AN ABSTRACT OF THE THESIS OF

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Title: Controls on Movement of Selected Landslides in the Coast Range and Western Cascades, Oregon

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The movement characteristics of five landslides are compared and interpreted based on records of approximately 10-years duration. Condon landslide in the Oregon Coast Range has consistently exhibited brief (1 - 8 days) movement episodes in wet winter months, separated by long periods of no movement. The translatory movement is probably controlled by the orientation and structure of the underlying sedimentary rocks. From 1981 to 1990, annual movement averaged 109 mm, and individual events varied from 1 to 187 mm. All major movement events (> 10 mm in 4-10 days) were precipitation-induced. A non-linear relationship exists between movement rates and Antecedent Precipitation Index, which has a daily recession coefficient of 0.87. The API threshold for movement initiation was estimated to be 160 mm, based on 16 documented major events between 1984 and 1990. Groundwater level at the landslide responded to

precipitation very quickly, with lag time usually less than 3 days. Movement started on days when the groundwater level rose above 2.5 m below ground surface, and a non-linear relationship exists between daily movement rate and groundwater level. Based on available data, there appears to be no change in movement characteristics of Condon landslide after two-third of it was clearcut in 1987.

Wilhelm landslide, located near Condon landslide, has a similar movement pattern, but smaller movement magnitude (averaged 34 mm per year, 1985-1990). The Mid-Santiam and Jude Creek landslides in the volcanic terrane of the western Cascade Range move at much faster rates, averaging 3.8 and 7.8 m per year from 1982 to 1990, respectively. Unlike the Condon and Wilhelm landslides, where individual movement events correspond with individual storms, these two western Cascades landslides exhibit prolonged movement. The Mid-Santiam landslide moves all year, and annual movement shows little variation over the year. The other studied landslides all have large intra- and interannual variation in movement rates, and movement generally stops in the summer dry period. The Lookout Creek landslide (average annual movement = 79 mm, 1981-1990) has slowed in the past four years, and has exhibited movement patterns similar to the storm-dominated Coast Range slides.

Geology and climatic patterns are the two most important factors contributing to the observed differences in timing and style of movement in the landslides studied.

Climatic patterns trigger movement events, and geology influences movement patterns through control on geotechnical properties of landslide materials. These factors can be used to classify landslide movement patterns on a regional scale.

Controls on Movement of Selected Landslides in the Coast Range and Western Cascades, Oregon

By

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CONTROLS ON MOVEMENT OF SELECTED LANDSLIDES IN THE COAST RANGE AND WESTERN CASCADES, OREGON

I. Introduction

Large-scale (1+ ha), deep-seated, slow-moving landslides (referred to as earthflows on the west coast of the U.S.) are common in the highly dissected Oregon Coast Range, as well as in the steep terrains in the Oregon Western Cascades. Most landslides appear to have histories spanning many years to millennia; they occur in forested, as well as clearcut and roaded lands. Indications of the recency or present rate of movement vary greatly across most larger landslide features.

The movement patterns of these landslides are complicated, and are believed to be controlled by factors like bedrock geology, geohydrology, landslide material properties, and climatic parameters (Swanson and Swanston, 1977, Keefer and Johnson, 1983, Ziemer, 1984, Graham, 1985, Iverson, 1986, and Iverson and Major, 1987). Geology and hydrology are the two major factors that determine the spatial occurrence of such landslides. Structural features of bedrock (e.g. bedding planes and faults) are important factors that determine the type, location, and geometry of landslides. Physical and mechanical properties of landslide materials are primarily controlled by bedrock geology and colluvial deposits such as glacial tills. Temporal and spatial patterns of precipitation on unstable hillslopes influence groundwater flow, thus modifying the effective stress states of the hillslope materials as well as the hydraulic gradients on the slope, which in turn contribute to hillslope instability and generate movement.

The Coast Range and western Cascades are situated in two distinctively different geologic terranes. The central Coast Range is underlain by thick (up to 4 m) Tertiary turbidite sandstone beds, with thin mudstone interbeds. These rock units typically dip 10-15° (Chan and Dott, 1983). Late Cenozoic basalt flows dot a small part of the Coast Range, and overlie the turbidite formations. The western Cascades are underlain by an assortment of Tertiary volcanic rocks, including andesitic flows, lahars, pyroclastic units, and water-reworked volcaniclastic sediments. Widespread hydrothermal alteration occurs in this region, resulting in formation of clay minerals and thus weaker rock formations and associated soil units.

Landslides in Oregon exhibit variable response to precipitation at storm event, and seasonal time scales. Previous research shows that most landslides with documented movement histories in the western Cascades typically move continuously throughout the wet season, and some landslides have been observed to advance more than five meters per year (Hicks, 1982, F. J. Swanson, pers. comm., 1989). Large landslides in the Oregon Coast Range generally move over shorter periods of time, and at greater instantaneous rates than their western Cascades counterparts. Most Coast Range landslides move less than one meter per year, except where large-scale catastrophic movement takes place, as in the case of the 20-ha Drift Creek Landslide, parts of which moved several hundred meters in a period of seconds to hours (Swanson and Lienkaemper, 1985). The western Cascades and the central Coast Range lie within the same climate region, but the western Cascades get ample snowfall in the winter while The Coast Range is covered by snow only occasionally. Therefore, the styles of landslide movement between these two areas could be compared on the bases of geology and climate parameters.

II. Objective

The overall objective of this study is to interpret movement characteristics (e.g. timing in relation to wet/dry seasons and individual storm events, and instantaneous movement rate) of selected landslides, based on analyses of geomorphology, properties of landslide materials, precipitation history and pattern, and geohydrology. The overall objective can be divided into two components :

(1) The Condon Creek landslide in central Oregon Coast Range is studied in detail, in terms of its physical characteristics, movement history, hydrologic properties, and the possible effects of clear-cutting of the forest on the landslide. Precipitation, groundwater, and movement data are compiled, and antecedent precipitation index method (API) is used to probe the relationship between the temporal occurrence and magnitude of the precipitation events and the timing of slide movement. Effects of logging on movement of the Condon Landslide, as well as comparing these characteristics before and after clear-cutting will be discussed. In addition to the author's own field investigation and interpretation, this study summarizes and interprets long-term field data sets collected by the Forestry Sciences Laboratory Research Work Unit 4356.

(2) The results from the Condon Landslide, along with observations from the Wilhelm Landslide, are compared with Lookout Creek, Mid-Santiam and Jude Creek Landslides in the western Cascades. Differences in movement characteristics are examined between the two regions in terms of geology, climate, and mechanical and physical properties of landslide materials. This project capitalizes in part on analysis of long-term monitoring of movement of selected landslides in western Oregon. The contribution of this study is comparative analysis of movement characteristics of these landslides.

III. Western Oregon Mass Movement Environment

Introduction

A great variety of mass movement features (landslides, debris avalanches, debris flows, etc.) is common in the mountainous areas of western Oregon. Small, rapid slides commonly occur in areas associated with human activities, such as roaded slopes and clear-cut areas; but large, deep-seated landslides occur naturally in steep, hilly, tectonically active regions underlain by unstable bedrock, such as the Pacific Rim. The western Oregon components of this system are western Cascades and Coast Range. Numerous researchers have conducted landslide mapping and detailed landslide investigations on the two regions; for example, Swanson and Swanston, 1977, Swanson and Lienkaemper, 1985, Ketcheson 1978, Graham 1985, Hicks 1982, Mills, 1983, Pyles et al, 1987. In this study, five landslides in the western Cascades and the Coast Range are selected for comparison: Mid-Santiam, Jude Creek, Lookout Creek, Condon, and Wilhelm.

Geography and Geomorphology

The Condon and Wilhelm landslides are located within the North Fork Siuslaw River basin, in Central Oregon Coast Range, about 14 km north-west of Mapleton (fig. 1 & fig. 2). The Oregon Coast Range is very rugged, and the topography is

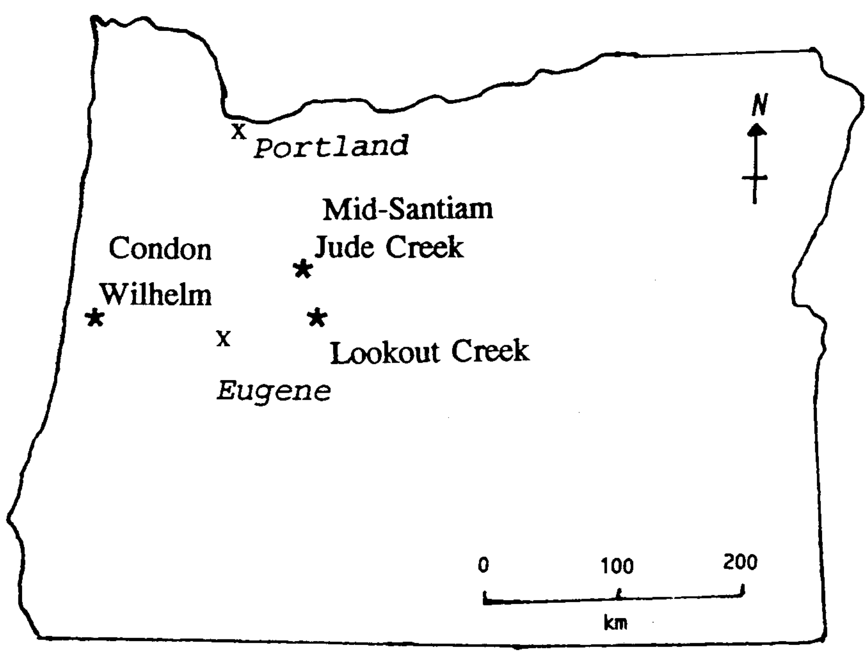

Figure 1 Location map of the five selected landslides described in this thesis.

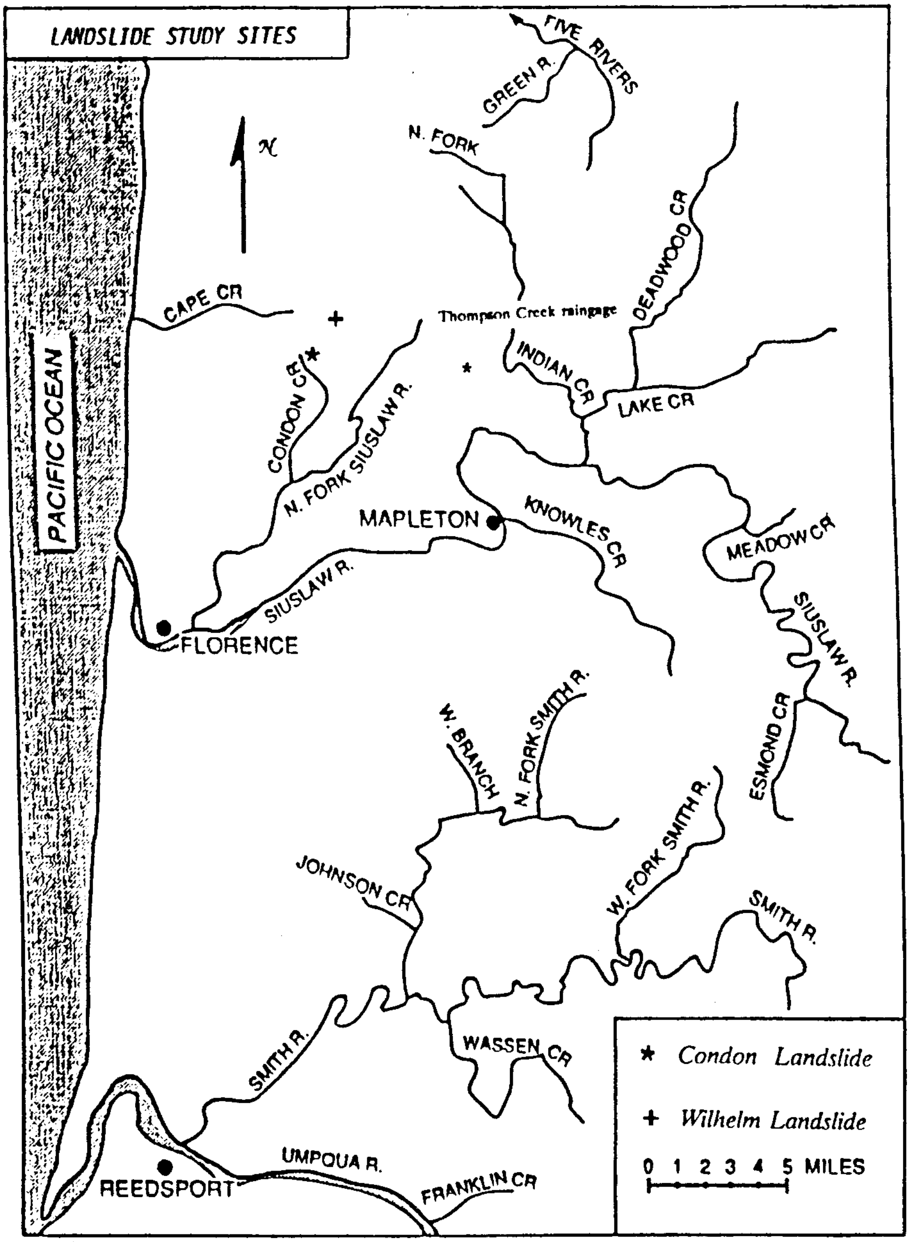

Figure 2 Location map of Condon and Wilhelm Landslides, and Thompson Creek and Mapleton raingages in the central Coast Range. The Condon raingage is located on the Condon landslide site.

dominated by short steep slopes and young V-shaped valleys. Hermans Peak, the highest mountain within 10 km of the Condon-Wilhelm study area, reaches 635 m above mean sea level. Fluvial processes have been actively dissecting the mountains, and the narrow valley bottoms are covered with Quaternary sediments. Graham (1985) identified and mapped a variety of mass movement features in this area, including slump-earthflows and blockslides (Varnes, 1978) on both forested and clear-cut slopes. The Condon and Wilhelm landslides have been monitored by the Forestry Sciences Laboratory. Graham (1985) also hypothesized that a major portion of the North Fork Siuslaw River Basin was underlain by dormant landslides, which may have been active during the Pleistocene.

The Mid-Santiam and Jude Creek Landslides are located in the upper part of the Middle Santiam River drainage basin (fig. 3). This part of the western Cascades has widespread mass movement activity. Mass movement processes, glaciation, and vigorous fluvial erosion have dissected the Middle Santiam River basin into an rugged terrain with high relief. Hicks (1982) mapped 61 km^2 of the basin, and reported approximately 22% of the land area in landslide landforms, although not all the landslides are active at the present time. The Mid-Santiam and Jude Creek Landslides (as named by Hicks 1982) are among the most active landslides

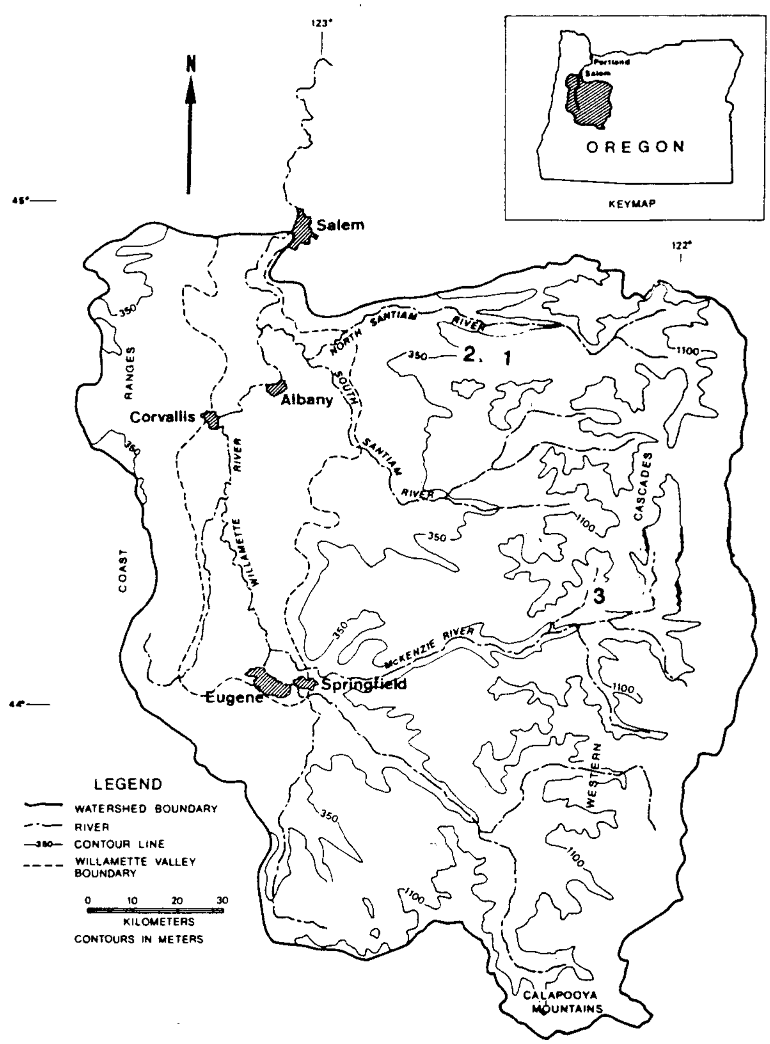

Figure 3 Location map of Mid-Santiam, Jude, and Lookout Creek Landslides in the western Cascade Range.

in the Middle Santiam drainage basin.

The Lookout Creek Landslide is located within the H. J. Andrews Experimental Forest in the western Cascades, about 80 km east of Eugene (fig. 3). It is situated at the north bank of the Lookout Creek, in the Blue River drainage basin (Swanson and James, 1975). The lower half of this 64 km^2 basin is underlain by hydrothermally altered volcaniclastic bedrock. About 25% of this area is in landforms generated by landslide activity at lower elevations. Hummocky landscapes, some more than 6,700 years old, have created deranged drainage systems.

Geology

The central Coast Range is underlain primarily by the Middle Eocene Tyee (Flournoy) Formation, which consists of sandstone beds interlayered with thin mudstone/siltstone beds. These repetitious sequences were deposited by turbidity currents with the outer fan facies (Snavely and Wagner, 1963, Chan and Dott, 1983). The Tyee Formation in the study area dips gently to the southwest, and no major geologic structures were reported there.

Yachats Basalt Formation of Upper Eocene age is found scattering around the western part of the study area: its exposure is very poor. It lies unconformably above the Tyee

(Flournoy) Formation, and intrudes into the Tyee Formation in the form of dikes and sills in numerous locations. It forms a lenticular pile of subaerial lava flows and flow breccia, and usually forms hilltops in the area (e.g. Hermans Peak). Few active landslides were found within this formation (Graham, 1985).

The two study areas in the western Cascades are underlain primarily by the late Oligocene to early Miocene Little Butte Volcanic Series, which is composed of a variety of basaltic to andesitic lava flows, and volcaniclastic materials such as volcanic breccia and welded tuff (Sherrod and Smith, 1989, Peck et al., 1964, Hicks, 1982). This formation covers more than half of the study sites. The Sardine Formation of Middle to Late Miocene overlies the Little Butte Volcanic Series, and is present at elevations above 850 m. Andesitic lava flows and pyroclastic volcanic rocks are the principal components of the Sardine Formation.

Hydrothermal alteration of the volcaniclastic rocks is widespread, especially in the Middle Santiam study area. Zeolitic alteration facies is predominant, and weathered materials from these altered volcanic rocks are rich in halloysite, smectites and montmorillonite (Taskey et al, 1978). The enrichment of clay and amorphorous materials is believed to contribute to development of mass movement topography of the region.

Soil

Soils in the Central Coast Range are mostly clayey silt, or silty clay, and they are poorly developed and shallow (Corliss, 1973) Soils originated from the Tyee Formation generally have high internal friction angle (35 - 41°) and are cohesionless (Schroeder and Alto, 1983).

Soils in the H. J. Andrews Experimental Forest generally have loamy surface horizons, ranging from silty-clays to sandy and gravelly loams, with over 50% of the volume being marcopores (Ranken, 1974). Soils vary greatly in thickness, and most have colluvial and glacial origins. In cases where soil was derived from altered volcanic rocks, such as soils covering the Middle Santiam River basin, clay is abundant in all soil horizons, with hydrated halloysite and amorphous materials being the major components (Taskey, 1978).

Climate

Both mountain ranges are influenced by modified Mediterranean-type maritime climate, with cool, wet winters and warm, dry summers. In the central Coast Range, average annual precipitation ranges from 1500 to more than 2500 mm. Mapleton has about 2200 mm of precipitation annually, but mountainous areas probably receive more precipitation because of orographic effects. Most of the precipitation falls in the winter months as rain. Snow falls rarely on higher ground, and it generally melts within days. High intensity precipitation in winter months contributes to development and movement of landslides in the area.

The central western Cascades receive about 1000 mm to over 2500 mm of precipitation annually, depending on location and elevation. Snow falls heavily in the winter at elevation above approximately 1000 m, and seasonal snowpacks more than a meter deep are common in open areas. The melting of seasonal snowpacks in spring, coupled with rain-on-snow events in winter, generate large runoff events, and may cause prolonged periods of landslide movement.

Vegetation

The western Cascades were once covered by extensive stands of mature (100-200 years old) and old-growth (200+ years old) forests, but several decades of logging have reduced their extent. Principal tree species found in the study areas are Douglas-fir (*Pseudotsuga menziesii*), western hemlock (*Tsuga heterophylla*), and western red cedar (*Thuja plicata*) at lower elevation. Higher elevation stands consist of subalpine fir (*Abies lasiocarpa*), Pacific silver fir (*Abies Amabalis*), and mountain hemlock (*Tsuga* *mertensiana*). Red alder (*Alnus rubra*) is often found as a pioneer species on disturbed sites and in riparian zones.

The forest stands in the Coast Range are similar to those found in lower elevations of the western Cascades, which consist mostly of Douglas fir, western hemlock and western red cedar. Understorey vegetation includes many species, and is dominated by devils club (*Oplopanax horridum*), vine maple (*Acer circinatum*), and rhododendron (*Rhododendron macrophyllum*).

IV. Geology of the Central Coast Range

Introduction

The central Oregon Coast Range is predominantly underlain by Cenozoic sedimentary and volcanic rocks (Fig. 4). The study area consists of more than 2000 meters of Lower Eocene Tyee and Flournoy Formations. This thick sandstone-siltstone sequence rests unconformably on the Siletz River Volcanics of early and middle Eocene age (Snavely and Wagner, 1963). The Tyee-Flournoy Formation is underlain by the younger Yachats Basalt at some locations (Snavely and Wagner, 1963, Graham, 1985). A small number of thin Cenozoic basaltic to felsic dikes cut into the Tyee-Flournoy Formation in Central Coast Range, especially at areas east of Mapleton (Schlicker and Deacon, 1974). The Cenozoic units were tilted gently, and were only slightly deformed by minor faulting and folding. The following section provides brief description of the geology of the study area, with emphasis on the Cenozoic sedimentary rocks, because they cover most of the area, and the Condon Creek drainage basin is underlain by them.

Tectonic Setting

The Cenozoic tectonic history of western Oregon was controlled primarily by interactions among the Farallon, Juan de Fuca and North American plates, as described by

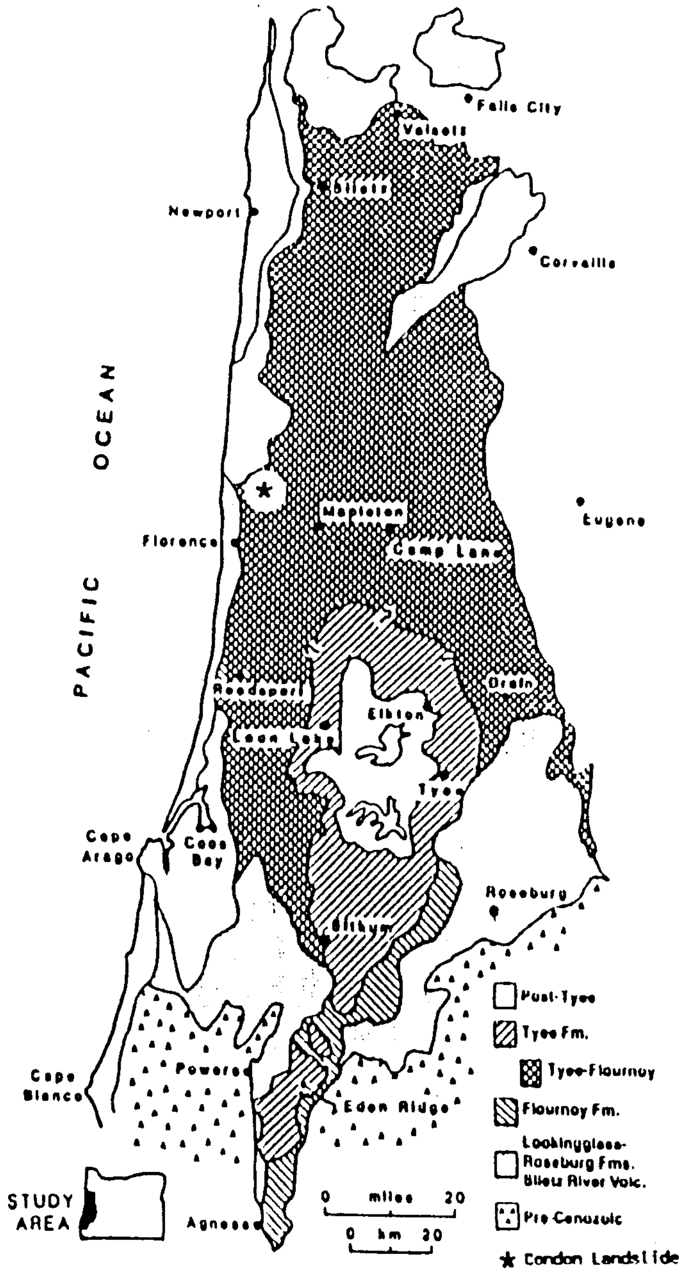

Figure 4 Simplified geologic map of Eocene formations in the Oregon Coast Range (Modified from Chan and Dott, 1983).

numerous authors (e.g. Baldwin, 1974; Snavely et al, 1980). During the Eocene Era, active subduction occurred to the west of the present-day Oregon Coast Range, and was the site of consequent rapid deposition of thick sediments in a deep forearc basin with an elongated geometry, bounded by an accretionary wedge of deformed oceanic sediments and basalts to the west (Snavely, et al, 1980, Chan and Dott, 1983). Tectonic models of western Oregon with rotations of more than 50° clockwise were suggested, based on paleomagnetic evidence (Simpson and Cox, 1977; Magill and Cox, 1980). During the post-Middle Miocene rotation, the narrow "Tyee-Flournoy basin" attached to the ancient Klamath Mountains terrane moved westward to the present location, due to back-arc spreading behind the new subduction zone between the Farallon and North American Plates (Simpson and Cox, 1977; Heller et al., 1985).

After sediment had filled the forearc basin, magma from vents and fissures formed basaltic dikes and sills in central Coast Range. Later shallow sea deposits, followed by tectonic uplift and subsequent volcanism, produced the present geology of the area.

<u>Stratigraphy</u>

In the central Coast Range, the Lower Eocene sequence of Lookingglass, Roseburg, and Flournoy Formations overlies the pre-Cenozoic complex sedimentary, metamorphic, and volcanic units of the northern Klamath Mountains (Baldwin, 1974). Near Mapleton only the Tyee-Flournoy Formations, and the Upper Eocene Yachats Basalt appear at the surface on the rugged central Oregon Coast Range.

Tyee-Flournoy Formations

Most of the central Coast Range is made up of a thick sequence of sedimentary beds named the Tyee Formation many years ago. But recently attempts have been made to distinguish the older Flournoy Formation as a separate stratigraphic unit underlying the Tyee Sandstones (Baldwin, 1974). The distinction was best observed at the southern Coast Range, but in the central and northern Coast Range, differentiation between the two units is difficult. Hence, the term "Tyee-Flournoy Formations" is used in this study, to reflect the undifferentiated nature of the two formations at Central Coast Range. Microfossils of Early Eocene age in the Tyee-Flournoy Formations are rare, and were described by numerous researchers (e.g. Mckeel and Lipps, 1975).

More recent sedimentary facies studies on the Tyee-Flournoy Formations revealed that both formations represented a prograding and shoaling sequence, that were accumulated in a narrow forearc basin resting on a tectonically active plate boundary (Chan and Dott, 1983).

Distinctive facies changes, from deltaic to basin plain environments developed within the Tyee-Flournoy Formations progressively from south to north (Chan and Dott, 1983). They concluded the Eocene Tyee-Flournoy strata were deposited in a submarine-fan setting characterized by abundant sand spilling or cascading across a very narrow shelf and slope, within the narrow-elongated-shape of the ancient forearc basin.

Paleocurrent and petrographic studies indicated that sediments of the Tyee-Flournoy Formations were derived in part from the Klamath Mountains, and the volcanic detritus was brought in from a volcanic arc system east of the Klamath Mountains (Dott, 1966; Chan and Dott, 1983). However, isotopic investigations of the Tyee-Flournoy sandstones indicate that most of the white mica and potassium feldspar detritus were derived from a unique granitoid source, probably from the Idaho batholith (Heller and others, 1985). A possible model utilizing the tectonic rotation hypothesis, was proposed by them, to accommodate both findings.

The lithology of the Tyee-Flournoy Formations varies greatly with different sedimentary facies, from coarse, thickly-bedded sandstone and conglomerate related to deltaic environment to the south, to beds of fine-grained mudstones of basin-plain facies at the northern end of the Formations. The Tyee-Flournoy rocks in the central Coast Range study area generally belong to the outer fan facies, which is characterized by repetitive, graded, and rhythmically bedded 0.5 to 1.0 meter thick sandstones, interlayered with thin (less than 0.5 meter) siltstone-mudstone intervals, which are rich in macerated plant debris and mica (Lovell, 1969; Chan and Dott, 1985). The sandstone/shale ratio averages 60 to 40% (Chan and Dott, 1983).

The Tyee-Flournoy sandstones are generally moderately- to poorly-sorted, arkosic, micaceous, and their mineralogic composition varies with respect to grain size (Fig. 5) (Chan and Dott, 1983). Diagenetic trend of the Tyee-Flournoy Formations was found coupling with the progressive change of the depositional facies within the Formations: Deltaic and inner fan facies have zeolite cement- and clay-rich framework and higher porosity, while the outer fan facies sandstones are characterized by lack of clay cement, and predominantly matrix is found in their poorly-sorted framework, which is thought to be the result of rapid deposition and basinal compaction (Chan, 1985).

Yachats Basalt

The Yachats Basalt crops out along the coastal area north of Florence, and forms a series of east-west linear ridges which terminate as prominent headlands, such as Cape Perpetua. It is composed of a series of subaerial and submarine flows and volcaniclastic rocks. Its heterogenous assemblage contains porphyritic basalt, basaltic andesite flows, and flow breccia, and is estimated to average 750 m thick (Schlicker and Deacon, 1974).

Basalt flows are the dominant lithology, and most flows are 3 to 8 m thick. Jointing is common, and the upper part of a typical flow is brecciated (Graham, 1985). Many of these flows are believed to have originated from local vents and fissures (Snavely and MacLeod, 1974)). The age of the Yachats Basalt is estimated to be Upper Eocene, based on its stratigraphic relationship with the underlying Early Eocene Tyee-Flournoy Formations and Middle Eocene Nestuca Formation (Snavely and MacLeod, 1974).

Structural Geology

The central Oregon Coast Range lies at the western flank of the Coast Range geanticline, which consists of northeast-trending anticlines and synclines developed within the Tyee-Flournoy Formations. The Tyee-Flournoy rocks in the vicinity of the Condon and Wilhelm landslides lie within the west flank of an anticline, and the sedimentary beds dip gently (5° to 15°) to the southwest. The structural relationship of the Yachats Basalt is obscure due to its a

Tyee-Flournoy sandstone composition for n = 68 samples

Mineral Constituent	Mean %	± Std. Deviation
quartz	19.0	4.0
K-feldspar	5.0	2.2
plagioclase	22.3	4.8
VRF	19.4	7.6
MRF	3.0	2.6
SRF	3.7	3.1
muscovite	0.9	0.7
biotite	2.8	1.8
matrix	9.2	7.3
clay cement	2.2	3.5
calcite	3.6	7.0
zeolite cement	1.9	3.5

b

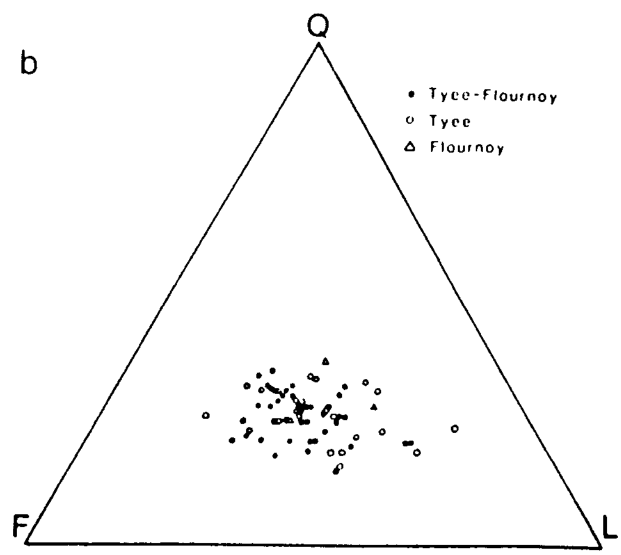

Figure 5 Grain size and composition of the Tyee and Flournoy Formations. a: Mean percentages of mineral constituents (VRF = volcanic rock fragments, MRF = metamorphic rock fragments, SRF = sedimentary rock fragments). b. Ternary diagram of normalized quartz (Q), feldspar (F), and lithic (L) components (from Dott and Chan, 1983).

limited exposure, but the orientation of its beds suggests that it was emplaced before the folding of older stratigraphic units (Schlicker and Deacon, 1974).

Very few faults are reported in the central Coast Range because of poor bedrock exposures and dense vegetation. MacLeod and Snavely (1968) used radar imagery to define physical lineations in the central Coast Range. Some of these prominent linear features were inferred to be faults. The Perkins Creek Fault, uncovered by Macleod and Snavely (1968), trends northwest along part of the Perkins Creek. Schlicker and Deacon (1974) speculated that the prevailing trend of most faults in the area would be northwest, as they suggested the northwest trending dikes within the Tyee-Flournoy Formations were emplaced along weak fault zones.

V. The Condon Landslide

Location

The Condon Landslide is located at the Mapleton Ranger District, Siuslaw National Forest, in the Central Oregon Coast Range, and is about 14 km northeast of Florence, in NE 1/4, sec. 10, T17S, R11W, Willamette Base Meridian, (fig. 2). It can be accessed by unpaved Forest Service roads connected to Oregon Highway 36 in the south, and to U.S. Highway 101 in the west. The landslide is in the middle of a southwest-facing slope near the headwaters of the Condon Creek.

On-site Geology

The entire Condon Landslide complex is underlain by the Tyee-Flournoy Formations. Local outcrops within or near the slide are very scare, and those observed are generally badly weathered, making stratigraphic measurement very difficult. Moreover, most outcrops of the bedrock were disturbed by mass movement activity, and the original orientations of bedding planes and structural features were masked or even destroyed. A badly weathered and decomposed Tyee section, 10 meters thick, was found at the top of the ridge, closes to where the landslide is located. The orientation of the poorly exposed sandstone beds hints that the rocks dip to the southwest, thus providing evidence that the landslide is sitting on the dip slope of the Tyee-Flournoy Formations. Roadcuts along the North Fork Siuslaw River furnish good exposures of the Tyee-Flournoy Formations. One section of the exposure, which is approximately 5.6 km southeast of the landslide, has repeated sandstone beds 1 to 3 m thick, interlayered with thin (0.3 m) siltstone beds, which trends N20°W, and dips 12° to the southwest, No apparent major structural features were observed in these outcrops.

The Tyee-Flournoy Formations within the landslide complex contain mostly massive sandstone, which has a predominantly brownish-gray color due to weathering. The sandstone beds are consistently one meter thick, and are interbedded with siltstone-mudstone material. The siltstone layers are typically only a few centimeters thick, grayish, and weather rapidly into brittle rock fragments (Fig. 6). These repetitive and graded sandstone beds with thin siltstone interlayers represent the outer fan facies of the Tyee-Flournoy Formations (Chan and Dott, 1983).

Hand specimens of the Tyee-Flournoy sandstone from the landslide site show they are medium- to coarse-grained rock, with little structural features. The rocks are made up mostly of plagioclase, quartz and volcanic fragments. The mineralogy of the sandstone beds here are apparently similar to the mineralogical content of other Tyee-

Figure 6 Outcrops of weathered Tyee-Flournoy strata at the western toe of the Condon Landslide. Photo shows massive sandstone units overlying disintegrated thin siltstone beds about 0.3 m thick. View is to the northeast.

