AN ABSTRACT OF THE THESIS OF

Jeffrey W. Lane for the degree of Master of Science in Geology

presented on May 19, 1987.

Title: Relations Between Geology and Mass Movement Features in a Part of the East Fork Coquille River Watershed, Southern Coast Range, Oregon

Signature redacted for privacy.

Abstract Approved: ________________________
Frederick J. Swanson

ABSTRACT

Various types of mass movement features are found in the drainage basin of the East Fork Coquille River in the southern Oregon Coast Range. The distribution and forms of mass movement features in the area are related to geologic factors and the resultant topography.

The Jurassic Otter Point Formation, a melange of low-grade metamorphic and marine sedimentary rocks, is present in scattered outcrops in the southwest portion of the study area but is not extensive. The Tertiary Roseburg Formation consists primarily of bedded siltstone and is compressed into a series of west to northwest-striking folds. The overlying Lookingglass, Flournoy, and Tyee Formations consist of rhythmically bedded sandstone and siltstone units with an east to northeasterly dip of 5-15° decreasing upward in the stratigraphic section. The units form cuesta ridges with up to 2000 feet of relief.

The distribution of mass movements is demonstrably related to the bedrock geology and the study area topography. Debris avalanches are more common on the steep slopes underlain by Flournoy Formation and Tyee Formation sandstones, on the obsequent slope of cuesta ridges, and on north-facing slopes.

Soil creep occurs throughout the study area and may be the primary mass movement form in siltstone terrane, though soil creep was not studied in detail. Slump-earthflows, rockfalls, and rock slumps also occur in the study area though less extensively than debris avalanches.

Stratigraphy and bedrock attitude contributed to the pre-historic occurrence of a major landslide involving Flournoy and Tyee Formation bedrock. The Sitkum landslide dammed the East Fork Coquille River, forming a substantial lake which is now filled with sediments. The form and size of the Sitkum landslide is similar to other landslides which have dammed drainages in the Coast Range, including Loon Lake, Triangle Lake, and Drift Creek.

Comparisons with the Loon Lake landslide, which has a known radiocarbon date, provide estimated dates of 3125 years B.P. for the Sitkum landslide and 10,300 years for the Triangle Lake landslide.

Relations Between Geology and Mass Movement Features in a Part of the East Fork Coquille River Watershed, Southern Coast Range, Oregon

by

Jeffrey W. Lane

A THESIS

submitted to

Oregon State University

in partial fulfillment of the
requirements for the degree of

Master of Science

Completed May 19, 1987

Commencement June 1988

APPROVED:

Signature redacted for privacy.

Assistant Professor of Geology in charge of major

Signature redacted for privacy.

Head of Department of Geology

Signature redacted for privacy.

Dean of Graduate School

Date thesis is presented May 19, 1987

ACKNOWLEDGEMENTS

Traveling from point A to point B would never be very interesting if the unusual and unexpected did not occur along the way, changing both plans and the planner. This project, from start to finish, has been no exception. Without the support and encouragement of many people the plans and objectives might have been lost in the detours.

The most significant detour was the acceptance of an employment offer from the Defense Mapping Agency in St. Louis, Missouri, in November of 1982, delaying completion of the thesis until the present time. Course work and field research had been conducted from 1980 to 1982. My current position is with the U. S. Geological Survey's National Mapping Division in Rolla, Missouri.

My thanks to Dr. Fred Swanson for the long-distance, long-term support, and to the USFS Forestry Sciences Lab RWU 1653. This research was partially supported through a memorandum of understanding between the US Forest Service and Oregon State University, through the support of Dr. Swanson and Dr. Logan Norris.

Thanks to the Amoco Production Company, Denver, for a generous grant, and exceptional patience.

Thanks to Rich and Linda Kirk, and Marilyn Milne, for providing shelter and friendship and hot showers in the metropolis of Sitkum.

My gratitude to my parents for providing support and encouragement.

And a special thanks to the midwest climate for providing a great deal of incentive to complete this project.

TABLE OF CONTENTS

TABLE OF CONTENTS (cont'd.)

LIST OF FIGURES

LIST OF FIGURES (cont'd.)

LIST OF TABLES

LIST OF PLATES

Relations Between Geology and Mass Movement Features in a Part of the East Fork Coquille River Watershed, Southern Coast Range, Oregon

INTRODUCTION

Purpose

Western Oregon is a region of dynamic geologic activity along the junction of the continental North American and oceanic Juan de Fuca Plates. The geologic record shows deposition of great quantities of marine sediments, folding and uplift of mountains, and volcanic activity that continues to the present. The climate of western Oregon is also a dynamic force, with annual precipitation of up to 140 inches in parts of the Coast Range and 100 inches in the Cascade Range.

The interaction of geology and climate produces a geomorphic response that is characterized by mass movements of various type. Slow processes such as creep and slump-earthflow are typical in areas of clay-rich bedrock and cohesive soils, such as in the volcaniclastic rocks of the Cascades. More rapid processes, such as debris avalanches, are common in the Coast Range, where relatively competent sedimentary bedrock produces steep slopes and cohesionless soils (Swanston and Swanson, 1976).

The temperate climate also produces high rates of bedrock weathering and a heavy vegetative cover of predominantly conifer forests.

Cover of vegetation and overburden of mass movement deposits limit the efficiency and reliability of reconnaissance geologic mapping. Knowing the character of mass movement activity and the relationship between geology and mass movement processes in a particular area is helpful in avoiding erroneous structural and stratigraphic interpretations, and in anticipating the impact of land management practices.

Studies which have looked at the relationship between geology and geomorphology, particularly with respect to mass movement activity, have been conducted by several researchers in the mountainous areas of western Oregon. Swanson (1975) and Amaranthus et al (1985) in the Klamath Mountains; Swanston and Swanson (1977), Schulz (1980), and Hicks (1982) in the western Cascades; and Gresswell et al (1979) and Graham (1986) in the central Coast Range, are a few of the mass movement studies in the region.

This thesis is concerned with the geology and mass movement activity in a selected area of the southern Oregon Coast Range. The bedrock is predominantly Eocene marine siltstones and sandstones of the Roseburg, Lookingglass, Flournoy, and Tyee Formations. The stratigraphy, structure, and topographic expression of the bedrock units set the stage for a variety of mass movement processes, ranging from soil creep to large complex landslides. Land management practices, particularly those associated with timber harvesting, are an added factor in the geomorphic response to the interaction of geology and climate.

Previous Work

Mapping and stratigraphic revision of the southern Coast Range geology has been an ongoing process since Diller (1901) surveyed the region. Allen and Baldwin (1944), Baldwin and Beaulieu (1973), and Baldwin (1974) published new observations and interpretations of the occurrence, structure, and stratigraphy of the Tertiary units in portions of Coos and western Douglas Counties. Baldwin's work is largely responsible for the present designations of the Tertiary sedimentary rocks and their distinction in the field.

Students of Baldwin mapped the geology of selected townships within and adjacent to the thesis study area. Magoon (1966) worked in township T. 29 S., R. 11 W., in the western portion of the study area, mapping rocks which he identified as the lower and middle members of the Umpqua Formation, and the Tyee Formation (the study area bedrock is discussed in the geology section of this thesis, pp. 25-35). These units were redefined by Baldwin (1974) as the Roseburg, Lookingglass (Bushnell Rock and Tenmile Valley Members), and Flournoy (White Tail Ridge Member) Formations. Nelson (1966), Trigger (1966), and Fairchild (1966) mapped adjacent townships to the north and south of the study area, delineating the occurrence and nature of the contacts between the Tertiary bedrock units. Trigger (1966) recognized an unconformable division between the Tyee and Flournoy sediments which becomes conformable northward into the study area.

Detailed geomorphic mapping of the southern Coast Range, particularly with respect to geologic structure and stratigraphy, has been neglected. Beaulieu and Hughes (1975), however, published environmental geology and geologic hazards maps for western Coos and Douglas Counties, including a portion of the study area. The report was designed to aid land-use planning agencies and was concerned with areas of flood susceptibility, mass movement, poor drainage, etc. Beaulieu and Hughes (1975) assigned qualitative ratings to the geologic units with regard to potential environmental hazards, but their mapping was not detailed.

Detailed mapping of geomorphic features has been conducted in the Elk and Sixes River basins of northern Curry County, in Jurassic and Cretaceous sediments, intrusives, and metasediments of the Klamath Mountains province (Swanson, 1975). The effort there was directed toward synthesizing land-use impact, geology, and geomorphology.

In the central Coast Range Gresswell, Heller, and Swanston (1979) studied mass movement activity, primarily debris avalanches and debris torrents, in the Mapleton Ranger District of the Siuslaw National Forest. They found increased mass movement activity as a result of forest management activities such as clear-cut logging and road construction. Slope steepness, slope aspect, slope dissection, and soil type were also considered, indirectly relating the geology to the geomorphic processes. Graham (1986) studied the geology (Flournoy Formation, Yachats Basalt) and mass movement relationships in the Smith River drainage basin in the Coast Range.

Schulz (1981) mapped the relationship between mass movements and bedrock in the Bull Run watershed of the Mt. Hood National Forest, in the western Cascades of Oregon, producing a mass movement hazard map for the late Tertiary sedimentary and volcanic units. Hicks (1982) did similar work in the Middle Santiam River drainage in the Willamette National Forest. His study included monitoring of several large, active landslide complexes. As elsewhere in the western Cascades, the juxtaposition of crystalline volcanic flow rock and tuffaceous bedrock is an important factor in the development of these mass movement features. Swanston and Swanson (1977) also discussed complex mass movement terrains in the western Cascades, and the relationships between geology and slope failure. Graham (1986) studied the geology (Flournoy Formation, Yachats Basalt) and mass movement relationships in the Smith River drainage basin in the Coast Range.

Methods

The gathering of geological and geomorphological data for the thesis project was accomplished by field reconnaissance, aerial photo analysis, and office compilation. Field work in the thesis area was conducted from July through September 1981, with additional work at Loon and Triangle Lakes in 1982. Mapping in these areas employed standard usage of a Brunton compass and rangefinder. Geological and geomorphological data were plotted on 1:15,840-scale maps (4 inches =

1 mile) enlarged from U.S. Geological Survey 15-minute quadrangle topographic bases.

A portion of the geomorphic data, and the detailed mapping of the complex landslide features, was derived from aerial photo interpretation. Black and white 1967 air photos (1:20,000 scale) from the U.S. Agricultural Stabilization and Conservation Service (ASCS) were used in the mapping process.

GEOGRAPHY

Location

The study area is located primarily within the watershed of the East Fork Coquille River, Coos County, Oregon (Figures 1 and 2). Physiographically it is situated in the southern portion of the Oregon Coast Range province. The East Fork Coquille River originates 8 miles to the east of the study area near the summit of the Coast Range, flows westerly for 26 miles through the study area, and continues another 45 miles to the Pacific Ocean (18 miles due west). The geographic center of the study area is approximately 17 air-miles southeast of Coos Bay, Oregon, and 30 air-miles west-southwest of Roseburg, Oregon.

The study area encompasses approximately 64 square miles, the western third of which is included within the USGS Coquille 15-minute quadrangle; the eastern two-thirds is covered by the Sitkum 15-minute quadrangle. The portion of the study area south of the East Fork Coquille River is within the US Bureau of Land Management's (BLM, Coos Bay District) Myrtlewood Resource Area, while the portion of the study area north of the East Fork is part of the BLM's Burnt Mountain Resource Area. The study area boundary generally follows drainage divides within the East Fork Coquille River watershed except along the northwestern and western margins where it parallels Cherry Creek, Middle Creek, and the North Fork Coquille River.

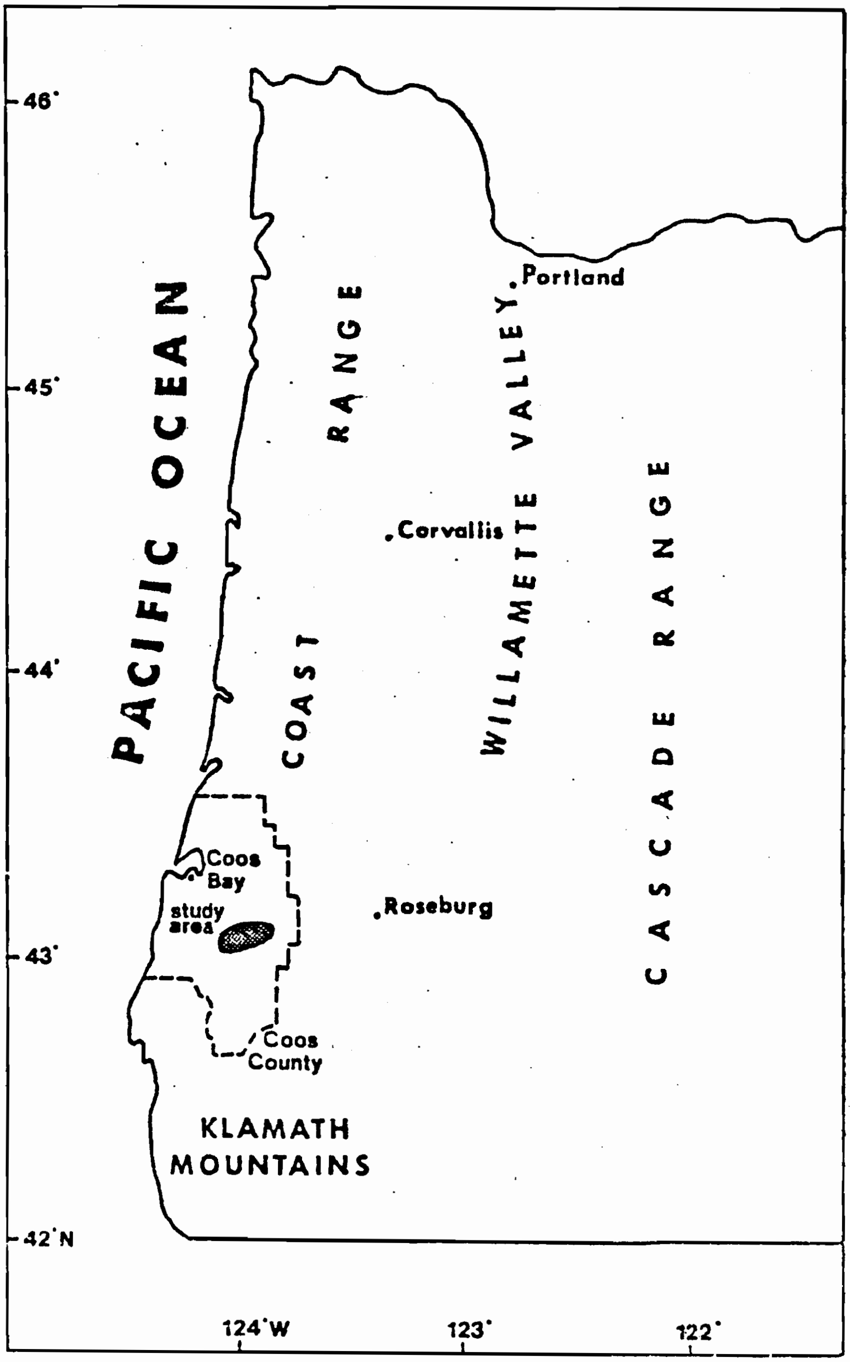


Figure 1. Location of the study area in western Oregon.

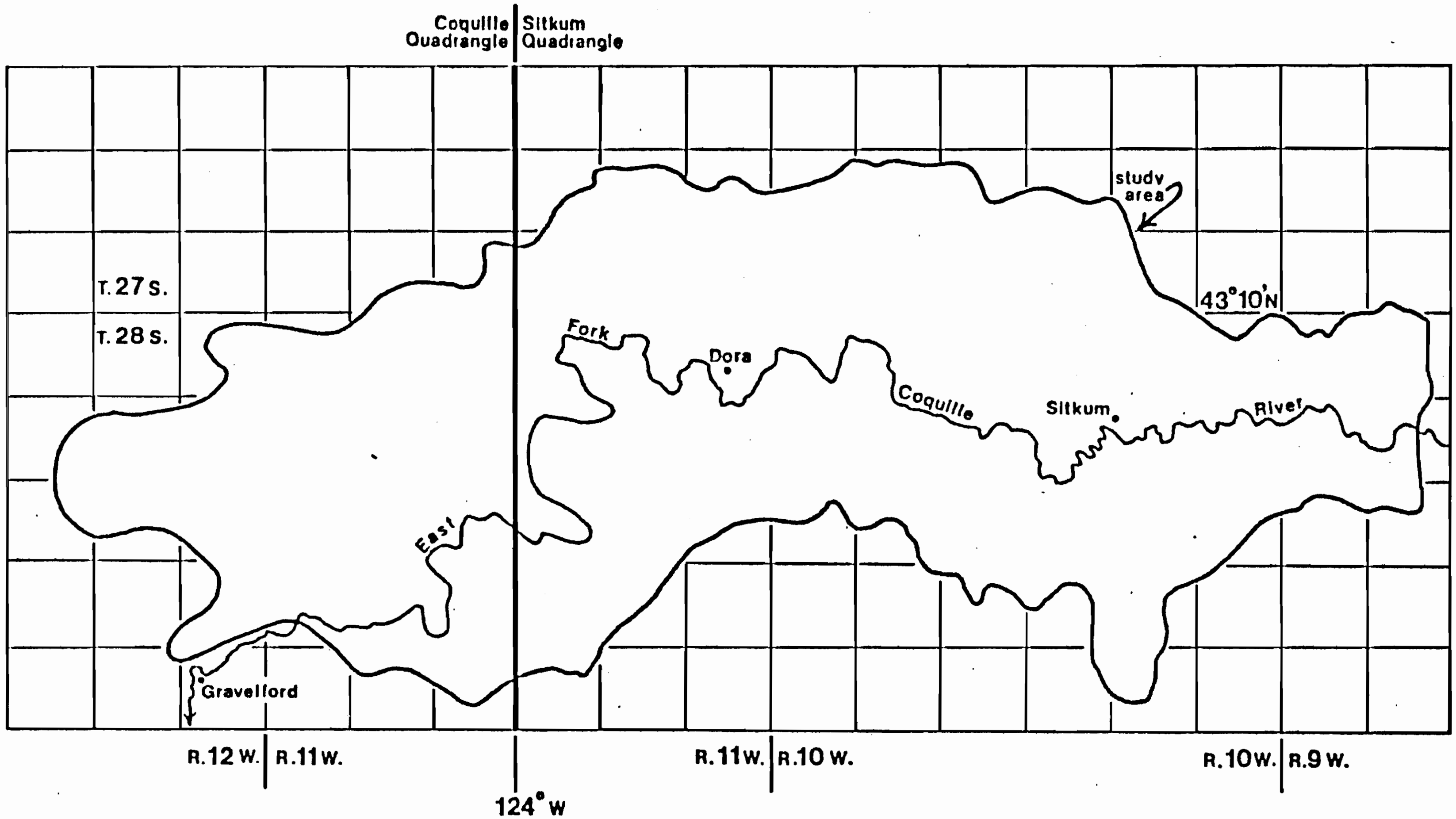


Figure 2. Study area boundary, with Land Office divisions and locations of communities; see Plates 1 and 2. Scale: 1 inch = 2 miles.

Paved county roads follow the East and North Forks of the Coquille River and connect the study area communities of Sitkum, Dora, and Gravelford with State Highway 42 at Myrtle Point, 10 miles southwest of Gravelford. Other, unpaved county roads connect the communities with Coquille, Coos Bay, and Roseburg. Unpaved county roads, BLM-maintained all-weather roads, as well as skid roads and trails provide additional access within the study area.

Climate

The study area, like the rest of the Coast Range, has a wet and temperate climate due to the westerly storm track from the Pacific Ocean and the orographic barrier of the Coast Range. Mild and wet winters with nearly all of the precipitation as rain, plus warm and dry summers, are typical of the region. Annual precipitation in the study area ranges from about 60 inches in the west to nearly 80 inches in the east, nearer to the Coast Range summit (Oregon State Water Resources Board, 1961). Seventy-five to 80% of the precipitation falls between October 1 and March 31 (Franklin and Dyrness, 1973), with an annual average of 160 to 180 days of precipitation and 40 to 50 days of 0.5 inches or more precipitation (Atlas of Oregon, 1976). Daily temperatures in January average 6° C, while the average in July is 17° C (Atlas of Oregon, 1976). The maritime influence of the nearby Pacific Ocean moderates the local climate during the summer months so that valleys on the windward side of the Coast Range are often blanketed by fog, while valleys on the leeward side of the Coast Range are warmer and drier.

Vegetation and Wildlife

The temperate, moist climate of the Coast Range results in a heavily vegetated landscape dominated by conifer forests and lush undergrowth. Vegetation in the study area is in the *Tsuga heterophylla* zone, covering most of western Oregon (Franklin and Dyrness, 1973). Because of widespread disruption by logging and fire, the climax species are not dominant; instead Douglas-fir (*Pseudotsuga menziesii*) is dominant, along with red alder (*Alnus rubra*), bigleaf maple (*Acer macrophyllum*), and western hemlock (*Tsuga heterophylla*). These species form dense stands that mantle the hill-slopes, with the deciduous species predominant in disturbed sites and riparian zones.

