grande Ronde basalt sill consists of 7% very fine sand- to silt- sized quartz, 1% plagioclase, 1% orthoclase, 1% muscovite, and 90% undifferentiated clay and silt size fragments. This sample is distinctly laminated and contains abundant porphyroblasts which in thin section appear to be composed of clay (fig. 85). Scanning electron microscopy confirms that the porphyroblasts are composed of clay and energy dispersive X-ray analysis suggests that the clay is either smectite or chlorite. A sequence of thick Grande Ronde basalt intrusions produced the contact metamorphism in this sample (Plate 1).

Framework grains in the ball park sandstones are subrounded to subangular. Well rounded quartz clasts are present but are very rare. Sieve analysis shows that the sandstones are very fine- to medium-grained (median 3.64 to 1.55 ø) and moderately to poorly sorted (std. dev. 0.50 to 1.59) (appendix 11).

Sandstones in the ball park unit were derived from acid igneous, metamorphic, intermediate volcanic, and minor sedimentary rock sources. Granitic rock fragments, microcline, orthoclase, myremekite, perthite, and euhedral zircon clasts suggest an acid igneous source terrain, such as the Mesozoic Klamath terrain of

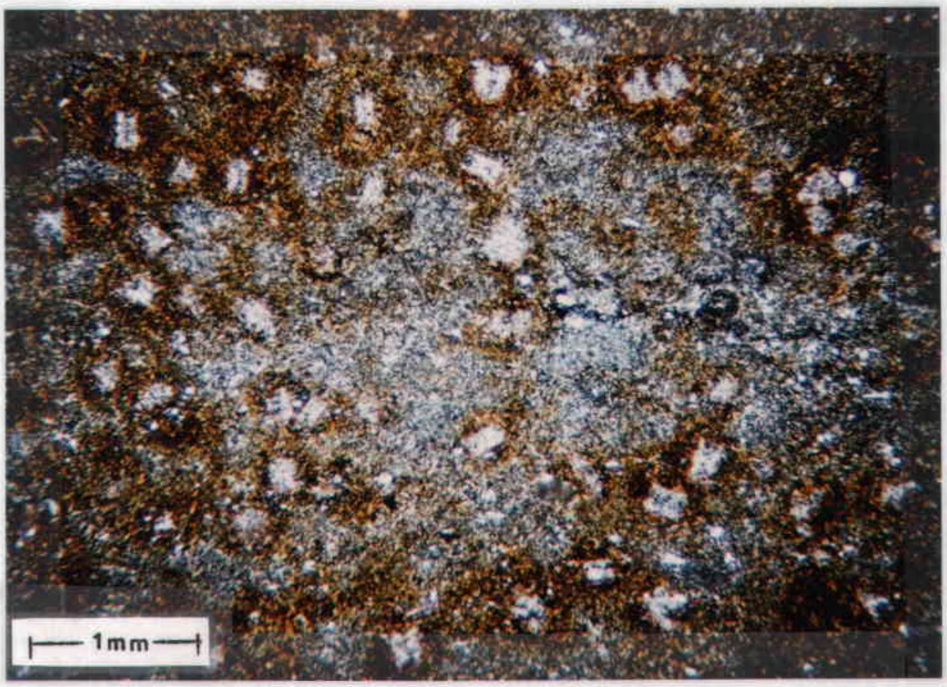


Fig. 85. A) Photomicrograph (plane polarized light) showing porphyroblastic texture of contact metamorphosed siltstone in the ball park unit of the Smuggler Cove fm. (sample 929, SE 1/4 NE 1/4 sec. 33, T4N, R9W).

southern Oregon, the Idaho batholith area, and the North Cascades of northeast Washington and British Columbia. The western Cascade arc is a likely source of the andesitic rock fragments, hypersthene, and green hornblende (e.g. Little Butte Volcanics). Rare well-rounded quartz clasts were were probably recycled from a sandstone source and siltstone clasts are more than likely intraformational.

Dickinson and Suczek (1979) devised a classification scheme of detrital framework modes that can be used as an aid in determining the plate tectonic setting of the provenance. The ball park unit sandstones plot within the recycled orogen and continental block provenances of Dickinson and Suczek (1979) (fig. 43). Recycled orogen provenances include deformed and uplifted sedimentary sequences in subduction zones, along collision orogens, or within foreland fold-thrust belts. Continental block provenances include continental platforms, shields, and plutonic basement rocks.

Uplifted Mesozoic basement blocks occur in the Idaho and Wallowa batholiths and foreland fold-thrust sequences occur in the Blue Mountains of Oregon (Mullen, 1978; Best, 1982). Therefore, the plate tectonic setting of the source area indicated by the plots of Dickinson and Suczek (1979) is consistent with the provenance areas identified by detailed petrography. The plots of Dickinson and Suczek (1979) however, can be misleading because the final depositional site of the ball park unit sandstones is a forearc basin, not a recycled orogen or continental block provenance.

Sandstones in the ball park unit were derived primarily from granitic and metamorphic source areas whereas sandstones in the older Pittsburg Bluff Formation were derived primarily from from an intermediate volcanic source (i.e., Cascade arc). The paucity of andesitic detritus in ball park unit sandstones could be explained by the presence of an ancestral Columbia River crossing the Cascade arc and depositing arkosic sands seaward of a "deltaic complex". The paucity of andesitic detritus could also be explained by a reduction of volcanism in the arc during deposition of the ball park unit.

Depositional Environment

Because of limited exposure area and extensive baking, relatively little can be deduced about the depositional environment of the ball park unit in the thesis area. No megafossils, trace fossils, or microfossils were found in the unit. The presence of well-laminated sandstones, rich in plant debris, with sharp basal and gradational upper contacts, however, suggests deposition by turbidity currents seaward of a "deltaic complex". Well-preserved plant debris and pyrite indicate a reducing environment. Reducing environments are common on the continental shelf and slope (McDougall, 1980). The lack of bioturbation in the laminated, carbonaceous mudstones of the ball park unit suggest either a high sedimentation rate or a low oxygen environment.

Peterson (1984) interpreted the lower Silver Point member, which Niem and Niem (in press) recently renamed the Northrup Creek formation to have been deposited on the outer shelf to upper slope. The ball park unit is thought to be correlative to the Northrup Creek formation. The above environment is indicated by sedimentary structures and by the paleoecology of the molluscan fauna and trace fossils. Sandstone beds were probably deposited on the slope by storm-generated turbidity currents. Deposition of alternating, graded to laminated sands and laminated muds has been attributed to storm induced turbidity currents (Hamblein and Walker, 1979).

Dott and Bird (1979) have interpreted the siltstones of the middle Eocene Elkton Formation as a transgressive slope facies between the deltaic Coaledo Formation and mid-fan turbidites of the Tyee Formation in southwestern Oregon. The Elkton Formation consists of dark gray laminated siltstone and mudstone with scattered thin beds (10-15 cm) of laminated fine-grained sandstone. Mudstone intraclasts occur sporadically within the sandstones. Locally, coarser-grained micaceous sandstones are present. The ball park unit in the thesis area is lithologically similar to much of the Elkton Formation. This suggests that the Elkton Formation and the ball park unit may have been deposited in similar upper bathyal environments. Large channels are present in the Elkton Formation, but exposures of the ball park unit in the thesis area are too limited to reveal the presence of large nested turbidite channel sequences. The fact that the ball park unit appears to rest on mudstone strata of differing ages however, could be explained by channeling. The Northrup Creek formation, which is thought to be a lateral correlative of the ball park unit, does contain thin, erosional, nested turbidite channel deposits (Niem, pers. comm., 1985)

The Scappoose Formation, as described by Warren and Norbisrath (1946) in the East Fork of the Nehalem River area, is Oligocene to early Miocene in age and is, in part, deltaic to shallow marine. The ball park unit and the Northrup Creek formation may represent an outer shelf to upper slope channel fill sequence related to the "Scappoose delta". In this model erosive turbidites, generated by slumping or storm activity near the "delta front" would move downslope across a narrow shelf producing a channel-fill sequence near the shelf-slope break. Such a model would explain the abundance of terrestrial carbonaceous plant debris, as well as other features, in the ball park unit.

PITTSBURG BLUFF FORMATION

Nomenclature and Distribution

The "Pittsburg Bluff sandstone" was first described by Schenck (1927). Weaver (1937) later proposed the term Pittsburg Bluff Formation for all middle Oligocene rocks in Columbia County, Oregon. Warren et al. (1945) and Warren and Norbisrath (1946) mapped the areal extent of the Pittsburg Bluff Formation in Columbia County and eastern Clatsop County. Olbinski (1983) and Nelson (1985) extended the Pittsburg Bluff Formation into central Clatsop County.

The term David Douglas tongue of the Pittsburg Bluff Formation is used informally in this report for a thin (15 m) sequence of very fine-grained tuffaceous sandstones which crops out in the extreme northeast corner of the thesis area at David Douglas State Park. The Pittsburg Bluff Formation in the thesis area is restricted to the David Douglas tongue. Siltstones, mudstones, and interbedded glauconitic sandstones that are age equivalent to the David Douglas tongue have been mapped as Smuggler Cove formation thereby restricting the Pittsburg Bluff Formation to a predominantly tuffaceous lithic to arkosic sandstone unit.

Peterson (1984) mapped the Klaskanine tongue of the Pittsburg Bluff Formation to the north of the thesis area. This unit is probably correlative to the David Douglas tongue. Both tongues represent a maximum westward extension of the sandy Pittsburg Bluff Formation. The tongues are distinguished from the Pittsburg Bluff Formation to the east by a dominance of very fine-grained tuffaceous sandstone and by the thin nature of the units.

The only exposures of the David Douglas tongue in the thesis area occur in road cuts along U.S. highway 26 at David Douglas State Park (e.g., localities 323a, 323b, 1200, 1201) and along logging roads adjacent to the highway (e.g., locality 320). The unit is approximately 15 meters thick in this area and pinches out rapidly (within 2 km.) into mudstones and siltstones of the Smuggler Cove formation to the south and west. Pittsburg Bluff Formation sandstones rapidly thicken and become coarser-grained and very fossiliferous to the east of the thesis area (Olbinski, 1983; Nelson, 1985; Moore, 1986). The David Douglas tongue forms a small bluff where it crops out in the thesis area.

Lithology

The David Douglas tongue consists of light gray (N 7), thick-bedded, moderately well-indurated, concretionary, bioturbated, very fine-grained tuffaceous sandstone and minor siltstone (fig. 83). The unit is characterized by ellipsoidal, 1 meter diameter calcareous concretions. Weathered pelecypods, gastropods, scaphopods, and large benthic foraminifera are common (appendix 1). The sandstones are very tuffaceous, contain minor amounts of glauconite and calcite cement. These massive very fine-grained sandstones are extensively bioturbated and, therefore, smaller sedimentary structures (e.g. cross-bedding) are not preserved. A few narrow 2 cm long clay-filled burrows are present.

Fig. 86. Typical exposure of Pittsburg Bluff Formation fine-grained sandstone (locality 323, SE 1/4 NW 1/4 sec. 21, T5N, R8W).

Contact Relations

Mapping relationships indicate that the David Douglas tongue interfingers with the Smuggler Cove formation and that the two units are conformable. Peterson (1984) noted a similar relationship to the north of the thesis area. The conformable and gradational nature of the contacts is evidenced by the similar depositional environments, ages, and structural attitudes of the units. At the type area, however, the shallow marine Pittsburg Bluff Formation disconformably overlies upper bathyal mudstones of the Keasey Formation (Warren and Norbisrath, 1946; Kadri, 1982). In northeast Clatsop County the Pittsburg Bluff formation disconformably overlies the Sager Creek formation (Olbinski, 1983; Niem and Niem, in press). Murphy (1981) and Kadri (1982) have demonstrated that to the north and east of the thesis area the Pittsburg Bluff Formation is unconformably overlain by the Astoria Formation and the Columbia River Basalt Group.

Age and Correlation

Foraminifera collected near the base of the David Douglas tongue (locality 323) show that the unit is Refugian (upper Eocene to possibly lower Oligocene) in age (McDougall pers. comm., 1983) (appendices 1 and 2). The fauna from this locality has some similarities to the Refugian Keasey Formation fauna (McDougall, pers. comm., 1983). The similarity in faunas probably can be attributed to similar depositional environments of the units.

Foraminifera collected from elsewere in the Pittsburg Bluff Formation indicate a Refugian to possible lower Zemmorian (upper Eocene to lower Oligocene) age (Peterson, 1984; Olbinski, 1983; Bruer et al., 1984; Nelson, 1985). Moore (1976), however, studied the molluscan fauna from the type Pittsburg Bluff Formation and concluded that the formation was of middle Oligocene age.

The slight discrepancy in ages indicated by foraminifera and mollusks probably can be attributed to the difficulty in correlating provincial foraminiferal and molluscan stages to global chronostratigraphic units. Prothero and Armentrout (1985) have recently suggested that the upper Refugian stage be extended into the Oligocene. Such an extension would eliminate the discrepencies.

The David Douglas tongue is correlative to the basal part of the Pittsburg Bluff Formation in eastern Clatsop County. The presence of glauconite, the Refugian age, and regional mapping support this correlation. To the west the glauconitic sandstone member of the Smuggler Cove formation correlates to the base of the David Douglas tongue. These thick glauconitic sandstones have been noted by Olbinski (1983) and by Goalen (in prep.) to the northeast at the base of the Pittsburg Bluff Formation. Reconaissance work suggests that tuffaceous sandstone mapped as Oswald West mudstone by Penoyer (1977) north of the thesis area should be included in the David Douglas tongue.

Regionally the Pittsburg Bluff Formation is correlative to the lower part of the Lincoln Creek Formation in southwest Washington and to the Tunnel Point Sandstone, the upper part of the Eugene Formation, and the lower part of the Alsea Formation in western

Oregon (Moore, 1976).

Petrology

Two samples from the David Douglas tongue along U.S. highway 26 were thin sectioned and examined with a petrographic microscope (samples 323a, 323b) (appendix 9). Heavy mineral analysis and grain size analysis were performed on samples 323b and 658c (appendices 10, 11).

Thin-sectioned samples are lithic wackes using the classification scheme of Williams et al. (1954) (fig. 72). Peterson (1984) reported that sandstones from the Klaskanine tongue are arkosic wackes using the same classification scheme. This discrepancy can be explained by the abundant diagenetic clay matrix reported by Peterson (1984). This matrix probably formed from diagenetic breakdown of unstable lithic fragments (e.g. volcanic glass shards). The sandstones in the thesis area are relatively fresh and the lithic fragments have been preserved.

Samples 323a and 323b are lithologically similar except that 323a contains abundant (22%) calcite cement whereas sample 323b has only 2% calcite cement. The calcite cement was precipitated soon after burial as evidenced by well-preserved unstable rock fragments and glass shards (fig. 84). Framework components of the two samples are in decreasing abundance: volcanic glass shards; quartz; glauconite; plagioclase (An 30-55); andesitic to basaltic rock fragments; biotite; muscovite; and hornblende (appendix 9). Sample 323a contains 27% clay-silt matrix. Fragmented diatoms are common in

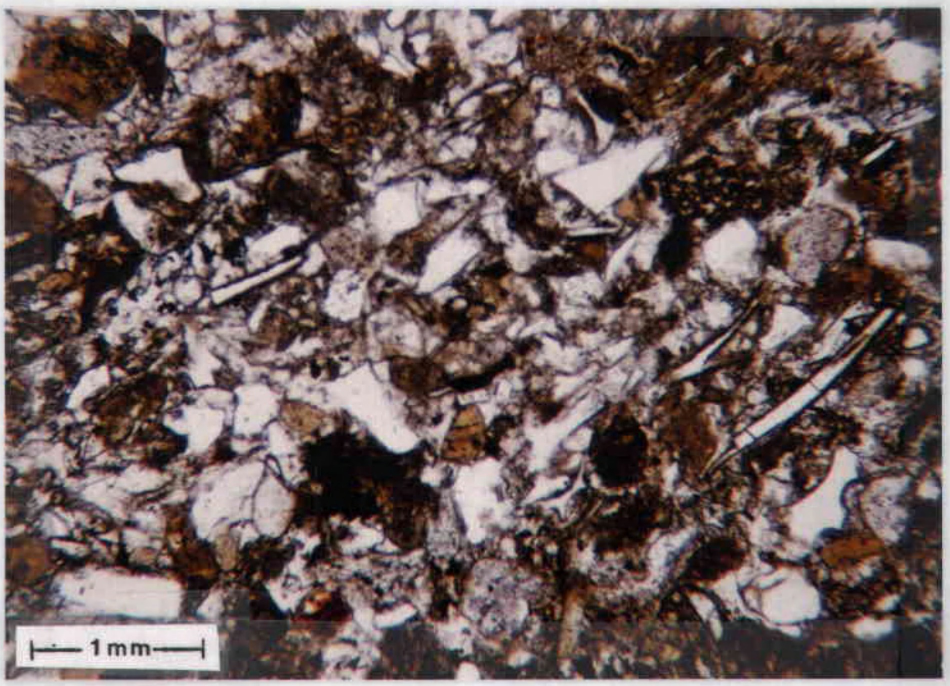


Fig. 87. Photomicrograph of calcite-cemented, very fine-grained Pittsburg Bluff Formation sandstone. Note the presence of abundant glass shards (sample 323a).

the matrix. Early calcite cement in this sample indicates that the matrix is detrital rather than authigenic in origin. Much of the matrix was probably introduced into the sandstones by extensive bioturbation.

Heavy minerals from the 3 ϕ and 4 ϕ size fraction of samples 323b and 658c were examined (appendix 10). Green hornblende, micas, and opaque minerals are the most common heavy minerals with minor amounts of lamprobolite, hypersthene, augite, garnet, tourmaline, apatite, euhedral zircon, epidote, kyanite, staurolite, monazite, rutile, and sphene. Abundant hornblende and mica combined with rare zircon and epidote serve to distinguish the heavy mineral suite of the David Douglas tongue from the heavy mineral suites of other units in the thesis area.

The David Douglas tongue was primarily derived from an explosive andesitic to rhyolitic source area as evidenced by abundant sickle-shaped glass shards, andesitic rock fragments, green hornblende, and hypersthene. The western Cascade arc of western Oregon and Washington is more than likely the source of the volcanic detritus. A minor metamorphic source is indicated by phyllitic and schistose rock fragments as well as kyanite. A minor granitic source area is indicated by the presence of euhedral zircon. None of the detritus can be positively assigned to a recycled sedimentary rock source (i.e. all abrasionally resistant detritus is angular). The sandstones are texturally and compositionally immature indicating rapid erosion of mountainous source areas and rapid deposition before significant winnowing by currents (Folk, 1980). The North Cascades of Washington, the Idaho batholith area, and the Klamath terrain are likely sources of the metamorphic and granitic detritus.

Depositional Environment

The David Douglas tongue was deposited in an outer shelf to upper slope environment and is a deeper marine facies equivalent of the type Pittsburg Bluff Formation of Columbia County. Foraminifera collected from near the base of the David Dougless tongue indicate bathyal deposition (McDougall pers. comm., (1983) (appendix 2). Glauconite and the very-fine grain size of the sandstone also suggests deposition on the outer shelf or upper slope (McRae, 1972; Kulm et al., 1975). Peterson (1984) also demonstrated that the Klaskanine tongue was deposited in an outer shelf to upper slope environment. The type Pittsburg Bluff Formation has been interpreted to have been deposited in middle shelf to deltaic environments (Moore, 1976).

The Pittsburg Bluff depositional setting has been compared to the modern Oregon continental shelf by Peterson (1984) and Nelson (1985). Kulm et al. (1975) describe four facies on the present Oregon shelf: 1) an inner to middle shelf sand facies; 2) a middle to outer shelf mixed sand and mud facies; 3) an outershelf mud facies; and 4) an outer shelf glauconitic sand facies. The type Pittsburg Bluff Formation corresponds to the inner shelf sand facies, parts of the Pittsburg Bluff and Smuggler Cove formations correspond to the outer shelf mud facies, the David Dougless and Klatskanine tongues correspond to the outer shelf mixed facies, and the glauconitic sandstone of the basal Pittsburg Bluff

Formation in eastern Clatsop County and the glauconitic sandstone unit of the Smuggler Cove formation correspond to the outer shelf glauconitic sand facies.

The Pittsburg Bluff Formation in Clatsop County appears to first shallow upward then deepen upward. Glauconitic sandstone at the base of the formation in eastern Clatsop County (Goalen, in prep.) and a bathyal fauna at the base of the David Dougless tongue suggest that the shelf-slope break was located east of the thesis area during initial deposition of the unit. The inner to middle shelf molluscan fossils above the glauconitic sandstone in eastern Clatsop County (Olbinski, 1983; Nelson, 1985) and correlative outer shelf sandstone and glauconite in the thesis area indicate that the shelf slope break was subsequently located within or to the west of the thesis area. Bathyal foraminifera faunas have been collected from upper Smuggler Cove formation siltstones above the David Douglas tongue and correlative units to the west indicating a gradual deepening depositional environment in the upper part of the Pittsburg Bluff Formation.

ASTORIA FORMATION

Introduction

The Astoria Formation crops out over a very limited part of the study area where it is exposed at three small localities (plate 1). Therefore, the Astoria Formation will be discussed briefly in this report. The reader is referred to Cooper (1981) for a thorough discussion of the Astoria Formation.

Nomenclature and Distribution

The Astoria Formation was first studied by T.A. Conrad in 1848. The type area Astoria Formation was divided into Oligocene and Miocene units by W.H. Dall in 1904 (Moore, 1963). Arnold and Hannibal (1913) extended the "Astoria series" to include strata exposed near Tillamook and Newport along the central Oregon coast. Howe (1926) described and redefined the type area at Astoria dividing the formation into three informal members: 1) a lower sandstone member; 2) a middle "shale" member; and 3) an upper sandstone member. Unfortunately the type area was subsequently covered and destroyed (Cooper, 1981).

Since 1926 the Astoria Formation has been mapped primarily on a biostratigraphic basis, which has caused confusion concerning the lithostratigraphy of the formation. Recent workers in the Astoria Formation (e.g., Cressy, 1974; Smith, 1975; M. Nelson, 1978; Cooper, 1981; Peterson, 1984) have attempted to define the formation on a lithostratigraphic basis and have divided it into several informal mappable members. In Clatsop County, the formation has been divided into four informal members: 1) the fluvial to shallow-marine Angora Peak member; 2) the shallow-marine Big Creek sandstone member; 3) the deep-marine Silver Point member; and 4) the deep-marine Pipeline member. The names Silver Point, Big Creek, and Pipeline are pre-empted according to the U.S. Geological Survey Geologic Names Committee, therefore Niem and Niem (in press) have recently renamed part of the Silver Point member the Cannon Beach member of the Astoria Formation. The Pipeline member has been renamed the Youngs Bay member and the Big Creek sandstone has been renamed the Wickiup Mountain member (Niem and Niem, in press). The name Angora Peak member of the Astoria Formation had already been reserved for this unit by the Geologic Names Committee of the U.S. Geological Survey and, therefore, was not changed by Niem and Niem (in press).

The Angora Peak member is the only member of the Astoria Formation in the thesis area. Cressy (1974) first proposed usage of the term Angora Peak member for a 1000 feet thick sequence of lower? to middle Miocene sandstone and conglomerate exposed near Angora Peak. The type section of the Angora Peak member is located several kilometers west of the thesis area and has been described in detail by Cressy (1974) and Cooper (1981). The unit pinches-out into the Cannon Beach member 10 kilometers north of the type section (Smith, 1975).

The Angora Peak member crops out in the southwestern corner of the thesis area (plate I) in three small exposures. The unit forms a low hill and is more erosionally resistant than surrounding

Smuggler Cove Formation mudstones. Approximately 80 meters of Angora Peak member are present in the study area. The top of the unit is not exposed exposed due to erosion. West of the thesis area the unit is approximately 250 meters thick (Cressy, 1974; Cooper, 1981).

Lithology

In the thesis area, the Angora Peak member consists of fine- to very fine-grained, moderately sorted, massive to cross-bedded, arkosic sandstone (localities 592, 662) (fig. 88) and subordinate moderately to poorly sorted polymict pebble conglomerates (locality 593). Neither Megafossils nor trace fossils were observed. The sandstones are pale yellowish orange (10YR 8/6) to light brown (5YR 5/6) due to iron oxide staining. Pebbles in the conglomerate are rounded and consist of volcanic (basaltic trhyolitic?), metamorphic, and plutonic? rock fragments. The pebbles are set in an arkosic sand-silt matrix. The exposure of the conglomerate is very small making it impossible to see large-scale sedimentary structures. To the west of the thesis area, the Angora Peak member consists of fine- to coarse-grained arkosic sandstone with subordinate micaceous and carbonaceous siltstone, coal, and conglomerate (Cressy, 1974). Large channels, conglomerate lenses, scour and fill structures, and large-scale trough cross-bedding are associated with the coarser-grained sandstone. Fine-grained feldspathic sandstones are thinly laminated to cross-bedded, thick bedded, and may contain shallow marine molluskan fossils. Thin subbituminous coal beds have been observed at several localities in the Angora Peak member

(Cressy, 1974).

Contact Relations

Cressy (1974) stated that "the Angora Peak sandstone unconformably overlies the Oswald West mudstone and is in turn unconformably overlain by the "Depoe Bay Basalts" (Grande Ronde Basalt of this study). The upper contact of the Astoria formation does not occur in the thesis area due to erosion and the lower contact is not exposed. Therefore, the contact relations are best deduced from regional information. Snavely et al. (1973) noted that the Astoria Formation at Depoe Bay Oregon (approx. 60 km south of the thesis area) unconformably overlies the Nye Mudstone and is unconformably overlain by Depoe Bay Basalt. Cressy (1974) and Cooper (1981) demonstrated that an angular unconformity exists between the Oswald West mudstone (renamed the Smuggler Cove formation by Niem and Niem, in press) and the Angora Peak member based on: 1) the fact that the Angora Peak member regionally overlies strata of different ages; 2) the sharp contact between the units; and 3) the distinctly different depositional environments of the units.

Age and Correlation

No age-diagnostic fossils were found in the Angora Peak member in the thesis area. However, the Angora Peak member unconformably overlies the Oligocene to lower Miocene Smuggler Cove formation and is intruded by middle Miocene Columbia River Basalts thereby providing age constraints. Molluskan fossils collected by Cressy (1974) and Cooper (1981) from the Angora Peak member adjacent to the thesis area have been assigned to the Pillarian and Newportian (early to middle Miocene) stages of Addicott (1976, 1981).

Based on megafossil ages, the Angora Peak member has been correlated to the Astoria Formation in the type area, and to the Astoria Formation in the Newport and Tillamook embayments (Cressy, 1974). Peterson (1984) considered the Big Creek member (Wickiup Mountain member of Niem and Niem, in press) to be chronologically correlative to the Angora Peak member.

Petrology

Two samples of fine-grained Angora Peak sandstone (589 and 662) were thin- sectioned and analyzed with a petrographic microscope. Heavy minerals from the 3ϕ and 4ϕ size fraction of these samples were examined (appendix 10) and the grain size distribution of the entire sample was determined (appendix 11).

The samples are arkosic wackes using the classification scheme of Williams et al. (1954) (fig. 72). Framework grains are subrounded to subangular and are composed of monocrystalline quartz (35%), polycrystalline quartz (5%), plagioclase (An 15-45)(5-6%), metamorphic rock fragments (2-4%), and minor amounts of microcline, chert, myrmekite, and heavy minerals (fig. 88). Heavy minerals include in decreasing abundance opaque minerals, biotite, muscovite, epidote, garnet, green hornblende, hypersthene, euhedral zircon, clinozoisite, tourmaline, staurolite, and rutile.

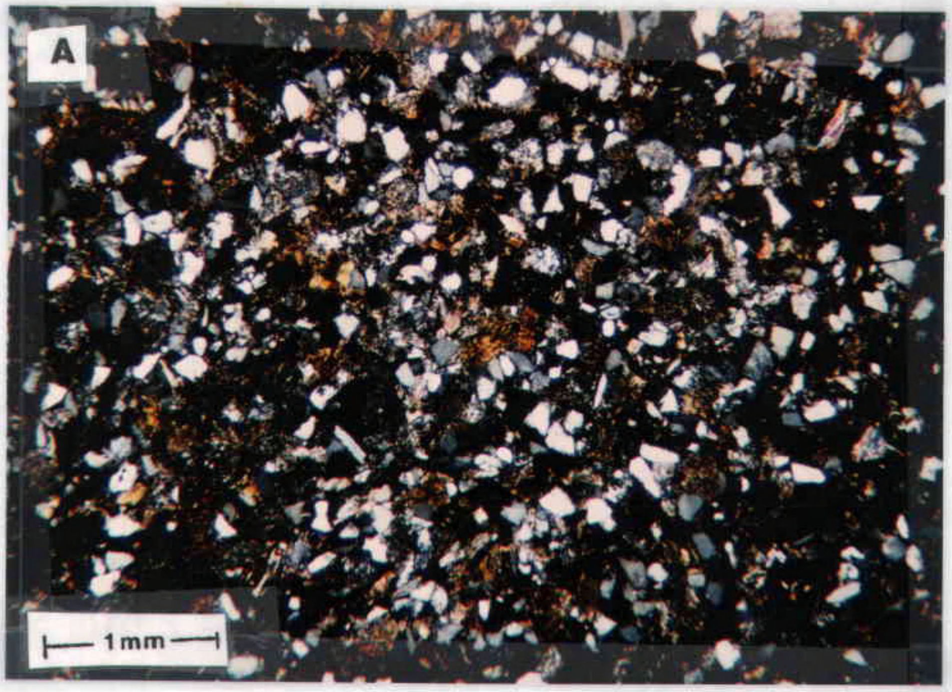


Fig. 88. Photomicrograph (crossed nicols) of fine-grained Angora Peak member sandstone. Note the abundant angular monocrystalline quartz, feldspar, and yellow-brown iron-stained clay (sample 662). B is exposure of cross-bedded sandstone where sample 662 was collected.

Framework grains are set in a clay-silt matrix which comprises 25% of the rock. Much of the matrix appears to be authigenic in origin. Potassium feldspar has been partially altered to clay and some metamorphic rock fragments have been altered to pseudomatrix. The percentage of authigenic versus detrital clay is unknown. Clay is the primary cement, but iron oxide cement is locally abundant. The iron oxide appears to be the result of teleogenetic (surficial) oxidation of iron-bearing minerals such as hornblende and biotite as "fresh" samples lack the iron oxide.

The sandstones are fine- to very fine-grained (median 3.32-2.45ϕ), poorly to moderately sorted (std. dev. 1.06-0.091), and positively skewed (simple skewness 1.07-1.21) (appendix 11). If the clay matrix, much of which is authigenic, is not included in the grain size calculations, the samples have standard deviations ranging from 0.46-0.51 and, therefore, classify as moderately to well sorted.

Angora Peak sandstones were derived from intermediate volcanic, acid plutonic, and metamorphic sources. The absence of well-rounded quartz and zircon clasts suggests that little of the detritus was derived from a recycled sedimentary rock source. A few andesitic rock fragments, grains of hypersthene, and green hornblende indicate input from an andesitic source, most probably the western Cascade arc. Clasts derived from a metamorphic terrain (e.g., metamorphic rock fragments, clinozoisite, and epidote) and clasts derived from an acid igneous terrain (e.g., euhedral zircon) may have been deposited via an ancestral Columbia River system that drained much of the northwestern United States. The Mesozoic Idaho batholith area, the North Cascades of Washington, and the Blue Mountains of eastern

Oregon could have supplied the metamorphic and igneous detritus in the Angora Peak member. Cressy (1974), Niem (1976) and Cooper (1981) have previously suggested an ancestral Columbia River source for the Angora Peak member.

Depositional Environment

The Angora Peak member exposed in the thesis area was deposited in both shallow marine and fluvial environments. The fine- to very fine-grained sandstone at localities 589 and 662 are moderately sorted and trough cross-bedded suggesting a high energy inner shelf depositional environment. The trough cross-bedding suggests development of sand waves on the shelf. Cressy (1974) collected shallow-marine (middle shelf to intertidal) mollusks from lithologically similar fine-grained Angora Peak sandstones adjacent to the thesis area. He concluded that the unfossiliferous fine-grained Angora Peak member sandstones were deposited as delta front sheet sands in a wave-dominated deltaic system.

The pebble conglomerate (locality 593) in the thesis area is interpreted to be fluvial in origin. The moderate to poor sorting observed in the thesis area and large-scale planar-tabular cross-beds and coal beds in the conglomerate adjacent to the thesis area (Cressy, 1974) indicate fluvial to distributary channel deltaic deposition.

Cressy (1974) and Cooper (1981) interpreted the Angora Peak member as a westward prograding, wave-dominated deltaic unit. Approximately 70% of the member was deposited in a wave dominated shallow-marine environment, the remainder in a fluvial environment (Cressy, 1974). Cooper (1981) interpreted all the members of the Astoria Formation as a complex of deltaic, prodeltaic, shallow-marine, and submarine canyon head deposits.

COLUMBIA RIVER BASALT GROUP

Nomenclature and Distribution

Introduction

Middle Miocene basaltic intrusions in the thesis area have previously been mapped as undifferentiated Tertiary intrusions (Beaulieu, 1973). In a regional study Snavely et al., (1973) described and mapped correlative middle Miocene basaltic intrusions and flows along the central to northern Oregon coast and along the southern Washington coast. They divided the middle Miocene basalts into three petrologic types: the Depoe Bay Basalt; the Cape Foulweather Basalt; and the basalt of Pack Sack Lookout. They noted that these coastal basalts were compositionally, petrologically, and chronostratigraphically "identical" to the Grande Ronde Basalt, the Frenchman Springs Member, and the Pomona Member of the Columbia River Basalt Group on the Columbia Plateau but considered the "coastal basalt" to have been erupted locally. The presence of dikes and sills was used to support a local origin for these basalts.

In 1979, however, Beeson et al., (1979) suggested that the "coastal basalts" were intially erupted on the Columbia Plateau, flowed down an ancesteral Columbia River into the marine environment forming submarine basalt piles and "invasive" dikes and sills. If the hypothesis of Beeson et al. (1979) is correct then the Miocene "coastal basalts" should be included in the Columbia River Basalt Group rather than the nomenclatural framework of Snavely et al.

(1973). Data from this study and from previous studies (e.g., Murphy, 1981; Peterson, 1984; Nelson, 1985) indicate that the "invasive" hypothesis is valid and, therefore, the Columbia River Basalt Group nomenclature is used for the "coastal basalts" in this report. Peterson (1984) was the first worker to use the Columbia River Basalt Group nomenclature for the "coastal basalts". Figure 89 compares the terminology of Snavely et al. (1973) to the Columbia River Basalt Group nomenclature used in this report. "Intrusive" is used in this report as a descriptive term refering to the presence of a younger igneous body in older strata and carries no genetic meaning. The origin of the middle Miocene basalts in the thesis area will be discussed in detail in a subsequent section.

Russell (1893) was the first to describe the Miocene basalts on the Columbia Plateau of eastern Washington, northeastern Oregon, and western Idaho. In 1901, he first used the term "Columbia River basalt" for these Miocene basalts and other older basalts (Waters, 1961). Merriam (1901) restricted the term Columbia River basalt to middle and upper Miocene basalts on the Columbia Plateau.

Until 1961 the Columbia River Basalts were considered to be a thick pile of monotonous flood basalts. Waters (1961) and Mackin (1961), however, were able to differentiate regional stratigraphic units through detailed geologic mapping and major oxide chemical analysis. Waters (1961) divided the Columbia River Basalt into two major units: the older Picture Gorge Basalt and the younger Yakima Basalt.

Swanson et al. (1979) recently revised the stratigraphic nomenclature of the Columbia River Basalt. They elevated the

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	Frenchman Springs
(WHALE COVE SANDSTONE)	(UNNAMED 400-ft THICK SANDSTONE INTERBED)
Depoe Bay	Grande Ronde

Fig. 89. Comparison of coastal basalt nomenclature of Snavely *et al*. (1973) to Columbia River Basalt nomenclature used in this report.

Columbia River Basalt to Group status and elevated the Yakima Basalt to the subgroup level. The older Picture Gorge Basalt and Imnaha Basalt were retained at the formational level. The Yakima Basalt subgroup consists of three formations which are, from oldest to youngest: Grande Ronde Basalt; Wanapum Basalt; and Saddle Mountains Basalt. The Grande Ronde Basalt was divided into four magnetostratigraphic units and the other two formations in the Yakima Basalt Subgroup were divided into numerous flows and members (fig. 90).

According to Beeson and Moran (1979) only the Grande Ronde Basalt, the Frenchman Springs and Priest Rapids members of the Wanapum Basalt, and the Pomona Member of the Saddle Mountains Basalt flowed out of the Columbia Plateau and into western Oregon and western Washington. These units have been further divided into different magnetostratigraphic and chemical subtypes which will be discussed in the following paragraphs.

Grande Ronde Basalt

The Grande Ronde Basalt on the Columbia Plateau was divided into a high MgO (>4.2% MgO) geochemical unit and a low MgO geochemical unit by Wright _et al._ (1973). In the Willamette Valley and in the northern Oregon Coast Range high MgO flows overlie low MgO flows (Beeson and Moran, 1979; Murphy, 1981). Long _et al._ (1980) used a plot of TiO2 versus MgO to define the Unitanum (high TiO2) subtype of low MgO Grande Ronde Basalt in the Pasco Basin of central Washington. Murphy (1981) and Goalen (in prep.) show that the low

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
					Elephant Mountain Member	10.5	N,T
	MIDDLE MIOCENE				erosional unconformity		
					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
				Grande Ronde Basalt		14.5-16.5	N_2
	LOWER MIOCENE		Picture Gorge Basalt				R_2
							N_1
							R_1
				Imnaha Basalt			R_1
							T
							N_0
							R_0

Fig. 90. Columbia River Basalt Group stratigraphy. Units present in the thesis area are marked with a heavy black vertical bar. Marked units plus the Priest Rapids Member, and N1 Grande Ronde Basalt are the only units present in western Oregon and western Washington (modified from Beeson and Moran, 1979).

MgO Grande Ronde Basalt flows in northeastern Clatsop County can also be subdivided into into high TiO_2 (>2.3% TiO_2) and low TiO_2 subunits. Using the three chemical subtypes discussed above and magnetostratigraphy (N=normal polarity, R=reverse polarity) the Grande Ronde Basalt of the northern Oregon Coast Range has been divided into four subunits which are from oldest to youngest: 1) R2, low MgO-high TiO_2 (Tgr1); 2) R2 low MgO-low TiO_2 (Tgr2); 3) N2, low MgO-low TiO_2 (Tgr3); and 4) N2, high MgO (Tgr4) (Murphy, 1981; Peterson, 1984; Nelson, 1985). Except for Tgr2 all of the above units occur in the thesis area. Tgr3 is the most common subunit of Grande Ronde Basalt in the thesis area. Since these units are intrusive in the thesis area and cannot be physically traced into a single type section they are refered to as petrologic types in this report.

Frenchman Springs Member of the Wanapum Basalt

The term Frenchman Springs member was informally proposed by Mackin (1961) for a thick sequence of glomeroporphyritic to porphyritic basalt flows within the Wanapum Basalt of eastern Washington. Beeson et al. (1985) have divided the Frenchman Springs Member into 6 major informal units which include at least 21 different flows. The basalt of Ginkgo and the basalt of Sand Hollow are the only Frenchman Springs units thought to have reached the Pacific Coast. These units consist of up to 11 flows (Beeson et al., 1985).

Peterson (1984) tentatively identified the Ginkgo petrologic type of Mackin (1961; equals basalt of Ginkgo of Beeson et al, 1985) and the Kelly Hollow petrologic type of Bently (1977; equals basalt of Sand Hollow of Beeson et al., 1985) in intrusive rocks of northwest Oregon. Murphy (1981) and Goalen (in prep.) recognized a subaerial Ginkgo flow and at least two aphyric Frenchman Springs Member flows (basalt of Sand Hollow of Beeson et al., 1985). The Frenchman Springs member in the thesis area consists of a single dike that has been tentatively assigned to the basalt of Ginkgo petrologic type based on major oxide chemistry.

Pomona Member of the Saddle Mountains Basalt.

The Pomona Member was named by Schmincke (1967) for a sequence of glomeroporphyritic basalts that crop out on the Columbia Plateau. Kienle (1971) and Schmincke (1967) traced the Pomona Member from where it was erupted in western Idaho as far west as Kelso in western Washington. Wells (1981) mapped an extensive area of subaerial Pomona Member basalts in the vicinity of Cathlamet, southwest Washington and Murphy (1981) mapped a small area of Pomona Member in northeastern Clatsop County.

Pomona Member petrologic type intrusives are present in southwest Washington (basalt of Pack Sack Lookout of Snavely et al., 1973) and have recently been found in Clatsop County (Rarey et al., 1984). In the thesis area Pomona Member petrologic type intrusions have been subdivided into microporphyritic (Tpm) and gabbroic (Tpg) subunits. The gabbroic subunit has a slightly different major oxide chemistry than Pomona Member flows but this is thought to be a result of differentiation and deuteric alteration during slow cooling. Therefore, these intrusions have been included in the Pomona Member petrologic type.

Distribution

Regional

The Columbia River Basalt group contains approximately 2 x 105 km3 of basaltic rock (Swanson and Wright, 1978) erupted from fissures in northeastern Oregon, southeastern Washington, and western Idaho (Taubeneck, 1970). The Grande Ronde Basalt, the Frenchman Springs and Priest Rapids members of the Wanapum Basalt, and the Pomona Member of the Saddle Mountains Basalt are some of the most extensive units in the Columbia River Basalt Group. Subaerial flows from these units, because of their tremendous volume and because of local basin configurations, were able to flow out of the Columbia Plateau, down an ancestral Columbia River Gorge, and into western Oregon and southwestern Washington. Other units of the Columbia River Basalt Group are confined to the Columbia Plateau region (Beeson and Moran, 1979). Figure 91 shows the inferred original distribution of the Columbia River Basalt Group as a whole and the inferred original distribution of the Grande Ronde Basalt, the Frenchman Springs Member, and the Pomona Member.

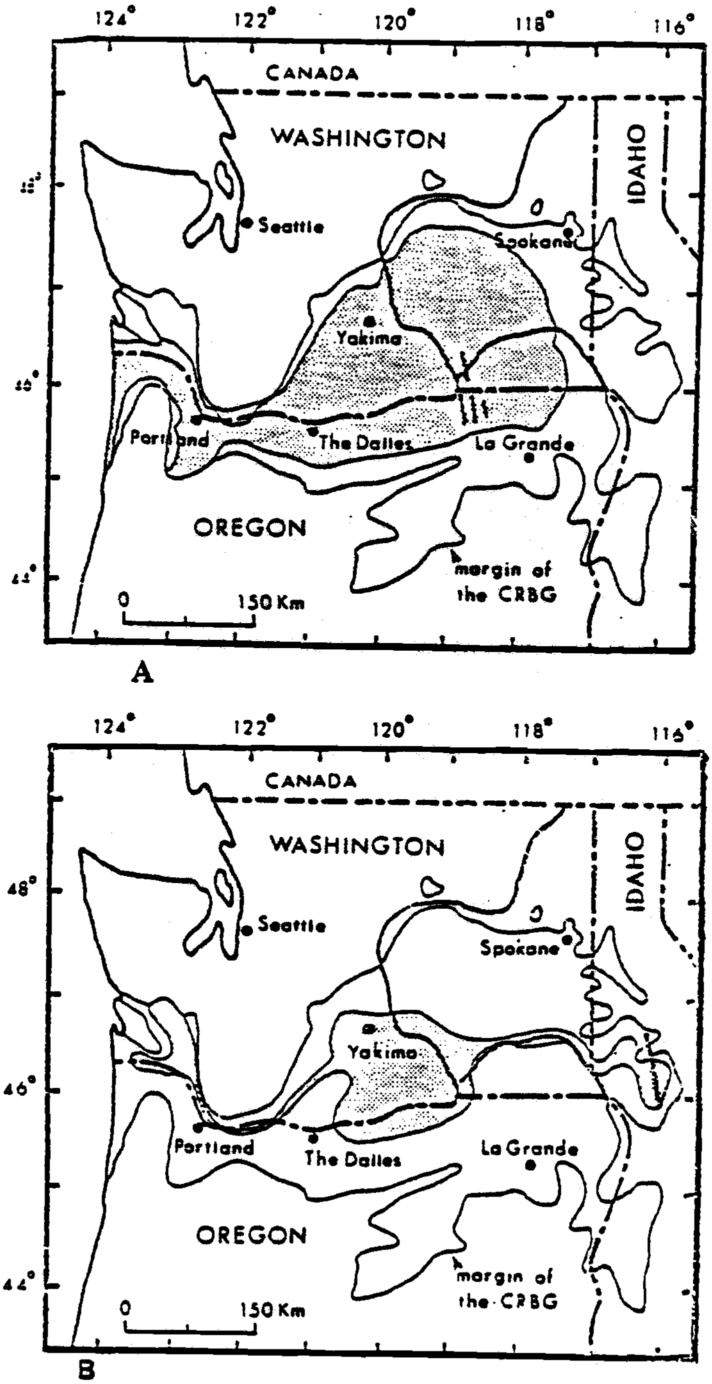


Fig. 91. Distribution of the Columbia River Basalt Group. A) distribution of the Frenchman Springs Member and B) distribution of the Pomona Member. Grande Ronde Basalt distribution is approximately the same as the margin of the CRBG shown in B (modifiedfrom Tolan and Beeson, 1984).