Flournoy sections. The micaceous and arkosic sandstone beds weather into silty-clay soil mantle, and are locally sandy. The soil layers are poorly developed at most slopes, partly due to their colluvial origins. The soil is generally cohesionless, light-weight, has low water content, and has relatively high internal friction angle (Schroeder and Alto, 1983).

Engineering geology properties of the Tyee-Flournoy Formations were described by Schlicker and Deacon (1974) and Graham (1985). Generally the dip slope of the Tyee Formation has Type II stability problems (Burroughs et al., 1976), characterized by slumps within the soil mantle, or blockslides with weathered bedrock and soil masses advancing along weak (failure) surfaces in a translatory manner. In the Condon Landslide, the weak surface may be located at thin siltstone-mudstone layers, over which the sandstone/siltstone beds are gliding. The relatively low cohesion in the Tyee sandstones is probably the result of small amount of clay present in the sandstones, which probably affects the movement characteristics of Condon Landslide.

Management Activities

The Condon landslide lies within the Siuslaw National

Forest, where stands of mature Douglas fir-western hemlock-red cedar once covered most of the area. "Multiple use" has been the management concept of the Forest Service, and forest areas surrounding the landslide have been modified by human activities. Networks of Forest Service roads and logging trails cut through the otherwise virgin forests, and scattered clearcut patches, along with leaves areas designated by the Forest Service to protect stability of steep slopes and wildlife habitat along streams. Several clearcuts are located within the Condon Creek watershed near the landslide site, and five- to fifteen-year old forest stands now occupy most of these sites.

The Condon Landslide was covered by indigenous forest of Douglas fir-western hemlock-red cedar, which originated after forest fire in the mid-1800's. Approximately two-thirds of the Condon landslide area was clearcut in 1987 (fig. 7), as part of the landslide research project administered by the Forest Service PNW Research Station at Corvallis, Oregon. The clearcut plot was then immediately slash-burned. The site is being revegetated with shrubs and Douglas-fir trees.

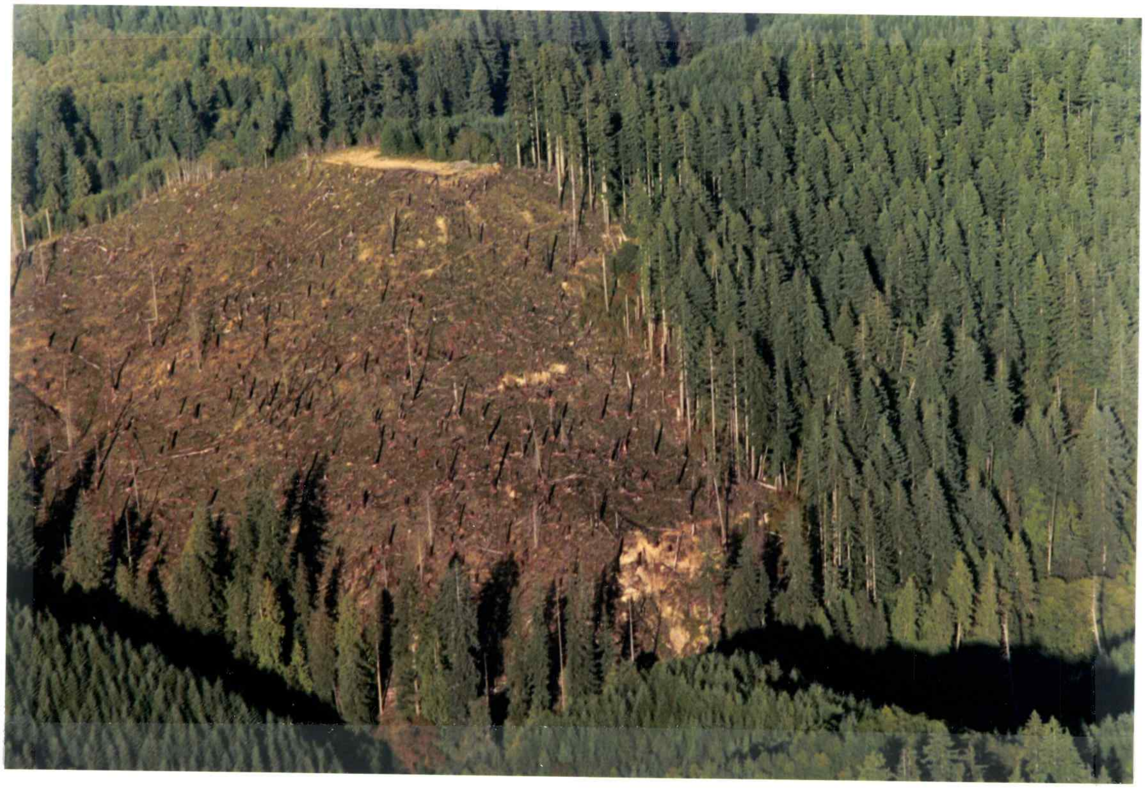

Figure 7 Aerial view of the Condon landslide. Note the clearcut part of the landslide at the left side of the photo. View is to the northeast.

Site Geomorphology

The 5.5-hectare landslide forms a large structural mid-slope bench, at an elevation between 180 and 250 meters (figs. 7 & 8). The entire slide rests on a flat surface, which was possibly formed by ancient blocksliding, and extends at least 100 m beyond the toe area of the Condon slide, almost to the bottom of the valley (Graham, 1985). The slope of the bench is gentle (5° to 10°), but becomes very steep in the toe and headscarp areas.

A refraction seismological investigation conducted by Graham (1985) estimated the thickness of the soil units to be 1.5 to 2 meters, and he calculated the base of the decomposed bedrock unit to be between 4.7 to 6.7 meters below ground surface. The blockslide configuration, and zones of weathered and disturbed bedrock may create thick layers of landslide colluvium in some areas. The bottom of the piezometer tube installed in 1989 rested on decomposed bedrock, which was 5 meters below ground surface.

Graham's (1985) landslide map provides an overall view of the slide (Fig. 8). Detailed descriptions of the geomorphology of the slide was done by Graham (1985), based on his field work conducted in the early 1980's. Since then, modifications of surface features have taken place, especially in the western part of the landslide, where

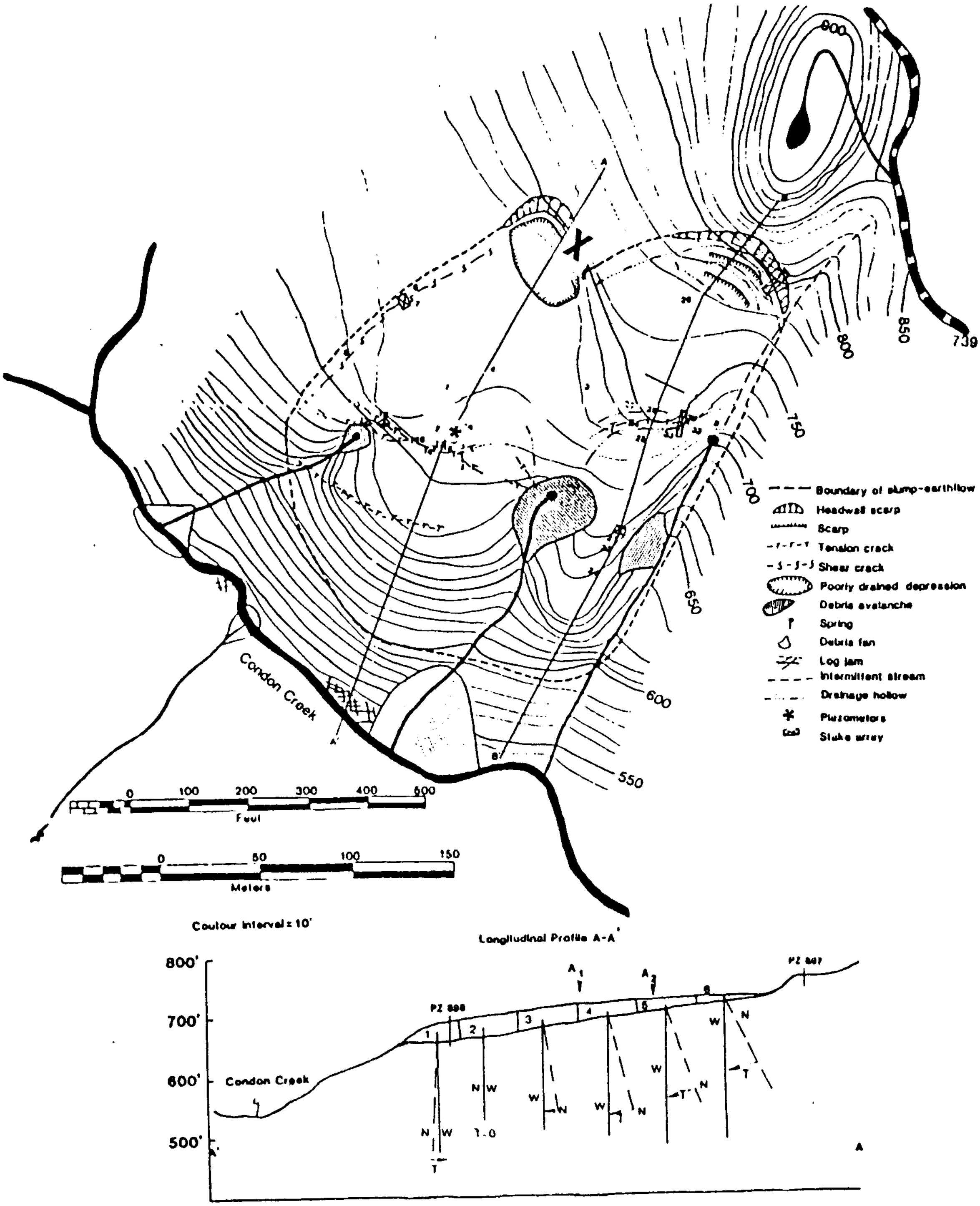

Figure 8 Topographic map and a simplified cross-section diagram of the Condon Landslide (From Graham, 1985). The X marks the site of the extensiometer-stake array pair providing movement record in this study.

clearcutting revealed features previously obscured by forest cover, and new sliding has created new features. The following is the account of the field observations done from winter 1989 through the winter of 1990.

The headwall of the Condon Landslide has two semicircular scarps that are concave downslope (figs. 8 & 9). The northeastern scarp consists of a series of smaller scarps with individual vertical displacements as much as 10 meters, but it appears to be presently inactive. Only soil rilling, and small scale slumps were noted. Small grabens are found below the northeastern scarp, and sag ponds were seen within these grabens during wet seasons (fig. 9). The north-central scarp in the clearcut area is currently very active, and is the site of the extensiometer installed on the landslide. Its central part is approximately 3 m high, and tension cracks occur above and below it. Beginning in the winter of 1989, a lateral-tension crack about 5 m long opened and widened up at the bottom of the central part of the north-central scarp, where the extensiometer is located (fig. 10). The central part of this headwall has since become unstable, and it is apparently slumping slowly downslope, as several parallel tension cracks are seen at the top of this head scarp.

The common boundary between the two slide bodies

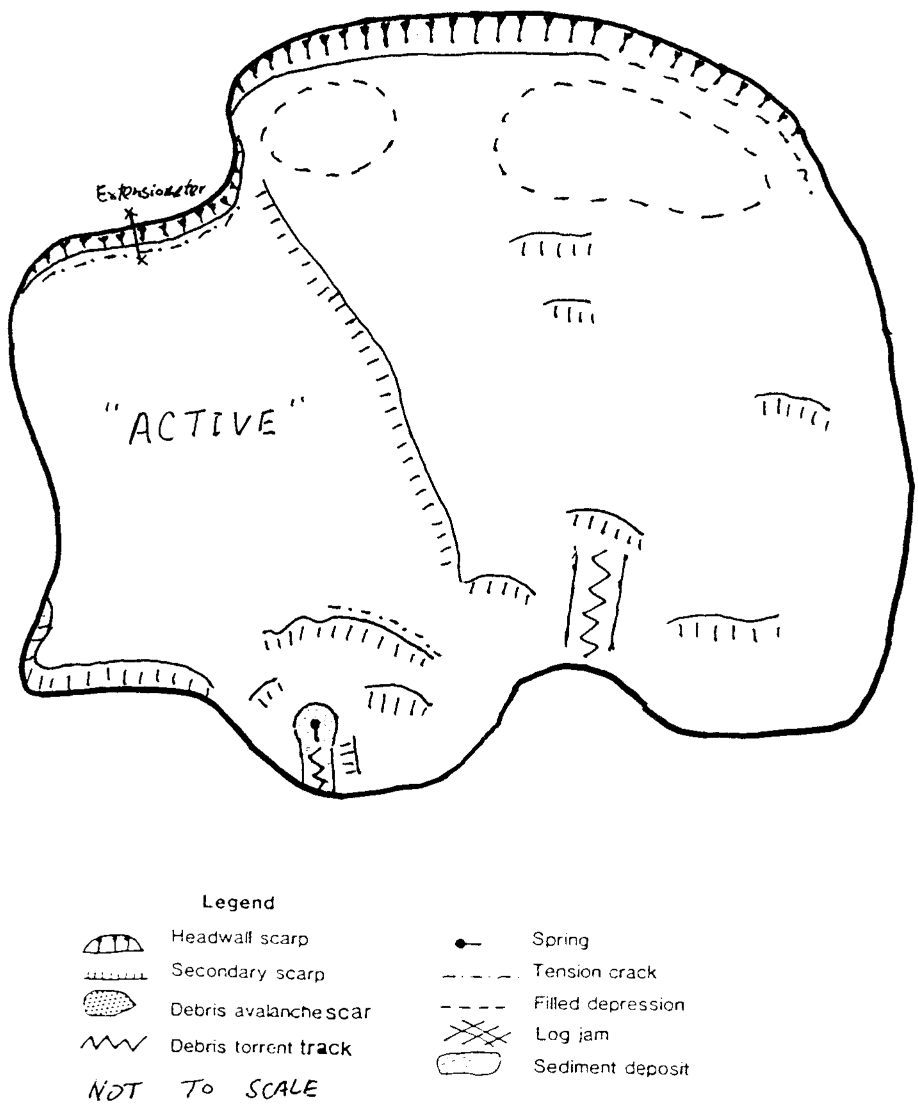

Figure 9 Geomorphic units of the Condon Landslide. "Active" on the aerial photo overlay identifies parts of the landslide which have shown signs of mass-movement activity between 1988 to 1990.

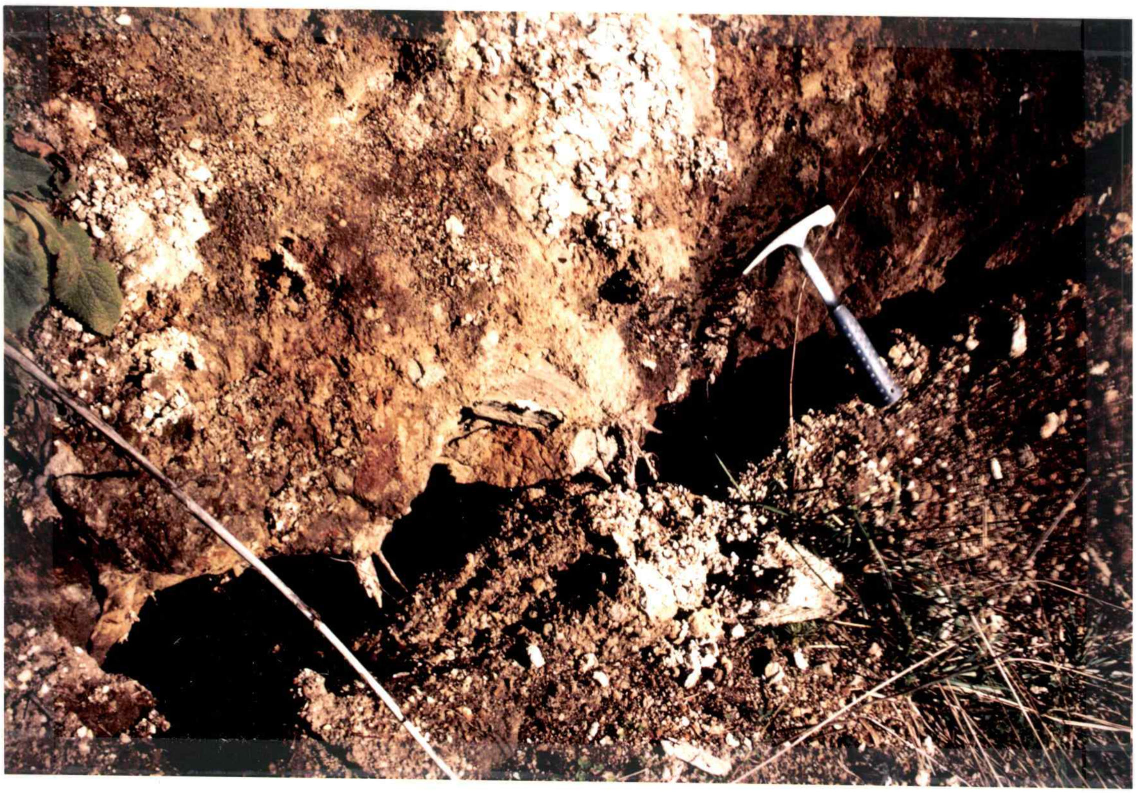

Figure 10 Tension crack at the Condon northwestern headscarp. The tension crack has been widening since January, 1989, and is still active.

described by Graham (1985) is marked by a discontinuous and complicated scarps from less than 0.5 m to approximately 1.5 m high (fig. 9). Part of the scarp has fresh soil exposures indicating recent activity, and this scarp roughly follows the boundary of the clearcut area. The lower part of the scarp makes a gradual transition to a shear crack, which is monitored by a stake array, and the movement has been very minor in recent years. Based on the measurement record, the stake array shows the western part of the slide is moving to the southwest, relative to the eastern part of the slide, which is apparently more stable. This scarp might have existed for a long time, but became more apparent after clearcutting, due to secondary erosion intensified by an absence of litter supply.

The middle part of the slide does not show major signs of deformation, except numerous small grabens at the north-eastern part of the slide, immediately below the northeastern head scarp (fig. 9). In areas with the small grabens, depressions and small scarps were observed, and trees were tipped, but no signs of current major movement were noted. The middle part of the slide appears to be gliding downslope as one complex mass block.

The toe area of the Condon landslide is complex, and current activity is focused on the western part of the toe area. The eastern part of the toe area consists of several scarp; one more than 10 m high. Evidence of old debris avalanches is visible at some locations near the bottom of the eastern toe area. These scarps show little activity in last two years, except soil rilling. Only small debris flows were observed originating from colluvial deposits resting at the bottom of the 10 m scarp.

The western toe area is currently very active and the landscape is highly disturbed (fig. 11). Active transverse cracks are present near the toe area. At the top of the major scarp above the toe, a linear depression about 5 m long formed from the widening of the transverse crack in 1989. Slumping of landslide material has uprooted dead trees, and has revealed bedrock at the headwall of a small drainage at the western end of the toe area (fig. 6). This small drainage is also the site of repeated debris avalanches, with the most recent one occurring in January, 1990 (fig. 12). Spring-fed streamflow exists during most of the year at mid-slope in this drainage, and its location and elevation suggest it is probably related to the predicted failure plane of the slide. The lower boundary (toe) of the slide is marked by rilling and slumping of fresh landslide materials on steep slopes, which border the sides of the small drainage at the western end (fig. 12). The failure plane may reach the surface along the lower boundary of the

Figure 11 Photo showing the northwestern toe area of Condon Landslide. Currently, most of the surface mass-movement activities have been focused in this area.

Figure 12 Debris avalanche site of the western toe area, Condon Landslide. Photo was taken in January 27, 1989, after a series of major storm events, which led to a major movement event in January, 1989.

slide, but surface features provided little evidence except groundwater seepage and different soil appearance.

Movement History

The first detailed description of the movement history of the Condon Landslide was done by Graham (1985). Dendrochronological studies completed by Graham (1985) suggested that between 1950 and 1980, there were three periods very active movement: 1957-58, 1964-65, and 1970-73, which corresponded with above average precipitation. The landslide may have been active for centuries, but its major movement events may have been interrupted by long periods of slight to no movement.

From the late 70's through early 80's, periodic monitoring of the landslide was carried out by Graham (1985). He reported four years (12/79 - 2/83) of movement records, based largely on regular measurements of stake arrays, which had total movement of 0 mm, 67 mm, 143 mm, and 205 mm respectively.

Monitoring work had been carried out from 1984 to 1990 by team members of the Forest Service RWU 4356, producing six years of continuous movement record. Chapter 8 will discuss in detail the movement patterns observed from the record available, and chapter 10 will discuss possible effects from the clearcutting operation.

VI. Field Methods and Instrumentation

Stake Array

Stake arrays (strain rhombuses) are used to measure the direction and magnitude of movement at active zones by manual repeated measurements over a period of time. Stake arrays consist of a group of four stakes arranged roughly in a rectangular formation across an active zone of differential movement, such as tension or shear cracks. The movement of one stake relative to the pair of stakes on the opposite side of a zone of movement can be calculated trigonometrically from repeated measurements of distances between the stakes. George Lienkaemper (person. comm., 1986) estimated the precision of measurement to be $\pm$ 5 mm, although the smallest distance detected by tape measurement is 1 mm.

In the Condon site, several stake arrays were set by Graham in 1980, and three stake arrays were monitored by Forest Service personnel from June, 1984, through June, 1990 (fig. 9). Stake arrays #1 and #3 recorded activity at the north-central and northeastern head scarps respectively, while stake array #2 registered activity along the shear crack related to the linear scarp at the border between the two landslide sub-units.

Extensiometer

Extensiometers utilized for landslide monitoring are modified water-level recorders. The instrument is composed of a cable with one end attached to an anchor on one side of a zone of differential movement, and the other end passes over a pulley tied to a counterweight in the instrument housing (Fig. 13). Pulling of the cable by landslide movement is recorded on a revolving drum or punch tape. When movement occurred, the cable pulled away from the instrument, and the exact timing and rates of movement events were noted on the recording drum. Exact amount of movement could be determined by extrapolating the total measured stake array values onto the extensiometer record over the same period of time. This is necessary because the orientation of the extensiometer tape may not be parallel with the direction of landslide movement, so the extensiometer may not give an accurate reading of absolute movement.

One extensiometer alongside stake array #1 has been logging movement activity at Condon Landslide continuously since June, 1984. The instrument is a modified AE-35 type water-level recorder, with the drum recorder housed in a hut on the active block of the slide, and the other end attached to an anchor at the top of the north-central headscarp, across the active tensional zone. This stake array/

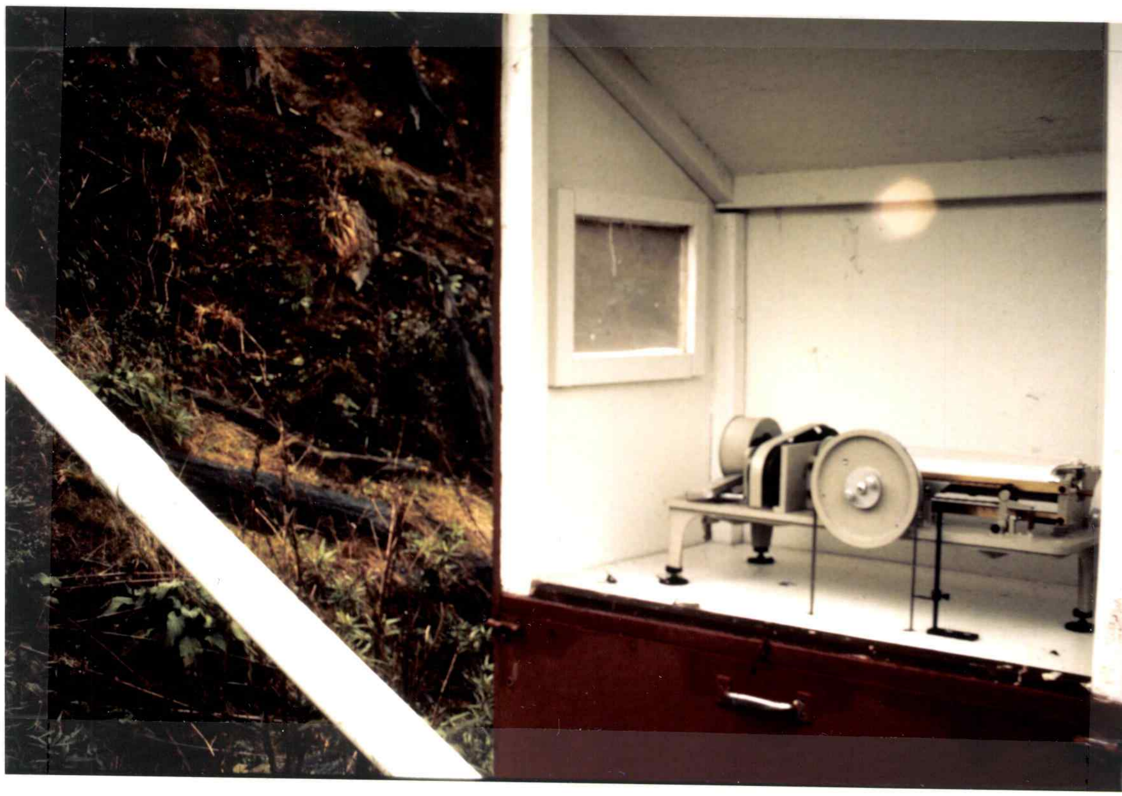

Figure 13 Extensiometer at Condon northwestern headscarp. The white pipe houses the cable where it passes from the instrument house to the anchor on the other side of the active crack.

extensiometer set provided complete movement record of the most active zone (north-central head scarp) at the Condon Landslide from 1984 through 1990.

Recording Piezometer

A recording groundwater piezometer was installed on the Condon Landslide in June, 1989 to monitor groundwater level changes within the landslide. The piezometer well was drilled into one of James Graham's old piezometer sites by using a 76 mm hand auger (fig. 14). The well was drilled to a depth of 5.1 meters, and a 25.4 mm PVC pipe with the bottom ten feet perforated was put in the well. Coarse sand was back-filled to four inches below soil surface. A bentonite cap was installed at the top of the piezometer, and a 76 mm thick concrete slab was placed around the pipe to provide foundation to the instrument hut.

The instrument consists of a pressure transducer attached to a 7.6 meter cable and a Emmos Data Recorder. The transducer received an input of 5 volts from the recorder, and returned a percentage of that voltage to the recorder, in which the returning voltage was controlled and adjusted by the amount of water pressure exerted on the transducer at the bottom of the well. The voltage was read by the recorder every half hour, and was converted from an analog to a digital value (AD unit), and then was stored

Figure 14 Piezometer and recording raingage at the Condon landslide. The instrument box at the left is the Emmos data recorder attached to the transducer below it. The collecting funnel of the raingage is at the top of the tall PVC pipe standing to the right of the hut.

in a Memory Module. The Memory Module was periodically retrieved from the field, and was decoded by a data reader in the lab. The raw data (AD unit) were then calibrated and interpreted by rating equations and were tabulated into daily records.

Recording Raingage

A recording raingage was installed along with the piezometer, and utilized the same Emmos Data Recorder. The pressure created by precipitation collected in the 203 mm diameter PVC pipe (fig. 14) was sensed by another transducer, and half-hourly signals were recorded, along with the piezometer readings. Data decoding processes of the raingage readings were the same as the piezometer data.

Topographic Surveying

Longitudinal and transverse survey lines, consisting of survey points on unstable ground, were installed on the landslide, always with reference points on stable ground adjacent to the landslide area. The lines are surveyed annually, using conventional theodolite or EDM transit. Yearly differences in positions of the survey points reflect landslide movement and changes of surficial features.

Two longitudinal survey lines were installed in 1984, by George Lienkaemper and his colleagues of the Forest

Service. The lines have been re-measured annually until June 1990. These survey lines were used to detect changes of the micro-topography of the landslide complex throughout the monitoring period.

All field data reside in the Forest Science Data Bank, Oregon State University, Corvallis, Oregon.

VII. Precipitation and Groundwater Level Characteristics

Climatic Pattern

The central Coast Range in Oregon is affected by maritime climate, with prevailing westerly winds bringing abundant moisture from the Pacific Ocean. Generally, the October to April period is cool, and the area receives 70 to 80 % of its annual precipitation then. The summer is warm and dry. The annual average precipitation of the area is from 1600 mm at Florence to 2200 mm at Mapleton, but may be much higher in mountainous areas because of orographic effect (e.g. 2415 mm at Thompson Creek during water year 1989) (Swanson & Roach, 1987, and K. Lautz, personal communication, 1990).

Winter storms coming from offshore sweep across the Coast Range, and dump a tremendous amount of moisture on the region. Precipitation distribution throughout the region is very uneven, even during individual storm events, due to the rugged terrain and relatively high relief found in the Coast Range. Duration of rainstorms varies, from a one-day minor event, to a persistent 7-8 days disastrous, torrential storm. Major individual storms can bring more than 200 mm of precipitation in several days.

Precipitation intensity is relatively low, compared to areas in other climatic regions, which have similar magnitude of annual precipitation. Mapleton precipitation records from 1975 to 1984 showed most storm events had 6-hour intensity of less than 50 mm, which had recurrence interval of two years. During the same period, typical large 24-hour events had intensities between 120 to 145 mm, which had return periods ranging from 2 to 3 years (Swanson and Roach, 1987). Records from 1985 to 1990 displayed daily precipitation seldom exceeding 50 mm (1.97 in.). The largest storm event occurred on 2/21-2/23/1986, when a total of 267 mm of precipitation was recorded in 72 hours, along with a 24-hour intensity of 114 mm. This storm coincided with the largest movement events recorded in the same period at Condon landslide, with 186.6 mm of displacement in 5 days. Most precipitation comes as rain, except at higher elevations, where snowfalls result in thin, ephemeral covers of snow on hillslopes infrequently.

Descriptions of Raingages

Precipitation records of three raingages in the Central Coast Range are used in this project: (1) on-site Condon, (2) Thompson Creek, and (3) Mapleton gages (figure 2).

The Condon gage (elevation 220 m) was installed in June, 1989 by the Forest Service. It is about 8 km east of the coast, and is surrounded by mountain ridges over 650 m in elevation. Accurate precipitation data are available from 8/22/89 to 4/12/90, in the form of both hourly and half-daily records, with a detection limit of 0.38 mm/hr.

The Mapleton gage is operated by NOAA, as part of the precipitation reporting network in Oregon. It is located at an elevation of 12 meters, in the Lower Siuslaw River Valley. The gage has been in operation since 1975, and its 1985 to 1990 record is used in this project. The record was obtained from the State of Oregon Climatic Research Institute in Corvallis, Oregon. The data set is in hourly format, and the minimum amount of precipitation detected is 2.54 mm (0.1").

The Thompson Creek gage is managed by the Oregon State University, Forest Engineering Department. It began functioning in 1988, and records from 1989 to 1990 water-year were obtained from Kevin Lautz, the current operator of the Forest Engineering Department raingage network. The gage is located at the upper valley of Thompson Creek (elevation 780 m). Data format is hourly, with precipitation detection limit of 1.02 mm. The record from this gage is used to back up the other two raingages, and to characterize the local precipitation pattern of the study area.

Precipitation Summary of the Three Selected Raingages

Table 1 provides a summary of the amounts of annual precipitation from the three raingages.

Precipitation varies temporally and spatially (Table 1 and Fig. 15). Variations of daily precipitation among the three gages are especially apparent in the major storm events, such as the February 6 to 11, 1990, event. During this storm period, Mapleton's precipitation was only 75% that of Condon, and Thompson Creek received only 65% of Condon's amount. In spite of these variations in precipitation amount, the timing of storm events was uniform among the three gages, with peaks coming within six hours.

Correlation of Precipitation Data Among Selected Raingages

The question was raised whether it was practical to combine records from three raingages for different periods of time, and use them together in an Antecedent Precipitation Index (API) analysis. Regression analyses were performed among the data sets from the three raingages, to examine the correlations among the three gages, and to develop a mathematical relationship between the on-site Condon record and the Mapleton site, where a long-term record exists. The mathematical relationship would be used to extrapolate the existing record from either the Mapleton or Thompson Creek gage to the half-month period of missing

****

	Mapleton	Thompson	Condon
1985	2134 (1689)		
1986	2400 (1966)e		
1987	2052 (1717)		
1988	2258 (1750)		
1989	2552 (2118)e	2415 (1934)	
1990	(1786)	(1616)	(1927)

****

Table 1 Summary of annual precipitation of the three selected raingages from water years 1985 to 1990. Figures in parentheses indicate the amount of precipitation received from October to March (wet season) of the water year (e = estimated values on some periods).

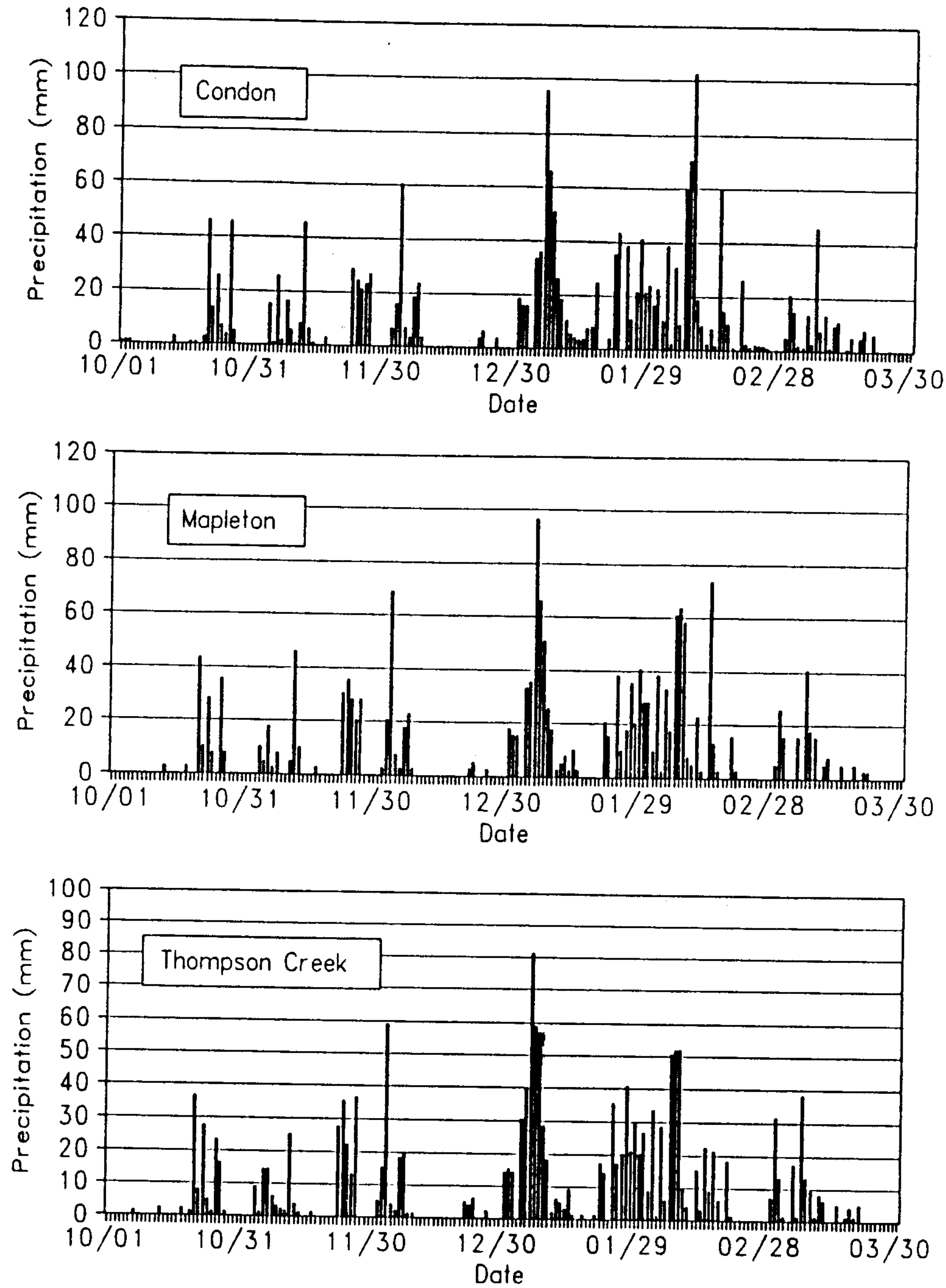

Figure 15 Precipitation Record from the three raingages near Condon landslide, Central Coast Range, 10/89-3/90.

for the Condon raingage, to complete the on-site API graph for comparison purposes.

With six months of data available for all three gages (10/89 to 3/90), two linear regression analyses were performed for: (1) Condon and Mapleton; (2) Condon and Thompson Creek (Figs. 16 & 17).