In the western portion of the study area, where the influence of the Pacific Ocean is more strongly felt, California-laurel (*Umbellularia californica*, or Oregon-myrtle) and Port Orford-cedar (*Chamaecyparis lawsoniana*) are characteristic species found along with the dominants. Eastward in the study area, with increasing elevation and decreasing maritime influence, western redcedar (*Thuja plicata*) and Pacific madrone (*Arbutus menziesii*) become distinctive members of the forest canopy.

Wildlife in the area include large numbers of Roosevelt elk (*Cervus elephus*) and blacktail deer (*Odocoileus hemionus colombianus*). A small rodent, the mountain beaver (*Aplodontia rufa*), causes damage to young tree seedlings, and may be a factor in soil mass movement features because of its burrowing activities (Pierson, 1977).

Topography

The study area lies on the western slopes of the Oregon Coast Range and so is characterized by moderate hillslope gradients and narrow valleys. Hillslope gradients in the western half of the study area are predominantly between 5% and 40%; elevations range from 60 feet at the confluence of the East and North Forks of the Coquille River, to about 1100 feet on nearby ridgetops. Hillslope gradients in the eastern half of the study area are generallly greater than 40%, and commonly greater than 50%; elevations range from 607 feet at Sitkum to about 2700 feet in section 26 (T. 28 S., R. 10 W.). The only areas of low relief are the floodplains and terraces of the lower East Fork Coquille River, the Sitkum landslide deposit, and the sediment-filled valleys upstream from the Sitkum landslide.

Structural bedrock control is evident in the topography and in the drainage system of the study area. In the western third of the study area, which is predominantly Roseburg Formation terrane, the ridges and streams have a southwest-northeast trend that roughly parallels the strike of the folded bedrock and the orientation of two mapped faults. In the eastern two-thirds of the study area underlain by the Lookingglass, Flournoy, and Tyee Formations, the strata are not strongly folded, but instead dip gently eastward. As a result the ridges have a cuesta form, with steep western obsequent slopes and gentler eastern dip slopes, and the streams have a trellis drainage pattern. Tributary streams have a general north-south orientation, while the East Fork Coquille River flows westward across the strike of the bedding as an antecedent or superposed stream.

It is probable that the main East Fork drainage system developed on the Eocene bedrock as the units were being folded and uplifted at a rate slow enough to allow the river to maintain a westward course. Tributaries of the East Fork, with insufficient erosive power to maintain their courses against the structural grain, have north-south orientations parallel to bedrock strike. If the East Fork were a superposed stream, it must have once flowed on post-Tyee beds that were relatively undisturbed or had an east-west structural grain. Erosion would then have removed the younger strata and dropped the East Fork onto the folded Eocene beds. The geologic record, however, doesn't indicate such a substantial break in structural orientation of Tertiary beds or such an extensive depositional basin for the post-Tyee sediments, so it is less likely that the East Fork is a superposed stream.

In addition to structure, the bedrock lithology of the study area also exerts an obvious influence on the topography since the Lookingglass, Flournoy, and Tyee Formations are comprised, in part, of resistant sandstone and/or conglomerate units. These bedrock units are typically expressed as cliffs on the steep western obsequent slopes of ridges, and as sites of falls or rapids in stream channels. Continued erosion in this portion of the study area produces homoclinal shifting of the ridges and valleys eastward. Such activity is not apparent in the western portion of the study area because of the thinness and great degree of folding of individual bedding units in the Roseburg Formation.

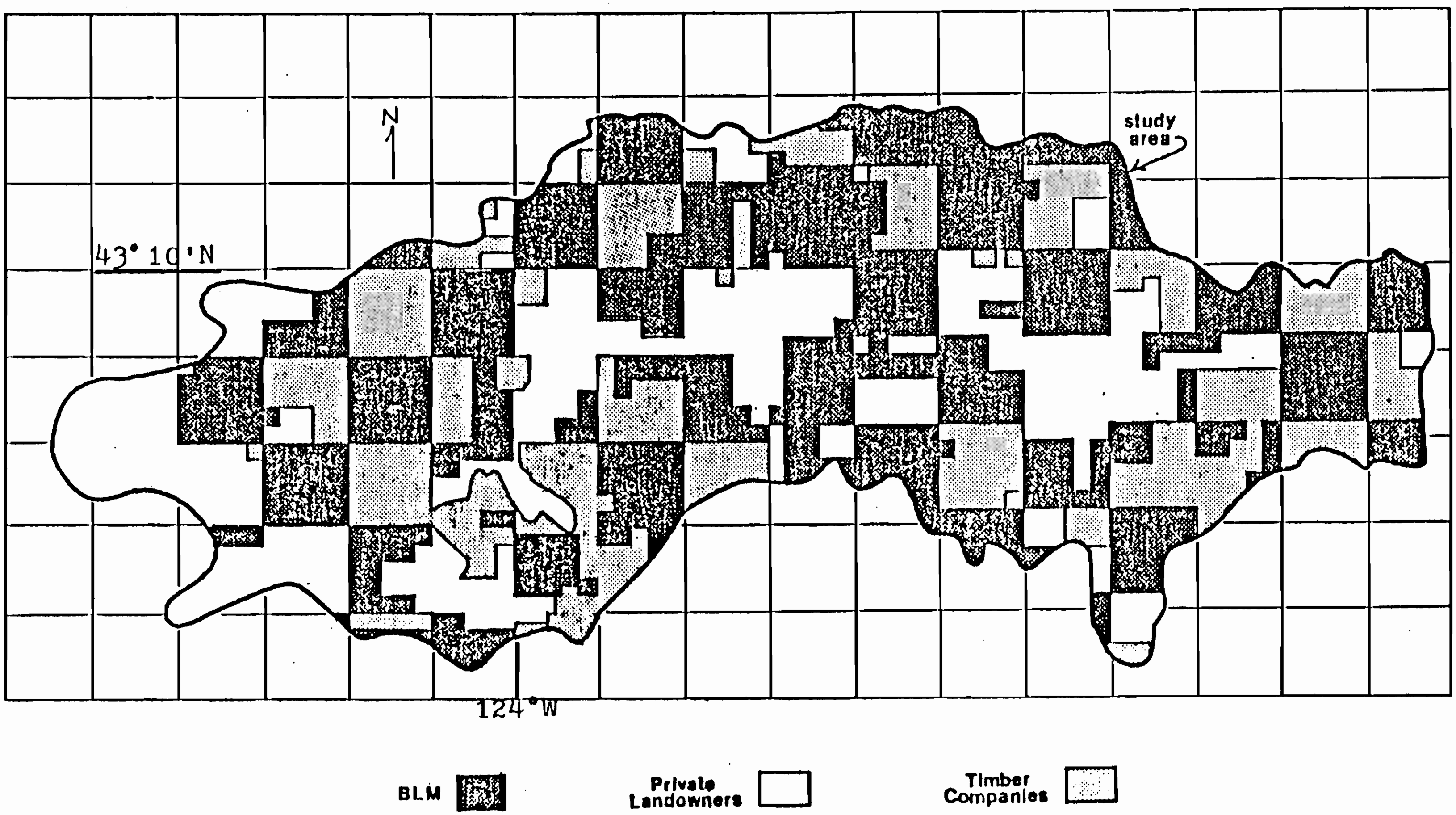
Land-Use History

Prior to settlement by white immigrants in the 1860's and 1870's (Dodge, 1898) the major disturbances in the natural environment were fire, flood, and mass movement activity. Coquille Indians may have occupied the East Fork valley, at least seasonally, but their numbers were apparently small (Dodge, 1898; Hall, 1984) and their impact on the environment unknown.

Since the mid-1800's human impact on the landscape has become a dominant force. Most of the floodplains and terraces have been cleared for rural homes, agriculture, and timber production. The impact of fire on forested hillslopes during the past 50 years has apparently been limited by wildfire suppression and controls on forest activities, though a major blaze in September 1936 destroyed much of the forest cover between Dora and Sitkum (M. Combs, pers. comm., 1982).

Logging activity, however, has replaced fire as the primary disturbance of the landscape. Clear-cut logging and slash burning remove the forest cover from hillslopes, restart vegetative succession, and may produce some temporary changes in the hydrologic cycle (Harr, 1976). Road-building, logging in steep terrain, and other factors often contribute to mass movement activity, especially debris avalanches (Marion, 1981). Soil erosion, water quality, stream and riparian habitats, and reforestation are all vulnerable to impact by such mass movement activity.

Figure 3. Land ownership patterns in the study area. BLM = U.S. Bureau of Land Management, federal agency lands; Private Landowners include rural residences, agricultural lands, and small-scale logging operations; Timber Companies include large-scale timber corporations.



N
study area
43° 10'N
124°W
BLM
Private Landowners
Timber Companies

Figure 3.

Land-use in the study area can be divided into residential-agricultural, roads, and forested areas with the plant community at various stages of development. Forest management is the principle factor in the quality of all these land classes except residential-agricultural, though even there it is important since the forest lands provide potable water for homes and crops, employment for many rural residents, and recreation. The timber industry owns approximately 25% of the land in the study area; another 25% is owned by small individual landowners; and the remaining 50% is in public domain under the US Bureau of Land Management (Figure 3).

The presence of young growth timber, stumps, dead snags and abundant pioneer species indicate nearly 80% of the land in the study area has been clear-cut or burned, probably since the 1930's. Roads, maintained and abandoned, reach to within 1 mile of 95% of the area.

Income-producing usage of the natural resources is limited to timber products, agriculture, hunting and fishing, and two rockquarry sites: one in Otter Point Formation metasediments and the other in basal Tyee Formation sandstone. There is no mineral or petroleum production in the study area, though low-grade coals are extensive in the late Eocene Bateman and Coaledo Formations of the Coos Bay area to the northwest.

GEOLOGY

Regional and Historical Geology

The Oregon Coast Range is a moderately deformed block of Tertiary marine sedimentary and volcanic rocks. It is bordered on the east by the Willamette Valley downwarp and western Cascade volcanic rocks; on the south by the pre-Tertiary Klamath Mountains complex; on the west by the continental shelf and slope; and extends northward to the Olympic Mountains in Washington. The study area, in the southern portion of the Coast Range, lies at the juncture of three geologic "influences". An extension of Klamath Mountains bedrock and structure is found in the southwest portion of the study area; intensely folded early Eocene marine sediments and submarine volcanics in the western portion of the study area; and gently tilted, early to middle Eocene marine sediments in the central and eastern portion of the study area, typical of the Oregon Coast Range.

The pre-Tertiary rocks of the Klamath Mountains in Oregon consist of marine sedimentary, volcanic, schist, serpentine, peridotite, and dioritic intrusive rocks disrupted by intense folding and faulting. Dott (1971) described the Klamaths as being analogous, structurally and lithologically, to the Sierra-Franciscan arc-to-trench sequence in California. Current research, however, views the geology of southwest Oregon in terms of micro-plate tectonics and accreted terranes. The Klamaths are regarded as consisting of several distinct terranes which were "rafted" in from distant sites by the complex collision of the North American and adjacent oceanic plates

(Dott and Bourgeois, 1980; Blake and Jayko, 1980; Bourgeois and Dott, 1985).

Included in the confused geologic picture are rocks identified as the Upper Jurassic Otter Point Formation. The Otter Point consists of a melange of shallow to deep water clastics, cherts, submarine volcanics and low-grade metamorphics, indicating an origin in or near a subduction trench. The Otter Point rocks, which outcrop within the study area, are associated with Upper Cretaceous conglomerates which Bourgeois and Dott (1985) postulate as having been derived from sediment sources in southern California, subsequently displaced northward and accreted to the Oregon margin.

The Otter Point and Upper Cretaceous rocks, which together are referred to as the Gold Beach terrane, are postulated as having an origin in a subduction setting. They would have been translated northward by a change to transcurrent faulting along the North American plate margin, and then accreted to the continent by renewed subduction along the Oregon coast during the Eocene (Bourgeois and Dott, 1985).

The change to subductive collision during the Eocene also meant renewed volcanism in the region. The oldest Cenozoic rocks exposed in the southern Coast Range are submarine basalts and rhythmically bedded sandstone and siltstone of the Paleocene to early Eocene Roseburg Formation, mapped and described by Baldwin (1974). The basalts are apparently of sea-floor or oceanic-island origin and interfinger with the overlying clastic sediments. The base of the Roseburg Formation is poorly exposed but at one locality rests directly on Otter

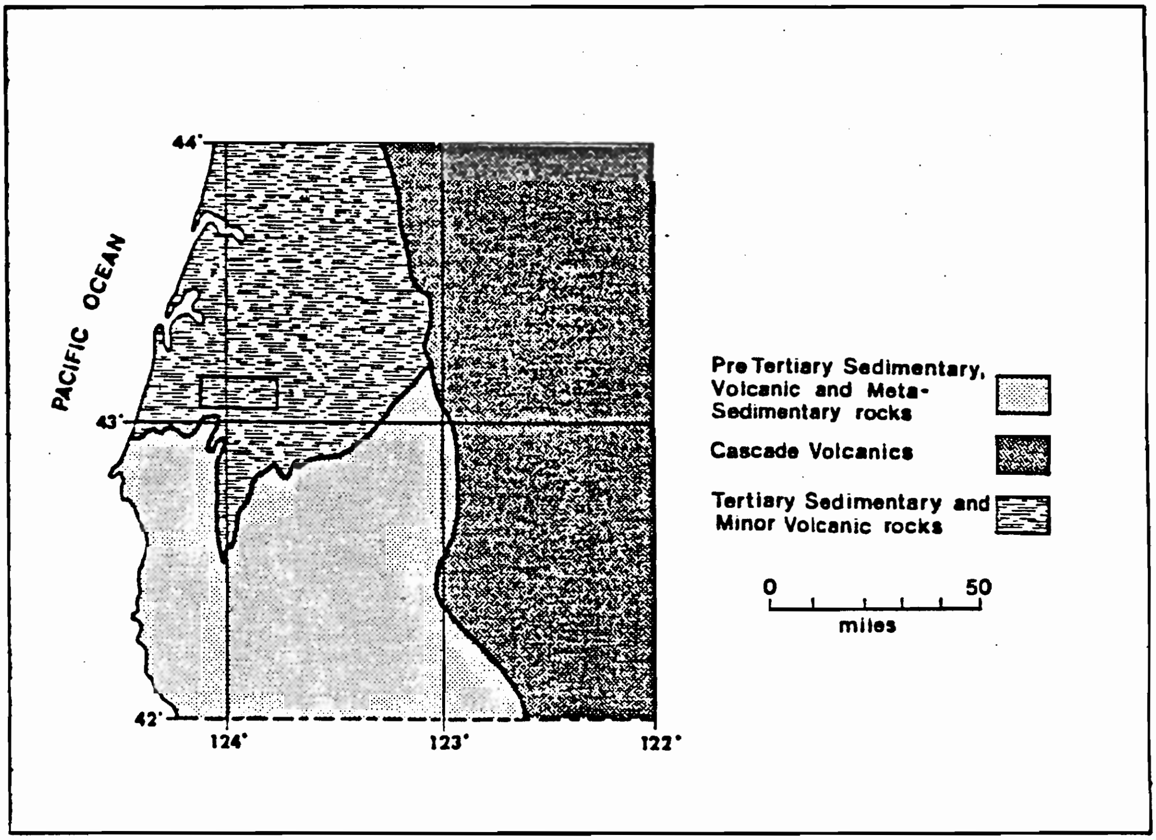


Figure 4. Generalized geologic map of southwest Oregon, after Walker and King (1967). Rectangle denotes location of study area.

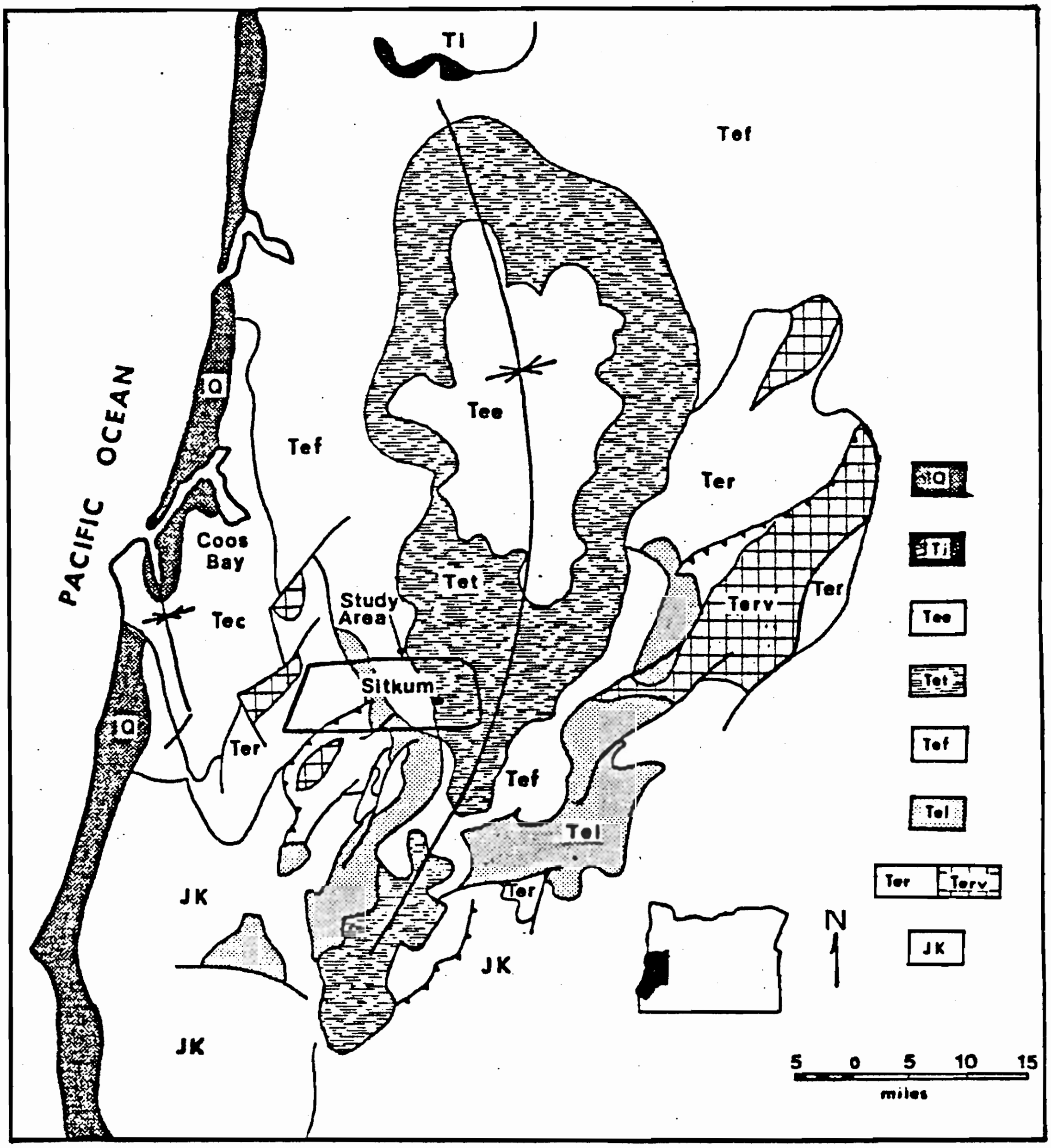


Figure 5. Geology of the southern Coast Range, after Baldwin (1981). Q= Quaternary sediments, Ti= intrusive basalts, Tee= Elkton Fm., Tef= Flournoy Fm., Tel= Lookingglass Fm., Ter= Roseburg Fm., Terv= Roseburg volcanics, and JK= pre-Tertiary sedimentary and metamorphic rocks.

Point bedrock. Roseburg deposition ceased with the onset of Eocene thrusting and folding parallel to structural trends of older units in the Klamath Mountains (Baldwin and Beaulieu, 1973).

The complex collision of North America with the oceanic plates resulted in the rotation, translation, and accretion of relative small geologic blocks having very different histories. These processes have been advanced to explain the occurrence of exotic terranes in the Klamath Mountains, apparent rotation of Coast Range basalts, and disparate source areas for Coast Range sediments. Drake (1982) has summarized current data and models for the evolution of the Oregon continental margin, and Blake and Jayko (1980) discussed the tectonostratigraphic terranes (microplates) of southwest Oregon. Simpson and Cox (1977), Magill and Cox (1981), and others have proposed post-Eocene tectonic rotation of the Coast Range block as an explanation of discordant magnetic declinations in Coast Range basalts, with the rotation resulting from the oblique subduction of the Farallon plate beneath the North American plate. The deformation of the Roseburg Formation also probably resulted from this subductive process (Baldwin and Beaulieu, 1973).

The hiatus following Roseburg deposition and deformation was apparently short, since both the Roseburg Formation and overlying Lookingglass Formation strata contain Penutian stage foraminifera (Baldwin and Beaulieu, 1973). The Lookingglass rests on the Roseburg with an angular unconformity and represents an early Eocene marine transgression over most of the present extent of the southern Coast Range and the northern margins of the Klamath Mountains. The

Lookingglass basal conglomerate grades upward into finer-grained sandstone and siltstone deposited farther from shore in deeper parts of the basin. A middle Eocene regression and subsequent transgression resulted in deposition of the overlying Flournoy Formation, which rests upon the Lookingglass with an angular unconformity. Regression and transgression occurred again later in the middle Eocene when the Tyee Formation was deposited with angular unconformity upon the Flournoy.