Clatsop County

Subaerial flows of Grande Ronde Basalt, Frenchman Springs Member, and Pomona Member crop out in the northeast corner of Clatsop County (Murphy, 1981; Goalen, in prep.) and thick piles of submarine Grande Ronde Basalt and submarine Frenchman Springs Member crop out in western Clatsop County (e.g., Cressy, 1974; Smith, 1975; Neel, 1976; Penoyer, 1977; Coryell, 1978; Niem and Niem, in press). Numerous intrusions of Grande Ronde Basalt petrologic type occur throughout western and central Clatsop County (e.g., Niem and Niem, in press). Frenchman Springs Member intrusive rocks are less abundant and appear to be restricted to the western parts of the county. Pomona Member petrologic type intrusions are apparently restricted to the southwestern corner of Clatsop County and the northwest corner of Tillamook County. Most Pomona Member petrologic type intrusions occur within the thesis area but reconnaissance work shows that they are present directly south of the thesis area and in the area studied by Smith (1975).

Only a few workers in Clatsop County have used extensive chemical analyses and magnetostratigraphy to divide the Grande Ronde Basalt into subunits (e.g., Murphy, 1981; Peterson, 1984; Nelson, 1985, this study) and, therefore, the distribution of the subunits is only partially understood. Dr. Alan R. Niem (Oregon State University) is currently undertaking a program to subdivide the remainder of Columbia River Basalt Group flows and intrusions in Clatsop County. Nelson (1985) mapped several very long, northeast-trending low MgO-high TiO_2 dikes that project towards the thesis area but terminate before reaching the thesis area.

Thesis Area

A number of Grande Ronde Basalt petrologic type dikes and sills occur throughout the northern and western parts of the thesis area (plate 1, fig. 4). The unit is very well exposed at a number of quarries and along several logging roads. The basalts are much more resistant to erosion than surrounding mudstones and, therefore, commonly form large hills and ridges. Seismic data and well data (CZ 11-28 well, Plate 3) show that one thick (200 ft.) low MgO-low TiO2 (Tgr3) sill and two underlying, smaller high MgO (Tgr4) sills are present in the subsurface at depths greater than 1,500 feet throughout the northern part of the thesis area (plate III). The numerous Grande Ronde Basalt dikes in the northern part of the thesis area range from 1 to 15 m (avg. 5 m) in thickness and appear to "sprout up" from these thick subsurface sills. Middle Miocene dikes throughout the thesis area generally trend northeast (Plate I).

In the southwestern part of the thesis area, a sequence of thick normally polarized low MgO-low TiO2 (Tgr3) sills and a thick underlying high MgO sill (Tgr4) intrude the upper Smuggler Cove formation and the ball park unit (plate 1). A few thin dikes are present above and below these sills.

The Frenchman Springs Member Petrologic type is restricted to a relatively thick (10 m) dike in the northwest corner of the thesis area (along Munce Road, NW1/4 sec. 30, T4N, R8W) (plate 1). The Pomona Member petrologic type, with the exception of one sill, is restricted to the southern part of the thesis area where it intrudes the Hamlet and Smuggler Cove formations. An isolated Pomona Member "sill" intrudes the Smuggler Cove formation in the north-central part of the thesis area (SW1/4 sec. 28, T5N, R8W).

The abundance of Columbia River Basalt Group intrusions and the lack of subaerial flows in the thesis area suggests that subaerial flows entered the marine environment several kilometers to the east of the thesis area and have subsequently been uplifted and eroded away. In all of Clatsop County, Pomona member intrusions have been recognized only in or very close to the southwestern part of the thesis area suggesting that a Pomona Member flow lobe was at one time present to the east of the thesis area. The nearest known Pomona Member subaerial flow outcrop is located in northeastern Clatsop County along the Columbia River some 50 km. northeast of the thesis area (Murphy, 1981). Kadri (1982) suggested that all Columbia River Basalt Group flows were confined to a structural low near the present Columbia River. This suggestion is not consistant with the presence of isolated Pomona member petrologic type intrusions in the thesis area and the occurrence of thick Columbia River Basalt Group flows in the Willamette Valley directly east of the thesis area across the modern day uplifted Coast Range structural high where only older volcanic rocks are exposed (Wells et al., 1983). Prior to emplacement of the basalts the thesis area was the site of the ancesteral Columbia River as is evidenced by the presence of lower to middle Miocene Angora Peak member. Therefore, it is probable that some subaerial Columbia River Basalt Group flows entered the marine environment directly east of the thesis area (i.e., in the vicinity of areas mapped by Safley, in prep. and Mumford, in prep., fig. 1). Other "invasive" dikes and sills were probably sourced from large Columbia River Basalt Group submarine basalt piles (e.g. at Saddle and Humbug mountains).

Lithology

Introduction

Although much of the Columbia River Basalt Group stratigraphy is based on chemistry and magnetostratigraphy, major units in northwest Oregon can initially be differentiated on a lithologic basis. Phenocryst content and size are the main criteria used in differentiating intrusive units. Grande Ronde Basalt is aphyric to very sparsely porphyritic, the Frenchman Springs member contains sparse (1/2-3%) large (up to 1 cm) plagioclase phenocrysts, and the Pomona member contains abundant small (1 mm) plagioclase microphenocrysts and a few larger (1 cm) glomerophenocrysts (Murphy, 1981; Peterson, 1984). Physical features of subaerial flows have been used to define units in the Columbia River Basalt Group, but these features cannot be used when examining intrusive rocks.

Grande Ronde Basalt

Grande Ronde Basalt in the thesis area is finely crystalline (aphyric to very sparsely porphyritic) and dark gray (N 2) to black (N 1). The subunits of the Grande Ronde basalt are lithologically very similar. Previous workers (e.g., Peterson, 1984; Nelson, 1985) have noted that high MgO intrusions tend to be more coarsely crystalline than low MgO intrusions of the same thickness. This relationship appears to be true in the thesis area, but differences are very minor and it is not always possible to deliniate the subunit based on crystallinity. Peterson (1984) noted that the R2 low MgO-high TiO2 petrologic subunit basalts contain a higher percentage of opaque minerals in the groundmass, which can be seen with a hand lens, than the other low MgO subunits. This very subtle difference was not observed in the thesis area.

Grande Ronde Basalt intrusions are typically slightly irregular, commonly pinching and swelling and changing orientation. Intrusions in the older units (i.e., lower Smuggler Cove formation and Keasey Formation) are usually tabular (fig. 92) whereas intrusions higher up in the section tend to be highly irregular and commonly consist of pod-like apophyses (fig. 93). The irregularity of the intrusions in the younger units is probably the result of intrusion into semi-consolidated, water-saturated, marine "mudstone". These "rocks" would tend to deform ductily rather than brittly preventing emplacement along linear fractures.

Cooling margins are thin (<10 cm) on all Grande Ronde Basalt intrusions in the thesis area. Margins may consist of dark glass but in many exposures the glass has been altered to clay. Mudstones adjacent to intrusions are typically baked and bleached light gray (N 7) to white (N 9). Porphyroblastic contact metamorphic textures are common in "mudstones" adjacent to large intrusions in the upper part of the stratigraphic section (fig. 87). The baked zone ranges from 10 cm near very small intrusions to over 20 meters adjacent to very

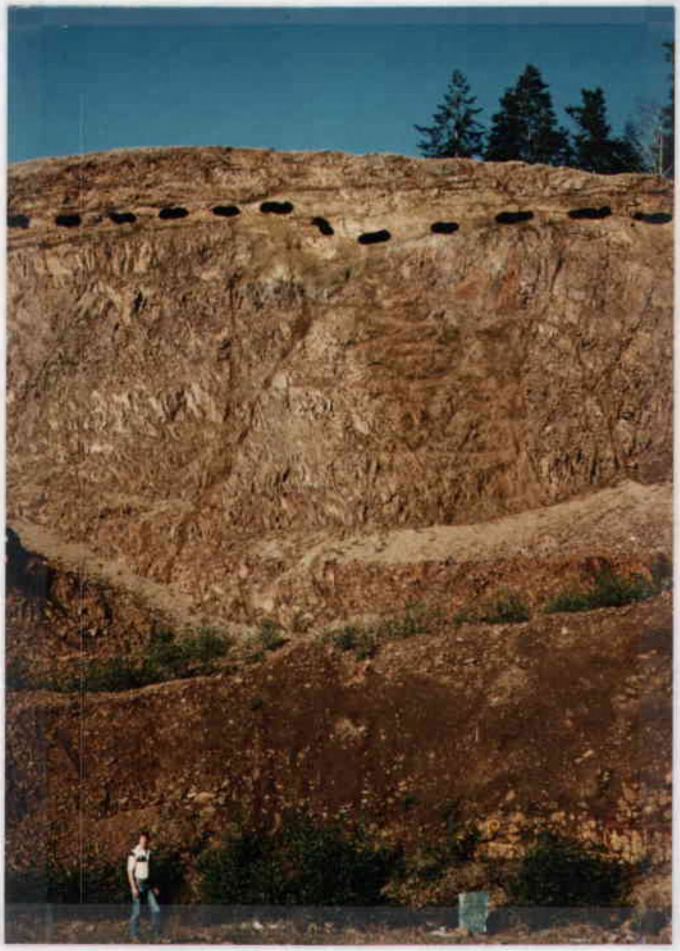


Fig. 92. Quarry exposing thick low MgO, low TiO_2 Grande Ronde Basalt (Tgr3) sill. Note the planar upper contact with the Smuggler Cove formation and the small normal birfricating fault offsetting the units (locality 242).

Fig. 93. Irregular intrusion of low MgO, low TiO2 Grande Ronde Basalt in mudstones of the upper Smuggler Cove formation. "Morman Joe" Lipka for scale (locality 40, SW 1/4 NE 1/4 sec. 19, T5N, R8W).

thick sills (*e.g.* CZ 11-28 well, fig. 80, plate 3). Mudstones adjacent to intrusions are usually deformed; but in some cases, emplacement of the basalts has resulted in little or no deformation.

Grande Ronde Basalt intrusions typically display blocky to platy jointing, but poorly developed columnar jointing is present locally. Smaller intrusions (<2 m thick) are usually highly fractured.

Frenchman Springs Member of the Wanapum Basalt

The Frenchman Springs Petrologic type in the thesis area consists of a single, roughly tabular dike that holds up a northeast trending ridge in the northwestern corner of the thesis area (Plate 1). The dike is relatively thick (approx. 15 m), has blocky jointing, and has baked the surrounding mudstones of the ball park unit of the Smuggler Cove formation a very light gray (N 8).

In hand sample the basalt is sparsely porphyritic (approx. 1% plagioclase phenocrysts), has a finely crystalline groundmass, and is black (N 1). The presence of "common" (1%) phenocrysts serve to distinguish the Frenchman Springs Petrologic type from the nearly aphyric Grande Ronde Basalt petrologic type.

Pomona Member of the Saddle Mountains Basalt

The Pomona Member petrologic type in the thesis area has been subdivided into two informal textural units (Plate I). These are the microporphyritic (Tpm) and the gabbroic (Tpg) subtypes. Two sills of the microporphyritic unit (Tpms) occur in the thesis area (localities

20 and 357). The sills are basaltic, dark gray (N 2) to black (N 1), and contain plagioclase as small glomerophenocrysts and microphenocrysts. Microphenocrysts are small (1 mm) but can readily be seen with a hand lens. On weathered surfaces the microphenocrysts appear as abundant small white laths. The white color is a result of alteration of the plagioclase to clay. One sill at quarry locality 357 is approximately 5 meters thick, has sharp contacts with the surrounding mudstones of the Sweet Home Creek member, and is roughly tabular. The other sill (locality 20) has a sharp but irregular contact with the surrounding mudstones of the Smuggler Cove formation and is approximately 7 meters thick. Jointing in both sills is blocky to faintly columnar.

The gabbroic subtype is typically well-weathered, has a mottled appearence, and is medium gray (N 4) to light gray (N 7). However, where the unit is fresh (i.e., stream bed localities such as 617) the rock is dark gray (N 2). The unit is characterized by a gabbroic to microgabbroic texture with individual crystals ranging from 1 mm to 1 cm in length. Pyroxene, plagioclase, and rare olivine can be observed in hand sample. Thick sills (5 m to 10 m) are most common (localities 297, 310, 596) but a thin (3 m) dike is also present (locality 313). Jointing is typically blocky but faint columnar jointing is also present. Cooling margins are thin (<2 cm) and glassy. Adjacent mudstones are typically bleached to a very light gray (N 8) and are well-indurated.

Petrography

Introduction

Twenty-seven surface and subsurface samples of middle Miocene Columbia River Basalt Group were thin-sectioned and analyzed with a petrographic microscope (appendix 8). Three of the samples are sidewall core samples from the CZ 11-28 well adjacent to the thesis area. Thirteen samples were point-counted (300 to 500 points) using a mechanical stage and modal abundances were visually estimated in the remaining samples. Because of the relatively low number of of points counted there is a several percent error in the abundances reported in appendix 8. All of the samples examined have the petrographic features of basalts or gabbros as described below.

Petrographic analysis was used to determine if the chemical differences between subunits are reflected in petrographic features. Thirteen of the Grande Ronde Basalt thin sections were from localities where chemical data were available and seven were from localities without chemical data. It was found that petrographic features can be used to make tentative subunit assignments but that chemical data and magnetic polarity data are needed to positively assign an intrusion to a specific subunit.

Petrographic analysis was also used to compare "invasive" dikes and sills of the Columbia River Basalt Group to presumably correlative subaerial flows described by Murphy (1981) to the northeast of the thesis area. When possible, petrographic and chemical samples were collected from the central part of intrusive bodies. Thin as well as thick intrusions were sampled and compared. There are no apparent mineralogical differences between subaerial flows and "invasive" dikes and sills. The only petrographic differences noted are the presence of occasional large vesicles in the flows and a coarser texture in several very thick sills. As was stated in previous sections, the Columbia River Basalt Group is petrographically distinct from the Eocene volcanic rocks in the thesis area.

Grande Ronde Basalt

The high MgO subunit of Grande Ronde Basalt (Tgr4) consists of very rare, small plagioclase phenocrysts (approx. 1 per square meter) set in an intergranular to subophitic groundmass. The groundmass consists of plagioclase microlites (45-50%), clinopyroxene (41-47%), and opaque minerals (9-11%) (fig. 92b). Plagioclase is labradorite (An 57-59) that varies in length from 0.1-0.8 mm. The clinopyroxene is augite (2vz 45-55) which is anhedral to subhedral, ranging from 0.1 to 1.5 mm in diameter. Chlorophaeite rims some augite crystals. Opaque minerals appear to be magnetite and ilmemite which are typically anhedral and range in size from 0.1 to 0.5 mm. Minor amounts of an interstitial mixture of quartz and alkaline feldspar is present in sample 523.

Low MgO petrologic type basalts (Tgr1 and Tgr3), where subunit assignment has been confirmed using chemical analysis, consist of very rare plagioclase phenocrysts (approx. 1 per square meter) set in an intersertal to hyalopilitic groundmass. Smaller intrusions (2-7 m thick) have a hyalopilitic texture whereas a sample from a thick (approx. 40 m) sill at locality 242 south of Cole Mountain has an intersertal texture. Groundmass of low MgO basalts consists of plagioclase microlites (30-55%), clinopyroxene (0-33%), opaque minerals (0-9%), and glass (4-55%) (fig. 94, appendix 8). Plagioclase microlites are labradorite (An 53-57) that range in length from 0.1-0.7 mm. Clinopyroxene is anhedral to subhedral augite (2vz 45-55) that is typically very small (<0.1 mm) but ranges in diameter from <0.1 to 1 mm. Opaque minerals, probably magnetite and ilmenite, occur as finely dissiminated dust in basaltic glass and as small intergranular grains. Glass is typically black to light brown and, therefore, probably consists of tackylite and sideromelane.

Peterson (1984) noted that low MgO-high TiO_2 (Tgr1) basalts contain very minor anhedral hypersthene and typically contain more than 10% opaque minerals whereas low MgO-low TiO_2 (Tgr3) basalts generally contain less than 7% opaque minerals. These differences were not observed in the thesis area. Only one high MgO-high TiO_2 intrusion occurs in the thesis area. A thin section from this intrusion (sample 2) reveals that it is petrographically "identical" to the low MgO-low TiO_2 basalt examined.

Low MgO basalts, as can be seen from the data above and on figure 94, tend to contain more glass and are more finely crystalline than the high MgO samples from intrusions of similar thickness. A comparison of holocrystalline or nearly holocrystalline samples shows that high MgO basalts contain less plagioclase and more augite than low MgO basalts. High MgO basalts also tend to contain a higher

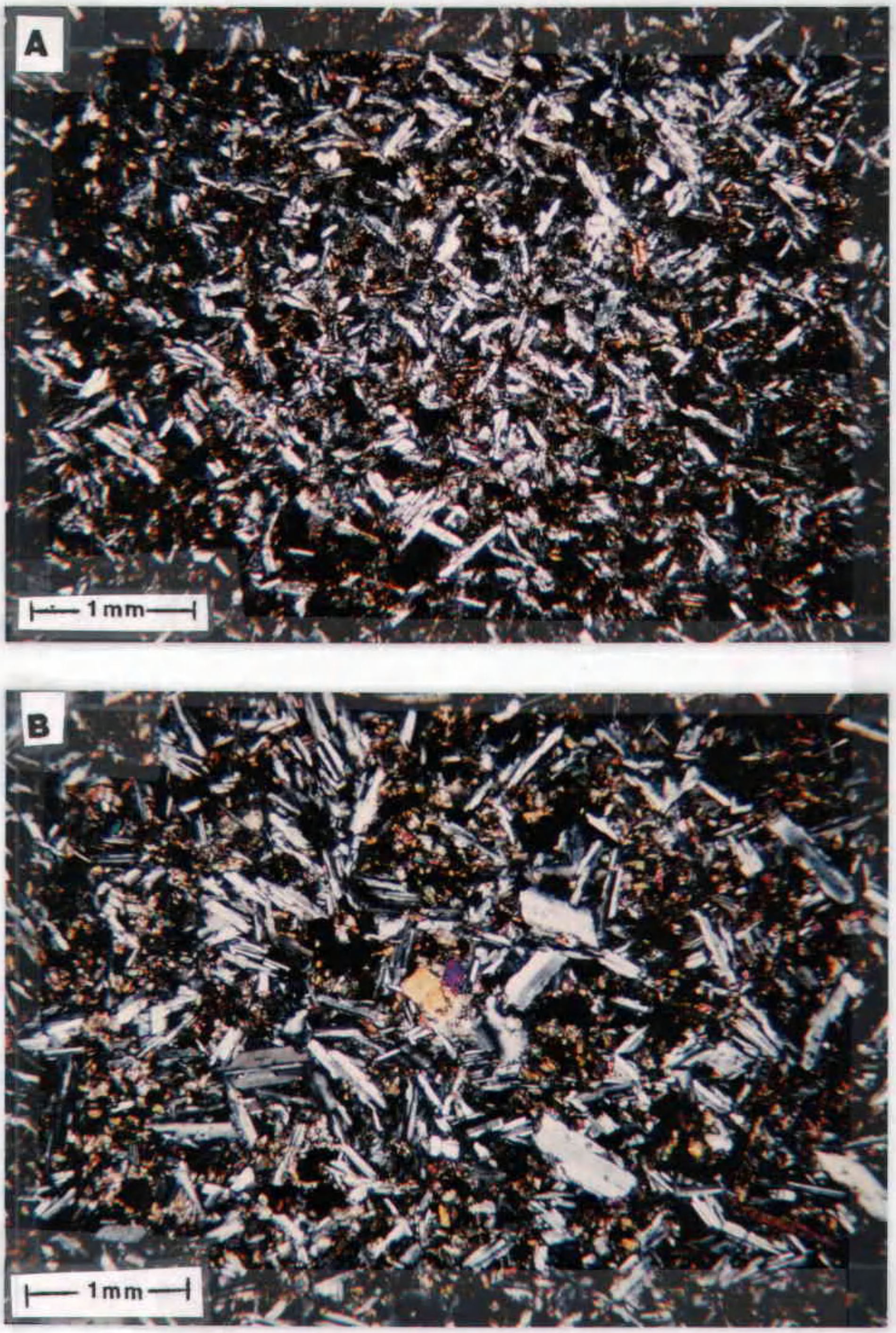


Fig. 94. Photomicrographs (crossed nicols) of A) glass-rich low MgO, low TiO2 Grande Ronde Basalt (Tgr3)(locality 40) and B) more coarsely crystalline high MgO Grande Ronde Basalt (Tgr4)(locality 131).

percentage of opaque minerals. The greater amount of MgO in the high MgO basalts is, therefore, probably reflected in the greater abundance of augite and opaque minerals.

Based on petrographic characteristics, six samples from localities without chemical data have been tentatively assigned to the low MgO subtype (appendix 8). All of these samples have normal polarity and, therefore, are probably low MgO-low TiO2 (Tgr3) basalts. One sample (520) has been tentatively assigned to the High MgO subunit based on petrographic characteristics and proximity to other high MgO localities.

Snavely et al. (1973) noted that the petrographic characteristics of the Depoe Bay Basalt (Grande Ronde Basalt petrologic type of this study) are "identical" to those of the Grande Ronde Basalt on the Columbia Plateau. Since 1973, the petrographic characteristics of the of the Depoe Bay Basalt have been studied by Tolson (1976), Penoyer (1977), Coryell (1978), Murphy (1981), Olbinski, 1983; Peterson (1984), and Nelson (1985). Data from the above studies and from this study show that the Grande Ronde Basalt petrologic type exposed on the coast is petrographically indistinguishable from the Grande Ronde Basalt of the Columbia Plateau studied by Waters (1961), Swanson (1967), and Wright et al. (1973) and from the Grande Ronde Basalt of the eastern Willamette Valley described by Anderson (1978).

Frenchman Springs Member of the Wanapum Basalt

One dike in the thesis area has been assigned to the Frenchman Springs Member. This basalt dike is sparsely porphyritic with a subophitic groundmass. It consists of labradorite phenocrysts (1/2%), labradorite microlites (46%), augite (38%), opaque minerals (9%), and minor amounts of basaltic glass (fig. 95). Labradorite phenocrysts range from 4 mm to 1 cm in length. The Frenchman Springs Member is petrographically similar to aphyric high MgO Grande Ronde Basalt but contains approximately 1/2% phenocrysts. Frenchman Springs Member subaerial flows studied on the Columbia Plateau (Waters, 1961; Wright _et al._, 1973) have similar petrographic characteristics as the Frenchman Springs Member petrologic type dike in the thesis area.

Beeson _et al._ (1985) have subdivided the Frenchman Springs member into several subunits based primarily on chemical composition. They noted that the petrographic characteristics of the different subunits were variable. The Frenchman Springs Member dike in the thesis area, on the basis of major oxide chemistry, has been tentatively assigned to the basalt of Ginkgo of Beeson _et al._ (1985). The dike in the thesis area is, however, less porphyritic than "typical" Ginkgo basalts which average approximately 5% labradorite phenocrysts.

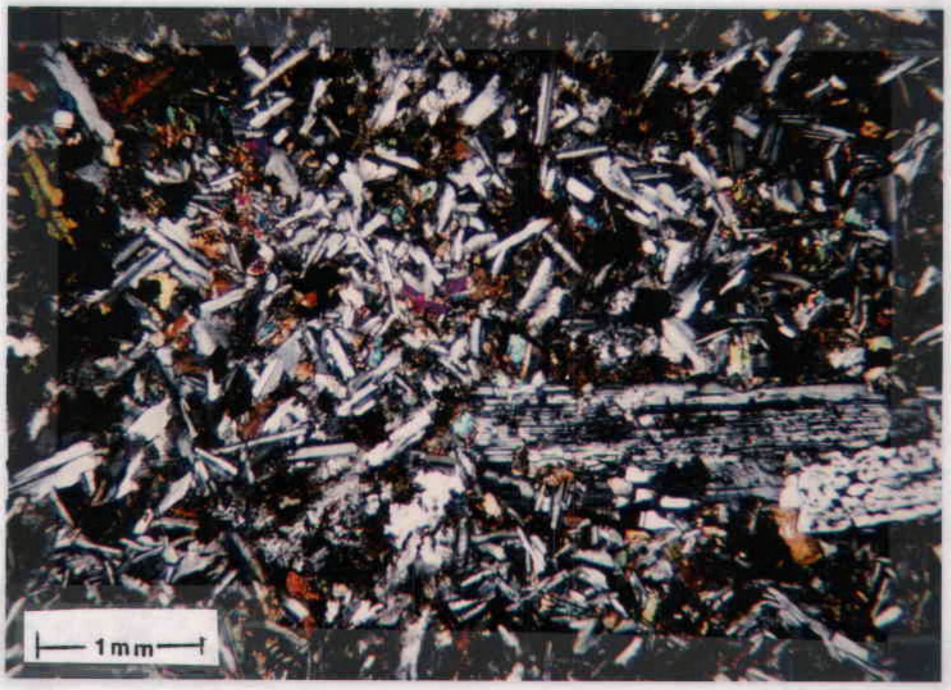


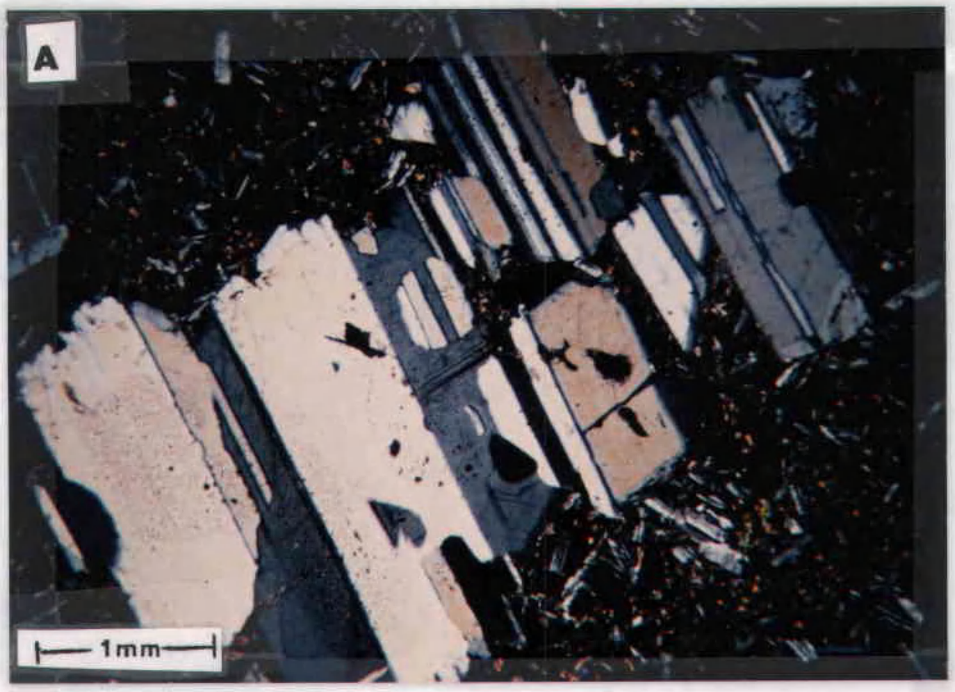
Fig. 95. Photomicrograph (crossed nicols) of Frenchman Springs Member basalt. Note the large incompletely crystallized labradorite phenocryst at the right surrounded by groundmass of labrodorite, muscolite, augite, opaque ilmenite, and basaltic glass (sample 76).

Pomona Member of the Saddle Mountains Basalt

The microporphyritic subunit of the Pomona Member petrologic type (Tpm) consists of plagioclase (45-50%), augite (25-42%), opaque minerals (5-7%), and glass (3-22%) (appendix 8). Textures are microporphyritic to sparsely glomeroporphyritic with an intersertal groundmass. The plagioclase is labradorite (An 54-63) which occurs in three distinct sizes: 1) as large (3 mm to 2 cm) phenocrysts and glomerophenocrysts; 2) as microphenocrysts (approx. 0.7 mm in length); and 3) as microlites (0.1 to 0.3 mm) (fig. 96). The presence of large glomerophenocrysts and abundant microphenocrysts serves to distinguish the Pomona Member microporphyritic subtype from other petrologic units in the Columbia River Basalt Group of northwestern Oregon.

The gabbroic subunit (Tpg) is compositionally similar to the microporphyritic subunit (Tpm) but has an ophitic texture and contains moderate amounts (up to 8%) of olivine (appendix 8, fig. 97). Subhedral to euhedral labradorite (An 58-64) comprises from 40-63% of the rock and ranges from 0.2 mm to 3 mm in length. Augite comprises from 22-29% of the rock and occurs as large (0.5 to 4 mm) subhedral crystals that frequently envelop labradorite microlites. Opaque minerals comprise from 5 to 7% of the rock and are relatively large (0.2-0.5 mm). Olivine is typically euhedral, fairly large (2-3 mm), and unaltered (fig. 97).

The texture and mineralogy of the gabbroic subunit (Tpg) can probably be attributed to slow cooling and crystal fractionation of the same "magma" that produced the microporphyritic subunit



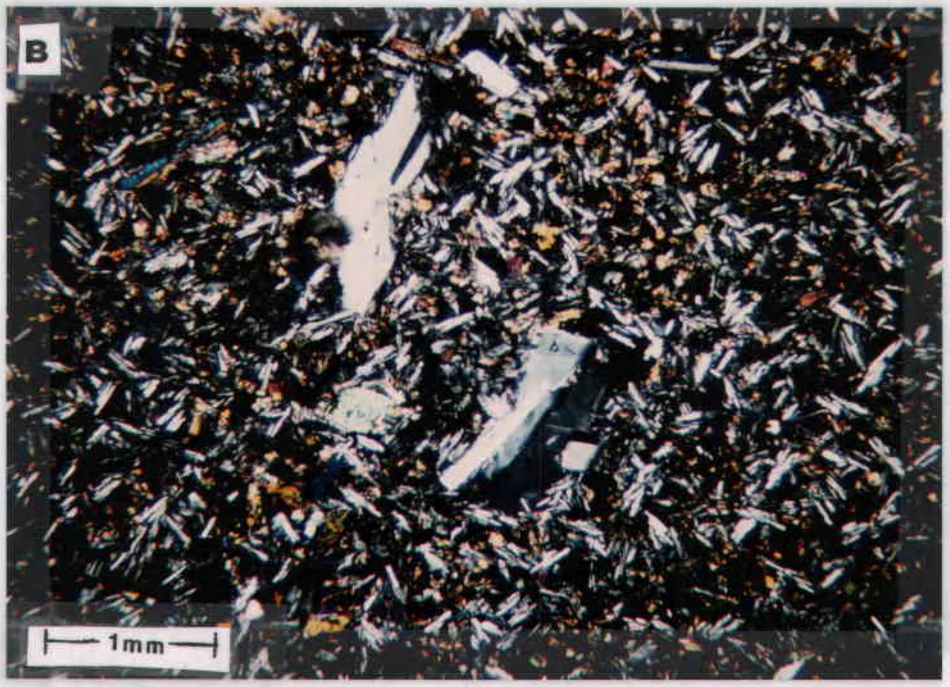


Fig. 96. Photomicrographs (crossed nicols) of the Pomona Member microporphyritic subunit. A) showing albite twinned labradorite glomerophenocryst (sample 20) and B) showing labradorite microphenocrysts in a groundmass of plagioclase, augite, and basaltic glass (sample 357).

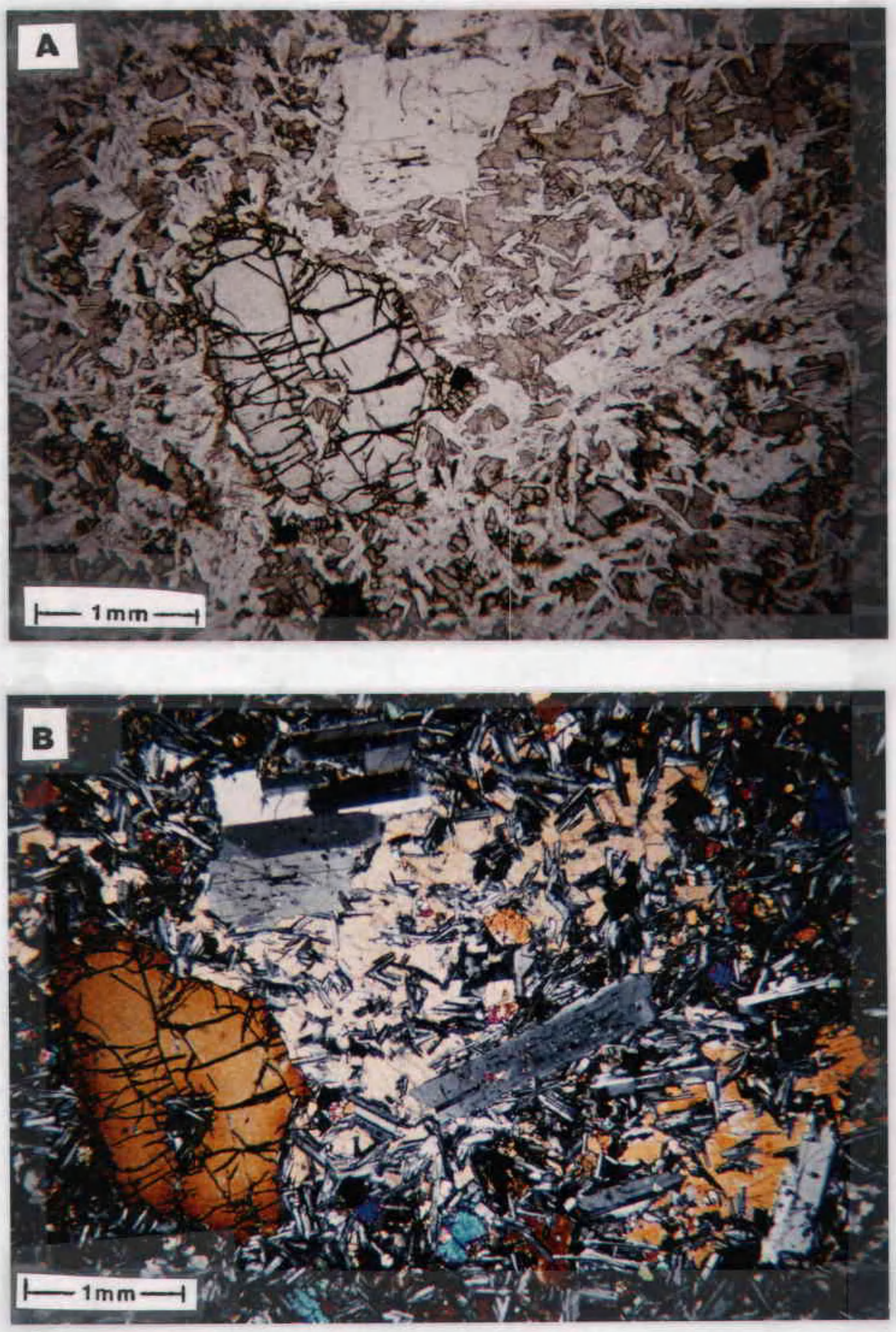


Fig. 97. Photomicrograph of the Pomona Member gabbroic subunit (Tpg). Note the ophitic glomerophyritic texture and the large fractured olivine phenocryst. A) plane polarized light and B) crossed nicols (sample 310).

(Tpm). In the extreme southwestern part of the thesis area just above the Tillamook Volcanics there are two exposures of Pomona Member sills which intrude the same stratigraphic horizon in the Sweet Home Creek member. A gabbroic subunit sill is exposed at locality 310 along Gods Valley Road and a microporphyritic subunit sill is exposed at locality 357 located approximately 2 km. to the east. These relationships suggest that the two exposures are of the same sill and that textural differences are the result of different cooling histories. The microporphyritic subunit exposure, as is evidenced by structural position, was probably emplaced at a higher elevation than the gabbroic subunit exposure and appears to be relatively thin (approx. 5 m) compared to the gabbroic exposure (approx. 10 m).

The Pomona Member in the type area on the Columbia Plateau consists of labradorite, augite, and olivine glomerophenocrysts and microphenocrysts set in an intersertal groundmass composed of labradorite, augite, opaque minerals, and basaltic glass (Schmincke, 1967). The microporphyritic subunit (Tpm) in the thesis area is petrographically similar to Pomona Member flows on the Columbia Plateau. The only significant difference is that olivine was not observed in the microporphyritic subunit. Murphy (1981) reported that a Pomona Member subaerial flow in northeastern Clatsop County contained approximately (3%) olivine. The gabbroic subunit (Tpg) is mineralogically similar to the Pomona Member flow described by Murphy (1981) except that it contains more olivine (up to 8%). This difference is probably due to differentiation during slow cooling.

Magnetostratigraphy

The polarity of 150 Columbia River Basalt Group samples from 52 localities was determined with a fluxgate magnetometer (appendix 7). From two to four samples were analyzed from each locality. In almost all cases samples from the same locality yielded the same polarity readings. Localities with conflicting or very weak readings have querried polarities in appendix 7.

Most localities were found to have normal polarities with reversed polarities being restricted to localities 2, 4, 314, 320, and 357. Localities 314, 320, and 357 are the Pomona Member petrologic type, indicating that the unit has a reversed polarity in the thesis area. Pomona Member locality 310, however, has a questionable normal polarity. The rock here is moderately weathered and microgabbroic. Slow cooling and/or weathering may have affected the polarity determination. Previous workers have shown that the Pomona Member in western Oregon, southwestern Washington, and on the Columbia Plateau is reversely polarized (e.g., Snavely et al., 1973; Beeson and Moran, 1979; Murphy, 1981).

The other remaining reversely polarized localities (2 and 4) are in a low MgO-high TiO_2 Grande Ronde Basalt dike. Chemical analysis was used to determine the subunit assignment of this dike. Murphy (1981), Peterson (1984), Nelson (1985) and Goalen (in prep.) have noted that the low MgO-high TiO_2 subunit in Clatsop County has a reversed polarity.

Intrusions with normal polarity include high MgO Grande Ronde Basalt (Tgr4), low MgO-low TiO_2 Grande Ronde Basalt (Tgr3), and the

Frenchman Springs Member Basalt. Numerous studies on the Columbia Plateau and in northwestern Oregon have shown that high MgO Grande Ronde basalt and Frenchman Springs Member have normal polarities (e.g., Beeson and Moran, 1979). Beeson and Moran (1979) recognized that subaerial flows of low MgO-low TiO2 Grande Ronde Basalt in western Oregon could be separated into two magnetostratigraphic units based on polarity. Peterson (1984) mapped both reversely polarized (Tgr2) and normally polarized (Tgr3) low MgO-low TiO2 intrusions to the north of the thesis area. Within the thesis area, however, all the low MgO-low TiO2 Grande Ronde Basalt has normal polarity. This indicates that the the reversely polarized subunit of Peterson (1984) (Tgr2) is not present in the thesis area.

Swanson and Wright (1978) and Swanson et al. (1979) devised a magnetic stratigraphy for the Columbia River Basalt Group. The Grande Ronde Basalt was divided into the R1, N1, R2, and N2 magnetostratigraphic units (fig. 90). R1 Grande Ronde Basalt flows are apparently restricted to the Columbia Plateau with N1, R2, and N2 flows being more voluminous and occurring on the Columbia Plateau as well as in western Oregon and southwestern Washington (Beeson and Moran, 1979). Therefore, the reversely polarized low MgO-high TiO2 subunit (Tgr1) in the thesis area is considered to be correlative to the R2 zone and both the High MgO (Tgr4) and the low MgO-low TiO2 (Tgr3) subunits are thought to be correlative to the N2 magnetic zone. The Frenchman Springs Member of the Wanapum Basalt has also been assigned to the N2 zone by Swanson et al. (1979).

In conclusion, paleomagnetic polarity studies of the middle Miocene intrusive rocks in the thesis area show that the chemical units of the Columbia River Basalt Group have consistent polarities. This indicates that the various chemical subtypes are good stratigraphic units (i.e. each subunit is genetically related and emplaced at approximately the same time) rather than an artificial grouping of chemically similar basalts.

Chemistry

Twenty seven Columbia River Basalt Group samples were collected and prepared for major oxide chemical analysis. The samples were analyzed using x-ray fluorescence and standardized against the U.S. Geological Survey silicate standard BCR-1 (Columbia River Basalt standard) by Dr. Peter Hooper's lab at Washington State University. The BCR-1 standard has been used on almost all published analyses of Columbia River Basalt, thereby allowing for accurate comparison of major oxide chemical data. Olbinski (1983), Peterson (1984), and Nelson (1985) had Columbia River Basalt Group samples analyzed against the "international" standard of Dr. Hooper. Goalen (in prep.) devised correction factors to convert international standard values to BCR-1 values. These correction factors are listed by Peterson (1984). The chemical analyses used in this report from Peterson (1984) and Nelson (1985) have been corrected to BCR-1 values using the correction factors.

Analyses from the thesis area were plotted on a silica variation diagram (fig. 98) and a plot of TiO_2 versus MgO (fig. 99). Snavely et al. (1973) defined compositional limits for the Grande Ronde Basalt (Depoe Bay Basalt), Frenchman Springs Member (Cape Foulweather

Silica Variation Diagram For Miocene Basalts

Hamlet Area, Northwest Oregon

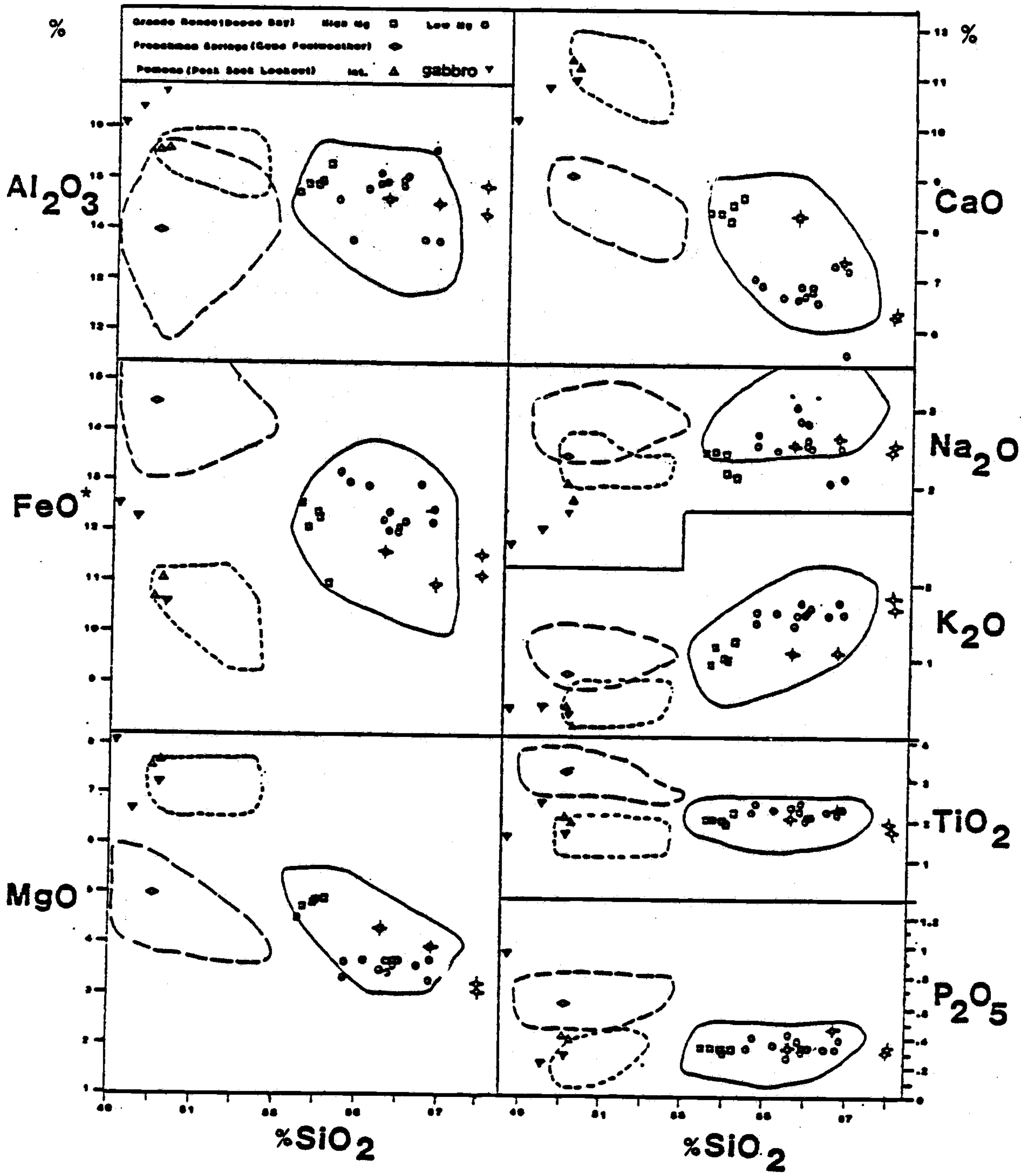


Fig. 98. Silica variatrion diagram with Miocene intrusive samples from the thesis area plotted. Grande Ronde-Depoe Bay, Frenchman Springs-Cape Foulweather, and Pomona-Pack Sack Lookout fields shown (fields from Snavely _et al._, 1973).