The regression analysis on Condon and Mapleton shows the two raingages matched closely during the period of record in terms of magnitude, even though the total precipitation of the Mapleton gage was only 93% that of the Condon gage. R^2 in this analysis is 0.85. The relationship between the Condon and Thompson Creek gages is also affirmative, with the R^2 being 0.80.

Definite correlation among the three raingages could be assumed from the regression analyses, as all three gages responded to winter precipitation events in similar patterns and timings, despite of differences in magnitudes of water they received. This permits the extrapolation of the Mapleton record to the missing record period of the Condon raingage (from 12/5/89 to 1/9/90). However, utilizing data from the three raingages at different times is not recommended, as the considerable variation of the magnitude among the raingages may interfere with the quantitative

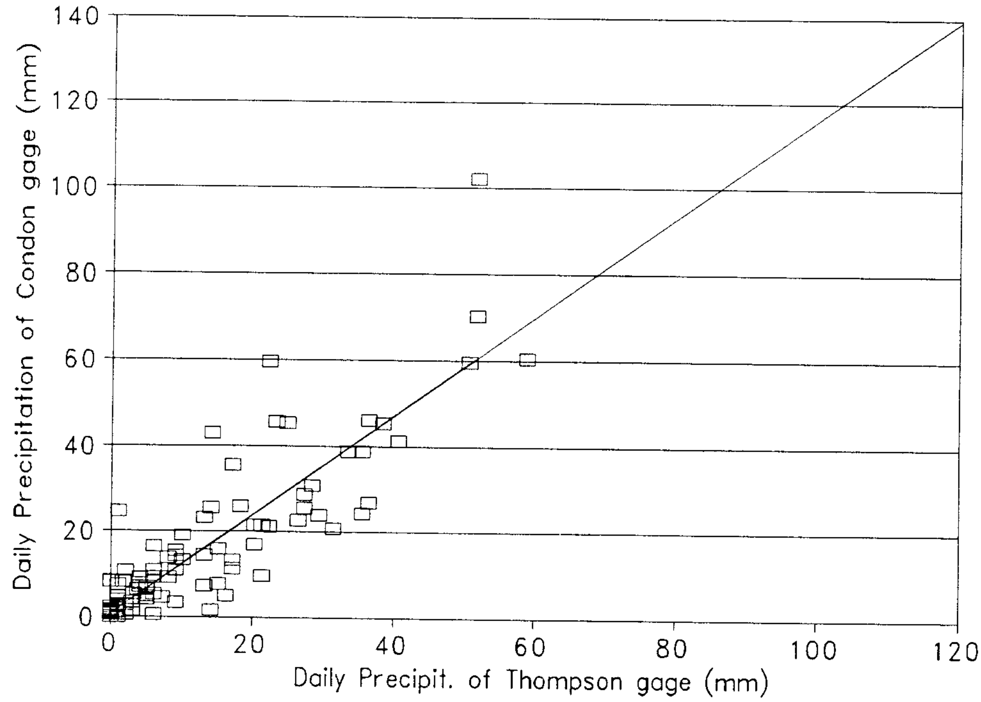

Figure 16 Linear regression analysis between Condon and Mapleton raingages. Periods of data are from 10/1/89 to 12/5/90, and 1/11/90 to 3/31/90. R^2 is 0.85. The equation of the resultant linear regression line is : $Y = 1.06 + 0.98*(X)$.

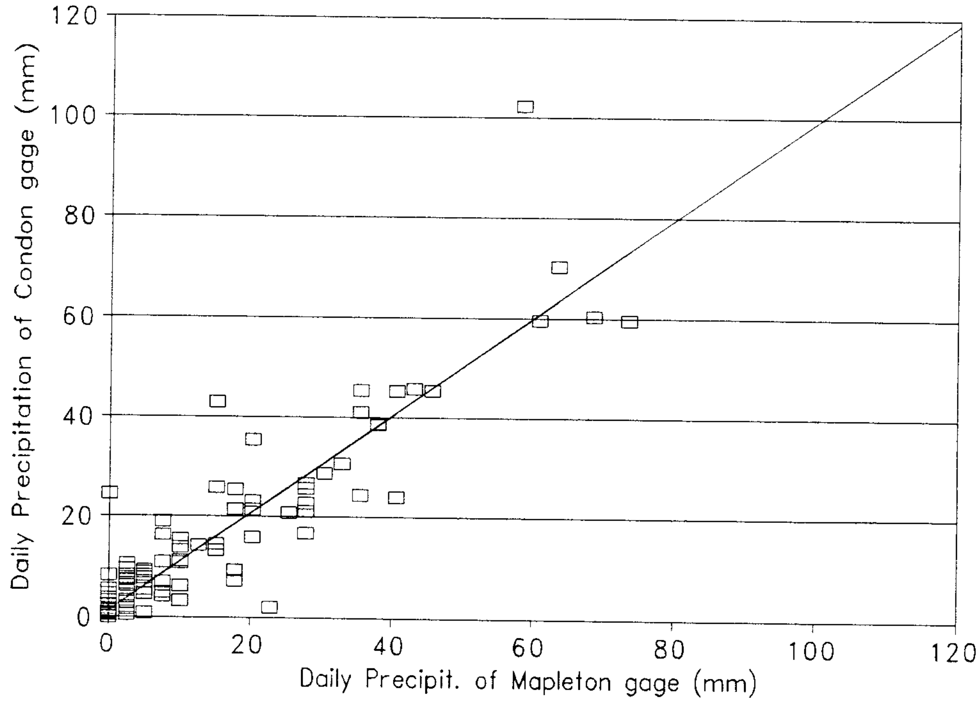

Figure 17 Linear regression analysis between Condon and Thompson Creek raingages. Periods of data are from 10/1/89 to 12/5/90, and 1/11/90 to 3/31/90. R^2 is 0.80. The equation of the resultant linear regression line is : $Y = 0.66 + 1.16*(X)$.

assessment of the threshold of the API values connected to movement events. Therefore, data from the Mapleton raingage alone are used in the API investigation.

Another linear regression analysis was run to develop a mathematical relationship between the precipitation data from the Mapleton and Thompson Creek raingages. The result is used to extrapolate the Thompson Creek record to the missing periods of the Mapleton record, to complete the precipitation data base of the Mapleton raingage for the Antecedent Precipitation Index investigation. Two periods of data with a total of nine months records were used for the analyses: October to December, 1988, and October to March, 1989-1990 (figure 18). The regression equation is:

$$Y = 0.738 + 1.040\ X$$

where X = Thompson Creek value

Y = Extrapolated Mapleton value

Generally, the timing and magnitude of precipitation events correlated well between the two raingages during the two data periods, with an $R^2 = 0.84$; and the equation was employed to estimate the amounts of precipitation received by the Mapleton raingage during its five periods of missing records from January to March, 1989.

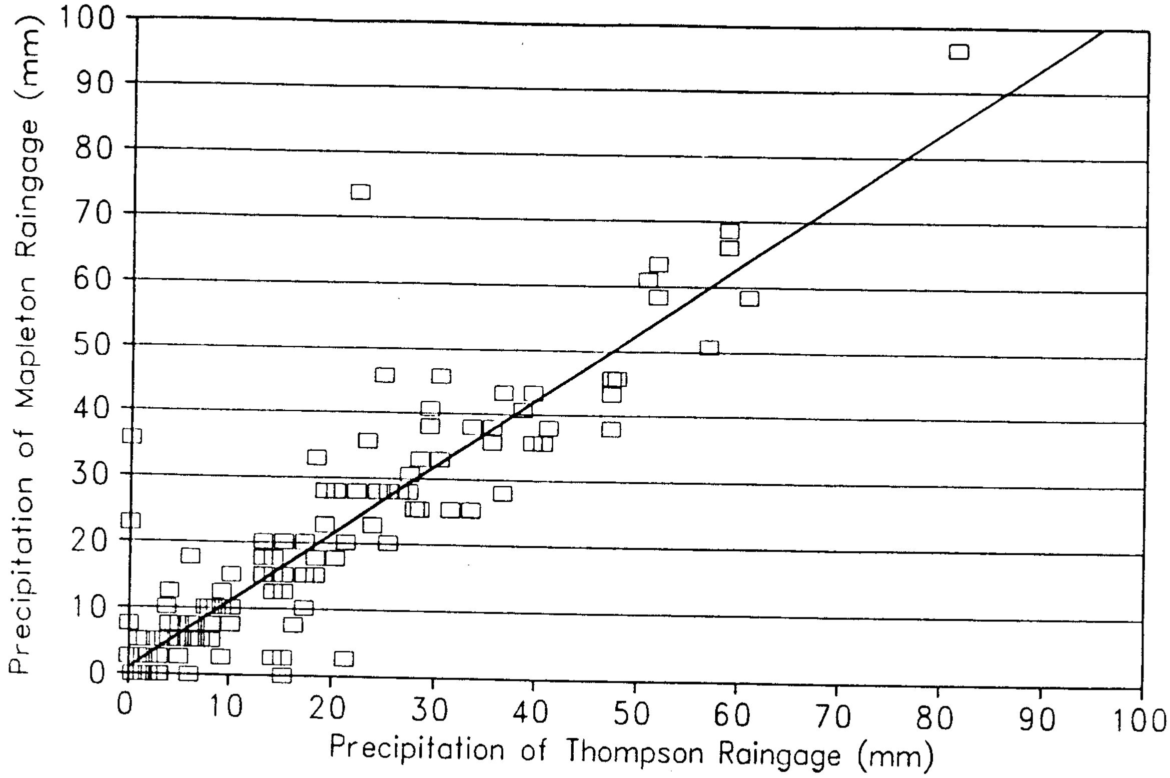

Figure 18 Linear regression analysis between Thompson Creek and Mapleton raingages. Periods of data are 10 to 12/88, and 10/89 to 3/90.

Groundwater level history of water year 1990

Condon on-site precipitation and piezometer records from 8/12/89 through 4/12/90 are summarized in Figure 19. The piezometer level is shown as the depth of the groundwater above the piezometer transducer, which is 5.11 meters below surface, situated roughly near the presumed failure surface. The amplitudes and profiles of the groundwater level graph illustrates how precipitation events shaped the landslide's groundwater flow. Approximately four meters below ground surface (one meter above piezometer transducer) appears to be the base of the groundwater level at the Condon Landslide, as groundwater level never dropped below this depth throughout the summer months of 1989.

During early September through middle October, the end of dry season, the groundwater level dropped slowly and consistently until the first major storm of water year 1990 (10/21/89) recharged the landslide (Fig. 19). This downward trend of groundwater level may be explained by drainage of groundwater to surface streams. Loss due to evapo-transpiration may not be significant due to clearcut and the fact that the groundwater level was well below root zone. Dry season rainfall generally induced little direct recharge of the groundwater level, as most water would be intercepted by the overlying vegetation, litter layer, and unsaturated soil zone, and later evaporated. This effect can be seen in

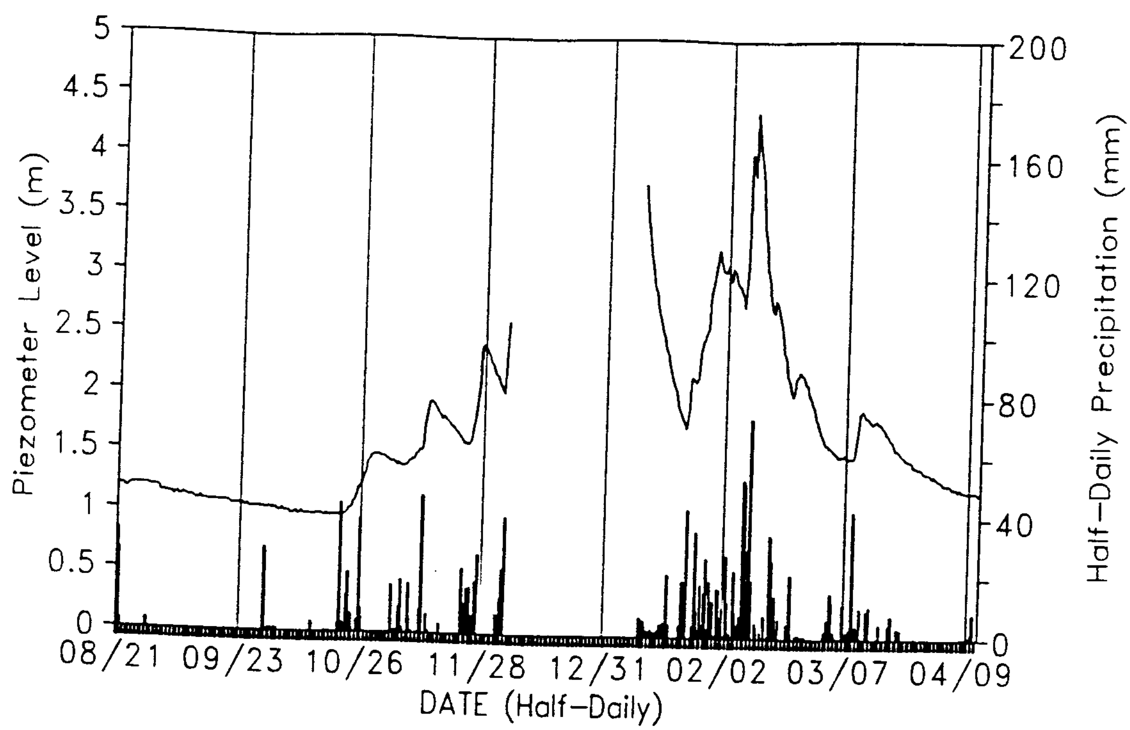

Figure 19 On-site precipitation record and groundwater level history of the Condon landslide piezometer site. The scale is half-daily, and the record stretches from 8/21/89 to 4/12/90, except between 12/5/89 to 1/10/90, when the data was lost. Note that the top of the plot is the ground surface.

late September 1989, when a small storm caused no modification of groundwater level (Figure 19).

During the wet season, individual storms with diverse magnitudes caused sharp fluctuations in groundwater levels. At the beginning of the 1990 wet season, discrete storms separated by three- to six-day periods of no precipitation provided smooth upward and downward trends of the water level (Fig. 19). The groundwater level went up abruptly, generally from 6 to 48 hours after the onset of large storms, and started going down gradually 24 to 48 hours after the storms had peaked or completely ceased. In the later part of the wet season, the groundwater level responded profoundly to water inputs, especially to large storm events, such as those of early February, 1990, which drove the groundwater level to its highest position of the wet season - only 0.6 meter below surface. After that, the dry spell stretching from 3/22/90 through 4/12/90 sent the groundwater level down to almost the pre-wet-season level.

<u>Characteristics of downward trends of the groundwater level</u>

In the case of the Condon Landslide, the rate of groundwater level rise depends on the amount of water input by individual storms, but the rate of lowering of the level after storm is a function of the height of the water level. That is, the higher the groundwater level elevation, the faster the rate of water level decline (Fig. 20). The linear regression of groundwater level to fall rate gave R^2 of 0.88. The roughly linear relationship between groundwater level and its falling rate indicates that the rate of groundwater flow depends mainly on hydraulic pressure, and the subsurface flow system is relatively simple.

The groundwater response time at the Condon Landslide, with lag times ranging from 24 to 48 hours, is slow compared to stream discharges observed at the same region (Fedora, 1987). The fast reaction of the groundwater level to precipitation events is coupled with a quick decrease of depth after the storms events ended. This indicates that the subsurface flow within the Condon Landslide is relatively quick, as compared to other landslides (Iverson & Major, 1987, Keefer & Johnson, 1983). This is due to the sandy landslide material, and extensive macropores in the fractured sandstone of the Condon slide.

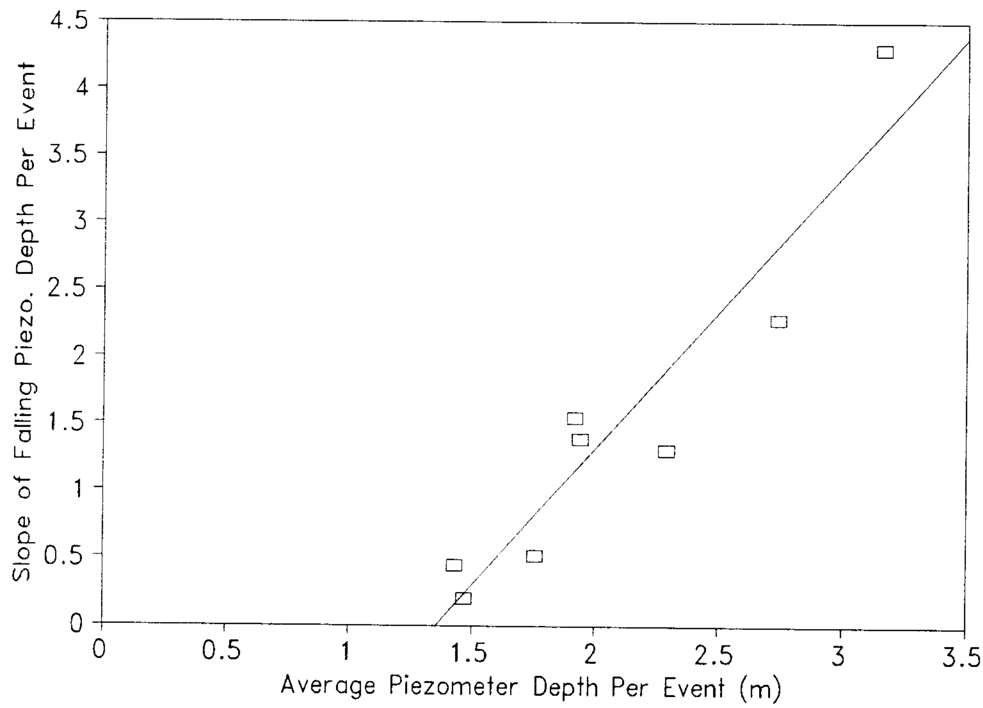

Figure 20 Slope of falling groundwater level with respect to average groundwater level for individual falling events. Events took place between 10/1/89 to 12/15/89, and 1/11/90 to 3/31/90.

VIII. Movement Characteristics of Condon Landslide

Introduction

The Condon Landslide displays a unique, yet consistent movement pattern. Movement events of the Condon Landslide usually occur in 1- to 5-day periods, and involve 10 to more than 100 mm of movement (fig. 21). All movement events took place between October and April. Although long periods of movement with small daily advances have been observed at the Condon Landslide, they were rare. Over the years, the Condon Landslide has been behaving in a predictable manner, with three major categories of movement observed: (1) small, (2) large-slow, and (3) large-fast. The following is the classification of movement events logged from 1984 to 1990. Effects on movement pattern by clearcutting at the landslide site in 1987 will be addressed in chapter 10.

The Small Movement Events

These events are short-lived, usually from one to three days, but could go on for as long as five days (Table 2). The total movement per event rarely exceeds 2 mm, and in most cases only 1 to 2 mm of movement were recorded. Rates of this movement group ranged from 0.3 mm/day to 2 mm/day. The number and average movement rates of these events are:

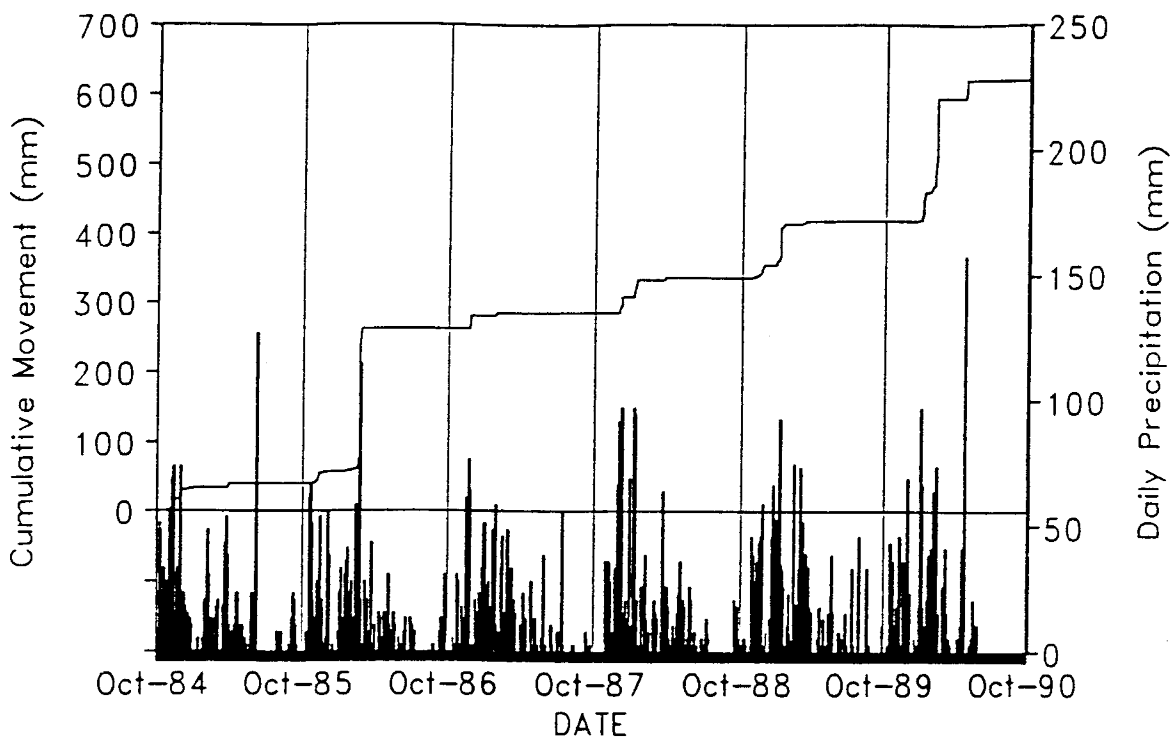

Figure 21 Cumulative movement and daily precipitation records of the Condon Landslide, water year 1984-1990. Precipitation record is from the Mapleton raingage, and movement record is based on data from the Condon extensiometer and stake array no. 1.

Date	Duration	Total Movement (mm)	Average Movement (mm/d)	Beginning Rate (mm/d)	Ending rate (mm/d)	Type
1985						
10/26	1	0.6	0.6			sm
10/29	1	1.0	1.0			sm
11/3-11/4	2	2.0	1.0			sm
11/10-11/14	5	12.3	2.5	5.8	1.7	Slow
11/16	1	0.7	0.7			sm
11/20	1	0.7	0.7			sm
11/25	1	0.6	0.6			sm
11/27-12/4	8	12.2	1.5	2.7	0.5	Slow
12/10	1	0.6	0.6			sm
12/14-12/15	2	1.2	0.6			sm
12/17-12/18	2	0.8	0.4			sm
12/30-1/1	3	0.9	0.3			sm
3/20-4/2	14	6.4	0.5			sm
1986						
10/28	1	2.0	2.0			sm
11/4	1	0.5	0.5			sm
11/7-11/11	5	12.9	2.6	2.9	1.0	Slow
11/22	1	1.0	1.0			sm
1/19-1/21	3	3.0	1.0			sm
2/7	1	1.0	1.0			sm
2/17-2/19	3	14.0	4.7	5.4	2.5	Fast
2/22-2/26	5	186.6	37.3	68.0	1.1	Fast
1987						
11/27-11/29	3	18.2	6.1	8.4	1.8	Fast
1/28-1/31	4	4.0	1.0			sm
(Landslide partly logged)						
1988						
12/3-12/7	5	9.3	1.9	4.0	1.6	Slow
12/9-12/11	3	14.0	4.7	8.8	1.9	Fast
1/10-1/13	4	13.9	3.5	7.5	0.3	Fast
1/15-1/16	2	10.0	5.0	6.6	1.8	Fast
3/25-3/27	3	5.0	1.7			sm
1989						
11/1-11/3	3	1.2	0.4			sm
11/5-11/7	3	0.9	0.3			sm
11/10-11/11	2	0.6	0.3			sm
11/17-11/19	3	2.2	0.7			sm
11/22-11/30	9	13.4	1.5	3.4	1.5	Slow
12/30-1/1	3	5.3	1.8			sm
1/8-1/12	5	49.3	9.9	33.0	1.0	Fast
1/14-1/18	5	4.3	0.9			sm
2/17-2/18	2	1.1	0.6			sm
3/6-3/7	2	0.8	0.4			sm
3/14-3/15	2	1.8	0.9			sm
3/27-3/28	2	1.2	0.6			sm
1990						
12/29	1	0.6	0.6			sm
1/6-1/12	7	39.6	5.7	7.0	3.5	Fast
1/26-2/1	7	7.5	1.1	1.0	1.9	Slow
2/4-2/6	3	1.2	0.4			sm
2/8-2/12	5	124.8	25.0	24.0	1.2	Fast
2/14-2/15	2	1.5	0.8			sm
4/27-4/29	3	26.0	8.7	20.3	3.0	Fast

Table 2 Individual movement events from water year 1985 to 1990, only large movement events are shown with detailed statistical data. (sm=small, Slow=large-slow-moving, Fast=large-fast-moving)

1985-87 (Pre-logging)	17	Average rate: 0.72mm/day ($8.3x10^{-9}$m/s)
1988-90 (Post-logging)	14	Average rate: 0.66mm/day ($7.6x10^{-9}$m/s)

These movement events are near the lower limit of detection. In this study events lasting for only one day with total movement less than 0.5 mm were included in the summation of annual movement, but were not counted as events here. Overall five events fall into this category, and account for only 0.3% (1.9 mm) of the total movement. In fact, some of the extremely small events might have actually been produced by measurement error or malfunctioning of the extensiometer (e.g. spooling of the steel cable). These small movement events (31 + 5 very small events) were distributed throughout the wet season. There were no apparent patterns of when and how they occurred, although they always happened on days when precipitation was registered.

In summary, this group of small events amounted to 57.2 mm of movement in six years of record, and represented only 9% of the total movement. In terms of the movement classification scheme developed by Pierson and Costa (1987), these events fall into the "mass creeping" category.

The Large Movement Events

This group can be divided into two categories: Long-persistent-slow-type; and brief-steep-step-fast type. These large movement events took place 16 times from 10/84 to 6/90. A total 56.4 cm of movement were recorded, and this contributed 90.4 % of the total amount of movement during the measurement period. Large movement events characterize the movement pattern of Condon and Wilhelm landslides.

Large Events that are persistent and slow

Three events of this type were recorded from 1985 to 1987, and three more were recorded from 1988 to 1990 (Table 2). Generally, these movement events had irregular movement rates throughout the period of movement, with the initial daily rate similar to the ending rate. Fig. 22 is a copy of the original daily records from the extensiometer, showing the pattern of typical long-persistent movement events. The beginning movement rates of these events ranged from 0.5 to 4.0 mm/day (<0.2mm/hour), and their ending rates were in the same range.

The maximum daily movement rate in these cases was between 3 to 5 mm, and their average movement rate was 1.8 mm/day (2.1×10^{-8} m/s). These events were also associated with periods of persistent precipitation with moderate intensity, around 50 mm in 6 hours (1.6-1.8 years return period).

Large Events that are brief and fast-moving

Ten fast-moving type events were recorded from 1984 to 1990. Three occurred before logging, and seven took place after clear-cutting in 1987. The duration of this kind of movement event ranged from two to seven days, with an average of four days. The total movement of these events varied from 10 mm to 187 mm (Table 2), and the average movement rate computed from all the recorded events was 13 mm/day, or 1.5×10^{-7} m/s.

Figure 23 shows a typical fast-moving-type event at Condon Landslide with rapid initial movement and slow ending rate, with smooth movement acceleration and deceleration within a short period of time, usually lasting from 6 hours to 2 days. An entire event could last from 1 to 5 days. The movement events always began abruptly, coinciding with large rainstorms, accelerated rapidly during the first few hours, advanced steadily for 12 to 24 hours, and then gradually slowed down, and came to a halt over 1 to 2 days.

The range of maximum daily rates is from 5 mm to over 125 mm. Since the landslide moves rapidly in the initial acceleration phase, the day with the fastest moving rate may

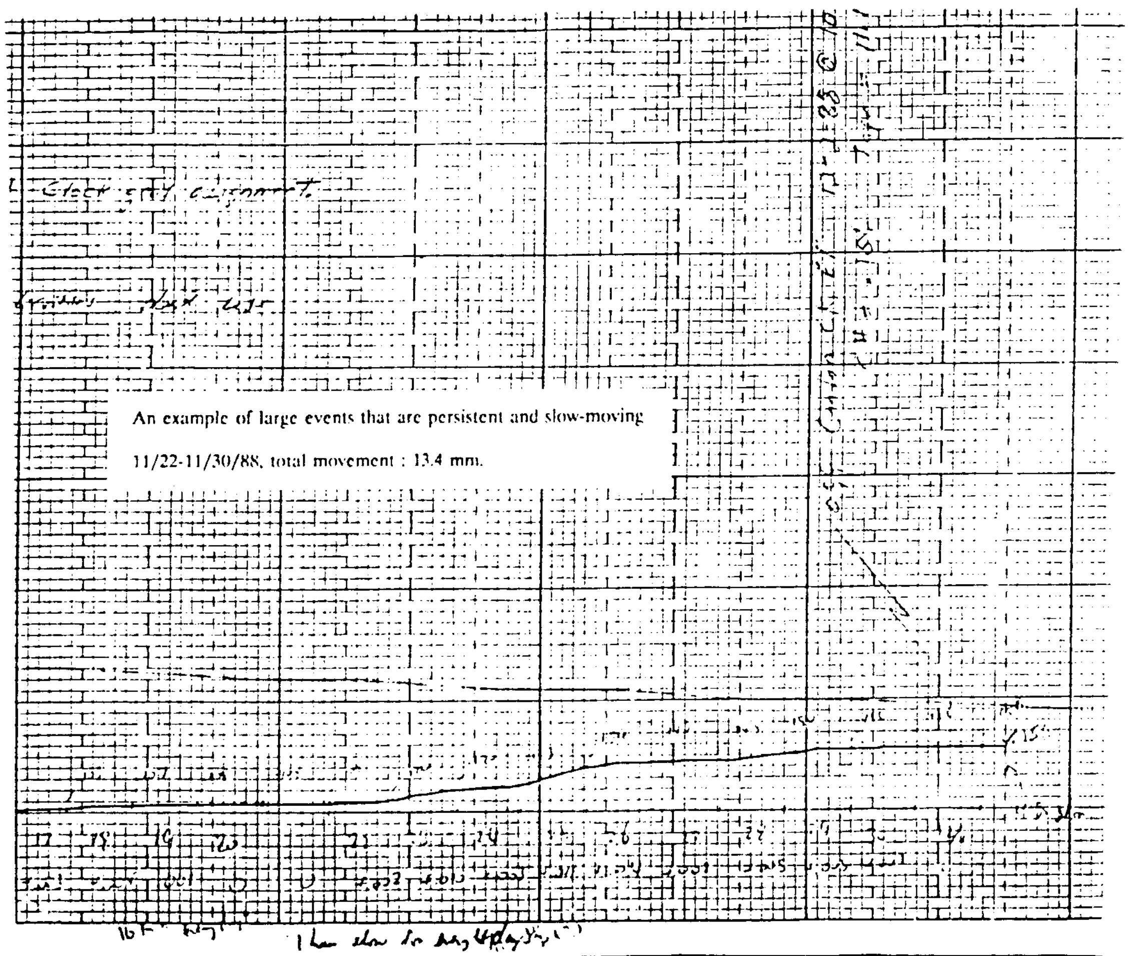

Figure 22 Copy of field data showing a large movement event that is persistent and slow-moving. The length of this event was 9 days, with a total movement of 13.4 mm. The beginning and ending movement rates are similar in this case. Horizontal axis is time (6 gridline-interval represent a day), and vertical axis is cumulative movement (1 gridline interval represents approximately 3 mm).

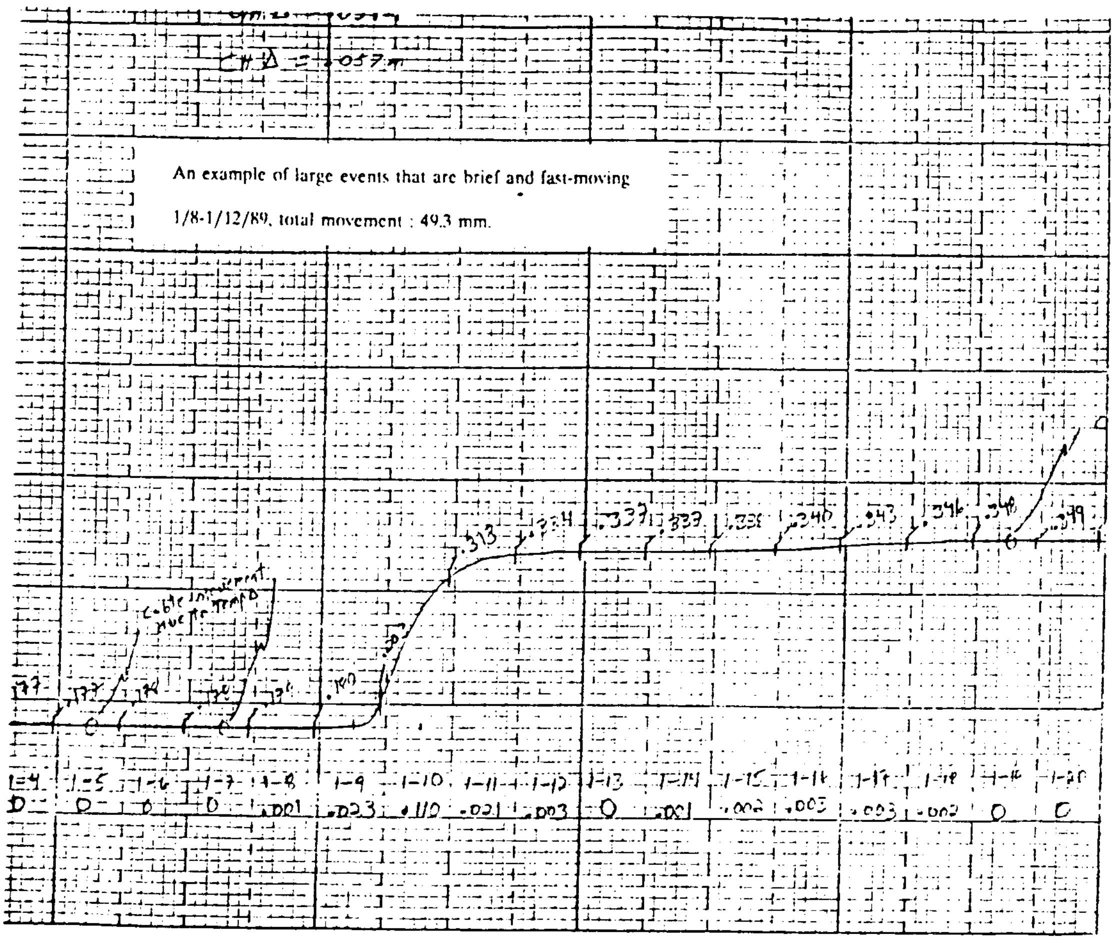

Figure 23 Copy of field data showing a large movement event that is brief and fast-moving. The length of this event was 5 days, with a total movement of 49.3 m. Note the smooth S-shape of the movement-rate plot, with abrupt and fast acceleration, and persistent deceleration.

contribute more than 50% of the total movement in the event. The normal beginning and ending rates of this type of movement are :

Beginning : 5.0 to over 30 mm/day, 0.2 to 4 mm/hour

ending : 1.0 to 3.0 mm/day, 0.05 to 0.15 mm/hour

For the example given in Fig. 23, the maximum daily movement amounted to 69% of the total movement.

These large movement events may produce significant morphological changes in the landslide. New tension cracks, for example, formed at the headscarp area after a major movement event (total 49.3 mm) in January 1989 (Fig. 10). Amount of input of water by winter storm events is the obvious driving condition.

<u>Relationship between movement and precipitation events</u>

All the fast-moving events with significant amounts of movement conformed well with high precipitation periods (fig. 21), which shows that movement episodes at Condon Landslide are driven by storm events. Some of them had a few days of intense precipitation before movement started, with a time lapse of 1 to 2 days between the beginning of major water inputs and the commencement of movement. H. Moriwaki (1991, person. comm.) has demonstrated the significance of a two-day lag time between storm water input and amount of movement. All large, fast-moving events stopped within one or two days after the corresponding rainstorm had ended. Most large storm events from 1984 to 1990 corresponded with movement episodes, but the relationship is not perfect (Figure 21); some large storm events did not produce movements, while several small storms triggered large movement events. In the following section, annual, monthly, as well as daily precipitation and movement records from 1984 to 1990 will be used to review the relationship between precipitation events and movement episodes at the Condon Landslide.