The Lookingglass and Flournoy Formations consist of sediments which were apparently transported from source areas in the Klamath highlands, and exhibit repeated transition from nearshore to deeper-water conditions prior to uplift and regression. The Lookingglass and Flournoy basins were limited primarily to the southern portions of the present Coast Range. The Tyee basin, however, was much more extensive and overlapped all other Tertiary units (Baldwin and Beaulieu, 1973).

Furthermore, the Tyee seems to have a different sediment source than the underlying formations. Isotope studies by Heller et al (1985) indicate the Idaho batholith as the likely source of Tyee sediments, and the position of the Coast Range basin much farther east than at present. Subsequent tectonic rotation and translation moved the Coast Range, and possibly the Klamaths also, to the present location. Rotation may have resulted during collision and accretion with a seamount terrane, or from post-collision extension within the adjacent continental plate (Heller et al, 1985).

Though the early and middle Eocene units are separated by angular unconformities, the post-Roseburg deformation has been mild (Dott, 1971), with gentle folding and little compressional faulting. Tertiary beds of the southern Coast Range are deformed by a broad north-south trending syncline, the southwest flank of which outcrops within the study area and exposes the Lookingglass, Flournoy and Tyee Formations. Continued mild Cenozoic tectonism permitted deposition of middle to late Eocene sediments over the Tyee beds, deposition of late Eocene and younger sediments along the present coastline, and folding of the South Slough syncline near Coos Bay since the Oligocene (Baldwin, 1974). Folding of the South Slough syncline continues to the present, as does late Cenozoic uplift of the Coast Range, evidenced by tilted and elevated marine terraces along the coastline and inland within the Coast Range (Baldwin and Beaulieu, 1973). Eustatic sea-level rises during the Holocene have drowned coastal valleys and formed wide alluvial floodplains.

Stratigraphy

Otter Point Formation

The Late Jurassic Otter Point Formation outcrops in the southwestern portion of the study area and underlies about 5% of the study area (Plate 1). The Otter Point was named and described by Koch (1966) for a series of highly sheared and lithologically diverse rocks which are exposed on the northern and western margins of the

Klamath Mountains. They extend northward at least as far as the study area and probably beneath the Tertiary cover of the southern Coast Range.

Dott (1971) discussed the geology of southwest Oregon in terms of the structural, lithologic, and plate tectonic evidence, and noted similarities between the Otter Point rocks and the Franciscan assemblage of California. He regarded the Otter Point as a "tectonic-stratigraphic melange rather than a true rock-stratigraphic unit." Bourgeois and Dott (1985) referred to the rocks as the Otter Point complex and identified them as part of a postulated exotic terrane which originated adjacent to sediment sources in coastal southern California. Included within the Otter Point are black mudstones and fine-grained sandstones (30-40%), coarse-grained sandstones and conglomerates (20-30%), plus cherts, volcanics, and low-grade metamorphic rocks such as tectonically emplaced greenstone and blueschist (Koch, 1966; Dott, 1971).

Within the study area the Otter Point contains low-grade metamorphic rocks of greenstone, serpentine, mica schists, and chert. None of the other characteristic Otter Point lithologies are observed, though they may be present. The outcrops occur as large blocks apparently surrounded by colluvium on ridges and hillsides. An active quarry in section 29 of T. 28 S., R. 11 W., is in Otter Point greenstone, schist, and chert. Exposures of Otter Point rocks elsewhere in the study area are limited in number and extent.

Magoon (1966) mapped the geology in the western portion of the study area, and noted the occurrence of pre-Tertiary metamorphic rocks as blocks ranging in diameter from several yards to 0.25 mile. He considered them to be either allochthonous blocks which had slid into the Roseburg depositional basin, or fault-breccia associated with thrust faulting. A major thrust fault was mapped by Baldwin and Beaulieu (1973) in the southwest portion of the study area, near the Otter Point rocks (Plate 1).

Baldwin and Beaulieu (1973), however, mapped the Otter Point rocks in the study area as stratigraphic units in contact with Roseburg strata, partly by faulting and partly depositional. If this is indeed the case, most Otter Point terrane in the study area is apparently composed of fine-grained Otter Point sediments which are concealed by the thick vegetation, alluvium, and colluvium, leaving only the resistant metamorphic rocks exposed.

The Otter Point/Roseburg contact is not exposed, and is not distinguishable in the field. Neither are the faults mapped by Baldwin and Beaulieu (1973) in this area readily apparent, except as extensions of features projected from outside the study area or as structural explanations of the local bedrock relationships. Baldwin (1975) noted that the Otter Point terrane is characterized by moderate slopes, which would provide little distinction from the Roseburg terrane; by deep weathering and soil development on the sheared sedimentary rock; and by earthflow activity, which would also obscure contacts and outcrops. The occurrence of the Otter Point, as noted on the study area map (Plate 1), is from Baldwin and Beaulieu (1973) and reflects their structural and stratigraphic interpretation.

The Otter Point sedimentary rocks are dated as Latest Jurassic (Tithonian age, 145-140 million years) (Dott, 1971). The included metamorphic rocks, however, probably owe their origin and emplacement to Cretaceous thrust faulting.

Roseburg Formation

The Roseburg Formation, named and described by Baldwin (1974), encompasses marine sedimentary and volcanic rocks which Diller (1898) named the Umpqua Formation and Baldwin (1965, 1969) mapped as lower Umpqua. They are apparently correlative with the Siletz River Volcanics in the northern and central Coast Range (Baldwin, 1981).

Roseburg strata underlie about 28% of the study area, outcropping west of a line running from Cherry Creek Park to Frona Park and south. Nearly all exposures of Roseburg Formation strata are in roadcuts along major streams and ridges. The occurrence and character of the Roseburg strata within the study area, as noted on the geologic map (Plate 1), is after Baldwin and Beaulieu (1973). Field work for this thesis generated additional structural data and resulted in minor adjustments of the Roseburg contacts.

Within the study area the strata consist of siltstone of a few inches to several feet in thickness, rhythmically-bedded sandstone and siltstone, and at least one locality of pillow basalts. The major basalt sections of the Roseburg Formation lie to the west and south, outside of the study area. Siltstone becomes predominant over the sandstone and siltstone beds in the eastern portion of Roseburg outcrops. The Roseburg sediments in the study area are intensely folded along the northeast-southwest fold axis, so that the total thickness is unknown. Approximately 8,000 feet, however, was measured in a section along Highway 42 between Coquille and Myrtle Point, 4 miles to the west (Baldwin, 1974).

The sandstone and siltstone beds in the study area are light to dark gray, tuffaceous, and weather to a yellow-brown color. Magoon (1966) described the sandstone as medium to coarse-grained, lithic and subfeldspathic wackes, having angular to subangular grains of quartz, feldspar, volcanic rock, chert, schistose metamorphic rock, quartzite and sedimentary rock and a matrix of crystalline chlorite. The Roseburg basalt in this region was described as calc-alkalic and fine-grained (Klohn, 1967). The occurrence of basalt within the study area is apparently limited to an area about two miles upstream from the mouth of Elk Creek; pillow basalt there is exposed adjacent to tuffaceous siltstones.

Age determinations for sedimentary units in the Roseburg Formation, through use of foraminifera, place it in the Penutian Age. The Roseburg as a whole is generally regarded as being of Paleocene through early Eocene in age (Baldwin, 1974), 54-52 million years.

Lookingglass Formation

The Lookingglass Formation, named and described by Baldwin (1974) and formerly identified as the middle member of the Umpqua Formation (Baldwin, 1965), lies stratigraphically above the Roseburg Formation.

An unconformable erosion surface separates the highly disturbed Roseburg from the gently tilted Lookingglass. The Lookingglass is exposed on the southwestern limb of a major north-south trending syncline and has a general eastward dip of about 10°.

Lookingglass bedrock outcrops in the 2 to 2.5-mile wide band running north-south through the west-central portion of the study area, and underlies about 20% of the study area. It is flanked on the west by Roseburg beds and on the east by stratigraphically higher Flournoy beds. Field work for this study confirmed Baldwin and Beaulieu's (1973) mapping of the Lookingglass, added structural data, and refined placement of contacts.

Baldwin (1974) divided the Lookingglass into three members which grade into one another. The basal Bushnell Rock Member has a thickness of 800 to 1000 feet at the type locality, and is composed of a pebble conglomerate and massive sandstone. The Bushnell Rock Member is not present in the study area, however, except on the ridge northwest of Pleasant Hill School (sects. 16 and 21, T. 28 S., R. 22 W.) where it occurs as an outlier separated from the main body of Lookingglass bedrock in the study area (Plate 1, site #50601).

Baldwin (1974) noted the occurrence of the Bushnell Rock Member primarily along the southern margins of the Coast Range basin, closer to the sediment source in the Klamaths. The Lookingglass outlier near Pleasant Hill School likely represents a depression in the Roseburg erosion surface which retained basal Lookingglass sediments while terrain immediately to the east received none. The Bushnell Rock Member is here represented by a coarse-grained pebbly sandstone with clasts of quartz, chert, and basalt, derived from intrusive and volcanic rocks.

The main body of Lookingglass bedrock outcropping in the study area is composed of thin, rhythmically-bedded, green-gray, fine-grained sandstone and siltstone beds, plus thin concretionary limestone beds. It apparently represents the local preservation of Baldwin's (1974) middle Tenmile Member. In the type area, on the east limb of the Coast Range syncline, Baldwin noted 3,200 feet of the middle member, while approximately 2,500 feet of section are preserved in the study area (Magoon, 1966). The change from conglomerate to fine-grained sandstone and siltstone indicates marine transgression, with the basin shoreline farther south than during deposition of the basal Bushnell Rock Member. The Tenmile sandstone is classified as a lithic wacke, with up to 30% rock fragments such as sandstone, schist, slate, and chert, pointing to the Klamath Mountains as the source (Magoon, 1966). All exposures of the Tenmile Member bedrock are restricted to stream beds and roadcuts. The fine-grained rocks weather deeply and commonly in a spheroidal manner, so that bedding planes are obscured.

The overlying Olalla Creek Member, composed of conglomerate and pebbly sandstone, is missing from the study area, having either been eroded or never deposited here. It does represent, however, a shallowing of the basin.

The Tenmile Member is dated by Thomas and by Rau (Baldwin, 1974) as possibly Penutian to early Ulatisian (early Eocene, 53-51 million years).

Flournoy Formation

The Flournoy Formation (Baldwin, 1974) lies unconformably upon the Lookingglass beds, but also dips gently eastward an average of 8°. It was formerly identified as the upper member of the Umpqua Formation (Baldwin, 1965, 1981; Diller, 1898, 1899), and in this locality was long confused with the overlying Tyee Formation (Magoon, 1966) because of similar lithology and the disconformable contact. Approximately 2300 feet of section, and possibly as much as 3000 feet, are represented by outcrops of Flournoy bedrock in the east-central portion of the study area between Dora and Sitkum. About 25% of the study area is underlain by Flournoy bedrock (Plate 2). Field work for this thesis added structural data and stratigraphic contacts, but essentially confirmed the mapping of Baldwin (1974).

Baldwin (1974) divides the Flournoy into the lower White Tail Ridge Member, which is composed of basal sandstone and conglomerate, and the upper Camas Valley Member, which is composed of fine-grained rhythmically-bedded sandstone and siltstone. Both are apparently represented in the study area, as the Flournoy bedrock becomes noticeably finer-grained up-section; cliffs of Flournoy sandstone are generally restricted to the lower, western beds. The Flournoy sandstone is gray, medium-grained, and like the Tyee beds, highly micaceous. Exposures occur as sandstone cliffs and along stream beds and roadcuts. Cliffs of Flournoy sandstone, ranging up to 30 feet in height and representing one or more bedding units, are exposed for distances of as much as 2,000 feet.

The contact between Flournoy and Lookingglass beds is not observable anywhere, and only near Sitkum can the Flournoy and Tyee contact be located within a few feet. Both the upper and lower contacts of the Flournoy are placed by recognition of topographic changes and stratigraphic and structural trends. The lower Flournoy beds are much more resistant than the siltstones of the underlying Lookingglass, so that the contact between steep, cliffy Flournoy slopes and the gentler, more subdued Lookingglass slopes is observable in the field, on the aerial photos, and on the topographic maps. The same can be said of the Tyee and Flournoy contact, with the coarse basal Tyee sandstones overlying the Camas Valley Member siltstones and fine-grained sandstones, and a very apparent change in slope gradient at the contact. The contact between Flournoy and Tyee beds, which in the study area is disconformable, becomes unconformable when traced southward (Trigger, 1966).

Baldwin (1974) placed the Flournoy beds in the early to middle Eocene (Ulatisian age; Laiming, 1940, and Mallory, 1959), 51.5 to 47 million years.

Tyee Formation

The Tyee Formation was named by Diller (1898) for a sequence of rhythmically-bedded sandstone and sandy siltstone beds which underlie much of the southern and central Coast Range. The formation is approximately 5,000 feet in thickness and is exposed on the edges and core of a broad north-south trending syncline. The rocks grade laterally from coarse to fine, and from shallow-water, nonturbidite facies to turbidite facies in a northward direction (Chan and Dott, 1983; Lovell, 1969). The Klamath Mountains were considered to be the likely source of the sediments (Snavely et al, 1964), based on traditional studies of paleocurrent indicators and lithofacies. Isotopic studies by Heller et al (1985), however, point to the Idaho batholith as the source area, indicating significant displacement of the Coast Range block since deposition of the Tyee sediments.

Snavely et al (1964), Lovell (1969), and others described the Tyee beds as bluish gray to gray micaceous arkosic and lithic wackes, with a good deal of lithologic variability throughout the depositional basin. Lovell (1969) recognized a continental shelf to deeper water transition in the Tyee and placed the study area rocks in his continental shelf (ie., nonturbidite) zone. Chan and Dott's (1983) model of a linear Flournoy-Tyee basin, with sediments issuing from multiple sources along the shelf, also placed the study area rocks in the delta-shelf facies. Baldwin (1974) divided the Tyee into three members on the eastern flanks of the Coast Range, but did not extended this division westward to where the study area is located.

The Tyee beds in the study area, as well as the local Lookingglass and Flournoy beds, are on the southwestern limb of the broad syncline which occupies the core of the southern Coast Range. The local Tyee beds dip gently east to northeastward an average of 7°, and underlie the eastern 25% of the study area (Plate 2). The field work in the study area confirmed mapping by Baldwin (1974), with the addition of structural data and the refinement of contact placement.

Sandstone units in the Tyee are generally greater than 20 feet in thickness and are characterized by crossbedding having amplitudes of 8 inches or more. Interbedded siltstones are also found in the Tyee sandstone, but arkosic wacke sandstone is the dominant lithology. The basal Tyee sandstone, exposed very well at Brewster Rock and on the high ridge south of Sitkum, is a medium-grained, micaceous, arkosic wacke with about 10% well-rounded pebbles (quartz and volcanic rock) and occasionally abundant carbonaceous woody debris. Siltstones and some thin-bedded sandstones higher in the section show more planar bedding and deposition of woody debris in quieter waters. The thick, cliff-forming sandstone units are still characteristic, however, throughout the Tyee section and are especially apparent on hillsides cleared of the forest cover.

Both Laiming (1940) and Snavely et al (1964) placed the Tyee in the early to middle Eocene (Ulatisian age), 51.5 to 47 million years, on the basis of microfauna.

Quaternary Deposits

Pleistocene and Holocene age deposits overlie nearly all of the study area bedrock, and are significant elements of the geomorphology and geography as well as being reflective of its geology and history. They include alluvium, mass movement debris, lacustrine sediment, and soils.

Alluvium

Within the study area the valleys of Cherry Creek, Middle Creek, the North Fork Coquille River and the lower reaches of the East Fork Coquille River are all occupied by fluvially deposited Holocene silts, sands, and gravels (Baldwin, 1975). The physical appearance of the alluvial deposits is the result, for the most part, of post-Wisconsin eustatic sea level rise, drowning coastal valleys and raising the base level for the drainage system (Baldwin, 1981). Only a portion of the alluvium filling the stream valleys is potentially transportable by the present day streams. The low stream gradient and high base level prevent the streams from transporting their entire sediment loads. Instead the aggrading streams have developed floodplains across which the stream courses meander.

Higher gradient streams such as Yankee Run, Camas Creek, and the upper reaches of the East Fork, where it traverses the more resistant Tyee and Flournoy Formations, deposit sediment temporarily within the stream channel. These streams, however, are not aggrading; ie., they are not developing wide floodplains. Streamflow periodically transports materials stored temporarily in the stream bed, channel bars, and other sites and has the potential to scour the stream bed bedrock. Fine-grained sediment is being transported farther during individual transport events and spending less time in local storage sites than cobble or boulder-sized particles (Dietrich and Dunne, 1978).

The character of the East Fork Coquille River is determined largely by its sediment load and the mass movement features which locally impact streambed elevation and location. Selected characteristics of the East Fork channel are noted on Plates 1 and 2, with symbols and codes indicating the major causative factors of rapids along the stream course.

Terraces

A series of terraces have been developed within the study area along the lower reaches of the North Fork and East Fork Coquille Rivers where those streams meander across their floodplains (Plate 1). The terrace deposits of sand, silt and gravel along the streams are probably Pleistocene in age (Baldwin, 1975), and reflect fluvial adjustment to base level changes during the Pleistocene and Holocene. Eustatic changes in sea level due to worldwide glacial advances and retreats, combined with the slow rise of the Oregon coastline (McKee, 1972), have alternately submerged and elevated the coastal margin. As a result, marine terraces can be identified at elevations ranging from below sea level along the present coastline to 1600 feet at Blue Ridge, northeast of Coquille (Baldwin and Beaulieu, 1973). The post-Wisconsin rise in sea level, however, has drowned most of the coastal valleys, extending tidal influences far upstream.

The East Fork Coquille has responded to these base level changes by alternately eroding and alluviating its valley. Consequently a remnant series of fluvial terraces can be identified along the East Fork Coquille downstream from Dora. These terraces range in elevation from about 180 feet to 120 feet above sea level, corresponding to the elevation drop of the East Fork Coquille floodplain along its course (Figure 6).

The downstream trend in elevation of the fluvial terraces allows a probable correlation with a marine terrace at Bandon, 20 miles due west near the mouth of the Coquille River. It is named the Whiskey Run terrace by Griggs (1945) and can be identified along the coastline between Cape Arago and Cape Blanco. The surface of the Whiskey Run terrace has been warped by the continued uplift of the Coast Range, so that its elevation varies from just below sea level at Floras Lake to 200 feet above sea level at Cape Blanco (Baldwin, 1981). At Bandon the Whiskey Run terrace has an elevation of 60 to 80 feet. Baldwin (1981) reported carbon 14 dates of 31,000 to 35,000 years (B.P.) for the top of the Whiskey Run terrace, providing a gauge of subsequent tectonic disturbance of the coastline.

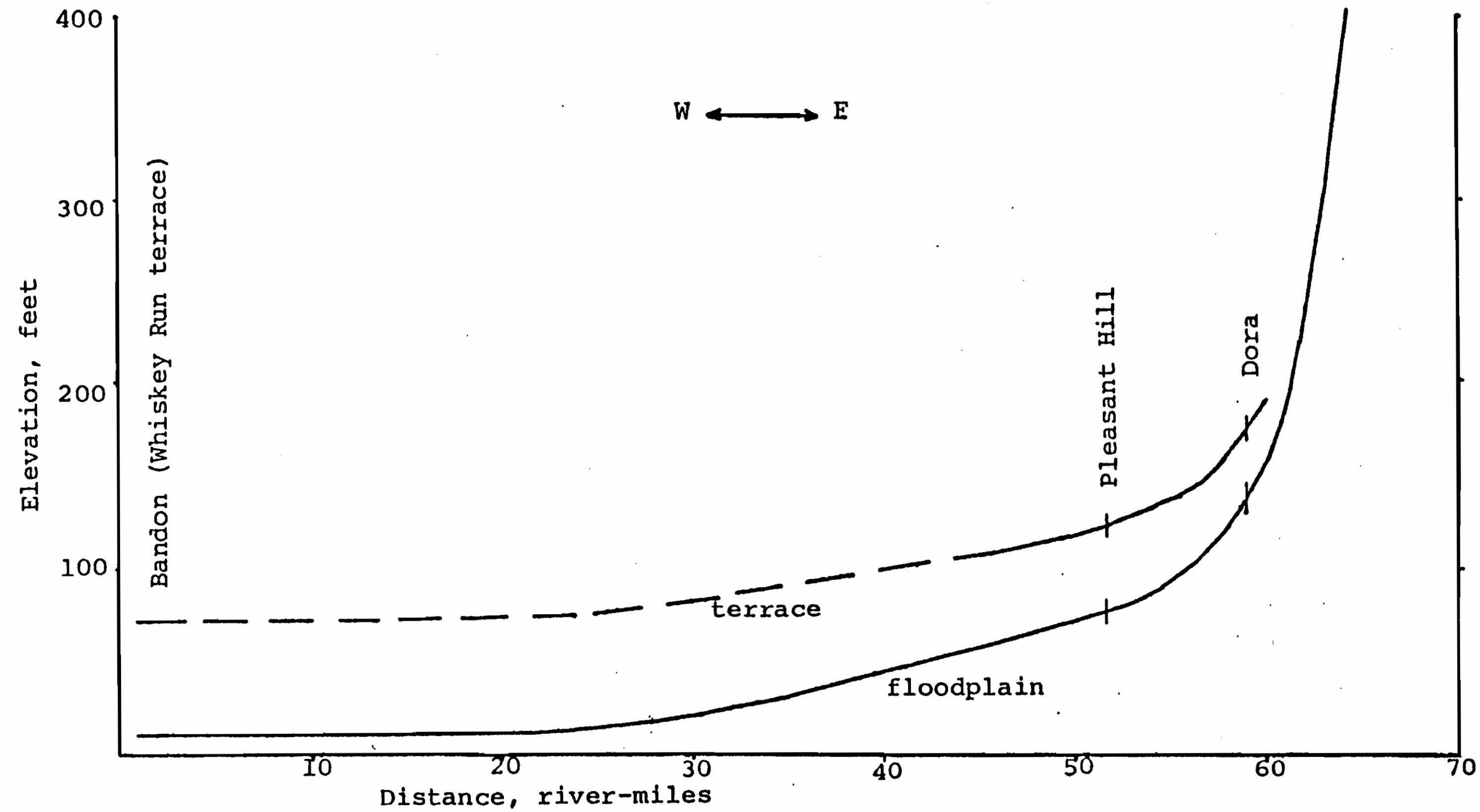


Figure 6. Comparison of longitudinal profiles of terrace and flood-plain along the Coquille River and the East Fork Coquille River. The terrace is not a continuous feature along the length of the rivers, as might be inferred from the figure.