Basalt), and the Pomona Member (basalt of Pack Sack Lookout) using a silica variation diagram. With several exceptions, samples from the thesis area plot within the compositional limits of Snavely et al. (1973) therby confirming the field identifications and demonstrating a certain degree of chemical homogenaeity for the various units. As Snavely et al. (1973) originally pointed out the plot also shows that the "coastal basalts" and the "plateau basalts" have essentially identical major element chemistries.

Two Grande Ronde Basalt well samples from the CZ 11-28 well and two samples from the gabbroic subunit of the Pomona Member plot significantly outside the fields defined by Snavely et al. (1973). The two anomolous Grande Ronde Basalt samples from the well are chemically similar to other Grande Ronde Basalt samples except that they contain higher percentages of of SiO_2. The higher SiO_2 values are probably due to either alteration or to contamination of the basalt samples with mudstone cuttings. One sample (310) of the gabbroic subunit of the Pomona member consistently plots within the the Pomona Member field of Snavely et al. (1973) whereas the other two samples of the gabbroic subunit plot near but outside the field. It is suggested that the chemical differences in these two coarsely crystalline samples is due to differentiation during slow cooling. It is possible, but not likely, that the intrusions mapped as the gabbroic subunit are not related to the Pomona Member. Pomona Member intrusives are located near and are chemically similar to the gabbroic intrusions and, therefore, it is this writer's opinion that the gabbroic intrusions are related to the Pomona Member. Gabbroic subunit rocks intrude late Oligocene to early

Miocene sedimentary rocks in the southwestern corner of the thesis area and, therefore, are at least approximately the same age as Pomona Member basalts (Plate I).

Murphy (1981), Olbinski (1983), Peterson (1984), and Nelson (1985) used a TiO_2 versus MgO diagram to differentiate the various subunits of Grande Ronde Basalt and to differentiate Grande Ronde Basalt from other units in the Columbia River Basalt Group (fig. 99). Samples from the thesis area plot within all of the fields outlined by Murphy (1981) with the majority of samples plotting in the low MgO-low TiO_2 Grande Ronde Basalt field. One sample from CZ 11-28 well cuttings (CZ-3) is considered to be high MgO basalt even though it plots within the low MgO field on figure 99. All samples from the well appear to be depleted in in MgO as a result of contamination or alteration, thereby shifting the plotted positions to the left on figure 99. Because of the apparent MgO depletion and because another sample (CZ-4) collected from the same well horizon plots in the high MgO field, sample CZ-3 has been assigned to the high MgO petrologic type. The two well samples assigned to the low MgO-low TiO_2 subunit plot slightly outside the low MgO field. This is probably a result of contamination of basalt cuttings by mudstone cuttings.

Two distinct clusters of samples occur within the low MgO-low TiO_2 field on figure 99. Four samples cluster at TiO_2 values of 2.3%. This cluster is labeled subunit S because all the samples are from intrusions in the southern part of the thesis area. Six samples from the northern part of the thesis area cluster at TiO_2 values of 2.1% and have been labeled subunit N on figure 99. One sample (51)

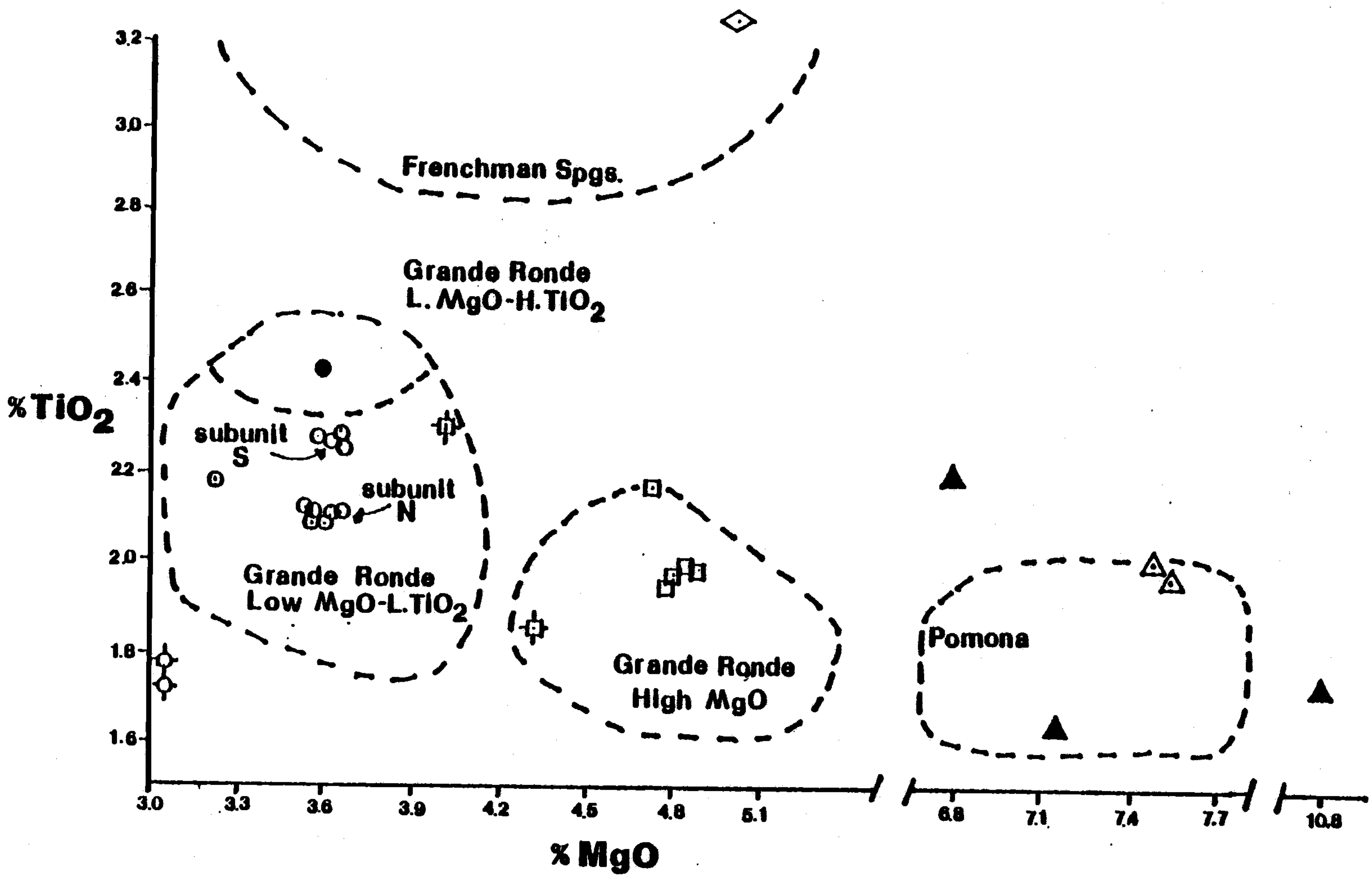


Fig. 99. TiO2 vs. Mgo plot for Miocene intrusives in the thesis area. Grande Ronde Basalt (low MgO, high TiO2 ●; low MgO, low TiO2 ○; and high MgO ◻). Frenchman Springs Member (◇), and Pomona Member (microporphyritic subunit ▲ and gabbroic subunit ▲) samples plotted. Fields from Murphy (1981).

from the northern part of the thesis area plots away from both clusters. Beeson et al. (1979) suggest that several different low MgO-low TiO2 flows entered the marine environment in the vicinity of the thesis area and formed "invasive" dikes and sills. The two subunits (N and S) in the low MgO-low TiO2 petrologic type may represent the invasion of two different normally polarized low MgO-low TiO2 flows with slightly different chemistries in different parts of the thesis area.

The two Pomona Member gabbroic subunit samples (297 and 617) that plotted outside of the Pomona Member field on the silica variation diagram (fig. 98) also plot outside the Pomona Member field on the TiO2 versus MgO diagram (fig. 99). This is probably due to differentiation and deuteric alteration during slow cooling as was previously mentioned.

The average major element chemical composition of the Columbia River Basalt Group intrusive units in the thesis area was compared to the average composition of the same units outside the thesis area (tables 5 and 6). This was done in order to establish average compositions of the various units and to compare the chemical composition of "coastal" intrusions to subaerial flows.

Table 5 shows that low Mgo-low TiO2 Grande Ronde Basalt (Tgr2 and Tgr3) and low MgO-high TiO2 Grande Ronde Basalt (Tgr1) have very similar major element chemistries. The only significant difference between the two units is in TiO2 content. Low TiO2 basalts average from 2.0 to 2.15% TiO2 whereas high TiO2 basalts average from 2.35 to 2.5% TiO2. There is also a tendency for high TiO2 basalts to contain more K2O than low TiO2 basalts. Even though there is

Table 5

Major Oxide Content of Subunits within the Grande Ronde Basalt: Comparison of Subaerial Flows to "coastal" intrusions

Unit	Low MgO, Low TiO_2						Low MgO, High TiO_2					High MgO					
	Intrusions+					Flows	Intrusions+				Flows	Intrusions+				Flows	
Col no	Snav.	Pet.	Nels.	Murp.	This	Bees.	Snav.	Pet.	Nels.	This	Bees.	Snav.	Pet.	Nels.	This	Bees.	Swan.
No. of anal.	26	8	9	6	11*	30	1	4	7	1*	7	13	5	4	5*	17	13
SiO_2	52.1	56.9	56.2	56.4	55.9	55.9	55.5	56.1	55.8	54.8	55.1	55.0	53.9	54.2	53.9	53.9	54.0
Al_2O_3	14.1	14.4	15.0	14.3	14.8	15.1	14.0	13.8	14.9	14.9	14.8	14.4	14.8	15.0	14.9	14.3	14.5
TiO_2	2.0	2.1	2.1	2.0	2.2	2.1	2.5	2.4	2.4	2.4	2.3	1.9	2.0	2.0	2.1	1.9	1.8
Fe_2O_3	2.6	----	----	----	2.0	2.0	2.4	----	----	2.0	2.0	2.1	----	----	2.0	2.0	----
FeO	9.8	----	----	----	10.4	9.7	9.9	----	----	11.0	10.4	9.5	----	----	10.1	9.7	----
MnO	0.21	0.18	0.19	----	0.19	0.20	0.22	0.19	0.18	0.19	0.20	0.21	0.24	0.22	0.21	0.20	----
CaO	6.9	7.0	7.0	6.9	6.7	7.0	6.6	7.2	5.2	6.9	6.8	7.9	8.9	8.4	8.3	8.5	9.1
MgO	3.4	3.2	3.4	3.8	3.6	3.4	3.5	3.2	3.4	3.6	3.3	4.3	4.4	4.5	4.8	5.0	5.3
K_2O	1.4	1.6	1.6	1.7	1.6	1.6	2.3	2.0	1.9	1.5	1.9	1.0	1.3	1.1	1.0	1.2	1.1
Na_2O	3.5	2.3	2.1	3.1	2.5	2.7	2.9	2.3	2.4	2.4	2.8	3.3	2.1	2.2	2.4	2.3	2.8
P_2O_5	0.40	0.30	0.30	----	0.33	0.33	0.34	0.33	0.32	0.40	0.40	0.32	0.26	0.27	0.30	0.28	----

Snav. - Snavely *et al.* (1973)
Pet. - Peterson (1984)
Nels. - Nelson (1985)
Murp. - Murphy (1981)
This - This Study

Bees. - Beeson, personal comm. to Peterson (1984)
Swan. - Swanson *et al.* (1979)

* Well samples not included because of possible contamination

\+ Includes several "coastal basalt" flows from Snavely, *et al.* (1973)

Table 6

Average Major Oxide Concentrations of Frenchman Springs and the Pomona Member: Comparison of Subaerial Flows to "Coastal" intrusions

Unit	Frenchman Springs Member						Pomona Member					
	Intrusions†				Flows		Intrusions			Flows		
Data	Snav.	Murp.	Pet.	This	Bees.	Swan.	Snav.	Hill	This	Murp.	Swan.	Wat.
No. of anal.	21	7	11	1	46	21	10	4	5	1	30	7
SiO_2	51.9	52.0	52.7	50.3	51.3	51.5	52.3	----	49.6	52.7	52.1	51.1
Al_2O_3	13.9	13.5	13.9	13.9	14.7	14.2	15.1	13.4	16.2	14.9	14.9	15.1
TiO_2	3.0	3.0	2.6	3.2	3.0	3.0	1.6	1.7	1.9	1.7	1.6	----
Fe_2O_3	3.6	----	----	2.0	2.0	----	1.7	----	2.0	----	----	2.5
FeO	10.9	----	----	12.5	12.5	----	8.8	----	9.4	----	----	8.6
MnO	0.22	----	0.23	0.23	0.23	----	0.18	----	0.18	----	----	0.20
CaO	7.9	8.2	8.2	9.0	8.2	8.6	10.6	10.0	10.7	10.3	10.7	10.9
MgO	4.1	4.3	3.7	5.0	4.1	4.6	6.9	6.7	8.0	7.2	7.0	7.0
K_2O	1.0	1.3	1.3	0.8	1.1	1.2	0.56	0.80	0.30	0.70	0.64	0.62
Na_2O	3.0	2.7	2.2	2.3	2.0	2.7	2.3	2.5	1.45	2.3	2.4	2.3
P_2O_5	0.69	----	0.60	0.61	0.54	----	0.28	----	0.28	----	----	0.16

† Includes several "coastal basalt" flows from Snavely, et. al. (1973)

Snav. - Snavely et. al. (1973)
Pet. - Peterson (1984)
Murp. - Murphy (1981)
This - This Study
Bees. - Beeson, personal comm.
Swan. - Swanson et. al. (1979)
Hill - Hill (1975)
Wat. - Waters, written comm. to Snavely, et. al. (1973)

relatively little chemical difference between the high TiO_2 basalts and the low TiO_2 basalts, it appears that they represent distinct units. This is evidenced by the reversed polarity of the high TiO_2 basalt and the normal polarity of the low TiO_2 basalt in the thesis area. In addition, TiO_2 values are generally very accurate, and variations in the values probably represent real differences rather than analytical error.

High MgO Grande Ronde Basalt (Tgr4) is higher in CaO and MgO and lower in SiO_2 and K_2O than low MgO Grande Ronde Basalt (table 5). Trace element data from the thesis area and from other workers show that high MgO Grande Ronde Basalt contains more Sc and Co and less Th, La, and Sm than low MgO basalts (table 7).

Frenchman Springs Member basalts are higher in total iron, TiO_2, and P_2O_5 and lower in SiO_2 than any units of the Grande Ronde Basalt (table 6, fig. 98). Peterson (1984) divided the Frenchman Springs Member in northwest Oregon into two informal units based on phenocryst abundance and P_2O_5 content. These subunits are the phenocryst-poor (<1%), lower P_2O_5 (0.46-0.51%) Kelly Hollow basalt and the phenocryst-rich (>1%), higher P_2O_5 (0.56-0.64%) Ginkgo basalt. Beeson et al. (1985) have recently recomended that the name Kelly Hollow be abandoned and that basalts formally assigned to this unit should be included in the basalt of Sand Hollow. Beeson et al. (1985) divided the basalt of Sand Hollow into low P_2O_5 (0.46% to 0.50%) and intermediate P_2O_5 (0.51% to 0.53%) chemical subtypes. The Ginkgo basalt has been renamed the basalt of Ginkgo by Beeson et al. (1985).

The Frenchman Springs Member dike in the thesis area contains

Table 7

Comparison of Trace Element Averages of Low MgO Grande Ronde Basalt to high Mgo Grande Ronde Basalt

Unit	Low MgO Grande Ronde Basalt						High MgO
	Intrusions				Flows		Flows
Data Source	Murp.	Hill	This	And.	Swan.	Murp.	And.
No. of anal.	5	3	1	17	1	3	7
Sc (ppm)	30.40	31.00	30.05 ± .08	34.00	31.20	29.00	38.00
Co	32.60	35.00	--------------	39.00	36.40	35.00	43.00
Th	5.32	5.30	5.73 ± .15	7.80	6.10	5.30	3.80
Ba	520.00	580.00	800.00 ± 100	900.00	783.00	686.00	800.00
La	23.80	24.30	21.48 ± .27	26.50	28.70	23.50	18.50
Sm	6.22	6.40	6.00 ± .021	6.60	7.70	6.19	5.40
Eu	1.65	1.79	2.30 ± .11	2.00	2.18	1.52	1.80
Ce	48.50	49.00	13.70 ± .099	50.00	58.00	48.00	40.00
Nd	------	------	22.19 ± 2.10	------	------	------	------
Tb	------	------	1.33 ± .081	------	------	------	------
Yb	------	------	3.33 ± .098	------	------	------	------
Lu	------	------	0.51 ± .022	------	------	------	------
Rb	------	------	48.66 ± 4.79	------	------	------	------
Hf	------	------	7.31 ± 0.36	------	------	------	------

Murp. - Murphy (1981)
Hill - Hill (1976)
This - This Study, sample 40
INAA analysis by Amy Hoover, Ore. St. Univ.

And. - Anderson (1978); Clackamas River Area
Swan. - Swanson and Wright (1979)

0.61% P2O5 and has been tentatively assigned to the basalt of Ginkgo. The dike in the thesis area has fewer phenocrysts than typical samples of the basalt of Ginkgo which average approximately 5% phenocrysts. The basalt of Ginkgo, however, has a highly variable phenocryst content (Beeson et al., 1985). The Frenchman Springs Member contains more iron (>13.5%) and less P2O5 (< 0.65%) than Tillamook Volcanics and, therefore, can usually be chemically differentiated from them.

The Pomona Member can be differentiated from other Columbia River Basalt Group units in western Oregon and western Washington by a very high CaO (>10%) and MgO (7-8%) contents as well as by a very low (0.8%) CaO content (table 6, fig. 98). The average major element composition of the Pomona Member in the thesis area is slightly different from the the average content of the unit elsewhere (table 6). The Pomona Member in the thesis area is slightly lower in SiO2 and K2O and slightly higher in MgO and Al2O3 than the Pomona member in other areas. These slight differences are due to the more mafic composition of the gabbroic subunit (Tpg) in the thesis area.

Tables 5 and 6 show that there is essentially no significant difference in major element composition between subaerial flows ("plateau basalts") and intrusives ("coastal basalts") when the same units and subunits are compared. Peterson (1984) reported a number of chemical analyses from Columbia River Basalt Group flows in Western Oregon. The samples were collected by Dr. Marvin Beeson of Portland State University and analyzed by Dr. Peter Hooper's lab at Washington State University. The subaerial flows sampled by Beeson can be traced through the Columbia River Gorge and onto the Columbia

Plateau. It is felt that the intrusions sampled in the thesis area and the flows sampled by Beeson offer the best comparison as they were both analyzed at the same lab on the same Columbia River Basalt standard. Table 5 shows that the average composition of the intrusions in the thesis area is essentially identical to the average composition of Columbia River Basalt Group subaerial flows sampled by Beeson. This suggests that the intrusions and the flows are petrographically related.

Age

Volcanic rocks of the Columbia River Basalt Group were erupted from approximately 18 Ma. to 6 Ma. (early to late Miocene) (Swanson et al., 1979). Flows reaching Clatsop County belong to the: Grande Ronde Basalt, dated from 16.5-14.5 Ma. on the Columbia Plateau; The Frenchman Springs Member of the Wanapum Basalt, dated at 13.4 Ma. on the Columbia Plateau; and the Pomona Member of the Saddle Mountains Basalt, dated at 12 Ma. on the Columbia Plateau (Swanson et al., 1979; Kienle et al., 1978).

Snavely et al. (1973) considered the "coastal basalts" to have been erupted locally during the same time interval as the chemically similar units of the middle to upper Miocene Columbia River Basalt Group. Beeson et al. (1979), however, suggested that the coastal intrusions represent the distal ends of plateau-derived flows that, upon reaching the marine environment, invaded the semi-consolidated forming "invasive" dikes and sills. If this hypothesis is correct, as data from this report suggests, then the intrusions are precisely the same age as the subaerial flows that sourced the intrusions.

All of the chemical types in the thesis area are Oligocene to early Miocene in age or younger as they intrude rocks of this age (plate 1). Peterson (1984), Neel (1976), and Cressy (1974) have shown that the Grande Ronde Basalt and the Frenchman Springs Member petrologic types intrude the lower to middle middle Miocene Astoria Formation. Snavely et al. (1973) reportrd a K/Ar age range of 14+2.7 Ma. to 16+0.6 Ma. for the Depoe Bay Basalt (Grande Ronde Petrologic type of this report) on the northern Oregon Coast. A very thick low MgO Grande Ronde basalt sill at Neahkanie Mountain, which forms a prominent headland 10 km west of the thesis area, has been radiometrically dated at 15.9+0.3 Ma. (Niem and Cressy, 1973). Snavely et al. (1973) reported a K/Ar age of 9.0+1.4 Ma. for the basalt at Pack Sack Lookout and stated that, "although a difference of about 4 m.y. is shown by the K/Ar ages of Pomona and Pack Sack Lookout basalts, the dates are from single samples and, therefore, no great reliance can be placed on them".

The previous discussion has shown that the Columbia River Basalt Group on the Columbia Plateau and the "coastal basalts" of Snavely et al. (1973) are approximately the same age. Paleomagnetic data indicate that that the chemical types on the plateau are precisely the same age as correlative chemical types on the Washington and Oregon coasts (Choinere and Swanson, 1979; Murphy, 1981; Peterson, 1984; Nelson, 1985; this study).

Correlation

At least ten subaerial Grande Ronde Basalt flows have been traced from the Columbia Plateau down the Columbia River Gorge and into western Oregon (Anderson, 1978; Beeson and Moran, 1979; Vogt, 1981). These include three R2, low MgO flows; five N2 low MgO flows; and 2 N2, high MgO flows. At least six of these flows reached northeastern Clatsop County including: at least 2 R2, low MgO flows; two N2, low Mgo flows; and two high MgO flows (Murphy, 1981; Goalen, in prep.). Two chemical subtypes (basalt of Sand Hollow and basalt of Ginkgo) of Frenchman Springs member subaerial flows have also been traced into northwest Oregon (Beeson, et al., 1985). The Pomona Member has been mapped as a single flow unit in northeastern Clatsop County and in southwest Washington (Wells, 1981; Murphy, 1981).

Peterson (1984) correlated the four Grande Ronde Basalt intrusive subunits in the Green Mountain-Youngs River area and two Frenchman Springs Member subunits to petrologically and paleomagnetically equivalent (i.e. same polarities) plateau derived Grande Ronde Basalt and Frenchman Springs Member subaerial and submarine flows in the Nicolai Mountain area (Murphy, 1981) and the Plympton Creek area (Goalen, in prep.) of northeastern Clatsop County. Data from the thesis area support the correlations of Peterson (1984). The Columbia River Basalt Group petrologic type intrusions in the thesis area have been correlated to the subaerial flows in northeastern Clatsop County as shown in figure 100. The units correlated in figure 100 have virtually identical chemical, petrographic, and magnetic polarity properties. The gabbroic subunit of the Pomona Member (Tpg) is the only exception to this, as it does have slight chemical and petrographic differences from the Pomona

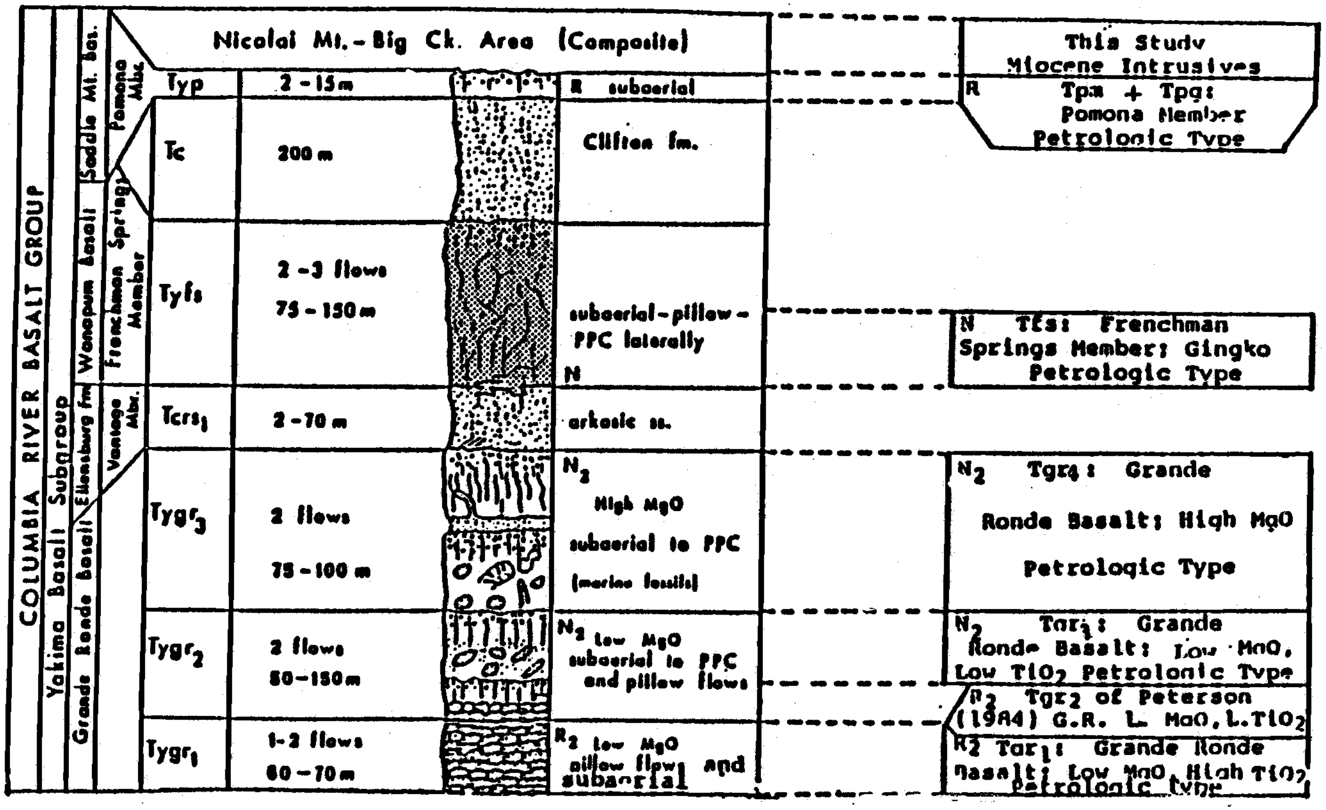


Fig. 100. Correlation of intrusive Columbia River Basalt units in the thesis area to subaerial Columbia River Basalt flows in northeastern Clatsop County (modified from Murphy, 1981).

Member studied by Murphy (1981). These differences are thought to be a result of differentiation during slow cooling. Rarey et al. (1983) correlated Pomona member intrusive rocks in the thesis area to a subaerial Pomona member flow mapped by Murphy (1981) in northeastern Clatsop County.

Origin

Snavely et al. (1973) noted that the Depoe Bay Basalt and the Grande Ronde Basalt, the Cape Foulweather Basalt and the Frenchman Springs Member, and the basalt at Pack Sack Lookout and the Pomona Member were consanguineous basalt groups. However, they considered the Columbia River Basalt Group ("plateau basalts") and the "coastal basalts" to have been erupted from separate vent areas approximately 500 kilometers apart. Therefore, Snavely et al. (1973) had to invoke regional magmatic processes to explain the identical age and chemistry of the two basalt groups. They based their conclusion that the "coastal basalts" had been erupted locally on the presence of dikes, sills, and eruptive features (e. g., "volcanic bombs") on the Oregon Coast. The three regional magmatic models suggested by Snavely et al. (1973) are: 1) partial melting of the Juan De Fuca Plate subducted beneath the American Plate; 2) partial melting along a horizontal shear at the base of the American Plate; and 3) partial melting of homogenous eclogite in the mantle of the subducted Juan De Fuca Plate. These models were sketched by Murphy (1981). Snavely et al. (1973) noted that there were two major problems with all of the above models: 1) they fail to account for the eruption of middle

Miocene calc-alkaline volcanics of the of the western Cascade arc between the two vent areas; and 2) they fail to explain how the magmas could maintain their compositional uniformities during upward ascent from the mantle to vent areas 500 kilometers apart.

The consanguineous nature of the "coastal basalts" and the "plateau basalts" led Beeson et al. (1979) to examine alternative models to those of Snavely et al. (1973). The hypothesis of Beeson et al. (1979) is that the "coastal basalts" are extensions of lava flows erupted on the Columbia Plateau. The presence of dikes and sills is thought to be a consequence of the interaction of thick flows of dense basaltic magma with soft, less dense, water-saturated marine sediments. Beeson et al. (1979) cited examples from Schmincke (1964) and Byerly and Swanson (1978) of Columbia River Basalt Group flows "invading" lacustrine and fluvial sediments in central Washington. Schmincke (1964) states that Columbia River Basalt Group flows,

> "may invade downward into the sediments at various levels in a smooth sill-like fashion with remarkably little deformation of the sediments, or it may occur in irregular forms, in dikes, lobes, or tongues of solid lava, as autobreccia, or peperite. Sediment layers may be gently lifted to the top of the basalt or may be bulldozed aside and intermixed with the sediment".

Beeson et al. (1979) suggested that the Columbia River Basalt Group flows interacted with marine sediments in a similar manner, but on a larger scale. The presence of dikes in thick sequences of submarine basalt (e.g., as mapped at Saddle Mountain by Penoyer, 1977) could be explained by local injection in the liquid basalt from lava interiors into shrinkage joints of cooled margins (autoinvasive intrusions).

Beeson et al. (1979) cited the following as evidence supportive of the "invasive hypothesis": 1) the inadequacy of the petrogenetic models of Snavely et al. (1973); 2) the fact that the "coastal basalts" are present only at the terminus of the "plateau flows" (fig. 101); 3) the association of coastal intrusions with deltaïc and shallow marine soft strata; 4) the absence of consistent dike orientation, indicating an absence of structural control; and 5) geophysical studies indicating that the coastal dikes do not extend to great depths.

Recent studies of the "coastal basalts" in Clatsop County by Murphy (1981), Rarey et al. (1983), Peterson (1984), Nelson (1985), and this study have provided additional data that is relavent to the origin of the "coastal basalts". Instead of the three coastal petrologic types defined by Snavely et al. (1973) seven different petrologic types have been defined. These intrusive petrologic types have been correlated to petrologically identical plateau derived subaerial and submarine flows in northeastern Clatsop County and appear to have been emplaced in the same sequence as the subaerial flows (fig. 100). The subaerial flows in northeastern Clatsop County can be physically traced through the Willamette Valley and onto the Columbia Plateau (Murphy, 1981; Beeson et al., 1979).

Murphy (1981) mapped in detail one of the few areas where the "plateau basalts" and the "coastal basalts" occur adjacent to one another. Late Miocene uplift of the Oregon and Washington coast ranges has resulted in erosion of this contact zone in most areas. Figure 102 is a cross section from Murphy (1981) through an area where the two basalt groups overlap. Although Murphy (1981) considered the the "coastal basalts" to have been locally erupted the

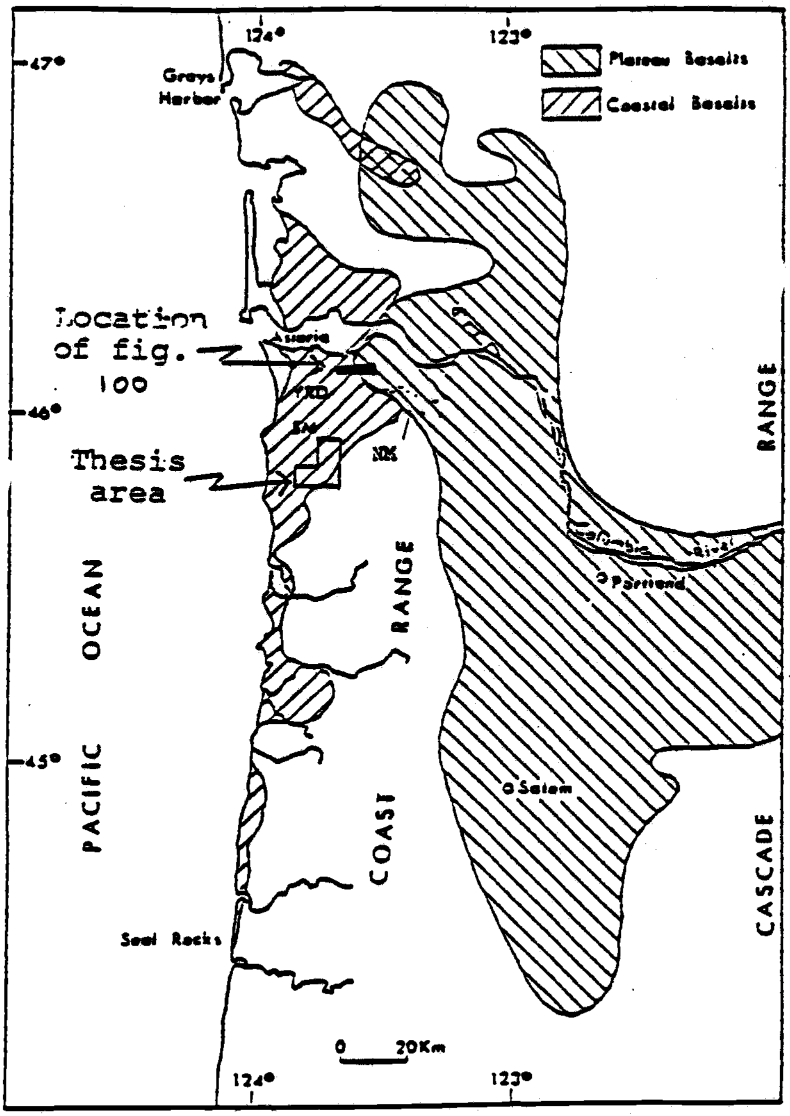


Fig. 101. Distribution of Columbia River "plateau basalts" compared to the distribution of "coastal basalts". In this report both the coastal and the plateau basalts are considered to be part of the Columbia River Basalt Group (figure modified from Beeson <u>et al</u>., 1979).

geology of the overlap area is highly supportive of an invasive origin for the "coastal basalts". In this area subaerial flows of Grande Ronde Basalt and Frenchman Springs Member grade westerly into into pillow lava deltas and eventually into dikes and sills (fig. 102). The subaerial Grande Ronde Basalt flow in figure 102 is chemically identical to and has the same polarity as the underlying Depoe Bay Basalt (Tdb). The two units were differentiated by the presence or absence of dikes. The same situation occurs between the Frenchman Springs Member (Tyfs) and the Cape Foulweather Basalt (Tcf). A local source for the Depoe Bay Basalt and the Cape Foulweather Basalt would require an extraordinary set of coincidences. Firstly, just enough lava would have to be erupted on the Columbia Plateau to cause the subaerial flows to travel hundreds of kilometers and yet stop right at the middle Miocene coastline. Secondly, a chemically identical basalt would have to be erupted at precisely the same time as the subaerial flow in the same spot where the subaerial flow terminated. Such a scenario seems unrealistic. Therefore, an invasive origin for the intrusions in Clatsop County appears more likely.

Extensive parts of the Eocene Tillamook Volcanics have been mapped in detail and no Miocene dikes have been observed intruding them (this study; Wells et al., 1983; Mumford, in prep.; Safley, in prep.). Since the Tillamook Volcanics are on trend with other Miocene dikes one would expect to find Miocene dikes withing the Tillamook Volcanics if the Miocene dikes originated from an underlying local coastal magma body.

Recent petrogenetic studies of the Columbia River Basalt

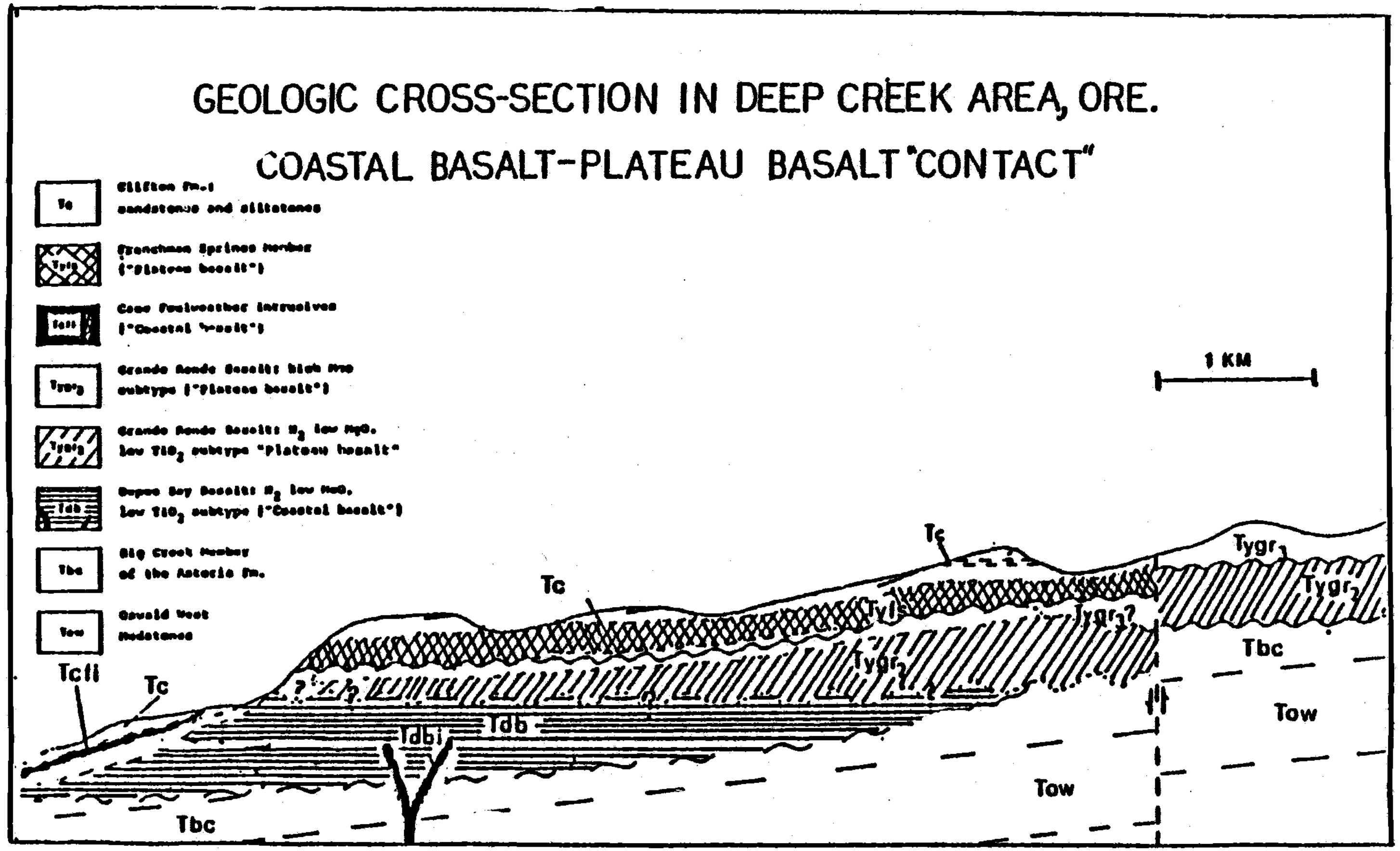


Fig. 102. Cross section through the Deep Creek area, northeastern Clatsop County, showing the relationships between the "coastal" basalts and the "plateau" basalts (modified from Murphy, 1981).

Group on the Columbia Plateau indicate that the lavas were originally derived from a mantle similar in isotopic composition to island arc and oceanic island basalts of the North Pacific (Carlson et al., 1981). This primary magma was then modified chemically and isotopically by crystal fractionation and and assimilation of sialic crustal rocks during transport through, or storage in, the continental crust. Carlson et al. (1981) concluded that the isotopic data are inconsistent with a primordial mantle source suggested by Snavely et al. (1973). Since interaction with continental crust with the magma is apparently required to produce the isotopic ratios in the Columbia River Basalt Group, it seems highly unlikely that precisely the same fractionation and assimilation events took place 500 kilometers apart at different vent areas. It seems especially unlikely because the "coastal basalts" are underlain by accreted oceanic crust.

In conclusion, it can be stated that the vast majority of recent data support the invasive hypothesis of Beeson et al. (1979). There are, however, several recent geologic observations that contradict statements made by Beeson et al. (1979). Studies of subsurface data (from exploration wells) in Clatsop County show that "coastal basalt" intrusions occur at depths exceeding 5,000 feet, some intrude "indurated" upper Eocene sedimentary rocks, and the intrusions may show reversed stratigraphic order (i.e., some Grande Ronde sills are at shallower depths than younger Frenchman Springs sills) (Olbinski, 1983; Peterson, 1984; this study; Martin et al., in press) (see fig. 66 and plate III). Although nearly every intrusion mapped at the surface intrude Oligocene to Miocene sedimentary rocks, a few dikes and sills intrude Eocene (upper Narizian) sedimentary rocks. In the thesis area, Pomona Member petrologic type sills intrude the Eocene Sweet Home Creek member at three localities including one (sample 617) that appears to intrude beneath the upper Eocene Cole Mountain basalt (plate I). Beeson et al. (1979) suggested that the Miocene sills were all emplaced at shallow depths and restricted to the Oligocene through Miocene sedimentary units in northwest Oregon.

Beeson et al. (1979) stated that there was no apparent structural alignment of the Miocene dikes. Recent studies in Clatsop County however, do show that three major dikes and many minor dikes have a preferential northeastward orientation indicating structural control of emplacememt (Olbinski, 1983; Peterson, 1984; Nelson, 1985; this study). Data from seismic sections through the northern portion of the thesis area and data from the CZ 11-28 well adjacent to the thesis area, however, indicate that the structural control is shallow rooted as the dikes are apparently sourced from underlying sills which intrude Refugian-aged strata.

Discussion

Data from this study and from other recent studies strongly support the invasive origin for the Miocene basalts of coastal Oregon and coastal Washington. However, the mechanics of invasion are poorly understood. As with other hypotheses that invoke large-scale mechanics (e.g., plate tectonics) the invasive hypothesis of Beeson et al. (1979) has been slow to gain acceptance in the geologic community. It is the contention of this report that the invasive hypothesis of Beeson et al. (1979) is the most logical and that the models of Snavely et al. (1973) are unrealistic.

Snavely et al. (1973) noted that more than 30 major petrologic units were erupted on the Columbia Plateauu and suggested that because of basin configurations and volumes of erupted lava, only six of these major petrologic units flowed into western Oregon and western Washington. Coincidentally, four of these major petrologic types were erupted at the same time, in the same sequence, at the terminus of the plateau-derived flows. This coincidence was explained by the presence of an extensive magma chamber feeding both areas. This does not, however, explain the absence of crustal contamination between the two vent areas or the reason why the only petrologic types erupted on the coast happen to correspond to the few subaerial flows that reached western Oregon and southwestern Washington. The extraordinary number of coincidences required by the local eruption hypothesis of Snavely et al. (1973) makes it untenable.

The general scenario of Beeson et al. (1979) is, however, a logical consequence of extraordinarily voluminous outpourings of lava adjacent to a continental margin. When huge flows, hundreds of times more voluminous than any historical flows, follow drainage systems into the ocean they might be expected to form thick piles of pillow basalt and breccia which may prograde far out onto the continental shel in a manner similar to fluvial deltas. Schmincke (1964) and Byerly and Swanson (1978) have shown that when fluid Columbia River Basalt flows came in contact with water and soft sediments, pillow basalts, hyaloclastites, and invasive dikes and sills formed locally. These same features should be expected where subaerial flows encountered the marine environment. It should also be expected that subaerial flows would have the same chemistry and magnetostratigraphy as the invasive dikes and sills.

"Unexpected" geologic features of the coastal basalts include deep intrusions into older sedimentary rocks, a structural alignment of some dikes, and local eruption features such as ring dikes. Deep Sea Drilling Project studies have shown that marine sediments can be poorly consolidated at great depths (e.g. 1000 m) below the sea floor (e.g., leg 63 D.S.D.P., Niem, pers. comm., 1985). This coupled with the high density contrast between molten basaltic lava (2.7 g/cm3) and water-saturated sediment (1.5-2 g/cm3) suggests that it may be possible for density driven invasion to occur to great depths.