Annual total movement and annual wet-season (October to April) precipitation from water year 1982 to 1990 portray a subtle implication that annual movement increases with increasing annual precipitation (fig. 24). The correlation between monthly movement and precipitation is positive yet not consistent (Fig. 25), which contrasted with landslides in northern California (Keefer and Johnson 1983). For purposes of further analysis, 16 individual large movement and storm events from 1984 to 1990 water-year are used. No apparent association exists between total movement and total precipitation of individual events (Figure 26), except all large events bigger than 30 mm had total precipitation greater than 150 mm. On a daily basis precipitation and daily movement are not related (fig. 27). Above all, these two graphs (Figs. 26 & 27) fail to consider an important

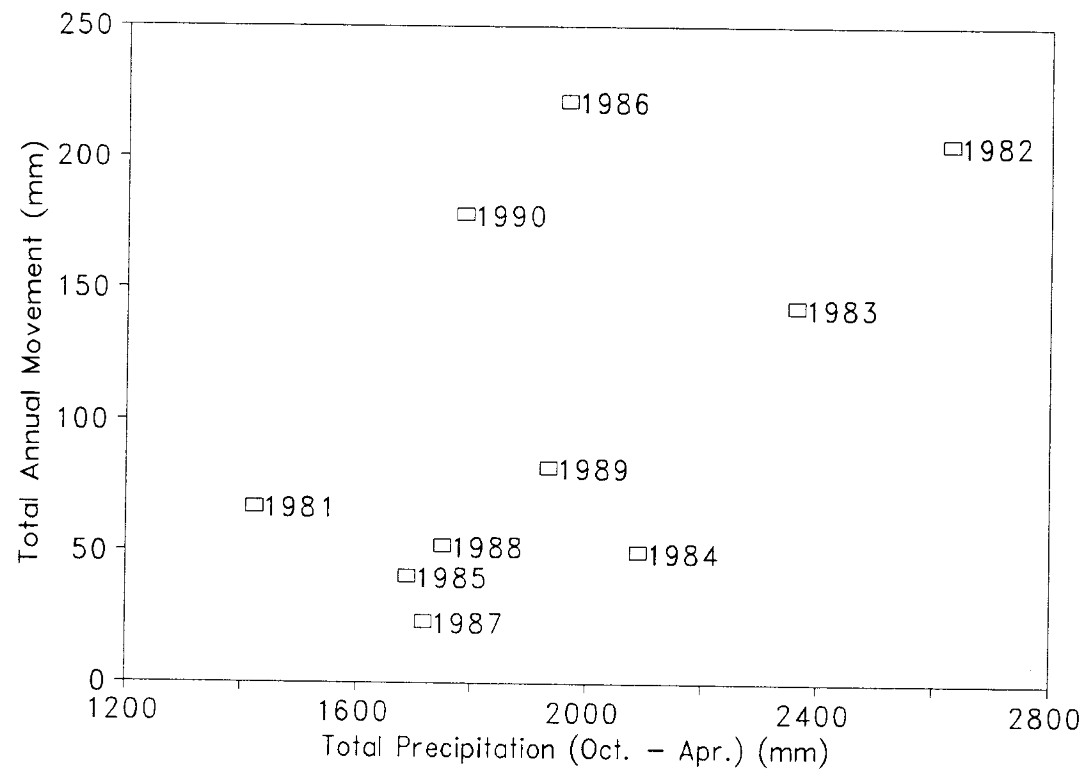

Fig. 24 Relationship between total movement and total precipitation from October to April at the Condon Landslide. Record period is from water years 1981 to 1990.

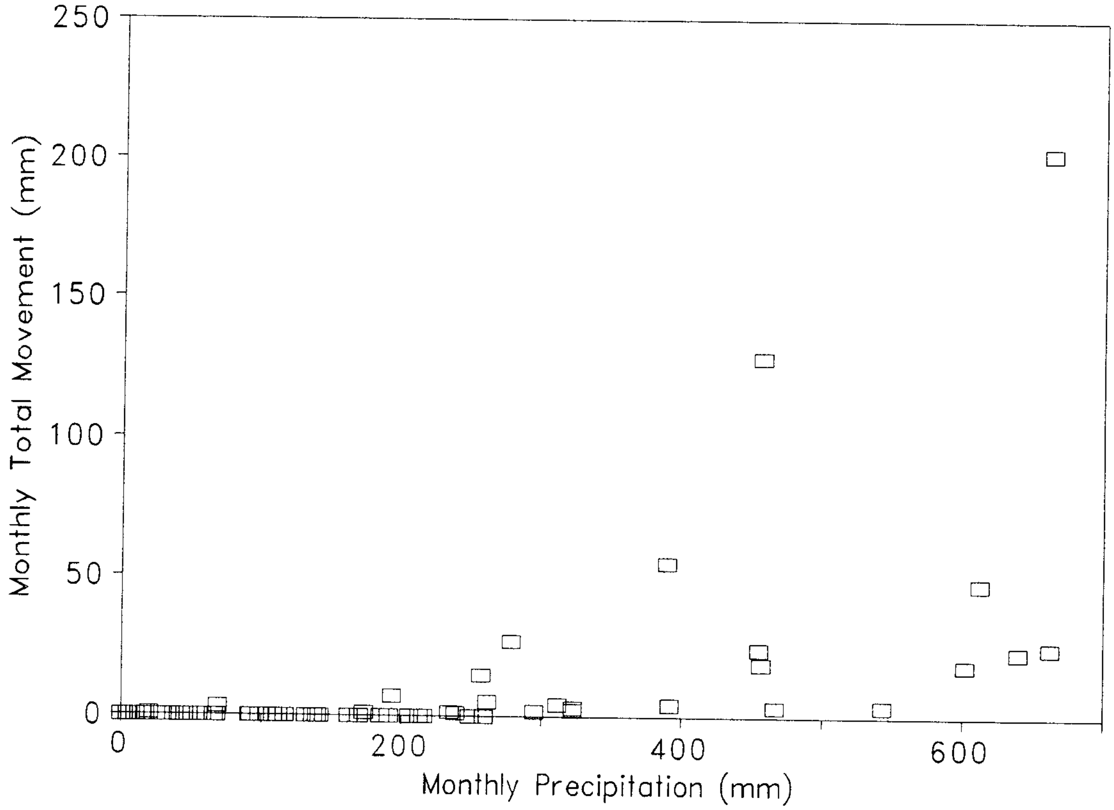

Figure 25 Relationship between monthly movement and precipitation of the Condon Landslide, from 10/85 to 2/87, and 10/87 to 6/90.

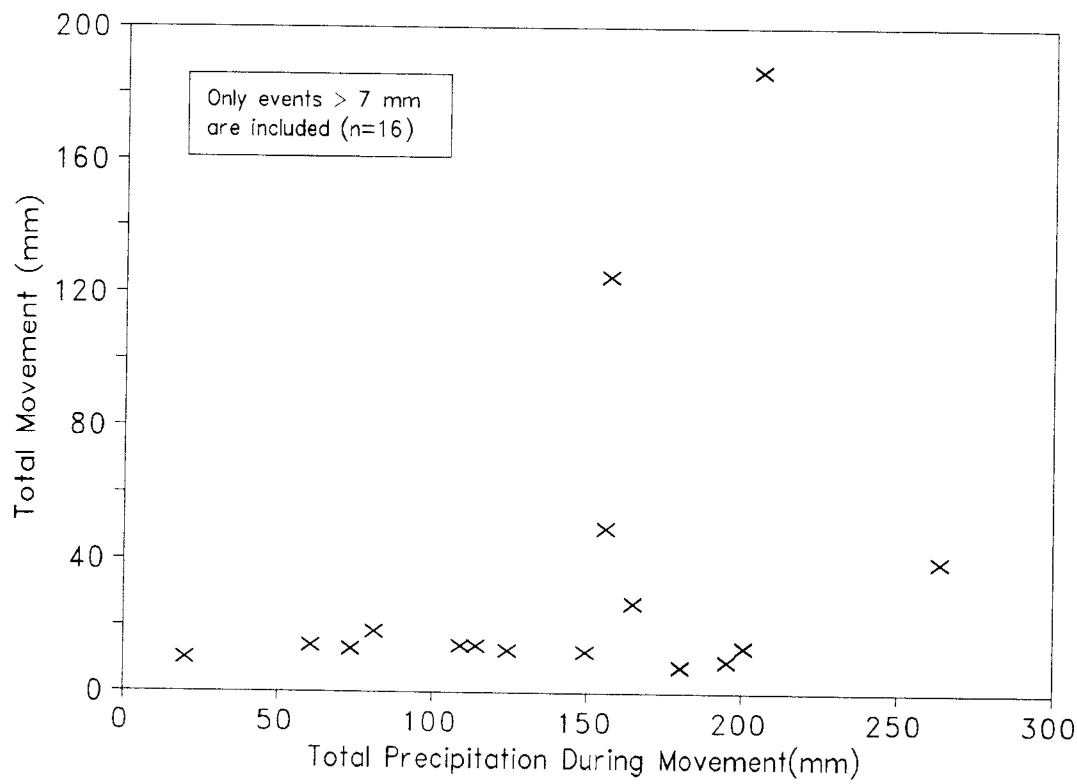

Figure 26 Relationship between total movement and precipitation for the 16 large movement events at Condon Landslide, water years 1985 to 1990.

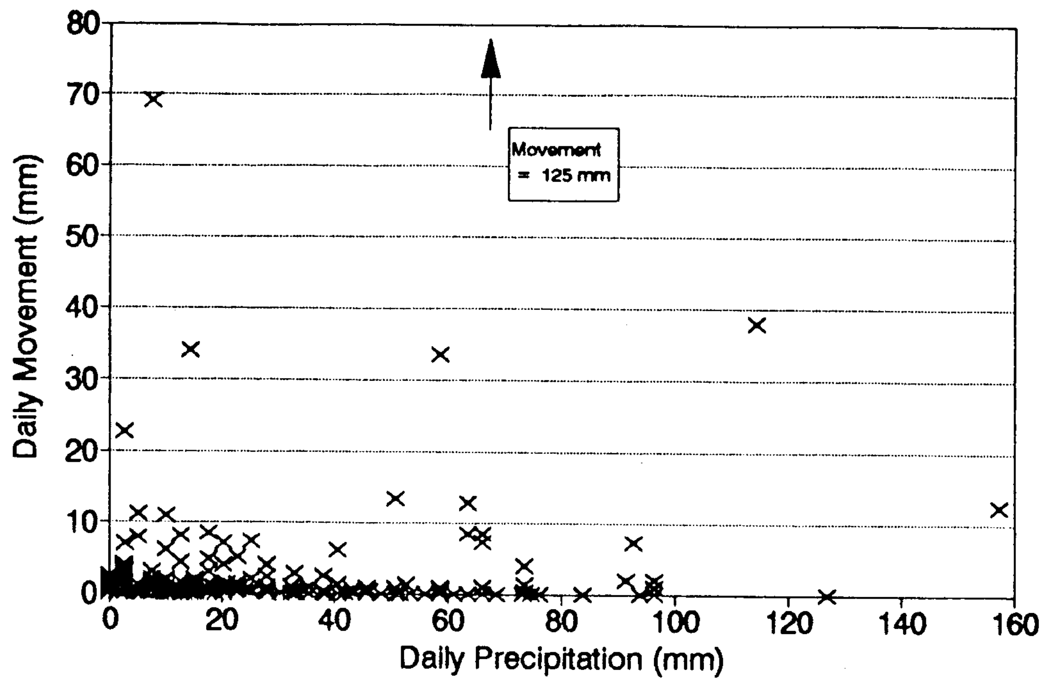

Figure 27 Relationship between daily movement and precipitation at Condon landslide, from 10/84 to 2/87, and 10/87 to 6/90.

element in the movement condition: high antecedent moisture retained in the landslide throughout the wet season.

The antecedent precipitation was also considered as precipitation for 72 hours before movement started, including precipitation received on the movement starting day. Only a weak positive relationship can be drawn between this variable and total movement per event (fig. 28).

Discussion

By using annual and monthly time scales to characterize the Condon Landslide movement pattern, the results do not perform well; because movement events at Condon Landslide are discrete episodes separated by long periods of no movement, not like landslides found in the western Cascades of Oregon and northern California, which move consistently throughout the wet seasons. However, figures 24 and 25 do show some correlation between amounts of precipitation and movement.

The quantitative analyses performed indicate two findings: (1) Precipitation is the major factor triggering movement at the Condon Landslide. This is especially true to most large movement events, which corresponded with major storms. (2) Although movement events at Condon Landslide are related to precipitation events, only a weak

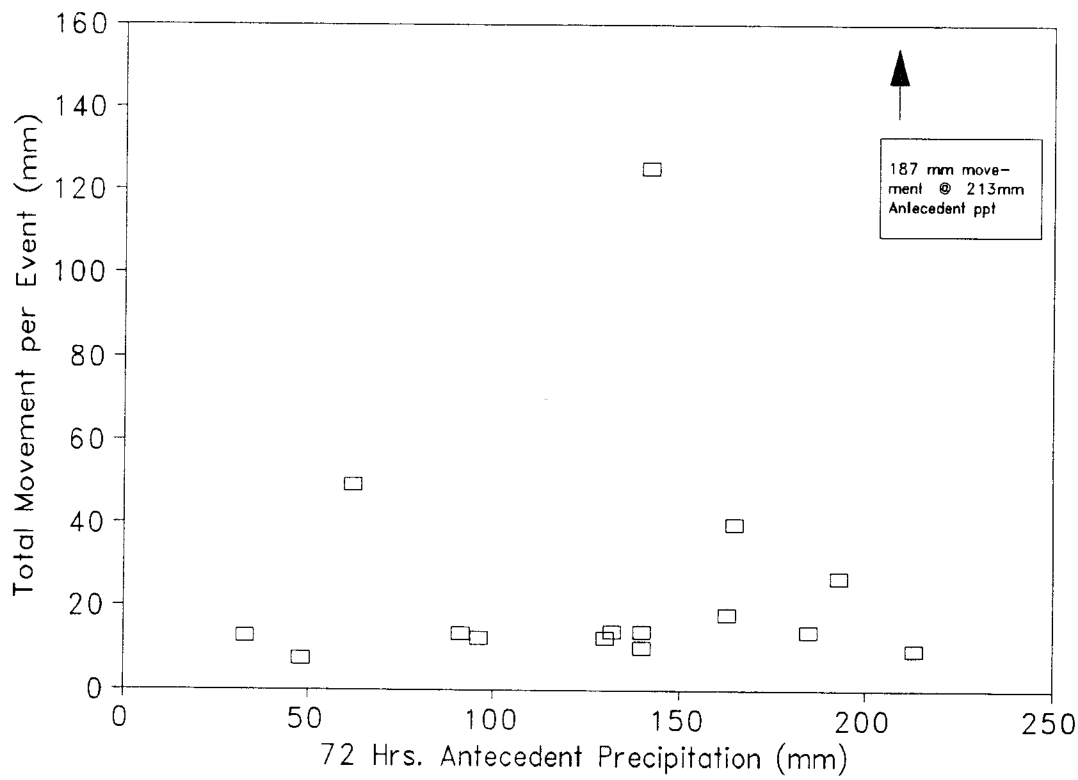

Figure 28 Relationship between total movement and 72-hour antecedent precipitation amounts for the 16 large movement events, Condon landslide, water years 1985-90.

relationship is observed between the two parameters quantitatively at a daily time scale, as well as using simple 72-hour antecedent rainfall condition. Later in this report, an Antecedent Precipitation Index (API) investigation is used in an attempt to derive a more precise relationship between precipitation and movement events.

Groundwater level changes and movement events

Changes of groundwater level within a landslide mass control the magnitude of pore water pressure. Therefore, groundwater strongly influences the effective stress state in a landslide mass, which can trigger hillslope failure (Iverson and Major, 1987). To understand the relationship between the groundwater level variations and movement events at the Condon Landslide, piezometer and movement data from October, 1989, to March, 1990, are used to determine if an empirical relationship exists between the two parameters, and to define probable critical (threshold) groundwater level associated with initiation of movement.

Five movement events spanning several days each between January and March, 1990, were used to formulate the relation between groundwater level and movement. A total of twenty days of movement had displacement rates ranging from 0.3 to 69.2 mm/day, while groundwater level varied from 2.47 to 4.50 m (approximately 2.6 and 0.5 m below surface) (Fig.

29). Most daily movement rates are less than 4 mm/day. Figure 29 depicts a positive correlation between groundwater level and daily movement rates.

Groundwater level determines the effective stress acting on the landslide, therefore, it is an important variable regulating landslide movement (Iverson, 1986). Figure 30 shows the daily groundwater level and movement records of the Condon landslide, and Figure 31 displays the frequencies of occurrence of movement with respect to groundwater level from the daily-record period. Displacements only occurred at water level above 2.0 meters, and movement happened during all days with groundwater level above 3.5 meters. Although three days with groundwater level above 3.0 meters did not have movement, they were days immediately after movement events in February, 1990.

Discussion

With only one wet-season's data available, it is difficult to draw reliable conclusions concerning the groundwater level threshold for movement events. However, when groundwater level exceeded 2.50 meters during periods of intense storms, the likelihood of displacement increased greatly. A groundwater level of 2.5 meters (2.5 m below ground surface) can be inferred as the critical water level (fig. 30), based on the following observations: (1) two days

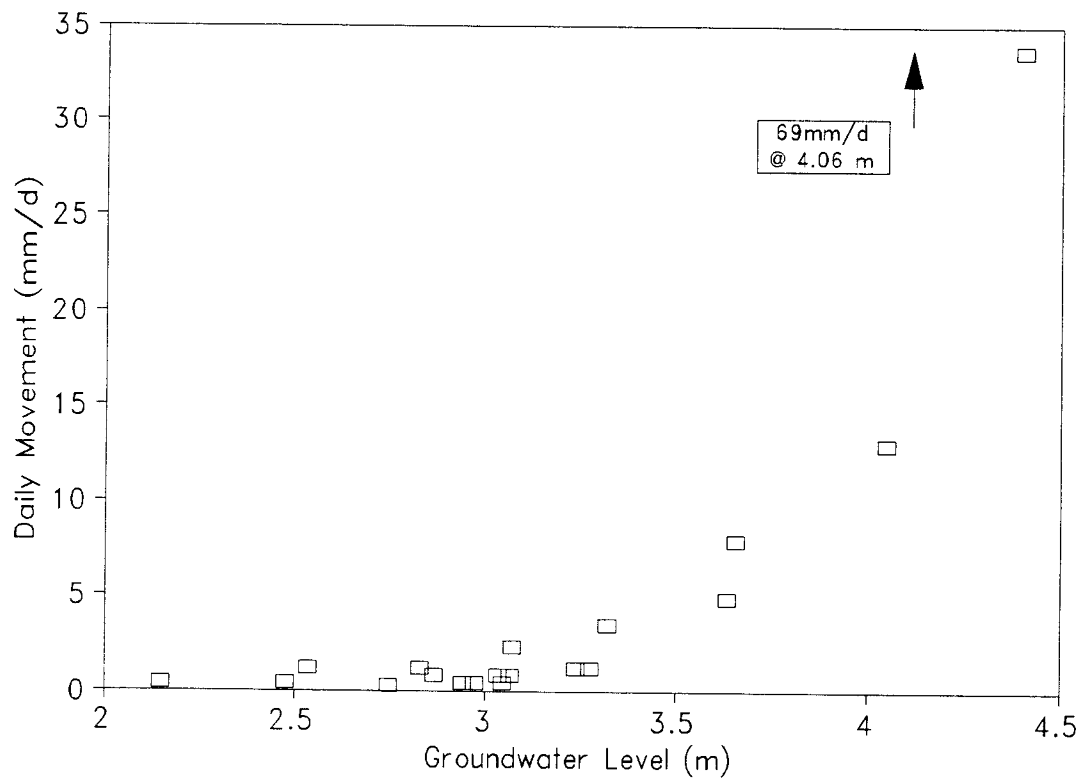

Figure 29 Correlation between daily movement rate and averaged daily groundwater level at Condon Landslide, 10/1-12/5/89 and 1/10-3/31/90. A non-linear relationship exists between the two parameters.

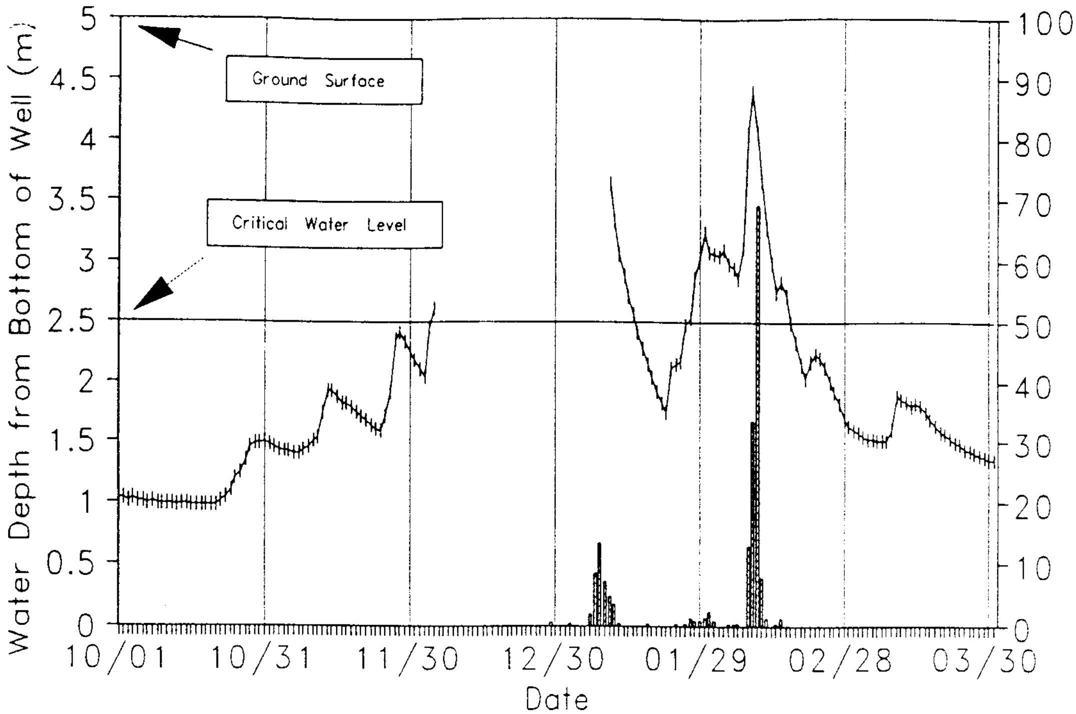

Figure 30 Daily groundwater level and movement record of Condon landslide from 10/89 to 3/90, showing the deduced critical groundwater level (2.5 m), in which most movement occurred above this level.

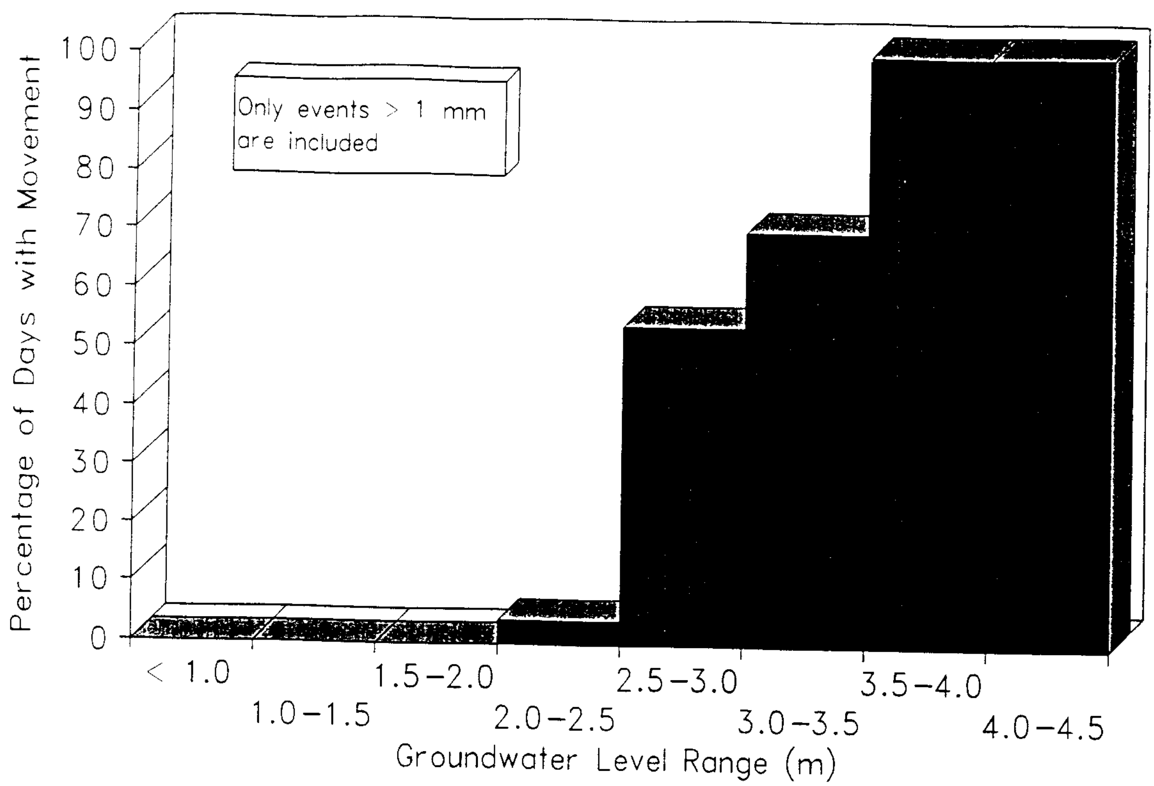

Figure 31 Frequency of movement at different groundwater levels, Condon Landslide, 10/89-3/90.

of movement with water levels at 2.14 meters were only one-day events with movement less than 0.5 mm. The magnitude of these two events is insignificant and might have been caused by instrument errors. Hence, 2.5 meters can be established as a reliable groundwater level for initiation of movement. (2) The major movement episodes between 1/26/90 to 2/15/90 started when the groundwater level reached 2.47 meters (fig. 30). Data were missing for the earlier episode which began at 1/6/90, so it could be assumed that movement commenced after the groundwater level reached 2.4 to 2.5 meters, based on analysis of the shape of the groundwater graph (fig. 30). The 2.5-meter groundwater threshold is drawn based only on empirical data; the lack of in-situ data of the mechanical properties of the landslide precludes calculation of a critical groundwater level in terms of the engineering behavior of the landslide.

Other questions concerning the relationship between variation in groundwater level and initiation of movement events remain to be answered. For example, since most movement events started when the groundwater level was rising, is there a correlation between the initial rate of movement and the rising rate of the water level?

IX. Antecedent Precipitation Index Analyses

Introduction

At Condon Landslide, movement events and precipitation occurrence are related, but based on comparison of precipitation and movement records, the relationship between them is not simple (see Chapter 8). In this study, an attempt has been made to probe the relationship between precipitation and movement events of the Condon Landslide, by using Antecedent Precipitation Index (API) as a surrogate for groundwater level variations. There are two basic objectives in this API analysis: (1) to correlate amounts of movement with magnitudes of API for large movement events; (2) to find if there is any relationship between the API values and the timing of movement, and to explore API for use in analysis of clearcut effects on timing of movement.

Generally, streamflow or groundwater flow occurring at any time is a function of magnitude and temporal distribution of preceding precipitation. A hydrologic system responding to precipitation would have a full "memory" of rain falling at that time, partial "memory" of rain that fell a short time before, and only very vague "memory" of rain that fell much earlier. Therefore, the influence of precipitation inputs to the system "decays" through time. This is the basic premise of the Antecedent

Precipitation Index (API) method (Fedora, 1987). The water output or loss from the system control the extent of the "memory" of that system. In the case of underground water content, this is controlled by numerous natural factors, including evapotranspiration rates, angle of the slope, soil properties, and underground flow regimes.

Previous Work

API has been used to model residual effect of preceding precipitation and soil water drainage on the magnitude of stream runoff or soil moisture. This method was utilized by Ziemer (1984) to examine the relationship between precipitation and hillslope deformation by creep and slow landslide processes, using borehole deformation and precipitation records in the Redwood Creek area of California. In regression analyses with a semi-annual time step, he concluded that earthflow movement increased in response to precipitation and the earthflow had an ability to "remember" precipitation conditions from the previous winter (regression factor derived to be 0.99). However, he observed no API threshold correlating with the beginning of movement events.

Using an API analysis, Istok and Boersma (1986) concluded that in climatic regions characterized by low-intensity rainfall, such as the Pacific Northwest, antecedent rainfall is more important in controlling the amount of runoff than is rainfall magnitude or intensity. An API analysis for simulating storm runoff events was done by Fedora (1987), who developed empirical equations based on API methodology, by using precipitation and watershed area as the only variables. Fedora demonstrated the usefulness of API models in simulating storm hydrographs, and estimating volumes of peak flows from small watersheds in central Oregon Coast Range. Nevertheless, the usefulness in predicting landslide motion has not been tested.

<u>Method of Data Analysis</u>

Values of API at time t are dependent upon precipitation occurring before that time. The API at time t can be expressed as :

$$API_t = API_{t-dt} * C + P_t$$

Where:

API_t = Antecedent Precipitation Index at time t (mm)

dt = Time interval of precipitation observations

C = Recession Coefficient (dimensionless)

P_t = Precipitation volume during the time interval before time t (mm)

In this study, a time interval of one day is used, because groundwater level in thick landslide colluvium responds more slowly to precipitation than stream runoff does, in which Fedora used an interval of two hours.

Recession Coefficient Factor

The recession coefficient (C) in the Antecedent Precipitation Index equation represents the decay factor of the water drainage effect of cumulative precipitation on runoff and groundwater level changes. It dictates how long the antecedent rainfall affects the magnitude of stream discharge or groundwater level (piezometer depth).

Fedora (1987) calculated recession coefficients for selected watersheds in the Oregon Coast Range using stream runoff data (p.49-50, and p.55-58). The values obtained range from 0.888 for a small watershed (9 km^2) to 0.949 for a relatively large (41 km^2) watershed (North Fork Siuslaw River), with a time interval of two hours. He found a positive correlation between the recession coefficient and watershed size, and he assumed that an extremely large watershed would have the C factor approaching 1.00.

The recession coefficient was determined by deriving the slope of the line formed by plotting the declining piezometer depth during periods of groundwater recession when there was very little (less than 10 mm/day) to no precipitation (fig. 32). Only the record from September, 1989 to March, 1990 was available for this derivation. For the groundwater condition at the Condon Landslide, the slope of the line is 0.87, that is, when no new precipitation was added to the soil, the groundwater level is 87 percent of the level of the previous day.

Assumptions

One complication in getting an accurate recession coefficient factor is the lack of understanding of the groundwater flow system within the Condon Landslide. The variation of the physical conditions of the groundwater system could greatly affect the linearity of the recession. Moreover, not enough climatic data are available to determine evapotranspiration condition at Condon Landslide, which is a major component of the recession coefficient. Despite these problems, a single-slope linear recession model was used in this study to simplify the simulation (Robert Beschta, 1989, person. comm.).

Selection of Precipitation Record

API analysis is based on precipitation data. Since the precipitation record at the Condon Landslide is available only from 9/89 through 4/90, API investigation for earlier

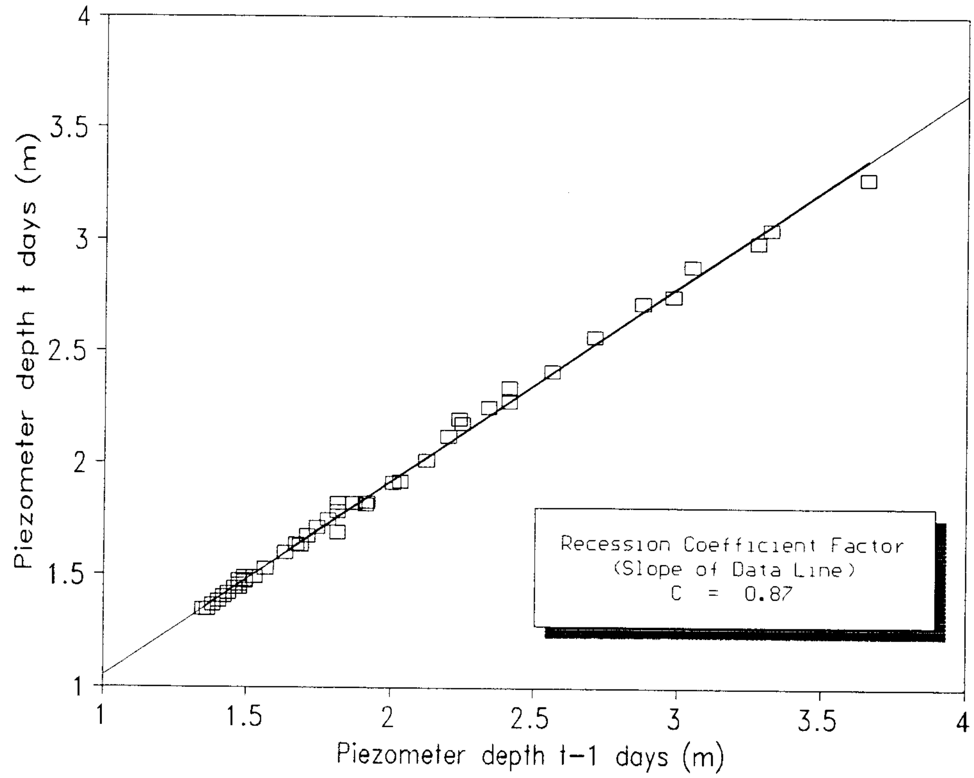

Fig. 32 Recession coefficient factor derived from the Condon Landslide piezometer. The time interval used is one day. The recession factor is equal to the slope of the regression line, which is 0.87.

periods must be based on nearby raingages. The Mapleton raingage provides a precipitation record throughout the investigation period, and is the natural choice for this study. However, the Mapleton raingage is 14 km from the Condon Landslide and is 12 m above sea level, while the landslide site is 160 to 280 m above sea level. Therefore, to test the suitability of the Mapleton record for this study, daily API with a recession factor of 0.87 was calculated from the on-site record, as well as records from Mapleton (p. 51) and Thompson Creek (p. 51) from 9/1/89 to 3/31/90 (fig. 33).

The API graphs of the three gages match well in terms of timing, although the magnitude of the storms differed among the three sites in most events (fig. 33). All three traces are virtually identical in certain short periods (e.g. December 1989, and mid-Jan. 1990), and most of their peaks coincide very well. All three gages received similar precipitation amounts in those periods, even during periods of persistent back-to-back storms stretching for more than two weeks (e.g. early January and early February, 1990).

The magnitudes of API among the three raingages vary throughout the record period. The variances among these gages were mostly due to differences of precipitation received by different gages rather than differences of

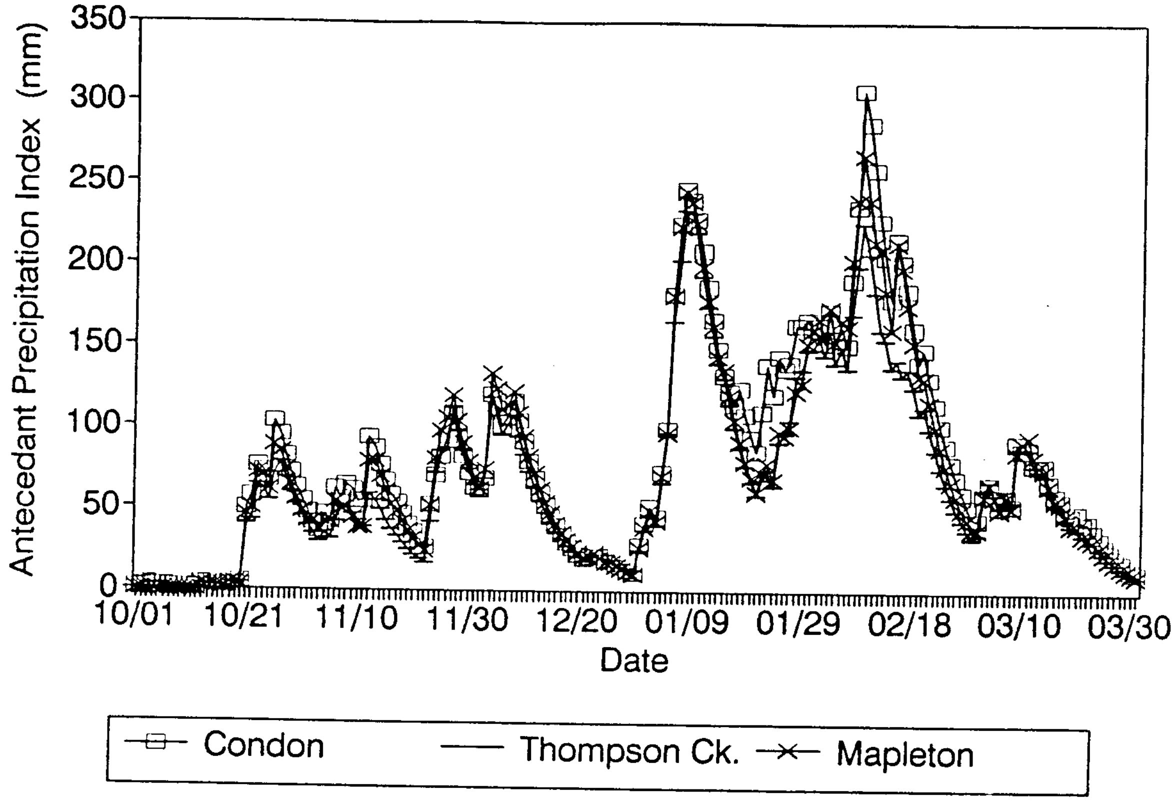

Figure 33 Calculated daily API for 3 recording raingages near Condon landslide, 10/89 to 3/90.

number and timing of storm events. Nevertheless, differences among the three API graphs are not significant (fig. 33). The Thompson Creek raingage received the least precipitation among the three gages during the comparison period, and this is reflected in its low overall API values. Nonetheless, the shape of its graph follows closely the traces of the other two data sets, except during the intense and persistent storms from 2/6/90 to 2/18/90. During that series of storm events, the total API values of the Thompson Creek gage was 75% of the API values of the Condon gage, which has consistently the highest API values among the three gages throughout the wet season. The total API values of the Mapleton gage were only 6% less than the Condon values during the same time period.