Mass Movement Debris

Major mass movement features within the study area, as sites of Quaternary sediment transport and storage can be divided into four general classes: debris avalanches and torrents, rock slumps, slump-earthflows, and landslide complexes. They are defined and discussed in greater detail in the Mass Movement section of this thesis.

Debris avalanches and torrents scar hillslopes, erode stream channels, and are a pervasive element of rugged, forested landscapes, and are subject to accelerated frequency as a result of human activities (Harr and Yee, 1975). Organic and inorganic sediments transported by debris avalanches and torrents are generally delivered to the fluvial system and incorporated as channel or alluvial fan deposits (Dietrich and Dunne, 1978). Debris avalanches and torrents occur predominantly in the more rugged eastern half of the study area.

Rock slumps in the study area are recognizable as discrete, competent units of bedrock which have moved downslope with a rotational motion, having a tipped bench-like form below a headscarp (Varnes, 1978). Within the study area 7 rock slump sites are mapped, ranging in areal extent from approximately 60,000 to 520,000 square feet.

Slump-earthflow and landslide complex deposits vary greatly in the study area, from features involving only the soil mantle to large-scale bedrock slides. The slump-earthflow sites generally involve only the soil mantle and vegetative cover (Dietrich and

Dunne, 1978). Slump-earthflows have depressed, spoon-shaped form, with head and lateral slump scarps, narrowing downslope. Earthflow sites in the study area, with only one exception, are in the rugged eastern portion, in Flournoy and Tyee terranes.

Quaternary deposits mapped as landslide complexes are the most significant mass movement features in the study area, in terms of volume and environmental impact. Their precise mode of emplacement is undetermined, but failure was probably catastrophic. Five sites are mapped in the study area, ranging in areal extent from 0.07 to 1.54 square miles. None of the sites are currently active, based on absence of evidence of ground or vegetation disturbance. These areas are characterized by an irregular ground surface with internal scarps and disrupted drainage, a major head scarp from which the landslide material was transported, and a lower slope gradient than the surrounding terrain.

The textural and lithologic character of the landslide deposits depends on the bedrock material involved, but could be expected to consist of rock fragments of all sizes, plus the former overlying soil mantle and vegetation, in an unstratified and poorly consolidated deposit. In addition, the wet temperate climate and heavy vegetative growth facilitates deep weathering of the deposits. Primarily because of dense vegetation, only the Sitkum landslide (Plate 2) was accessible for extensive field investigation. Its large areal extent and numerous exposures in clear-cuts, roads, and pastureland allowed relatively detailed field and aerial photo mapping.

Figure 7. View northeast from SE 1/4 of Sec. 16, T. 28 S., R. 10 W.; Sitkum landslide deposit is in foreground, cliffs of basal Tyee sandstone in background, and lake sediment fill at intermediate distance. See site #51202 on Plate 2.

Lacustrine Deposits

The sediment fill which now occupies the former lake created by the Sitkum landslide (Plate 2) has two apparent origins (Baldwin, 1958; Adams, 1981). One is deposition of fine-grained suspended particles in the quiet waters as muddy, thin, horizontally bedded lacustrine sediments. The other is deposition of coarser-grained bedload and suspended material by streams flowing into the lake, as topset, foreset, and bottomset-bedded deltaic sediments at the mouths of streams. The lacustrine and deltaic facies should interfinger as sediment input varied through time and the deltas extended farther into the lake. They should also rest unconformably on the Flournoy and Tyee bedrock which forms the former lower valley walls, and on landslide debris at the western and southwestern margins.

Additional elements of the sedimentary fill are channel levees, floodplain deposits, channel deposits, and alluvial fans. The East Fork and Brummit Creek flood often during the winter months, depositing sediment on the agriculturally developed floodplain and the channel levees. Aerial photos document the former channels of the streams, showing natural re-routing of streams, including meander cutoffs. It would seem likely that the streams have been slowly re-working the entire alluvial surface, capping the delta/lacustrine sediments with a veneer of fluvial sediments as the landslide dam is lowered by fluvial erosion. The rate at which the East Fork can incise the dam is limited by the very coarse, resistant boulders which choke the channel (Figure 8), and by the low gradient of the

East Fork upstream from the landslide dam (Adams, 1981; Baldwin, 1958).

Short, steep-gradient streams from the surrounding hills have built alluvial fans on the margins of the former lake, particularly southwest and southeast of Sitkum (Plate 2; Figures 10 and 11). The alluvial fans are the subaerial expression of former deltas, but now as fans they interfinger with the floodplain deposits. Alluvial fans have developed elsewhere in the eastern portion of the study area where debris torrents in high gradient tributary streams deliver material to lower gradient streams. Rapids form along the East Fork at tributary-mainstream junctures, since some of the material delivered by tributary streams is too coarse for bedload transport by the main stream. Fan growth, however, is limited by high stream competence and narrow valley floors. In the western portion of the study area, where fans might form at channel junctions, the relief is apparently too subdued and the sediment too fine for fan development.

Figure 8. East Fork Coquille River course across Sitkum landslide deposit. Boulder near center has diameter of approximately 10 feet. See site #41208 on Plate 2.

Soils

Soil characteristics at any particular site are a function of the parent material, slope, precipitation and ground water circulation, aspect, and biological activity. The distribution of various soil types within the study area reflects primarily the distribution of bedrock or parent material types and the slope gradient. Residual soils, derived in situ from weathering of the parent material, develop best on stable sites such as flat ridgetops or fluvial terraces. Hillslope processes such as mass movement, important in the Coast Range due to steep slopes and high precipitation, transport soils downslope and mix residual soil types, forming colluvial soils. The predominance of steep slopes in the study area also preclude development of great soil depths or well-developed soil profiles, despite the temperate and humid climate (Corliss, 1973). Most of the study area soils are gravelly loams and silty clay loams (Baldwin, 1975; Harr and Yee, 1975, for Tyee sandstone in the central Coast Range).

Structure

Folding

Bedding attitudes of sandstone and siltstone units were measured throughout the area, including measurements which confirmed attitudes recorded by Magoon (1966) and Baldwin and Beaulieu (1973). These measurements illustrate strong folding and overturning of Roseburg strata, with attitudes striking northeast-southwest; and gentle eastward dipping of the Lookingglass, Flournoy, and Tyee strata with attitudes striking north-south. Dips tend to decrease upsection, from Lookingglass through Tyee, from a maximum of 30° to nearly horizontal.

Faulting

Baldwin and Beaulieu (1973) mapped three faults within the study area, one thrust fault and two of unspecified type. These faults are restricted to Roseburg and Otter Point strata and have a northeast-southwest trend. They are continuations of faults mapped to the south and southwest of the study area, and trend parallel to the general pattern of faulting in the southern Coast Range and northern Klamaths (Baldwin, 1974). Thrust faulting in the region appears to be restricted to Roseburg and pre-Tertiary strata, dating the faults as probably early Eocene (post-Roseburg, pre-Lookingglass). Normal and reverse faults, also with a northeast-southwest trend, disrupt strata no younger than the Flournoy Formation. All of the faults appear to have upthrown southeastern blocks, indicating a northwest-southeast axis of compression during the early Eocene; perhaps due to southeasterly-plunging subduction of nearby oceanic crust (Baldwin, 1974). Post-Roseburg compression rotated to a west-east axis, and gentle folding and uplift gradually replaced faulting.

MASS MOVEMENTS

Classification of Mass Movements

Mass movement features result from several factors acting together to produce instability and movement of soil and/or bedrock. The form of such mass movements is varied, depending on the interaction and relative importance of factors contributing to instability; ie., slope, pre-cipitation, bedrock structure, etc. Varnes (1978) classified mass movements by form and behavior, based on (1) dominant type of movement, (2) rate of movement, and (3) type of material involved. Mass movement processes occurring within the East Fork Coquille watershed include creep, rockfall, debris avalanche, debris torrent, slump-earthflow, rock slump, and landslide complex.

Creep

The large annual rainfall, mild temperatures, and lush vegetation of the Coast Range are conducive to formation of a soil mantle over the bedrock. The steep slopes typical of the Coast Range, however, tend to preclude thick, well developed soils (Corliss, 1973), because of the pervasiveness of gravity-induced soil creep and other erosion processes. Creep occurs as slow, quasi-viscous movement downslope in response to constant gravitational stress. For any particular soil the steeper the slope of the ground surface, the greater the gravitional stress and potential for creep (Swanston, 1981). Animal burrowing, frost heaving, root throw, and other processes also contribute to downslope movement of the upper soil column.

The shear strength of a soil generally increases with depth so that the degree of transport by soil creep decreases with depth (Carson and Kirkby, 1972). The base of the creep zone is generally gradational, without formation of a discrete shear plan, and may be at a depth of several feet. Shear stresses acting upon the soil column may, however, lead to failure along discrete shear planes, with soil creep grading into more disruptive mass movement activity.

Creep is believed to occur throughout the study area on all soil mantled slopes. It is an important source of sediment for streams, either directly by bank encroachment on channels or by other mass movement forms, such as debris abalanches and slump-earthflows. Soil creep is most often recognized by bowed or leaning conifers on hillslopes, and by disruption of road surfaces.

Rockfall

Varnes (1978) defined rockfalls as rapid descent of bedrock material by freefall, leaping, bounding, or rolling. The mass of rock involved in the fall is detached from exposed bedrock along a surface with little or no shear displacement involved in the movement. The exposure of steep bedrock slopes, the nature and orientation of the bedding units, and the nature and orientation of tectonic fractures in the bedrock largely determine the occurrence of rockfalls. As weathering or mechanical processes enlarge fractures or weaknesses in the bedrock, the stability of a rock fragment will decrease and a fall may occur, depending on the orientation of the separation surface relative to the slope (Carson and Kirkby, 1972).

The cliffs of resistant sandstone and conglomerate in the Lookingglass, Flournoy, and Tyee Formations are sites of rockfall within the study area. Few of the cliffs, principally resistant sandstone units, are of sufficient height, however, to accumulate significant talus slopes at the cliff base. Instead, rockfall debris is incorporated into the colluvial forest soil downslope from the cliff base, or is delivered directly into stream channels where it contributes to development of rapids in the streams. The latter case is particularly evident along the East Fork Coquille where it flows beneath cliffs of Flournoy and Tyee sandstone. Fluvial erosion and creep processes maintain cliffs by removing talus, and softer material underlying cliff-forming units, from the cliff base. Where rockfall occurs near the head of a drainage it may transform into a debris avalanche.

Debris Avalanche

A complete gradation can be recognized between debris slides, debris flows, and debris avalanches. Distinctions are difficult to define in the field (Blong, 1973) so that all are designated by the term debris avalanche in this discussion.

Debris avalanches are characterized by rapid, shallow translational movement of rock, soil, and associated organic material (Varnes, 1978) with stages of failure, transport, and deposition. Rock material may or may not be involved in the initial failure of the debris avalanche or may be included during the erosive stage of transport. Organic material, particularly standing and downed timber, also may or may not be included in the initial debris avalanche or may be incorporated during transport of the debris.

Water content contributes to the fluid behavior of moving masses; pore water pressures are important elements of the failure and transport processes. Failure is generally progressive, though rapid, and involves cohesionless materials. A distinct failure surface, or surfaces, separates the moving debris from the underlying resistant stratum which may be scoured down to bedrock. Debris avalanches commonly have a long, narrow form, an upper eroded scar area, and a lower area of debris deposition, unless removed by erosion (Dietrich and Dunne, 1978).

Within the study area debris avalanches are located on steep slopes and occur in colluvial soils, road sidecast materials, and piled organic debris, particularly on steep drainages within recently logged areas. Debris avalanche sites are recognized both as recent movements and as scars from past activity, on both managed and undisturbed lands. Other mass movement features may grade into or contribute to debris avalanches, which in turn may contribute to stream sediment loads or alluvial fan development (Dietrich and Dunne, 1978).

Debris Torrent

Debris torrents are very rapid, water saturated mass movements of organic debris, rock, and soil, initiated and confined primarily to high gradient stream channels. Most debris torrents involve the movement of debris avalanches from slopes to channels (Swanson and Lienkaemper, 1978). They may also result when the toes of slump-earthflows deliver material to a stream channel, creating a dam of organic debris and sediment which fails and sends a debris-laden flood surge downstream. Logging or road building practices which leave unstable material in channels or on nearby slopes may also lead to development of debris torrents through formation of debris dams or other mechanisms. The failure of these dams, during or after heavy precipitation, releases a slurry of water and debris which may scour the stream channel down to bedrock. The subsequent redeposition of materials occurs in alluvial fans or as sediment and organic debris dams in the channel.

Debris torrents may leave their channels, especially at abrupt changes in channel direction, eroding adjacent slopes or depositing debris. Most debris torrents in the mountainous terrain of western Oregon are the result of debris avalanche activity (Swanson, Leinkaemper, and Sedell, 1976). They severely impact stream channel habitats, soil cover, and riparian vegetation and may destroy bridges, homes, and other man-made features (Swanston and Swanson, 1976).

Slump-Earthflow

Varnes (1978) made a distinction between slump and earthflow mass movements, but also recognized that the two processes can occur together as a single mass movement feature. Slump-earthflows are characterized by an upper area of rotational failure with main and lateral scarps, and a lower area of earthflow activity which narrows as it moves downslope, from relatively large blocks of mass movement debris to smaller and smaller units nearer the foot of the feature. The mass moves slowly as it breaks up into smaller parts with variable orientation and motion (Varnes, 1978).

Mapped, active slump-earthflows within the study area are recognized by main and lateral scarps exposing bare mineral soil and rock, by the surface appearance of the moving mass, and by disruption of vegetation. Trees growing on active slump-earthflow sites show unusual stem curvature, and may lean in random directions.

Slump-earthflows in the study area are much smaller than those described in the western Cascade Mountains (Hicks, 1982), and lack grabens or surface drainage development. They are generally less than 1 acre in areal extent.

Rock Slump

Varnes (1978) used the term rock slump to describe bedrock failures which move about a horizontal axis of rotation, at extremely slow to moderate velocities, with movement by shear displacement along a descrete surface or surfaces. The intersection of the displacement surface with the ground surface indicates a line of rupture which is convex when viewed in a downslope direction. The rock slump generally moves as a single competent unit with a bench-like surface that slopes at a lower angle than the surrounding hill-side, or slopes back into the hillside.

Within the study area, rock slumps are easily identified by their topographic expression: bench-like form, lower slope angle than the adjacent hiddside, and concave headscarp. All of the rock slumps in the study area appear as discrete entities not in direct association with other mass movement features, and all lack evidence of current or recent activity. They range in areal extent from 1.5 to 12 acres.

Complex Landslides

Single mass movement features may consist of elements having different forms of movement, velocities, and different types of materials involved. Such features are complex in their development and are herein described as complex landslides. The form of movement is the primary feature of concern; Varnes (1978) defines a complex as a mass movement feature that exhibits more than one form of move-ment. Features such as slump-earthflows, exhibiting both rotational and translational movements, are complex.

Complex landslides in the study area include those features which show evidence of multiple modes and multiple phases of movement. These are the most extensive of the mass movement features in the study area, with mappable surface features showing internal scarps, patterns, and somewhat variable direction of movement. Slump-earthflows are excluded from this category because of their distinctive form and occurrence in the study area. Some of the complex landslides are apparently of a catastrophic origin, though more than one episode of failure may have occurred at each feature site. None of the sites, however, are active at present.

Mass Movement Distribution in the Study Area

The mode and distribution of mass movement activity is not uniform throughout the study area. Slope aspect, slope angle, parent bedrock lithology, stratigraphy, structure, and land management activities influence the susceptibility of slopes to soil or bedrock failure. The variability of these factors produces variability in the distribution of mass movement features over a given area. While occurrence is not uniform it is fairly systematic, so that the controlling factors can be identified, particularly with respect to geology and land management activities.

Creep

In general terms the occurrence or non-occurrence of soil creep is dependent on the presence of a soil column and the angle of the hillslope surface. For soils of similar character and attitude an increase in slope angle is believed to result in greater creep activity. Other factors, such as soil hydrology, vegetation and root strength, and micro-climate, also have to be considered when studying soil creep impact on local hillslope form (Swanston, 1981). Direct observation and measurement of individual factors, however, is difficult and not well documented.

For the purpose of this study, creep is considered to occur on all hillslopes, and to be relatively more active on steeper hillslopes. It can therefore be expected that creep is more pronounced in the portion of the study area underlain by Flournoy and Tyee strata because of the steeper terrain. The predominantly sandstone lithology of the Flournoy and Tyee Formations, and the 10-20° eastward dip of the bedrock, produces high relief and steep hillslopes, particularly on the obsequent slopes of cuesta ridges.

In contrast the western portion of the study area is underlain by the predominantly siltstone lithology of the Roseburg and Lookingglass Formations, and is characterized by relatively low relief and gentler slopes. While soil creep may be less active because of the low slope angles, its relative importance as a mass movement process is high because the low relief precludes significant debris avalanche and landslide activity.

The influence of land management activities on soil creep has not been well documented, and their impact in the study area is unknown. There is speculation, however, that such activities as road building and clear-cut logging may alter the hydrologic cycle on hillslopes and thus alter the soil creep rate (Swanston, 1981; Gray, 1970).

Rockfall

Rockfall occurs at several sites in the study area where bedrock units are exposed as cliffs. Sandstone units in the Tyee Formation, lower Flournoy Formation, and conglomerates in the Lookingglass Formation form cliffs, mostly in the eastern portion of the study area. The eastward dip of the Flournoy and Tyee bedrock results in major cliff exposures occurring on the western slopes of cuesta ridges (Plates 1 and 2). Major cliffs also occur on hillslopes above the East Fork Coquille River where its valley cuts across the Flournoy and Tyee bedrock, and where fluvial erosion maintains slope steepness.

The most notable rockfall sites in the study area are at Brewster Rock (basal Tyee sandstone) and Pleasant Hill Ridge (basal Lookingglass conglomeratic sandstone). Rockfall undoubtedly occurs at all cliff sites, but development of significant talus slopes is limited to these two locations. The thickness of the resistant sandstone units and the rates of rockfall activity are insufficient to overcome weathering rates and vegetation growth. As a result, talus debris is overgrown by vegetation and incorporated into colluvial soils downslope of cliffs. The rockfall at Pleasant Hill is maintained, in part, by fluvial erosion at the hillslope base, while the Brewster Rock rockfall site is maintained by the great thickness of the basal Tyee sandstone and the large volume of resulting rockfall debris (Plate 1, site #50601; Plate 2, site #31205). Both sites also occur on relatively dry southern exposures, which may inhibit chemical weathering somewhat, but might contribute to rapid physical weathering.

Debris Avalanches and Debris Torrents

The distinction between debris avalanches and debris torrents is not always apparent in the field, since debris avalanches often enter stream channels and behave as debris torrents, so the two processes were catalogued and analyzed together in this study. Distinct debris torrents were difficult to identify on aerial photographs of the study area and were not observed except in association with debris avalanches.

A total of 45 debris avalanche/torrent sites were catalogued in the study area, ranging in age from less than 1 year to about 14 years (1967-1981). Data for making age determinations, such as sequential aerial photo series, were unavailable so the decision to include sites was subjective, based primarily on vegetation regrowth. Because of rugged terrain and dense vegetation, 31 of the 45 sites were catalogued by distant field observation and by interpretation of aerial photos (1967 ASCS black and white, 1:20,000).

Debris avalanche/torrent sites occur primarily in the resistant sandstone units in the eastern portion of the study area. Three-fourths of the 45 catalogued sites are in Tyee Formation terrane, and density of sites in Tyee terrane is about 10 times that of Roseburg/Otter Point terrane (Table 1). The predominant occurrence of sites in the Tyee Formation terrane can be explained by the high relief and steep slopes typical of the areas underlain by Tyee sandstone. The siltstones which comprise much of the Roseburg, Lookingglass, and Flournoy strata are susceptible to weathering and erosion processes other than debris avalanches and torrents.