The northeast structural alignment of shallow dikes and the presence of eruptive features could be explained by a mechanism analogous to injection of clastic dikes into overlying muddy sediments. In the thesis area clastic dikes in the Keasey Formation have a preferred northwesterly orientation. Other workers have noted that overpressuring of sand or mud can result in dike emplacement and the "eruption" of mud or sand volcanoes (Reineck and Singh, 1980; Walton and O'Sullivan, 1950). If an invasive sill is overlain by a rapidly forming pile of pillow basalts overpressuring would occur that may result in "magma" being injected along prexisting structural weaknesses, forming dikes. Some of these might reach the ocean floor forming small, local eruptive centers like Haystack Rock in western Clatsop County (see Neel, 1976)

Future Studies

Although nearly all the data support the invasive hypothesis of Beeson _et al_. (1979) there at least two studies that should be done before the origin of the "coastal basalts" can be "positively" determined. These are: 1) to further map the distribution of the petrologic subunits and to evaluate, in detail, the relationships between pillowed flows, breccias, and intrusions; and 2) to construct theoretical models for the mechanism of invasion. Dr. Alan R. Niem of Oregon State University is presently engaged in a detailed study of the middle Miocene basaltic rocks in Clatsop County and their relationship to the coeval Columbia River Basalt Group.

QUATERNARY DEPOSITS

Quaternary deposits in the thesis area have been divided into three major units: 1) alluvial terrace gravel (Qt1 and Qt2); 2) undifferentiated recent alluvial and estuarine deposits (Qae); and 3) rock fall/landslide deposits (Qls) (Plate I). With the exception of landslide deposits significant thick Quaternary deposits are limited to the southwest corner of the thesis area in the North Fork of the Nehalem River Valley (fig. 1, plate 1). Deposits in this area consist of older alluvial terrace deposits and younger undifferentiated alluvial and estuarine deposits. Terrace deposits are 1-4 m thick and are composed primarily of poorly sorted rounded Columbia River Basalt Group cobbles and boulders in a sandy matrix (fig. 103). Basalt cobbles appear to be locally derived from adjacent eroded dikes and sills. Upper Smuggler Cove formation mudstone cobbles, which are locally derived, are rarely present in the terrace deposits. The terraces are erosional (as opposed to depositional terraces) and occur at two levels: one (Qt1) approximately 7 m above present river level; and the other (Qt2) approximately 15 m above present river level. Recent alluvial deposits consist of poorly sorted, subangular to rounded basalt cobbles, minor sand, and minor silt. Estuarine deposits are very poorly exposed and consist of clay, peat, silt, and minor sand.

Mass movement deposits are ubiquitous throughout the thesis area. Most deposits are related to soil creep and slumping of sedimentary rock units. Tilting of trees and slump scars evidence these mass movement processes. Earthflow and slumping have resulted

Fig. 103. Quaternary alluvial terrace deposit (Qtg2) overlying upper Smuggler Cove formation (locality 504).

in formation of several "sag ponds" (*e.g.* Soapstone lake) in the west-central part of the thesis area (plate 1). Logging roads cut into the sides of mudstone hills commonly display tension cracks and are locally destroyed by landsliding. Soil creep may destroy logging roads underlain by volcanic rocks if road cuts are not deep. Because of the ubiquitous nature of mass movement deposits, in the thesis area only the larger "landslide" deposits have been mapped on plate 1.

Alluvial terrace deposits in the thesis are probably correlative to Pleistocene? marine terrace deposits mapped near Cannon Beach by Neel (1976). Quaternary uplift of the Coast Range resulted in a relative lowering of base level and probably caused renewed downcutting and entrenchment of the North Fork of the Nehalem River. This river was able to erode the well-indurated strata in section 32 (T4N, R9W) more quickly than Grassy Lake Creek, a small tributary. This resulted in knickpoint retreat of Grassy Lake Creek from its confluence of the North Fork of the Nehalem River. The Grassy Lake Creek knickpoint has retreated approximately 60 meters and, because of the well-indurated upper Smuggler Cove formation strata in this area, forms a waterfall (fig. 80). The presence of similar alluvial terraces adjacent to the shoreline and the knickpoint retreat of Grassy lake Creek indicate that the thesis area has been uplifted at least 10 meters, and probably much more, during the Quaternary.

STRUCTURAL GEOLOGY

Regional Structure

Paleomagnetic studies in the western Cordillera show that large-scale rotation and translation of crustal blocks has occurred along the Pacific Margin since at least Jurassic time (Beck, 1980). Simpson and Cox (1977) demonstrated as much as 75 of clockwise rotation for the Oregon Coast Range (relative to the stable North American craton) since the middle Eocene whereas Beck and Burr (1979) and Wells and Coe (1985) determined significantly less rotation for the Washington Coast Range during the same time interval (approximately 22). Several models have been proposed to account for the paleomagnetic rotations. These models include: 1) rigid microplate rotation of the lower to middle Eocene Oregon Coast Range oceanic plate during accretion to the North American continent (Simpson and Cox, 1977); 2) rigid plate rotation caused by asymmetric back arc spreading (Simpson and Cox, 1977; Magill et al., 1981); 3) small block rotation in a broad zone of right-lateral shear generated by the relative northward motion of the Farallon plate (Beck, 1976, 1980); and 4) a combination of the above mechanisms (Magill et al., 1981, 1982).

Post late middle Eocene rotations of the Oregon and Washington Coast Ranges and Cascades are very similar whereas pre-late middle Eocene rotations of the Oregon and Washington Coast Ranges are quite different (Wells and Coe, 1985). This suggests that the Oregon and Washington Coast Ranges were seperate oceanic fragments prior to accretion during the middle Eocene. The similar post-Eocene rotation of the two regions has been interpreted as evidence for rigid microplate rotation of the two areas after accretion (Magill et al., 1981). Wells and Coe (1985), however, have suggested that post-Eocene rotations are, at least in part, a result of shear coupling along transcurrent fault zones. In this model simple shear is produced by partial coupling of the forearc region with the obliquely subducting Farallon plate. The shear is taken up by members of a conjugate secondary shear set (Riedel shears) which rotate in response to the external shear couple. Asymmetric back arc spreading (i.e., Basin and Range extension) may have, however, contributed to the observed post-Eocene rotations (Wells and Coe, 1985).

Wells and Coe (1985) suggested that the post-Eocene structure of southwest Washington consists of major northwest-trending right-lateral faults and northwest to southeast-trending R' (Riedel shears) left-lateral faults. Such a fault pattern is consistent with the shear coupling model of Freund (1974) according to Wells and Coe (1985) (fig. 104). The pre-late Eocene structure of southwest Washington is fairly complex consisting of broad folds and faults of various orientations (Wells and Coe, 1985; Wells, 1981).

In Clatsop County, northwest Oregon, two major fault patterns have been mapped: 1) an older sequence of east-west trending high angle faults; and 2) a younger sequence of northeast- and northwest-trending conjugate shears with oblique left-lateral and oblique right-lateral motion respectively (e.g., Olbinski, 1983; Peterson, 1984; Nelson, 1985; this study, Niem and Niem, in press).

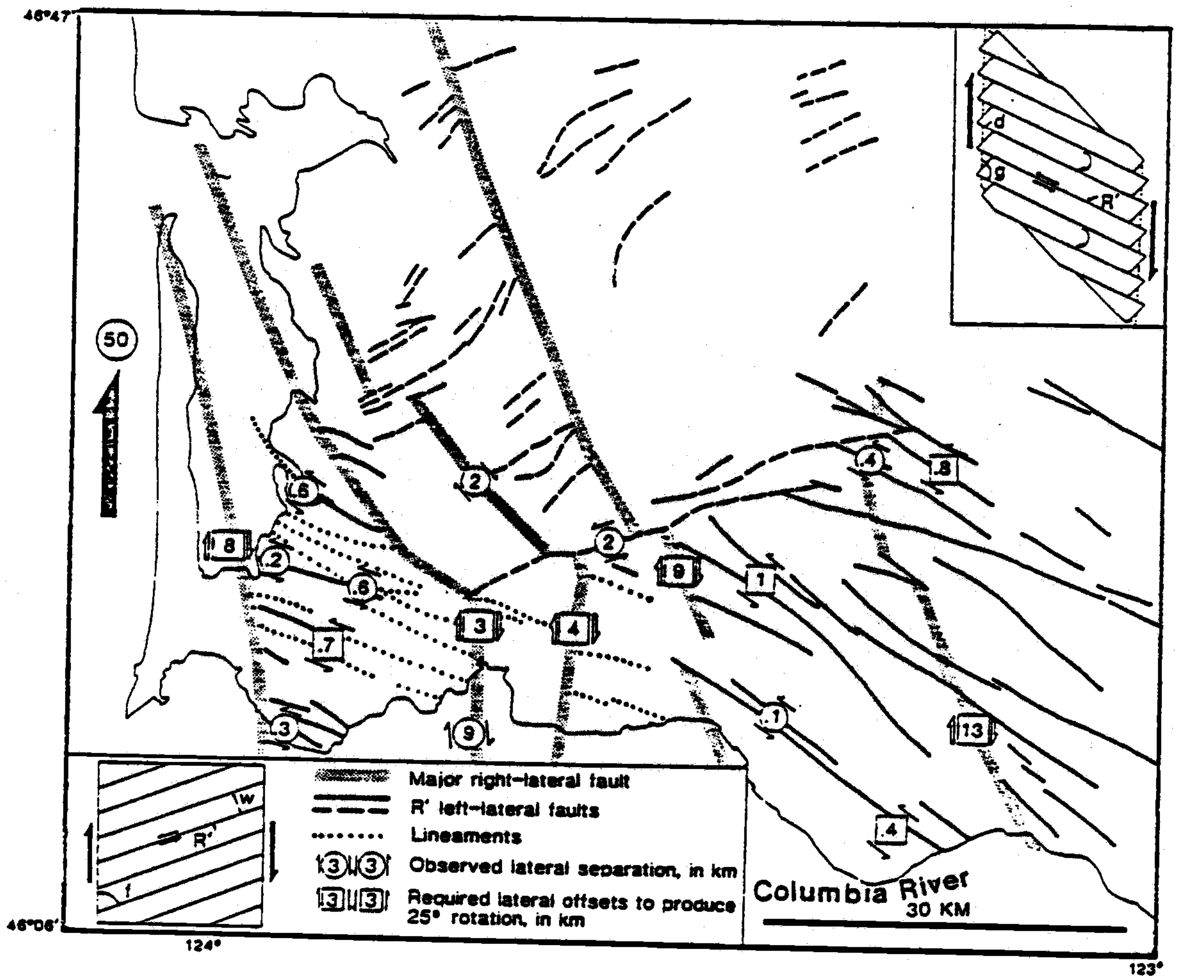


Fig. 104. Comparison of shear deformation model of Fruend (1974)(corner boxes) to strike-slip interpretation of post late Eocene faulting in southwest Washington (from Wells and Coe, 1985).

A few north-south faults are locally associated with the conjugate system (Olbinski, 1983; Peterson, 1984; Nelson, 1985). These fault systems are post middle Eocene, and for the most part, post-late Eocene in age (Peterson, 1984; Nelson, 1985).

Nelson (1985) in a paleomagnetic study using a spinner magnetometer demonstrated that faulting of middle Miocene intrusions in Clatsop County resulted in significant small block (<1 km2) clockwise rotation (approx. 11 or greater). This indicates that at least a part of the post-accretion tectonic rotation in northwest Oregon is due to a distributed shear mechanism similar to that proposed by Wells and Coe (1985) for southwest Washington.

The structure in northwest Oregon can be interpreted to reflect a shear coupling model. In this model the north-south and the northwest-trending (approx. N 30 W) right-lateral faults would be the primary expression of simple dextral shear and the northeast-trending left-lateral faults and the east-west trending faults would be sinistral R' secondary shears that have accomodated the clockwise rotations. This model would require the east-west trending faults to have left-lateral motion or would require this faults to be older and unrelated to the shear coupling. Nelson (1985) and Peterson (1984) have suggested a model in which the east-west trending faults are older and the northwest- and the northeast-trending faults represent a conjugate system caused by north-south compression. This model fits the structural pattern but does not adequately explain the observed clockwise rotation of Miocene dikes.

To the south of the thesis area, in Tillamook County, a series of northwest-trending faults (approx. N 45 W) has been mapped by

Wells *et al.* (1983). These faults may be R' left-lateral faults (Wells and Coe, 1985). It is also possible that some of these northwest-trending faults with more northerly orientations (approx. N 25 W) are right lateral faults acting as the primary expression of simple dextral shear. The presence of several major north-south-trending faults offshore of northwest Oregon, as suggested by Snavely *et al.* (1973), could in part account for the apparent rarity of these faults in Clatsop County. It should also be noted that the fault pattern in the shear coupling models of Wells and Coe (1985) and Freund (1974) consists primarily of northeast and northwest oriented faults with north-south faults being relatively rare (fig. 104).

Local Structure

The thesis area is located in the forearc on the northwest flank of the Coast Range uplift which is cored by Tillamook Volcanics and Siletz River Volcanics (Wells and Peck, 1961). Strata in the thesis area have a regional dip of 3-15 to the northwest off of the west flank of this uplift (Plate II). High-angle northeast- and northwest-trending faults and large east-west-trending faults locally disrupt the regional dip (plate 1). The fault pattern in the thesis area is similar to that noted to the north and east of the thesis area by Olbinski (1983), Peterson (1984), Nelson (1985), and Niem and Niem (in press).

Extreme northwest Oregon has been divided into two basins. The Astoria basin is located west of the Coast Range structural high

(Nehalem arch); the Nehalem (or Willamette) basin is located east of the high (Wells and Peck, 1961; Bruer et al., 1984; Peterson, 1984; Armentrout and Suek, 1985; Neim and Niem, in press). The Coast Range structural high has been defined as the present-day axis of uplift as determined by gravity data. The thesis area is located to the west of the axis of uplift in the Astoria basin. The degree to which the Nehalem structural arch has been present through time, thereby dividing the region into two distinct basins through time, is debatable.

The upper Narizian strata in northwest Oregon (Cowlitz Formation and Hamlet formation) show some facies changes that can be correlated with the structural arch. For example, Cowlitz Formation shelf sandstones pinch out into bathyal mudstones near but not precisely at the structural arch axis (Olbinski, 1983; Nelson, 1985; Niem and Niem, in press) (Plate IV). Fauna from the Cowlitz Formation (Hamlet formation correlative) east of the arch, however, show that open marine conditions were present in the region (Yett, 1974; Jackson, 1983). Therefore, if the structural arch was present during the late Narizian it was not large enough to cause restriction of normal oceanic currents. Depositional evidence from the early Refugian Sager Creek formation (Vesper Church formation of Olbinski, 1983) indicates that there was not a structural arch affecting sea floor topography during this part of the Refugian. The Sager Creek formation submarine canyon head deposits cuts across the region of the present day structural arch and is thickest over the present day structural arch (Kadri, 1982; Olbinski, 1983; Rarey, 1984). Therefore, in this writer's opinion, the present-day axis of the

Coast Range structural arch differs in position, at least slightly, from older structural axes in the region. In addition, it is unlikely that a structural high has existed continuously in northwest Oregon since the Narizian.

A number of well-documented faults occur in the thesis area but because of thick vegetation and limited exposures it is not possible to determine all of the structural features present. Structural features are more easily studied and recognized in the southern part of the the thesis area because of better exposures and the presence of thin sedimentary units (e.g., Roy Creek member) which can be used to demonstrate stratigraphic separation.

Strikes and dips taken from mudstone-siltstone units in road cuts have questionable regional structural significance. In addition to the lack of distinct bedding planes in the thick mudstone units, numerous intrusions, penecontemporaneous deformation, and the ubiquitous modern slumps contribute to a variable pattern of strike and dip. The most reliable attitudes in the thesis area are from stream bed exposures. Columbia River Basalt dikes have been used as piercing points in mudstone units by previous workers in Clatsop County (e.g., Olbinski, 1983; Nelson, 1985). Dikes in the thesis area, on a local scale (i.e., at individual exposures), are irregular and frequently change orientation and vertical height. Some dikes are fairly long and maintain a relatively constant orientation but others are short and highly irregular (Plate I). Therefore, the dikes should be used with caution as piercing points.

East-West Faults

Four major east-west trending faults have been mapped in the thesis area. These four faults have the largest separation and the widest shear zones of all the faults mapped in the thesis area. The east-west faults appear to be offset by a set of northwest- and northeast-trending faults. These faults occur in the central and southern parts of the thesis area and include Helloffa fault, Gods Valley fault, Sweet Home fault, and Cole Mountain fault (fig. 105, plate I). All of these faults are high-angle normal separation faults that show from approximately 60 to 200 m of dip separation. Strike separation is difficult to estimate due to a lack of piercement structures. Based on stratigraphic continuity strike separation is definitely less than 10 km and probably less than several hundred meters. The east-west faults juxtapose Tillamook Volcanics against Keasey Formation, Hamlet formation, and Cole Mountain basalt in the eastern part of the area and may involve strata of the Smuggler Cove formation in the western part of the area. The faults in the thesis area have been correlated to faults mapped to the east and to the west by Mumford (in prep.) and by Cressey (1974). The mapped lengths of these faults range from 1 to over 10 km.

Helloffa fault parallels Helloffa Creek and Rector Ridge in the extreme southern part of the thesis area. The fault plane is not exposed but, based on existing exposures, can be limited to a very small area in the Helloffa Creek valley. In section 11 (T3N, R9W) the Roy Creek member shows approximately 200 m of stratigraphic

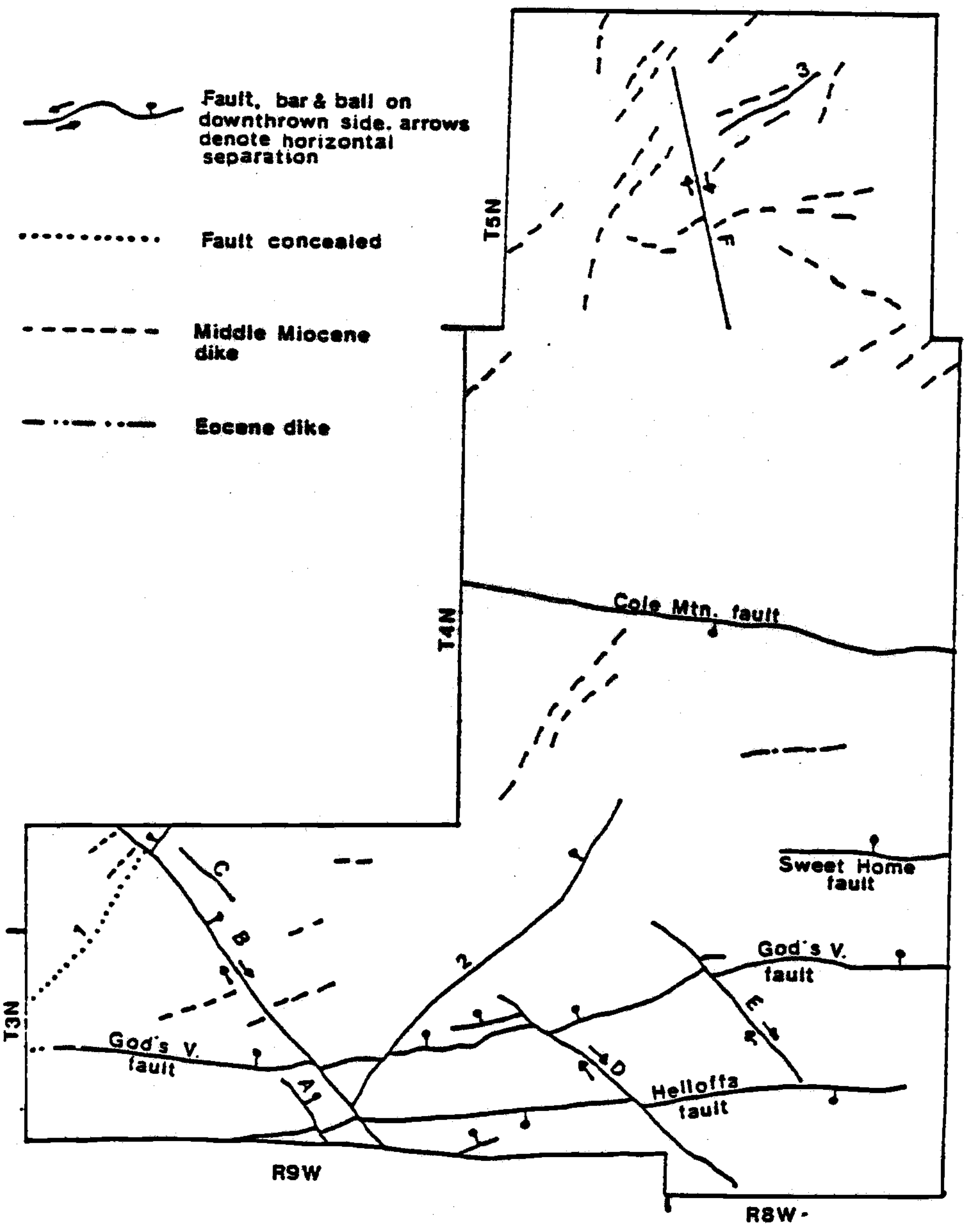


Fig. 105. Sketch map of the thesis area showing the location of major faults and dikes.

separation. Displacement of strata shows that the fault is probably a high-angle normal fault. Helloffa fault can be traced westward into a sequence of major east-west trending faults mapped by Cressy (1974).

Gods Valley fault parallels Helloffa fault 2 km to the north. The fault plane is well exposed at locality 287 north of Rector Ridge (fig. 106). At this location strata of the lower Jewell member are faulted against older upthrown Tillamook Volcanics. The brecciated fault zone is approximately 10 m wide and consists of intermixed tuffaceous gouge and angular basaltic clasts. The fault plane at this locality dips 79 to the north. Slickensides are common and indicate that last motion on the fault was oblique slip (slickensides plunge 15 to the west). "Steps" on the slickensides suggest that last motion was left-lateral. The downfaulted Jewell member mudstones dip as much as 80 to the north adjacent to the fault plane. Differential erosion of mudstones and volcanic rocks has resulted in exhumation of the fault plane along Rector Ridge (plate I). Several smaller subsidiary east-west faults closely parallel Gods Valley fault (fig. 105).

Gods Valley fault can be traced eastward into the thesis area of Mumford (in prep.) where well exposed fault breccias (approx. 5 feet thick) occur in the Tillamook Volcanics. This correlative fault zone also has subhorizontal slickensides. Data from this study and from Niem and Niem (in press) indicate that the Gods Valley fault can be traced westward into to a major east-west trending oblique slip fault mapped by Cressy (1974). On the north flank of Rock Mountain (south of Angora Peak), the fault zone displaces middle Miocene basalts and

Fig. 106. Exposure of Gods Valley fault (locality 827, NW 1/4 NW 1/4 sec. 32, T4N, R8W). Brecciated Tillamook Volcanics are faulted against lower Jewell member mudstones. Fault plane is nearly vertical and contains subhorizontal slickenslides.

middle Miocene Astoria Formation. The fault appears to have left-lateral separation. Northwest-trending folds mapped in Miocene Astoria Formation adjacent to the fault between Rock Mountain and Angora Peak may be en echelon folds produced by left-lateral motion in the late Miocene (Niem, pers. comm., 1985).

The Sweet Home Creek fault shows less separation (approximately 60 m of dip separation) than the other east-west faults mapped in the thesis area. The fault plane is exposed in Sweet Home Creek (locality 626) where Sweet Home Creek mudstones are downfaulted against Tillamook Volcanics. The fault zone at this locality is approximately 2 m wide.

The Cole Mountain fault, located in the central part of the thesis area, shows approximately 200 m of vertical stratigraphic separation. The best evidence for this fault is the apparent displacement of the Cole Mountain basalt-Keasey Formation contact in the west and the presence of an upthrown block of Tillamook Volcanics to the east (Plate 1). Fault plane breccia (2 m wide) is exposed in the North Fork of the Nehalem River (locality 568) where Tillamook Volcanics and Roy Creek member are faulted against Sweet Home Creek member. The Cole Mountain fault can be traced eastward into the thesis areas of Mumford (in prep.) and Safley (in prep.) (named Quartz Creek fault in this area) where the sense of vertical displacement is opposite. This indicates that the fault has significant strike-slip displacement. The fault shows up as a distinct lineament on SLAR and U-2 high altitude imagery. Niem and Niem (in press) show the fault ending at the unconformity between the Smuggler Cove formation and the Angora Peak member of the Astoria

Formation, the west of the thesis area. This suggests that the fault was not active after the Oligocene.

Northwest- and Northeast-Trending Faults

Five major northwest-trending faults (faults A-F) and three major northeast-trending faults (faults 1-3) have been mapped in the thesis area (fig. 105). These faults show both strike separation and dip separation. Dip separations range from several meters (fault F) to over 80 m (fault 1). Strike separations, where faults appear to offset middle Miocene dikes (faults B and F), are approximately 200 m. Slickensides associated with the northwest-trending faults (e.g. locality 398) are horizontal to subhorizontal. Because of this, the faults are thought to be primarily strike-slip although there was more than likely a component of dip-slip. Northeast-trending faults appear to be primarily left-lateral, and northwest-trending faults appear to be primarily right-lateral. This is based on apparent offset of middle Miocene dikes and other faults. Locally "steps" on the slickensides also suggest the above motions. Brecciated fault zones associated with the these faults, where exposed, are generally less than 2 m wide. The northwest- and northeast-trending faults in the thesis area appear to have less total motion than the east-west faults based on thiner fault zones, shorter lengths, and less dip separation.

The fault plane of fault A is well exposed at a quarry in the southern part of the thesis area (locality 398) (fig. 107). The fault plane is nearly vertical and contains well-developed

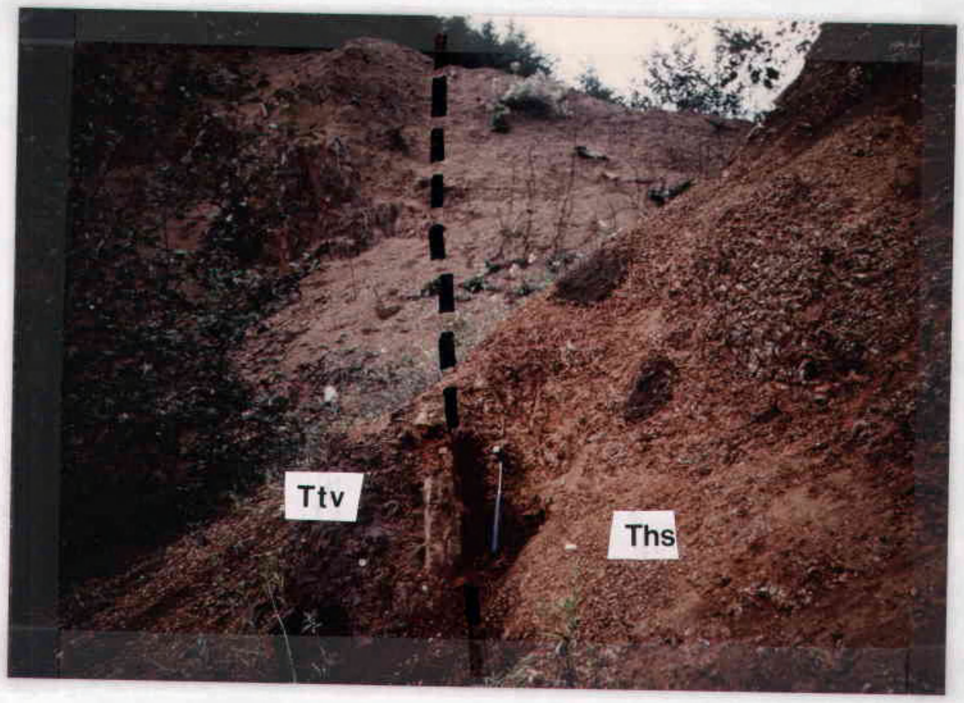


Fig. 107. Exposure of high-angle northwest-trending fault (locality 398, SW 1/4 SE 1/4 sec. 9, T3N, R9W). Tillamook Volcanics and overlying Roy Creek member (left) are faulted against Sweet Home member (right).

subhorizontal (plunge 15 northwest) slickensides. "Steps" on these slickensides are inconclusive as to sense of motion. Tillamook Volcanics and Roy Creek member are upfaulted against Sweet Home Creek mudstone at this locality. Approximately 30 m of dip separation occurs. Fault B is on trend with the linear valley of Grassy Lake Creek to the northwest of the thesis area. Niem and Niem (in press) have mapped the Grassy Lake Creek lineament as a major fault based on juxtaposition of Miocene Angora Peak member sandstones against Oligocene upper Smuggler Cove formation mudstones.

The northwest- and northeast-trending faults mapped in the thesis area are, in general, well documented. Exceptions to this include fault 2, which was mapped on the basis of a linear valley and apparent offset of mudstone units, and fault F, which was mapped on the basis of a linear valley and the apparent right-lateral offset of middle Miocene dikes. Fault 1 was mapped on the basis of offset middle Miocene sills and because of apparent juxtaposition of ball park unit sandstones against upper Smuggler Cove formation mudstones (Plate 1). Much of the fault plane is, however, covered by Quaternary alluvial and estuarine deposits.

Basalt Dikes

Middle Miocene Columbia River basalt dikes in thesis area show a preferential northeast orientation (fig. 104). These dikes were probably emplaced along pre-existing structures. Olbinski (1983), Peterson, (1984), and Nelson (1985) noted a similar relationship between Miocene dikes and northeast-trending faults to the north and east of the thesis area. Stereonet s-pole plots of fault and intrusion orientations measured or estimated in the thesis area indicate some parallelism (fig. 108). The s-pole plot of faults and small shears indicates that the majority strike northwest with some striking northeast and dip steeply and that the majority of intrusions strike northeast and dip steeply.

Wilcox et al. (1973) noted that en echelon tension "fractures" commonly develop in areas undergoing horizontal shear stress in "wrench tectonic" settings. If the shear is a north-south right-lateral shear, as is thought to occur along the Pacific Margin (Beck, 1980; Wells and Coe, 1985), the tension fractures can be theoretically expected to have a northeast orientation. Heyl et al. (1966) mapped a sequence of short mafic dikes and parallel faults, similar to those in the thesis area, emplaced in tension fracture zones. Therefore, it is possible that the Miocene dikes in the thesis area were emplaced along en echelon "tension fracture zones" that developed in response to a north-south right-lateral shear.

A late Eocene Cole Mountain basalt dike, located in the east-central part of the thesis area, has an east-west orientation (fig. 105). The dike parallels the major east-west faults in the southern portion of the thesis area. Cole Mountain dikes also tend to subparallel major east-west faults to the east of the thesis area (Safley, in prep., this study). This suggests that the late Eocene dikes were emplaced along pre-existing late Eocene east-west faults.

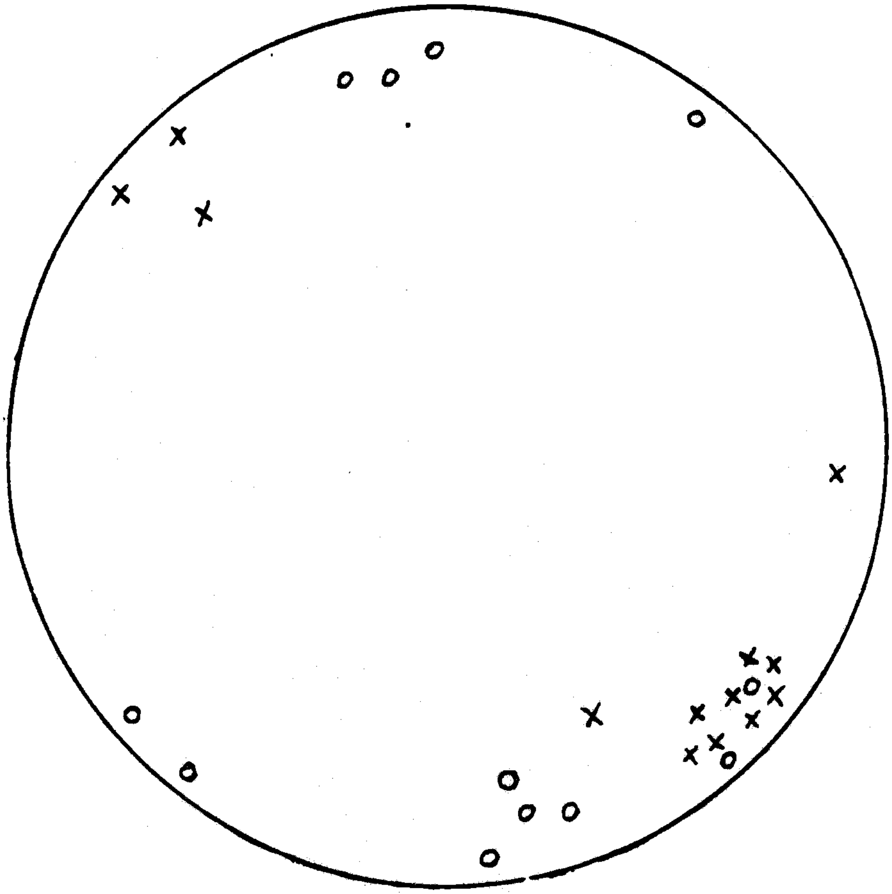


Fig. 108: Stereonet plot of poles to planes representing dikes and faults in the thesis area.

Lineations

Low altitude air photos, U-2 high altitude air photos, Landsat imagery, and side looking airborne radar (SLAR) imagery were examined for lineations in the thesis area. Faint lineations occur along the Little North Fork of the Nehalem River, and along the North Fork of the Nehalem River in the southwestern portion of the thesis area. Faults F and 1, respectively, roughly correspond to these lineations.

Age of Faulting

The east-west-trending faults in the thesis area are, apparently, older than the northeast- and northwest-trending faults. This is evidenced by the apparent offsetting of the east-west faults by the northwest- and northeast-trending faults (fig. 102, Plate 1). Late Eocene Cole Mountain basalt dikes parallel east-west faults suggesting that this faulting style preceeded or was active during emplacement of the dikes. Some east-west faults can be traced west into Cressy's (1974) and Smith's (1975) thesis areas where these faults truncate middle Miocene dikes (see Niem and Niem, in press). Cole Mountain fault, however, appears to die at the Oligocene-Miocene unconformity to the west. Therefore, activity on the faults occurred during the late Eocene and during the Miocene. The long period of faulting indicates that reactivation of some faults occurred. The east-west faults do not truncate Quaternary alluvial terraces in the North Fork of the Nehalem River Valley (Plate I).

Northwest- and northeast-trending conjugate faults were active during the middle to late Miocene as evidenced by emplacement of middle Miocene dikes parallel to the faults and by offset of these dikes by the conjugate faults. Quaternary alluvial terraces are not cut by these faults indicating that activity on the faults had ceased by the end of the Pliocene or possibly by the end of the Miocene. Previous workers in Clatsop County have suggested that the northwest and northeast faults were active during the middle to late Miocene and are somewhat younger than the east-west faults (Olbinski, 1983; Peterson, 1984; Nelson, 1985). Pavlis and Bruhn (1983) have suggested that major uplift of the Coast Range forearc ridge occurred during the middle to late Miocene. Therefore, activity on the conjugate fault system was apparently contemporaneous with uplift of the Oregon Coast Range.

Discussion

Wells and Coe (1985) noted that, "the fault pattern in southwest Washington is complex and subject to interpretation because of poor exposures". The same can be said for northwest Oregon.

The structure in northwest Oregon can be interpreted in a similar way as the post-Eocene structure of southwest Washington (see Wells and Coe, 1985). The areas have similar faulting patterns and are located in adjacent forearc tectonic settings.. The major differences are that in northwest Oregon there are several major east-west faults and an apparent absence of major northwest-trending R' left-lateral faults. In addition, north-south to northwest-trending right-lateral faults (primary transcurrent faults of Wells and Coe, 1985) are relatively rare and tend to show less displacement in northwest Oregon than in southwest Washington. The east-west faults can be explained by the shear coupling model if they have left-lateral slip. The large amount of dip separation on these faults could be explained by an initial episode of late Eocene to Oligocene dip-slip faulting and a later Miocene reactivation as R' shear. It is possible that some of the northwest-trending faults in northwest Oregon (including those in the thesis area) mapped as right-lateral, because of poor exposure, are actually left-lateral.

Nelson (1985) suggested that some of the structure in northwest Oregon could be related to wrench tectonics. The east-west faults mapped in the thesis area have some features similar to wrench faults in so far as they apparently have strike slip displacement and have "associated faults". The east-west faults in the thesis area do, however, differ from classical wrench fault systems described by Wilcox et al. (1973) in a number of significant ways: 1) by definition wrench faults have 10's of kilometers of strike slip displacement whereas the east-west faults in the thesis area have a maximum strike slip displacement of less than 10 kilometers and a probable displacement of several hundred to a thousand meters; 2) although the east-west faults mapped in the thesis area can be traced a fair distance (1-15km, see Niem and Niem, in press) they are shorter than wrench faults described by Wilcox et al. (1973); 3) the presence of late Eocene to Miocene east-west oriented wrench faults cannot be easily explained by the regional tectonic setting of northwest Oregon; 4) there are no associated structural wrench basins (e.g. pull apart basins) in the region as is evidenced by a lack of abnormally thick stratigraphic sections in the vicinity of the east-west faults; 5) although the east-west faults are apparently the major faults in the thesis area, adjacent to the thesis area northwest-trending faults with similar or greater displacement have been mapped (*e.g.* Gales Creek fault of Jackson, 1983 and Green Mountain fault of Timmons, 1981); and 6) classical well developed *en echelon* structures are not associated with the east-west faults in the thesis area. The northwest- and northeast-trending faults in the thesis area differ from *en echelon* faults associated with throughgoing wrench faults described by Wilcox *et al.* (1973). The *en echelon* faults form at relatively low angles (*e.g.* 25) to the wrench fault and stop at the wrench fault whereas the northwest and northeast faults in the thesis area form at relatively high angles (*e.g.* 50) to the east-west faults and truncate the faults. This truncation is significant in that by definition wrench faults are major throughgoing structures. Therefore, it is this authors opinion that it would be inproper to call the east-west faults in the thesis area wrench faults and to interpret associated structures in terms of classical wrench tectonics.

In conclusion, the structural features observed in northwest Oregon and in the thesis area could have been produced by a shear couple similar to that proposed by Wells and Coe (1985) for southwest Washington. An early episode of movement on the east-west faults is, however, probably unrelated to shear coupling. Shear coupling could also have produced much of the post-Eocene tectonic rotation observed in northwest Oregon. However, it is possible that other mechanisms and different stress regimes contributed to or caused the structural features observed in the thesis area. The major problem with the shear coupling model is that it does not properly address the tectonics associated with uplift of the Oregon and Washington coast ranges forearc ridge, which is the major post Eocene structural event in the region. For example, if uplift of the forearc ridge is related to deep seated flow, as suggested by Pauvlis and Bruhn (1983), then the regional stress regime would be significantly modified from that interpreted by Wells and Coe (1985). Another problem with the shear coupling model is that it does not adequately explain all movement on the major east-west faults in the thesis area. It is clear that a more detailed regional structural analysis of northwest is needed.

GEOPHYSICS

Introduction

Gravity data, aeromagnetic data, ground-based proton precession magnetometer traverses (appendix 14), electric and sonic logs from a gas exploration well (Plate IV), and seismic lines across the northern part of the thesis area were used as an aid in surface and subsurface mapping.

Gravity Data

A Bouguer gravity anomaly map drawn by Finn et al.(1984) shows that the thesis area is located on the northwest flank of the Coast Range gravity high (fig. 109). The presence of a gravity high in the southeastern corner of the thesis area and a gentle downward gradient to the northwest is consistant with the geology mapped in the thesis area. That is, gravity values are highest where the Tillamook Volcanics crop out in the southeast corner of the study area. Values decrease gradually northwestward as the thickness of sedimentary rocks that overlie the Tillamook Volcanics increases. The gravity gradient is primarily controlled by the dense, northwesterly dipping Tillamook Volcanics and underlying basement rocks. A distinctive peninsula-shaped high in the northern part of the thesis area may reflect an uplifted basement block of Tillamook Volcanics or an abnormally thick succession of middle Miocene intrusive basalts. It should, however, be kept in mind that the peninsula shaped anomaly was contoured on the basis of one gravity measurement and, therefore,

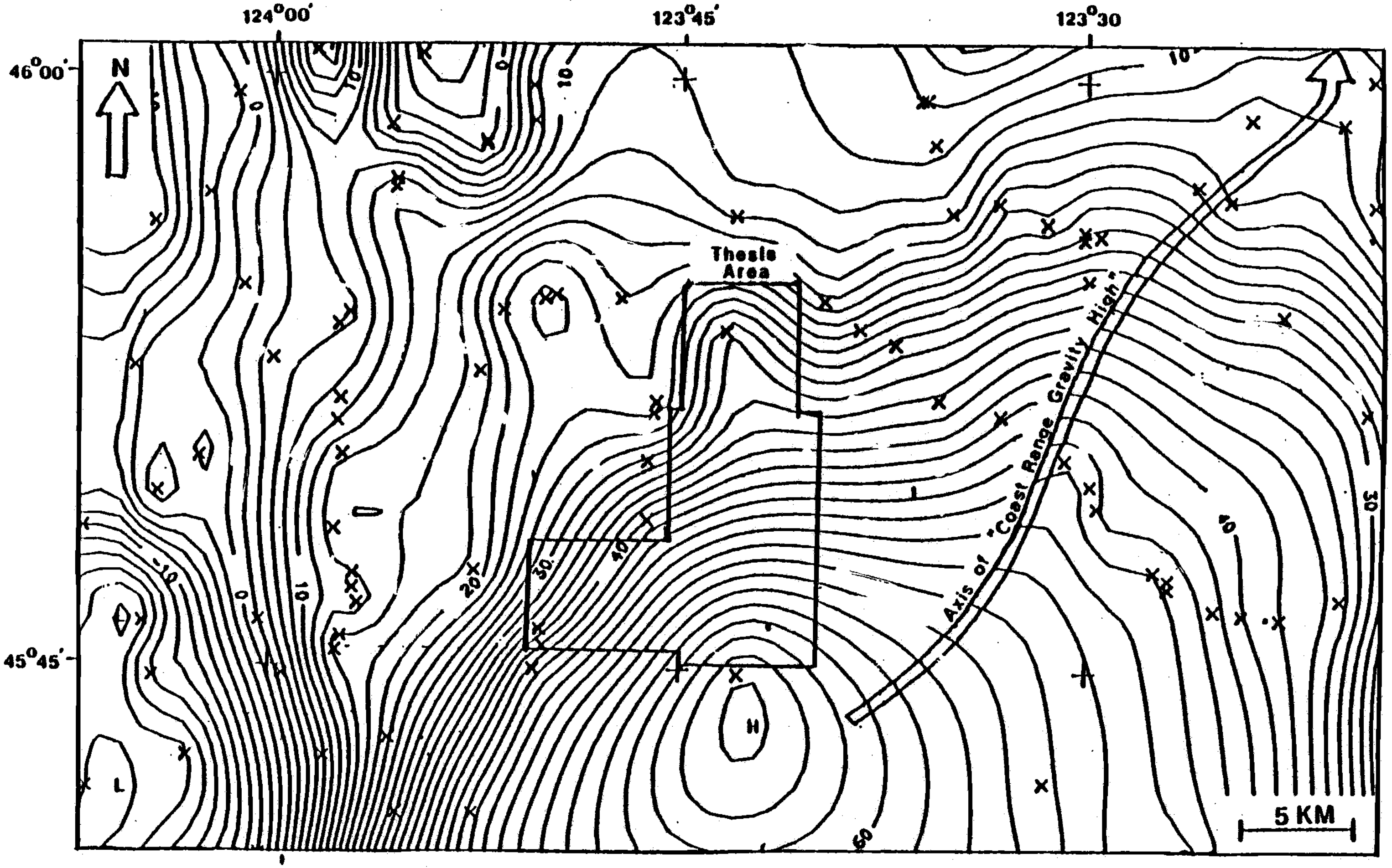


Fig. 109. Bouguer gravity anomaly map showing the location of the thesis area and the localtion of the Coast Range gravity high. Contour interval is 2 milligals. Gravity stations are marked (X)(modified from Finn _et al._, 1984).

the contours could be drawn in many different ways. Snavely (pers. comm. to Niem, 1983) has suggested that the relatively steep gravity gradient to the southwest of the thesis area may be reflecting a large northeast-trending basement fault. It is, however, this authors opinion that this steep downward gradient is primarily a result of more steeply dipping volcanic "basement" rocks (e.g. Tillamook Volcanics) and a thinning of the Tillamook Volcanics to the west at the flanks of the volcanic island complex. Regional mapping shows that the Tillamook Volcanics dip steeply to the west in the vicinity of the steep gravity gradient. Niem and Niem, in press and Wells et al. (1983) did not map any major faults paralleling the gradient. In the southwest corner of the thesis area a fault was mapped along the gradient in the North Fork of the Nehalem River Valley (Plate 1). This fault does not have sufficient displacement, at least in the Miocene strata, to cause the observed gravity gradient. In summary, Finn et al. had only three gravity stations in the thesis area (fig. 109) and, therefore, gravity data are insufficient to delineate secondary structural features (i.e., faults).