Even though the three gages are located at different elevations, and are 8 to 14 km apart, their API graphs are strikingly similar. This may be explained in part that the timing of the major and minor storm events move rapidly across the area containing the three gages and spatial variation in precipitation within the storms is not great at the scale of the raingage networks. An additional consideration is that the antecedent parameter smoothed the precipitation record, because the decay rate of hydrologic effects of previous storms is slow enough that previous storm events may substantially influence the daily API values, even from large storms.

The major weakness of this analysis is the short period of precipitation record used in API calculation. Data from only one wet season are obviously limited for establishing a firm relationship among the data sets from the three raingages. Despite these limitations, I used the Mapleton record for API calculations from 1985 to 1990, while records from the Condon gage were employed for on-site hydrological analyses (e.g. groundwater level correlation).

Antecedent Precipitation Index and Movement Plots

The API record along with cumulative movement plot from 10/1/84 to 5/30/90 is shown in Figure 34. Half-yearly daily API and movement rate plots, which have higher resolution, can be found in Appendix A. Only half-yearly API was presented in detail because: (1) only one major movement episode has even been recorded from April 5 to September 30 (the 4/27-29/90 event); (2) the total summer precipitation at the central Oregon Coast Range is very low, less than 20% of total yearly precipitation in most year. Therefore, with a recession rate of 0.87, the API carried over from previous water year ranges only from 4 to 44 mm, and have little effect on total API at the beginning of the next movement season.

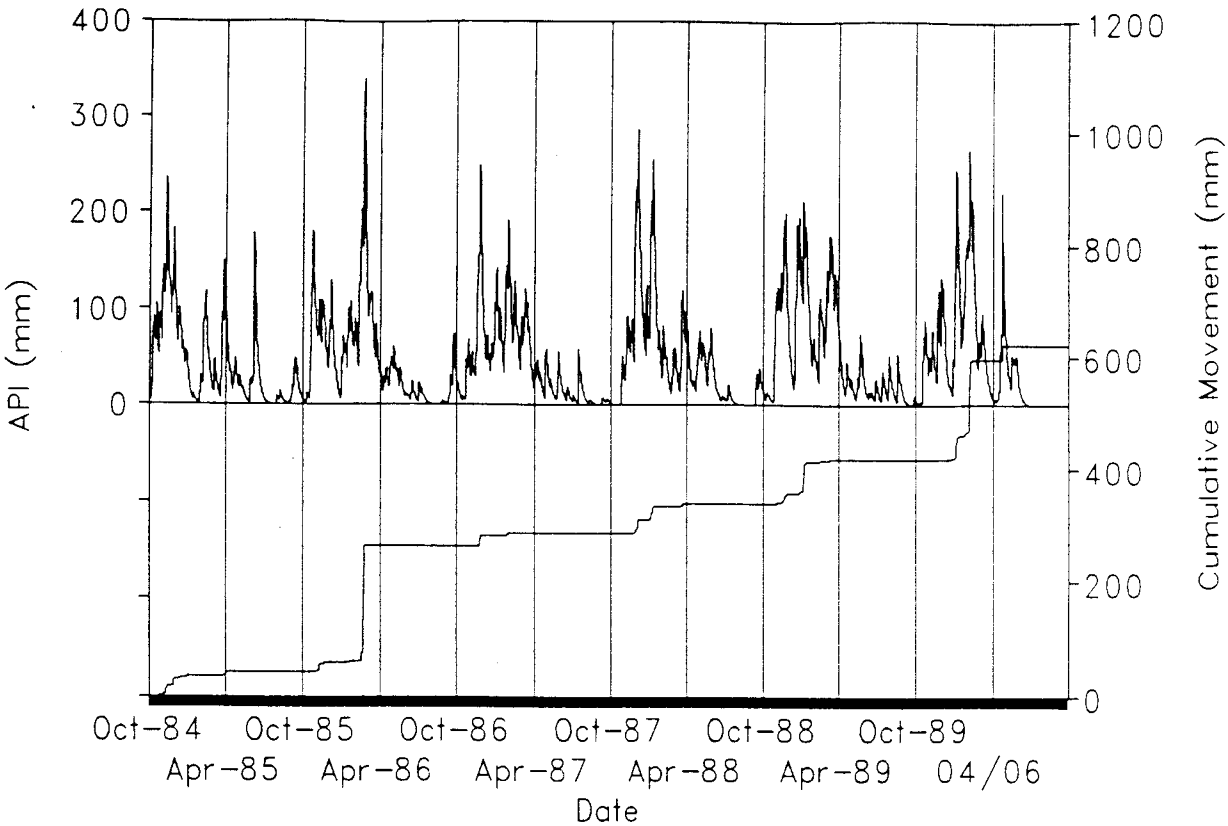

Figure 34 Cumulative movement and daily API plots of the Condon landslide, 10/84 to 6/90.

The six-year combined plot (fig. 34), and half-yearly API plots (Appendix A) reveal that most major movement events occurred with periods of high API: 13 of the 16 (81 %) major movement events had API of over 150 mm when movement began. API values at the cessation of movement are much more variable, ranging from 85 mm to 280 mm, which have values of 61 % to 176 % compared to their starting API.

Correlation of Antecedent Precipitation Index to Movement Events

The API analysis can be divided into two parts : (1) to study the relationship between the amounts of movement and magnitudes of API for large movement events; (2) to find if there is any relationship between the magnitude of API and the timing of movement, formulating an API threshold for movement initiation if possible.

Correlation of magnitude of movement and API

Monthly and daily API and movement rates were used to investigate the possible relationship between the magnitudes of API and movement events at Condon Landslide. The monthly and daily plots using API and movement amounts accounted for periods 10/84-2/87, and 10/87-5/90 (Figs. 35 & 36). These two plots demonstrate the positive relationship between amounts of movement and API, although the relationship is non-linear. Figure 35 shows that most months having movement had total API exceeding 2000 mm, while Figure 36 shows that the majority of days with movement had a daily API of more than 180 mm.

To further the analysis, 16 large events were used to test the observed API-movement relationship (Figs. 37 & 38). API values vary greatly for events less than 20 mm total or 5 mm per day. However, the major movement events (total movement > 30mm) lie within high API regime. Again, a positive and non-linear association exists between the API and movement events. This confirms that antecedent rainfall affects movement of Condon Landslide, although no consistent quantitative correlation can be established.

Relationship between magnitude of API and timing of movement

API analysis was performed to investigate the relationship between daily API values and the initiation of movement events, and to determine a movement threshold if possible. Figure 39 displays the API at the first movement day of the large movement events versus total movement. Only 16 large movement events from 10/84 to 5/90 were included in this plot because the large events dominate the movement patterns. Small movement events show no patterns with respect to API values (Appendix A). In this plot, no apparent relationship is observed between total movement

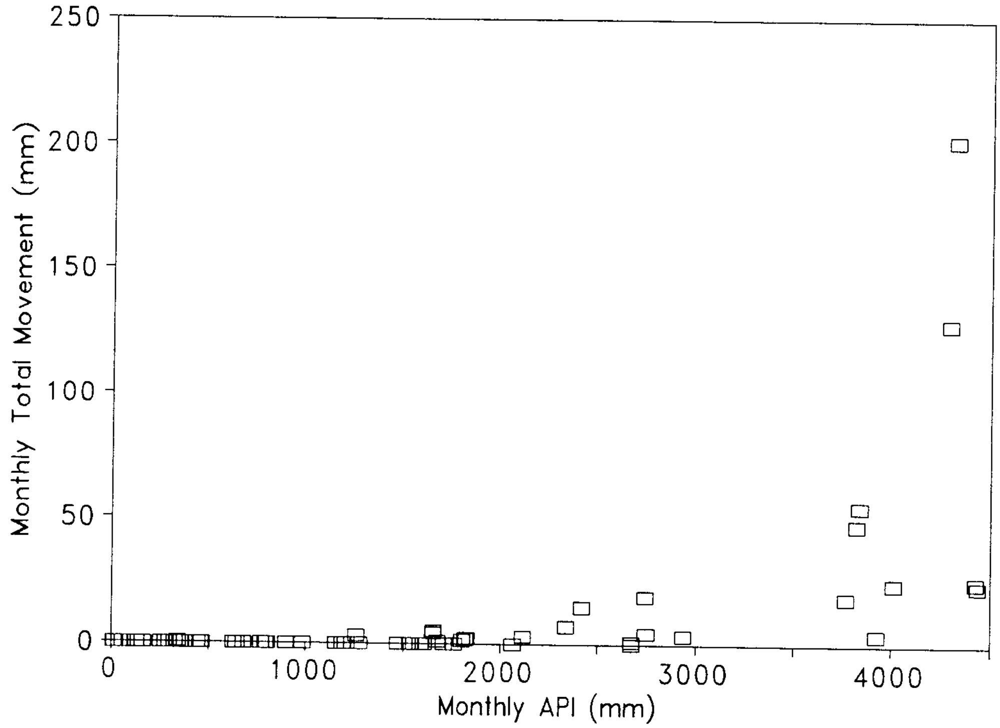

Fig. 35 Relationship between Monthly API and movement of Condon landslide. Periods of data are 10/84 to 2/87, and 10/87 to 6/90.

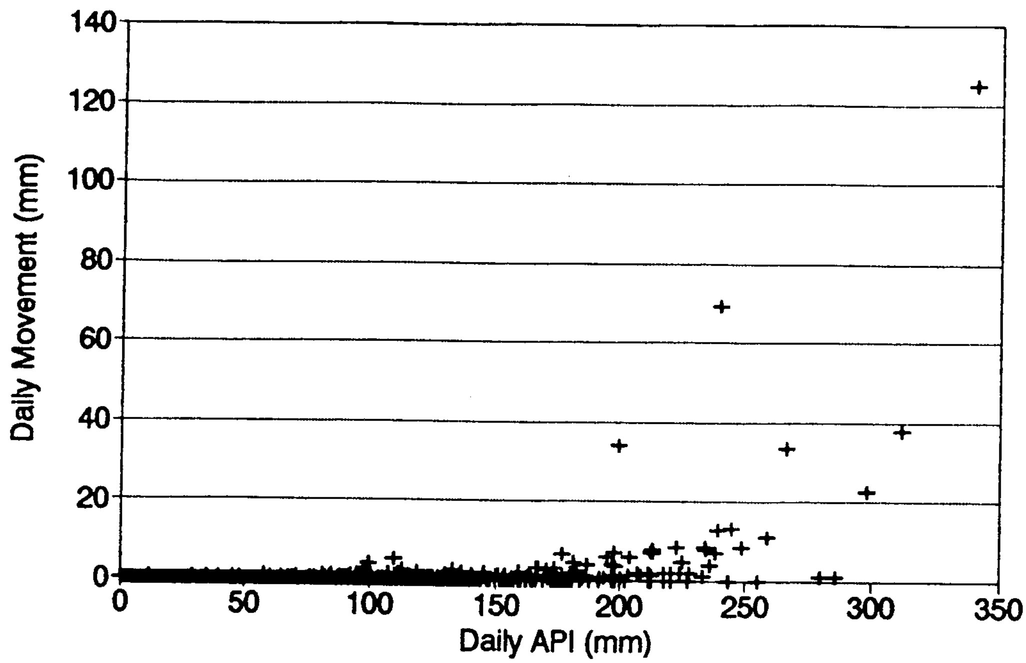

Figure 36 Relationship between daily API and movement of Condon Landslide, water years 1985-90. This plots shows movement generally occurred at high API ranges.

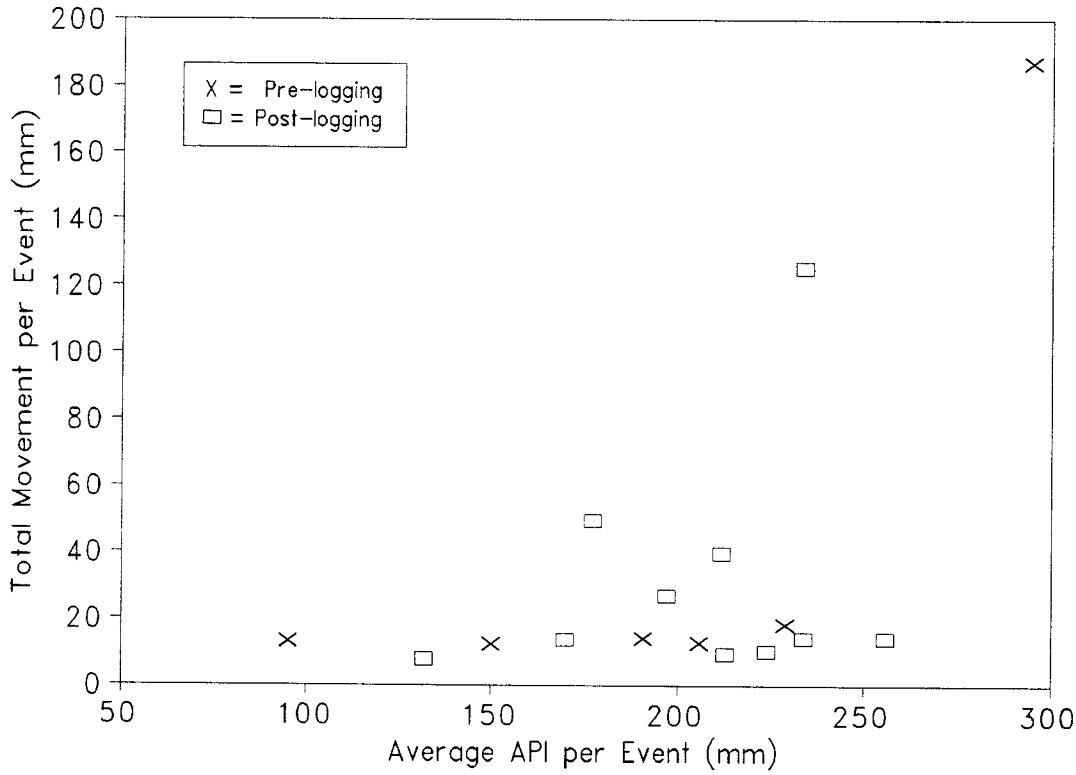

Figure 37 Relationship between average daily API and total movement per event for the 16 large movement events that at Condon landslide, 10/84 to 6/90.

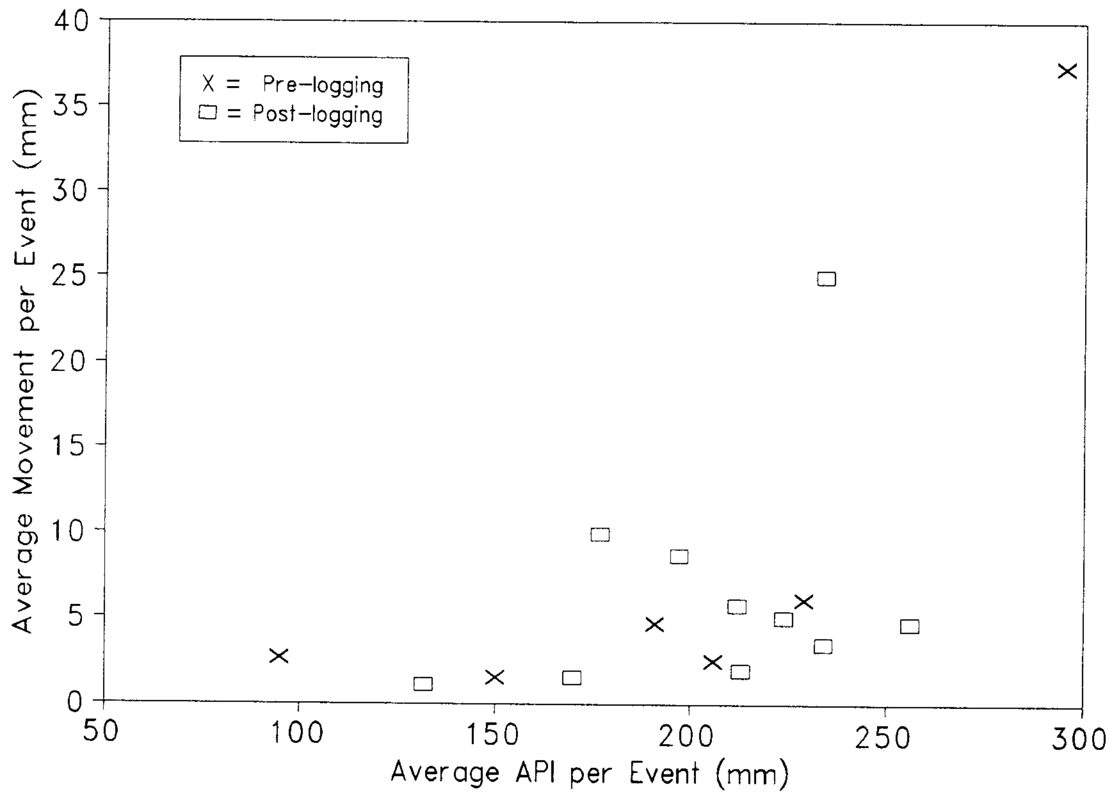

Figure 38 Relationship between average daily API and average daily movement for the 16 large movement events at Condon Landslide, 10/84 to 6/90.

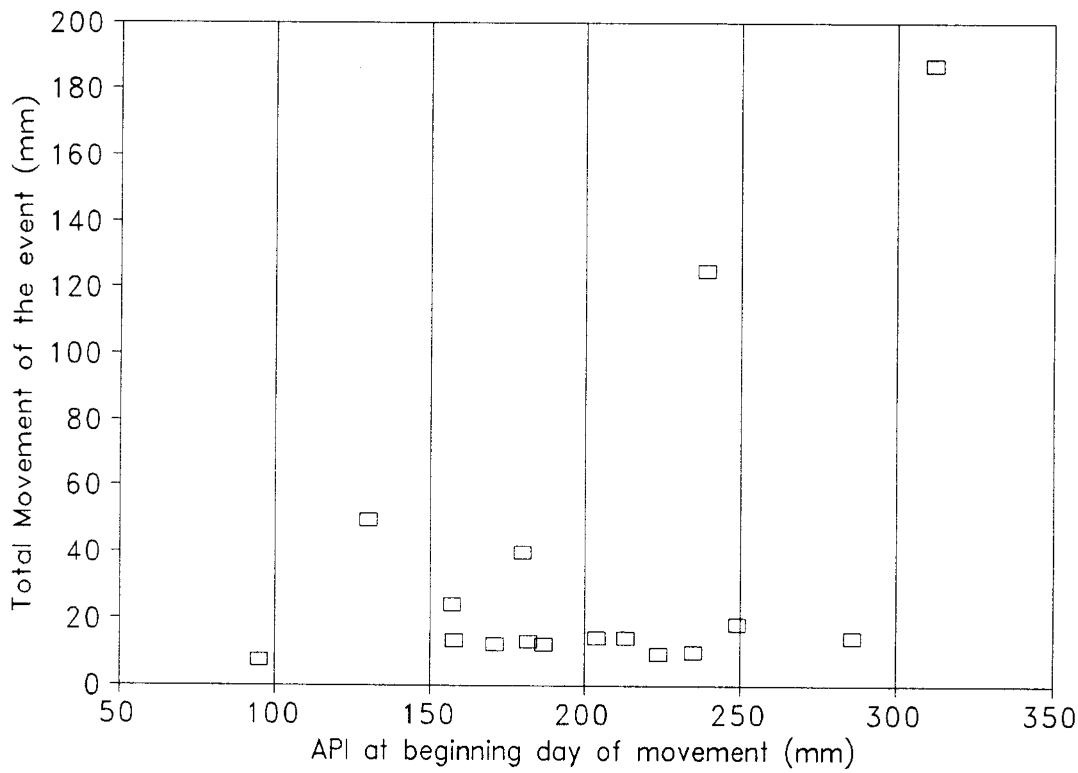

Figure 39 API at on-set of large movement events versus total movement, Condon Landslide, water years 1985-90.

amounts and the beginning API of the events. However, 14 out of 16 (88 %) large events had their beginning API exceeding 150 mm.

Frequency plots of daily API with and without movement also reveal an increase of days (as much as 37 percentile change for 40 mm API groupings) with movement above the 121 to 160 mm API group (Figs 40a & 40b). The two frequency plots, which represent pre- and post-logging conditions respectively, have similar results: both demonstrate that high daily API values are associated with movement in most occasions, while more than 50% of days with API above 160 mm had movement in both cases. These two plots concur with Figure 39, in which 88 % of the large movement events began after API had risen above 150 mm. A threshold of 160 mm can therefore be established as the minimum API level that could initiate large movement events at the Condon Landslide, during periods of pre- and post-logging.

Another study based on Crozier's (1986) antecedent method was done for comparison. Figure 41 is based on his methodology (Crozier, 1986, p. 185-186), in which the daily antecedent precipitation factor (x-axis) included precipitation only up to the previous day. The analysis contains two variables: daily precipitation and daily antecedent precipitation conditions. The Condon data set

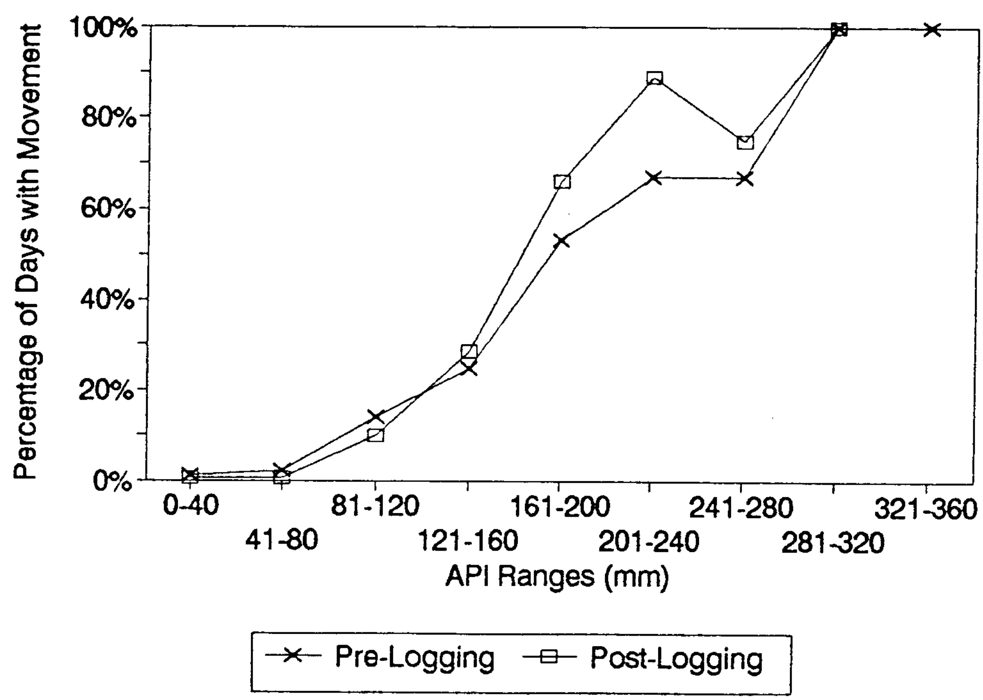

Figure 40 Frequency of movement at different API ranges, pre-logging and post-logging conditions, Condon Landslide, Water Year 1985 to 1990.

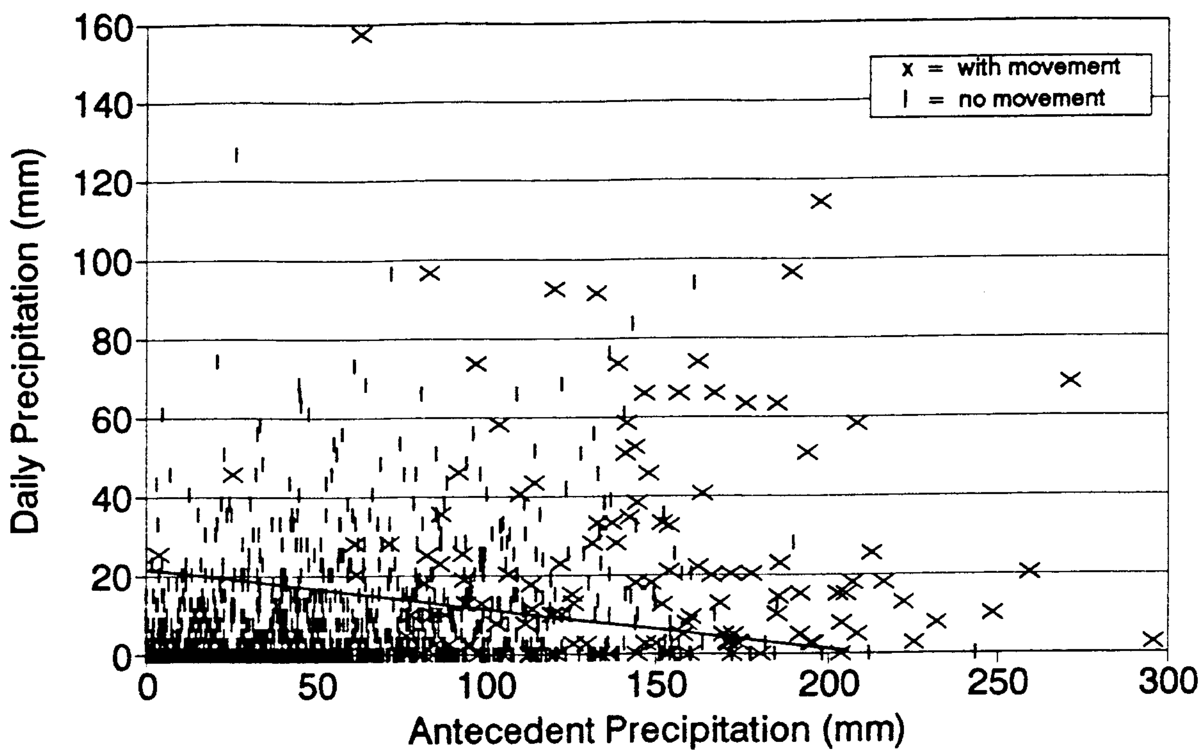

Figure 41 Antecedent rainfall condition and precipitation during movement, Condon Landslide, Crozier's method (1986). The straight line marks the threshold condition in which roughly 80% of the days to the right of the line experienced movement.

did not produce a clear-cut result, as opposed to what Crozier obtained from shallow slips in the Otago Peninsula of New Zealand. The straight line in the plot represents a probability threshold line, in which there is an 80 % chance that movement would occur when climatic conditions fall to the right of the line. Thus, the line shows that with daily precipitation as little as 15 mm, movement is likely to commence when API is above 160 mm on that day.

Discussion

Since most major movement events (87 %) stopped when the API began to decrease, continuation of movement of the Condon Landslide depends on continued high levels of water input.

Based on the monthly API and movement plot (Fig. 35), total monthly movement increases as a result of increasing precipitation, and under extreme conditions of continuous water input (> 3500 mm API in this case), total monthly movement could increase greatly. Using API on a monthly scale seems to characterize the movement pattern better than using original monthly precipitation (compare to Fig. 25).

All plots of API and movement amounts confirm the relationship between the two parameters: large amounts of movement, either on a daily scale, or as individual events, were associated with high API values (Figs. 36, 37 & 38). Since high API periods are often the result of large, persistent winter storms, movement events at Condon Landslide are primarily driven by sudden increase of water inputs into the soil mantle of the slide. However, no linear relationship is observed between magnitude of movements and API in this analysis.

The frequency plots (figs. 40a & b) and figure 39 demonstrate that using API to predict the timing of the initiation of movement events of Condon Landslide is more accurate than by using precipitation data alone. The API threshold derived from the analysis is 160 mm, and when API jumps up to more than 200 mm, the likelihood to have movement increases tremendously. With API usually ranging from 50 to 80 mm during the wet seasons, it takes only a moderate storm (100 mm/day) to propel the API over the 160 mm threshold. Storms with maximum intensity over 140 mm/day (2-year return period) could very likely send the landslide moving (such as the 4/27-29/90 event), provided that appropriate antecedent conditions exist.

Crozier's antecedent method of predicting onset of movement uses two climatic variables, which account for effects of precipitation intensity on antecedent conditions on a daily basis. His method is most suitable on predicting shallow soil-slip failure and debris flow, because they respond quickly to individual storms and storm intensity. In the case of Condon landslide, the threshold of movement concluded from his method contains uncertainties that make it no better than the simpler API model. Therefore, an API threshold of 160 mm is favored over the 80 % probability threshold from Crozier's method.

Testing of the Estimated Threshold

The predicted API thresholds were tested independently on Graham's record, to assess their practicability and consistency of predicting the onset of movement. Unfortunately, most of Graham's (1985) movement records from water year 1980 to 1983 were derived from stake array measurements, which only gave the total amount of movement between measurements. From December 1981 through February 1982, a total of 205 mm of movement were recorded. He did document one specific movement event (1/24-26/82), which was recorded by the extensiometer with daily movement rates (Graham, 1985, p.56). This major movement lasted for only two days, but had a total movement of 135 mm. API with a recession factor of 0.87 was calculated from 10/1/81 to 2/13/82, using the Mapleton precipitation record (fig. 42). Daily precipitation record was missing from 2/14 to 2/28, when only 37 mm of precipitation were recorded. No major storms occurred in March 1982, in which only 236 mm (daily

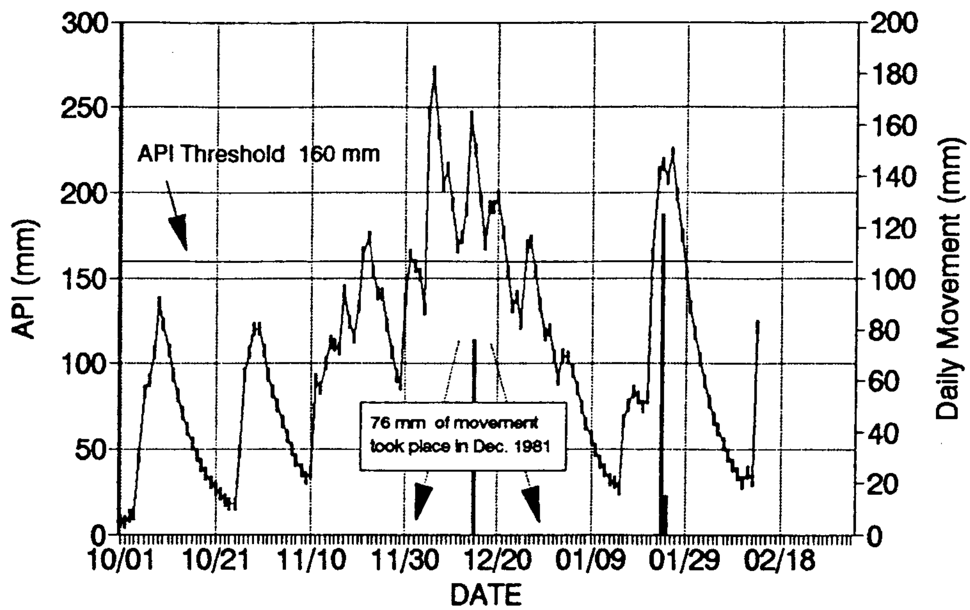

Figure 42 Testing of the application of API method on Condon Landslide, from 10/1/81 to 2/13/82. The calculated threshold line (API = 160 mm) is shown.

maximum: 30.5 mm) of precipitation were recorded. Therefore, days after 2/13/82 were left out of the calculation.

The plot shows two major periods of high API, corresponding to intense precipitation: 12/5-28/81, and 1/21-26/82. The second period coincides with the documented major movement event. During this event, API had risen above the threshold level (160 mm) on 1/22/82, and movement began on 1/24/82, when API exceeded 220 mm. Therefore, the API-threshold method predicted only one day earlier than the beginning of the actual movement. In December, 1981, the API plot crossed the threshold line three times, and 20 out of 31 days had API more than 160 mm. During the same period of time, Graham (1985) had recorded a sum of 76 mm of movement from repeated stake array measurement. Even though the daily record was lacking during this period, I assume that the estimated API threshold was exceeded significantly, and that major movement events occurred during the 12/5/81 to 12/20/81 period.

Sources of Error

Since the API model considers only precipitation input to the hydrologic system of the landslide, its accuracy could be impaired by other hydrologic parameters, such as groundwater flow regime and rain-on-snow effects, as well as the local variations of the mechanical behavior of the landslide body. Despite these considerations, precipitation seems to be the most important driving force at the Condon Landslide, and using precipitation as the sole variable in the API model is assumed to be acceptable for semi-quantitative assessment.

Another problem within the model deals with the recession coefficient constant (C). To simplify the model, a single recession coefficient was used throughout the wet season, and in both pre- and post-logging periods. However, the rate of groundwater decrease may change through time at the Condon site, because movement could induce changes in subsurface hydrology. Therefore, a single recession coefficient may not represent the true groundwater recession rate at the Condon Landslide in both short (days) and long (year-to-year) time scales, and this may over- or under-estimate the API, and affect the accuracy of the predicted threshold. Formulating non-linear recession coefficients throughout the wet season is difficult here, due to the lack of full understanding of the hydrological behavior of the Condon Landslide.

Plant phenology is another factor that affects C, as seasonal variation in plant growth and leaf area alter evapotranspiration (ET) rates, although the significance of differences of C throughout the year is unclear. Moreover, clearcutting on part of the landslide could alter the evapotranspiration rates averaged over the landslide mass; therefore, C might be different in the pre- and post-logging periods.

The distance and elevation differences between the Condon Landslide and the Mapleton raingage could make the calculated API depart from the realistic situation at the landslide site. Nevertheless, this problem seems to have little effect on this analysis, due to the temporal resolution (daily) of the analysis. However, this problem should be considered, if a more refined time-scale is used for movement prediction.

Applicability of the API Model to Predict On-set of Movement

Much landslide literature has described the use of precipitation records in the prediction of amounts and initiation of landslide movement (e.g. Crozier, 1987). However, the relationships found between precipitation patterns and movement at different landslides are often weak and inconsistent (e.g. Iverson and Major, 1987, and Keefer, 1983). The API model tested at Condon Landslide is an improved technique over the "traditional" methods which use precipitation records only in predicting the initiation of landslide events. However, the API model failed to relate magnitude of movement to specific ranges of API.

Although the API model accounts only for precipitation input and possible groundwater condition at a specific locality, the empirical nature of the model need not be restricted to specific sites. Since the model is best used to predict initiation of movement events but not for a quantitative assessment of landslide movement, a specific model, like the one developed for the Condon Landslide, could be used throughout the region with similar geologic and climatic conditions. However, this model has limited usefulness in areas with a strong influence of snow, and in terranes with landslides characterized by prolonged periods of movement, such as in the western Cascades.

Generally, with accurate on-site movement observation, some measured hydrologic parameters, such as groundwater level and precipitation, and careful model calibration, the API method could be a simple and practical tool to predict landslide movement.

X. Effects of Clearcutting on Movement Patterns

Method

One of the most important effects of clearcutting on hillslope hydrologic system is alteration of evapotranspiration (ET). Evapotranspiration consists of two components: (1) evaporation of moisture from all surfaces (plants, ground, etc.) and shallow subsurface soil layers; and (2) transpiration: the uptake of water by plants from soil, and the subsequent release of the moisture to the atmosphere from stomata or through evaporation. Various atmospheric (e.g. temperature), vegetative, and soil properties affect the rate of evapotranspiration (Brown et al, 1972). Generally, researchers believe the removal of trees reduces the evapotranspiration rate because of the elimination of the transpiration component. This results in higher soil moisture content in the soil mantle throughout the year (especially early fall), and increased streamflow and subsequent soil erosion during early winter months (Rothacher, 1973). In theory, this increase in soil moisture content could affect landslide stability and movement patterns by maintaining relatively higher groundwater level and thus higher pore water pressure.