The relationship between debris avalanche/torrent occurrence and hillslope gradient is shown more clearly in Table 2. The land area in each gradient classification, and each aspect classification (Table 3), was calculated by sampling slope gradients and slope aspects with a dot grid (quarter section corners) over the study area base map, excluding terrace and alluvial deposits along the stream valleys. From the grid measurements percentages were calculated for each classification, and then the inferred total area for each classification.

The frequency of debris avalanches/torrents increases with increasing slope gradients (Table 2). The Table also shows the distribution of slope gradients among the different geologic units in the study area. The Tyee and Flournoy Formations, which are predominantly sandstone, have much greater percentages of land area with steeper slopes than do the Roseburg and Lookingglass formations. Consequently the frequency of debris avalanche/ torrent occurrence is higher in the Flournoy and Tyee. The apparent relationship is that more resistant bedrock units have steeper slopes over a greater area and, therefore, greater frequency of slope failure.

The influence of geologic structure on the distribution of debris avalanche/torrent sites in the study area is associated closely with that of bedrock lithology. Uplift of the region is irregular drainage

terrane	sites	area of study (sq.mi.)	areal frequency (sites/sq.mi.)	annual frequency (events/ sq.mi./yr)
Roseburg Fm.	4	21.0	0.19	0.014
Lookingglass Fm.	1	8.5	0.12	0.008
Flournoy Fm.	6	13.75	0.44	0.031
Tyee Fm.	34	14.25	2.38	0.170
total	45	57.5	0.78	0.056

Table 1. Distribution and frequency of debris avalanche/torrent sites catalogued in the study area. The area of each terrane was derived from the base maps and includes areas covered by mass movement deposits. Land areas covered by major alluvial/terrace/lacustrine deposits are not included in the calculations on Tables 1-4. Failures occurred within the 1967-1981 period.

slope gradient (%)	area (sq.mi.) per gradient class. Roseburg	Lookingglass	Flournoy	Tyee	total	sites per slope class	areal frequency (sites/ sq.mi.)
<40	14.3	6.4	4.4	2.45	27.6		
41-50	3.1	1.5	2.75	3.85	11.2		
51-60	1.3	0.4	2.75	3.15	7.6	3	0.4
61-70	1.9	0.2	1.8	1.4	5.3	12	2.3
71-80	0.4		1.2	2.0	3.6	13	3.6
81-90			0.4	1.0	1.4	8	5.7
>91			0.4	0.4	0.8	9	11.2
total	21.0	8.5	13.7	14.25	57.5	45	

Table 2. Distribution of slope area (square miles) by slope gradient, and frequency of debris avalanche/torrent sites for each gradient classification.

slope aspect	area (sq.mi.) per aspect class.					sites per aspect class.	frequency (sites/ sq.mi.)
	Roseburg	Lookingglass	Flournoy	Tyee	total		
North	6.5	2.1	4.4	4.7	17.7	17	1.0
West	6.1	2.0	3.45	3.7	15.2	21	1.4
South	4.6	2.1	1.9	3.45	12.1	5	0.4
East	3.8	2.3	4.0	2.4	12.5	2	0.2
total	21.0	8.5	13.75	14.25	57.5	45	

Table 3. Distribution of slope area (square miles) by slope aspect, and frequency of debris avalanche/torrent sites for each aspect classification.

responsible for the mountainous terrain, and the location of the study area on the west limb of the Coast Range syncline results in ridges having a cuesta form. The tabular bedrock units, dipping east and northeastward, produce steep west-aspect slopes and gentler east-aspect dip slopes. Consequently the frequency of debris avalanche/ torrent sites on slopes with a west-aspect (bearing 45° west to 135° west) is approximately seven times the frequency of sites on the east-aspect slopes (Table 3). Table 3 also shows the irregular distribution of land area within each slope aspect and the individual geologic units, a function of the geologic structure of the study area.

Geologic structure may also be responsible for the relatively high frequency of sites on north-aspect slopes. Groundwater movement along the bedrock dip slope in a north-eastward direction would keep north-aspect slopes wetter. The true impact of this hydrological factor is uncertain, however. South-facing slopes tend to be drier, and in the study area the frequency of debris avalanche/ torrent sites on south-facing slopes is only 40% of the frequency on north-facing slopes. Dyrness (1967) documented the relationship between aspect and slope failure for the 1964-1965 winter storms in the Andrews Experimental Forest in the western Cascades and noted very little activity on southern aspects. He speculated that drier slope aspects, because of lower rates of rock weathering and soil formation, were more stable. Slopes with deep soils and well-weathered bedrock, such as on northern aspects, are apparently more susceptible to mass movement activity.

Such land use factors as logging and road construction may also be important for individual slope failure sites. In general terms logging and road construction, the principal soil disturbances by management, have produced different frequencies of slope failure in the study area (Table 4). Although roads occupy only 2% of the study area they account for 11% of the inventoried debris avalanche/ torrent sites. Logged slopes, harvested during the 1967-1981 period, occupy 13% of the area and account for 36% of the sites.

Areas classified as "undisturbed" include clear-cuts judged to be greater than 10 years old, where the timber removal appears not to have had any influence on current slope failure. These undisturbed lands, 85% of the study area, account for 53% of the inventoried failure sites. Slope failure on these lands can not be attributed to any apparent management activities. The lack of sequential aerial photos of the study area prevents a more concise comparison of failure frequencies for roads, clear-cuts, and undisturbed slopes.

However, studies by Amaranthus et al (1985), Swanson et al (1981), and Gresswell et al (1979) also show disparate frequencies of slope failure depending on the source area conditions. Differences in local geology, road-building methods, and logging practices influence the relative frequencies in the studies, though all show the highest frequency of slope failure on hillslopes disturbed by road-building and logging.

source area of failure	area (sq.mi.)	sites	frequency (sites/sq.mi.)
clear-cut (< 10 years)	7.7	16	2.1
road-fill	1.2	5	4.2
undisturbed (+ clear-cut >10 years in age)	48.6	24	0.5
total	57.5	45	0.78

Table 4. Distribution of debris avalanche/torrent sites with respect to source area of failure.

Figure 9. Debris torrent in Tyee Formation terrane, north aspect slope (site #41701 on Plate 2). Photo taken August, 1981; failure probably occurred the previous winter.

Slump-Earthflows

There are 12 slump-earthflow sites in the study area, ranging in area from 0.25 to 0.5 acres. All of the sites occur in Flournoy and Tyee terrane, with one exception in Otter Point and Roseburg terrane in the southwest corner of the study area. The mudstone component of the Otter Point Formation is reported to be particularly susceptible to mass movement (Baldwin, 1975; Swanson, 1975), although in the study area the extent of Otter Point outcrop is too small to note this characteristic.

The majority of the sites, however, occur on hillslopes underlain by sandstone and siltstone bedrock in the eastern portion of the study area, and are characterized by two different settings. Five of the slump-earthflows occur on west-aspect hillslopes, the obsequent slope of the cuesta ridges. These all appear to be of natural origin, and have a slope angle of approximately 30°. The other group includes 6 slump-earthflow sites occurring on slopes of approximately 20° and east to northeast aspect.

The slump-earthflows with westerly aspect occur primarily along contacts between siltstone and overlying sandstone beds. The over-steepening of the slopes by the resistant sandstone apparently set the stage for failure. For the slump-earthflows of easterly aspect, however, other factors are involved. At 5 of the 6 sites, tractor logging activity has disturbed the area and road materials are involved in the earthflow. All 6 of the sites are characterized by very wet ground and/or springs. The location of the slump-earthflows on the dip slope of ridges, assuming hydrologic movement of groundwater eastward down the dip slope, is possibly a factor in their development.

Rock Slumps

Bedrock slumps occur at 8 sites in the study area, distributed equally between the Roseburg and Flournoy strata. The 4 failures of Roseburg bedrock range in areal extent from 0.25 to 10 acres; failures of Flournoy bedrock range from 1.5 to 10 acres. In both areas the rock slumps occur in interbedded sandstone and siltstone, though structural differences are apparent in the distribution of the features. They all appear to be simple rock slumps, as defined earlier, without involvement of rock topple.

The siltstone bedrock of the Lookingglass Formation is apparently not competent or resistant enough to fail as slump units. In contrast the very resistant and massive sandstone beds of the Tyee Formation tend to fail catastrophically by debris avalanches. The strata of the Roseburg and Flournoy Formations, however, are susceptible to rock slump failure, apparently because of the stratigraphic succession of relatively thin sandstone and siltstone beds.

Structural trends in the Roseburg and Flournoy strata appear to be at least partially responsible for the occurrence of rock slumps in the two areas. Fold axes in the Roseburg Formation are generally aligned in a northeast-southwest direction, with bedrock planes dipping southeast and northwest. As a result the rock slump sites in Roseburg terrane occur on slopes with a north, northeast, or east aspect. In the Flournoy Formation, which dips uniformly east-northeast, the rock slump sites all occur on slopes with south or east aspect.

The structural reasons for the non-random occurrence of the rock slumps, however, do not seem conclusive. Other factors, such as hydrology, slope steepness, and stream erosion, may influence the distribution. The small sample of such features in the study area limits the scope of inference about these apparent trends.

Land management influence on the occurrence and distribution of the rock slumps in the study area is not apparent. All of the features appear to be inactive, including 5 sites where timber removal has occurred, so they likely have a natural origin pre-dating any management activity.

Complex Landslides

Mass movement features at 5 locations in the study area are classified as complex landslides, based on apparent multiple forms of movement and multiple failure units. The sites are all distinct topographic features, and are attributable to stratigraphic and structural geologic factors. Land management activities, such as logging and road building, appear to post-date all of the landslide failures and to present no current threat of reactivation of the sites.

Two of the complex landslide sites occur along the lower course of the East Fork Coquille River and involve the failure of fluvial terraces and adjacent siltstone bedrock (Plate 2, sites #40904 and #40702). The larger site, about 65 acres in area, occurs in Roseburg Formation siltstone and terrace alluvium; the smaller site covers about 50 acres and occurs in Lookingglass Formation siltstone and terrace alluvium. Both sites are evidenced by scarps, hummocky and slumped ground, and discontinuity with adjacent terraces.

It is likely that failure occurred when streambank erosion by the East Fork oversteepened or undercut the valley walls. The Roseburg and Lookingglass siltstones weather easily and might be expected to fail in a slump or earthflow fashion under such circumstances. Involvement of the terrace materials would be secondary to that of the siltstone bedrock.

A third landslide site occurs near the center of the study area and involves Lookingglass Formation and basal Flournoy Formation bedrock (Plate 2, site #30901). The feature, at the Lookingglass/ Flournoy contact, covers about 45 acres and is deposited partially on a terrace of the East Fork Coquille River. Failure probably occurred because of oversteepening of the hillslope by the resistant basal Flournoy Formation sandstone. Both the Lookingglass and Flournoy strata dip in an east or east-northeast direction here, so that the Flournoy sandstone caps a relatively steep cuesta slope underlain by Lookingglass siltstone.

The two largest complex landslide sites occur in the eastern portion of the study area, near Sitkum and Brummit Creek. The

Brummit Creek slide has an areal extent of about 0.3 square mile, with the West Fork of Brummit Creek flowing along the base of the slide in an eastward arc. The slide headscarp, the apparent lack of internal disruption, and the location on the dip slope of the eastward-dipping Tyee Formation bedrock suggest that the landslide failed as a block glide (Varnes, 1978) of Tyee sandstone. The competent sandstone bedrock, with siltstone interbeds possibly acting as failure planes, might have moved downslope at an unknown rate, displacing the West Fork of Brummit Creek eastward. However, the lack of field access and bedrock exposure prevent confirmation of the form of failure. In addition, such large block glide failures are not documented elsewhere in Tyee Formation bedrock.

The Sitkum landslide is the largest mass movement feature in the study area, having a surface area of about 1.54 square miles (Figures 7, 10, and 11). The slide deposit consists of Flournoy Formation siltstone and sandstone, and basal Tyee sandstone, which failed on a north-aspect slope and extends nearly 2 miles from headscarp to toe. The undisturbed Flournoy bedrock here strikes N 15° W and dips 12° NE, and the overlying Tyee beds strike N 5° W and dip 13° NE, so that the landslide failed obliquely in a downdip direction. The failure plane or planes are in Flournoy bedrock, but whether the failure surfaces coincide with bedrock planes is unknown.

An engineering report on a similar landslide in New Zealand, with similar stratigraphy and geologic structure, showed failure occurring as a catastrophic blockglide (Read, 1979). The New Zealand slide dammed a major river and moved a short distance down the drainage

Figure 10. Panorama of the Sitkum landslide, view south from Brewster Rock (Sec. 4, T. 28 S., R. 10 W.; site #31205 on Plate 2). Refer to Figure 11 for a diagram of the panorama.

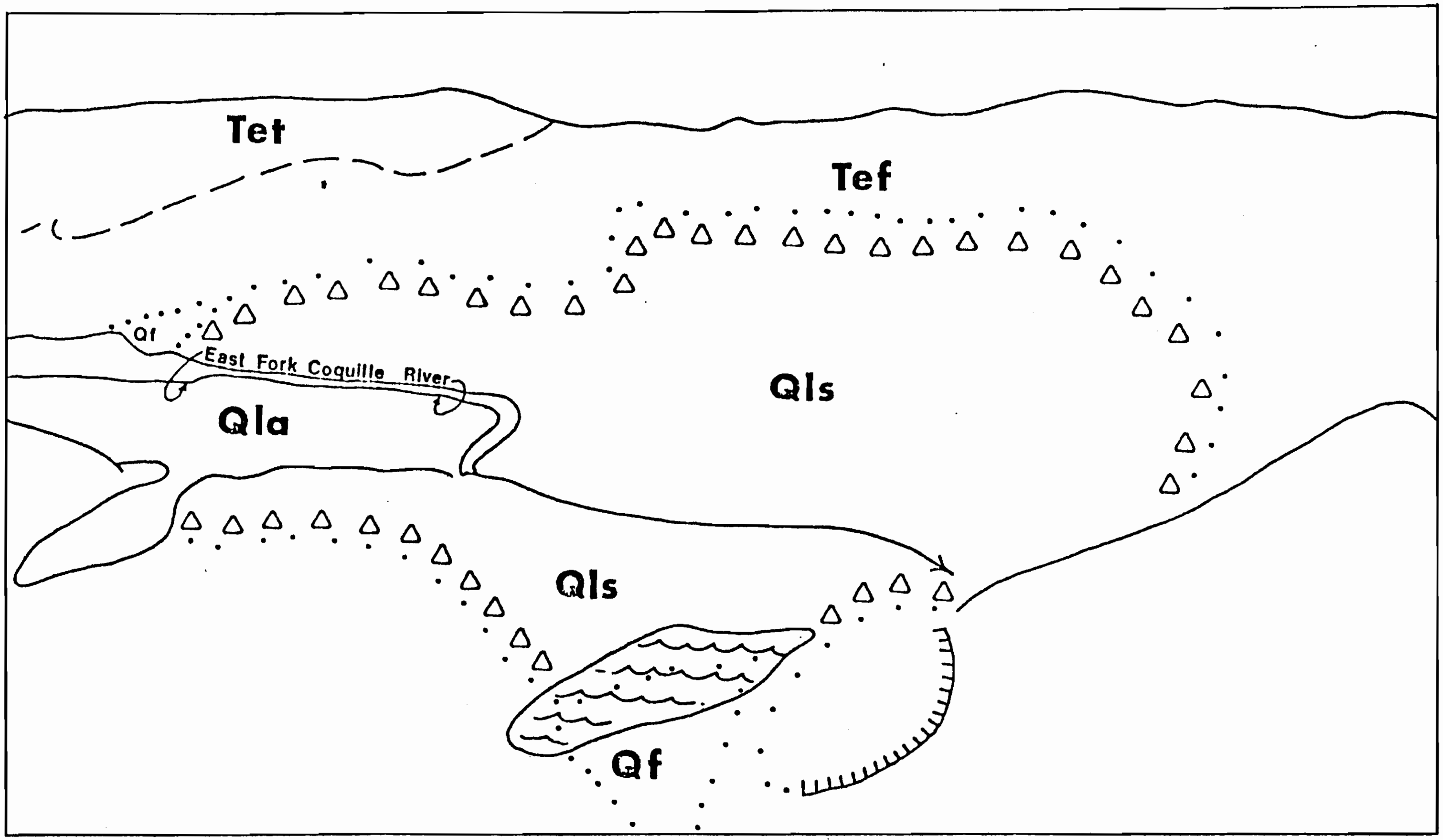


Figure 11. Diagram of the Sitkum landslide, as viewed in Figure 10. Tet= Tyee Fm., Tef= Flournoy Fm., Qls= landslide debris, Qla= lacustrine sedi-ments, Qf= alluvial fans. The East Fork Coquille River is shown flowing left to right (east to west) across the lacustrine sediments and landslide debris.

course as a debris flow. The Sitkum slide also dammed a stream, the East Fork, but it does not appear to have traveled downstream. The form of failure is uncertain, though apparently there was enough momentum involved to carry slide debris, including basal Tyee sandstone, across the former valley floor and approximately 150 feet up the north valley wall.

The landslide dam now has an estimated height of 280 feet and the landslide toe has an elevation approximately 150 feet higher than the original dam crest, based on stream and landslide toe profiles (Figures 12 and 13). The original dam crest was probably about 350 to 400 feet high, at the level of the landslide surface which now flanks the East Fork Coquille River. The East Fork cuts across the landslide toe and drops 270 feet over a 0.9 mile boulder-filled course (Figure 8).

The Sitkum landslide has an estimated volume of 1.29×10^{10} cubic feet, and originally impounded a lake estimated at 5.32×10^{9} cubic feet in volume. The lake extended nearly three miles up the valleys of the East Fork Coquille and Brummit Creek, and it probably had a maximum depth of about 300 feet prior to incision of the landslide dam and infilling of the lake. Much of the former valley floor beneath the lake-fill is probably covered by landslide debris to a considerable depth.

The geologic influence on the occurrence of the Sitkum landslide is evident by the attitudes of the Flournoy and Tyee bedding units which are here exposed on the southwest limb of the Coast Range syncline. The massive basal Tyee sandstone apparently formerly capped

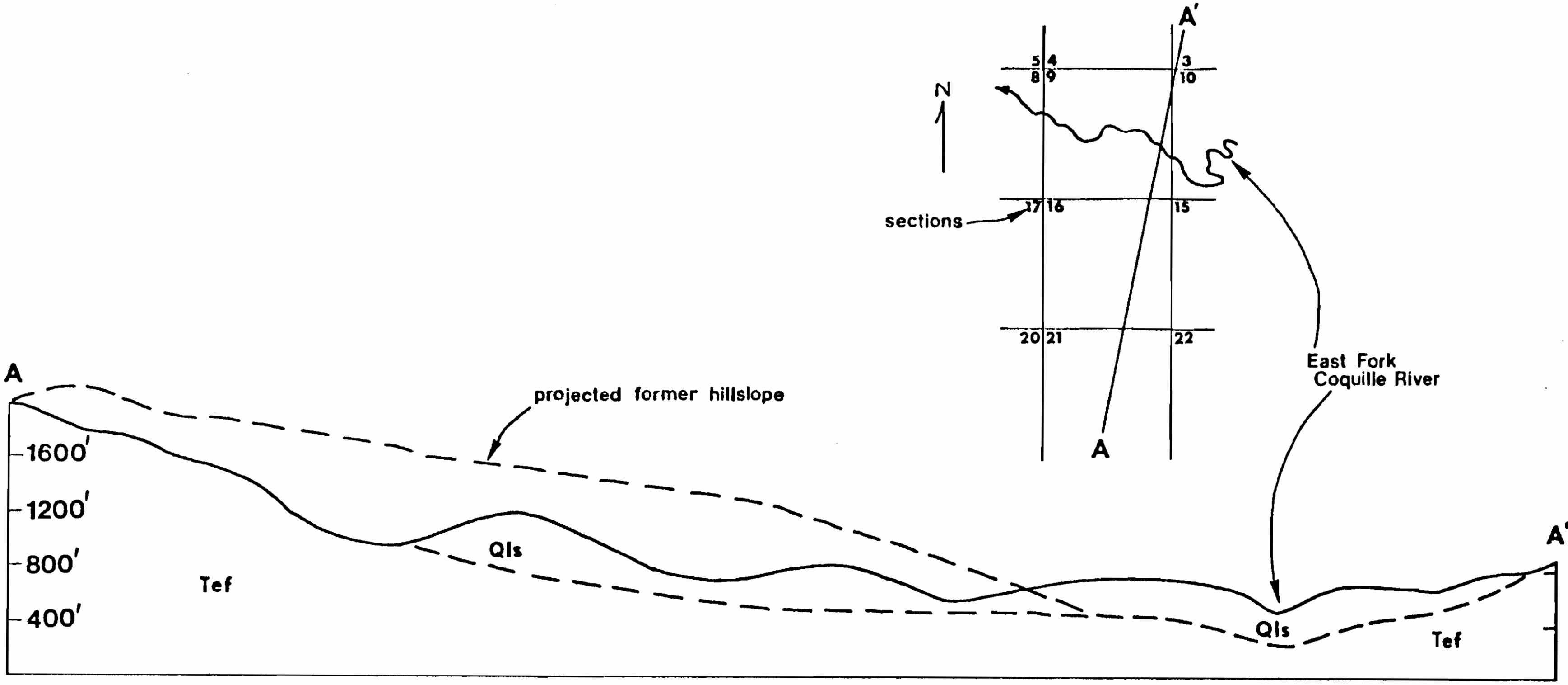


Figure 12. Cross-section of the Sitkum landslide, with reference to planimetric index (see Plate 2). The direction from A to A' is from south to north. The former hillslope is projected from adjacent undisturbed hillslopes. Qls= landslide debris, Tef= Flournoy Fm.