Aeromagnetic Data

Aeromagnetic data from the U.S. Geological Survey (open file report 84-205) was available for the northern part of the thesis area (fig. 110). The outcropping Tillamook Volcanics form a cluster

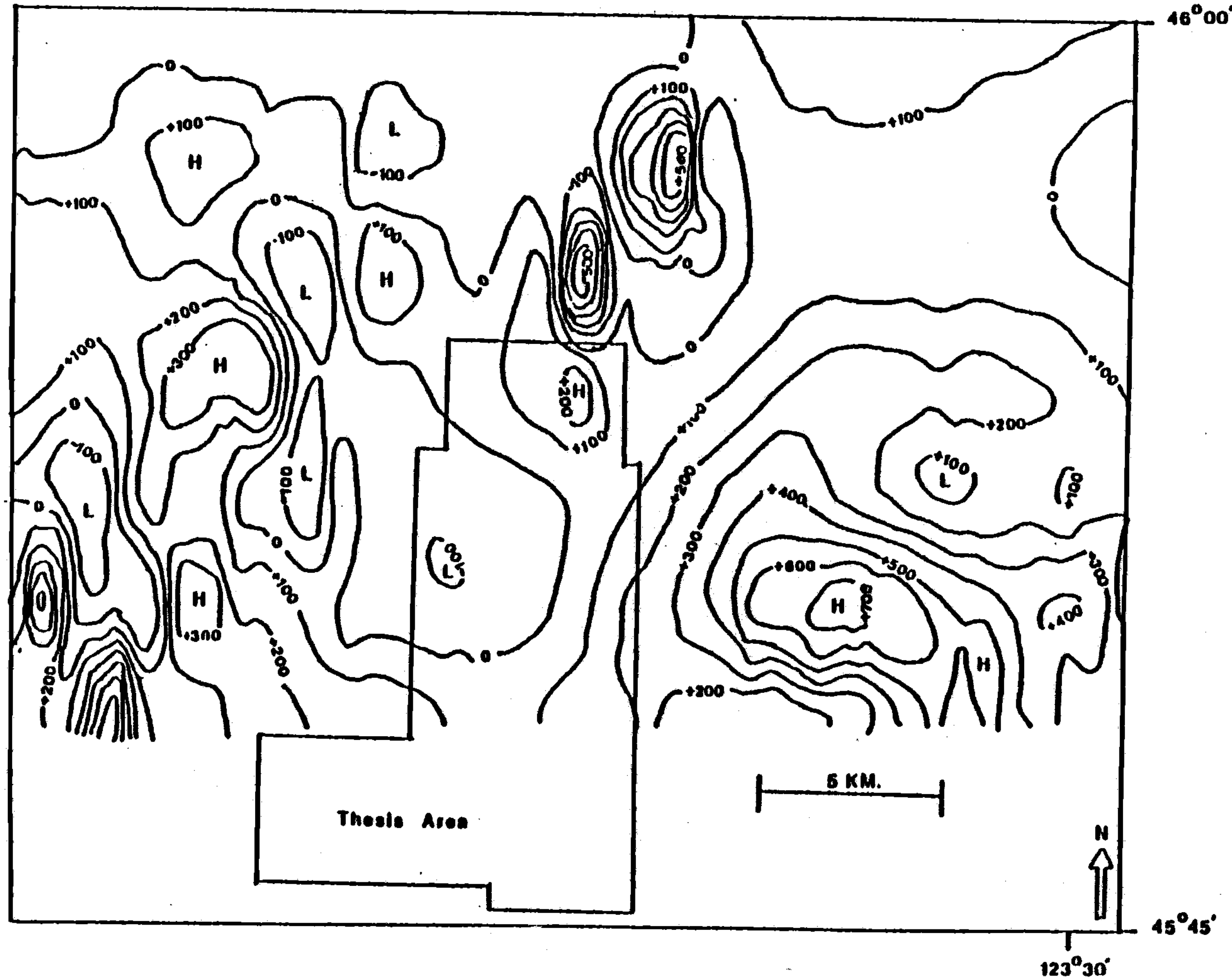


Fig. 110. Aeromagnetic map of the thesis area and surrounding areas. Contour interval is 100 gammas with a datum of 54,619.15 gammas (modified from U.S. Geological Survey open file report 84-205).

of magnetic highs to the east of the thesis area and middle Miocene basalt sequences form magnetic highs and lows to the north and west of the thesis area (fig. 110). The two magnetic anomalies north of the thesis area correspond to Saddle Mountain and Humbug Mountain which consist of more than 1500 feet of submarine middle Miocene basalt. The magnetic data indicate that Humbug Mountain is reversely polarized and that Saddle Mountain is normally polarized. Niem (pers. comm., 1985), however, has obtained Miocene basalt samples from Humbug Mountain which have normal polarities using a fluxgate magnetometer. As part of this thesis a reversely polarized low MgO-high TiO2 Grande Ronde Basalt dike has been mapped in the northeast corner of the thesis area. This dike can be mapped into Humbug Mountain. Regional studies (e.g. Peterson, 1984; Nelson, 1985) have shown that low MgO-low TiO2 dikes have a reverse polarity. The above data coupled with the large magnetic low on figure 106 indicate that either Humbug Mountain and the area beneath it consists of both normally and reversely polarized basalts or that the fluxgate data are unreliable and that it consists almost entirely of reversely polarized basalts. It is extremely unlikely that the large negative anomaly is caused by an unusually thick stratigraphic sequence.

Within the thesis area a gentle downward gradient occurs from southeast to northwest. This gradient is primarily due to the northwestward dipping Tillamook Volcanics. A 200 milligal magnetic high in the northeastern corner of the thesis area is probably due to the presence of several thick, normally polarized Grande Ronde Basalt sills and/or the presence of upfaulted Tillamook

Volcanics in the subsurface. This same region shows up as a small gravity high on the Bouguer gravity anomaly map of figure 105. Surficial mapping, reflection seismic data, and well data indicate that thick dikes and sills are present in this area. Seismic data is not, however, available for the area directly above the anomaly. The 100 milligal magnetic low in the west-central part of the thesis area is not easily explained. Directly east of this low an uplifted block of Cole Mountain basalt has been mapped. This suggests that either the Cole Mountain basalt in this area is reversely polarized or has a weak remnant magnetism. A weak normal remnant magnetism is indicated by fluxgate magnetometer data (see Cole Mountain basalt section). The presence of reversely polarized Tillamook Volcanics beneath the Cole Mountain basalt would also explain the magnetic low. Another explanation is that the Hamlet formation mudstones are unusually thick in this area. Well data (Plate III), however, suggests that the Hamlet formation is relatively thin in this area. The presence of a relatively gentle gravity gradient in the area argues against any major changes in stratigraphic thickness.

Relatively high (+200 gammas) magnetic contours project into the southwestern part of the thesis area. The presence of very thick (> 50 m), normal polarity Grande Ronde Basalt sills in this area is a likely cause of the magnetic high.

Proton Precession Magnetometer Traverses

Fourteen proton precession magnetometer traverses were made within the thesis area to help delineate faults and to project dikes and sills in the subsurface. In addition to these traverses a number of measurements were taken from areas were the underlying lithology was positively known. This was done to determine typical responses of lithologies. The spacing between magnetic field measurements on traverses ranged from 1 to 4 meters depending on the length of the traverse. On most traverses more than 100 magnetic field measurements were made.

Seven traverses were made across areas where Columbia River Basalt Group basalts were thought to intrude mudstones (traverses A-E). These traverses are characterized by a relatively flat profile over mudstone and a positive 100-200 gamma anomaly over basalt intrusions. The majority of Grande Ronde basalt intrusives in the thesis area have a normal polarity. In most cases it was possible to delineate the position of an intrusion based on magnetic profiles. Where this was not possibles lithologies have been querried in appendix 14.

Seven traverses were made across areas where Tillamook Volcanics and Cole Mountain basalt are in contact with sedimentary rocks (profiles F-L, appendix 14). The Tillamook Volcanics are characterized by rapidly fluctuating magnetic readings (100-500 gamma anomalies) whereas mudstones have a relatively flat magnetic profile. Tillamook Volcanics consist of both normally and reversely polarized flows but the fluctuation in magnetic readings is too closely spaced to be reflecting the polarity changes. Fluctuations are probably due to changes in magnetic properties from the base to the top of flows.

Traverses across projections of Gods Valley fault (profiles

G,J, adn L) show a well defined magnetic break where the fault had been mapped. Traverse L was made across Gods Valley fault in an area where Tillamook Volcanics have been faulted against Cole Mountain basalt. On this profile the break is less distinctive because of the similar magnetic signatures of the two basalt units. Two traverses were made across projections of Helloffa fault where Tillamook Volcanics have been upfaulted against sandstones and mudstones of the Hamlet formation (profiles H and I). Distinctive magnetic breaks occur where the fault had been mapped. Two profiles crossed areas underlain by Roy Creek member sandstone and conglomerate (profiles H and K). On both of these profiles, there is a strong suggestion of an "upthrown" block of Tillamook Volcanics in the Roy Creek member section. This could be caused either by faulting or by the presence of an errosional remnant of Tillamook Volcanics (i.e., "sea stack").

Well Data

Electric logs, sonic logs, and well cuttings from the CZ 11-28 well, located three kilometers to the west of the thesis area, were examined and interpreted (plate III). The well data were then used as an aid in constructing cross sections (plate II). Martin et al. (1985) used the interpreted well section of this report in a regional well correlation project of Clatsop County. Gamma ray and sp (self potential) curves show that the well encountered primarily mudstones and occasional volcanic rocks. Sidewall cores and cuttings were used to further refine interpretations. Microfossil assemblages were collected and subsequently identified by

Rau, McDougall, and Bukry (appendices 1 and 2). The microfossils showed that the well encountered upper Narizian through upper Refugian bathyal sedimentary rocks. These sedimentary rocks were assigned to the Hamlet, Keasey, and Smuggler Cove formations. Major oxide analysis was used to determine the volcanic units encountered in the well. These analyses showed that the well bottomed in Tillamook Volcanics and encountered three Grande Ronde Basalt sills upsection. The absence of any significant reservoir sandstones in this well explains why it was a dry hole.

Seismic Data

Seismic reflection lines which were run across the northern part of the thesis were made available for examination. These seismic sections were correlated to the CZ 11-28 well and interpreted. In general, the seismic data are useful in defining sedimentary-basalt sill contacts because of high acoustic impedience. Contacts between sedimentary rocks could not be delineated on the seismic profiles. Two important pieces of information were delineated from the seismic data: 1) the the strata in the northern part of the thesis area dip gently (2-6) to the northwest; and 2) a thick, relatively continuous series of Columbia River Basalt Group sills intrude the Refugian Smuggler Cove formation and the Refugian Jewell member in the northern part of the thesis area. The latter data is supportive of an "invasive sill" origin for the Columbia River Basalts and is useful in constraining models on the mechanics of invasion.

GEOLOGIC HISTORY

The following geologic history is reconstructed from the rocks in the thesis area and in surrounding areas. Oceanic island basalts of the Siletz River Volcanics and the Crescent Formation were accreted to the North American continent during the middle Eocene (Wells et al., 1984). This resulted in a westward jump of the subduction zone and a westward migration of the volcanic arc. A thick sequence of bathyal turbidites and mudstones (Tyee, Yamhill, and McIntosh Formations) was deposited over the accreted volcanic rocks in a developing forearc basin (Wells et al.,1984).

In northwest Oregon a thick sequence of late middle Eocene volcanic rocks (Tillamook Volcanics) was erupted in the developing forearc forming a large-island complex (fig. 111a). These volcanic rocks constitute the oldest unit exposed in the thesis area. Chemical data suggest that the Tillamook Volcanics constitute an extensional suite formed in a spreading-center island tectonic environment.

Cessation of Tillamook volcanism resulted in rapid thermal subsidence and sea level transgression (fig. 111b). During this time the Narizian basaltic conglomerates and basaltic sandstones of the Roy Creek member were deposited in nearshore high energy conditions in the thesis area, but as transgression continued to the east, arkosic detritus from the continent mixed with the basaltic detritus (fig. 111c). Bathyal mudstones of the Sweet Home Creek member were deposited above the sandstones (fig. 111d).

Sea level regression during the late Narizian (late Eocene)

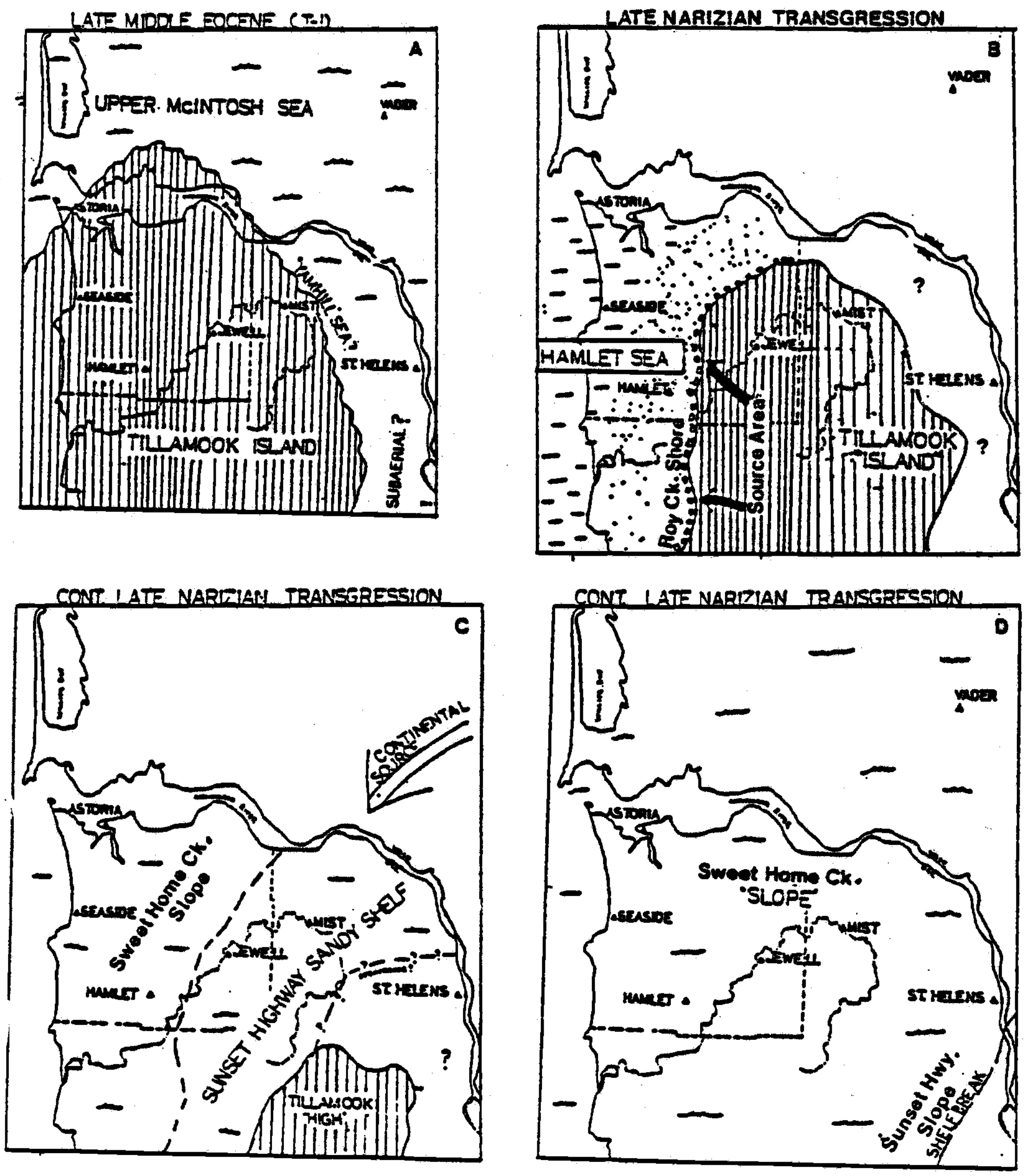


Fig. 111. Paleogeographic maps of northwest Oregon and southwest Washington for the middle Eocene to late Miocene. The thesis area is located at Hamlet on these maps.

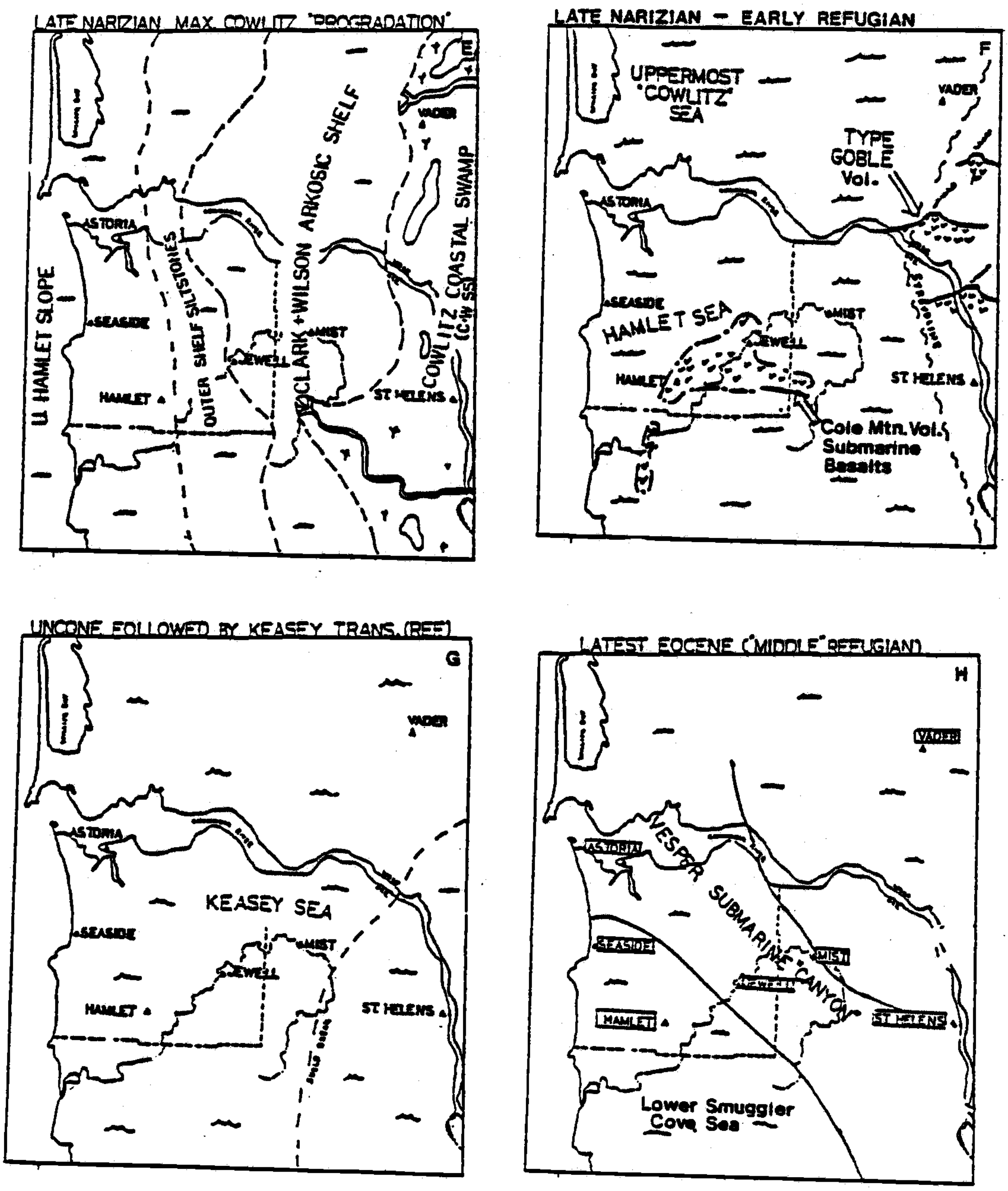


Fig. 111. Paleogeographic maps continued.

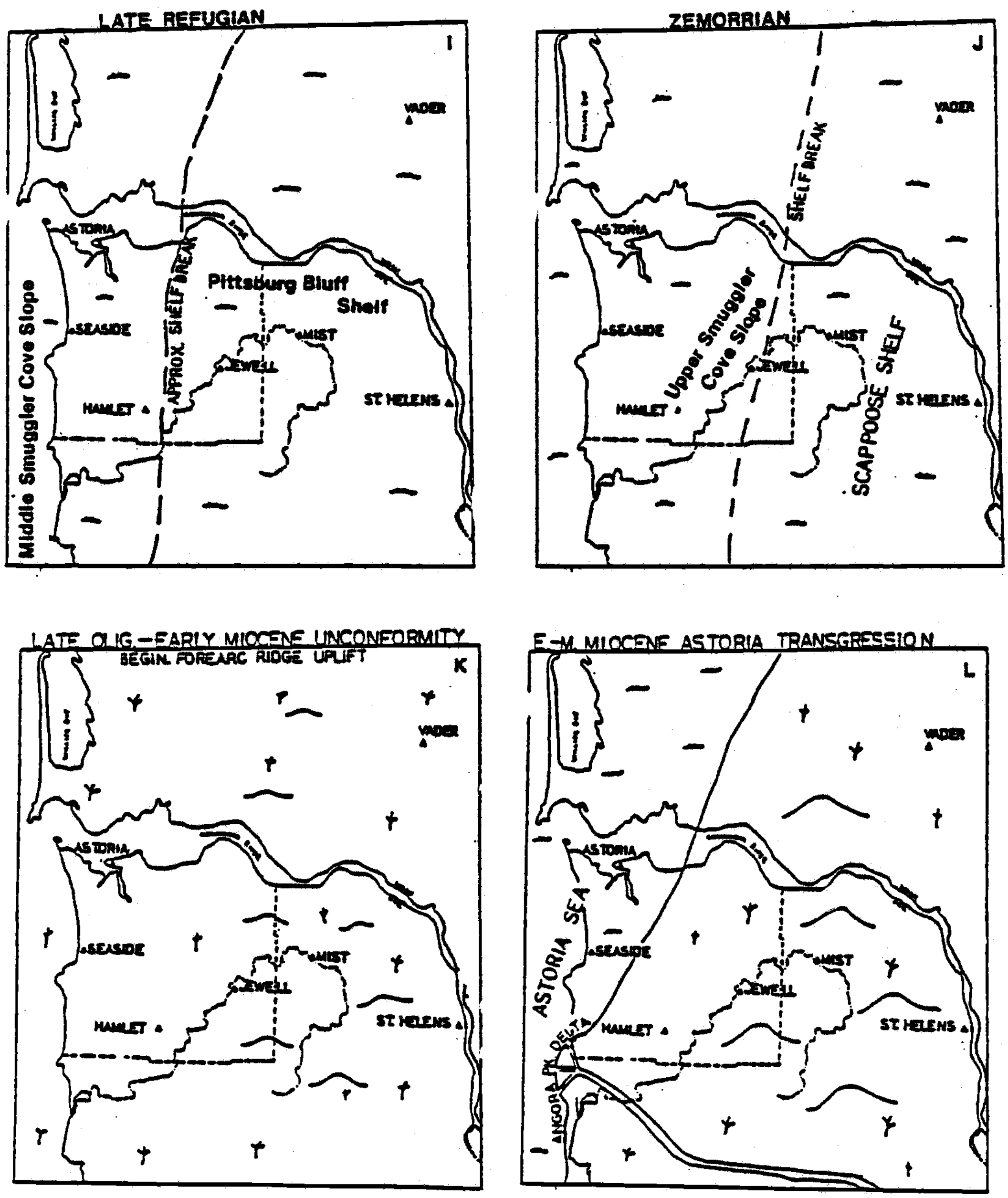


Fig. 111. Paleogeographic maps continued.

resulted in deposition of a thick sequence of nearshore to shelf arkosic sandstones (Cowlitz Formation) east of the thesis area (fig. 111e). Time-correlative outer shelf to upper slope mudstones of the upper Sweet Home Creek member were deposited in the thesis area.

Local Submarine eruption and intrusion of Cole Mountain basalts in the Hamlet-Jewell area followed deposition of the Sweet Home Creek member. Correlative calk-alkaline, subaerial and submarine basalts of the type Goble Volcanics were erupted to the east.

A thick sequence of Refugian outer shelf to bathyal tuffaceous mudstones (Keasey Formation) was deposited in the forearc basin unconformably over the Cole Mountain basalt, Hamlet formation, and Cowlitz Formation in Clatsop and Columbia counties (fig. 111g). From late Refugian (early Oligocene?) until early Saucesian (early Miocene) a sequence of shallow-marine units was deposited east of the thesis area in Columbia County (Pittsburg Bluff Formation and Scappoose Formation). The Smuggler Cove formation, David Douglas tongue of the Pittsburg Bluff Formation, and the Northrup Creek formation mapped in the thesis area are time-correlative to the shallow marine units, but were deposited in deeper water (outer shelf to middle slope) (fig. 111 h,i,j). The Cascade arc was active during deposition of these units and supplied abundant tuffaceous detritus to the forearc basin. An ancestral Columbia River, located east of the thesis area, contributed lesser amounts of continentally arkosic detritus to the basin.

Initial uplift of the Oregon and Washington Coast Ranges, possibly due to deep-seated flow (Paulis and Bruhn, 1983), is reflected by the late Oligocene-early Miocene unconformity (fig. 111k). Early to middle Miocene, shallow-marine to "deltaic", sandstones and mudstones of the Astoria Formation unconformably overlie the Oligocene to lower Miocene units in Clatsop County (fig. 111l). Much of Columbia County was subaerially exposed during at this time.

During the middle to late Miocene, flood basalts erupted on the Columbia Plateau, flowed down an ancestral Columbia River gorge and entered the marine environment near the Clatsop County-Columbia County line (fig. 111m). These flows formed thick sequences of submarine breccia in western Clatsop County (e.g., Saddle Mountain and Humbug Mountain north of the thesis area). Some of the dense basalt invaded the water-saturated, semiconsolidated sediments as well as moderately indurated rocks, forming dikes and sills.

Major uplift of the Coast Range forearc ridge began in the late Miocene resulting in extensive erosion of rock units in the thesis area. Thin deposits of alluvial gravel were deposited in the thesis area during the Quaternary.

North-south compression and shear caused by partial coupling of the obliquely subducting Juan de Fuca plate to the North American plate may have produced the numerous east-west trending, northwest-trending, and northeast-trending high-angle conjugate faults observed in the Eocene through Miocene strata of the thesis area. An earlier episode of late Eocene faulting may have produced the large east-west faults observed in the thesis area.

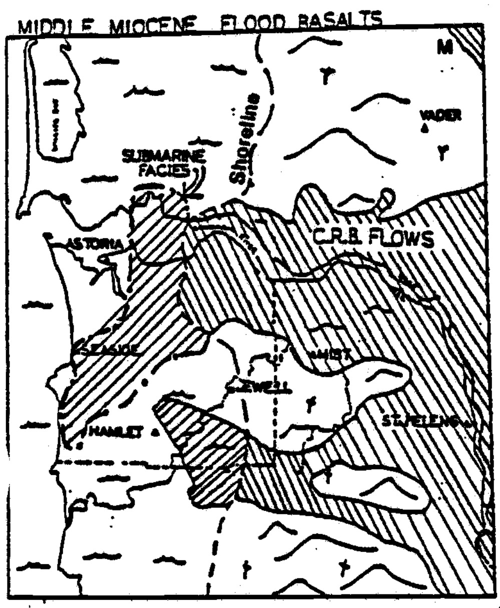


Fig. 111. Paleogeographic maps continued.

HYDROCARBON GEOLOGY

The discovery, in 1979, and develpoment of the Mist gas field, located 35 kilometers to the east of the thesis area, demonstrated the presence of commercial quantities of hydrocarbons in northwest Oregon. As of January, 1985, the Mist field consisted of 16 wells producing from 13 pools. Cumulative gas production for the field is 19,219,324 mcf (Armentrout and Suek, 1985). Reservoir sandstones in the Mist area are in the middle to late Eocene Cowlitz Formation which consist of very permeable micaceous arkosic sandstones (Armentrout and Suek, 1985).

One of the purposes of this study was to delineate the southwestward extent of the of the Cowlitz Formation reservoir sandstones. Diamond Shamrock Corp. drilled an exploration well (CZ 11-28) 4 kilometers west of the thesis area with Cowlitz Formation reservoir sandstone as the target. This dry well did not penetrate any significant reservoir sandstone and bottomed out in Tillamook Volcanics (plate III). Data from this study show that Cowlitz Formation sandstones pinch out into bathyal mudstones of the Sweet Home Creek member of the Hamlet formation before reaching the thesis area. The pinch-out is located some 10 kilometers east of the thesis area (plate IV).

Surficial mapping shows that sandstones compose a very small part of the stratigraphic section in the thesis area. The only sandstones are: basaltic sandstones of the Roy Creek member; small arkosic sandstone channels in the Jewell member; tuffaceous sandstones of the Pittsburg Bluff Formation; thin arkosic sandstone beds in the ball park unit in the Smuggler Cove formation; and arkosic sandstones of the Astoria Formation.

The Jewell member and ball park unit sandstones are very thin (<2m) and contain abundant diagenetic clay matrix. A Smuggler Cove formation (ball park unit) sandstone has a total porosity of 29.4% and an effective porosity of 16.4% (Jill Schlaefer, pers. comm., 1983). The permeability of this sample, as calculated by experimentally derived equations, is low (0.12 m.d.; appendix 13).

Pittsburg Bluff Formation sandstones in the thesis area are very fine-grained, moderately sorted, lithic rich, and contain abundant volcanic glass. Because of this they have little porosity (approx. 3%) and are probably relatively impermeable. The Pittsburg Bluff Formation in the thesis area is not an attractive reservoir sandstone.

The only thick sandstone units in the thesis area are the Roy Creek member of the Hamlet formation and the Angora Peak member of the Astoria Formation. Roy Creek member basaltic sandstones contain abundant pore-filling chloritic cement. Because of the chloritic pore-fill porosity has been reduced to several percent. Galloway (1979) suggested, that once significant chloritic pore-fill develops in a basaltic sandstone, the sandstone should no longer be considered a viable reservoir.

Arkosic sandstones of the Angora Peak member of the Astoria Formation are not present in the subsurface of the thesis area and, therefore, cannot be considered a potential reservoir in this area. However, the deltaic to shallow marine Angora Peak member is considered to be a prime reservoir target on the nearby inner continental shelf (Snavely et al. 1979; Cooper, 1981; Niem and Niem, in press).

Bruer et al. (1984) and Armentrout and Suek (1985) have described arkosic "Yamhill sandstone" (or "lower Cowlitz sandstone") beneath the Tillamook Volcanics in the subsurface of the Mist area. Data from this thesis suggest that the "Yamhill sandstone" is correlative to the Cowlitz Formation sandstone mapped by Wells (1981) beneath the Grays River area Goble Volcanics in southwest Washington. This indicates that the sandstone may have a wide distribution and may be present beneath the Tillamook Volcanics in the subsurface of the thesis area. The Tillamook Volcanics in this region are greater than 2,000 m thick, making it very expensive to test this hypothetical reservoir.

The northern part of the thesis area has been actively explored for hydrocarbons as evidenced by seismic lines and nearby drilling activity by Diamond Shamrock Corporation. Data from this study show that significant reservoir sandstones do not occur in the subsurface of the thesis area and, therefore, the area offers little potential of hydrocarbon production. Mudstones in the thesis area are very ductile and it is highly doubtful that they contain significant fracture porosity.

Source rock geochemical analysis was performed on five mudstone samples from the thesis area by AMOCO Production Company and the University of British Columbia (appendix 13). Samples were collected from fresh stream bed exposures of the Hamlet and Keasey formations and from a "coaly" bed in the ball park unit. Total organic carbon ranged 0.9-1.1% except for the "coaly" sample which probably contained more than 35% organic matter. Carbonaceous sandstones in the ball park unit may contain as much as 5% organic plant debris.

Vitrinite reflectance values ranged from 0.53-0.72% Ro. The sample with 0.72% Ro is Jewell mudstones collected near a thick Columbia River Basalt sill. The relatively high Ro value (increased thermal maturation) of this sample is probably due to baking by the intrusion. The generation of the samples ranges from non-source to marginal source, and they fall into the early gas to early peak gas stage of diagenesis. Vitrinite reflectance values of the samples fall above the gas window of Dow (1977) (fig. 112).

Armentrout and Suek (1985) have suggested that the gas in the Mist field is thermogenic in origin and migrated into the field from the north Willamette basin. Eocene mudstone samples from the thesis area contain more total organic carbon and have higher vitrinite reflectance values than Eocene mudstones sampled by Armentrout and Suek (1985) (avg. 1% compared to average of 0.65%; 0.53-0.58 compared to <0.50% Ro). An additional factor to be considered in evaluating the hydrocarbon-generating postential of the Astoria basin is the abundance of intrusions (Niem and Niem, 1985). According to Reverdatta and Melenevskii (1983), intrusion of numerous dikes and sills into thermally immature mudstones can generate "significant" amounts of thermogenic gas. The abundant Eocene and Miocene intrusions in the thesis area and in adjacent areas may have generated thermogenic gas in Clatsop County. The volume of this gas however, would tend to be relatively small. The possibility of intrusion-generated thermogenic gas in conjuction with the marginal

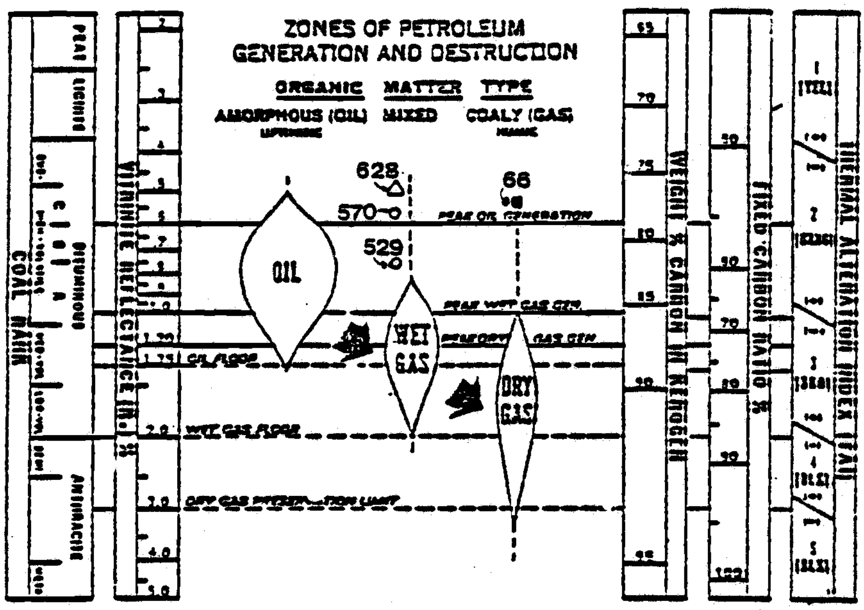


Fig. 112. Zoned of petroleum generation and destruction (Dow, 1977). Samples from the thesis area (Sweet Home Creek member , Jewell member , and Ballpark member of the Smuggler Cove formation) plot above and near the wet gas and dry gas windows.

generation capacity of non-baked mudstones in the thesis area indicate that Eocene mudstones in the thesis area are potential source rocks (or marginal source rocks) for thermogenic gas. Moderately abundant organic material throughout the stratigraphic section indicates that biogenic gas could also be generated. Therefore, it is possible that mudstones in the Astoria "basin" (including the thesis area) contributed thermogenic gas to Cowlitz Formation reservoirs to the east.

As previously mentioned, Astoria Formation sandstones on the inner continental shelf form a potential reservoir. The middle to upper Eocene mudstones in the thesis area that were deposited in an upper slope environment may project into middle to lower slope mudstones offshore. These mudstones would tend to be thicker and would have undergone deeper burial than the mudstones in the thesis area. Therefore, they could have generated significant thermogenic gas and filled the overlying Astoria Formation reservoirs. The seal for the Astoria reservoirs could be overlying unnamed upper Miocene and Pliocene mudstones as shown on a geologic cross section of the Oregon continental margin by Snavely et al. (1980).

MINERAL AND CRUSHED ROCK RESOURCES

Basaltic rocks in northwest Oregon are frequently quarried for road aggregate for use on local logging roads. The middle Miocene Columbia River Basalt Group intrusions in the thesis area are generally unweathered and hard and, therefore, make good road aggregate. Many of the intrusions are relatively thick (> 10 m), especially the thick sills, allowing for large volumes of rock to be quarried. Several road aggregate quarries in Columbia River basalt are present in the thesis area (plate I). Future quarries could be chosen on the basis of mapped extent of intrusions.

The Eocene Tillamook Volcanics are much more voluminous than the Miocene Columbia River Basalts in the thesis area. The Tillamook Volcanics, however, consist of subaerial flows which contain brecciated and vesiculated flow tops and have been moderately weathered. Therefore, the Tillamook Volcanics are of marginal quality for road aggregate. The Cole Mountain basalt is generally well weathered and soft making it of little use for road aggregate.

Mumford (in prep.) and Safley (in prep.) have found a small exploritory adit along the Quartz Creek- Cole Mountain fault zone near Quartz Creek (located some 15 km east of the thesis area). There does not, however, appear to have been any significant mineralization other than pyrite in the Quartz Creek area. Within the thesis area there are no signs of mining activity nor is there any obvious evidence of potentially economic mineral deposits.

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APPENDICES

APPENDIX 1: FOSSIL CHECKLIST
CZ 11-28 Well

SPECIES \ LOCATION	2116-01	2116-03	2116-04	2116-06	2116-10	2116-12	2116-15	2116-17	2116-21	2116-24	2116-26	2116-29	2116-34	2116-40				
Alabamina kernensis		X					X	X	X									
Ammodiscus incertus	X	X		X														
Astigerina crassiformis	X																	
Bathysiphon eocenicus					X													
Bathysiphon sp.				X	X		X	X	X			X		X				
Boldia hodgei												X						
Bolivina kleinpelli		X																
Bolivina marginata	X																	
Bolivina scabrata	X							X										
Bulimina corrugata					X													
Bulimina macilenta	X	X				X												
Bulimina scu?lptilis lacinata								X						X				
Cassidulina laevigata												X						
Caucasina schencki									X									
Ceratobulimina washburnei					X													
Chilostomella oolina							X		X			X						
Cibicides fortunatus										X								
Cibicides sp.	X						X					X		X				
Cyclammina japonica	X																	
Cyclammina pacifica	X	X	X	X				X	X	X	X			X				
Dentalina cooperensis	X																	
Dentalina sp.	X		X															
Dorothia sp.			X															
Eggerella elongata	X								X									
Eggerella subconica		X																
Fissurina marginata		X																
Glandulina laevigata						X												
Globobulimina pacifica		X		X			X		X		X	X		X				
Globocassidulina globosa	X	X	X			X			X	X		X	X					
Gyroidina condoni		X										X						
Gyroidina aff. G. rosaformis	X																	
Gyroidina soldanii			X	X		X	X											
Gyroidina soldanii octocamerata	X																	
Gyroidina sp.	X																	
Haplophragmoides deflata	X	X																
Haplophragmoides obliquacamerata			X	X							X							
Hoeglundina eocenica											X							
Karreriella washingtonensis							X											
Lagena hispida obliquicamerata				X														
Lenticulina inornata												X						
Lenticulina sp.		X	X									X						
Marginulina subbullata												X						
Martinotiella communis				X														
Melonis cf. M. soldanii	X																	
Melonis sp. of McDougall	X					X												
Nodosaria delicata						X		X	X									
Nodosaria hispida								X						X				
Nodosaria longiscata				X		X	X		X		X	X		X				
Nodosaria ovrula						X	X	X				X						

APPENDIX 1 (cont.)
CZ 11-28 Well (cont.)

SPECIES \ LOCATION	2116-01	2116-03	2116-04	2116-06	2116-10	2116-12	2116-15	2116-17	2116-21	2116-24	2116-26	2116-29	2116-34	2116-40
Nodosaria sp.		X	X			X	X		X					
Oridorsalis umbonatus				X		X		X						
Orthomorphina rohri												X		X
Planulina markleyana							X							
Planulina tolmani												X		
Plectofrondicularia packardi	X	X												
Plectofrondicularia packardi multilineata					X									
Praeglobobulimina pupoides							X							
Praeglobobulimina sp.							X							
Pseudenodosaria inflata				X										X
Pullenia bulloides							X							
Quinqueloculina imperialis	X	X	X											
Saracenaria schencki														X
Sigmoilina tenuis				X		X								
Spiroplactammina tejonensis						X								
Stilostomella adolphina								X						
Stilostomella lepidula	X	X		X		X								
Triloculina sp.		X												
Trochamina globigeriniformis	X								X					
Uvigerina cocoaensis species group	X	X	X	X	X	X		X				X		X
Uvigerina garzaensis	X	X	X	X	X	X								X
Valvulineria jacksonensis	X													

APPENDIX 1 (cont.)
Hamlet formation

SPECIES \ LOCATION	R-221	R-398	R-548	R-557	R-562	R-564	R-608	R-612	R-615	R-620	R-621	R-628	R-629	R-2116-26	R-2116-29	R-2116-34'	R-2116-40'	R-559
FORAMINIFERA (Benthic, surface samples)																		
Allamorphina trigonia						X												
Amphimorphina jenkinsi						X		X										
Anomalina californiensis								X		X								
Bathysiphon eocenica			X												X		X	
Boldia hodgei									X		X		X		X			
Bolivina huneri			X															
Bolivina marginata										X								
Bolivina pisciformis			X															
Bolivina scabrata						X						X	X					
Bulimina microcostata						X												
Caucasina schenki									X			X						
Cibicides fortunatus										X								
Cibicides natlandi									X									
Cibicidina walli										X								
Cibicides sp.												X			X			
Cyclammina pacifica												X					X	
Cyclammina japonica														X				
Dentalina communis			X															
Dentalina jacksonensis												X						
Fissurina marginata				X														
Fursenkoina bramletti						X												
Globobulimina pacifica		X	X			X			X		X	X	X	X	X		X	
Globocassidulina globosa									X	X	X		X		X	X		
Globocassidulina margareta											X							
Gyroidina condoni		X							X		X	X	X		X			
Gyroidina soldannii									X									
Gyroidina soldannii octocamerata										X								
Haplophragmoides obliquicamerata		X						X						X				
Karreriella washingtonensis						X					X							
Lenticulina budensis										X		X	X					
Lenticulina chehalisensis											X	X	X					
Lenticulina crassa									X									
Lenticulina inornata		X	X								X	X			X			
Lenticulina limbosus hockelvensis			X						X									
?Lenticulina rotulata		X																
Lenticulina sp.											X	X			X			
Lenticulina welchi			X					X			X	X	X					
Marginulina subbullata											X				X			
Nodosaria longiscata		X		X	X	X						X		X	X			X
Nodosaria pyrula						X									X			
Nodosaria sp.										X								
Nonionella jacksonensis											X		X					
Oridorsalis umbovatus						X												
Orthomorphina rohri				X											X		X	
Planulina tolmani			X										X					
Plectofrondicularia minuta						X												
Plectofrondicularia oregonensis									X									X
Plectofrondicularia packardi		X	X						X		X	X	X					
Plectofrondicularia vaughani																		X
Praeglobobulimina pupoides		X		X					X		X	X	X					X

APPENDIX 1 (cont.)
Hamlet formation (cont.)

SPECIES \ LOCATION	R-221	R-398	R-548	R-557	R-562	R-564	R-608	R-612	R-615	R-620	R-621	R-628	R-629	R-2116-26	R-2116-29	R-2116-34	R-2116-40	R-558
Pseudonodosaria conica												X						
Pseudonodosaria inflata				X													X	
Pullenia salisburyi						X						X						
Quinqueloculina imperialis			X									X						
Quinqueloculina weaveri								X										
Rectobolivina sp.												X						
Saracenaria hantkeni											X							
Spiroloculina texana								X		X								
Stilostomella aduena																		X
Stilostomella lepidula		X				X		X	X	X	X	X	X					
Stilostomella sp.																		X
Trifarina nannai									X			X						
Valvulineria jacksonensis welcominsis		X		X		X			X		X		X					
PLANKTIC FORAMINIFERA-Undifferentiated			X			X				X	X							
Globigerina spp.								X		X								
FORAMINIFERA OCCURRING ONLY IN WELL SAM.																		
Cassidulina laevigata															X			
Chilostomella oolina															X			
Hoeglundina eocenica														X				
Nodosaria hispida																	X	
Saracenaria schencki																	X	
Uvigerina cocoansis species group															X		X	
Uvigerina garzaensis																	X	
DIATOMS																		
Coscinodiscus spp.																		X
Rhizosolenia spp.																		X
Stephanopyxis spp.				X														X
Stephanopyxis turris	X			X														X
Resting spores	X																	X
CALCAREOUS NANNOFOSSILS*																		
Coccolithus formosus																X		
Chiasmolithus solitus								X		X								
Cyclicargolithus floridanus																	X	
Discoaster														X				
Reticulofenestra sp.								X										
Reticulofenestra umbilica					X			X		X				X		X		
Sphenolithus														X				
MOLLUSKS																		
Scaphandrid gastropod							X											
Delectopecten? (pelecypod)	X										X		X					
OTHER																		
Echinoid spine			X															
Fish tooth												X						
Megafossil fragments			X									X						
Radiolarians	X																	X

* Only representative species are listed.