One-third of the Condon landslide area (the active part) was clearcut and slash-burned during the summer of

1987, and continuous movement records before and after clearcutting are used herein to detect any influence on timing and magnitude of landslide movement by the removal of trees on slopes at the Oregon Coast Range (fig. 7). Extensiometer records are available for water years 1985 to 1990, with 3 years of pre-logging and 3 years of post-logging periods for comparison. Stake array records from water years 1981 to 1984 are used to assist this analysis. Total and average annual movement and precipitation, frequencies and initiation of movement of different types of movement events, and API values are used to assess the logging effects on movement patterns of the Condon Landslide.

Results

The three post-logging water years registered slightly more precipitation (4 %), with more numerous precipitation events that exceeded the API threshold for movement initiation, and more movement (18 %) than the three pre-logging years (Table 3). The three post-logging years also had more large movement events that are brief and fast-moving, although the two comparison periods had similar numbers of small and large slow-moving events. The movement rates of the three types of events are comparable between the pre- and post-logging periods (Table 2). The API analysis shows no remarkable differences in timing of

	Annual Ppt. (Oct.-Apr.) (mm/yr.)	Annual Movement (mm/yr.)	No. of movement events			Precipitation events resulted in API >160mm	No. of large movement events after API>160mm
			Small	Large Slow	Large Fast		
Pre-logging (WY 1985-1987)	1895	95	17	3	3	7	5
Post-logging (WY 1988-1990)	1978	112	14	3	7	11	10

Table 3 Comparison of types and frequency of movement events, average total movement and precipitation, and API values between pre- and post-logging conditions.

movement initiation, and average API per large events between pre- and post-logging periods (see also Chapter 9). The largest apparent difference between the pre- and post-logging periods is in the number of large movement events that are brief and fast-moving (from 3 to 7) over each three-year period.

Discussion

Although removal of trees on the landslide might effectively reduce the shear strength of the landslide mass by increasing the moisture content within the slide body because of decreasing evapotranspiration, thus making it more prompt to unstability and movement, movement records from the Condon Landslide do not demonstrate that this effect is significant. Total annual movement of pre- and post-logging years all fall within a similar range, in which the post-logging years had 18% more movement (Table 3); while at the same time average precipitation of the post-logging years was only 4% higher than pre-logging years. Therefore, clearcutting of the Condon landslide does not appear to have accelerated movement, based on available information.

Clearcutting a large part of the Condon Landslide might also alter the sensitivity of API analysis because the responsiveness of the landslide movement to rainfall events could be altered by changes of the landslide hydrology caused by clearcutting. After clearcutting the API threshold on movement initiation might become more sensitive to smaller rainstorms because more moisture is retained within the landslide body. Based on comparison of data from pre- and post-logging periods, clearcutting of the landslide site apparently has not significantly modified the API value corresponding with movement initiation. From 10/84 to 2/87, the API threshold was exceeded 7 times (Appendix A), and 5 of those occasions (71%) movement was initiated. During the post-logging years (1988-1990), the API threshold was exceeded 11 times (Appendix A) and on 10 of those occasions (91%) movement was initiated. However, the patterns of landslide initiation between pre- and post-logging periods do not differ greatly; movement during both periods was initiated 0 to 1 day after the threshold was exceeded, except the threshold apparently worked better after logging. Therefore, logging two-thirds of the landslide site may not have had a significant impact on the internal hydrologic system of the slide, based on the resolution of the analytical approach used here. With only a relatively short period of continuous record and lack of extensive groundwater data, this conclusion is limited.

Another difference between pre- and post-logging periods is that a bigger percentage (33% vs. 10%) of movement events began at API values less that 160 mm during pre-logging period than in the post-logging years (see also figure 39). Furthermore, the pre-logging period had more small events (total movement < 5 mm, and total duration less than 5 days) (17 vs. 14), and they always corresponded with low API values (< 150mm) (Appendix A). The post-logging years have experienced fewer small movement events, but have more major events, reflecting in part more numerous precipitation periods with API > 160 mm (Table 3 and Appendix A).

In an effort to examine possible effects of clearcutting on hydrology and landslide movement, an attempt was made to utilize existing hydrologic models to evaluate changes of clearcutting on evapotranspiration rates and soil moisture in the central Coast Range. Unfortunately, due to problems in modelling and model parameterization, no reliable results were produced by the two tested models. However, preliminary findings from one model (PRMS) using Alsea Watershed data indicated that ET rates did not change significantly on a small Coast Range watershed during the first four years after clearcutting. Other researchers have found evidence of minor effects of logging on stream runoff. Rothacher (1973) reported increased average peak streamflows of small runoff events early in the wet season after clearcutting from small experimental watersheds in the western Cascades. However, he concluded logging did not increase peak flows of very large storm events that occur later in the wet season, after the soil mantle has been thoroughly wetted. Harr and McCorison (1979) described similar findings in other small watersheds on the H.J. Andrews Experimental Forest in the western Cascades, and they observed no significant changes in size or timing of peak flows from rainfall events alone. They concluded that clearcutting could alter snow accumulation and melt rate, thus modifying the hydrologic response to winter storms. In the central Coast Range, where snowfall is infrequent and small, this effect is negligible.

These published findings indicate that clearcutting may have only minor effects on altering the hydrologic system of hillslopes in western Oregon, and therefore may not greatly modify the very large, rainfall-triggered landslide movement events. Based on this reasoning, clearcutting may not have a profound impact on timing and magnitude of movement of deep-seated landslides in the central Coast Range, although clearcutting may promote small scale mass movement and surface erosion, and this is observed in Condon landslide. While the movement magnitude and pattern have not changed drastically after clearcutting, the landslide toe has experienced debris slides and widespread slumping after the removal of vegetative cover (Figs. 11 and 12).

XI. Physical Characteristics of Selected Landslides

Locations

The Condon and Wilhelm Landslides are located at the Mapleton Ranger District, Siuslaw National Forest, central Oregon Coast Range, Oregon (figs. 1 & 2). The Wilhelm Landslide is located approximately 6 km ENE of the Condon Landslide (fig. 2). The Lookout Creek Landslide is in the H. J. Andrews Experimental Forest, about 80 km east of Eugene (fig. 3). The Mid-Santiam and Jude Creek landslides are at the north-central part of the western Cascade Range, approximately 32 km north of the Lookout Creek Landslide (fig. 3).

Physical Features of the Selected Landslides

Common geomorphological features, such as hummocky topography, tension and shear cracks, and active grabens, characterize the selected landslides from both mountain ranges, yet the detailed configuration of individual slides is unique, due to site specific effects of geology, geomorphology, and soil properties. Descriptions of the Condon and Wilhelm Landslides are given by Graham (1985). More recent observations of the surface topography of the two slides are noted in this thesis. Hicks (1982) and George Lienkaemper (per. comm., 1986) described the Mid-Santiam and Jude Creek Landslides. The Lookout Creek

Landslide is described by Swanson and Swanston (1977) and Pyles et al. (1987). The following are brief descriptions of the surface geomorphology of the selected landslides, except Condon Landslide, which is described in chapter 5 of this thesis.

The Wilhelm Landslide covers approximately 8.0 hectares of forested land, and is elongated northwest-southeast (fig. 43) (Graham, 1985). Its upper end is bounded by a scarp 12 meters high. The main body of the slide consists of large grabens, and shear and tension cracks that show signs of translatory movement. The landslide is more than 12 m thick at some locations (Graham, 1985). The toe area of the slide is characterized by oversteepened slopes covered by landslide debris, large blocks of weathered basalt and sandstone, downed trees, and detachment sites of debris flows, which are now bare soil or partially vegetated with red alder, and subjected to surface and small scale mass erosion. Graham (1985) reported these debris flows severely affected the Wilhelm Creek channels for several kilometers downstream in the past few decades.

The Middle Santiam Landslide has a nearly vertical bedrock headscarp up to 30 meters high (fig. 44). Immediately below the headscarp is a 4-hectare, gently-sloping (4°) bench, which is cut by numerous tension cracks

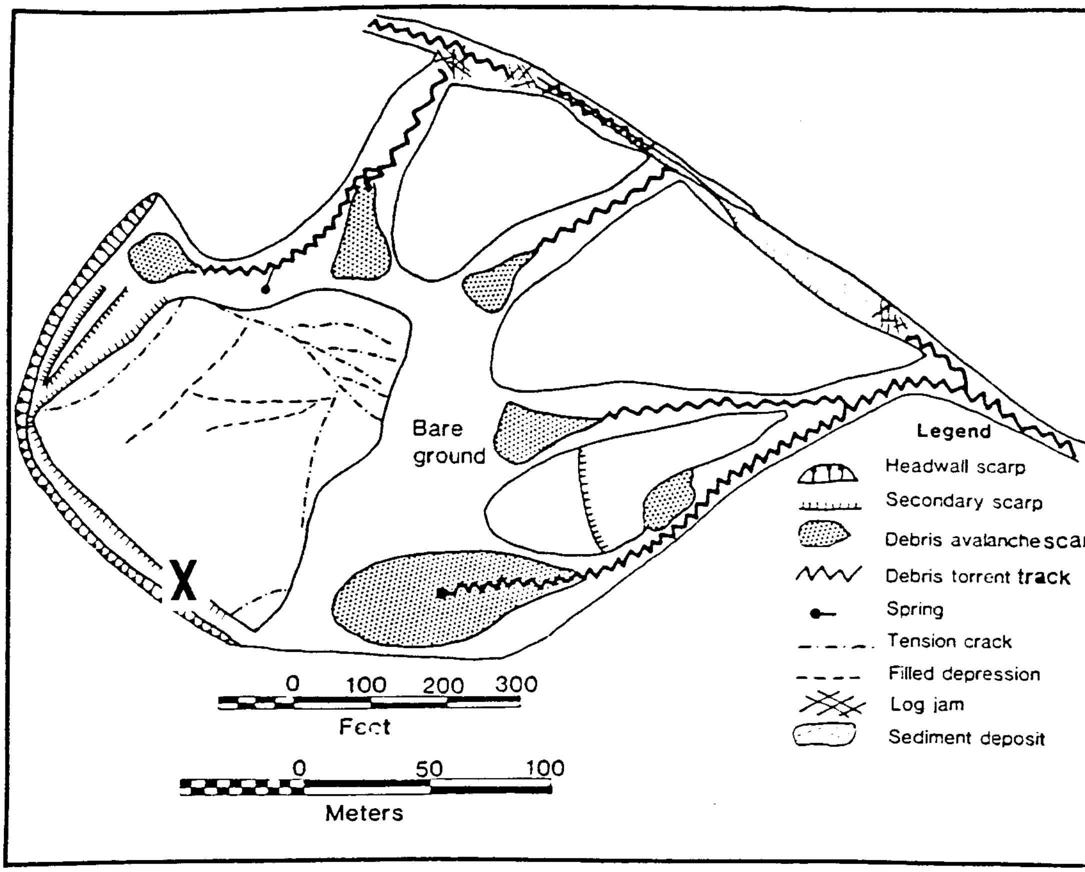

Figure 43 Geomorphic map of the Wilhelm landslide, X is the site of the stake array and extensiometer used in this study (modified from Graham, 1985).

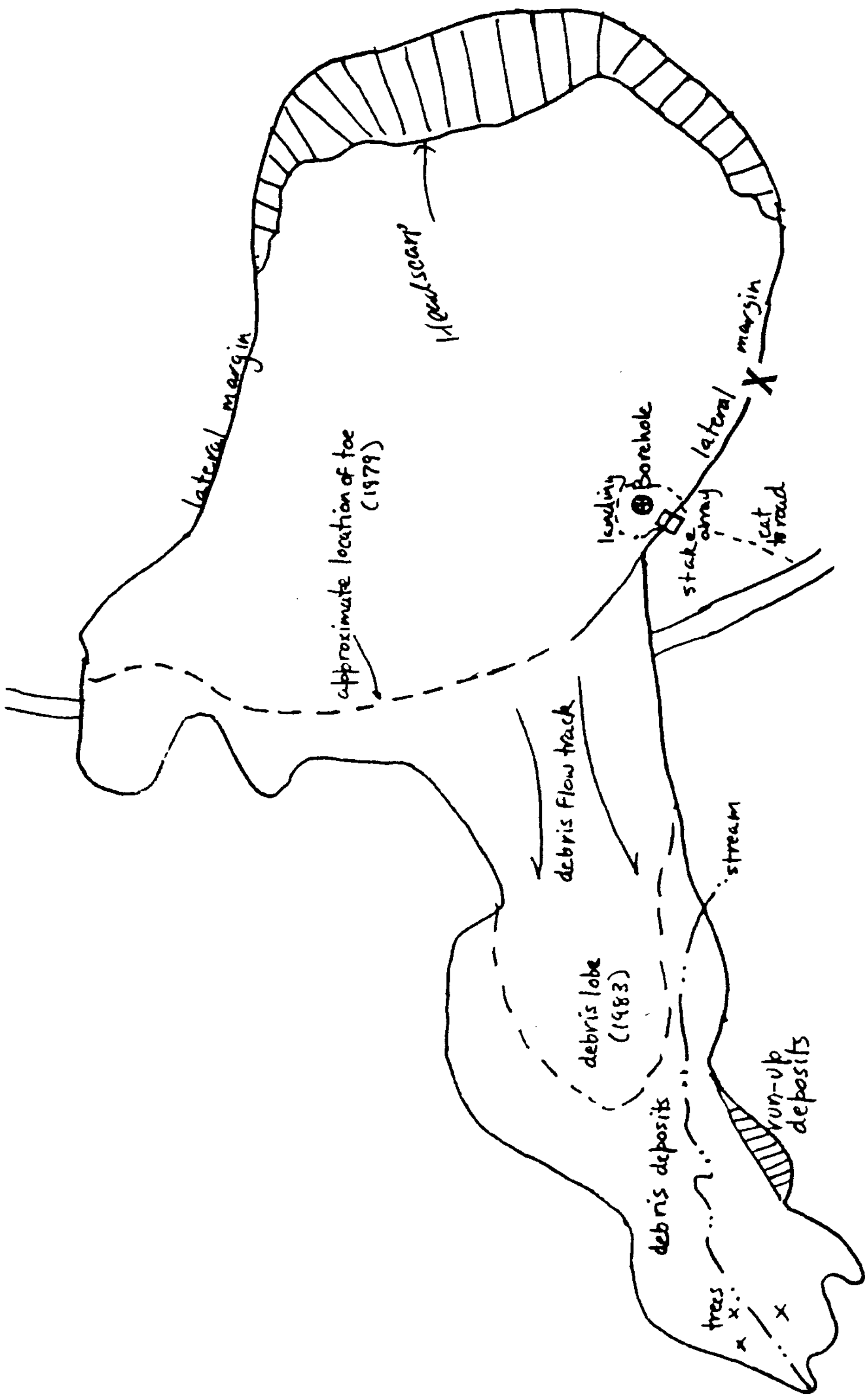

Figure 44 Simplified geomorphic map of the Mid-Santiam Landslide. X is the location of the stake array and extensiometer used in this study (modified from Hicks, 1982).

and scarps. The middle and lower parts of the bench are characterized by hummocky topography, which is an indication of considerable extentional deformation (fig. 45). Below the bench is a steep slope (21°), which passes into several debris lobes, extending onto a flat river terrace.

The toe area has been constantly active for more than a decade (Craig Creel, Forestry Sci. Lab, per. comm., 1990). The upper part of the landslide has a basal shear zone 14 to 15 m below ground surface, based on inclinometer tube observation (Hicks, 1982). The landslide toe lobes engulfed the access road in 1989 (M. Grayer, Sweet Home Ranger Dist., Willamette Nat. Forest, per. comm., 1991).

The Jude Creek landslide advances into Jude Creek, a tributary of the Middle Santiam River. The entire landslide occupies 30 ha of land in a small basin, covered by old-growth forest, and is bounded by extensive bedrock and soil scarps (fig. 46) (Hicks, 1982). In the past decade, only the lower 3.4 hectares of the slide has been active. This area is highly disturbed by slide movement of 6 to 9 m/year. Springs, tension and shear cracks, and a deranged drainage network characterize the landslide surface topography and hydrology. The forest canopy at the lower part of the slide is largely destroyed, while tipped and deformed trees in other parts of the slide reflect a history of mass movement

Figure 45 Hummocky landscape at the middle part of the Mid -Santiam landslide. Photo was taken in December, 1990.

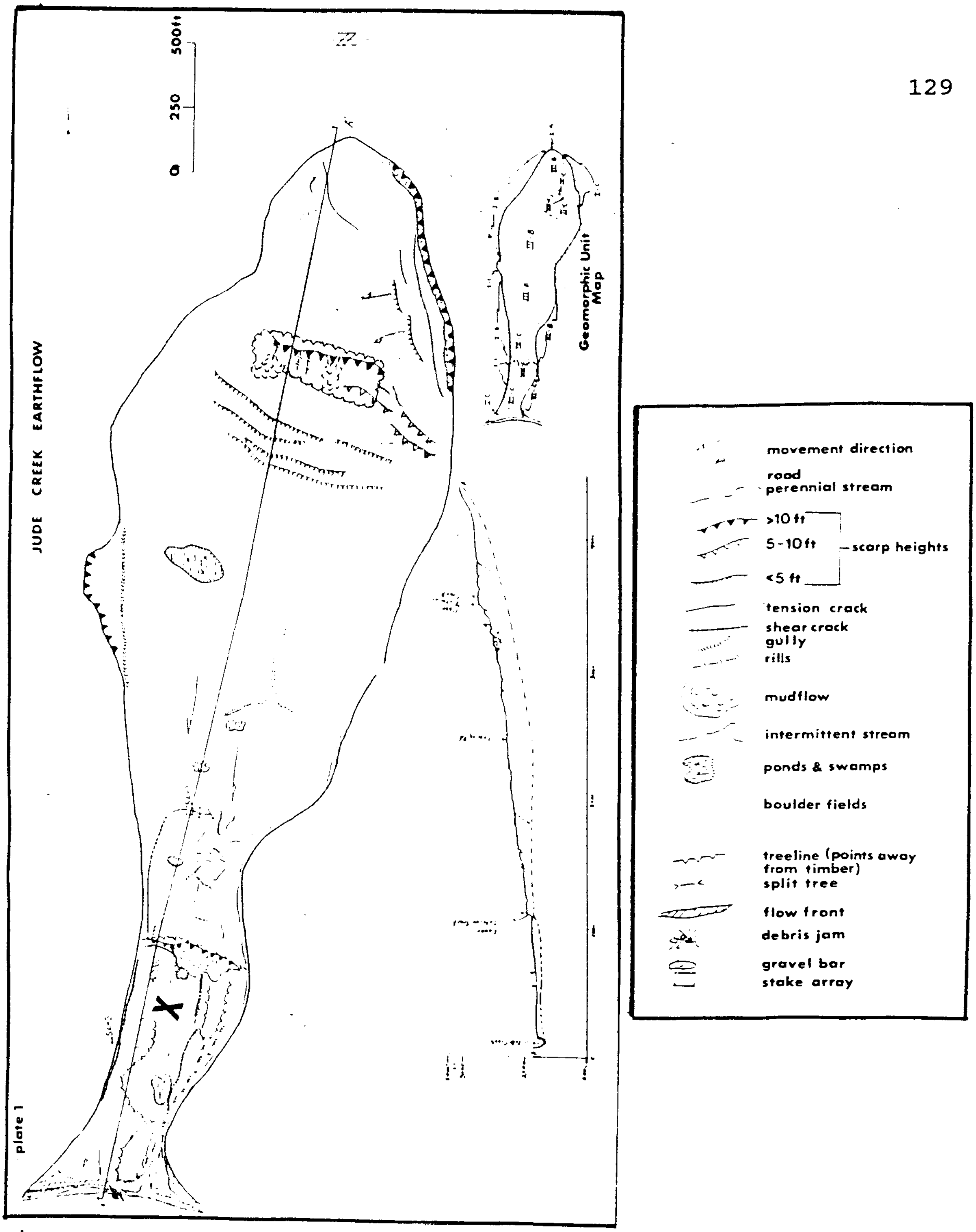

Figure 46 Map of the Jude Creek landslide. Most of the discussion in this study is focused on the active (lower) part of the landslide, which is highlighted in this figure. X is the location of the extensiometer and stake array used in this study (modified from Hicks, 1982).

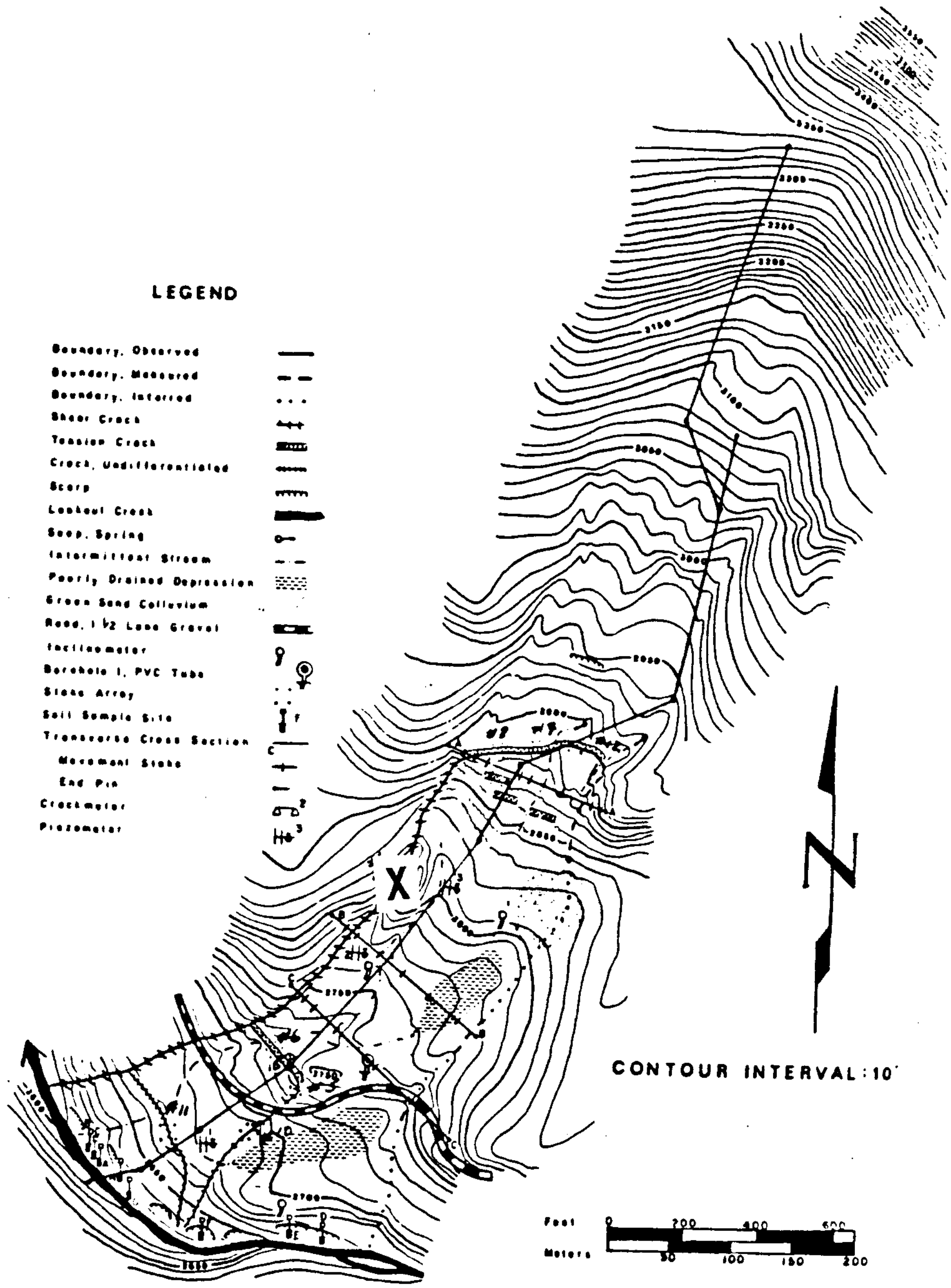

Figure 47 Topographic map of the Lookout Creek landslide, showing major geomorphic features. The stake array and extensiometer used in this study is marked by an X (modified from Mills, 1983).

stretching over centuries (Vest, 1989).

The toe of the landslide is constantly eroded by Jude Creek, and forms a steep, gullied colluvium front. A massive accumulation of large woody debris and boulders from the slide persistently constricts and periodically blocks Jude Creek where the toe enters the stream. Immediately downstream the channel is scoured by occasional debris flows originated from the toe area, while aggradation is observed upstream of the slide. Hicks (1982) reported that about 60% of Jude Creek's main channel, including all of the channel downstream of this slide, is directly affected by activity of various landslides.

The 13-hectare Lookout Creek landslide flows south into Lookout Creek. This 900 m long and 150 m wide slide is characterized by stepped topography, with a succession of steep and gentle slope segments, poorly drained depressions, and systems of shear and tension cracks (fig. 47) (Swanson and Swanston, 1977, Pyles et al, 1987). Currently only 40 % of the landslide area is active (up to 490 m from Lookout Creek).

XII. Movement History and Characteristics of Selected Landslides

Field Methods of Movement Measurement

Three principal methods have been employed to monitor movement of the five selected landslides: stake arrays, extensiometers, and surveying. Details of the instrumentation are provided by Hicks (1982), Graham (1985), Mills (1983), and Pyles et al (1987). A summary of methods is given in Chapter 6.

The Central Oregon Coast Range

Condon Landslide

Detailed descriptions and discussions of the movement history and characteristics of the Condon Landslide can be found in Chapters 5 and 8, and Tables 2 and 4 of this thesis.

Wilhelm Landslide

A brief movement history of this landslide was given by Graham (1985). Based on aerial photo and tree ring interpretations, Graham noted that a major movement episode took place in 1971. Other than this anomaly, he did not describe additional incidents. He attributed this to a lack of evidence from dendrochronological studies, probably due to the translatory nature of movement, which resulted in

Water Year	Condon	Wilhelm	Mid-Santiam	Lower Jude	Lookout
		(Unit of total movement = millimeter)			
1981	67	n/a	n/a	n/a	83
1982	205	n/a	6280	7250	206
1983	143	n/a	3290	17610	65
1984	50	n/a	2290	7260	180
1985	41	16	2450	1420	71
1986	221	95	3330	6770	133
1987	23	44	3680	4050	12
1988	52	18	3990	5530	11
1989	82	21	4580	7710	22
1990	202	12	4520	12980	08
Average (mm/year)	109	34	3823	7842	79

Table 4 Annual movement records of the five selected landslides, from water year 1981 to 1990. n\a = record not available or missing. Measurement sites (stake-arrays and extensiometers) are shown in figures 8, 43, 44, 46 & 47.

minor tipping of trees. Surface features of the landslides, such as extensive development of small scarps overprinting older, larger scarps, and mature (20+ years) alder stands found at part of the toe area of the slide, suggest that the slide may have been active at least intermittently for possibly more than half a century.

Continuous monitoring of the Wilhelm Landslide with extensiometer and stake arrays began in 1984 and continued until June, 1990. The total amount of movement is 206 mm (33 % of Condon's total movement) in this monitoring period (Table 4). In general, movement events of the Wilhelm Landslide are smaller and less frequent than the Condon Landslide (fig. 48). This is reflected in its annual sum of movement. Most movement events were short, stretching for 1 to 4 days and commonly having total movement of 1 to 3 mm. Movement was generally distributed evenly throughout an individual event, with brief periods (1/2 to 2 days) of faster movement intervals. The typical movement rate of those events was 0.5 mm/day. Fig. 49 shows a typical movement event as seen from the original cumulative movement data chart.

The western Cascade Range

Mid-Santiam Landslide

The Mid-Santiam landslide is a deep-seated and old

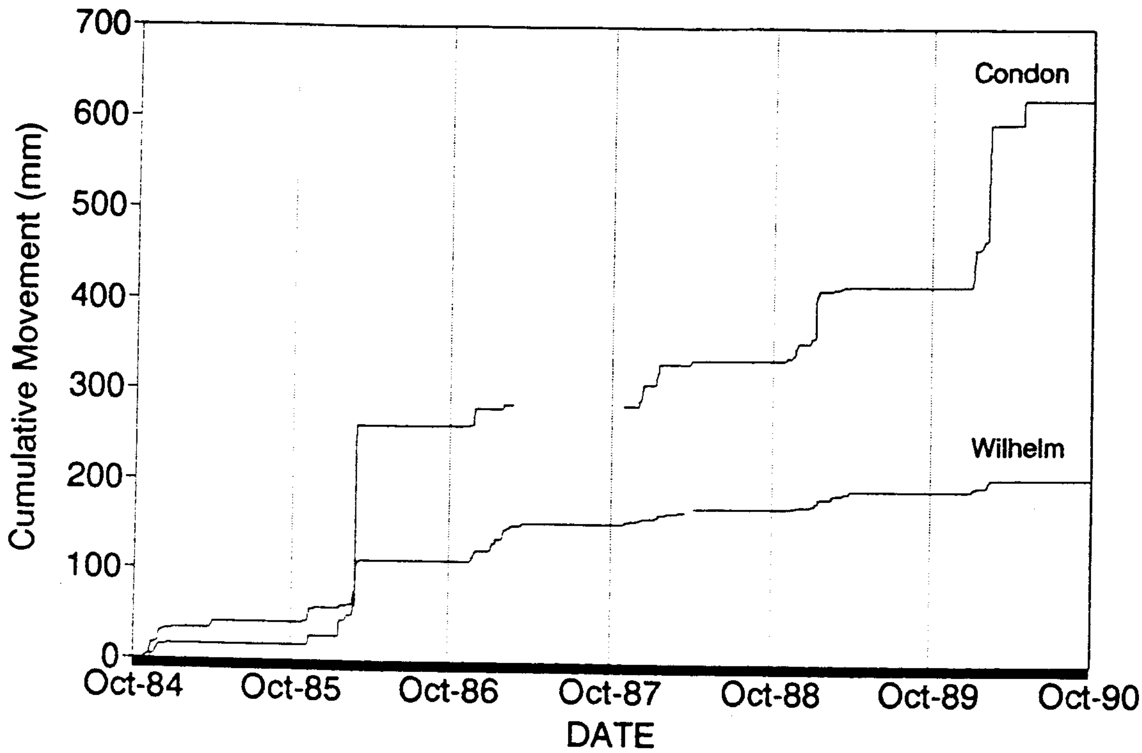

Figure 48 Daily cumulative movement records of the Condon, and Wilhelm landslides, from 10/84 to 9/90.

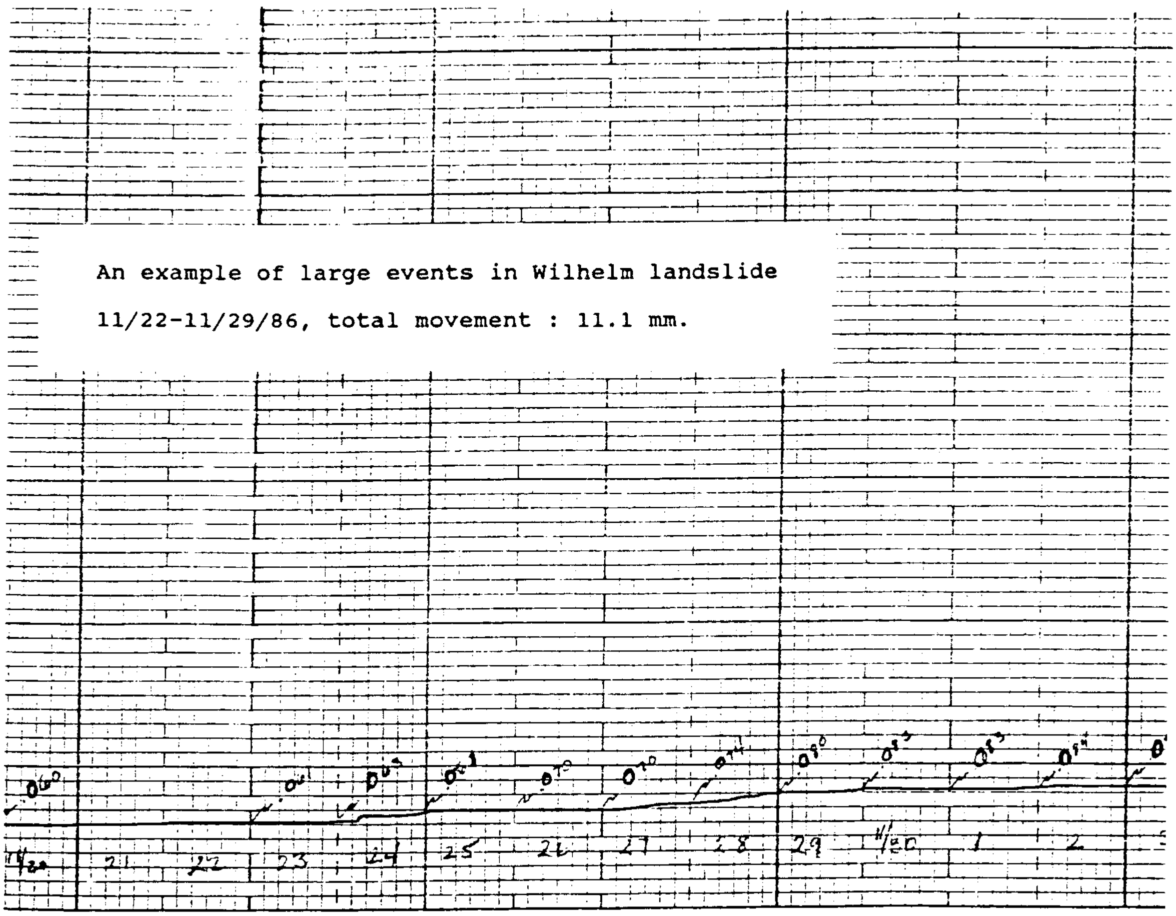

Figure 49 Field data plot showing a typical Wilhelm movement event. The event is similar to the persistent and slow-moving events recorded at Condon landslide.

landslide (Hicks, 1982). Based on the vertical form of the mature forest stand on part of the landslide, the landslide was relatively stable in the century preceding 1970 (Hicks, 1982). Road construction through the toe of the landslide in 1965 modified the configuration of the mass of the landslide, and possibly reactivated movement. Clearcutting of 42 percent of the landslide area occurred in the late 1960's and early 1970's. The original Forest Service road through the landslide toe was obliterated by landsliding in the mid-1970's. A new road was built on the river terrace around the toe of the landslide, but continued landslide movement throughout the 1970's and 1980's and numerous debris flows, have led to the engulfment of the second road in 1989.

The Mid-Santiam landslide moves throughout the year, with fast movement (20 to 70 mm/d) during wet, winter months; and relatively slow daily movement in summer months, typically less than 10 mm/d (Fig. 50a). The landslide tends to advance continuously and at uniform rate during winter months, with base rates ranging from 8 to 20 mm per day, which are interrupted by periods of accelerations, with maximum rates over 50 mm per day. The maximum daily movement recorded from water years 1982 to 1990 was 294 mm (11/24/88). These large, individual events were caused by snowmelt and rain-on-snow events in the spring (George

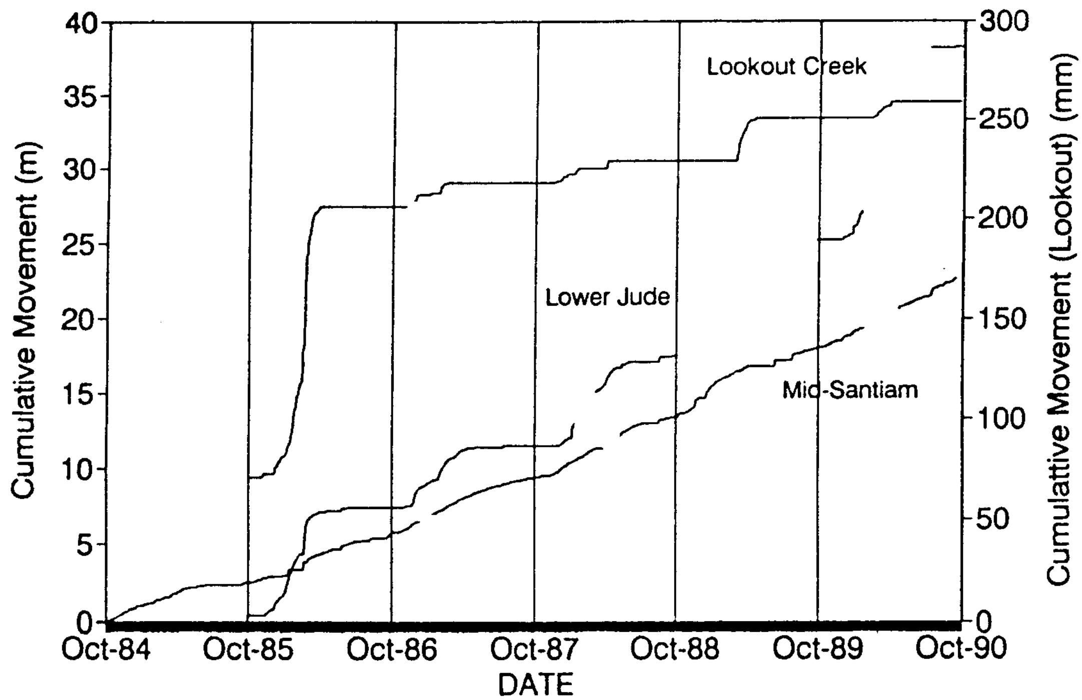

Figure 50a Cumulative records of Mid-Santiam and Lower Jude (left side), and Lookout Creek Landslides, from 10/84 to 9/90. The Lookout Creek Landslide has much smaller total movement than the other two slides (Note different scale).