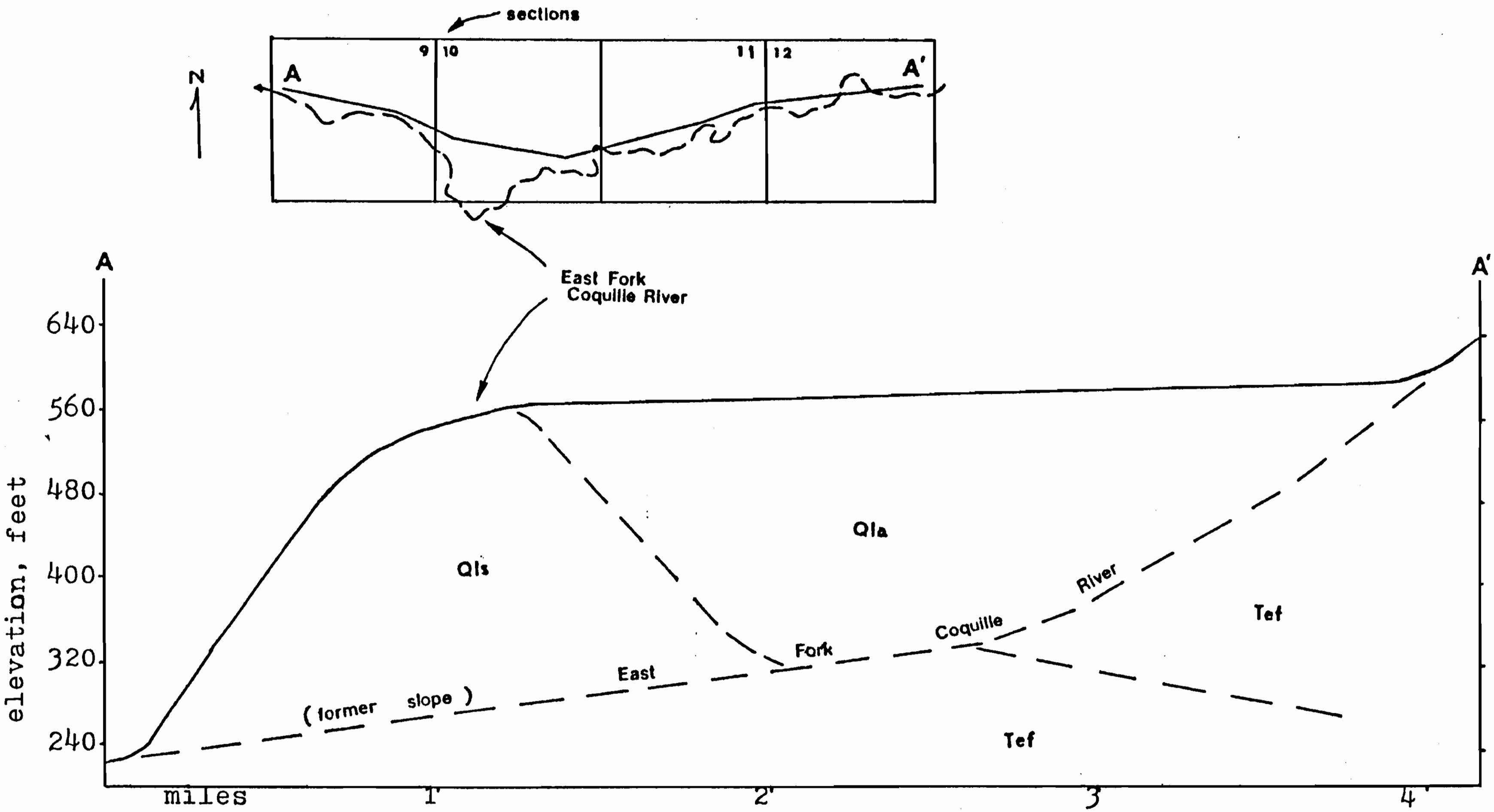


Figure 13. Profile of Sitkum landslide dam and lake-fill, through sections 9, 10, 11, and 12, T. 28 S., R. 10 W. Qls= landslide deposit, Qla= lake-fill, Tef= Flournoy Fm., and Tet= Tyee Fm.

the ridge crest, overlying interbedded siltstones and sandstones of the Flournoy Formation. The resistant Tyee sandstone probably led to oversteepening of the former hillslope and now resists erosion of the landslide dam. Blocks of Tyee sandstone, ranging up to 10 feet in diameter, are exposed in the center of the landslide deposit, the East Fork's stream course across the landslide dam, and the landslide margin north of the East Fork.

There are no radiocarbon dates for the Sitkum land-slide available. However, the Loon Lake slide, 30 miles north of Sitkum, was dated by radiocarbon methods at 1460 years B.P. (Baldwin, 1981). This date, combined with calculations of sedimentation rates for the Loon Lake basin, when applied to the Sitkum landslide, gives a minimum age of 3125 years B.P. for the Sitkum landslide, based on an estimate of the time required to fill the former lake impoundment with sediment (see Appendix A).

LANDSLIDE LAKES

Mass movement activity in the Oregon Coast Range is characterized by debris avalanches/torrents (Swanston and Swanson, 1977; Gresswell et al, 1979), by block glide failures (Graham, 1986) and by large bedrock landslides which effectively obstruct stream drainages and result in the formation of lakes. Landslide-dammed lakes typically occur in areas of the Coast Range underlain by resistant sandstone bedrock, such as the Flournoy and Tyee Formations. The bedrock contributes to the formation of landslide lakes by producing rugged, high relief terrains, and by forming deposits of very coarse debris when failure does occur. Lower overall clay content of the siltstone/mudstone interbeds, and typical dips of 10° to 20° of the Flournoy and Tyee units, also contribute to the opportunity for significant failures. Thus there is in the Coast Range greater potential for large, relatively rapid slides that impound drainages. In contrast, in the Cascades alteration of volcanic bedrock typically leads to development of slow moving earthflows which are easily eroded by streams, though local base level may be raised.

Landslide lakes in the Coast Range vary in size, probably depending on the volume of the slide debris, the height of the dam, and valley geometry. The lifetime of a lake is related to the capacity of the impounded stream to incise or remove the dam, and the rate of sediment deposition from the drainage basin contributing to the lake. Most landslide lakes are quite small and are located near the heads of drainages, such as Elk and Wasson Lakes in the

Scottsburg quadrangle (Baldwin, 1958). However, three of the larger sites--Loon Lake, Triangle Lake, and Drift Creek Slide--are comparable to the Sitkum landslide area and are considered in this study (Figure 14).

Loon Lake is located in the Scottsburg 7.5-minute USGS quad rangle, about 30 miles north of Sitkum, and was formed by a landslide 1460 ± 80 years B.P. (radiocarbon date on wood from standing, drowned trees, from Baldwin, 1981). The slide consists of Tyee Formation sandstone which failed on an east-aspect slope and moved approximately 0.5 mile down-slope to dam the drainage of Mill Creek (Figures 15 and 16). Undisturbed bedrock nearby has an attitude of N 15 E and dips 12 southeast. The failure occurred on the bedrock dip slope.

The dam formed by the landslide debris at Loon Lake is about 235 feet high and originally impounded a lake of about 6 miles in length and 2.6 square miles in surface area (based on area of sediment deposition). Sediment has filled most of the impoundment, reducing Loon Lake to its present surface area of 218 acres (0.34 square miles). The lake outlet flows over coarse landslide debris, consisting of boulder to car-size sandstone blocks, so that incision of the dam has been minor, probably less than 50 feet. The rate of sediment production by the watershed (92 square miles in area) compared to the rate of dam incision over the past 1460 years indicates the lake will be filled long before the landslide dam is removed completely.

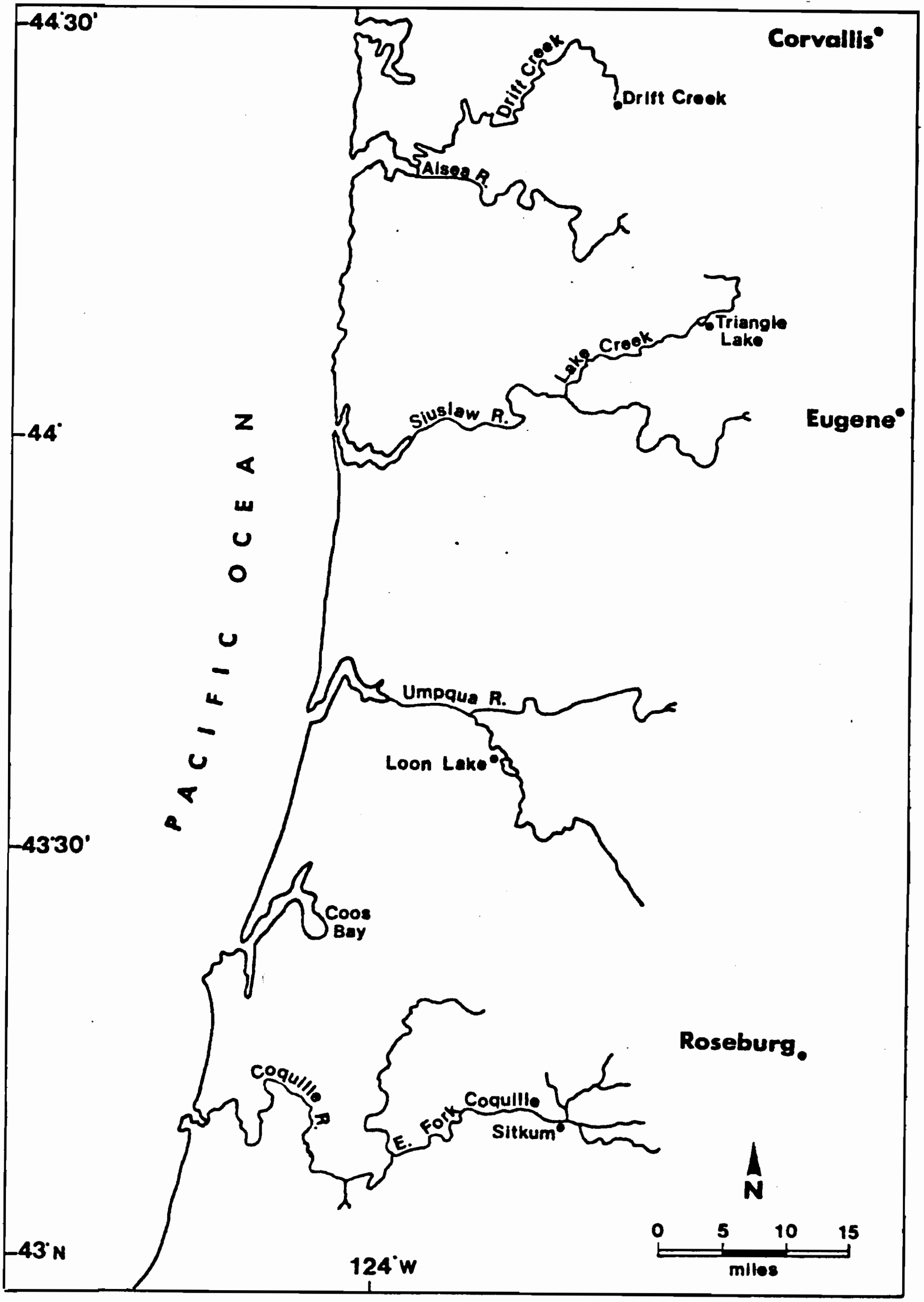


Figure 14. Location of major landslide-dammed lakes in the Coast Range, as discussed in the text; at Sitkum, Loon Lake, Triangle Lake, and Drift Creek.

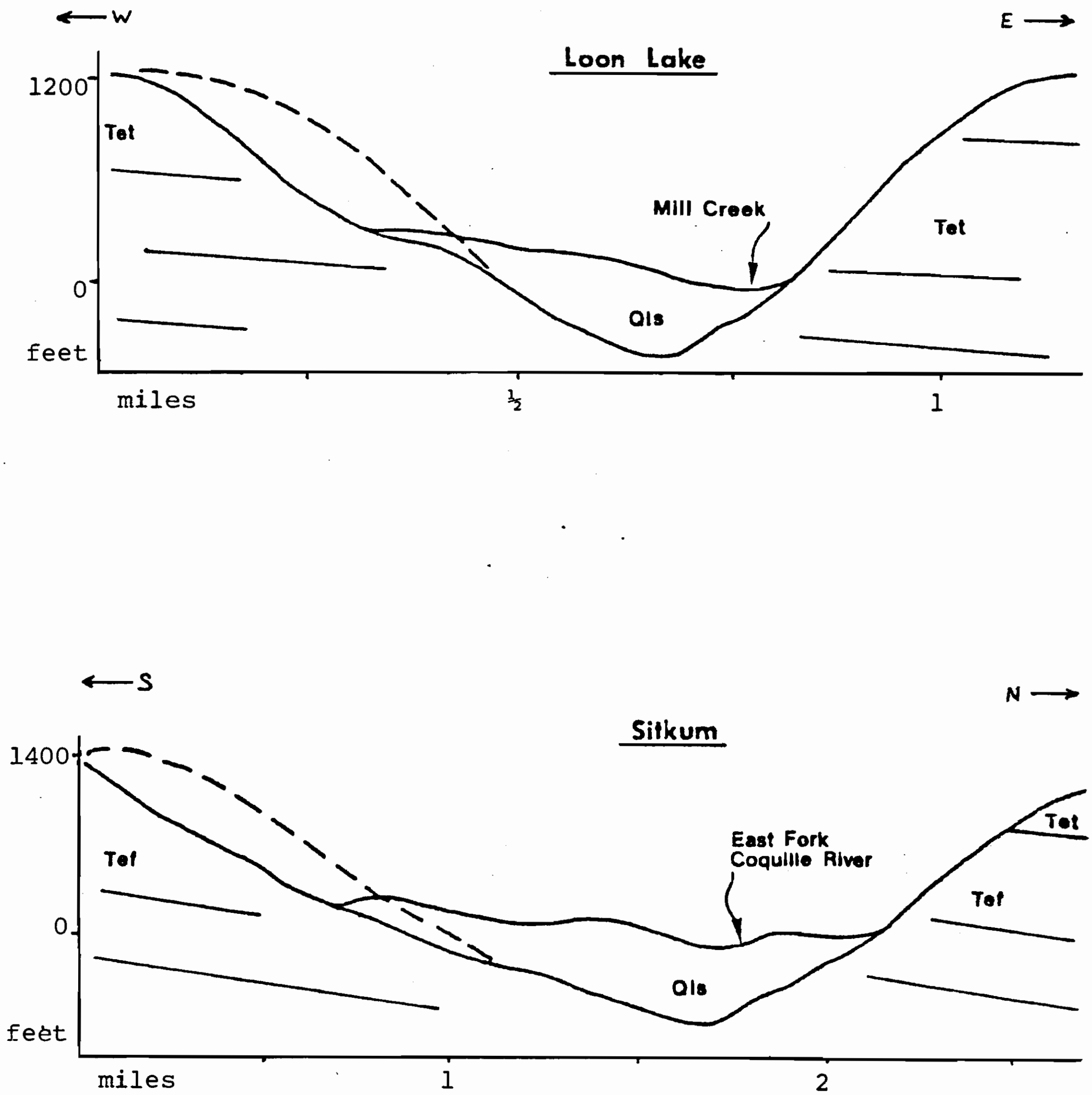


Figure 15. Valley profiles of the Loon Lake and Sitkum landslide sites, both viewed in a downstream direction. The large arrows indicate the location of the streams at the crest of the landslide dams. Qls= landslide debris, Tef= Flournoy Fm., Tet= Tyee Fm. The apparent dip of the bedrock is indicated. The dashed lines approximate the former hillslope prior to failure, and indicate the valley wall from which the landslide moved.

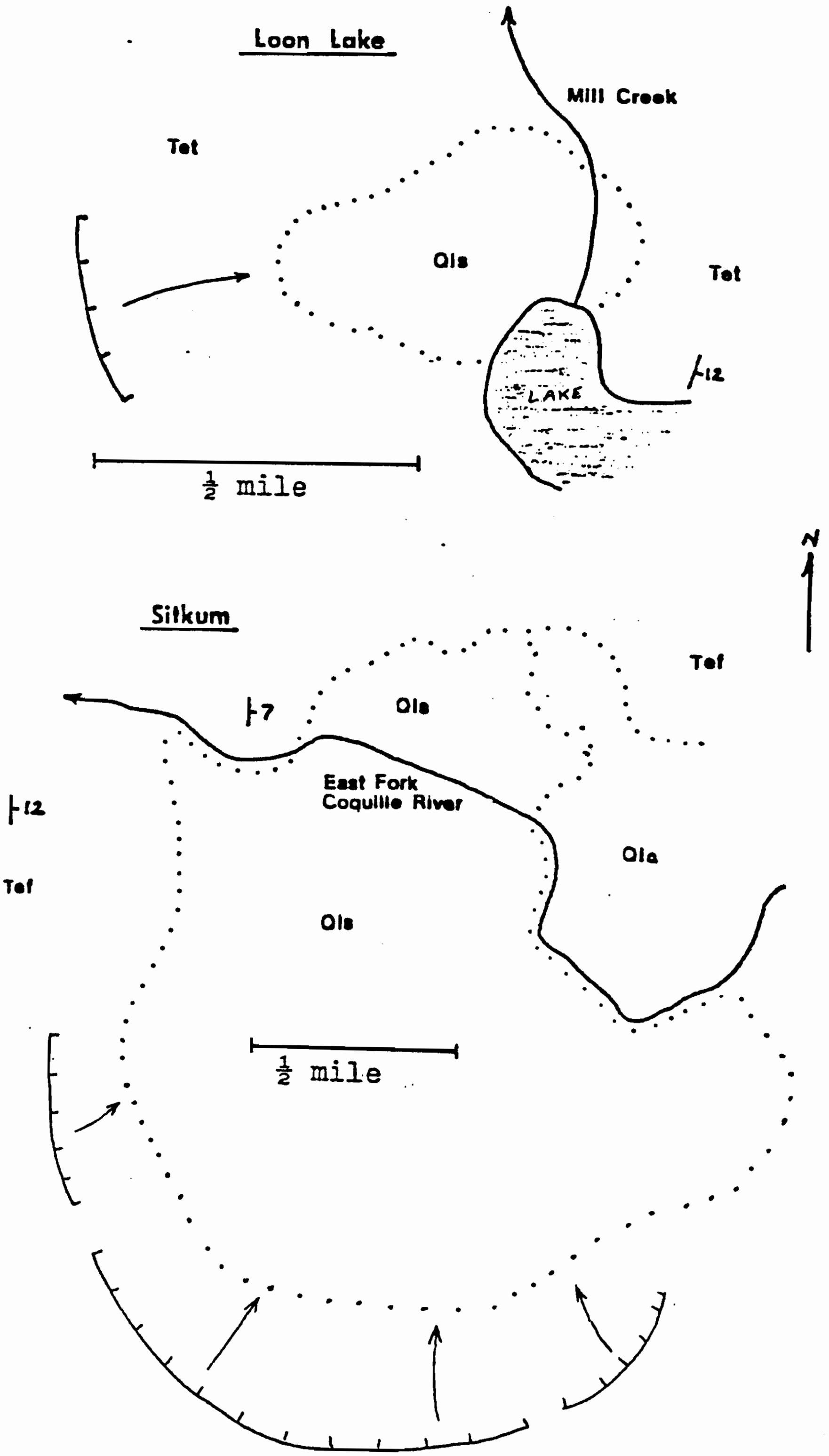


Figure 16. Sketch maps of Loon Lake and Sitkum landslides.

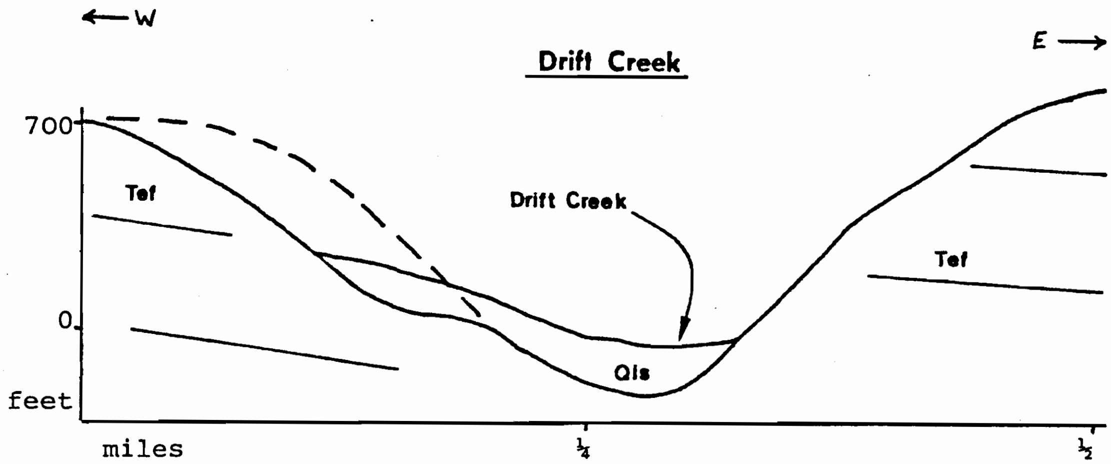
Triangle Lake, in the Blachly 7.5-minute USGS quadrangle, about 60 miles north of Sitkum, is dammed by landslide debris composed of Flournoy Formation sandstone and siltstone. The landslide dam, which is about 250 feet in height, was formed when bedrock failed on a northwest-aspect slope and moved about 0.5 mile downslope, blocking the drainage of Lake Creek (Figures 17 and 18). Undisturbed bedrock near the landslide dam has an attitude of N 50° E, dipping 20° northwest, so the Triangle Lake slide also occurred on the bedrock dip slope.