APPENDIX 1 (cont.)

Jewell member

SPECIES \ LOCATION	R-528	R-529	R-531	R-533	R-535	R-659	2116-4	2116-6	2116-10	2116-12	2116-15	2116-17	2116-21	2116-24				
FORAMINIFERA (Benthic, surface samples)																		
Alabamina kernensis	X										X	X	X					
Bathysiphon sp.						X		X	X		X	X	X					
Bulimina macilenta	X								X									
Cibicidina walli						X												
Cibicides elemansis						X												
Cyclamina pacifica						X	X	X			X	X	X	X				
Cyclamina sp.						X												
?Furkenkoina bramletti					X													
Globobulimina pacifica						X		X			X		X					
Gyroidina condoni	X																	
Lenticulina sp.	X						X											
Melonis sp. of McDougall (1980)						X				X								
Nodosaria longiscata	X				X			X		X	X		X					
Nodosari pyrula	X									X	X	X						
Plectofrondicularia packardi					X													
Praeglobobulimina pupoides	X				X						X							
Pseudonodosaria inflata						X		X										
Pullenia salisburyi					X													
Stilostomella lepidula	X							X		X								
Uvigerina cocoaensis species group					X		X	X	X	X		X						
Uvigerina garzaensis	X				X		X	X	X	X								
PLANKTIC FORAMINIFERA																		
Globigerina sp.	X																	
FORAMINIFERA OCCURRING ONLY IN WELL SAM.																		
Ammodiscus incertus								X										
Bathysiphon eocenicus									X									
Bolivina scabrata												X						
Bulimina corrugata									X									
Bulimina sculptilus lacinata												X						
Caucasina schencki													X					
Ceratobulimina washburnei									X									
Chilostomella oolina											X		X					
Cibicides fortunatus														X				
Cibicides sp.											X							
Dentalina sp.							X											
Dorthia sp.							X											
Eggerella elongata													X					
Glandulina laevigata										X								
Globocassidulina globosa							X			X			X	X				
Gyroidina soldanii							X	X		X	X							
Haplophragmoides obliquacamerata							X	X										
Karreriella washingtonensis											X							
Lagena hispida obliquicamerata								X										
Martinotiella communis								X										
Nodosaria delicata										X		X	X					
Nodosaria hispida												X						
Nodosaria sp.							X			X	X		X					
Oridorsalis umbonatus								X		X		X						
Planulina marhleyana											X							
Plectofrondicularia packardi multilin.									X									

APPENDIX 1 (cont.)

Jewell member (cont.)

SPECIES \ LOCATION	R-528	R-532	R-531	R-533	R-535	R-659	2116-4	2116-6	2116-10	2116-12	2116-15	2116-17	2116-21	2116-24
Praeglobobulimina sp.											X			
Pullenia bulloides											X			
Quinqueloculina imperialis							X							
Sigmolina tenuis								X		X				
Spiroplectammina tejonensis										X				
Stilostomella adolphina												X		
Trochamina globigeriniformis													X	
DIATOMS														
Stephanopyxis sp.				X										
Resting spores				X										
CALCAREOUS NANNOFOSSILS*														
Braarudosphaera bigelowii												X		
Dictyococcites scrippsae												X		
Dictyococcites bisectus		X												
Isthmolithus recurvus									X					
Reticulofenestra sp.									X			X		
Reticulofenestra umbilica												X		
OTHER														
Crinoid fragments			X											
Fish tooth						X								
Geodites (fish ear bones)	X													

Pittsburg Bluff Formation

SPECIES \ LOCATION	R-320	R-323	R-1200	R-1201
FORAMINIFERA (Benthic, surface samples)				
Cibicides fortunatus	X			
Cyclammina spp.			X	X
Globocassidulina globosa	X			
Gyroidina orbicularis	X			
Haplophragmoides spp.			X	X
Melonis pompilioides?			X	
Planulina tolmani	X			
Plectofrondicularia minuta	X			
Plectofrondicularia oregonensis	X			
Plectofrondicularia packardi	X			
Praeglobobulimina cf. P. ovata?			X	
Pseudoglandulina inflata?			X	
Spiroloculina texana	X			
Stilostomella lepidula	X			
PLANKTIC FORAMINIFERA				
Globigerina spp.	X			
Globorotalia spp.	X			
MOLLUSKS				
Tellinid pelecypod		X		
Gastropods		X		
Scaphopods		X		

APPENDIX 1 (cont.)
Smuggler Cove formation

SPECIES \ LOCATION	R-83-140	R-83-141	R-40a	R-46a	R-116	R-663	R-761	2116-1	2116-3	SAMPLES WITH ONLY MOLLUSKS PRESENT R-13	R-22	R-34	R-35	R-41	R-72	R-98	R-151	R-499
FORAMINIFERA (Benthic, surface samples)																		
Amodiscus cf. A. incertis		X						X	X									
Anomolina sp.		X																
Bathysiphon sp.						X												
Bolivina sp.		X																
Chilostomella sp.		X																
Cyclamina japonica					X			X										
Cyclamina pacifica					X	X	X	X	X									
Cyclamina sp.	X	X	X	X														
Dentalina jacksonensis						X												
Dentalina sp. A of Rau (1948)		X																
Gyroidina orbicularis planata		X																
Haplophragmoides deflata					X													
Haplophragmoides sp.	X		X	X														
Hoeglundina eocenica		X																
Karreriella washingtonensis		X																
Lenticulina cf. L. limbosus		X																
Lenticulina inornatus		X																
Martinotiella communis							X											
Nodogenerina cocoaensis		X																
Nodosaria cf. N. pyrula		X																
Nodosaria longiscata						X												
Pseudoglandulina inflata	X																	
Pseudonodosaria conica					X													
Pyrgo sp.		X																
Quinqueloculina imperialis		X						X	X									
Sigmoilina tenuis		X																
PLANKTIC FORAMINIFERA																		
Globigerina sp.		X																
FORAMINIFERA OCCURRING ONLY IN WELL SAM																		
Alabamina kernensis									X									
Astigerina crassiformis								X										
Bolivina kleinpelli									X									
Bolivina marginata								X										
Bolivina scabrata								X										
Bulimina macilenta								X	X									
Cibicides sp.								X										
Dentalina cooperensis								X										
Dentalina sp.								X										
Eggerella elongata								X										
Eggerella subconica									X									
Fissurina marginata									X									
Globobulimina pacifica									X									
Globocassidulina globosa								X	X									
Gyrodina condoni									X									
Gyrodina aff. G. rosaformis								X										
Gyrodina soldanii octocamerata								X										
Gyrodina sp.								X										
Lenticulina sp.									X									
Melonis cf. M. soldanii								X										
Melonis sp. of McDougall								X										

APPENDIX 1 (cont.)
Smuggler Cove formation (cont.)

SPECIES \ LOCATION	R-83-140	R-83-141	R-40a	R-46a	R-116	R-663	R-761	2116-1	2116-3	SAMPLES WITH ONLY MOLLUSKS PRESENT R-13	R-22	R-34	R-35	R-41	R-72	R-98	R-151	R-499
Nodosaria sp.									X									
Plectofrondicularia packardi								X	X									
Stilostomella lepidula								X	X									
Triloculina sp.									X									
Trochamina globigeriniformis								X										
Uvigerina cocoaensis species group								X	X									
Uvigerina garzaensis								X	X									
Valvulineria jacksonensis								X										
CALCAREOUS NANNOFOSSILS																		
Braarudosphaera bigelowii								X										
Dictyococcites bisectus		X																
Isthmolithus recurvus								X										
Reticulofenestra umbilica								X										
MOLLUSKS																		
PELECYPODS																		
Delectopecten?					X					X				X			X	X
Delectopecten cf. D. peckhami?											X							
Limaria sp.?										X								
Macoma sp.?												X						
Undifferentiated pelecypods													X		X			
GASTROPODS																		
Turrinosyrinx cf. T. borgenae?																X		
Gemmula cf. G. rockcreekensis?											X							
Naticid?										X								
Epitoniid?										X								
Undifferentiated gastropods												X						
SCAPHOPODS																		
Dentalium sp.?										X								

APPENDIX 2: AGE AND ENVIRONMENTAL DETERMINATIONS

CZ 11-28 Well

STRATIGRAPHIC UNIT	SAMPLE NO.	DEPTH (FEET)	FORAMINIFERA DATA* AGE	FORAMINIFERA DATA* ECOLOGY	CALC. NANNOFOSSIL AGES
Lower Smuggler Cove fm.	2116-01	1050-1080	Late Refugian ("Osw. W.fauna)	Bathyal	Subzone CP-15b
	2116-03	1290-1320			
—?—?—?—?—?—?					
Keasey Fm. (Parts may be time equiv. to Vesper)	2116-04	1410-1440	Refugian ("Keasey fauna," upper Ref. of McD. and lower Ref. of McKeel)	Bathyal	
	2116-06	1650-1680			
	2116-10	2150-2160			Prob. subzone CP15b
	2116-12	2390-2400			
	2116-15	2730-2740			
Jewell member of Keasey Formation	2116-17	2970-2980	Prob. lower Refugian	Bathyal	La. mid. Eocene. Olig.
	2116-21	3450-3460			
	2116-24	3810-3820	Poss. Narizian	Bathyal	
Hamlet formation	2116-26	4050-4060	Narizian, prob. late Narizian	Bathyal	Prob. subzone CP14b (Pos. subzone CP15a)
	2116-29	4410-4420			
Tillamook Volcanics	2116-30	4550-4560	Narizian		
	2116-34	5010-5020			
	2116-40	5690-5700			

Roy Creek member

SAMPLE NO.	STRATIGRAPHIC POSITION	FOSSIL GROUP	AGE DETERMINATION	ECOLOGIC DETERMINATION
R-339	"middle"	Mollusks	N.D.	Prob. intertidal to subtidal

Note: Molluscan determinations by Ellen Moore, Calcareous nannofossils by Dave Bukry, and foraminifera by Kristin McDougall (McD.) and Weldon Rau (Rau). N.D.= non-diagnostic.

APPENDIX 2 (cont.)

Sweet Home Creek member

SAMPLE NO.	STRATIGRAPHIC POSITION	FOSSIL GROUP	AGE DETERMINATION	ECOLOGIC DETERMINATION
R-221	Uppermost portion	Diatoms	N.D.	Neritic to shelf edge
		Mollusks	N.D.	Prob. 25-2,000m
R-398	Lower portion	Forams (McD.)	Late Nariz.-e.Refug.	Up. bathy. (200-500m)
		Forams (Rau)	Narizian (Cow.-Nest)	Middle bathyal
R-548	Upper portion	Forams (McD.)	Late Narizian	Up. bathy. (200-500m)
		Forams (Rau)	Narizian(Cow.-Nest.)	Low. mid. to up. bathyal
R-557	Upper portion	Diatoms	N.D.	Neritic to shelf edge
		Forams (McD.)	Narizian-Refugian	Up. bathy. (200-500m)
		Forams (Rau)	N.D.	Bathyal
R-558	Upper portion	Diatoms	Mid.Eoc.-Low. Olig.	Neritic to shelf edge
		Forams (McD.)	Narizian-Refugian	Outer neritic-up. bathy.
		Forams (Rau)	Prob. Narizian	Middle bathyal
R-562	Upper portion	Calc. nannofossil	Late mid. Eoc.-Olig	Open marine
R-564	Middle portion?	Forams (McD.)	Late Narizian	Up. mid. bathy.(500-1000m)
		Forams (Rau)	N.D.	Middle to upper bathyal
R-612	Upper portion	Calc. nannofossil	Subzone CP 14a (Eo.)	Open marine
		Forams (McD.)	Late Narizian	Up. bathyal (200-500m)
		Forams (Rau)	Narizian (Cow-Nest.)	Bathyal
R-615	Middle portion?	Forams (McD.)	Late Narizian	Up. bathyal (200-500m)
		Forams (Rau)	Narizian (Cow-Nest.)	Middle bathyal
R-620	Lower-middle portion	Calc. nannofossil	Subzone CP 14a (Eo.)	Open marine
		Forams (McD.)	Late Narizian	Up. bathyal (200-500m)
		Forams (Rau)	Narizian (Cow-Nest.)	Middle bathyal
R-621	Lower portion	Mollusks	N.D.	Prob. 25-2,000 meters
		Forams (McD.)	Late Narizian	Outer shelf to up. bathy.
		Forams (Rau)	Narizian (Cow-Nest.)	Up. mid. to mid. bathyal
R-628	Lower-middle portion	Forams (McD.)	Late Narizian	Up. bathyal (200-500m)
		Forams (Rau)	Narizian (Cow-Nest.)	Up. mid. to up. bathyal
R-629	Lower portion	Mollusks	N.D.	Prob. 25-2,000 meters
		Forams (McD.)	Late Narizian	Up. bathyal (200-500m)
		Forams (Rau)	Narizian (Cow-Nest.)	Up. mid. to up. bathyal

APPENDIX 2 (cont.)

Jewell member of the Keasey Formation

SAMPLE NO.	STRATIGRAPHIC POSITION	FOSSIL GROUP	AGE DETERMINATION	ECOLOGIC DETERMINATION
528	Uppermost Jewell mbr.?	Forams (McD.) Forams (Rau)	Late Refugian N.D.	Low.bathyal(1,200-2,000m) Middle bathyal
529	Uppermost Jewell mbr.?	Calc. nannofossil	Late mid. Eoc-Olig.	Open marine
533	Upper Jewell mbr.	Diatoms	N.D.	Outer neritic-shelf edge
535	Lower Jewell mbr.	Forams (McD.) Forams (Rau)	Refugian N.D.	Mid. bathy.(500-1,200m) Bathyal
659	Lowermost Jewell mbr.	Forams (McD.) Forams (Rau)	Refugian Refugian-Zemorrian	Up. bathy. or deeper Low. mid. to low. bathy.

Pittsburg Bluff Formation

SAMPLE NO.	STRATIGRAPHIC POSITION	FOSSIL GROUP	AGE DETERMINATION	ECOLOGIC DETERMINATION
R-320	Lower?	Forams (McD.) Forams (Rau)	Refugian N.D.	Mid.-low.bathy(500-2000m) Middle bathyal
R-1200	Lower?	Forams (Rau)	N.D.	Bathyal

Smuggler Cove Formation

SAMPLE NO.	STRATIGRAPHIC POSITION	FOSSIL GROUP	AGE DETERMINATION	ECOLOGIC DETERMINATION
R-83-140	Middle (1m above glau)	Forams (Rau)	N.D.	Bathyal?
R-83-141	Middle to upper portion (Approx. 8m above gla.)	Calc. nannofossil Forams (Rau)	Late mid. Eoc-Olig. Refugian	Open marine Upper bathyal
R-116	Middle portion	Forams (McD.) Forams (Rau)	Late Eocene-Olig. N.D.	N.D. Middle bathyal
R-663	Upper portion	Forams (McD.)	N.D.	Bathyal
R-761	Lower portion	Forams (McD.)	N.D.	Prob. bathyal
R-41	Lower portion	Mollusks	N.D.	Prob. 25-2,000m
R-98	Middle portion	Mollusks	Prob. Matlockian	N.D.
R-151	Middle portion	Mollusks	N.D.	Prob. 25-2,000m
R-13	Middle portion	Mollusks	N.D.	Poss. 15-100m
R-22	Middle portion	Mollusks	Possibly Galvinian	N.D.

APPENDIX 3: FOSSIL LOCALITIES

Sample Number	Location	Outcrop
R-13	SW¼ NW¼ sec. 19, T5N, R8W	Road cut
R-22	NW¼ SE¼ sec. 29, T5N, R8W	Road cut
R-34	SE¼ SW¼ sec. 29, T5N, R8W	Road cut
R-35	SW¼ SW¼ sec. 29, T5N, R8W	Road cut
R-40a	SW¼ NE¼ sec. 19, T5N, R8W	Road cut
R-41	NW¼ SE¼ sec. 19, T5N, R8W	Road cut
R-46a	NE¼ SW¼ sec. 20, T5N, R8W	Road cut
R-72	SE¼ NW¼ sec. 30, T5N, R8W	Road cut
R-98	NE¼ NE¼ sec. 35, T5N, R9W	Quarry
R-116	NW¼ SE¼ sec. 36, T5N, R9W	Road cut
R-151	NW¼ SW¼ sec. 2, T4N, R9W	Road cut
R-221	NW¼ SW¼ sec. 8, T4N, R8W	Road cut
R-320	SE¼ NW¼ sec. 21, T5N, R8W	Road cut
R-323	SE¼ NW¼ sec. 21, T5N, R8W	Road cut
R-339	SW¼ SW¼ sec. 8, T3N, R9W	Road cut
R-365	SE¼ SE¼ sec. 9, T3N, R9W	Road cut
R-398	SW¼ SE¼ sec. 9, T3N, R9W	Quarry
R-419	NE¼ SE¼ sec. 5, T3N, R8W	Road cut
R-499	NE¼ NE¼ sec. 2, T4N, R9W	Road cut
R-528	NE¼ NE¼ sec. 26, T4N, R9W	Stream cut
R-529	NE¼ NE¼ sec. 26, T4N, R9W	Stream cut
R-531	NW¼ NW¼ sec. 25, T4N, R9W	Stream cut
R-533	SE¼ NW¼ sec. 25, T4N, R9W	Stream cut
R-535	NW¼ NE¼ sec. 25, T4N, R9W	Stream cut
R-548	NE¼ SE¼ sec. 19, T4N, R8W	Stream cut
R-557	SW¼ SE¼ sec. 17, T4N, R8W	Stream cut
R-558	SW¼ SE¼ sec. 17, T4N, R8W	Stream cut
R-559	SW¼ SE¼ sec. 17, T4N, R8W	Stream cut
R-562	SW¼ SW¼ sec. 17, T4N, R8W	Stream cut
R-564	SW¼ SW¼ sec. 17, T4N, R8W	Stream cut
R-608	NW¼ SW¼ sec. 20, T4N, R8W	Stream cut
R-612	NE¼ SW¼ sec. 20, T4N, R8W	Stream cut
R-615	SE¼ SW¼ sec. 20, T4N, R8W	Stream cut
R-620	NE¼ NE¼ sec. 29, T4N, R8W	Stream cut
R-621	SE¼ NE¼ sec. 29, T4N, R8W	Stream cut
R-628	NW¼ SW¼ sec. 28, T4N, R8W	Stream cut
R-629	SE¼ SW¼ sec. 28, T4N, R8W	Stream cut
R-631	NW¼ SW¼ sec. 33, T4N, R8W	Stream cut
R-633	SE¼ SW¼ sec. 22, T4N, R9W	Road cut
R-761	NW¼ SW¼ sec. 36, T5N, R9W	Road cut
R-1200	SE¼ NW¼ sec. 21, T5N, R8W	Road cut
R-1201	NE¼ NW¼ sec. 21, T5N, R8W	Road cut
R-83-140	SE¼ NE¼ sec. 9, T4N, R9W	Road cut
R-83-141	SE¼ SW¼ sec. 9, T4N, R9W	Road cut
R-83-CR4		Stream cut

APPENDIX 4: GEOCHEMICAL ANALYSES

Tillamook Volcanics

Sample	SiO_2	Al_2O_3	TiO_2	Fe_2O_3	FeO	MnO	CaO	MgO	K_2O	Na_2O	P_2O_5	FeO*
R-339	49.15	14.84	3.46	6.21	7.12	0.24	10.22	4.23	0.82	3.14	0.57	12.71
R-362	49.72	14.69	3.77	5.74	6.58	0.21	10.39	4.64	0.78	2.89	0.60	11.75
R-366	49.15	14.89	3.86	6.42	7.36	0.20	9.41	4.56	0.81	2.87	0.47	13.14
R-395	50.18	14.91	3.54	5.72	6.55	0.29	10.11	4.53	0.94	2.73	0.50	11.70
R-398	53.54	14.70	2.65	5.66	6.48	0.20	7.58	3.11	1.39	3.71	1.18	11.57
R-418	49.55	15.69	3.14	5.74	6.58	0.23	10.40	4.87	0.67	2.72	0.42	11.74
R-424[2]	50.71	15.11	3.78	5.76	6.60	0.25	8.22	4.09	1.11	3.73	0.64	11.78
R-626	48.94	14.23	3.72	6.75	7.73	0.25	9.70	4.57	0.61	2.96	0.54	13.81
R-645	50.43	16.30	3.14	5.22	5.98	0.18	10.12	3.94	0.93	3.26	0.51	10.68
R-667	49.31	14.34	3.01	6.03	6.90	0.19	9.79	6.71	0.79	2.56	0.38	13.11
R-672	50.70	16.79	3.27	5.06	5.80	0.17	9.60	3.74	1.01	3.31	0.54	10.36
R-674	56.16	14.75	2.35	5.27	6.04	0.19	6.30	2.73	1.76	3.67	0.79	10.78
R-685a	49.91	14.70	3.30	6.24	7.15	0.18	10.07	4.56	0.64	2.82	0.44	12.77
R-907	53.24	14.57	2.66	5.71	6.54	0.22	7.48	3.16	1.38	3.85	1.19	11.68
MS-42	49.49	16.06	3.01	5.40	6.19	0.26	10.19	5.61	0.77	2.61	0.42	11.05
MS-335	49.01	14.06	3.63	6.55	7.50	0.21	9.67	5.29	0.87	2.72	0.50	13.40
MS-643	49.91	14.34	3.59	6.07	6.95	0.19	9.73	5.06	0.83	2.85	0.49	12.41
MS-983	48.48	14.43	2.98	6.23	7.13	0.21	10.26	6.79	0.60	2.50	0.39	12.74
CZ-5+	51.54	14.85	3.37	5.72	6.55	0.15	7.56	4.28	1.13	4.14	0.52	11.70
Avg.	50.48	14.96	3.28	5.86	6.72	0.21	9.31	4.55	0.94	3.11	0.59	12.05
Std.Dev.	σ1.93	σ0.74	σ.42	σ.46	σ.53	σ.03	σ1.23	σ1.07	σ.30	σ.50	σ.23	σ0.97

Analyses by Dr. Peter Hooper of Wash. St. Univ. using XRF with his "international standard".

[2] Tillamook Volcanics clast in the Roy Creek mbr. of the Hamlet formation.

+Well sample from 4,900' in the CZ 11-28 well.

$FeO^* = (.9\ Fe_2O_3) + FeO$

APPENDIX 4 (cont.)

Sample	SiO_2	Al_2O_3	TiO_2	Fe_2O_3	FeO	MnO	CaO	MgO	K_2O	Na_2O	P_2O_5	FeO*
Cole Mtn. basalt[1] (thesis area)												
R-251	56.64	16.49	1.63	4.57	5.24	0.12	7.85	3.01	0.75	3.40	0.29	9.35
R-561	58.25	15.99	1.43	4.35	4.99	0.16	6.71	3.67	0.84	3.30	0.32	8.90
R-614	55.89	16.37	2.11	4.58	5.25	0.16	7.05	3.67	0.78	3.85	0.30	9.37
RQ-850	56.46	15.97	1.49	4.02	4.60	0.14	10.15	2.25	0.81	3.78	0.32	8.22
Avg.	56.81	16.21	1.67	4.38	5.02	0.15	7.94	3.15	0.80	3.58	0.31	8.96
Std.Dev.	σ1.01	σ0.26	σ.31	σ.26	σ.30	σ.02	σ1.55	σ.68	σ.04	σ.27	σ.02	σ.54
Cole Mtn. basalt[1] (Washington Co.)												
RCC1	51.44	17.95	1.33	4.51	5.17	0.14	10.83	5.25	0.16	2.94	0.28	9.23
Type Area Goble Volcanics[1]												
RTG-1	55.04	17.43	1.29	4.32	4.95	0.14	8.98	3.60	0.67	3.36	0.22	8.84
RTG-5	53.99	18.28	1.13	4.02	4.60	0.15	9.90	4.09	0.51	3.17	0.16	8.22
Avg.	54.52	17.85	1.21	4.17	4.76	0.15	9.44	3.85	0.59	3.27	0.19	8.53
Gray's River Area Goble Volcanics[1]												
RG1	49.85	16.26	3.18	5.79	6.63	0.19	9.57	3.96	0.84	3.20	0.52	11.84
RG2	50.31	15.46	3.29	6.04	6.92	0.20	9.35	4.12	0.82	3.00	0.49	12.36
RG3	55.03	14.75	2.90	5.64	6.46	0.19	6.78	2.36	1.44	3.50	0.97	11.54
Avg.	51.73	15.49	3.12	5.82	6.67	0.19	8.57	3.48	1.03	3.23	0.66	11.90
Std.Dev.	σ2.87	σ0.76	σ.20	σ.20	σ.23	σ.01	σ1.55	σ.97	σ.35	σ.25	σ.27	σ0.41

[1]Analyses by Dr. Peter Hooper of Wash. St. Univ. using XRF with his "international standard".

$FeO^* = (.9\ Fe_2O_3) + FeO$

APPENDIX 4 (cont.)

COLUMBIA RIVER BASALT GROUP[2]											
Grande Ronde Basalt (Low MgO-Low TiO_2)											
Sample	SiO_2	Al_2O_3	TiO_2	Fe_2O_3	FeO	MnO	CaO	MgO	K_2O	Na_2O	P_2O_5
R-38	55.96	14.93	2.10	2.00	10.04	0.21	6.72	3.64	1.62	2.46	0.33
R-40b	55.42	14.95	2.08	2.00	10.49	0.20	6.55	3.50	1.56	2.93	0.31
R-40c	55.77	15.10	2.11	2.00	10.16	0.18	6.61	3.61	1.49	2.65	0.31
R-45	56.06	15.02	2.06	2.00	10.16	0.18	6.52	3.60	1.66	2.41	0.32
R-51	56.79	15.52	2.17	2.00	10.16	0.18	5.49	3.23	1.73	2.40	0.32
R-60	55.83	15.09	2.10	2.00	10.11	0.18	6.55	3.54	1.68	2.62	0.31
R-98	55.95	14.81	2.07	2.00	10.03	0.18	6.82	3.55	1.67	2.61	0.31
R-242	55.23	14.72	2.27	2.00	10.86	0.21	6.67	3.64	1.61	2.43	0.34
R-597	56.48	13.77	2.23	2.00	10.93	0.21	7.28	3.58	1.58	1.96	0.32
R-911	54.63	14.68	2.30	2.00	11.16	0.21	6.84	3.57	1.54	2.74	0.34
R-917	56.88	13.68	2.29	2.00	10.41	0.20	7.18	3.64	1.59	2.07	0.39
CZ-1*	58.50	14.77	1.76	2.00	9.10	0.15	6.28	3.03	1.68	2.49	0.28
CZ-2*	58.24	14.48	1.73	2.00	9.53	0.17	6.25	3.06	1.84	2.62	0.27
Avg.++	55.91	14.75	2.16	2.00	10.41	0.19	6.66	3.55	1.61	2.48	0.33
Std.Dev.	σ0.67	σ0.56	σ.09	σ.00	σ0.40	σ.01	σ.46	σ.12	σ.07	σ.28	σ.02
Grande Ronde Basalt (Low MgO-High TiO_2)											
R-2	54.75	14.86	2.42	2.00	10.97	0.19	6.88	3.61	1.47	2.44	0.40
Grande Ronde Basalt (High MgO)											
R-26a	53.62	14.91	2.04	2.00	10.21	0.20	8.16	4.92	1.02	2.62	0.30
R-71	53.94	14.85	2.00	2.00	10.35	0.23	8.15	4.79	1.01	2.39	0.31
R-78	53.48	14.77	2.05	2.00	10.66	0.21	8.17	4.83	0.93	2.57	0.31
R-131	54.00	14.90	1.98	2.00	10.26	0.21	8.49	4.77	1.01	2.10	0.29
R-523	54.28	15.26	2.21	2.00	8.93	0.18	8.60	4.91	1.25	2.07	0.31
CZ-3*	56.87	14.46	2.26	2.00	8.45	0.17	7.32	3.94	1.15	2.73	0.45
CZ-4*	55.67	14.55	1.85	2.00	9.05	0.20	8.29	4.31	1.14	2.46	0.26
Avg.++	53.86	14.94	2.06	2.00	10.08	0.21	8.31	4.84	1.04	2.35	0.30
Std.Dev.	σ0.32	σ0.19	σ.09	σ.00	σ0.67	σ.02	σ.21	σ.07	σ.12	σ.26	σ.01

[2] Analyses by Hooper using CRB std. *Well sample ++Avg. exclud. well samples.

COLUMBIA RIVER BASALT GROUP (cont.)[2]

Pomona Member

Sample	SiO_2	Al_2O_3	TiO_2	Fe_2O_3	FeO	MnO	CaO	MgO	K_2O	Na_2O	P_2O_5
R-20	50.05	15.52	2.05	2.00	8.63	0.18	11.33	7.52	0.34	1.98	0.40
R-310**	50.11	17.39	1.66	2.00	8.78	0.18	10.77	7.17	0.38	1.53	0.27
R-357	50.29	15.57	1.97	2.00	9.05	0.17	11.18	7.61	0.13	1.65	0.37
R-297**	49.69	16.66	2.21	2.00	10.28	0.18	10.79	6.81	0.32	1.18	0.20
R-617**	47.97	16.06	1.73	2.00	10.46	0.18	9.71	10.81	0.35	0.89	0.14
Avg.	49.62	16.24	1.92	2.00	9.44	0.18	10.74	7.98	0.30	1.45	0.28
Std.Dev.	σ0.95	σ0.79	σ.23	σ.00	σ.86	σ.00	σ0.61	σ1.61	σ.10	σ.42	σ.11
Frenchman Springs Member											
R-76	50.34	13.94	3.23	2.00	12.54	0.23	9.02	4.97	0.80	2.33	0.61
"Umatilla Member" analysis[3]											
R-40[3]	54.86	15.20	2.73	2.00	9.38	0.24	6.24	2.82	2.69	2.96	0.89

[2] Analyses by Dr. Peter Hooper of Wash. St. Univ. using XRF with the Columbia River Basalt standard.

**Gabbroic basalts that have been assigned to the Pomona Member.

[3] Incorrect analysis that was resubmitted and found to have a Grande Ronde Basalt chemistry. This analysis was probably the result of "computor error" at Wash. St.

APPENDIX 5: GEOCHEMICAL ANALYSES (DATA FROM OTHER WORKERS)

Tillamook Volcanics

Sample	SiO_2	Al_2O_3	TiO_2	Fe_2O_3	FeO	MnO	CaO	MgO	K_2O	Na_2O	P_2O_5
82472	49.60	14.91	3.16	6.30	7.21	0.18	10.35	4.24	0.64	2.99	0.41
82672	50.66	14.27	3.56	6.20	7.11	0.21	9.40	3.99	0.82	3.27	0.51
72052	49.60	18.33	2.98	5.13	5.88	0.19	10.46	3.16	0.76	3.07	0.45
719101	51.57	15.10	3.21	5.74	6.58	0.22	8.36	3.86	1.07	3.68	0.61
7773	49.51	14.20	3.88	6.59	7.55	0.23	9.03	4.26	0.94	3.23	0.58
372021	48.52	16.00	4.43	5.80	6.64	0.23	9.10	4.08	1.02	3.27	0.92
372022	51.08	17.65	3.41	5.28	6.05	0.16	7.06	3.21	1.24	3.57	1.29
372023	51.72	16.96	4.05	5.20	5.96	0.24	6.67	2.19	2.39	4.01	0.62
372113	49.19	10.07	1.92	5.16	5.91	0.17	11.49	13.81	0.42	1.64	0.23
372112	48.70	14.16	3.64	6.33	7.25	0.28	10.07	5.09	0.86	3.08	0.51
372142	58.42	15.76	1.68	4.44	5.08	0.18	5.59	2.01	1.92	4.30	0.63
372144	58.15	15.89	1.67	4.36	4.99	0.17	5.58	2.11	1.93	4.50	0.64
372311	53.30	15.59	2.56	5.35	6.13	0.27	7.59	2.85	1.44	3.83	1.08
372312	50.33	16.31	3.67	5.13	5.87	0.22	8.48	2.88	1.27	3.81	2.03
372313	52.21	16.06	2.73	5.45	6.24	0.23	7.55	2.95	1.52	3.93	1.14
372314	51.96	16.13	3.54	4.90	5.62	0.22	8.77	3.19	1.17	3.80	0.69
362361	50.58	14.86	3.62	5.85	6.70	0.21	9.19	4.18	0.83	3.42	0.57
82222	52.92	15.31	2.96	5.42	6.20	0.24	7.70	3.68	1.11	3.60	0.87
91962	49.98	15.09	3.17	5.94	6.81	0.18	10.26	4.62	0.59	2.94	0.42
91932	51.12	14.78	3.56	5.94	6.81	0.24	9.41	3.65	0.87	3.13	0.50
7772	49.47	13.93	3.78	6.64	7.61	0.23	9.26	4.25	0.91	3.34	0.58
71991	52.47	14.31	3.01	6.18	7.08	0.24	7.88	3.53	1.12	3.50	0.69
82612	50.45	16.61	3.15	5.40	6.18	0.18	9.66	3.86	0.82	3.23	0.45
7742	53.16	14.83	2.85	5.75	6.59	0.25	7.53	3.42	1.33	3.57	0.72
73034	49.53	14.69	3.39	5.89	6.75	0.20	11.13	4.24	0.76	2.89	0.51
Avg. of	51.37	15.27	3.18	5.62	6.44	0.21	8.69	4.00	1.11	3.40	0.71
Mumford	σ2.49	σ1.54	σ.68	σ.59	σ.68	σ.03	σ1.60	σ2.15	σ.45	σ.56	σ.37
2ESSP1	53.57	16.04	2.63	5.08	5.82	0.26	7.75	3.36	1.35	3.38	0.76
2ESRR2	52.64	15.16	2.73	5.60	6.41	0.20	7.75	3.64	1.40	3.59	0.90
2ESWCC	52.78	14.97	2.67	5.39	6.17	0.21	7.65	3.40	1.52	3.71	1.54
2ESB1	53.51	16.01	3.31	4.79	5.48	0.17	7.44	1.63	1.59	4.16	1.91
2ESPB2	59.96	15.91	1.31	4.35	4.98	0.16	4.74	1.26	2.39	4.47	0.44
2ESPB3	54.00	14.62	2.76	5.38	6.16	0.19	7.67	3.34	1.92	3.18	0.77
2ES30	63.41	16.37	1.16	2.90	3.32	0.15	3.50	0.85	2.85	5.16	0.34
2ES54	54.47	14.73	2.79	5.24	6.00	0.19	7.71	3.25	1.80	3.08	0.75
2ES78	54.99	15.55	2.23	5.15	5.90	0.21	6.91	3.00	1.53	3.65	0.88
2ES86	53.29	16.22	2.40	5.22	5.98	0.22	7.89	3.29	1.20	3.54	0.74
2ES93	50.23	17.69	3.36	5.19	5.95	0.23	9.80	3.06	0.90	3.10	0.49
2ES108	66.18	15.74	0.79	2.66	3.04	0.14	3.05	0.13	3.38	4.73	0.17
2ES109	50.08	16.27	3.32	5.78	6.62	0.18	9.71	3.48	0.92	3.02	0.62
2ES114	60.86	15.58	1.31	4.27	4.89	0.16	4.60	1.18	2.38	4.32	0.45
2ES138	58.02	15.94	1.73	4.50	5.15	0.19	6.30	1.78	1.43	4.29	0.67
2ES181	52.54	14.89	2.72	5.50	6.30	0.23	7.64	3.63	1.41	3.65	1.49
2ES206	51.18	14.62	3.54	5.90	6.75	0.17	8.76	4.25	1.02	3.21	0.60
2ES225	48.92	17.46	2.79	5.47	6.26	0.19	10.73	4.52	0.63	2.66	0.37
2ES228	52.80	14.64	3.34	5.57	6.38	0.20	8.32	3.51	1.18	3.40	0.66
2ES236	67.44	15.32	0.67	2.85	3.27	0.09	1.57	0.12	3.43	5.11	0.13
2ES241	53.40	15.60	2.65	5.18	5.94	0.26	7.54	2.63	1.52	3.86	1.43
2ES258	49.24	15.80	3.36	5.92	6.78	0.19	9.86	4.29	0.85	3.16	0.56
2ES266	53.32	15.00	2.40	5.55	6.36	0.21	7.31	3.25	1.39	3.96	1.26
2ES270	58.44	14.97	1.74	4.86	5.56	0.20	5.56	1.83	2.12	4.01	0.69
2ES353	52.40	15.37	2.81	5.66	6.49	0.20	8.07	3.43	1.17	3.51	0.89
Avg. of	55.11	15.62	2.42	4.96	5.68	0.19	7.11	2.72	1.65	3.76	0.78
Safley	σ5.02	σ.80	σ.84	σ.92	σ1.06	σ.03	σ2.22	σ1.26	σ.74	σ.65	σ.44

Above data from Mumford (in prep.) and Safley (in prep.).
σ equals one standard deviation

APPENDIX 5 (cont.)

Tillamook Volcanics (cont.)

Sample	SiO_2	Al_2O_3	TiO_2	Fe_2O_3	FeO	MnO	CaO	MgO	K_2O	Na_2O	P_2O_5
8-21-5	51.03	13.87	3.49	6.11	7.00	0.32	9.10	3.34	1.62	3.52	0.61
11-20-2	53.87	13.91	2.85	5.43	6.22	0.27	7.67	3.01	1.32	4.38	1.06
11-20-3	51.60	13.98	3.33	5.76	6.59	0.41	8.83	3.28	1.40	4.04	0.78
9-4-3	53.34	13.97	2.59	5.79	6.63	0.34	7.45	3.11	1.26	4.30	1.22
11-21-1	53.72	14.62	2.65	5.32	6.10	0.20	8.21	2.85	1.32	4.31	0.69
11-21-5	53.49	13.10	2.91	5.93	6.79	0.23	7.94	3.02	1.41	4.03	1.15
11-21-6	52.85	13.27	2.90	5.74	6.58	0.21	8.22	3.31	1.60	4.13	1.17
10-4-1	52.21	14.29	2.73	5.85	6.70	0.26	8.16	2.87	1.40	4.27	1.25
10-4-5	53.04	13.71	3.01	5.48	6.28	0.20	8.31	3.25	1.56	4.13	1.03
10-8-4	52.37	13.81	3.03	5.80	6.64	0.16	7.83	3.16	2.18	4.04	0.96
9-17-3	52.11	14.33	3.14	5.93	6.80	0.16	7.56	3.05	1.91	4.01	1.01
10-17-1	59.75	14.27	1.64	4.85	5.56	0.21	5.45	1.61	0.98	5.02	0.66
11-18-2	50.25	12.81	3.55	6.83	7.82	0.24	9.34	4.36	0.99	3.30	0.49
11-28-1	50.09	13.38	3.32	6.33	7.25	0.24	10.45	4.33	0.89	3.21	0.50
11-28-2	51.27	13.05	3.29	6.15	7.05	0.24	9.83	4.34	1.04	3.23	0.52
11-20-1	49.92	13.59	3.31	6.29	7.21	0.22	9.88	4.37	1.51	3.18	0.51
Avg. of	52.56	13.76	2.98	5.83	6.70	0.24	8.39	3.32	1.40	3.94	0.85
Jackson	σ2.30	σ.52	σ.46	σ.46	σ.53	σ.06	σ1.20	σ.73	σ.34	σ.52	σ.28
9(O)	54.19	16.72	2.34	4.85	5.55	0.18	7.20	2.18	1.70	4.25	0.77
10(O)	54.81	16.80	2.27	4.65	5.32	0.18	7.06	2.19	1.85	4.13	0.75
11(O)	54.48	17.60	1.93	4.25	4.87	0.19	6.06	1.91	3.24	4.21	0.72
12(O)	64.77	16.20	0.79	3.29	3.77	0.12	3.29	3.77	0.12	2.72	0.93
13(O)	55.67	17.10	2.25	4.43	5.07	0.19	6.64	1.92	1.68	4.26	0.78
Gv-1(N)	62.28	16.04	1.12	3.42	3.92	0.19	3.50	1.30	3.13	4.79	0.31
Gv-2(N)	62.60	15.57	1.04	3.60	4.12	0.15	3.78	1.07	3.12	4.66	0.31
265(N)	56.92	16.09	1.89	4.48	5.13	0.22	5.84	1.94	2.08	4.71	0.72
9-4-1(T)	54.59	16.19	2.14	5.20	5.95	0.19	6.66	2.86	1.61	3.67	0.94
9-4-2(T)	53.06	16.42	2.16	5.44	6.23	0.22	6.99	3.36	1.61	3.55	0.97
9-4-3(T)	51.10	15.08	3.20	6.17	7.07	0.16	8.31	3.63	1.24	3.32	0.71
9-4-4(T)	53.99	15.90	2.00	5.06	5.79	0.22	5.93	2.38	2.13	3.78	0.81
9-114(T)	52.05	15.44	2.86	5.63	6.44	0.22	7.88	3.36	1.29	3.67	1.16
10-181(T)	50.22	16.60	3.49	5.26	6.02	0.35	9.16	3.92	0.96	3.50	0.52
1-25-2(T)	54.80	16.51	2.40	2.00	8.61	0.19	7.21	3.27	1.75	2.61	0.64
T-C&W(K)	50.17	15.77	3.00	5.74	6.60	0.19	9.79	4.02	0.48	3.46	0.45
Avg. of	55.35	16.25	2.18	4.59	5.65	0.20	6.55	2.52	1.87	3.98	0.67
N. Grn.	σ4.35	σ0.64	σ.75	σ1.08	σ1.24	σ.05	σ1.94	σ1.00	σ.77	σ.65	σ.26
Mtn. Area											
KAC-40	48.58	14.20	3.17	6.10	8.32	0.20	10.02	5.13	0.68	3.04	0.36
KAC-13	46.70	14.45	2.61	5.71	7.02	0.18	11.40	6.39	0.64	3.20	0.32
KAC-39+	47.29	13.83	2.85	5.85	7.15	0.25	11.14	6.43	1.11	2.79	0.31
Avg.of	47.52	14.16	2.88	5.89	7.44	0.21	10.85	5.98	0.81	3.01	0.33
Cameron	σ0.96	σ0.31	σ.28	σ.19	σ.56	σ.04	σ0.73	σ.74	σ.26	σ.21	σ.03

+Submarine basalt

Data from Jackson (1983), Olbinski(O) (1983), Timmons(T) (1981), Kadri(K) (1982), Nelson(N) (1985), and Cameron (1980).

APPENDIX 5 (cont.)