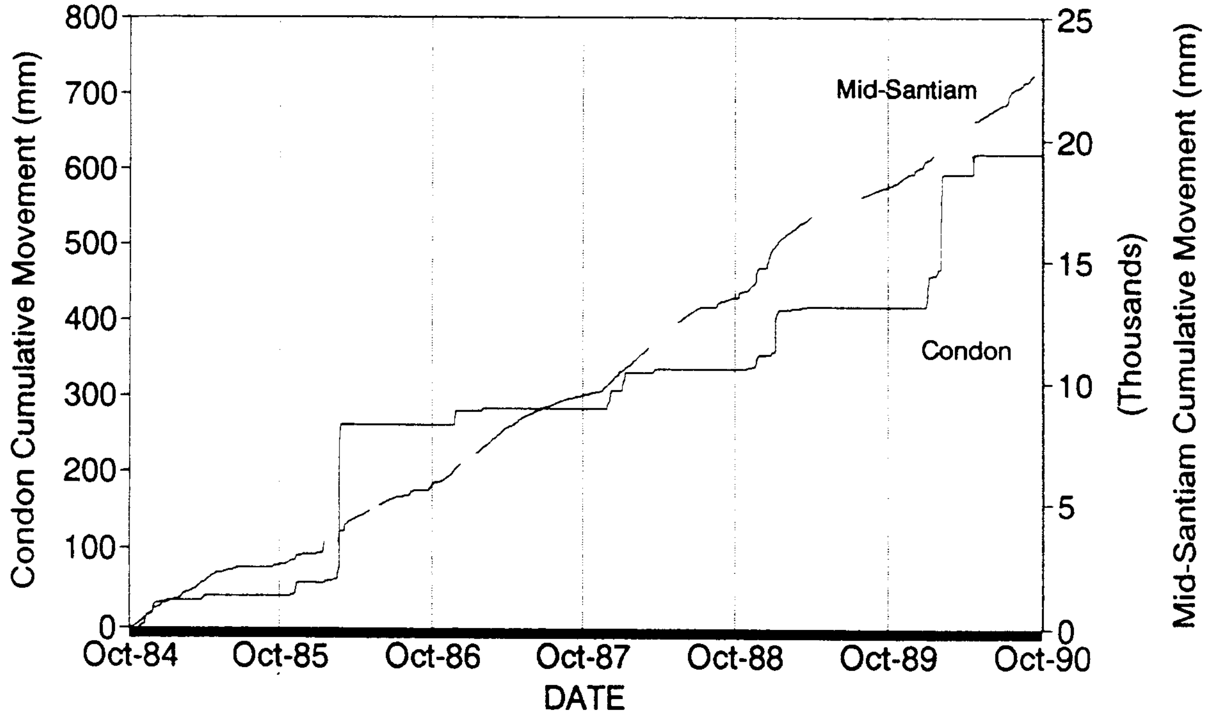

Figure 50b Cumulative movement records of Condon and Mid-Santiam landslides from 10/84 to 9/90. Showing the contrasting movement patterns found at monitored landslides from the central Coast Range and the western Cascades Range.

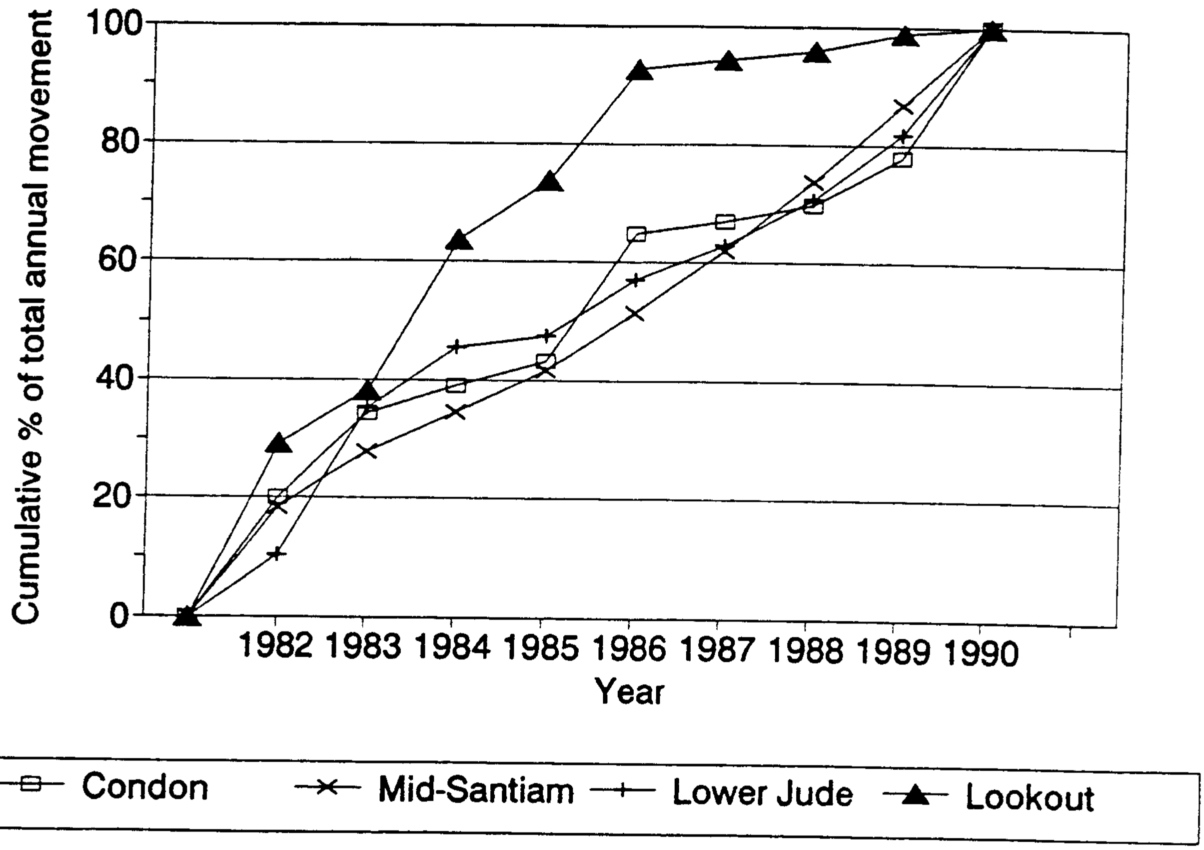

Figure 51 Cumulative percentages of movement of 4 landslides from water year 1982 to 1990. Annual percentage is calculated from the annual movement of a particular year divided by the total movement from 1982 to 1990.

Lienkaemper, per. comm., 1986). Movement rates varied from year to year, ranging from 2.3 to 6.3 m/yr from 1982 to 1990, and averaged 3.8 m (3826 mm) per year (table 4). Compared to other landslides, the Mid-Santiam landslide has had a relatively uniform trend of movement in recent years (fig. 51).

Jude Creek Landslide

The deep-seated, Jude Creek Landslide was monitored by Forestry Sciences Laboratory personnel over the last decade. Most of the landslide is covered by natural forest, and small clearcut patches are found outside the active slide area, but within the Jude Creek watershed. In contrast with the Mid-Santiam slide, the Jude Creek landslide has not been strongly influenced by management activities. Instead, its movement reflects, in part, interaction with Jude Creek, which cuts and modifies the toe of the landslide (Hicks, 1982, George Lienkaemper, 1986 per. comm.).

The Jude Creek Landslide probably has had a long history of activity. A steep, vegetated scarp in the upper part of the slide indicates movement before about 1600 A.D. and subsequent stabilization (Hicks, 1982). Based on observations of movement in the 1980's, the Jude Creek landslide can be divided into two blocks with distinctive movement rates. The upper part of the landslide is relatively stable with annual movement of less than 1200 mm, while the lower block of the slide is moving rapidly, with annual rates sometimes exceeding 6000 mm (table 4). Two reasons were proposed by George Lienkaemper (1986, per. comm.) to explain the comparatively slow movement rate of the upper block: (1) the upper block was better drained because it was underlain by andesite rather than the clay-rich, pyroclastic rocks, that underlie the lower block; (2) the upper block was far from the Jude Creek, which continuously erodes the toe of the lower block, and was separated from the lower block by an active scarp about 4 m high in the middle of the slide.

The timing of movement of the upper and lower blocks of Jude Creek landslide are similar (George Lienkaemper, per. comm., 1986), although their magnitudes are very different. The lower Jude Creek slide block moves rapidly, averaging about 632 mm per month from October through March in recent years (Table 6). Normal daily rates range from 6 mm to 30 mm, and rates over 50 mm per day are not uncommon. Only short movement periods occur in summer months, separated by prolonged periods (2 weeks to more than 2 months) of no movement (fig. 50a). The overall movement pattern of Jude Creek Landslide is similar to the Mid-Santiam Landslide, except that Jude Creek slide stops completely in summer months. Furthermore, the movement rates of Jude Creek

	Condon	Wilhelm	Mid-Santiam	Lower Jude	Lookout Ck.
ANNUAL (WY 81-90)					
Average (mm)	109	34	3840	7880	80
Maximum (mm) (yr.)	221 (86)	95 (86)	6280 (82)	17610 (83)	206 (82)
Minimum (mm) (yr.)	23 (87)	12 (90)	2290 (84)	1420 (85)	8 (90)
MONTHLY					
Avg. no. of month per year w/ movement	4	4	12	10	4
Average rate for WY 1981-90 (mm/mo.)	11	3.2	320	633	6.7
Average winter rates (Oct.-Mar.)	17	6.3	395	745	6.8
Maximum monthly rate (mm)	202 (2/86)	67 (2/86)	987 (3/82)	2452 (2/86)	75 (2/86)
DAILY					
Percentage of days without movement	92	94	22	59	87
Average rate for WY 1981-90 (mm/d)	0.3	0.1	10	22	0.2
Maximum Daily rate (mm)	125	16	294	476	9.4
(Date)	2/23/86	2/23/86	11/24/89	1/1/90	2/23/86
Data periods used for monthly & daily comparisons (WY)	85-90	86-90	82-90	86-88 & 90	86 & 88-90

Table 5 Summary comparison of movement rates among the five selected landslides

Landslide had greater intraannual variation than the Mid-Santiam Landslide, which had a distinctly smooth movement rate curve (fig. 50a).

Lookout Creek Landslide

The Lookout Creek Landslide site has complex mass-movement topography, of which only the lower part is currently active. The slide was first described in detail by Swanson and Swanston (1977), and subsequent researchers have studied the activities and movement mechanisms of the landslide (Mills, 1987, Pyles et al., 1987). By studying split trees and scar tissue in tree rings within the landslide site, Swanson and Swanston (1977) estimated the recent activities had occur since at least 1900 AD. They also analyzed landforms of the Lookout Creek valley, and concluded that the landslide may have begun movement more than three centuries ago.

The monitoring of landslide movement began in 1974 when stake arrays were installed across shear and tension cracks in the active areas. Extensiometers were put on the landslide in 1977 to complement the stake arrays, and records from the most active stake array (#2) are shown in table 4. In 1982, three transverse stake lines were placed across the landslide. Based on two years of measurement from these stake lines, Pyles et. al. (1987) reported that the west margin had moved somewhat faster than the east margin. In 1976, several piezometers were installed on the landslide, and have since then supplied periods of continuous groundwater level records.

The landslide moved an average of 80 mm per year from water year 1981 to 1990 (table 4). The landslide exhibits slow (0.5 to 1.5 mm/d), continuous movement in winter months with faster movement during storms or snowmelts, and no movement at all during the summer (fig. 50a). The maximum daily rate between 10/85 and 11/90 was 9.4 mm (2/23/86). With similar movement patterns, the major difference found between the Lookout Creek slide, and the Jude Creek and Mid-Santiam slides is the magnitude of movement rates. The Lookout Creek slide moves much slower than its western Cascade counterparts (10 cm/yr. vs. 400 to 900 cm/yr.), despite the fact that they occur in the same geologic and climatic regions. The reason behind the difference in movement magnitude among the three slides is not clear, but could possibly relate to the physical configuration of movement masses and mechanical characteristics of the failure materials. Annual movement rate of the landslide declined considerably during the sample period, which has an average of only 13 mm/year, from water year 1987 to 1990 (figs. 50a & 51).

Comparison of Movement Characteristics Among Selected Landslides

Table 5 lists important movement characteristics of the five selected landslides and how they compare with each other in terms of timing, frequency, magnitude, rate, and style of movement. Table 6 compares movement rates among the five selected landslide, in annual, monthly and daily scales. Two groups with different movement ratings are observed: (1) "fast-rates" which includes the Mid-Santiam and Lower Jude Creek landslides; (2) "slow-rates" which includes Condon, Wilhelm and Lookout Creek slides, all with rates 10 to 100 times less than the "fast" group.

Figure 52 shows the frequency distribution of daily movement of the five selected landslides. Two distinctive sets of movement rates are observed: (1) the Mid-Santiam and Lower Jude Creek landslides have only 22 % and 59 % of days with movement less than 1 mm, and most of their movement days have rates between 8 and 50 mm. (2) the Condon, Wilhelm, and Lookout Creek landslides all have more than 90 % of days with movement less than 1 mm, and their movement rates seldom exceed 20 mm per day.

Discussion

An interesting note is that maximum daily movements of

Landslide	Timing	Movement Patterns Frequency	Duration	Style of Movement
Condon	moves only from October to April Storm response only	averaged 3-4 major events (> 7.5 mm) per year	large events: 3-9 days minor events: 1-3 days	most large events (63%) are brief and fast-moving, few (37%) are persistent
Wilhelm	moves in Oct. to March only, storm response only	estimated 2-4 events per year	large events estimated 1-5 days	similar to Condon but shorter & smaller events
Mid-Santiam	moves all year except brief periods (1-5 days) in summer and fall	persistent, events not correlated with storms, rates are on seasonal scale	movement spans full year	consistently fast in Oct.-Mar., and slow in Apr.-Sept.
Jude Creek	moves continuously in Oct. to Apr., largely stops in summer	persistent events (> 1 month) in winter, some are storm-related individual (1-5 days) events in dry season	movement could span months in winter, but only days in summer	pattern similar to Mid-Santiam, more variable daily rates, with periods of no movement in between
Lookout Creek	moves in winter months only, storm and sometimes seasonal responses	movement usually begins in late fall (Nov.), persistent events esp. in late winter, few brief events	Rare persistent events averaged 1/2 to 1 month, brief storm events usually last 1 - 5 days	similar to Condon, with longer events and smaller daily rates. Events shortened in recent years

Table 6 Comparison of movement characteristics among the five selected landslides

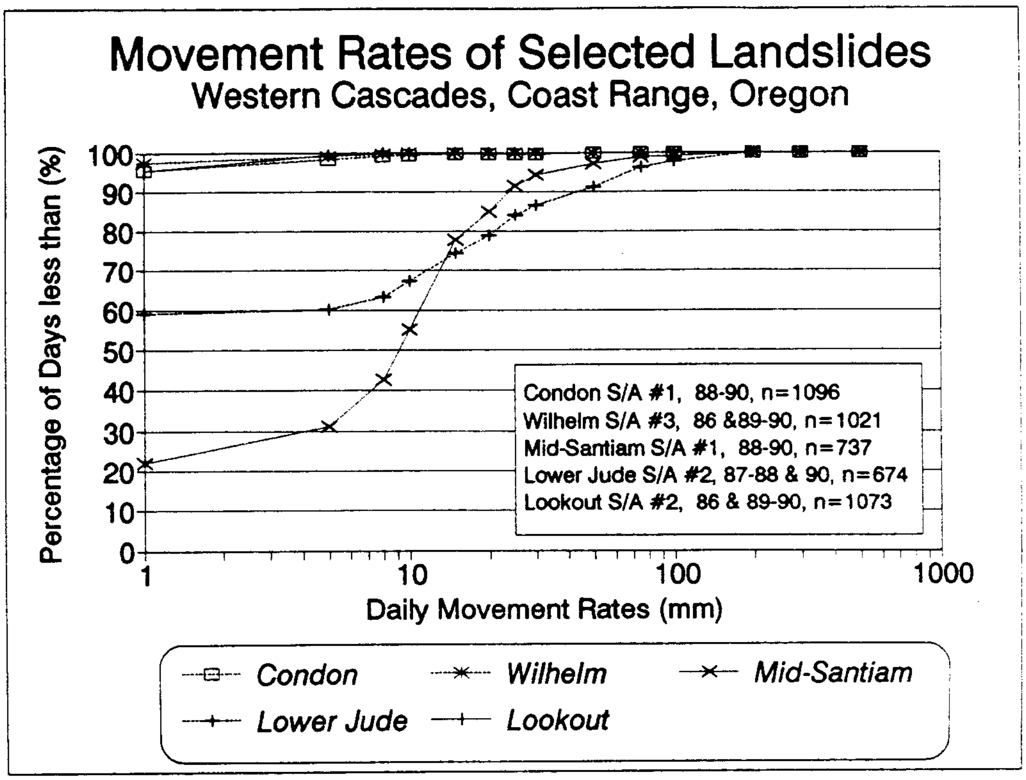

Figure 52a Cumulative distribution of daily movement magnitude of the five selected landslides. Length of records used here is 3 years, but different years were chosen for different landslides due to missing and unreliable data periods. This also explains the different number of days used in the five landslide records.

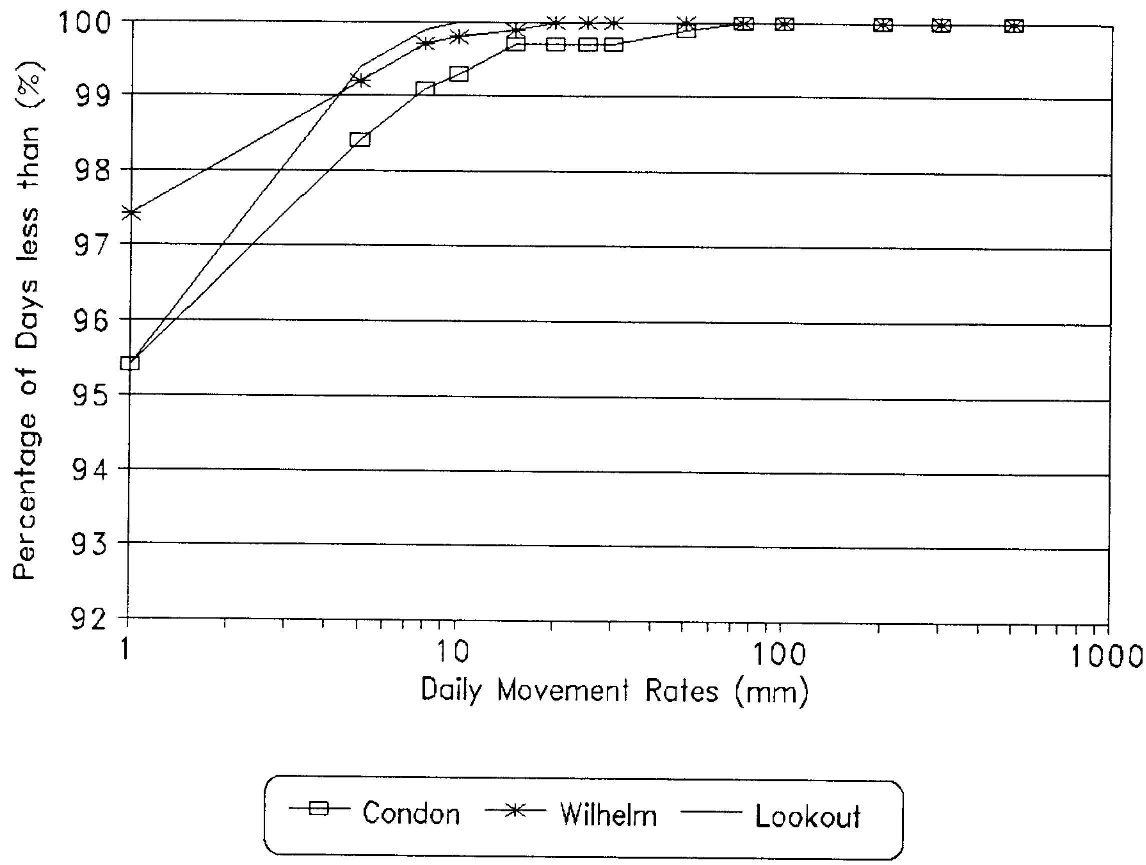

Figure 52b Cumulative distribution of daily movement magnitude of Condon, Wilhelm and Lookout Creek Landslides. This figure is the upper part of figure 52a, using different scale to show the details of the cumulative daily movement plots of the three landslides.

the Condon, Wilhelm and Lookout Creek Landslides were all on February 23, 1986 (Table 6). They also had their highest monthly movement in February, 1986. The precipitation intensity in that event had a 24-hr. return period of 4 years at several Coast Range sampling stations (Swanson and Roach, 1986). On that day Jude Creek landslide had an unusually high movement rate (387 mm), and it also had its highest monthly movement at February, 1986. The daily movement data from the Mid-Santiam landslide was lost during that storm event; however, the stake array recorded only 220 mm of movement during the entire storm period.

One proposed explanation is that Condon, Wilhelm, and possibly Lookout landslides are primarily precipitation (water input) driven, and they respond rapidly to individual large winter storms. The two Coast Range landslides are composed of rigid sandstone blocks sliding on weak siltstone-mudstone shear zones, brief, rapid movement may open macro pores, thereby enhancing drainage, and possibly serving as negative feedback to movement. Relatively long periods of no movement after major movement events (e.g. 2/86) as well as no movement in summer months at Condon Landslide may also reflect this hypothetical mechanism (fig. 21).

The Mid-Santiam and Jude Creek Landslide have more persistent movement in the winter, and they respond most strongly to precipitation at the seasonal scale. Their daily movement rates are dictated by effects of long term water inputs and the hydrologic systems of the entire slide. Therefore, their daily movement rates do not depend highly on individual events, possibly because of very low transmissivity of the clay-rich landslide materials. The clay-rich materials in these two landslides do not appear to operate with the negative feedback movement-drainage mechanism discussed for the other three landslides.

Landslide Movement in Dry Season

Among the three selected landslides in western Cascades, the Mid-Santiam Landslide is the only one that always moves in the summer. Although the Jude Creek Landslide moves occasionally in brief periods (1 to 5 days) in summer months, the movement rates are modest compared to winter movement. The highly disrupted nature of the Mid-Santiam landslide may contribute to the fact that it moves during summer dry months. This is because the thick slide mass and its highly deformable material is highly cracked in the surface, but cracks may close up just below the surface, like crevasses in a glacier. Therefore, drainage is poor within the slide mass, and moisture is easily retained in the slide mass for long periods of time. Moisture retention within the slide mass could be aided by the lack of vegetation cover at Mid-Santiam Landslide. The forest stand on the Jude Creek and Lookout Creek Landslides may reduce groundwater levels in the summer, although hydrologic modelling of evapotranspiration rates on forested and clearcut hillslope in the Coast Range suggest that this effect is not profound.

As the movement rates and patterns of the Jude Creek Landslide are related to the erosion of its toe by Jude Creek, and summer low streamflow has little effect on landslide stability, the Jude Creek slide is relatively immobile in the summer.

Neither Condon nor Wilhelm Landslide moved in the summer. Their movement records show a strong relation between movement and precipitation events (fig. 21). Antecedent moisture conditions are important to movement of Condon Landslide. This is discussed in the negative feedback mechanism discussed earlier in this section.

Effects of landslide runout areas on movement patterns

Differences in the relation of landslides to streams and topographic features of landslide runout areas may influence movement patterns. Mid-Santiam Landslide moves over a relatively flat river terrace, while the toe of the

Jude Creek Landslide goes into Jude Creek, and is constantly eroded and modified by this third-order stream. The toe materials of the Mid-Santiam Landslide spread out freely on the terrace by slumping and debris flows, without encountering obstructions.

The erosion of the toe by Jude Creek might be the major factor causing the large variation of movement rates observed at the Jude Creek Landslide. The different amounts and timing of flow of Jude Creek causes variation in erosion of landslide materials at the toe. The loss of toe materials is due to two mechanisms: (1) stream erosion in which materials are transported away by streamflow; and (2) debris flows originating from the toe area transport landslide materials down the main channel of Jude Creek. The rates and durations of removal of the toe materials might have sent pressure waves upward to the tensional upper part of the slide, and induced movement pulses with different movement rates at the lower part of Jude Creek Landslide, while the whole active slide block tried to re-established the mass balance disturbed by the removal of toe materials. This process has been described by Iverson (1986) for the Minor Creek Landslide in northern California.

Slowing down of Lookout Creek Landslide

During the past decade until 1987, the Lookout Creek

Landslide had been moving persistently throughout winter months, with annual rates averaging 123 mm (Table 4). However, from 1987 onward, the annual movement of the landslide decreased drastically (to 13 mm/yr.). Mid-Santiam and Jude Creek Landslides maintained annual movement rates during the period 1987-1990 similar to rates earlier in the decade (Table 4, and fig. 51). The duration of annual movement of the Lookout Creek slide has been shortened as well (fig. 50a).

Reasons leading to slowing of Lookout Creek Landslide are not clear, but it may be related to the interaction between Lookout Creek and the slide mass. Swanson and Swanston (1977) noted the impacts on Lookout Creek from the landslide, and the feedback mechanism from the removal of toe materials by stream erosion that influence landslide movement. The severe storms of the winter of 1964-1965 produced large-scale streamside debris avalanches and slides that could affect the stability of the Lookout Creek Landslides for decades. In the past few years, however, without extreme storm events, large-scale streamside failures have not occurred since the 1964-1965 winter. Consequently progressive slide movement may have lead to a buttressing effect at the landslide toe. Long-term monitoring continues with the intent of examining effects of the next major episode of toe erosion.

Summary

Movement rates vary substantially over each landslide area. In most cases there are few observation points on any one landslide feature, and each landslide sampled and discussed in this thesis has both slower and faster components than the sites for which we have movement data. However, I believe that the movement data described here reflect the timing and relative rates of movement of these landslides in manners useful for characterizing and analyzing the frequency, duration, and rates of movement in terms of geological and hydrological controls.

Apart from explanations given above, differences of movement characteristics at selected landslides in central Coast Range and western Cascades can be interpreted in terms of climate, geology, and mechanical properties of landslide materials. These parameters will be covered in the following chapter.

XIII. Physical and Mechanical Factors Influencing Movement Characteristics of Landslides in the central Coast Range and western Cascade Range

Introduction

The movement rate, style, magnitude and timing of individual landslides are unique, and this uniqueness is the result of many internal and external factors that control the activity of the particular landslides. Some of these factors include geology, soil, climatic regimes, groundwater condition, and human activities that affect hydrology or mass distribution. Although no two landslides have an identical movement record, landslides tend to behave similarly if they lie within similar physical dominions, like the same climatic regime and geologic terrane; this phenomenon has been probed by landslide researchers. Drawing on work by Japanese landslide researchers, Swanson and Lienkaemper (1985) have explored a semi-quantitative approach to classify the differences of mass movement occurrence, frequency, rate, and volume of disturbed materials in terms of geologic zoning within western Oregon. Their study showed that different geologic terranes exhibit different mass movement patterns. Keefer and Johnson (1983) have compared indexes of soil properties from different landslides, and confirmed the effects of physical properties of soil on the movement characteristics of different landslides. This chapter describes the physical and mechanical factors affecting the five selected landslides, and compares the effectiveness of these factors on controlling the magnitude and timing of movement events on both events and seasonal time scales.

Climatic Differences

The central parts of western Cascades and Coast Range are situated at similar latitude, and are characterized by wet cool winter, and dry warm summer (see also Chapter 7). Precipitation records from Mapleton (central Coast Range) and H. J. Andrews Experimental Forest (western Cascades) show the two regions experience similar precipitation amounts and seasonality (Fig. 53), although the storm magnitudes and the role of snow varies between the two locations. From water year 1985 to 1989, average annual precipitation received by the Mapleton gage was 2292 mm, and the Andrews gage was 2080 mm.

Effects of snow

Another important climatic factor is snow which may accumulate and then melt, possibly initiating or accelerating movement on the time scale of individual storm events, or throughout the spring period of melting of a seasonal snowpack. Melting of thick snowpacks in late

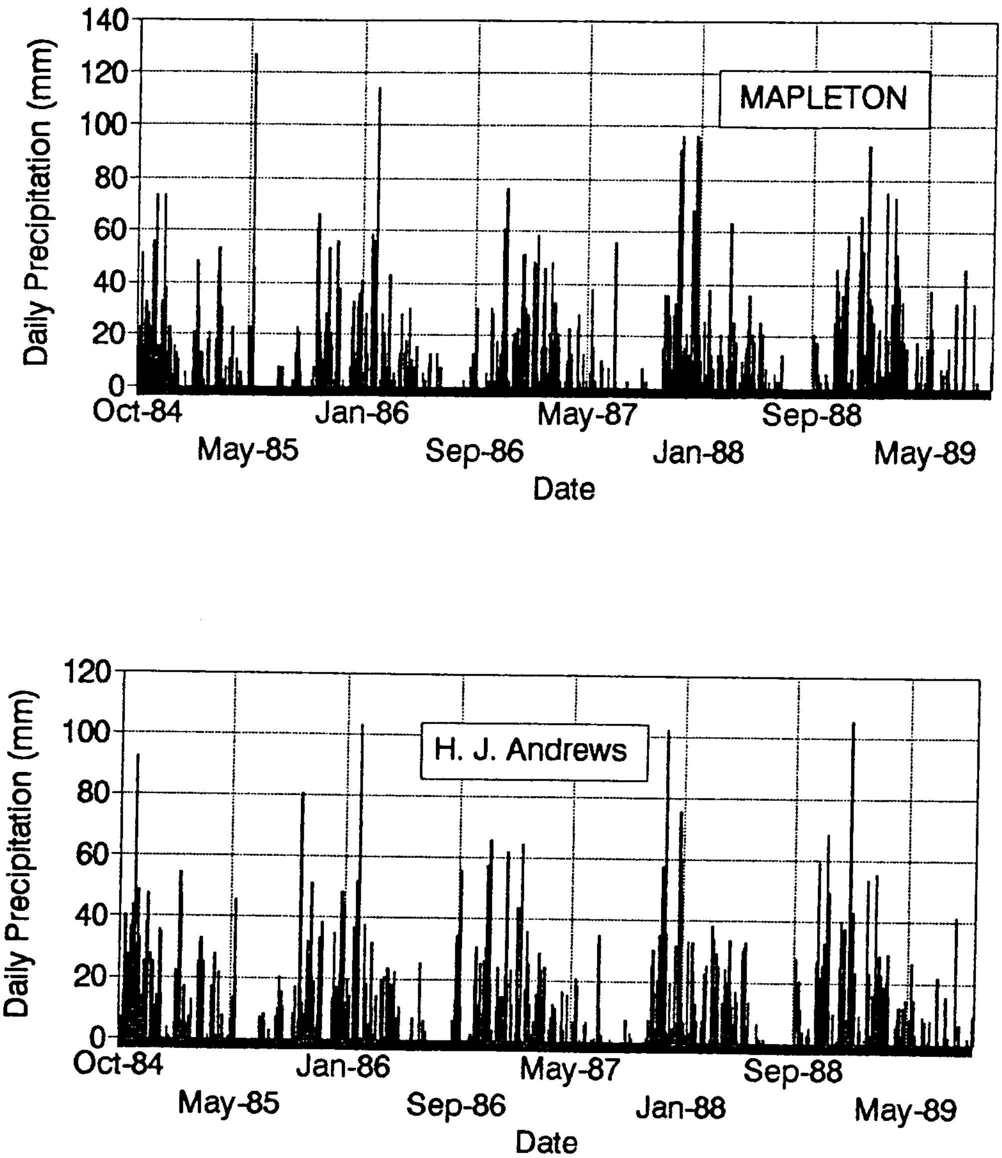

Figure 53 Daily precipitation records of the Mapleton and H. J. Andrews raingages, 10/84 to 9/89.

winter-early spring is a common phenomenon only at higher elevations of the Cascade Range. At mid-elevation zone (350 - 1100 m), thinner (typically <0.5 m), transient snowpacks are common. Melting of these snowpacks during warm rain events produces floods and mass movement activity (Harr, 1981). Snowmelt caused by individual storms events in this so-called "rain-on-snow" zone are common in December through February and typically occur several times each year. The interior of these snowpacks generally remains at or near 0°C, and it requires little energy (such as from rain) to start melting (Harr, 1986). During warm rainy periods, these transient snowpacks can melt very quickly, and supply additional water to the hydrologic system in short time spans. Harr (1981) estimated that in most rain-on-snow events, snowmelt can account for 10 to 25 % of water input to the hydrologic systems, depending on various climatic and physical conditions. Based on studies conducted by Fred Swanson, Harr (1981) concluded that 85% of all small, rapid slope failures documented in the H.J. Andrews Experimental Forest were triggered during rain-on-snow events.

The central part of the Coast Range only receives small, infrequent snowfalls, and snowpacks usually melt within days after the storm. Snowmelt in late winter and rain-on-snow effects are of very limited extent in the Oregon Coast Range. Therefore, movement patterns of the

Condon and Wilhelm Landslides presumably are not affected significantly by snow hydrology.

All three Cascade Landslides are located at the upper end of the transient snow zone, where rain-on-snow events cause rapid inputs of water into soil (Harr, 1981, 1986). The sudden increase in pore water pressure by rapid water input could reduce the effective strength of landslide materials, and could initiate or accelerate the movement of landslides. The rain-on-snow effect is especially pronounced in clearcut and open areas, such as the landslide-disturbed hillslopes, because thicker snowpacks are formed on unforested lands, without interception by forest stands (Harr, 1981). Moreover, snowpacks in unforested areas melt more rapidly during rain-on-snow events, because they receive higher energy transfer from the warm, moist airmass, and more latent heat exchange due to their exposed conditions (Harr, 1981).

Probable effects of rain-on-snow storm events om movement of thick, persistently moving landslides in the Cascades appears to be variable. The timing of movement of the Mid-Santiam and Jude Landslides depend on the seasonal rather than storm-event supply of moisture. In this case, amounts of precipitation, and general size and melt rate of the seasonal snowpacks may be more important than individual storm events on controlling movement. The Lookout Creek Landslide, on the other hand, has had a mainly storm-response characteristics over the 15-year period of observation, although movement rate has changed. In the late 1970's and early 1980's, movement involved both storm and seasonal components, but in the late 1980's movement became more exclusively storm-event related. The movement records of the Lookout Creek Landslide since 1987 have been characterized by short movement events that lasted from 2 to 5 days. These individual events occurred during rain-on-snow events. The movement record of the Lookout Creek Landslide also shows increased activities from February to April in recent years, sometimes accounting for more than 50% of the total annual movement; but whether this activity was caused by seasonal snowmelt is unclear. Although Harr (1981) and F. J. Swanson (per. comm., 1991) have noted the possible effects of snowmelt and rain-on-snow effects on Landslide activity, no quantitative assessment has tested this hypothesis.

Geology

Relationships between bedrock geology and mass movement activity have long been established by numerous scientists (e.g. Swanson and Lienkaemper, 1985). The general geology of the two study areas is discussed in Chapters 3 and 4, and Chapter 5 gives more detailed information on the regional and on-site geology of the central Coast Range, where Condon and Wilhelm Landslides are located. Table 7 summarizes the geology of five selected landslides, and geologic features that influence the geomorphology and movement patterns of the landslides. The following are summary notes on particular geologic properties that shape the distinctive movement patterns found in the central Coast Range and western Cascade Range.