Unlike the Sitkum and Loon Lake slides, however, the stream outlet from Triangle Lake flows on bedrock between the landslide dam and the hillslope on which failure occurred. The outlet, Lake Creek, apparently was entrenched in undisturbed bedrock of the hillslope after breaching the dam crest at the headward edge of the landslide debris. The original lake formed by the dam extended for 6 miles upstream and had an area of approximately 3200 acres (5.0 square miles). Most of the former basin is filled with sediment, leaving a 256 acre lake near the landslide dam. No radio-isotopic age determinations have been made for the Triangle Lake slide. However, an age estimate of 10,300 years B.P. is obtained for Triangle Lake by applying the sedimentation rates and known age of Loon Lake (see Appendix A).

The Drift Creek slide (T. 13 S., R. 9 W., Sec. 9 of the Tidewater 15-minute quadrangle) is 24 miles northwest of Triangle Lake. It is by far the smallest of the sites discussed here, but is included because it is the only one of these lakes to have formed since settlement by Europeans in the area. The landslide consists of Flournoy Formation siltstones and sandstones which failed on an east-aspect slope near the headwaters of Drift Creek in December, 1975, during a period of high precipitation (Thrall et al, 1980). The lake formed by the landslide dam is approximately 10 acres in size. The lake outlet flows across the toe of the slide, which has not been deeply incised apparently because of the small discharge of the stream (Figures 17, 18 and 19).

The slide, which failed during a known period of less than 36 hours, possibly within minutes, has a volume of approximately 1.7×10^7 cubic yards, and a surface area of about 50 acres. In contrast with the Sitkum and Loon Lake sites the Drift Creek slide deposit is more finely textured, with sandstone boulders of only a few feet in maximum diameter. A portion of the initial slide flowed down the stream channel after reaching the hillslope base and damming the stream. The upper third of the slide area failed by block glide. It now consists of several large competent blocks of siltstone bedrock, and a head scarp that is about 200 feet in height (Thrall et al, 1980).

Similarities between the Drift Creek, Triangle Lake, Loon Lake, and Sitkum slides include: failure of bedrock on a dip slope, involvement of Tyee Formation and/or Flournoy Formation bedrock, and the creation of relatively long-lived dams. In addition, all appear to have failed rapidly since the outlet streams flow through the slide debris and not against the opposite valley wall. Slow failure, such as is typical of an earthflow, would have pushed the stream



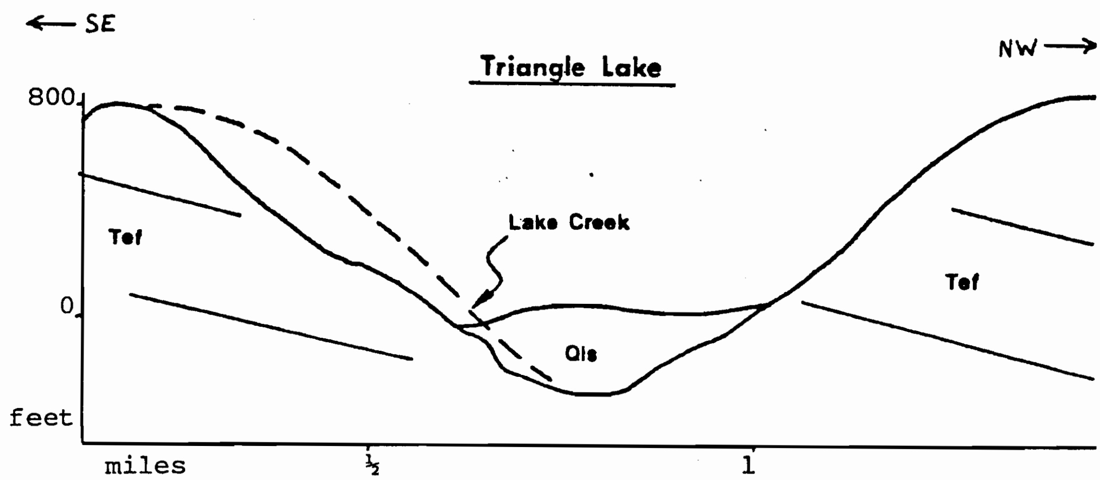


Figure 17. Valley profiles of the Drift Creek and Triangle Lake landslide sites, both viewed in a downstream direction. The large arrows indicate the location of the streams at the crest of the landslide dams. Qls= landslide debris, Tef= Flournoy Fm. The apparent dip of the bedrock is indicated. The dashed lines approximate the former hillslope prior to failure, and indicate the valley wall from which the landslide moved.

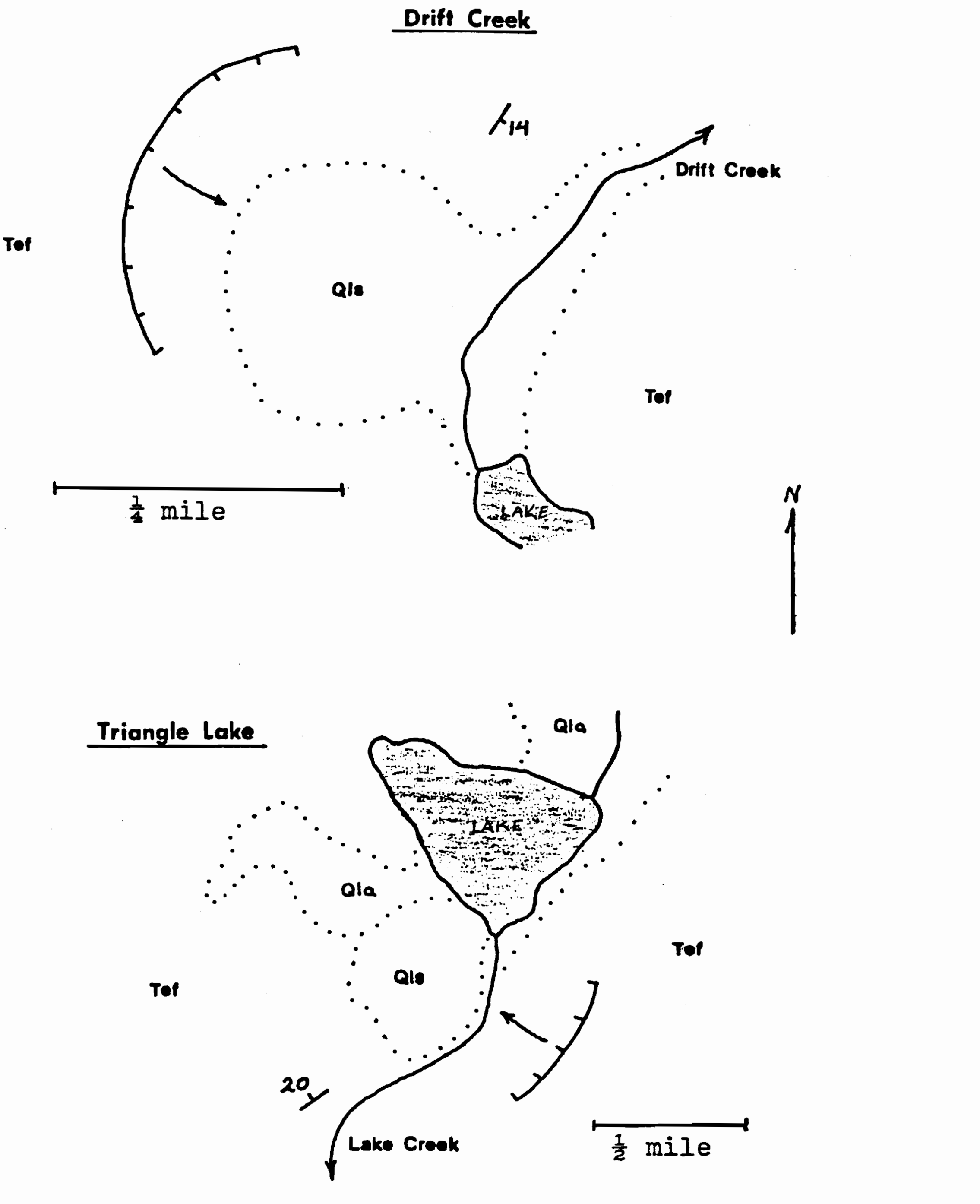


Figure 18. Sketch maps of Drift Creek and Triangle Lake landslides. Drift Creek sketch is after Thrall et al (1980).

Figure 19. Upper: Drift Creek landslide and headscarp, view is to the southwest. Lower: landslide dam and lake at Drift Creek site, view is south (landslide failed on right-hand slope, dam is at lower left). Photos taken April, 1982.

ahead of the landslide mass as it moved across the valley floor. In this case the channel would flow between the slide debris and the opposite valley wall.

Swanson et al (1986) studied landslide dams in Japan and classified them on the basis of either rapid (more than 1.5 meters/day) or slow (less than 1.5 meters/day) landslide movement and landslide-channel interaction. A channel constriction ratio of landslide toe speed to channel width gives an expression of channel response to the landslide. An annual constriction ratio of greater than 100 seems necessary to form a lake behind a landslide dam (Swanson et al, 1986). Figure 20 shows data from Swanson et al (1986) combined with data for Sitkum, Loon Lake, Triangle Lake, and Drift Creek, comparing landslide volume with drainage basin area. The figure relates the occurrence of dam failure to the interaction of landslide volume and fluvial erosion potential. For example, a land slide of small volume would likely produce a short-lived lake if the contributing drainage basin is large enough to produce streamflow that can remove the dam. However, many other factors influence the lifespan of a landslide dam, such as the coarseness of the landslide debris and the steepness of the downstream slope on the dam surface.

The occurrence of landslide-dammed lakes in the central and southern Coast Range results from a particular geologic setting and the hydrologic influence of high precipitation. Any additional causative factor has not been determined. Adams (1981) studied landslide-dammed lakes in New Zealand whose formation was triggered by historic earthquakes. He also examined prehistoric landslide-

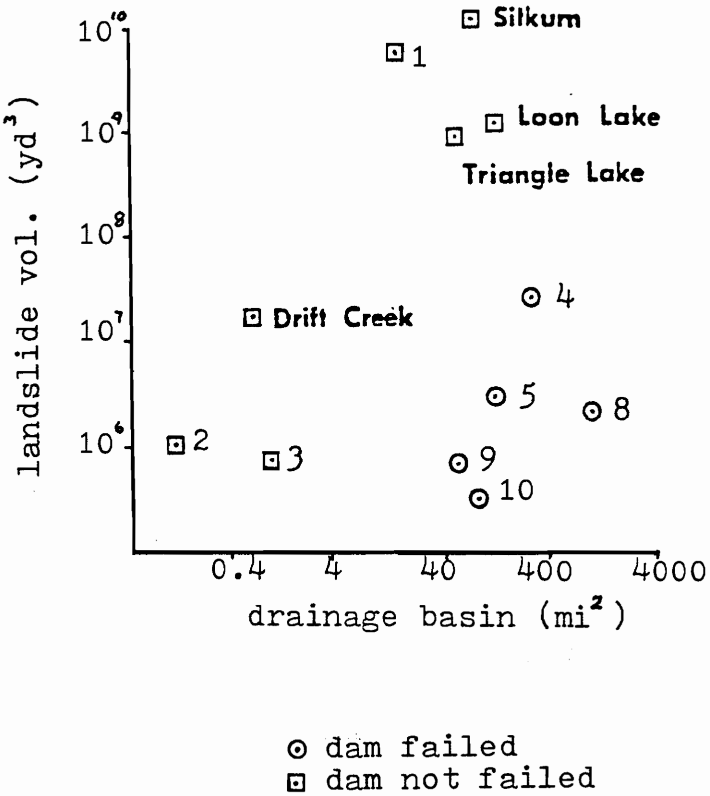


Figure 20. Distribution of landslide dams in relation to the landslide volume and the drainage basin area above the landslide dam. Dam sites 1 through 10 are from Swanson et al (1986), "Landslide Dams in Japan," and the site numbers correspond to Swanson et al's (1986) figure 7.

dammed lakes which he interpreted as probably being formed by earthquakes. He based conclusions with regard to the prehistoric events on similar landslide morphology, synchronous age, and areal distribution of the "earthquake-dammed" lakes.

The Oregon Coast Range, being on the North American continental margin, is an area of active structural deformation and uplift. Earthquakes in the Coast Range are rare, however, at least in historical time, so this mechanism may not be applicable to the landslide-dammed lakes in the Coast Range. In addition, a synchronous age for the major landslide sites has not been determined, and the Drift Creek slide is known to not have been triggered by an earthquake.

However, studies by Heaton and Hartzell (1987) indicate that seismicity along the North American and Juan de Fuca plate margin may be characterized by long periods of quiet broken by short periods of large subduction earthquake activity. They present evidence that the Cascadia subduction zone, off the coasts of Oregon and Washington, may store elastic energy for intervals of several hundred years between occurrences of rapid plate motion. Since historical records in the Pacific Northwest are available for only the past 150 years, it is possible that subduction of the Juan de Fuca plate along the Cascadia subduction zone may trigger landslides in the Oregon Coast Range.

SUMMARY

1. The base of the stratigraphic section in the study area is represented by the Late Jurassic Otter Point Formation and the Eocene Roseburg Formation. The Otter Point occurs as isolated blocks of metamorphic and sedimentary rocks over a small portion of the study area. The Roseburg siltstones and sandstones are tightly folded, with a northeast-southwest fold axis, and outcrop in the western third of the study area. The overlying Lookingglass, Flournoy, and Tyee Formations outcrop in the eastern two-thirds of the study area and are all Eocene in age. They are all composed of rhythmically bedded sandstones and siltstones which dip gently eastward. The entire study area lies on the western limb of a major north-south syncline in the southern Coast Range.

2. The bedrock lithology and structure in the study area is evidenced by the north to northwest-southeast trend of cuesta ridges in Lookingglass, Flournoy, and Tyee terranes, and by northeast-southwest trending ridges in Roseburg terrane. In addition, the greater percentage of sandstone in the Flournoy and Tyee Formations produces steeper slopes and greater relief than in areas underlain by the Roseburg and Lookingglass Formations, which are predominantly siltstone.

3. The geologic history of the study area, combined with the region's high precipitation, has resulted in most bedrock being obscured by Quaternary alluvial deposits, deep soils, and mass movement deposits. The mass movement activity includes creep, rockfall,

debris avalanches/ torrents, slump-earthflows, rock slumps, and complex land-slides.

4. Rockfall occurs where significant sandstone or conglomerate units outcrop, forming cliffs of exposed bed-rock on hillslopes in Flournoy, Tyee, and basal Looking-glass terranes. Rapid weathering processes and dense vegetation prevent the development of unvegetated talus slopes; instead, rockfall debris is incorporated downslope into colluvial soils.

5. The distribution of 45 debris avalanche/torrent sites in the study area indicates that slope failure is more likely to occur on steeper slopes. Slopes with a gradient of 91% or greater have a failure frequency nearly twice that of slopes between 81-90%. Debris avalanche/ torrent distribution is also skewed toward Tyee Formation terrane since the Tyee is characterized by a larger percentage of resistant sandstone units and steeper slopes.

6. Debris avalanche/torrent sites in the study area occur more frequently on west-aspect slopes. The principal factor in this relationship is the topographic expression of the bedrock geology, which produces cuesta ridges with steep west-aspect, obsequent slopes.

7. The correlation of debris avalanche/torrent sites with management history indicates that failures are most likely to occur in road-fill materials and in clear-cuts less than 10 years old. Undisturbed slopes and older clear-cut areas have much lower frequencies of failure sites.

8. Slump-earthflows in the study area are much smaller and less frequent than those which characterize mass movement activity

documented in the western Cascades. All can be attributed to over-steepened slopes at sandstone-siltstone contacts, or to logging and road-building on perennially wet areas on cuesta dip slopes.

9. Rock slumps do not occur extensively in the study area and structural or stratigraphic explanations for their distribution are indistinct.

10. Significant landslide features of a catastrophic origin are identified at 5 sites. All of the failures pre-date white settlement of the region. The largest site, the Sitkum landslide, involved a massive amount of material and dammed the East Fork Coquille River, forming a large lake which is now filled with alluvial and lacustrine sediments.

11. Comparisons between the Sitkum landslide and other landslide-dammed lakes in the Coast Range (Loon Lake, Triangle Lake, and Drift Creek) provides some generalizations. Failure of resistant Tyee and/or Flournoy Formation sand-stone occurred along the bedrock dip slope; failure was probably rapid, as indicated by run-up of landslide debris on the opposite valley wall; and the landslide dams are composed of materials coarse enough to prevent rapid removal by the streams.

12. Comparisons between the Sitkum landslide and Loon Lake landslide (which has a known age) provide an estimated minimum age of 3125 years B.P. for the Sitkum landslide. Comparisons between the Loon Lake and Triangle Lake landslides provide an estimated age of 10,300 years B.P. for the Triangle Lake landslide.

BIBLIOGRAPHY

Adams, J., 1981, Earthquake-dammed lakes in New Zealand: Geology, v. 9, p. 215-219.

Allen, J.E., and Baldwin, E.M., 1944, Geology and Coal Resources of the Coos Bay quadrangle: Oregon Dept. of Geol. and Min. Indus. Bull. 27, 153 p.

Amaranthus, M.P., Rice, R.M., Barr, N.R., and Ziemer, R.R., 1985, Logging and forest roads related to increased debris slides in SW Oregon: Journal of Forestry, v. 83, no. 4, p. 229-233.

Atlas of Oregon, 1976, University of Oregon Press: Eugene, Oregon, p. 130-139.

Baldwin, E.M., 1958, Landslide Lakes in the Coast Range of Oregon: Geol. Soc. Oregon Country, Geol. News Letter, v. 24, no. 4, p. 23-24.

Baldwin, E.M., 1965, Geology of the south end of the Oregon Coast Range Tertiary Basin: Northwest Science, v. 39, no. 3, p. 93-103.

Baldwin, E.M., 1969, Geologic map of the Myrtle Point area, Coos County Oregon: U. S. Geological Survey Mineral Inv. Map MF-302.

Baldwin, E.M., 1974, Eocene stratigraphy of southwestern Oregon: Oregon Dept. of Geology and Min. Indus. Bull. 83, 40 p.

Baldwin, E.M., 1975, Revision of the Eocene stratigraphy of southwestern Oregon, in Weaver, D.W., Hornaday, G.R., and Tipton, A. (eds.), Paleogene Symposium: Pacific Section of Am. Assoc. Petrol. Geologists, SEPM, Ann. Mtg., p. 49-64.

Baldwin, E.M., 1981, Geology of Oregon, 3rd edition: Kendall-Hunt Publ. Co., Dubuque, Iowa, 170 p.

Baldwin, E.M., and Beaulieu, J.D., 1973, Geology and mineral resources of Coos County, Oregon: Oregon Dept. of Geology and Min. Indus. Bull. 80, p. 1-40.

Beaulieu, J.D., and Hughes, P.W., 1975, Environmental Geology of Western Coos and Douglas Counties, Oregon: Oregon Dept. of Geology and Min. Indus. Bull. 87, 148 p.

Blake, M.C., Jr., and Jayko, A., 1980, Late Mesozoic tectonostratigraphic terranes (microplates) in southwestern Oregon: Geol. Soc. Am. Abstracts with Programs, v. 12(3), p. 98.

Blong, R.J., 1973, A numerical classification of selected landslides of the debris slide-avalanche-flow type: Eng. Geol., v. 7, p. 99-114.

Bourgeois, J., and Dott, R.H. Jr., 1985, Stratigraphy and sedimentology of Upper Cretaceous rocks in coastal southwest Oregon: Evidence for wrench-fault tectonics in a postulated accretionary terrane; Geol. Soc. Am. Bulletin, v. 96, p. 1007-1019.

Carson, M.A., and Kirkby, M. J., 1972, Hillslope form and process: Cambridge University Press, 475 p.

Chan, M.A., and Dott, R. H., Jr., 1983, Shelf and deep-sea sedimentation in Eocene forearc basin, western Oregon--fan or non-fan?: American Assoc. Petrol. Geologists Bull., v. 67, no. 11, p. 2100-2116.

Combs, Mapril, 1982, personal communication, Myrtle Point, Oregon.

Corliss, J.F., 1973, Soil survey of Alsea area, Oregon: USDA Soil Cons. Service and Forest Service survey, 82 p.

Dietrich, W.E., and Dunne, T., 1978, Sediment budget for a small catchment in mountainous terrain: Zeit. fur Geomorph. Suppl. Bd. 29, p. 191-206.

Diller, J.S., 1898, Roseburg folio, Oregon: U. S. Geol. Survey, Geol. Atlas of U. S., Folio no. 49.