Sample	SiO_2	Al_2O_3	TiO_2	Fe_2O_3	FeO	MnO	CaO	MgO	K_2O	Na_2O	P_2O_5
Cole Mountain basalt											
362411	58.00	15.81	1.47	4.19	4.80	0.18	7.55	3.80	0.60	3.29	0.31
7321	56.41	16.42	1.42	4.25	4.87	0.15	8.15	4.29	0.60	3.16	0.27
9961	56.86	17.75	1.55	3.85	4.41	0.10	8.22	3.32	0.70	2.96	0.27
63013	58.62	16.53	1.51	3.96	4.54	0.12	6.93	3.26	0.78	3.44	0.33
73042	56.93	15.92	1.47	4.53	5.19	0.17	7.85	3.98	0.41	3.21	0.33
73034	58.10	15.92	1.45	4.31	4.94	0.17	7.28	3.61	0.35	3.56	0.32
7373	56.51	15.08	1.40	4.41	5.06	0.16	9.01	4.02	0.69	3.34	0.31
717103	55.74	16.93	1.50	4.63	5.31	0.13	8.12	3.80	0.70	2.87	0.26
Avg.of	57.15	16.30	1.47	4.27	4.89	0.15	7.89	3.76	0.60	2.85	0.30
Mumford	σ0.99	σ0.81	σ.05	σ.27	σ.31	σ.03	σ.64	σ.35	σ.15	σ1.06	σ.03
2ES17	54.03	15.89	1.66	4.99	5.72	0.16	9.11	4.77	0.62	2.74	0.31
2ES21	54.30	16.06	1.71	4.88	5.59	0.16	8.45	4.70	0.49	3.39	0.28
2ES91	55.91	16.81	1.38	4.47	5.12	0.11	6.59	4.65	1.50	3.23	0.23
2ES277	56.87	16.34	1.46	4.67	5.35	0.16	6.59	3.79	0.79	3.75	0.23
Avg. of	55.28	16.28	1.55	4.75	5.45	0.15	7.69	4.48	0.85	3.28	0.26
Safley	σ1.35	σ0.40	σ.16	σ.23	σ.27	σ.03	σ1.29	σ.46	σ.45	σ.42	σ.04
QW4560(O)	54.83	19.69	1.05	3.32	3.81	0.11	8.45	3.26	0.86	4.39	0.22
Gv-3(N)	54.78	16.65	1.45	4.51	5.16	0.15	7.06	2.83	0.55	3.83	0.32
9-6-8(T)	53.41	18.34	1.33	4.25	4.86	0.17	9.44	4.63	0.14	3.17	0.27
10188(T)	53.79	16.64	1.87	5.04	5.77	0.17	8.02	4.37	0.66	3.37	0.39
Avg.of	54.88	17.83	1.42	4.28	4.90	0.15	8.24	3.77	0.55	3.69	0.30
N. Grn.	σ1.84	σ1.48	σ.34	σ.72	σ.82	σ.03	σ.99	σ.86	σ.30	σ.54	σ.07
Mtn. Area											
Altered Cole Mountain basalt											
2ES504	54.62	15.05	2.39	5.55	6.36	0.23	7.48	3.44	0.96	3.08	0.83
2ES23	56.75	14.54	2.16	5.40	6.18	0.15	6.61	3.62	0.96	2.91	0.72
2ES61	54.86	14.03	2.38	6.00	6.87	0.13	6.08	4.65	1.02	3.00	0.97
2ES351	58.76	13.56	2.10	5.39	6.17	0.17	5.56	2.82	1.22	3.18	1.07
Avg. of	56.24	14.30	2.26	5.59	6.40	0.17	6.43	3.63	1.04	3.04	0.90
Safley	σ1.93	σ0.64	σ.15	σ.29	σ.33	σ.04	σ.82	σ.76	σ.12	σ.11	σ.15
Type area Goble Volcanics											
CBa	55.70	16.69	1.98		7.73*	0.12	8.65	4.86	0.53	2.91	—
CBb	52.68	15.56	1.39		7.57*	0.06	9.25	7.17	0.41	2.61	—
77-06	57.61	16.37	1.68		7.43*	0.03	7.34	3.38	1.05	3.03	—
77-10	52.54	17.40	2.42		8.90*	0.14	11.34	5.80	0.50	2.70	—
77-14	57.52	18.77	2.18		7.87*	0.11	9.15	2.98	0.70	3.05	—
77-20	56.17	16.35	2.34		8.32*	0.07	6.55	3.52	1.10	3.01	—
77-44	52.78	16.07	1.88		8.70*	0.13	11.10	5.70	0.35	2.64	—
77-46	54.33	16.48	1.79		8.29*	0.13	9.41	5.34	0.54	2.86	—
Avg.of	54.92	16.71	1.96		8.10*	0.10	9.10	4.85	0.65	2.85	—
Burr	σ2.12	σ0.98	σ.34		σ.54	σ.04	σ1.64	σ1.45	σ.28	σ.18	—
Grays River area Goble Volcanics											
5	50.6	16.8	3.1	6.3	6.3	0.22	8.6	2.7	1.1	3.6	0.64
6	51.1	15.4	3.6	7.1	6.5	0.22	8.1	3.2	0.93	3.2	0.60
7	50.8	15.7	3.6	6.0	6.8	0.19	8.2	3.3	1.2	3.6	0.75
8	48.6	15.8	3.3	5.5	7.6	0.21	10.3	4.7	0.86	2.8	0.46
9	49.4	15.2	3.3	3.4	9.9	0.18	10.4	4.7	0.59	2.7	0.46
10+	49.1	13.8	3.8	3.7	11.6	0.22	9.7	4.1	0.36	2.9	0.77
11+	49.0	16.4	2.6	3.8	7.4	0.15	11.9	5.5	0.29	2.6	0.39
12+	48.3	16.0	3.4	4.9	8.4	0.21	10.6	4.6	0.36	2.8	0.46
13+	46.7	15.8	4.0	6.6	7.6	0.53	10.4	5.0	0.28	2.5	0.60
Avg.of	49.3	15.7	3.4	5.3	8.0	0.24	9.8	4.2	0.66	3.0	0.57
Wolf.	σ1.4	σ0.8	σ.4	σ1.4	σ1.7	σ.11	σ1.3	σ.9	σ.36	σ.4	σ.14

equals FeO(.9 Fe_2O_3+FeO) +equals submarine basalt. Data from Mumford(in prep.) Safley(in prep), Olbinski(1983), Timmons(1981), Nelson(N)(1985), Burr(1978), and Wolf and McKee(1972).

APPENDIX 5 (cont.)

Siletz River Volcanics

Sample	SiO_2	Al_2O_3	TiO_2	Fe_2O_3	FeO	MnO	CaO	MgO	K_2O	Na_2O	P_2O_5
1	48.6	14.7	1.7	4.3	6.7	0.19	12.9	8.6	0.10	2.1	0.16
2	49.2	14.7	1.6	3.2	7.8	0.19	11.9	8.7	0.26	2.3	0.14
3	49.2	14.1	1.4	4.2	8.6	0.19	11.9	7.7	0.16	2.5	0.15
4	48.2	16.1	2.7	4.8	7.9	0.19	11.0	6.2	0.06	2.4	0.38
5	48.5	14.7	2.5	5.9	8.1	0.23	11.4	5.8	0.12	2.5	0.27
6	48.4	14.1	2.9	5.2	8.5	0.30	11.9	5.7	0.17	2.7	0.29
7	47.9	14.3	2.9	5.5	8.6	0.26	11.8	5.6	0.19	2.6	0.30
8	48.4	14.0	2.7	5.2	9.2	0.23	11.5	5.7	0.15	2.7	0.31
Avg.of	48.6	14.6	2.3	4.8	8.2	0.22	11.8	6.7	0.15	2.5	0.25
Snavely	σ.5	σ0.7	σ.6	σ.9	σ.7	σ.04	σ0.6	σ1.4	σ.06	σ.2	σ.09
					Crescent Formation						
1	49.3	15.8	2.1	5.1	6.9	0.25	13.6	5.2	0.30	2.4	0.25
2	48.6	14.9	2.3	6.2	6.3	0.34	13.4	5.1	0.17	2.5	0.30
3	49.2	14.9	1.7	4.2	7.1	0.20	12.1	8.1	0.17	2.0	0.31
Avg.of	49.0	15.2	2.0	5.2	6.8	0.26	12.7	6.1	0.21	2.3	0.29
Wolf.	σ0.4	σ0.5	σ.3	σ1.0	σ.4	σ.07	σ0.7	σ1.7	σ.08	σ.3	σ.03

Avg.of 10 samples (Snavely)	48.5	14.6	2.2	4.3	8.0	0.21	11.8	7.2	0.22	2.6	0.26

Data from Snavely and others (1968) and Wolf and McKee (1972).

APPENDIX 6: CIPW NORMATIVE COMPOSITIONS

Tillamook Volcanics

Sample	Qtz.	Orth.	Albite	Anorth.	Diop.	Hed.	Ens.	Ferr.	Oliv.	Mag.	Ilm.	Apat.	C.I.*
R-339	4.91	4.86	26.56	23.96	15.10	3.30	3.62	0.91	0	8.99	6.57	1.35	38.5
R-362	6.70	4.62	24.45	24.79	16.29	1.77	4.10	0.51	0	8.31	7.15	1.42	38.1
R-366	6.66	4.80	24.28	25.34	12.46	2.08	5.67	1.09	0	9.29	7.32	1.11	37.9
R-395	7.38	5.57	23.09	25.63	14.66	2.22	4.57	0.80	0	8.28	6.72	1.18	37.3
R-398	9.72	8.23	31.39	19.33	6.30	2.25	4.88	2.01	0	8.19	5.03	2.79	28.7
R-418	6.60	3.97	23.01	28.61	13.47	2.69	4.97	1.14	0	8.31	5.96	0.99	36.5
R-424	5.48	6.57	31.55	21.19	10.79	1.44	5.26	0.80	0	8.33	7.17	1.52	33.8
R-626	6.48	3.61	25.04	23.72	13.66	3.03	5.14	1.31	0	9.77	7.06	1.28	39.9
R-645	5.24	5.51	27.58	27.08	13.58	2.16	3.59	0.66	0	7.55	5.96	1.21	33.5
R-667	4.56	4.68	21.66	25.29	14.28	2.26	10.22	1.86	0	8.72	5.71	0.90	43.1
R-672	5.82	5.98	28.00	27.95	11.36	1.45	4.12	0.60	0	7.32	6.20	1.28	31.5
R-674	12.92	10.42	31.05	18.55	4.36	1.77	4.83	2.25	0	7.63	4.46	1.87	25.3
R-685	8.20	2.07	23.86	26.40	13.59	2.94	5.15	1.28	0	9.03	6.26	1.04	38.3
R-907	9.06	8.17	32.57	18.38	6.52	2.36	4.91	2.04	0	8.26	5.05	2.82	29.1
MS-42	5.12	4.56	22.08	29.81	12.45	1.85	8.31	1.42	0	7.81	5.71	0.99	37.6
MS-335	5.98	5.15	23.01	23.57	14.27	2.57	6.66	1.38	0	9.48	6.89	1.18	41.3
MS-643	6.96	4.92	24.11	23.87	15.08	1.70	5.71	0.74	0	9.07	6.81	1.16	39.1
MS-983	3.98	3.55	21.15	26.36	15.14	2.33	10.03	1.77	0	9.01	5.65	0.92	43.9
Avg.	6.77	5.40	25.80	24.43	12.41	2.23	5.65	1.25	0	8.52	6.20	1.39	36.3
Std.Dev	σ2.15	σ1.96	σ3.70	σ3.31	σ3.40	σ.53	σ1.96	σ.54	0	σ.69	σ.83	σ.57	σ5.1

Cole Mtn. basalt

Sample	Qtz.	Orth.	Albite	Anorth.	Diop.	Hed.	Ens.	Ferr.	Oliv.	Mag.	Ilm.	Apat.	C.I.*
R-561	16.36	4.97	27.92	26.32	2.98	0.98	7.83	2.97	0	6.29	2.71	0.76	23.8
R-614	11.30	4.62	32.57	25.06	5.11	1.28	6.84	1.97	0	6.63	4.00	0.71	25.8
Avg.	13.83	4.80	30.25	25.69	4.05	1.13	7.34	2.47	0	6.46	3.36	0.74	24.8

*Normative color index = Ol+Opx+Cpx+Mt+Il+Hm

APPENDIX 6 (cont.)

COLUMBIA RIVER BASALT GROUP												
Grande Ronde Basalt (Low MgO-Low TiO_2)												
Sample	Qtz.	Orth.	Albite	Anorth.	Diop.	Hed.	Ens.	Ferr.	Oliv.	Mag.	Ilm.	Apat.
R-38	11.69	9.59	20.81	24.89	2.31	3.01	8.07	12.07	0	2.89	3.98	0.78
R-48	12.01	9.83	20.34	25.24	1.82	2.44	8.19	12.61	0	2.89	3.91	0.76
R-51	14.28	10.25	20.30	25.19	0.00	0.00	8.11	13.73	0	2.89	4.12	0.76
R-98	10.88	9.89	22.08	23.74	2.92	3.90	7.56	11.58	0	2.89	3.93	0.73
R-242	10.77	9.53	20.56	24.48	2.24	3.18	8.10	13.21	0	2.89	4.31	0.80
R-597	14.69	9.24	25.63	23.87	4.04	3.67	5.91	10.57	0	2.89	3.97	0.88
R-917	14.88	9.30	23.56	24.21	3.97	3.32	5.01	11.33	0	2.89	4.08	1.04
Avg.	12.74	9.66	22.33	22.64	3.65	2.79	7.14	11.84	0	2.89	4.03	0.82
Grande Ronde Basalt (Low MgO-High TiO_2)												
R-2	10.54	8.71	20.64	25.23	2.20	3.14	8.04	13.18	0	2.89	4.59	0.95
Grande Ronde Basalt (High MgO)												
R-71	8.37	5.98	20.22	26.79	4.75	4.97	9.82	11.80	0	2.89	3.80	0.73
R-131	9.52	5.98	17.77	28.23	4.92	5.11	9.69	11.56	0	2.89	3.76	0.69
R-523	9.97	7.40	17.51	28.63	5.51	4.45	9.77	9.04	0	2.89	4.19	0.73
Avg.	9.29	6.45	18.50	27.88	5.06	4.84	9.76	10.80	0	2.89	3.92	0.72
Frenchman Springs Mbr.												
R-76	4.58	4.74	19.71	25.20	6.08	6.98	9.66	12.72	0	2.89	6.13	1.44
Pomona Mbr.												
R-20	2.34	2.01	16.75	32.44	11.42	5.86	13.58	8.00	0	2.89	3.89	0.95
R-357	4.40	0.77	13.96	34.68	9.71	5.33	14.60	9.19	0	2.89	3.74	0.88
R-297+	5.59	1.78	19.03	32.98	11.66	2.30	11.36	2.57	0	8.19	3.93	0.59
R-310+	2.65	2.25	22.00	33.17	11.28	2.36	12.77	3.07	0	7.18	2.88	0.64
R-617+	0.00	2.07	16.58	32.50	8.90	1.46	22.47	4.22	0.18	8.31	3.02	0.45
Avg.	3.00	1.78	17.66	33.15	10.60	3.46	14.96	5.41	0.04	5.89	3.50	0.70

+Gabbroic texture with norm calculated using "international std." chemical analy.

APPENDIX 7: POLARITY AND LOCATION OF BASALT SAMPLES

TILLAMOOK VOLCANICS

Sample No.	Polarity	Location	Exp
(X) R-339	N	SW¼ SW¼ sec. 8, T3N, R9W	q
R-340	N	SE¼ SW¼ sec. 8, T3N, R9W	r
(X) R-362	N	SE¼ SE¼ sec. 8, T3N, R9W	q
(X) R-366	N	NW¼ SW¼ sec. 10, T3N, R9W	r
(P) R-370	N	NE¼ SW¼ sec. 10, T3N, R9W	r
(P) R-375	N	NE¼ NW¼ sec. 11, T3N, R9W	r
R-386	N	NE¼ NE¼ sec. 11, T3N, R9W	r
R-387	N	NW¼ NW¼ sec. 12, T3N, R9W	r
(P) R-392	-	SE¼ SE¼ sec. 1, T3N, R9W	r
(X) R-395	-	SW¼ SW¼ sec. 8, T3N, R9W	r
(X) R-398	N	SW¼ SE¼ sec. 9, T3N, R9W	q
(P) R-409	-	SW¼ SE¼ sec. 6, T3N, R8W	r
(X) R-418	N	NW¼ NW¼ sec. 4, T3N, R8W	r
R-420	N	SW¼ SW¼ sec. 4, T3N, R8W	r
(P) R-423	N	SE¼ SW¼ sec. 4, T3N, R8W	r
(X) R-424*	-	NE¼ SE¼ sec. 5, T3N, R8W	r
(P) R-427	N	NE¼ NE¼ sec. 6, T3N, R8W	r
(X) R-626	-	NW¼ SW¼ sec. 28, T4N, R8W	s
R-641	R	NE¼ SW¼ sec. 10, T3N, R8W	r
R-644	R	SE¼ SE¼ sec. 9, T3N, R8W	r
(X) R-645a	R?	SW¼ SE¼ sec. 9, T3N, R8W	r
(P) R-645b	R	SW¼ SE¼ sec. 9, T3N, R8W	r
(P) R-646	N	SE¼ SW¼ sec. 9, T3N, R9W	r
R-650	R	NE¼ NE¼ sec. 17, T3N, R8W	r
(P) R-651	R	NW¼ NE¼ sec. 17, T3N, R8W	r
(X) R-667	N	SE¼ SW¼ sec. 7, T3N, R8W	r
(X) R-672	N	SE¼ SW¼ sec. 7, T3N, R8W	r
(P) R-673	N	SE¼ SW¼ sec. 7, T3N, R8W	r
(X) R-674	N	NE¼ SW¼ sec. 7, T3N, R8W	r
R-677	N	NE¼ SW¼ sec. 7, T3N, R8W	r
R-684a	N	NE¼ SW¼ sec. 7, T3N, R8W	r
(X) R-685ab	N	SW¼ NE¼ sec. 7, T3N, R8W	r
R-685c	R	NE¼ NW¼ sec. 7, T3N, R8W	r
(X) R-907	-	NW¼ SW¼ sec. 10, T3N, R9W	r
MS-7	R	NE¼ NE¼ sec. 12, T3N, R9W	r
MS-12	N	NE¼ NE¼ sec. 12, T3N, R9W	r
(X) MS-42	R	NE¼ NE¼ sec. 12, T3N, R9W	r
MS-74	N	NE¼ NE¼ sec. 12, T3N, R9W	r
(P) MS-85	-	NE¼ NE¼ sec. 12, T3N, R9W	r
MS-194	N	NE¼ NE¼ sec. 12, T3N, R9W	r
(P) MS-250	R?	SW¼ SW¼ sec. 6, T3N, R8W	r
(X) MS-335	-	SW¼ SW¼ sec. 6, T3N, R8W	r
MS-490	N	SW¼ SW¼ sec. 6, T3N, R8W	r
MS-580	R?	SW¼ SW¼ sec. 6, T3N, R8W	r
MS-592,604	N	SW¼ SW¼ sec. 6, T3N, R8W	r
(X) MS-643	N?	SW¼ SW¼ sec. 6, T3N, R8W	r
(P) MS-662	-	SW¼ SW¼ sec. 6, T3N, R8W	r
MS-684d	R	SW¼ SW¼ sec. 6, T3N, R8W	r
MS-808	N?	SE¼ SW¼ sec. 6, T3N, R8W	r
(X) MS-983	R	SE¼ SW¼ sec. 6, T3N, R8W	r
MS-1013	N	SE¼ SW¼ sec. 6, T3N, R8W	r

APPENDIX 7 (cont.)

Sample No.	Polarity	Location	Exp
POMONA MEMBER			
(X) R-20	R?	NE¼ SW¼ sec. 28, T5N, R8W	r
(X) R-297b**	-	SW¼ NE¼ sec. 32, T4N, R9W	q
(X) R-310**	N?	SW¼ SW¼ sec. 8, T3N, R9W	r
R-314b**	R?	SW¼ SE¼ sec. 29, T4N, R9W	r
(X) R-357	R	NW¼ SE¼ sec. 9, T3N, R9W	q
(X) R-617**	-	NW¼ NE¼ sec. 29, T4N, R8W	s
FRENCHMAN SPRINGS MEMBER			
(X) R-76	N?	SW¼ NW¼ sec. 30, T5N, R8W	q
R-68	N?	NE¼ NW¼ sec. 30, T5N, R8W	r
COLE MOUNTAIN BASALT			
(P) R-159	-	NE¼ SW¼ sec. 11, T4N, R9W	r
R-162	N	SW¼ SW¼ sec. 12, T4N, R9W	r
(P) R-192	N	NW¼ SW¼ sec. 4, T4N, R8W	r
(P) R-229	N	SE¼ NE¼ sec. 18, T4N, R8W	r
(X) R-251	N	NE¼ SE¼ sec. 8, T4N, R8W	q
R-259a	N	NE¼ NW¼ sec. 20, T4N, R8W	r
(P) R-470	N?	NW¼ SE¼ sec. 25, T4N, R9W	r
(P) R-546	N?	NE¼ SW¼ sec. 19, T4N, R8W	s
(X) R-561	N	SW¼ SE¼ sec. 17, T4N, R8W	s
(P) R-599	-	NE¼ NW¼ sec. 20, T4N, R8W	s
R-605	R?	SE¼ NW¼ sec. 20, T4N, R8W	s
(P) R-610	-	SE¼ NW¼ sec. 20, T4N, R8W	s
(X) R-614	N	NW¼ NE¼ sec. 18, T4N, R8W	r
(P) Q-180	-	SE¼ NE¼ sec. 25, T4N, R9W	s
(X) Q-850	N	SE¼ NE¼ sec. 25, T4N, R9W	s
(P) QQ-257	-	NW¼ SE¼ sec. 19, T4N, R8W	q
WELL SAMPLES- CZ 11-28		NW¼ sec. 28, T5N, R9W	
(C) CZ-1		1940-1950'	
(X) CZ-2		1950-1960'	
(C) CZ-3		3050-3060'	
(X) CZ-4		3070-3080'	
(P) 2116-18b		3250-3270'	
(P) 2116-30a		4500-4520'	
(P) 2116-30b		4500-4520'	
(P) 2116-32		4800-4820'	
(C) CZ-5		4900-4910'	
(P) 2116-37a		5400-5420'	
(P) 2116-37b		5400-5420'	
(P) 2116-37c		5400-5420'	

APPENDIX 7: (cont.)

GRANDE RONDE BASALT

	Sample No.	Polarity	Location	Exp
(X)	R-2	R?	SE¼ NE¼ sec. 19, T5N, R8W	r
	R-4	R?	NW¼ NW¼ sec. 20, T5N, R8W	r
	R-8a	N	NE¼ NW¼ sec. 20, T5N, R8W	r
	R-10	N	SW¼ NE¼ sec. 19, T5N, R8W	r
(P)	R-16	-	SW¼ NE¼ sec. 24, T5N, R9W	r
	R-26	N	SW¼ SE¼ sec. 29, T5N, R8W	q
(X)	R-26a	N	SW¼ SE¼ sec. 29, T5N, R8W	q
	R-31	N	NW¼ NE¼ sec. 32, T5N, R8W	r
	R-33	N	SE¼ SW¼ sec. 29, T5N, R8W	r
	R-36	N	SE¼ SE¼ sec. 30, T5N, R8W	q
(X)	R-38	N?	NW¼ SW¼ sec. 21, T5N, R8W	q
(X)	R-40bc	N	NE¼ SW¼ sec. 19, T5N, R8W	q
(X)	R-45	N	SW¼ SW¼ sec. 20, T5N, R8W	r
	R-47a	N	SW¼ NE¼ sec. 20, T5N, R8W	r
(X)	R-51	N	SE¼ NW¼ sec. 19, T5N, R8W	r
(P)	R-56	N	NW¼ SW¼ sec. 19, T5N, R8W	r
(X)	R-60	N	NW¼ NW¼ sec. 30, T5N, R8W	r
(X)	R-71	N	NW¼ NE¼ sec. 30, T5N, R8W	r
	R-72	N	SE¼ NW¼ sec. 30, T5N, R8W	r
(X)	R-78	N	NW¼ SW¼ sec. 30, T5N, R8W	r
(X)	R-98	N	NE¼ NE¼ sec. 35, T5N, R9W	q
	R-117	N?	SW¼ SE¼ sec. 36, T5N, R9W	r
	R-123	N	NW¼ SE¼ sec. 33, T5N, R8W	r
	R-127	N?	SW¼ SE¼ sec. 33, T5N, R8W	r
(X)	R-131	N	NW¼ NE¼ sec. 31, T5N, R8W	q
	R-176	N	NE¼ NE¼ sec. 4, T4N, R8W	r
(P)	R-297	N?	SW¼ NE¼ sec. 32, T4N, R9W	q
(X)	R-242	N	NW¼ NE¼ sec. 23, T4N, R9W	q
	R-351	N	SW¼ SE¼ sec. 33, T4N, R9W	r
(P)	R-448	N?	NE¼ NW¼ sec. 7, T3N, R9W	q
	R-504	N	SE¼ SE¼ sec. 36, T5N, R9W	r
	R-510	N	SW¼ NW¼ sec. 5, T4N, R8W	r
(P)	R-517	N	NE¼ NE¼ sec. 23, T5N, R9W	r
(P)	R-520	N	SW¼ NW¼ sec. 26, T4N, R9W	s
	R-522	N	NE¼ NW¼ sec. 26, T4N, R9W	s
(X)	R-523	N	NW¼ NE¼ sec. 26, T4N, R9W	s
(P)	R-581	N	NE¼ NW¼ sec. 32, T4N, R9W	s
(X)	R-597	N	NE¼ NW¼ sec. 32, T4N, R9W	q
(X)	R-911	N	NW¼ SW¼ sec. 27, T4N, R9W	q
	R-915	N	NE¼ NE¼ sec. 33, T4N, R9W	r
	R-916	N	SW¼ NE¼ sec. 33, T4N, R9W	r
(X)	R-917	N	SW¼ NE¼ sec. 33, T4N, R9W	r
	R-924	N	NE¼ NW¼ sec. 34, T4N, R9W	r
	R-1005	N	NE¼ NE¼ sec. 24, T5N, R9W	r

APPENDIX 7 (cont.)

Sample No.	Location
Cole Mountain basalt (Wash. Co.)	
(C) RCC1	NW¼ NW¼ sec. 2, T3N, R5W
Type area Goble Volcanics	
(X) RTG-1	SW¼ NE¼ sec. 12, T6N, R2W
(C) RTG-5	SW¼ NW¼ sec. 17, T6N, R1W
Grays River area Goble Volcanics	
(X) RG-1	SE¼ SW¼ sec. 9, T10N, R6W
(C) RG-2	SE¼ NE¼ sec. 8, T10N, R6W
(X) RG-3	SE¼ SE¼ sec. 23, T10N, R5W

**Microgabbroic texture, tentatively assigned to Pomona B.
(P) Petrography sample (C) Chemistry sample
(X) Petrography and chemistry sample
r=roadcut s=stream-cut q=quarry

APPENDIX 8: PETROGRAPHIC SUMMARY OF IGNEOUS ROCKS

Tillamook Volcanics

SAMPLE	TEXTURE[1]	PHENOCRYSTS					GROUNDMASS							Alt.	COMMENTS
		% Plag	An	Size‡	% Cpx†	Size‡	% Plag.	An	Size‡	% Cpx.†	Size‡	% Opaq.	Size‡		
R-339*	Glomeropor., Pilo.	6.3	60	1-3	2.6	.8-2	44.3	54	.1-.2	23.3	≪.1	18.3	<.1	VS	Zoned Plag. Phenos
R-362	Porphyritic, Pilo.	3	?	.6-1½	2	.6-2	47	61	.1-.3	30	≪.1	13	≪.1	VS	
R-366	Microporphy, Pilo.	1	?	.4-.7			53	60	.1-.3	35	≤.1	12	≤.1	VS	Zoned Plag. Phenos
R-370	Microporphy, Pilo.	6	64	.5-1	1	.5	42	63	.1	28	≪.1	18	≪.1	S	
R-375	Glomeropor., Faint.Pilo.	1	65	1	1	1	55	61	.1-.2	28	.1	12	.1	S	Leucox+Pigeon.Phenos
R-392	Porphyritic, Faint.Pilo.	2	68	1-2	1	.5	47	50	.1	30	<.1	15	<.1	VS	Zoned Plag. Phenos
R-395*	Glomeropor., Pilo.	8.7	65	.5-2	2.3	1	41.3	59	.1-.3	30.7	≤.1	12.7	≤.1	VS	Embayed Plag. Phenos
R-398*	Glomeropor., Pilo.	0.6	64	.1	0.2	2	56.7	53	.1-.5	24.7	.1-.2	14.3	.1	VS	Basaltic Hbl. Pheno.
R-409*	Porphyritic, Faint.Pilo.	7.3	66	1-2.5	1.6	1	39.7	57	.2	33.3	<.1	12.3	.1	VS	Embayed Plag. Phenos
R-418*	Glomeropor., Faint.Pilo.	13.6	66	.5-1½	0.6	.8	38.7	52	.1-.3	29.7	<.1	14.3	<.1	VS	Zoned Plag. Phenos
R-423*	Glomeropor., Pilo.	9.3	64	1-2	1.3	.5-.7	38.7	54	.1-.2	39.3	<.1	9.7	.1	VS	
R-424*	Microglomeropor., Pilo.	0.6	?	.6	0.3	.6	48.7	53	.1-.2	34.3	<.1	15.7	<.1	VS	Boulder in Roy Ck. Mbr.
R-427*	Porphyritic, Pilo.	0.6	?	1-3	0.3	1	52.3	55	.4-.5	35.7	.1-.2	11.7	.1	VS	Close to sub-ophitic
R-626	Aphyric, Pilo.						47	63	.1-.2	37	≤.1	16	.1	VS	Distinctly Pilotaxitic
R-645a	Glomeropor., Pilo	10	63	2	1	1	45	56	.1-.3	28	.1	15	<.1	S	Opaque Phenocryst
R-645b*	Porphyritic, Faint.Pilo.	17.3	61	1-4	1.7	1	32.7	60	.1-.3	28.7	≪.1	18.6	<.1	VS	Opaque Phenocryst
R-646*	Porphyritic, Pilo.	13.6	63	1-3	0.3	1	45.7	57	.1-.4	28.3	<.1	9.3	.1	S	Opaque Phenocryst
R-907*	Glomeropor., Pilo.	1.7	65	1-2	0.3	1	52.7	52	.1-.5	29.3	.1	12.6	.1	VS	Opaque Phenocryst
R-651	Porphyritic, Faint.Pilo.	15	67	2			40	60	.1-.3	30	<.1	10	<.1	M	Alt. Pheno. (Augite?)
R-667*	Glomeropor., Pilo.	6.3	62	1-7	5.3	1-7	29.7	55	.1-.2	38.3	<.1	18.7	.1	S	
R-672*	Glomeropor., Faint.Pilo.	20.3	66	1-4	2.3	1-2	37.7	56	.2-.5	27.3	<.1	11.7	.1	VS	Opaque Phenocryst
R-673	Glomeropor., Faint.Pilo.	22	68	1-3½	2	1-2	36	56	.1-.5	26	≤.1	10	.1	VS	Opaque Phenocryst
R-674*	Microporphy., Pilo.	.7	58	.5-1	0.3	1	62.3	50	.1-.3	22.3	.1	10.7	.1	S	
R-685*	Glomeropor., Faint.Pilo.	3.3	63	1-2½	0.7	1	49.3	61	.2-.4	30.7	.1-.3	11.7	.1	S	
MS42	Porphyritic, Faint.Pilo.	10	64	1-2	4	1-2	38	62	.1-.2	32	<.1	12	≤.1	M	
MS85	Glomeropor., Pilo.	10	67	1-2	3	.5-2	43	57	.1-.3	33	<.1	8	.1-.2	M	
MS250*	Glomeropor., Pilo.	13.3	63	2-2½	3.8	2	38	59	.1-.3	31	<.1	10.5	≤.1	S	
MS335*	Microglomero., Pilo.	0.5	66	.7	0.2	.6	47.3	63	.1-.4	33.7	<.1	12.3	<.1	S	
MS643*	Glomeropor., Pilo.	0.3	?	1	0.2	1	55.3	60	.1-.4	32.3	<.1	10.7	.1-.2	M	
MS662	Glomeropor., Pilo.	2	?	2	0.5	1	50	58	.2-.3	32	≤.1	13	.1-.2	S	
MS983	Porphyritic, Pilo.	0.5	?	1	3	.5-2	44	68	.1-.4	36	.1-.2	13	.1-.2	S	

[1]All samples have an intergranular groundmass. †Most clinopyroxene is augite. *Samples which were point-counted. ‡Size in mm.
Notes: Pilo.=Pilotaxitic, Alt.=Degree of alteration, VS=Very slight, S=Slight, and M=Moderate.

COLUMBIA RIVER BASALT GROUP

SAMPLE	UNIT	TEXTURE	PHENOCRYSTS Plag %	An	Size (mm)	Cpx %	Siz mm	GROUNDMASS Plag. %	An	Size (mm)	Cpx+ %	Size (mm)	Opaq. %	Size (mm)	Glass %	Oliv. %	ALT.	COMMENTS
R-20*	Pomona	Glomerop.,Interser.	1	64	3			49.3	62	.1-1	42.7	.2	5.3	.2	2.7	?	VS	
R-357*	Pomona	Microglom.,Interse.	4.5	62	1	.7	1½	40.5	60	.2-.3	25.2	.2	6.5	.2	21.5	?	S	
R-310*	Pomona?	Glomerop., Ophitic	18.3	64	1½-3			44.7	63	.3	22.3	.2-4	5.3	.2		8.3	VS	Unaltered Olivine
R-297	Pomona?	Ophitic						40	58	.2-.5	29	.1-3	4	.2		?	M-E	
R-617*	Pomona?	Ophitic						50.5	60	.5-1	21.7	.2-3	7.6	.5		6.3	S	Unaltered Olivine
R-76*	French.Sp.	Porphy., Subophit.	0.5	62	1-3			46.3	55	.2-.6	37.7	.1-.5	8.7	.2	3.0		S	Calcite alt.
R-71*	GR H-Mg	Intergranular						45.3	59	.1-1	43.7	.1-.3	10.7	.1-.5			VS	
R-131*	GR H-Mg	Intergranular						46.5	57	.1-1	43.3	.1-.3	9.5	.1-.4			VS	
R-523*	GR H-Mg	Subophitic						49.5	59	.1-1	41.0	.1-1½	9.0	.1-.5			S	
R-38*	GR L-Mg	Hyalopilitic						32.7	56	.1-.2	22.3	.1	9.3		36.3		S	
R-45	GR L-Mg	Hyalopilitic						30	57	.1-.2	20	.1	3		45		VS	
R-51	GR L-Mg	Hyalopilitic						34	55	.1-.2			7	.1	55		S	
R-98*	GR L-Mg	Hyalopilitic						39.7	54	.2-.7	28.3	.1	8.3	.1	24.0		S	
R-242	GR L-Mg	Intersertal						55	57	.1-1	33	.1-1	8	.3	4		S	
R-597*	GR L-Mg	Hyalopilitic						33.0	53	.1-1	11.3	.1			54.7		S	
RR-40	GR L-Mg	Hyalopilitic						36	55	.1-.7	23	.1	4	.1	37.3		S	
R-2*	GR L-MgHT	Hyalopilitic						41.3	56	.1-.3	30.3	.1	8.7	.2	20.3		S	
R-16	GR L-Mg?	Intersertal						50	51	.2-1	36	.2	8	.2	6		S	
R-56	GR L-Mg?	Intersertal						44	60	.1-1	30	.2	8	.1	18		M	
R-297	GR L-Mg?	Hyalopilitic						42	54	.1-.2	21	.1	9	.1	30		S	
R-517	GR L-Mg?	Hyalopilitic	0.2	64	3			30	56	.1-.8	25	.1-.7			45		M	
R-520*	GR H-Mg?	Subophitic						54.3	55	.4-1½	37.3	.3-1½	8.3	.3-1			VS	
R-581*	GR L-Mg?	Subophitic						54.0	57	.1-.3	33.7	.1-.5	9.3	.2-.4			S	
RR-448	GR L-Mg?	Hyalopilitic						40	52	.1-.3	28	.1	7	.2	25		S	

+Most clinopyroxene is augite. *Samples which were point-counted.

Notes: Glomerop.=Glomeroporphyritic, Intere.=Intersertal, Alt.=Degree of alteration, VS=Very Slight, S=Slight, M=moderate.

APPENDIX 8: (cont.)

SAMPLE	TEXTURE	PHENOCRYSTS Plag. %	An	Size (mm)	Cpx+ %	Size (mm)	GROUNDMASS Plag %	An	Size (mm)	Cpx+ %	Size (mm)	Opaq. %	Size (mm)	Glass++ %	Vesic.1	Alter.	UNIT
R-159*	Glomeropor., Hyalopilitic	23.3	65	.5-1.5	2.3	.5-1	18.0	52	.1-.2	3.5	.1	5.5	.1-.3	28.5	17	VE	
R-192	Hyalopilitic						30	55	.2-.8	6	.2	4	.1	60	1	VE	
R-229	Hyalopilitic						60	54	.1-2	10	.2-1	5	.2	25		M	
R-251*	Porphyritic, Hyalopilitic	1.0	?	2			42.7	56	.4-.8	12.7	.2-.6	4.7	.2	37.2	3	S	COLE
R-470	Intersertal						45	57	.1-.8	15	.1-2	5	.2	20	15	M	
R-546	Intersertal						45	53	.1-.5	20	.1-1	2	.2	18	7	M	MOUNTAIN
R-561*	Glomeropor., Intersertal	20.5	58	1-2.5	2.5	1-2	35	?	.1	2	.1	8.5	.1-.5	23.0		S	
R-599	Glomeropor., Hyalopilitic	17	60	.5-2	4	.4-1½	8	?	.1			0.5	.1	71		S	
R-610*	Glomeropor., Hyalopilitic	3	69				41.5	56	.1-.3			2.3	.1	32.0		S	BASALT
R-614*	Intersertal						61.3	57	.3-2	8.7	.3-.5	5.3	.1	25.7		S	
Q180	Porphyritic, Hyalopilitic	15	62	.5-1	2	.4								73		VE	
Q850*	Glomeropor., Hyalopilitic	20.5	66	.5-2.5	4.5	.3-3	21.3	53	.1	4.3	.1	6.5	.1	29.5		S	
QQ257	Hyalopilitic						40	55	.3-.5	1	.1			59		S	
RTG-1	Porphy., Intergran., F.Pilo	29	63	1-3	4	1	36	?	.2	26	.1	7	.1			VS	TYP.GOBLE V.
RG-1	Glomerop., Intergran., Pilo.	10	64	1.5	3	1	43	?	.1-.3	32	.1	13	.1			VS	GRAYS RIVER
RG-3	Porphyritic, Interg., Pilo.	2	60	1.5			49	56	.1-.3	37	.1	12	.1			VS	GOBLE VOL.
CZ-2	Intersertal						55	57	.1-.5	33	.2	6	.1-.4	4		VS	GR Low-Mg (WELL SAMPLES@)
CZ-4	Porphyritic, Intersertal	0.5	?	2			52	54	.2-.5	30	.2	11	.2	7		VS	Grande R.
211618	Intersertal						55	?	.2-.7	31	.2	10	.2	4		S	High-Mg
211630a	Glomerop., Intergran., Pilo.	2	60	2			53	?	.1	33	.1	12	.1			S	
211630b	Glomerop., Intergran., Pilo.	2	?	2			50	56?	.1	35	.1	13	.1			S	Tillamook
211632	Porphy., Intergran., Pilo.	3	?	1			50	?	.1	35	.1	12	.1			S	
211637a	Intergran., Pilotaxitic						55	?	.1	35	.1	10	.1			S	Volcanics
211637b	Glomerop., Intergran., Pilo.	2	?	1			53	?	.2-.3	36	.1	9	.1			S	

+Most clinopyroxene is augite. ++Percent Glass, may also include unidentifiable groundmass and partially devitrified glass.
1Percent vesicles and amygdules. * Samples which were point counted. @Small samples, difficult to estimate percentages.
Notes: Glomerop.=Glomeroporphyritic, Pilo.=Pilotaxitic. Interg.=Intergranular, Alter.=Degree of alteration. VS=Very slight
S=Slight, M=Moderate, and VE=Very Extensive.

APPENDIX 9: PETROGRAPHIC SUMMARY OF SEDIMENTARY ROCKS

SAMPLE NUMBER	VRF(BAS.+AND.)% Total % VRF	Inter-gran.	Inter-sertal	Vesstc-glassy	%PLAG.[1]	%CPX.[2]	%OPAQ.	%HBL.	%QUARTZ	%FOSSIL.	%POR.[3]	%CALCITE CEMENT	%CLAY[4]	ROUNDING	SAND SIZE	COMMENTS	UNIT
R-338a	63	45	3	15	8	1	<1		<1	1	2	12	10	Subrounded+	C	Cont. foram and mollusk frag.	
R-339	67	44	5	20	2	1	<1				2	15	10	Subrounded+	VC	Common pyrite	ROY
R-365	80	63	7	10	1	.5			<1		2		16	Subrounded+	C	Plastically def. VRF	
R-398b	78	65	5	8	4	1	<1		<1	1	2	3	12	Subrounded+	C	Fractured plag.,mollusk frag.	
R-625*	82	61	15	5½	1½	½	<1				3	1	15	Subrounded+	C	Well develop. radial chlorite cement	CREEK
R-287	77	50	12	15	5	1	<1				2	1	15	Subrounded+	F	Contains occ. large (4mm) scoria clasts	
R-289	78	52	11	15	8		<1				2	1	12	Subrounded+	F	Common clay-filled burrows	
R-398a	62++				25	1	<1				2		8	Subrounded+	F	Well compacted	MEMBER
R-443	83++				5	1	<1				2		10	Subrounded+	F	VRF are extensively altered	
R-630*	72	57½		6	10½	½	<1				3		15	Subrounded+	F	Well developed radial chlorite cement.	
R-637	79	51	18	10	8	1	<1				3		11	Subrounded+	F		
R-792	80	60	10	10	5	1	<1				2		12	Subrounded+	F	Well developed radial chlorite cement.	
R-604	58	31	15	12	25						2		17	Subrounded+	F	Distinctly laminated.	SWEET
R-616	75	55	10	10	4		<1				2	18	1	Subrounded+	F	Minor pyrite	HOME CK.
R-221	62++		2	60	3	1	<1				1		32	Subangular	VF	VRF are extensively altered to clay.	MEMBER
RQ-350	77		17	60	3	<1		<1			1	20		Subang-Ang.	C	Norm. graded, contains glauconite.	COLF MTN. [illegible]

[1]Plagioclase ranges from An 45 to An 65. [2]Most or all clinopyroxene is augite. [3]Porosity is difficult to estimate due to very small pore spaces.[4]Much of the clay is chloritic cement, surficial weathering makes it difficult to distinguish detrital clays from diagenetic clays. *Samples which have been point-counted. +VRF are subrounded to rounded and plagioclase is subangular. ++VRF are too altered to determine original textures.

SAMPLE NUMBER	% QUARTZ	%POLYCRYST. QUARTZ	% PLAGIOCLASE	%ORTHOCLASE	%MICROCLINE	% CHERT	% BIOTITE	%MUSCOVITE	%HORNBLENDE	%HEAVY MIN.	% MRF	% VRF	TYPE OF VRF	%GLAUCONITE	% POROSITY	% CALCITE CEMENT	% MATRIX	ROUNDNESS	SAND SIZE	COMMENTS	UNIT
R-145*	13	1	P	P	P	P	P	P	P	P	?	?			1		60	SA-SR	VF	Extensively bioturbated.	JEWELL MEMBER of the KEASEY FORMATION
R-179*	36.3	2.5	2.8	12.5	?	1	6.5	3.5	1	1	3.5	4.0	I		4.0		15.5	SA-SR	VF	Occ. well-rounded qtz.	
R-180	34	3	4	11	1	1	6	3	1	1	8	3	I		4		15	SA-SR	VF		
R-181	37	2	4	10	1		5	3	1	2	8	3	I		3		13	SA-SR	F		
R-353	38	8	3	11	1		3	2		1	9	5	I		4		11	SA-SR	F	Occ. burrows.	
R-377	34	7	3	10	1		7	2	1	1	7	3	I		5		15	SA-SR	F	Feldspars alt. to clay.	
R-382	36	5	3	6	1		6	2		1	8	3	I		4		18	SA-SR	F		
R-532	33	2	5	5	1		7	3		1	9	3	I		3		16	SA-SR	F		
R-623	37.3	4.5	3.7	8.2	1		4.5	0.5	1	2	14.0	3.5	I		3		16.7	SA-SR	F	Rare well-rounded qtz.	
R-323a*	8.0	1.5	6.6	2.3			1.3	1	1	1	2.0	3.5	IM	9	2	22½	27.5	SA	VF	12.5% glass shards. Also has foram. and diatom frag.	PITTSBURG BLUFF FORMATION
R-323b	9	1	5	2			2	1	1	1	3	5	IM	9	2	2	55	SA	VF	Common alt. glass shards.	
R-451	2	1	8									60+	IM	1	2		17	SR	F	Rare siltstone clasts(1 cm)	SMUGGLER COVE FORMATION
R-580	30	9	7	4		1	1	2		1	5	3	I		3		18	SA-SR	F	Baked. Common alt. clasts.	
R-758*	1	1	5.5	?			1	1	1	1		4	IM	47	1	9	29	A-R	F	Rare carbonaceous debris.	
R-872			3									35	I				60	SR	S	VRF have felted texture	
R-311*	29.5	8.5	6.2	17.5	1	1	1	1.5		1	4	6	MS		2	15	5	SR-SA	F	Also 1% each myrmekite,pyrite, granitic RF,perthite,organics	SMUGGLER COVE FORMATION BALLPARK UNIT
R-925*	27.3	5.0	4.2	5.5	1	1	1	1		1	3	4	I		4		31	SR-SA	VF	Common organic debris.	
R-928	29	3	4	9	1		1	1		1	3	4	IS		5		35	SR-SA	VF	Occ. organic debris.	
R-929	7		1	1				1									90	SA	S	Baked, Lam., has porphyroblasts	
R-930*	25.5	8.2	4.0	8.3	1		1	3		1	4	4	I		7		18	SA-SR	M	7% siltstone rip-up clasts.	
R-662*	35.7	4.7	5.7	12.7	1	1	2.5	2		1	1.5	7.3	I		3		25	SA-SR	F	Myrmekite, com. iron oxide.	ANGORA PK MEMBER
R-589	35	5	5	11	1		2	2		1	4	6	I		4		25	SA-SR	VF	Common iron oxide.	