Central Coast Range

A major portion of the central Coast Range is underlain by the Tyee-Flournoy formations, in which massive competent sandstone beds interlayered with less resistant thin (0.1 to 0.3 m) siltstone-mudstone beds. These weak and less permeable thin beds could become slipping surfaces (shear zones) at landslide bases. The gently dipping Tyee-Flournoy strata correspond with hillslopes in many locations. These dip slopes are usual sites for deep-seated landslides, where translatory "block slides" are common, with large bedrock blocks detached and moved in short and rapid rates. The Condon and Drift Creek Landslides are typical examples.

The contacts between early Eocene Tyee-Flournoy formations and late Eocene Yachats Basalt always contain weakened (metamorphosed) rocks and abundant clay, thus are

	Condon	Wilhelm	Mid-Santiam and Jude Creek	Lookout Ck.
Geologic Units	Tyee-Flournoy	Tyee-Flournoy Yachats Basalt	Little Butte Volcanic Series	Sardine Formation Quaternary glacial deposits
Age	Early Eocene	Early Eocene Late Eocene	Oligocene to early Miocene	Middle to late Miocene
Rock types	Sandstone Siltstone	Sandstone Siltstone Basalt	Andesite to basalt, volcanic & tuff breccia, crystal-lithic tuff tuffaceous sandstone and altered pyroclastic materials	Quaternary glacial debris from volcanic and pyro-clastics materials
Alteration	None	Minor alterat-ion on contact with the Yachats basalt	widespread zeolitic-grade alteration especially in pyroclastic materials, with zeolite and montmorillonite	minor extent in the study area
Geologic Structures	Bedrock dips gently along slope	Contact between Tyee & Yachats units within landslide	Assortment of thinly bedded pyroclastic materials interbedded with jointed andesitic to basaltic flows (avg. 15 m thick)	lava flows and pyroclastics rocks overlain by unconsolidated glacial deposits

Table 7 Summary of geology of the central Coast Range, Mid-Santiam drainage basin and the H. J. Andrews Experimental Forest. Major references: Chan and Dott (1983), Graham (1985), Baldwin (1974), Swanson and James (1975), Peck et al, 1964, and Sherrod and Smith 1989.

sites of hillslope failure. The Wilhelm Landslide is an example of this type of landslide occurrence.

Clay contents of the Tyee turbidite sandstone are low due to the provenance of the sediment and low degree of diagenesis (Chan, 1985). Weathered materials derived from Tyee rock units generally have low plasticity. Therefore, slow-moving landslides with highly deformed bodies are not common in the central Coast Range.

Western Cascade Range

Unlike the regional uniformity of one or two rock types present throughout the central Coast Range, the western Cascades have different rock units present in a local scale. Contacts between different rock types with contrasting strength are commonly sites susceptible to deep-seated slope failure. This is especially true in zones between competent and weaker rock types, such as between lava flows and pyroclastic rocks.

The pyroclastic rocks and tuffs are less resistant to weathering, and decompose easily to clay-rich soil. Many hillslope failures in the Mid-Santiam basin are associated with pyroclastic rocks. Joint block failure is common in the more competent andesitic flows, forming rockfalls at steep slopes, cliffs, and large headscarps of deep-seated landslides, like the Middle-Santiam Landslides.

Widespread hydrothermal alteration of the volcanic rocks resulted in the formation of clay minerals and amorphous materials. The presence of abundant clays makes soil more plastic, and could explain the slow deformation of landslide materials and the persistent movement pattern observed in the Mid-Santiam and Jude Landslides.

Conclusion

Geology controls physical and mechanical properties of landslide materials. Therefore, geology is an appropriate element to be used to classify landslide movement patterns on a regional scale, with a qualitative approach.

Landslide Materials Properties

Large, deep-seated landslides typically occur on slopes with thick soil mantle (colluvium), and/or highly decomposed bedrock. The stability of the landslides is affected by the strength and permeability of the landslide colluvium, which are related to soil moisture content, soil texture and clay mineralogy. The properties of landslide materials at the Mid-Santiam and Lookout Creek Landslides were studied by Hicks (1982) and Mills (1983), and the clay mineralogy of the soils found in the western Cascades were studied by Taskey et al (1978). Schroeder and Alto (1983) analyzed soil samples from the Mapleton area, where the Condon and Wilhelm Landslides are located. No definitive failure-plane-samples were discussed by those researchers, due to the difficulty in locating the failure planes in the field. Table 8 lists important soil properties from those two study areas. Atterberg limits of samples from the two study area plus one area investigated by Keefer and Johnson (1983), and soil classification diagram are shown in Figure 54.

Discussion

The five selected landslides occur in various soil types, with diverse physical and engineering properties (Table 8). The Mid-Santiam and Lookout Creek Landslides share some properties (e.g. ranges of Atterberg Limits), because they lie within the same geologic terrane. Compared to its western Cascade counterparts, the Mapleton area has distinctive properties. Although soils from all the landslides contain mostly silt and clay, the Mapleton samples (Condon and Wilhelm Landslides) have higher silt contents (all MH and ML soils, Fig. 54) than the two sample areas in the western Cascades. The Mapleton samples have low plasticity indices (3 to 15) (Fig. 54), compared to other sample areas (7 to 30). Therefore, landslide materials found at the two Coast Range landslides have

Properties (Averaged)	Mapleton[a]	Mid-Santiam[b,c]	Lookout[d]
Bedrock Geology	Sandstone	volcaniclastic	volcanics
Natural Water Content	28.0	25 - 45	53.8
Liquid Limit	48.8	44.8	65.8
Plasticity Index	8.7	15.6	17.6
Liquidity Index	-3.2 - -0.39	(0.37)*	-0.16 - 0.65
Activity Index	n/a	0.9 - 5.7	0.22 - 1.25
Unified Class.	MH - ML	MH, ML, CH CL, SM, SC	MH - ML
Clay Mineralogy	chlorite	montmoril-lonite, kaolinite	halloysite montmoril-lonite, chlorite (kaolinite)
Engineering Properties			
Cohesion c' (kPa)	0 - 6.9	n/a	12.4 - 34.5
c' (residual)	n/a	n/a	0 - 14.4
Friction angle	35.3 - 41.1	n/a	23.2 - 31.0
(residual)	n/a	n/a	27.8 - 35.0

Table 8 Comparison of engineering properties of soils from Mapleton, Mid-Santiam and Lookout Creek landslide areas. Single values are averaged values. Data from Schroeder and Alto (1983)[a], Hicks (1982)[b], Taskey (1978)[c], and Mills (1983)[d]. * = Estimated values.

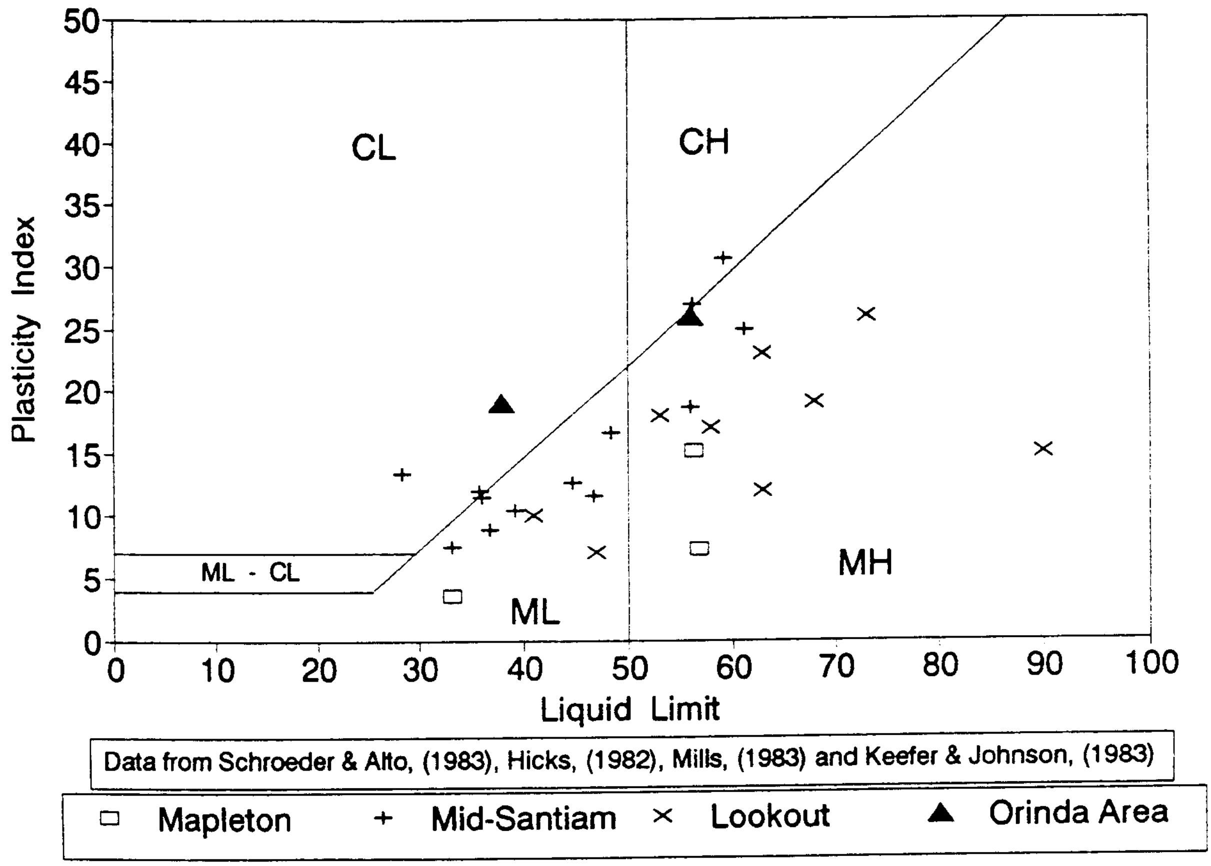

Figure 54 Soil classification chart for soil data from various landslides. Data from Hicks (1982), Schroeder and Alto (1983), Mills (1983), and Keefer and Johnson, 1983.

greater tendency to behave non-plastically under most physical conditions. The negative Liquidity Indexes from the Mapleton samples (Table 8) further indicate that the soils from that area experience brittle fracture when sheared. These two index properties help explain the general movement patterns of Condon and Wilhelm Landslides, where movement events in brief periods between long periods of relatively stable conditions.

The scattered data points of samples from the Mid-Santiam and Lookout Creek areas signify the diverse soil and rock types (Fig. 54). The widely variable index properties of these soils can also be caused by the complex nature of the deep colluvial materials, which range from soils weathered from bedrock, colluvium, to glacial tills (Swanson and James, 1975). Nevertheless, most of the samples are grouped right below the A-Line, which indicates they are not very plastic (soils above A-Line are classified as plastic). Furthermore, the wide range of plasticity indexes and the positive values (between 0 and 1) of the liquidity indexes (Fig. 54 and Table 8) indicate that soil materials of landslides in parts of the western Cascades could become plastic when soil water content raises above certain threshold values, or when the materials are sheared. This could be the case of the Lookout Creek Landslide, where a shear zone about 25 cm thick was observed 6.6 m below the surface of the slide in a test well (Mills, 1983). The plastic behavior of the landslide materials under some physical conditions, such as high water content could account for prolonged movement periods of the three western Cascades Landslides in wet periods.

In Figure 54, data from the Orinda area in northern California (Keefer and Johnson, 1983) are incorporated for comparison. The Orinda area is underlain by series of marine sandstone, siltstone, mudstone, conglomerate, and minor limestone, lignite, and tuff. The soils from that area are mostly clayey silts, and have high plasticity index, high natural water content, and low strength. When certain pore water pressures were met, the landslides in the area behaved like plastic bodies, moving slowly and continuously over months in wet periods (Keefer and Johnson, 1983). The slides in that area have similar movement patterns to the Jude Creek Landslide, except that their movement rates are more consistent and faster (averaged 32 to 120 mm per day), probably reflecting higher plasticity.

The above discussion reveals the possibility of using plasticity index as a crude guide to classify the general movement patterns of landslides. Figure 54 demonstrates such classification scheme: plasticity indexes of the Mapleton area are lowest, where landslides in that area move in brief periods (minutes to days) with large displacements (catastrophic failure). Landslides in materials with higher plasticity indexes tend to move over longer periods (weeks to months) and with more persistent rates (Lookout Creek, Mid-Santiam, and Jude Creek Landslides). Where plasticity indexes are high (> 20), and soils are plastic, such as slides in the Orinda area, landslides can move constantly and fast as long as soil water content (pore water pressure) remains above certain limits. Since landslides occur under a wide range of physical and mechanical conditions of soils, plasticity index can be used only as a rough predictor of landslide movement patterns found at different soil and geologic types.

Clay content of landslide colluvium is an important factor in slope stability studies. Generally, soils with high clay content are cohesive and behave plastically when water contents are high. Unstable slopes and active mass wasting processes are associated with soils with smectite (sometimes called montmorillonite) (Schulz, 1980) . In the five selected landslides, only materials from the western Cascades landslides contain significant amounts of clays. Montmorillonite, kaolinite (halloysite), and amorphous compounds (gel, imogolite strands, etc.) are common components present in the Mid-Santiam and Lookout Creek areas (Taskey et al, 1978, Hicks, 1982). In the shearing zones within volcaniclastic units, Taskey et. al. (1978) observed very abundant smectite, along with minor zeolite and kaolinite (?).

Although clay content and mineralogy alone cannot be used as reliable guidelines on slope stability analyses due to large variations of physical and mechanical properties under natural conditions, they can be used to support the plasticity index classification scheme. High clay contents (30 to 44 %) of the soils from the Orinda area, California, define the high plasticity indexes of the soil materials, and the relatively long movement periods of the landslides there. The low clay contents (low plasticity index) and non-cohesive behavior of the soil materials in the Mapleton area (Schroeder and Alto, 1983) are associated with the brief-large displacement landslide movement pattern which is typical in that geological terrane.

Table 8 shows two engineering properties of soils samples from the two study regions: cohesion and friction angle of soil samples. Strength of the soils from these landslide sites is not compared here, due to the large variations of the strength parameters of the soil samples under different physical conditions, even within a single landslide body, as well as differences on sample handling and testing methods employed by researchers. The low cohesion of soils from the Mapleton area is probably due to low clay content of soils derived from the Tyee-Flournoy sandstones. Cohesionless materials favor brittle fracturing of soil masses, and therefore catastrophic failure of slopes. The relatively high cohesion of soils from the Lookout Creek area is identified with plastic behavior of the soils having high water content, and is related to prolonged periods of progressive slope failure with slow displacement rates. The internal friction angles of soils from the Mapleton area are higher than in samples from the Lookout Creek Landslide. However, these values are immaterial, as progressive hillslope failures in deep-seated landslides (earthflows) could reduce the friction angle of the failure mass to residual ranges, which are well below the original friction angle (approximately 16° in the Perkins Peak area, central Coast Range, reported by Pierson, 1977).

Summary

Climatic and geologic (soil) characteristics are the two most important factors that control the landslide movement patterns. Regimes of movement pattern of the deep-seated landslides can be classified based on geologic terranes and associated surficial-deposit, and theirg eotechnical properties reflected in various indices. Climatic patterns are significant modifying factors of the timing and rate of movement. Such a qualitative approach is demonstrated well in this study in comparing the deep-seated landslides of the western Cascade and the central Coast Range.

XIV. Conclusions

(1) The Condon Landslide is characterized by brief (1 - 9 days) movement episodes with wide range of daily movement rates (from < 1 mm/d to > 120 mm/d) with most of the daily movements less than 5 mm; and are separated by long periods of no movement. All major events were triggered by large winter storm events. Annual movement ranged from 23 mm in 1987 to 221 mm in 1986, and averaged 106 mm per water year from 1982 to 1990. Similar movement pattern is observed at the nearby Wilhelm Landslide, except the magnitudes of movement events were much smaller (typically < 3 mm). From 1986 to 1989, the Wilhelm Landslide averaged 38 mm of annual movement.

(2) The movement pattern of the Condon Landslide is controlled by its geology, in which the Tyee-Flournoy strata control the blockslide configuration of the slide. Rigid sandstone blocks slide on a weak siltstone-mudstone shear zone along the dip slope. Currently only the northwestern part of the landslide is active, which was clearcut in 1987.

(3) The groundwater level within the Condon Landslide responds to precipitation events very quickly, with a lag time usually less than 2 days. The quick response is

possibly due to the thickness (approximately 5 to 8 m) of the slide body, low plasticity of the landslide materials, and the expected presence of macropores and crack systems within the landslide.

(4) Higher rates of daily movement occurred at higher groundwater levels. Most days with movement (90%) occurred when groundwater level was within 2.5 m of the ground surface, and movement occurred everyday when groundwater level rose to within 1.5 m of the surface.

(5) Clearcutting apparently did not change the movement pattern at Condon Landslide, however, it does appeared to have increased surface mass wasting and shallow slope failures, especially at the western toe of the slide.

(6) Data on precipitation amount alone are not effective in predicting the timing and magnitude of movement at the Condon Landslide.

(7) Antecedent Precipitation Index (API) is a practical basis for predicting the initiation of movement events. For Condon Landslide, an API value of 160 mm functions as a threshold in that movement is very likely above that. From 1985 to 1990 the 160 mm threshold was exceeded 18 times, which resulted in 15 (83 %) movement events. However, API

values were not useful in predicting movement magnitudes.

(8) Landslides in the western Cascade Range move in a more persistent manner, and generally respond to precipitation on a seasonal-scale, contrasting to the studied central Coast Range landslides, which generally respond to individual storm events.

(7) The Mid-Santiam Landslide moved about 3800 mm per year, in 1981 to 1990. It moves all year long, with consistently fast daily rates in wet periods, and slow dry season rates. Annual movement rates were similar from 1981 to 1990.

(8) The Lower Jude Creek Landslide averaged 7800 mm of movement per year from 1982 to 1990, and annual movement varied greatly over the same period. Continuous interaction with Jude Creek probably contributes to the substantial activity observed in the lower block of Jude Creek Landslide. The landslide usually moves consistently throughout winter months, but halts completely in the summer.

(9) During the late 1970's and early 1980's, movement of the Lookout Creek Landslide responded at both storm and seasonal-scales. Starting in 1987, movement became dominantly storm-event related, and the landslide has been

slowing down since then, with annual movement only averaged 13 mm for 1987 to 1990. This may reflect lack of toe erosion since 1965 and some buttressing of the toe area as a result of encroachment into Lookout Creek since then.

(10) Plastic deformation of the landslide masses are common in the western Cascade Range landslides, because of relatively high clay contents found in landslide materials. The clay is derived from hydrothermally altered volcaniclastic rocks and weathering process. These clays contribute to the long and consistent movement in winter months typical of the western Cascades slides.

(11) Geology and climate pattern are the two most important factors dictating movement characteristics, and can be used to classify regimes of deep-seated landslide movement patterns. Geology controls the structure of the landslide and dictates the physical and mechanical properties of landslide materials. Climate controls the timing and magnitude of water availability (storms), which is vital to initiating movement.

XV. Bibliography

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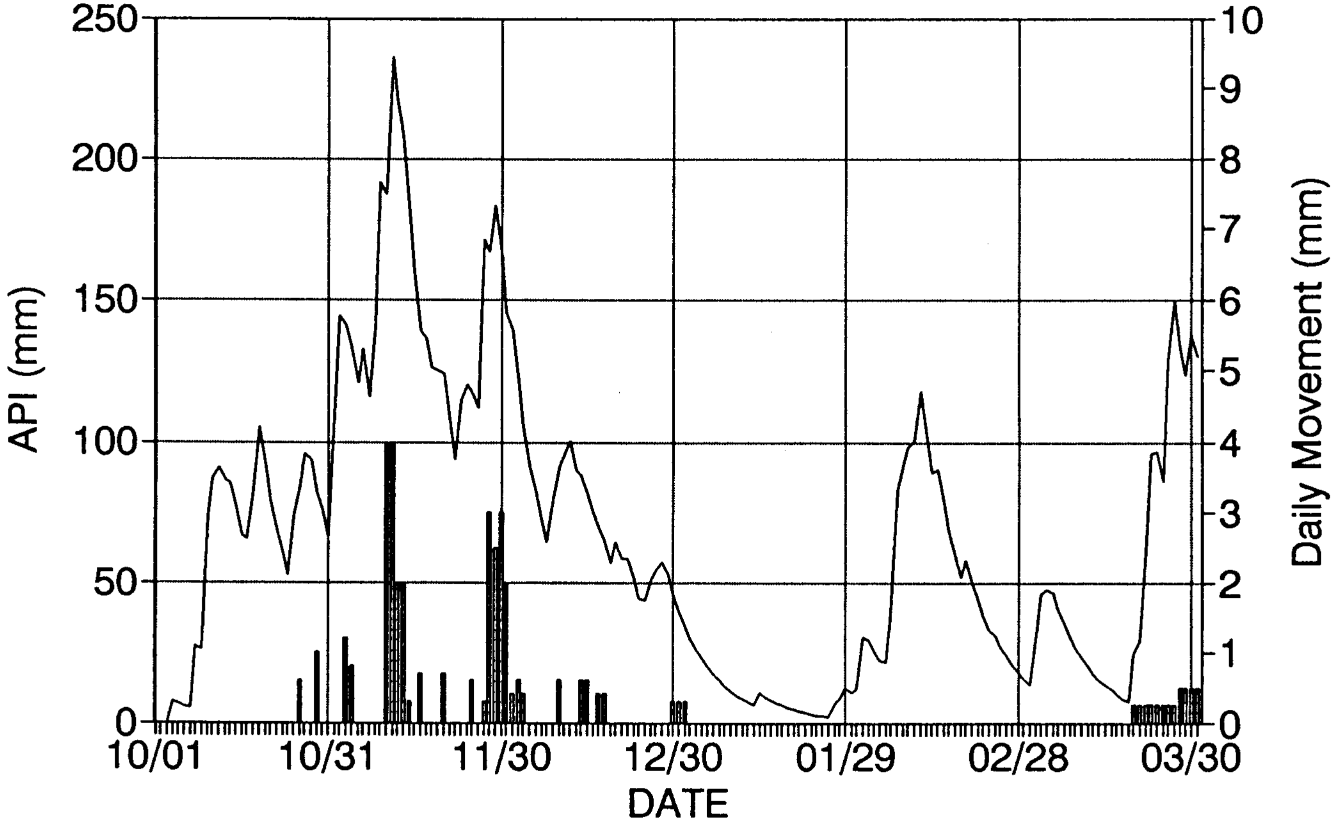
XVI. Appendix

Half-yearly records of API and Movement events

Condon Landslide, water year 1985 to 1990

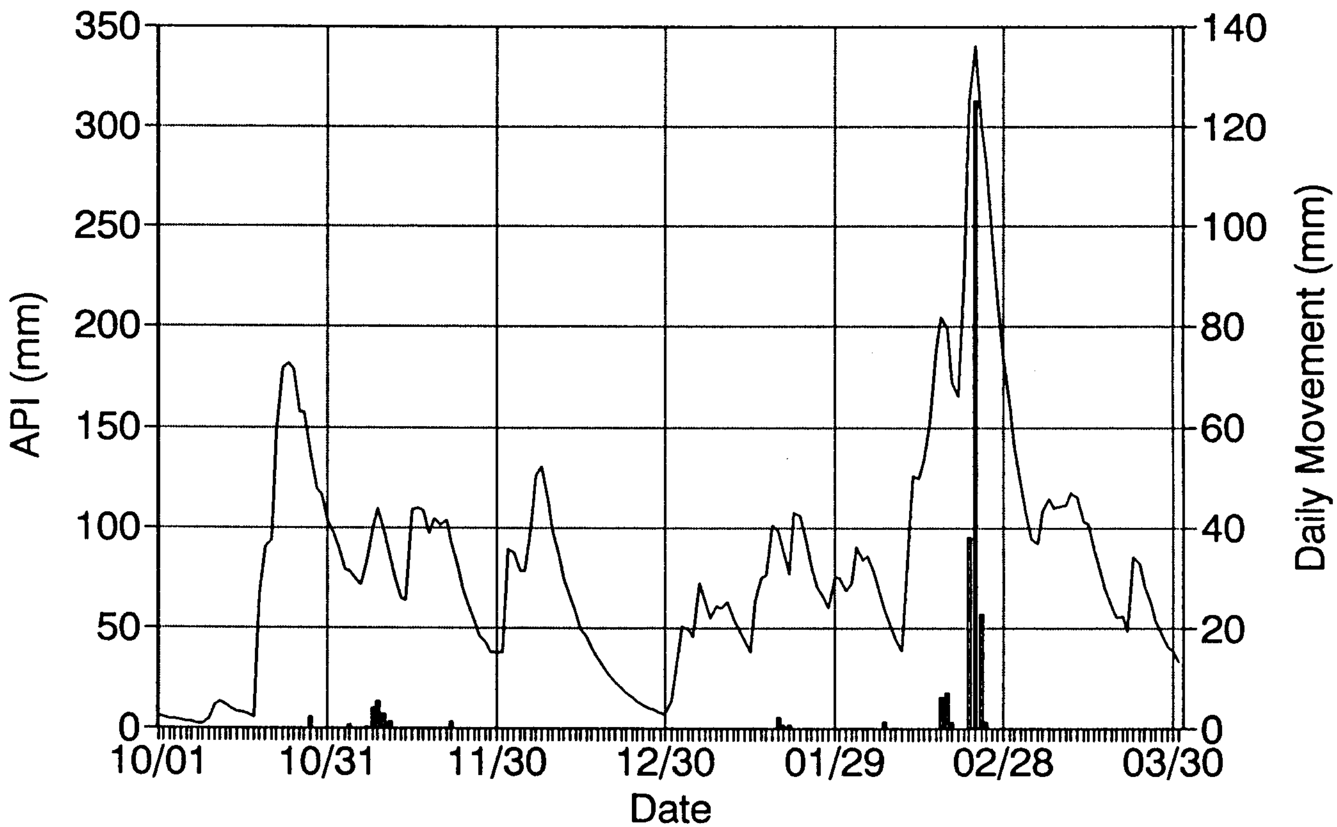
API and Movement of 1985 WY

Condon Landslide, Oregon, Oct.-Mar.



API and Movement of 1986 WY

Condon Landslide, Oregon, 10/85-3/86



API and Movement of 1987 WY
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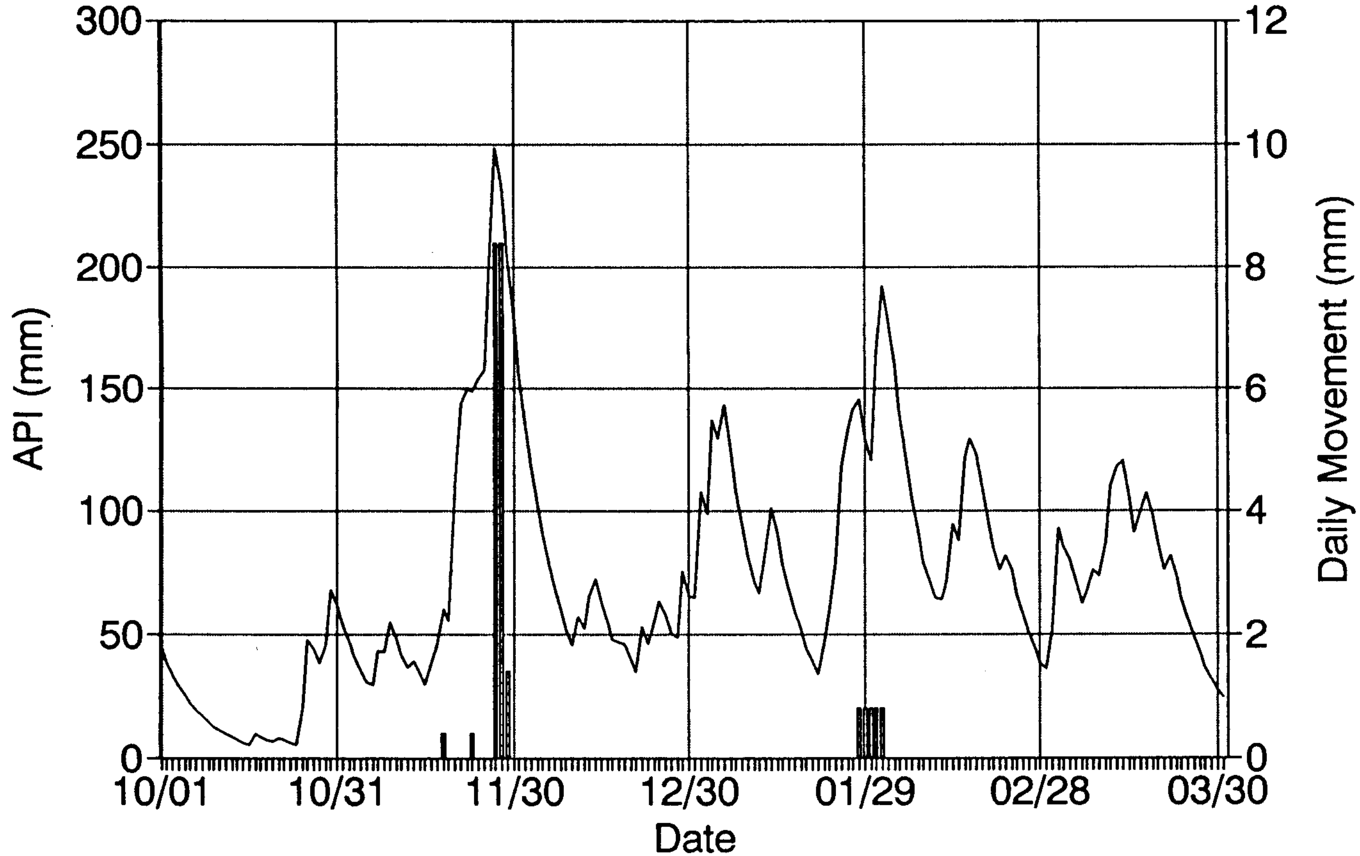

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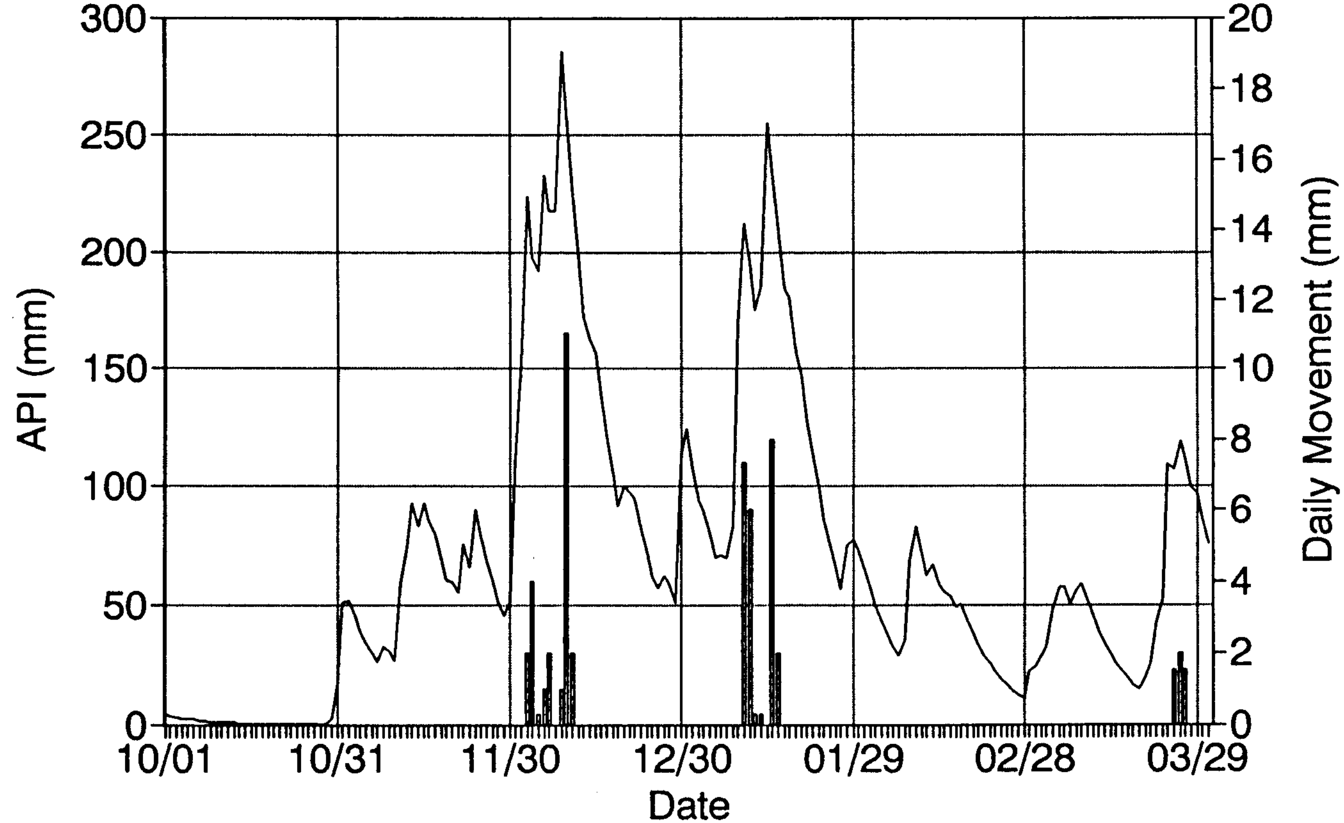

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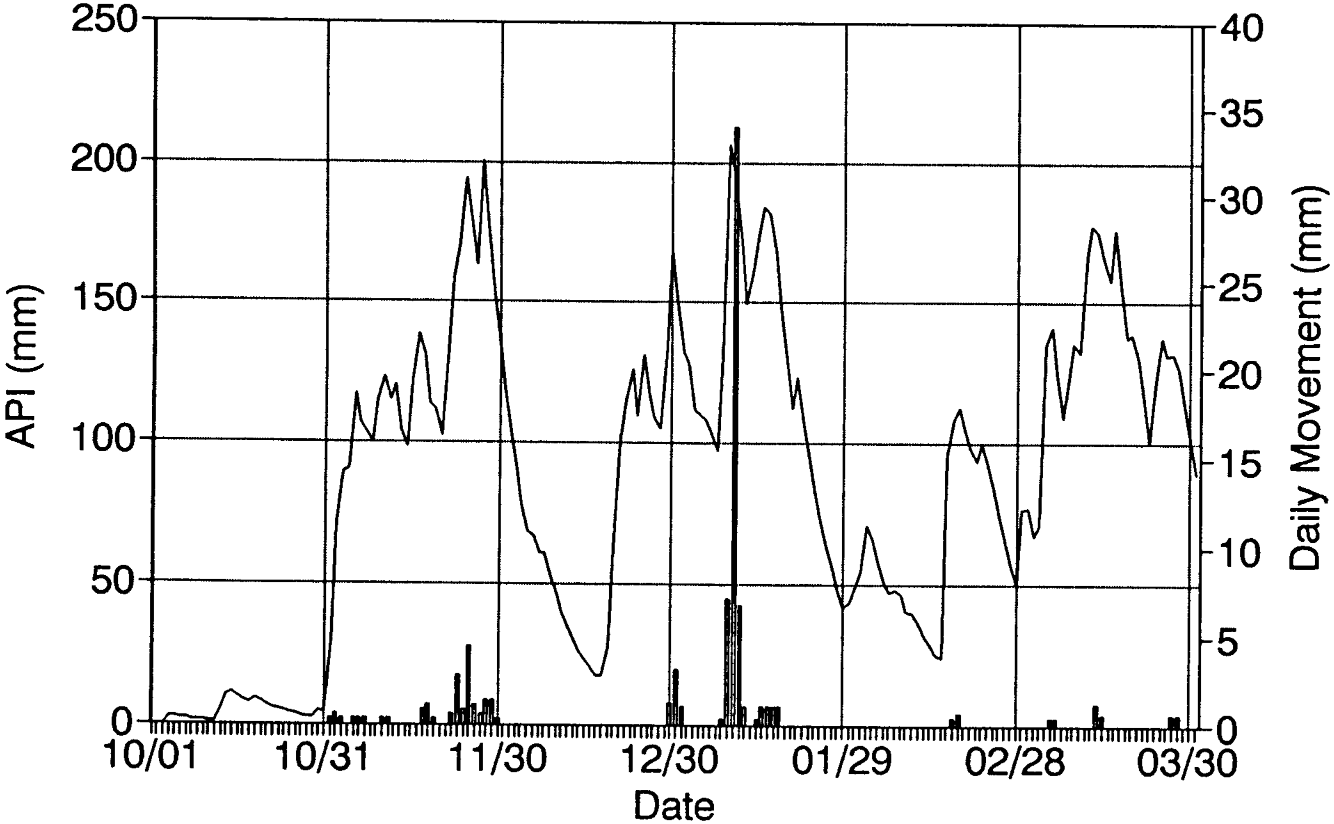

API and Movement of WY 1990
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