Diller, J.S., 1899, The Coos Bay Coal Field, Oregon: U. S. Geological Survey Ann. Report, Pt. 3, p. 309-370.

Diller, J.S., 1901, Coos Bay folio, Oregon: U. S. Geol. Survey, Geological Atlas of the U. S., Folio no. 73.

Dodge, O., 1898, Pioneer history of Coos and Curry Counties, Capitol Printing Co., Salem, Oregon, 185 p.

Dott, R.H., Jr., 1971, Geology of the southwestern Oregon coast west of the 124th meridian: Oregon Dept. Geol. and Min. Indus. Bull. 69, 63 p.

Dott, R.H. Jr., and Bourgeois, J., 1980, Late Mesozoic tectonics and sedimentation along the southwestern Oregon coast: Geol. Soc. Am. Abs. with Programs, v. 12 (3), p. 104.

Drake, E.T., 1982, Tectonic evolution of the Oregon continental margin: Oregon Geology, v. 44, no. 2, p. 15-21.

Dyrness, C.T., 1967, Mass soil movements in the H. J. Andrews Experimental Forest: USDA Forest Service Res. Paper, PNW-42, 12 p.

Fairchild, R.W., 1966, Geology of T. 29 S., R. 11 W., of the Sitkum and Coquille quadrangles, Oregon: unpub. master's thesis, Univ. of Oregon, Eugene, 73 p.

Franklin, J.F., and Dyrness, C. T., 1973, Natural vegetation of Oregon and Washington: USDA Forest Service Gen. Tech. Report PNW-8, 417 p.

Graham, J.D., 1986, Mass Movement dynamics, geomorphology and their relationship to geology in the North Fork Siuslaw River drainage basin, Oregon Coast Range, unpub. master's thesis, Oregon State University, Corvallis, 96 p.

Gray, D.H., 1970, Effects of forest clearcutting on the stability of natural slopes: Bull. Assoc. Eng. Geol., v. 7, p. 45-66.

Gresswell, S., Heller, D., and Swanston, D. N., 1979, Mass movement response to forest management in the central Oregon Coast Ranges: USDA Forest Service Res. Bull. PNW-84, 26 p.

Griggs, A.B., 1945, Chromite-bearing sands of the southern part of the coast of Oregon, U.S. Geol. Survey Bull. 945-E, 150 p.

Hall, R.L., 1984, The Coquille Indians: Yesterday, today, and tomorrow; Smith, Smith and Smith Publ. Co., Lake Oswego, Oregon, 203 p.

Harr, R.D., 1976, Forest practices and streamflow in western Oregon: U. S. Forest Service, Pacific N. W. Forest and Range Exp. Sta., Gen. Tech. Report, PNW-49, 18 p.

Harr, R.D., and Yee, C. S., 1975, Soil and hydrologic factors affecting the stability of natural slopes in the Oregon Coast Range: Oregon State Univ., Corvallis, Water Res. Research Inst. Report, WRRI-33, 204 p.

Heaton, T.H., and Hartzell, S.H., 1987, Earthquake hazards on the Cascadia subduction zone: Science, v. 236, p. 162-168.

Heller, P.L., Peterman, Z.E., O'Neil, J.R., and Shafiqullah, M., 1985, Isotopic provenance of sandstones from the Eocene Tyee Formation, Oregon Coast Range: Geol. Soc. Am. Bull., v. 96, p. 770-780.

Hicks, B.A., 1982, Geology, geomorphology, and dynamics of mass movement in parts of the Middle Santiam River drainage, western Cascades, Oregon: unpub. master's thesis, Oregon State Univ., Corvallis, 153 p.

Karlin, R., 1980, Sediment sources and clay mineral distributions off the Oregon Coast: Journal of Sediment. Petrology, v. 50, no. 2, p. 543-560.

Klohn, M.S., 1967, Geology of the north-central part of the Coos Bay quadrangle, Oregon: unpub. master's thesis, Univ. of Oregon, Eugene, 59 p.

Koch, J.G., 1966, Late Mesozoic stratigraphy and tectonic history, Port Orford-Gold Beach area, southwestern Oregon coast: Am. Assoc. Petrol. Geologists Bull., v. 50, no. 1, p. 25-71.

Laiming, B.G., 1940, Some foraminiferal correlation in the Eocene of the San Joaquin Valley, Calif.: 6th Pac. Sci. Congress, 1939, Proc., v. 2, p. 535-568.

Lovell, J.T.B., 1969, Tyee Fm., undeformed turbidites and their lateral equivalents, mineralogy and paleo-geography: Geol. Soc. Am. Bull., v. 80, no. 1, p. 9-22.

Magill, J.R., and Cox, A., 1981, Post-Oligocene tectonic rotation of the Oregon western Cascade Range and the Klamath Mountains: Geology, v. 9, p. 127-131.

Magoon, L.B., 1966, Geology of T. 28 S., R. 11 W., of the Sitkum and Coquille quadrangles, Oregon: unpub. master's thesis, Univ. of Oregon, Eugene, 73 p.

Mallory, V.S., 1959, Lower Tertiary biostratigraphy of the California Coast Ranges: Am. Assoc. Petrol. Geologists Monograph, 416 p.

Marion, D.A., 1981, Landslide occurrence in the Blue River drainage, Oregon: unpub. master's thesis, Oregon State Univ., Corvallis, 114 p.

McKee, B., 1972, Cascadia -- The Geological Evolution of the Pacific Northwest: New York, McGraw-Hill, 394 p.

Nelson, E.B., 1966, The geology of the Fairview-McKinley area, central Coos County, Oregon: unpub. master's thesis, Univ. of Oregon, Eugene, 59 p.

Oregon State Dept. of Revenue, 1976, Assessment and Appraisal Div., Timber Mapping Section, map of Coos County: Salem, Oregon.

Oregon State Water Resources Board, 1961, map of Hydrological Stations and Annual Precip., Coos-Coquille drainage basin: Salem, Oregon, File No. 17.62.

Pierson, T.C., 1977, Factors controlling debris flow initiation on forested hillslopes in the Oregon Coast Range, Ph.D. thesis, Univ. of Washington, Seattle, 166 p.

Read, S.A.L., 1979, Lake Waikaremoana outlet, engineering geological studies of factors related to leakage through the natural dam: New Zealand Geological Survey, Engineering Geology Report 336, 42 p.

Roden, G., 1967, On river discharge into the Northeast Pacific Ocean and the Bering Sea: Journal of Geophys. Res., v. 72, p. 5613-5629.

Schulz, M.G., 1980, The quantification of soil mass movements and their relationahip to bedrock geology in the Bull Run Watershed, Multnomah and Clackamas Counties, Oregon: unpub. master's thesis, Oregon State Univ., Corvallis, 170 p.

Simpson, R.W., and Cox, A., 1977, Paleomagnetic evidence for tectonic rotation of the Oregon Coast Range: Geology, v. 5, p. 585-589.

Snavely, P.D., Jr., Wagner, H. C., and MacLeod, N. S., 1964, Rhythmic-bedded eugeosynclinal deposits of the Tyee Formation, Oregon Coast Range, in Symposium on cyclic sed., Merriam, D. F. (ed.): Kansas Geol. Survey Bull. 169, v. 11, p. 461-480.

Swanson, F.J., 1975, Geology, geomorphology, and forest land use in the Elk and Sixes River basins, southwest Oregon: Proc. of Oregon Academy of Science, v. 3, p. 85.

Swanson, F.J., Harr, R. D., and Frederiksen, R. L., 1980, Field trip guide: Geomorphology and hydrology in the H. J. Andrews Exp. Forest, western Cascades; Oregon Dept. of Geology and Min. Indus. Bull. 101, p.217-232.

Swanson, F.J., and James, M.E., 1975, Geology and geomorphology of the H.J. Andrews Exp. Forest, western Cascades, Oregon: USDA Forest Service Res. Paper PNW-188, 14 p.

Swanson, F.J., and Lienkaemper, G.W., 1978, Physical consequences of large organic debris in Pacific Northwest streams: USDA Forest Service, Gen. Tech. Report PNW-69, 12 p.

Swanson, F.J., Lienkaemper, G.W., and Sedell, J.R., 1976, History, physical effects, and management implications of large organic debris in western Oregon streams: USDA Forest Service Gen. Tech. Report PNW-56, 15 p.

Swanson, F.J., Oyagi, N., and Tominaga, M., 1986, Landslide Dams in Japan, in Landslide Dams: Processes, Risks, and Mitigation; ed. by R. Schuster; Geotechnical Spec. Pub. 3, Am. Soc. of Civil Eng., N.Y.

Swanson, F.J., Swanson, M.M., and Woods, C., 1981, Analysis of debris avalanche erosion in steep forested lands: an example from Mapleton, Oregon, USA, in Erosion and Sed. Transport in Pacific Rim Steeplands: IAHS-AISH Pub. 132, p. 67-75.

Swanston, D.N., 1981, Creep and earthflow erosion from undisturbed and management impacted slopes in the Coast and Cascade Ranges of the Pacific Northwest, USA, in Erosion and Sediment Transport in Pacific Rim Steeplands, IAHS-AISH Pub. 132, p. 76-96.

Swanston, D.N., and Swanson, F.J., 1976, Timber harvesting, mass erosion, and steepland forest geomorphology in the Pacific Northwest, in Geomorphology and Engineering, Coates, D.R. (ed.): Dowden, Hutchinson and Ross, Inc., p. 199-221.

Swanston, D.N., and Swanson, F.J., 1977, Complex mass movement terrains in the western Cascade Range, Oregon: Geol. Soc. of Am. Reviews in Eng. Geol., v.3, p. 113-124.

Thrall, F.G., Jack, R., Johnson, J.J., and Stanley, D.A., 1980, Failure mechanisms of the Drift Creek slide: presented at Ann. Mtg., Cordilleran Section of Geol. Soc. Am., Corvallis, Oregon.

Trigger, J.K., 1966, Geology of the south-central part of the Sitkum quadrangle, Coos County, Oregon: unpub. master's thesis, Univ. of Oregon, Eugene, 79 p.

U.S. Geological Survey Water Data Reports, 1965-1976 (pub. annually), Water resources data for California, Oregon, and Washington.

Varnes, D.J., 1978, Slope movement types and processes, in Landslides analysis and control, Schuster, R. L. and Krizek, R. J. (eds): National Acad. Sciences, Trans. Bd. Spec. Report 176, p. 12-80.

Walker, G.W., and King, P.B., 1967, Geologic Map of Oregon: U. S. Dept. of Interior, Misc. Geologic Invest. Map I-595.

APPENDICES

APPENDIX A:

AGE-DATING OF SITKUM LANDSLIDE

The Loon Lake landslide is located in the Scottsburg 7.5-minute USGS quadrangle, approximately 30 miles north of the Sitkum landslide. The Loon Lake and Sitkum landslides are comparable in their geologic settings and in the formation of long-lived lakes behind the landslide dams, though the lake at Sitkum is now completely filled with alluvial and lacustrine sediments.

Age determinations have not been reported for the Sitkum landslide. However, Baldwin (1981) gave a radiocarbon date of 1460 ± 80 years B.P. for the failure of the Loon Lake landslide. By determining the lake-fill sediment volumes at the two sites, and applying estimated sediment yield rates for the drainage basins, a minimum age for the Sitkum landslide can be extrapolated from the Loon Lake date.

The drainage basin area for Loon Lake is 92 square miles (238 km^2). From cross-section and surface area calculations on the USGS topographic map, the sediment-fill behind the Loon Lake landslide dam is estimated at 2.9×10^9 cubic feet (8.2×10^7 m^3). The Sitkum landslide drainage basin is 79 square miles (205 km^2) in area, and the sediment-fill trapped by the dam is estimated at 5.3×10^9 cubic feet (1.5×10^8 m^3; derived from calculations of the sediment-fill surface area and depths).

Dividing the Loon Lake sediment-fill volume by the age of the landslide provides a sediment yield rate for the Loon Lake basin:

$$(1) \qquad \frac{8.2 \times 10^7 \ m^3}{1460 \text{ years}} = 5.6 \times 10^4 \ m^3/yr$$

$$(2) \qquad \frac{5.6 \times 10^4 \ m^3/yr}{238 \ km^2} = 2.35 \times 10^2 \ m^3/yr\text{-}km^2$$

And dividing the Sitkum sediment-fill volume by the Loon Lake basin sediment yield rate gives an estimate of the minimum age of the Sitkum landslide:

$$(3) \qquad 2.35 \times 10^2 \ m^3/yr\text{-}km^2 \times 205 \ km^2 = 4.8 \times 10^4 \ m^3/yr$$

$$(4) \qquad \frac{1.5 \times 10^8 \ m^3}{4.8 \times 10^4 \ m^3/yr} = 3125 \text{ years}$$

This is an estimate of the time required for alluvial processes to fill the former lake dammed by Sitkum landslide, if sediment yield rates have been historically similar in the Loon Lake and Sitkum basins. The actual age of the Sitkum landslide would be 3125 years plus the interval from final sediment infilling of the lake to the present.

In comparison, Karlin (1980) summarized data on the sediment yields of coastal rivers in Washington, Oregon, and northern California; including the Coquille and Coos Rivers (USGS, 1965-1976; Roden, 1967), both at 125 tonnes/km^2-yr. Using this value, a bulk density of 1.6 tonnes/m^3, the area of the Sitkum drainage basin (205 km^2), and the sediment-fill volume at Sitkum (1.5 x 10^8 m^3), the minimum age of the Sitkum landslide can be estimated in the following manner:

(5) $$205 \text{ km}^2 \times 125 \text{ tonnes/km}^2\text{-yr} = 2.6 \times 10^4 \text{ tonnes/yr}$$

(6) $$\frac{2.6 \times 10^4 \text{ tonnes/yr}}{1.6 \text{ tonnes/m}^3} = 1.6 \times 10^4 \text{ m}^3\text{/yr}$$

(7) $$\frac{1.5 \times 10^8 \text{ m}^3}{1.6 \times 10^4 \text{ m}^3\text{/yr}} = 9375 \text{ years}$$

This value differs from the first age estimate by over 6200 years. However, the validity of both figures depends on the historical sediment yield rates in the Sitkum and Loon Lake basins, especially prior to modern (non-Indian) settlement of the region. Karlin (1980) gave widely divergent values for the modern sediment yield rates from Pacific Coast streams, indicating that adjacent basins can be quite different in their sediment output. Also, Karlin's (1980) figures are for suspended sediment loads, while the landslide-dammed lakes also trap bedload sediments. Fire and windfall are two pre-logging factors which could have impacted the sediment yield rates in the Sitkum and Loon Lake basins, but their actual importance is unknown since the sedimentation histories of the two sites have not been studied.

The age of the Loon Lake landslide and the apparent basin sediment yield rate can also be extrapolated to the Triangle Lake landslide, as in equations 1 through 4. Substituting the drainage basin area and the estimated sediment-fill volume of Triangle Lake ($132\ km^2$; $3.2 \times 10^8\ m^3$), results in a date of 10,300 years B.P. for the failure of the Triangle Lake landslide.

(3a) $$2.35 \times 10^2\ m^3/yr\text{-}km^2 \times 132\ km^2 = 3.1 \times 10^4\ m^3/yr$$

(4a) $$\frac{3.2 \times 10^8\ m^3}{3.1 \times 10^4\ m^3/yr} = 10{,}300 \text{ years}$$

In comparison, application of the data from Karlin (1980), as in equations 5 through 7, gives an age of 32,000 years B.P. for the Triangle Lake landslide:

(5a) $$132\ km^2 \times 125\ tonnes/km^2\text{-}yr = 1.6 \times 10^4\ tonnes/yr$$

(6a) $$\frac{1.6 \times 10^4\ tonnes/yr}{1.6\ tonnes/m^3} = 1.0 \times 10^4\ m^3/yr$$

(7a) $$\frac{3.2 \times 10^8\ m^3}{1.0 \times 10^4\ m^3/yr} = 32{,}000 \text{ years}$$

APPENDIX B:

INDEX OF STUDY AREA SITES

The collection and organization of data for this project required the development of an index system to identify and locate the geographic positions of such data on the study area base map. This system involved the assignment of unique numbers to each of 119 General Land Office sections covering the study area (a 7 x 17-mile grid) and to each data site within the sections, giving each data site a 5-digit index number. The first digit of each index number refers to the section row number (1 through 7, north to south) while the second two digits refer to the section column number (1 through 17, west to east). The last 2 digits of each index number refer to the individual data site located within a particular section. Brewster Rock, for example, has index number 31204; it is located in the third row and 12th column of sections, and is site number 4 in that particular section.

Because index numbers were assigned to a wide variety of data during the fieldwork and office investigations, only a portion of the index data are listed in this Appendix. Data such as bedrock attitudes, bedrock samples, and various field notes are not included. Those sites which are described in the following list can be referenced by the index numbers to Plates 1 and 2, as well as to the text of this thesis. The sites in this list have also been organized by the type of feature being described.

In the following list the "source" category refers to the condition of the slope area where the failure occurred; the "slope %" refers to the slope steepness at the site of failure. Under "aspect" N = north, S = south, E = east, and W = west. Under "terrane" T = Tyee Formation, F = Flournoy Formation, L = Lookingglass Formation, R = Roseburg Formation, and O = Otter Point Formation.

Debris Avalanches/Torrents

index #	source	slope %	aspect	terrane	est.vol.(ft.^{3})
10903	clear-cut	65	W	F	inaccessible
21202	undisturbed	100	W	T	"
21203	undisturbed	100	W	T	"
21205	undisturbed	85	N	T	"
21208	undisturbed	120	W	T	"
21403	undisturbed	75	W	T	"
30501	undisturbed	65	N	R	"
30605	clear-cut	82	W	R	1.8×10^3
31202	clear-cut	65	W	F	inaccessible
31203	road-fill	75	N	T	1.7×10^3
31206	undisturbed	75	S	T	inaccessible
31207	undisturbed	75	S	T	"
31208	undisturbed	72	W	T	"
31304	road-fill	52	E	T	3.0×10^3
31701	undisturbed	52	W	T	inaccessible
31702	undisturbed	115	W	T	inaccessible
40802	undisturbed	62	W	L	"

Debris Avalanches/Torrents

index #	source	slope %	aspect	terrane	est.vol.(ft.^{3})
41007	road-fill	75	W	F	"
41301	undisturbed	85	W	F	"
41302	undisturbed	85	W	F	"
41406	undisturbed	75	N	T	"
41407	undisturbed	75	N	T	"
41408	undisturbed	80	N	T	"
41409	undisturbed	80	N	T	"
41502	undisturbed	100	S	T	"
41507	undisturbed	88	N	T	"
41508	undisturbed	88	N	T	"
41606	road-fill	95	S	T	3.3×10^3
41701	clear-cut	62	N	T	2.0×10^4
41704	clear-cut	110	S	T	3.0×10^3
41711	undisturbed	62	W	T	8.4×10^2
41712	undisturbed	65	N	T	9.0×10^3
51303	clear-cut	85	W	T	inaccessible
51401	clear-cut	100	W	T	"
51402	clear-cut	55	N	T	"
51403	clear-cut	82	W	T	1.8×10^4
51404	clear-cut	62	W	T	1.0×10^4
51405	clear-cut	62	N	T	inaccessible
51501	undisturbed	105	W	T	6.0×10^3
61301	clear-cut	65	E	F	inaccessible
61302	clear-cut	75	N	T	"

Debris Avalanches/Torrents

index #	source	slope %	aspect	terrane	est.vol.($ft.^3$)
61303	clear-cut	62	N	T	"
61401	road-fill	70	W	T	1.5×10^4
70502	clear-cut	75	N	R	4.5×10^3
70504	clear-cut	75	N	R	inaccessible

Slump-earthflows

index #	source	slope %	aspect	terrane	area ($ft.^2$)
21201	road-fill	40	N	T	3.0×10^4
21204	undisturbed	90	W	T	4.0×10^4
21206	undisturbed	90	W	T	3.6×10^4
21207	undisturbed	90	W	T	6.0×10^4
31102	clear-cut	40	E	F	3.0×10^4
31303	undisturbed	50	E	T	1.3×10^6
31706	road-fill	60	E	T	3.6×10^4
41101	road-fill	70	E	T	1.5×10^5
41107	undisturbed	70	W	F	3.0×10^4
41509	undisturbed	40	E	T	5.0×10^3
51101	undisturbed	70	W	F	3.0×10^4
60504	undisturbed	60	E	O	5.0×10^3

Rock Slumps, Complex Landslides, Rockfalls

index #	feature	slope %	aspect	terrane	area (acres)
11104	rock slump	70	S	F	3.0
21102	"	40	S	F	3.0

Rock Slumps, Complex Landslides, Rockfalls

index #	feature	slope %	aspect	terrane	area (acres)
30403	"	40	E	R	11.9
31201	"	40	S	R	1.4
41201	"	70	E	F	11.9
50403	"	40	N	R	3.2
50501	rock slump	40	N	R	2.5
60505	"	60	N	R/O	0.2
21304	landslide	50	E	T	230
30901	"	55	S	F/L	45
40702	"	15	N	R	64
40904	"	20	N	L	48
51201	"	40	N	T/F	985 (1.5 sq.mi.)
31205	rockfall		S	T	
50601	"		S	L	

miscellaneous

index #	
31204	view south toward Sitkum landslide from Brewster Rock; Figure 10.
41208	view of East Fork Coquille River course over landslide debris; Figure 8.