*Samples which have been point-counted +Highly altered volcanic rock fragments (VRF).
Notes: A=Angular, SA=Subangular, SR=Subrounded, R=Rounded, S=Silt-sized.

APPENDIX 10: HEAVY MINERAL ANALYSES

UNIT	ROY CREEK MEMBER										SWEET HOME CK.M				JEWELL MEMBER OF KEASEY FORMATION													
SAMPLE	289		339		365		398		419		364	620	628	629	137		145		179		181		353		377		382	
MINERAL	4ø	3ø	4ø	3ø	4ø	3ø	4ø	3ø	4ø	3ø	*	*	*	*	4ø	3ø	4ø	3ø	4ø	3ø	4ø	3ø	4ø	3ø	4ø	3ø	4ø	3ø
AMPHIBOLES																												
Hornblende		VR															R	R	C	O	A	R	C	R	O	O	O	
Lamprobolite																						VR						
Actinolite																												
Tremolite															VR				R		R	VR	R		VR			
PYROXENES																												
Hypersthene																	VR		VR	R			VR					VR
Augite	O	O	A	C	A	A	A	C	VA	A		O	R	VR			VR		VR			VR		VR				
Diopside																							R					
MICAS																												
Biotite	VR	VR									O	C	R		VR	R	O	O	R	A	R	A	O	O	A	VA	O	C
Muscovite	R										O	VR	VR		R	R		R		O	R	O	O	O	A	A	O	C
Chloritic	R			R	R	R		R	R	R			R			R		O		O		O		R			R	R
EPIDOTE																												
Undiff.	VR	VR										R	VR		A	A	A	C	R	R		O	O		A	R	A	R
Clinozoisite												VR			VR	R	R					VR	VR		VR	VR	R	
ZIRCON																												
Colorless	R										VR	O	VR		C	R	O	O	C	O	O	O	C	R	R		O	
Pink	VR														VR		VR		VR		VR		VR	VR			R	
Orange																	VR											
GARNET																												
Colorless	VR						VR	VR				A	R	VR	O	O	O	O	O	O	A	O	O	O	R		C	O
Pink	VR											O							R	R			R				VR	VR
TOURMALINE																												
Blue																												
Brown-Grn.	R											VR			R	VR	R	VR	R	R	R	R	R	R	R		R	
KYANITE																					VR							
STAUROLITE															VR	VR	VR		VR		VR		VR				VR	
MONAZITR																				R				VR				
RUTILE												VR	VR		VR		VR				VR		VR				VR	
SPHENE												R	VR		R	R			VR	VR	R	R	VR	VR				
APATITE	VR				VR						R	C	O	VR					VR		VR							
ANDALUSITE																		VR			VR	VR	VR					
SILLIMANITE																				VR								
OPAQUES																												
Magnetite	C	O	C	A	C	C	A	A	C	A	O	'C	O	R	C	O	C	C	VC	VC	O	A	VC	C	O	O	C	VC
Hematite				R		R				O												C						
Ilmenite	VA	A	VC	A	A	A	A	A	A	A	R	O	R	R	O	O	C	C	VC	VC	O	A	VC	C	O	O	C	VC
Leucoxene	A	A	O	O	O	O	O	O	O	C	R	O	VR		O	O	C	O	R	R	R	R	O	A	R	VC	A	A
Pyrite	R	R	VA	VA	O	O	O	C	R	R	A	VC	VC	C														
Pyritiz.Forams											A	VC	A	A														
Pyritiz.Diatoms											A	VC	A	A														
%HEAVY MINERAL	.3	.2	8	2	.5	1	5	6	6	9	<.1	<.1	<.1	<.1	.8	.2	.2	.2	.4	.2	.6	.2	2	.8	1	.7	.3	.2

VA(Very Abundant)>45%, A(Abundant) 25-45%; VC(Very Common) 15-25%, C(Common) 10-15%, O(Occasional) 5-10%, R(Rare) 1-5%, VR(Very Rare)<1%. *Combined 4.5ø and 4ø

APPENDIX 10 (cont.)

UNIT	PITT.BLUFF				S.C.f.		Ballpark unit S.C. fm.								ANGORA PK.				CZ11-28 WELL			
SAMPLE #	323		658c		451		311		312		928		930		589		662		CZ 06*	CZ 21*	CZ 24*	CZ 26*
MINERAL	4φ	3φ	4φ	3φ	4φ	3φ	4φ	3φ	4φ	3φ	4φ	3φ	4φ	3φ	4φ	3φ	4φ	3φ				
AMPHIBOLES																						
Hornblende	A		A	O	O						R			R	O		O					
Lamprobolite			R				VR						VR			VR						
Actinolite																						
Tremolite																						
PYROXENES																						
Hyperathene	VR			R		VR	VR	VR			C		C	R	VR	VR	VR					
Augite		R		O			VR	R				O	R	R							VR	VR
Diopside																						
MICAS																						
Biotite	C	VC		O		R	O	C	C	C		VR	VR	VR		VC		C		VR		
Muscovite	C	VC		R			C	VC	C	C			VR			C		C				
Chloritic																						
EPIDOTE																						
Undiff.	VR	VR	VR		C										C	O	VC	O				VR
Clinozoisite																	R					
ZIRCON																						
Colorless			VR		A		O	VR			VC		VC	C	C		O					
Pink											VR		VR		VR		VR					
Orange																						
GARNET																						
Colorless	VR	R	O	O	VR	R	VC	O	O		O	C	O	O	R		R	R				VR
Pink	VR						R															
TOURMALINE																						
Blue																						
Brown-Grn.			VR				VR	VR			VR				O		O					
KYANITE	VR																					
STAUROLITE			VR								VR				VR		VR					
MONAZITE				R										VR								
RUTILE	VR												VR		VR		VR				VR	
SPHENE		VR					R	VR	VR		VR			VR	VR			VR				
APATITE	O							VR	VR						VR					VR		VR
ANDALUSITE																						
SILLIMANITE																	VR					
OPAQUES																						
Magnetite	O	O	C	VC	VC	A	VC	C	O	C	VC	VC	VC	A	O	C	C	VC	O	O	O	O
Hematite																						
Ilmenite	O	O	C	VC	C	A	VC	VC	O	C	VC	VC	VC	A	O	C	VC	VC	O	O	O	O
Leucoxene			O	R	O	VC	VC	VC					C	C	A	VC	VC	A	O	O	O	O
Pyrite					C				VA	VA	C	C							VC	A	C	VC
Pyritiz.Forams																			A	A	A	A
Pyritiz.Diatoms																			A	A	A	A
%HEAVY MINERALS	.2	.2	.2	.2	.7	.2	.8	.3	1.3	4	.2	.2	.5	.4	.2	.2	2	.7	<.1	<.1	<.1	<.1

VA(Very abundant)>45%, A(Abundant) 25-45%, VC(Very Common) 15-25%, C(Common) 10-15%, O(Occasional) 5-10%, R(Rare) 1-5%, VR(Very Rare)<1%. *Combined 4.5φ and 4φ

APPENDIX 11: GRAIN SIZE ANALYSIS

Unit	Sample Number	Modal Class (ø)	Median(ø)	Mean (ø)	Std.Dev. (sorting)	Simple Sorting	Inclusive Graphic Skewness	Simple Skewness	Kurtosis	Coarsest 1% (mm)	Percent Smaller Than 4.5ø	VERBAL TERMS(1) Median	Mean	Sorting	Skewness	Kurtosis	Trask Sorting Coeff.
Hol.Beach	Yac1	-0.5	-0.98	-0.99	0.50	0.89	-0.03	-0.03	1.25	5.1		VC--	VC	W.-M.	N.Sym	Lepto	1.22
	339	-2.0	-2.54	-1.91	1.83	3.38	0.66	4.92	1.58	12.5		VF.Grav	VF.Grav	Poor	V.Pos	V.Lepto	1.83
ROY	339*	-2.0	-2.91	-2.59	0.80	1.39	0.61	1.67	1.01	12.5		VC	VF.Grav	Mod	V.Pos	Meso	1.47
	365	1.0	0.52	0.58	0.74	1.54	0.27	1.16	2.21	2.6		Coarse	Coarse	Mod	Pos	V.Lepto	1.22
CREEK	365*	1.0	0.52	0.58	0.57	0.98	0.09	0.03	1.40	2.6		Coarse	Coarse	Mod	N.Sym	Lepto	1.22
	398	-0.5	0.00	0.28	1.34	2.32	0.38	2.05	1.03	3.9		VC-C	Coarse	Poor	V.pos	Meso	1.73
FORMATION	419	1.0	0.74	0.94	1.29	2.12	0.14	1.07	1.05	2.2		Coarse	Coarse	Poor	Pos	Meso	1.77
	289	2.5	2.50	2.59	0.60	1.05	0.27	0.75	1.11	0.35		Fine	Fine	Mod	Pos	Mes-Lept	1.30
	630	3.0	2.36	2.28	1.12	1.69	-0.09	-0.28	0.84	0.80		Fine	Fine	Poor	N.Sym	Platy	1.77
	137	3.5	3.40	3.66	0.84	1.38	0.43	1.05	1.15	0.23	25.6	V.Fine	V.Fine	Mod	V.Pos	Lepto	
	137*	3.5	3.19	3.18	0.42	0.67	-0.11	-0.24	0.86	0.25		V.Fine	V.Fine	Well	Neg-Sym	Platy	
	145	3.5	3.80	3.82	0.77	1.28	0.11	0.49	0.95	0.19	42.3	V.Fine	V.Fine	Mod	Pos	Meso	
	145*	3.5	3.34	3.31	0.52	0.58	-0.17	-0.25	0.88	0.21		V.Fine	V.Fine	Mod	Neg	Platy	
JEWELL	179	3.0	3.32	3.51	0.87	1.60	0.33	0.97	0.95	0.25	25.1	V.Fine	V.Fine	Mod	V.Pos	Meso	
	179*	3.0	3.00	3.04	0.45	0.75	0.06	-0.01	0.93	0.27		F.-VF	V.Fine	Well	N.Sym	Meso	
	181	3.0	3.19	3.41	0.95	1.56	0.34	1.11	1.00	0.30	27.3	V.Fine	V.Fine	Mod	V.Pos	Meso	
	181*	3.0	2.95	2.97	0.50	0.84	0.00	-0.10	1.01	0.32		Fine	Fine	W.-M.	Sym	Meso	
MEMBER	353	2.5	2.84	2.48	1.16	1.78	0.35	1.36	0.73	0.37	23.4	Fine	V.Fine	Poor	V.Pos	Platy	
	353*	2.5	3.02	2.52	0.62	1.00	0.20	0.57	0.82	0.40		Fine	Fine	Mod	Pos	Platy	
	377	3.0	3.05	3.31	0.98	1.65	0.50	1.20	1.13	0.32	22.3	V.Fine	V.Fine	Mod	V.Pos	Lepto	
	377*	3.0	2.89	2.90	0.52	0.89	-0.05	-0.22	0.89	0.35		Fine	Fine	Mod	N.Sym	Meso	
	382	3.0	3.32	2.84	1.00	1.59	0.18	0.78	0.89	0.38	33.9	V.Fine	V.Fine	M-P	Pos	Platy	
	382*	3.0	3.40	2.86	0.58	0.90	0.01	-0.10	0.82	0.45		Fine	Fine	Mod	Sym	Platy	
	532	3.0	3.06	3.31	0.90	1.53	0.44	1.43	1.39	0.27	20.2	V.Fine	V.Fine	Mod	V.Pos	Lepto	
	532*	3.0	2.89	2.94	0.49	0.82	0.10	0.07	1.10	0.30		Fine	Fine	Well	N.Sym	Meso	
	623	3.0	3.12	3.50	1.37	1.59	0.44	1.18	1.03	0.31	26.8	V.Fine	V.Fine	Poor	V.Pos	Meso	
	623*	3.0	2.95	2.91	0.63	0.78	-0.17	-0.35	0.90	0.34		Fine	Fine	Mod	Neg	Meso	
PITTSBURG	323	4.0	3.83	3.85	0.50	0.91	0.14	0.39	1.07	0.19		V.Fine	V.Fine	Mod	Pos	Meso	
BLUFF FM.	658c	4.0	3.74	3.83	0.68	1.10	0.18	0.37	0.98	0.19		V.Fine	V.Fine	Mod	Pos	Meso	
SMUGGLER	451	3.0	3.05	3.23	0.89	1.53	0.41	1.45	1.05	0.24	21.2	V.Fine	V.Fine	Mod	Pos	Meso	
COVE fm.	451*	3.0	2.95	2.95	0.50	0.74	-0.02	-0.07	0.77	0.25		Fine	Fine	W.-M.	Sym	Platy	
	311	3.0	2.78	3.05	1.03	1.72	0.43	1.59	1.14	0.35	19.4	Fine	V.Fine	Poor	V.Pos	Lepto	
SMUGGLER	311*	3.0	2.64	2.67	0.63	0.99	0.06	0.08	0.93	0.38		Fine	Fine	Mod	N.Sym	Meso	
COVE	312a	4.0	3.64	3.68	0.78	1.33	0.41	0.17	1.07	0.23	31.6	V.Fine	V.Fine	Mod	V.Pos	Meso	
FORMATION	312a*	4.0	3.32	3.33	0.50	0.82	-0.70	-0.39	0.97	0.25		V.Fine	V.Fine	W.-M.	V.Neg	Meso	
BALLPARK	928	3.5	3.63	3.51	0.67	1.20	-0.32	-0.85	1.28	0.38	19.9	V.Fine	V.Fine	Mod	V.Neg	Lepto	
UNIT	928*	3.5	3.45	3.28	0.66	1.14	-2.25	-1.15	1.18	0.43		V.Fine	V.Fine	Mod	V.Neg	Lepto	
	930	1.5	1.85	2.44	1.59	2.37	0.48	2.16	1.01	0.80	21.3	Medium	Fine	Poor	V.Pos	Meso	
	930*	1.5	1.55	1.63	0.83	1.42	0.20	0.71	1.17	0.86		Medium	Medium	Mod	Pos	Lepto	
ANGORA	589	3.0	3.32	3.68	1.06	1.56	0.43	1.21	0.74	0.24	38.2	V.Fine	V.Fine	Poor	V.Pos	Platy	
PEAK	589*	3.0	3.00	3.00	0.46	0.75	0.02	-0.01	0.96	0.26		F.-VF	F.-VF	Well	N.Sym	Meso	
MEMBER	662	3.0	2.95	3.18	0.91	1.60	1.80	1.07	1.40	0.34	18.7	Fine	V.Fine	Mod	Pos	Lepto	
(Astoria F.)	662*	3.0	2.84	2.82	0.51	0.87	-0.06	-0.14	1.25	0.35		Fine	Fine	Mod	Neg	Lepto	

[1]For explanation of terms see Folk and Ward (1957).
*Recalculated without the less than 4.5ø size fraction.

APPENDIX 12: SANDSTONE SAMPLE LOCALITIES

Sample no.	Location		
R-248 (T)	SW¼ NW¼ sec. 16, T4N, R8W	r	ROY CREEK MEMBER of the HAMLET FORMATION
R-287 (T)	SW¼ SE¼ sec. 28, T4N, R8W	r	
R-289 (T,H,S)	NE¼ NE¼ sec. 33, T4N, R8W	r	
R-338a(T)	SW¼ SW¼ sec. 8, T3N, R8W	r	
R-339 (T,H,S)	SW¼ SW¼ sec. 8, T3N, R9W	r	
R-365 (T,H,S)	SE¼ SE¼ sec. 9, T3N, R9W	r	
R-398a(T,H,S)	SW¼ SE¼ sec. 9, T3N, R9W	q	
R-398b(T)	SW¼ SE¼ sec. 9, T3N, R9W	q	
R-419 (H,S)	NE¼ SE¼ sec. 5, T3N, R8W	r	
R-443 (T)	NE¼ SE¼ sec. 10, T3N, R9W	r	
R-625 (T,E)	SW¼ NW¼ sec. 28, T4N, R8W	s	
R-630 (T,S)	SE¼ SW¼ sec. 28, T4N, R8W	s	
R-637 (T)	SW¼ SE¼ sec. 28, T4N, R8W	r	
R-792 (T)	SE¼ SW¼ sec. 28, T4N, R8W	r	
R-221 (T)	NW¼ SW¼ sec. 8, T4N, R8W	r	SWEET HOME CREEK MEMBER
R-364 (H)	NE¼ SE¼ sec. 9, T3N, R9W	r	
R-604 (T)	SE¼ NW¼ sec. 20, T4N, R8W	s	
R-616 (T)	SE¼ SW¼ sec. 20, T4N, R8W	s	
R-620 (H,E)	NE¼ NE¼ sec. 29, T4N, R8W	s	
R-628 (H)	NW¼ SW¼ sec. 28, T4N, R8W	s	
R-629 (H)	SW¼ SW¼ sec. 28, T4N, R8W	s	
RQ-350(T)	SE¼ NE¼ sec. 25, T4N, R9W	s	COLE MTN. BASALT
R-137 (H,S)	NW¼ NW¼ sec. 11, T4N, R9W	r	JEWELL MEMBER of the KEASEY FORMATION
R-145 (T,H,S)	NW¼ NW¼ sec. 11, T4N, R9W	r	
R-179 (T,H,S,E)	SW¼ NE¼ sec. 4, T4N, R8W	r	
R-180 (T)	SW¼ NE¼ sec. 4, T4N, R8W	r	
R-181 (T,H,S)	SE¼ NE¼ sec. 5, T4N, R8W	r	
R-353 (T,H,S)	SW¼ SE¼ sec. 4, T3N, R9W	r	
R-377 (T,H,S)	SE¼ SE¼ sec. 2, T3N, R9W	r	
R-382 (T,H,S)	SW¼ SE¼ sec. 2, T3N, R9W	r	
R-532 (T,S)	NW¼ NW¼ sec. 25, T4N, R9W	s	
R-623 (T,S)	NW¼ SW¼ sec. 6, T4N, R8W	r	
R-323a(T)	SE¼ NW¼ sec. 21, T5N, R8W	r	PITTSBURG BLUFF FORMATION
R-323b(T,H,S)	SE¼ NW¼ sec. 21, T5N, R8W	r	
R-658c(H,S)	SE¼ NW¼ sec. 21, T5N, R8W	r	
R-451 (T,S,H)	NE¼ SW¼ sec. 32, T4N, R9W	r	SMUGGLER COVE FORMATION
R-580 (T)	NW¼ NW¼ sec. 32, T4N, R9W	r	
R-758 (T)	SW¼ NW¼ sec. 36, T5N, R9W	r	
R-872 (T)	NW¼ SW¼ sec. 26, T4N, R9W	r	
R-311 (T,H,S)	SE¼ NE¼ sec. 31, T4N, R9W	r	
R-312 (T,H,S)	NE¼ SE¼ sec. 31, T4N, R9W	r	
R-925 (T,E)	NW¼ NW¼ sec. 34, T4N, R9W	r	
R-928 (T,H,S)	SE¼ NE¼ sec. 33, T4N, R9W	r	
R-929 (T,E)	SE¼ NE¼ sec. 33, T4N, R9W	r	
R-930 (T,H,S)	SE¼ NE¼ sec. 33, T4N, R9W	r	
R-589c(T,H,S)	NW¼ NW¼ sec. 31, T4N, R9W	r	ANGORA PEAK MEMBER
R-662 (T,H,S)	SE¼ NW¼ sec. 31, T4N, R9W	r	

(T)=Thin-section (H)=Heavy mineral (S)=Sieve (E)=SEM

APPENDIX 13: SOURCE ROCK GEOCHEMISTRY AND POROSITY

Sample	Unit	%TOC	V.R. %RO	Kero. Type*	Kerogen Oil/Gas	Generation Rating	Stage of Diagenesis
R-628	S.H.C.	1.1	0.53	M	Gas	Non-Source	Early Gas
R-559+	S.H.C.	0.9	--	-	---	---	---
Q-570	L. Jew.	0.9	0.58	M	Wet Gas	Marginal	Early Gas
R-529	U. Jew.	0.9	0.72	A	Wet Gas	Marginal	E. Peak Gas
R-66++	B^{u}_{S}.C.fm.	-	0.57	-	---	---	---

* Visual kerogen type: A=Amorphous M=Mixed
\+ Extensively baked near a sill.
++Sample from "coaly" bed. Total organic carbon is approximately 70%. Analysis by Tim England, Univ. of British Columbia.

Sample	Unit	Total Porosity(%)	Effective Porosity(%)	Permeab.* (m.d.)
R-928	Ballpark unit S.C.f.	29.4	16.4	0.12
R-658c	Pittsburg Bluff F.	Too Friable for Analysis		

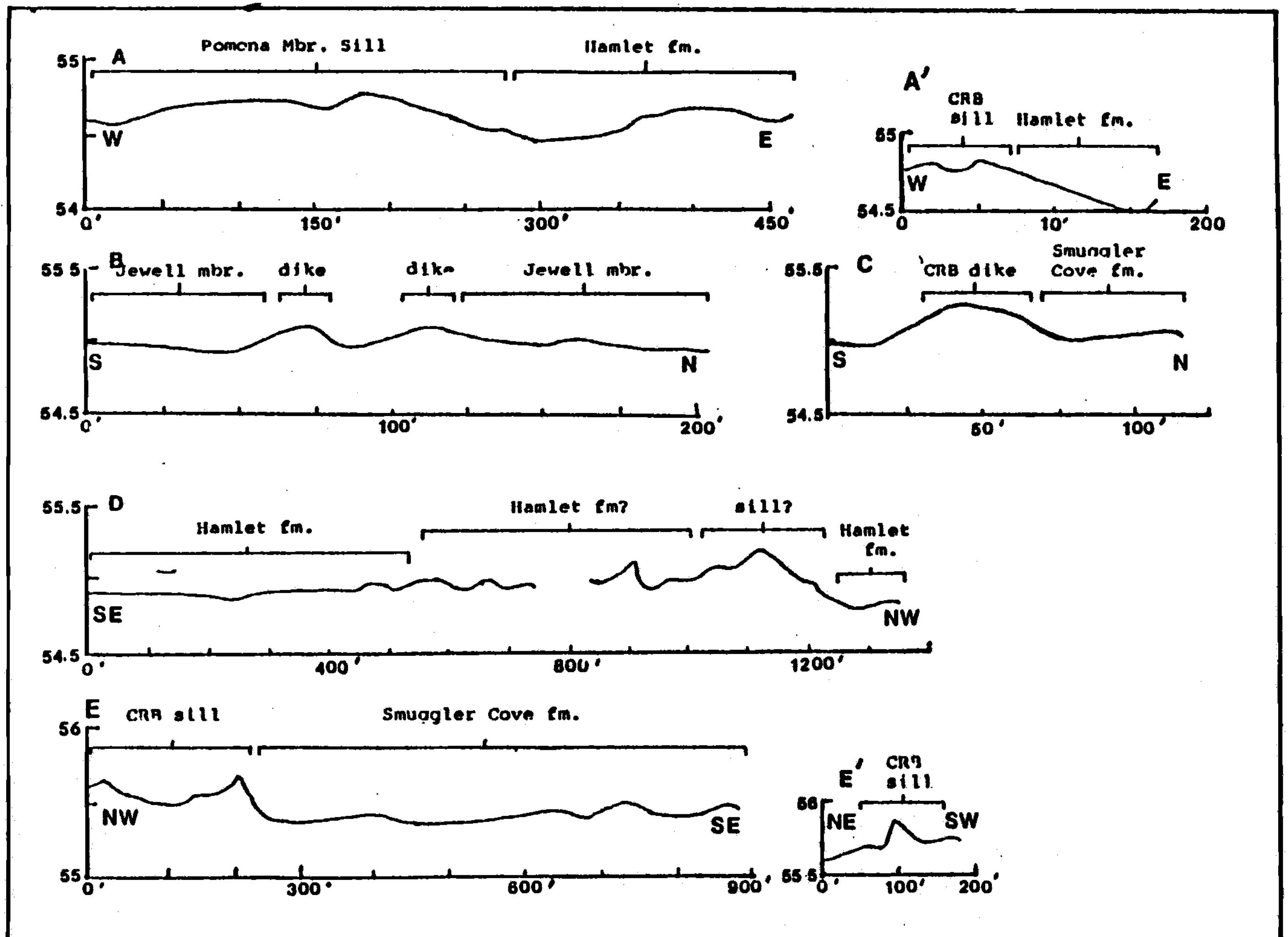
*Permeability was determined from effective porosity using experimentally derived equations for "shaley sandstones" (perm.= eff, por. X $0.03e^{.57}$). Therefore, the permeability value is very approximate.

Note: All of the above analyses were performed by AMOCO Production Company through Terry Mitchell and Jill Schlaefer unless otherwise noted.

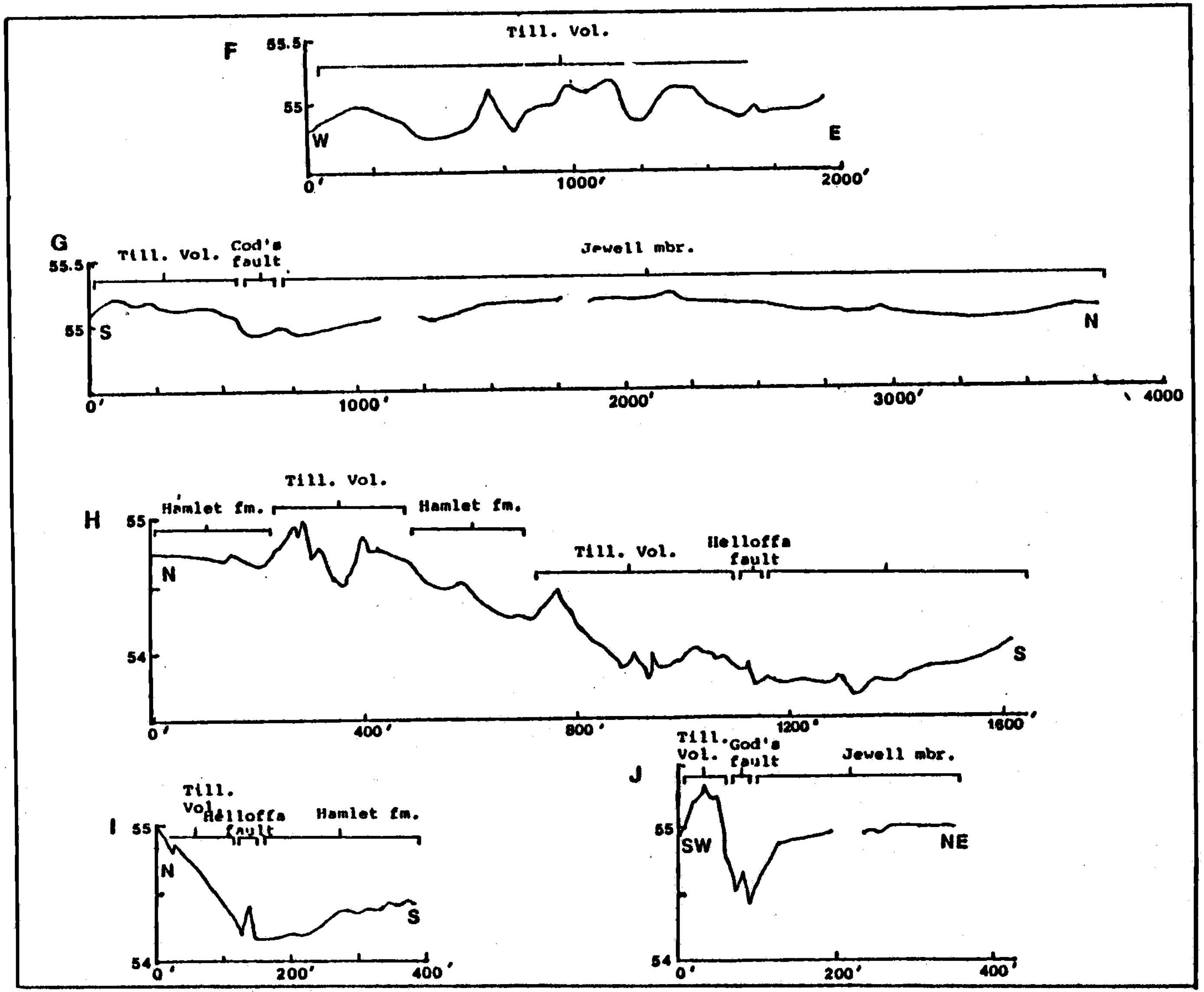
Location of Samples

Sample	Exposure	Location
R-66	road	NW¼ NW¼ sec. 30, T5N, R8W
R-529	stream	NW¼ NW¼ sec. 25, T4N, R9W
R-559	stream	SW¼ SE¼ sec. 17, T4N, R8W
Q-570	stream	SE¼ NE¼ sec. 25, T4N, R9W
R-628	stream	NE¼ SW¼ sec. 28, T4N, R8W
R-928	road	SE¼ NE¼ sec. 33, T4N, R9W
R-658c	road	SE¼ NW¼ sec. 21, T5N, R8W

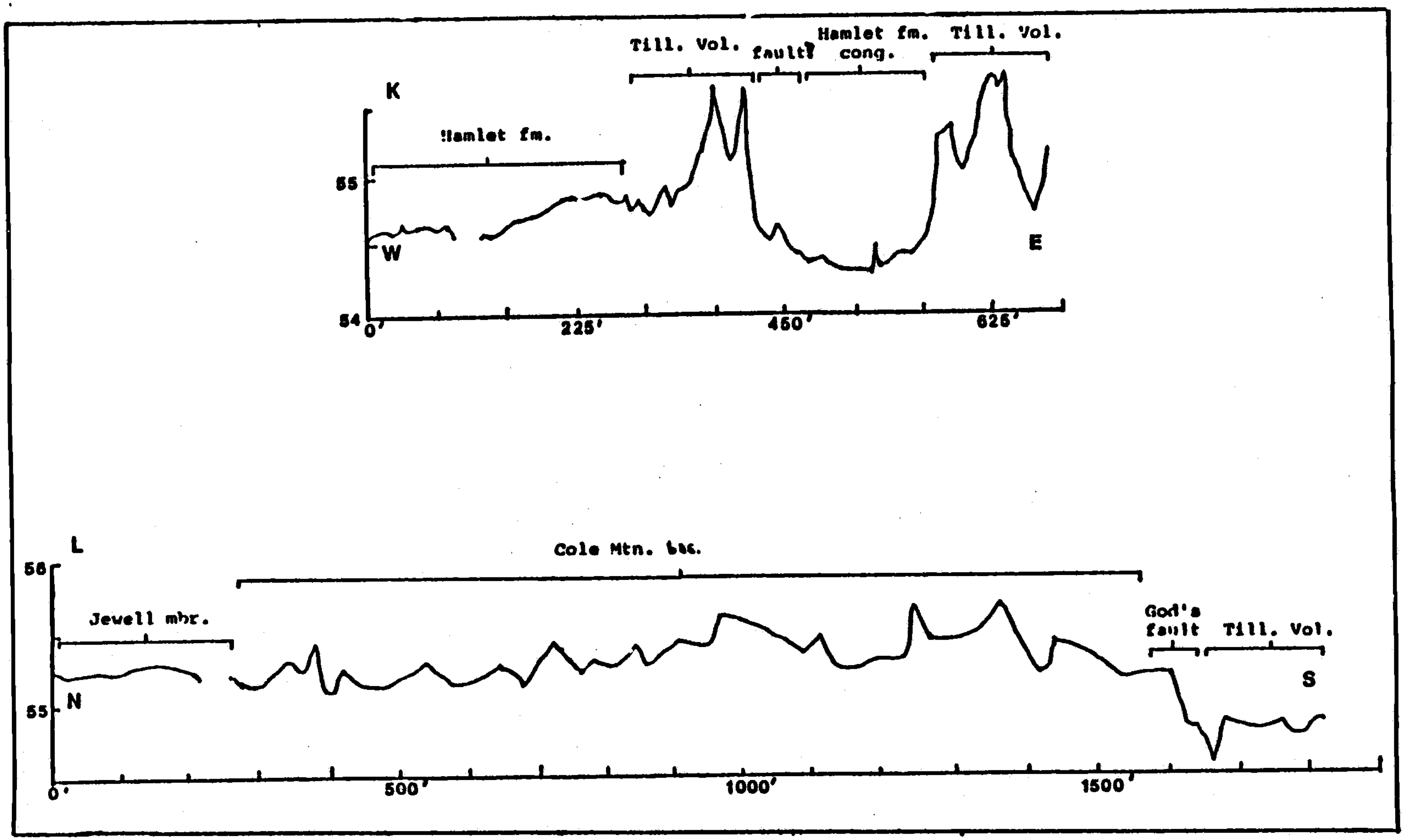
APPENDIX 14: PROTON PROCESSION MAGNETOMETER TRAVERSES



APPENDIX 14: (cont.)



APPENDIX 14: (cont.)



APPENDIX 15: LOCATION OF MAGNETOMETER TRAVERSES

Traverse	Location
A	Along God's Valley Road from NE¼ SE¼ sec. 7, T3N, R9W to SW¼ SW¼ sec. 8, T3N, R9W.
A'	Short traverse along logging road located in SW¼ SW¼ sec. 8, T3N, R9W.
B	Short traverse along logging road located in NW¼ SE¼ sec. 4, T3N, R9W.
C	Along logging road from NW¼ NE¼ sec. 4, T3N, R9W to SW¼ SE¼ sec. 33, T4N, R9W.
D	Along Rector Ridge Road from NE¼ SE¼ sec. 9, T3N. R9W to SE¼ NW¼ sec. 9, T3N, R9W.
E	Along logging road from SE¼ SW¼ sec. 33, T4N, R9W to SW¼ NW¼ sec. 33, T4N, R9W.
E'	Short traverse in NW¼ SW¼ sec. 33, T4N, R9W.
F	Along Crawford Road in NE¼ SW¼ sec. 2, T3N, R9W.
G	Along logging road from NW¼ SW¼ sec. 2, T3N, R9W to NW¼ NE¼ sec. 2, T3N, R9W.
H	Along logging road from NE¼ SE¼ sec. 9, T3N, R9W to NE¼ SW¼ sec. 10, T3N, R9W.
I	Short traverse along logging road in NE¼ SE¼ sec. 9, T3N, R9W.
J	Short traverse along logging road in NW¼ SW¼ sec. 22, T4N, R8W.
K	Along God's Valley Road from SW¼ SW¼ sec. 8, T3N, R9W to SE¼ SW¼ sec. 8, T3N, R9W.
L	Along logging road from NW¼ SW¼ sec. 32, T4N, R8W to NW¼ SE¼ sec. 32, T4N, R8W.

APPENDIX 16: LOCALITIES MENTIONED IN TEXT

Locality Number	Location	Outcrop
2	SE¼ NE¼ sec. 19, T5N, R8W	RC
4	NE¼ NW¼ sec. 20, T5N, R8W	RC
8a	NW¼ NW¼ sec. 20, T5N, R8W	RC
13	SW¼ NW¼ sec. 19, T5N, R8W	RC
20	NE¼ SW¼ sec. 28, T5N, R8W	RC
22	NW¼ SE¼ sec. 29, T5N, R8W	RC
29	SE¼ SE¼ sec. 29, T5N, R8W	RC
40a	SW¼ NE¼ sec. 19, T5N, R8W	RC
41	NW¼ SE¼ sec. 19, T5N, R8W	RC
46	NE¼ SW¼ sec. 20, T5N, R8W	RC
48	SE¼ NE¼ sec. 20, T5N, R8W	RC
66	NW¼ NW¼ sec. 30, T5N, R8W	RC
93	SW¼ SE¼ sec. 24, T5N, R9W	RC
98	NE¼ NE¼ sec. 35, T5N, R9W	Q
99	SW¼ NW¼ sec. 36, T5N, R9W	RC
116	NW¼ SW¼ sec. 36, T5N, R9W	RC
117	SW¼ SW¼ sec. 36, T5N, R9W	RC
124	SW¼ NE¼ sec. 33, T5N, R8W	RC
137	NW¼ NW¼ sec. 11, T4N, R9W	RC
142	SW¼ SE¼ sec. 36, T5N, R9W	RC
145	NW¼ NW¼ sec. 11, T4N, R9W	RC
163	SW¼ SW¼ sec. 12, T4N, R9W	RC
164	SW¼ SW¼ sec. 12, T4N, R9W	RC
165	SW¼ SW¼ sec. 12, T4N, R9W	RC
179	SW¼ NE¼ sec. 4, T4N, R8W	RC
180	SW¼ NE¼ sec. 4, T4N, R8W	RC
181	SE¼ NE¼ sec. 5, T4N, R8W	RC
191	NE¼ SW¼ sec. 4, T4N, R8W	RC
192	NW¼ SW¼ sec. 4, T4N, R8W	RC
193	SE¼ SE¼ sec. 5, T4N, R8W	RC
198	NE¼ NE¼ sec. 8, T4N, R8W	RC
206	SW¼ SW¼ sec. 12, T4N, R9W	RC
212	NW¼ SW¼ sec. 7, T4N, R8W	RC
219	SW¼ SW¼ sec. 6, T4N, R8W	RC
221	NW¼ SW¼ sec. 8, T4N, R8W	RC
226	NE¼ SW¼ sec. 8, T4N, R8W	RC
248	SW¼ NW½ sec. 16, T4N, R8W	RC
251	NE¼ SE¼ sec. 8, T4N, R8W	Q
257	NW¼ SE¼ sec. 19, T4N, R8W	RC
258	NW¼ SW¼ sec. 20, T4N, R8W	RC
287	SW¼ AE¼ sec. 28, T4N, R9W	RC
289	NE¼ NE¼ sec. 33, T4N, R9W	RC
297	SW¼ NE¼ sec. 32, T4N, R9W	Q
310	NW¼ SW¼ sec. 8, T3N, R9W	RC
311	SW½ NW¼ sec. 35, T4N, R9W	RC
313	NE¼ SE¼ sec. 31, T4N, R9W	RC

APPENDIX 16 (cont.)

Locality Number	Location	Outcrop
314	NE¼ NW¼ sec. 32, T4N, R9W	Q
320	SE¼ NW¼ sec. 21, T5N, R8W	RC
323a	SE¼ NW¼ sec. 21, T5N, R8W	RC
323b	SE¼ NW¼ sec. 21, T5N, R8W	RC
327	SE¼ SE¼ sec. 9, T4N, R8W	RC
328	NW¼ NE¼ sec. 16, T4N, R8W	RC
329	NW¼ NE¼ sec. 16, T4N, R8W	RC
330	NE¼ NE¼ sec. 16, T4N, R8W	RC
332	SE¼ SE¼ sec. 9, T4N, R8W	RC
339	SW¼ SW¼ sec. 8, T3N, R9W	RC
353	NW¼ SE¼ sec. 4, T3N, R9W	RC
357	NW¼ SE¼ sec. 9, T3N, R9W	Q
362	SE¼ SE¼ sec. 8, T3N, R9W	Q
365	SE¼ SE¼ sec. 9, T3N, R9W	RC
366	NW¼ SW¼ sec. 10, T3N, R9W	RC
375	NE¼ NW¼ sec. 11, T3N, R9W	RC
377	SE¼ SE¼ sec. 2, T3N, R9W	RC
382	SE¼ SE¼ sec. 2, T3N, R9W	RC
398	SW¼ SE¼ sec. 9, T3N, R9W	Q
401	SW¼ SW¼ sec. 10, T3N, R9W	RC
408	NE¼ NE¼ sec. 12, T3N, R9W	RC
409	SE¼ SW¼ sec. 6, T3N, R8W	RC
419	NE¼ SE¼ sec. 5, T3N, R8W	RC
424	NE¼ SE¼ sec. 5, T3N, R8W	RC
425	NE¼ NW¼ sec. 5, T3N, R8W	RC
451	NW¼ SW¼ sec. 32, T4N, R9W	RC
522	NW¼ NW¼ sec. 26, T4N, R9W	SC
528	NE¼ NE¼ sec. 26, T4N, R9W	SC
529	NW¼ NW¼ sec. 25, T4N, R9W	SC
532	SW¼ NW¼ sec. 25, T4N, R9W	SC
533	SE¼ NW¼ sec. 25, T4N, R9W	SC
535	NW¼ NE¼ sec. 25, T4N, R9W	SC
542	NE¼ SE¼ sec. 25, T4N, R9W	SC
548	NW¼ SW¼ sec. 20, T4N, R8W	SC
550	NW¼ SW¼ sec. 20, T4N, R8W	SC
552	NW¼ SW¼ sec. 20, T4N, R8W	SC
553	NE¼ SW¼ sec. 20, T4N, R8W	SC
554	NE¼ NW¼ sec. 20, T4N, R8W	SC
558	SW¼ SE¼ sec. 17, T4N, R8W	SC
560	SW¼ SE¼ sec. 17, T4N, R8W	SC
563	SE¼ SE¼ sec. 17, T4N, R8W	SC
575	SW¼ SW¼ sec. 16, T4N, R8W	RC
580	NW¼ NW¼ sec. 32, T4N, R9W	SC
584a	NW¼ NW¼ sec. 32, T4N, R9W	SC
589	SW¼ SW¼ sec. 31, T4N, R9W	RC
592	SW¼ NE¼ sec. 31, T4N, R9W	RC
593	NE¼ NW¼ sec. 31, T4N, R9W	RC
596	NW¼ NE¼ sec. 32, T4N, R9W	RC
604	SE¼ NW¼ sec. 20, T4N, R8W	SC
605	SE¼ NW¼ sec. 20, T4N, R8W	SC
607	SE¼ NW¼ sec. 20, T4N, R8W	SC
610	NE¼ SW¼ sec. 20, T4N, R8W	SC

APPENDIX 16 (cont.)

Locality Number	Location	Outcrop
611	NE¼ SW¼ sec. 20, T4N, R8W	SC
612	NE¼ SW¼ sec. 20, T4N, R8W	SC
615	SE¼ SW¼ sec. 20, T4N, R8W	SC
616	SE¼ SW¼ sec. 20, T4N, R8W	SC
620	SE¼ NE¼ sec. 29, T4N, R8W	SC
621	SE¼ NE¼ sec. 29, T4N, R8W	SC
622	SE¼ NE¼ sec. 29, T4N, R8W	SC
623	SW¼ NW¼ sec. 28, T4N, R8W	SC
626	NW¼ SW¼ sec. 28, T4N, R8W	SC
628	NW¼ SW¼ sec. 28, T4N, R8W	SC
629	SW¼ SW¼ sec. 28, T4N, R8W	SC
630	SE¼ SW¼ sec. 28, T4N, R8W	SC
631	NW¼ NE¼ sec. 33, T4N, R8W	SC
645a	SW¼ SE¼ sec. 9, T3N, R8W	RR cut
645b	SW¼ SE¼ sec. 9, T3N, R8W	RR cut
646	SE¼ SW¼ sec. 9, T3N, R8W	RR cut
648	SE¼ SE¼ sec. 8, T3N, R8W	RR cut
656	SW¼ NE¼ sec. 16, T3N, R8W	SC
659	NE¼ SE¼ sec. 6, T4N, R8W	RC
662	SW¼ NE¼ sec. 31, T4N, R9W	RC
672	SE¼ SW¼ sec. 7, T3N, R8W	RC
673	SE¼ SW¼ sec. 7, T3N, R8W	RC
674	SE¼ SW¼ sec. 7, T3N, R8W	RC
746	SW¼ NW¼ sec. 16, T4N, R8W	RC
750	NE¼ SW¼ sec. 8, T4N, R8W	RC
766	SW¼ SE¼ sec. 36, T5N, R9W	RC
827	NW¼ SW¼ sec. 32, T4N, R8W	RC
854	SE¼ SW¼ sec. 11, T3N, R9W	RC
907	NW¼ SW¼ sec. 10, T3N, R9W	RC
925	SW¼ NW¼ sec. 34, T4N, R9W	RC
927	SE¼ NE¼ sec. 33, T4N, R9W	RC
930	SE¼ NE¼ sec. 33, T4N, T9W	RC
1005	NW¼ NE¼ sec. 24, T5N, R9W	RC
1200	SE¼ NW¼ sec. 21, T5N, R8W	RC
1201	SE¼ NW¼ sec. 21, T5N, R8W	RC
M-250	NE¼ NE¼ sec. 12, T3N, R9W	RC
M-585	SW¼ SW¼ sec. 6, T3N, R8W	RC
Q-350	SE¼ NE¼ sec. 25, T4N, R9W	RC
Q-850	SE¼ NE¼ sec. 25, T4N, R9W	RC
R-83-140	SE¼ NE¼ sec. 9, T4N, R9W	RC
R-83-141	SE¼ SW¼ sec. 9, T4N, R9W	RC

RC = road cut
SC = stream cut
RR cut = railroad cut
Q = quarry