# THE LOCATION AND ANALYSIS OF LANDSLIDES ALONG THE LEWIS OVERTHRUST FAULT, GLACIER NATIONAL PARK, MONTANA


By

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## PREFACE

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# CHAPTER I

# INTRODUCTION

## Overview

As development in the mountainous areas of the western United States accelerates, the exposure of humans and structures to natural hazards from high magnitude mass-wasting events increases. Encroachment of human activity into previously uninhabited, undeveloped mountain areas creates situations whereby mass-wasting events threaten human activity.

Glacier National Park, a mountainous Park in northwest Montana, is one such location where the potential is high for high magnitude mass-wasting events. The mountains in this area commonly have steep slopes, high relief, and weathered bedrock features. These features, combined with a mid-latitude climate conducive to extensive frost-weathering action, have resulted in a landscape possessing many deposits attributable to slope movement. Mass-wasting features such as snow avalanches, slumps, rockslides, rockfalls, earthflows, and talus slopes are common along the slopes of the mountains. Rock glaciers are present, although less common (Butler, 1976). Along one area in particular, the Lewis Overthrust fault along the eastern front of the Rockies, extensive mass-wasting deposits are found.

Although landslide deposits are widespread throughout the Rocky Mountains, and indeed, throughout most mountainous regions in the world, little work has been done concerning the landslide deposits in northwestern Montana. Mudge (1965) studied rockslide and rockfall deposits along the Sawtooth Ridge area of Montana, approximately 120 km south of Glacier National Park; and Butler (1979a), as part of a study concerned primarily with the distribution of snow avalanche paths, mapped surficial mass-wasting deposits along part of the Lewis Overthrust in Glacier National Park. No other detailed studies concerning landslides in the Rocky Mountains of northwestern Montana appear in the professional literature.

Recognition that high magnitude landslide hazards exist in the Rocky Mountains of northwestern Montana is evident in the literature. Varnes (1975) noted the presence of large rockfall-debris flows along the front of large thrust sheets of Paleozoic limestone in northwest Montana. Similarly, Radbruch-Hall _et al_. (1981), noted the hazards of large and rapid slope failures in fractured rock in the Northern Rocky Mountains of the U.S. They also noted that information on landslides and their relation to geologic conditions is sparse for parts of the western United States. Butler (1983a) briefly described two high magnitude mass movements - a rockslide avalanche and debris-flow - along the Lewis Overthrust fault in Glacier National Park. Neither deposit was analyzed in detail.

## Objectives

The intent and purpose of this study is essentially threefold: 1) provide an inventory and description of the spatial distribution of landslide activity along the Lewis Overthrust fault in Glacier National Park, Montana. Environmental variables which may be important controlling factors in the landslide distribution will be examined; 2) include a detailed analysis of two high magnitude landslide events along this fault, including the determination of the year of each event, possible trigger mechanisms of each event, and general site characteristics of each deposit. Landslide analysis will involve dendrogeomorphological and relative age-dating techniques, as well as an examination of historical records. Climatological and seismic data will be examined to attempt to postulate the trigger mechanisms of each landslide event; and 3) place the landslide activity from this area into a natural hazards context. The potential impacts of landslide activity along the Lewis Overthrust fault to human activity in this area will be discussed.

## Research Context of This Study

As is common to many research studies completed under the broad discipline of "Geography,", this study incorporates ideas and concepts from several other disciplines. Ives and Bovis (1978) discuss the interdisciplinary nature of natural hazards research, pointing out that one international natural hazards research group is comprised of professional climatologists, engineers, geographers, and geologists. This current research effort is primarily a study in physical geography and geomorphology. This study contains an analysis of mass-wasting features (primarily landslides), describes the accumulative effects of these high magnitude erosive and depositional agents, and maps the landslide deposits along the Lewis Overthrust fault. Climatology and seismology are included in discussions of the potential trigger mechanisms of the two landslide events.

## Justification of Study

White and Haas (1975) concluded that the vulnerability of the United States to natural hazards is being increased by the following factors: 1) a shift in population from county and city to suburban and exurban locations, i.e. to higher risk areas; 2) more people now live in new and unfamiliar environments where they are totally unaware of potential risks; 3) the increasing size of corporations enlarges their capacity to absorb risks, which may result in plants being located in high risk areas. The location of these firms attracts job-seekers and housing developments to the same dangerous locations; 4) the enlargement of new housing starts accounted for by mobile homes means more families are living in dwellings which are easily damaged by natural hazards.

Concerning landslides, White and Haas (1975) believed the most immediate research priority lies in the identification of landslide hazards on a nationwide basis. Initially, this involves no more than the synthesis of existing geomorphic and related information and its correlation with human occupance of the hazard zone. White and Haas recommended that relatively small areas be selected as representative of major landslide hazard regions for studies to identify the distribution and type of existing landslides and their causal mechanisms.

A relative paucity of research information exists concerning landslide activity in the Rocky Mountains of northwest Montana. Radbruch-Hall _et al_. (1981) noted the sparseness of landslide information in parts of the western United States. Butler (1983a) concluded that additional mapping of high magnitude mass movements is needed to expand the natural hazards inventory for the Rockies of Canada and extreme northern United States.

A research study of this type is beneficial to the natural hazards inventory in northwest Montana. Mapping the landslide deposits along the Lewis Overthrust fault, discussing the potential hazards from landslide activity in this area, and attempting to determine the causal mechanisms of the two high magnitude landslides will increase the natural hazards inventory in this area.

# CHAPTER II

# STUDY AREA

## Glacier National Park

Glacier National Park is located in the northwestern corner of Montana (Figure 1). Approximately 4,100 square kilometers comprise the Park; it was established in 1910 (Glacier Natural History Association, 1983). The Park is part of the Waterton-Glacier International Peace Park. Glacier Park is bounded on the east by the Blackfoot Indian Reservation; on the southeast and southwest by the Burlington Northern Railway or U.S. Highway 2, beyond which lies National Forest land; on the west by the North Fork of the Flathead River; and on the north by the U.S.-Canadian international border ($49^{\circ}$ N).

Two major mountain ranges run roughly north-south through Glacier National Park (Figure 2). The Livingston Range extends southward from the Canadian boundary to the McDonald Valley. To the east lies the Lewis Range, which extends from the Canadian boundary south-southeastward through the southern tip of the Park. The Continental Divide initially follows the Livingston Range southward, and then crosses to the Lewis Range at approximately $48^{\circ}$ 50' North. The Divide follows the Lewis Range southward as it exits the Park.

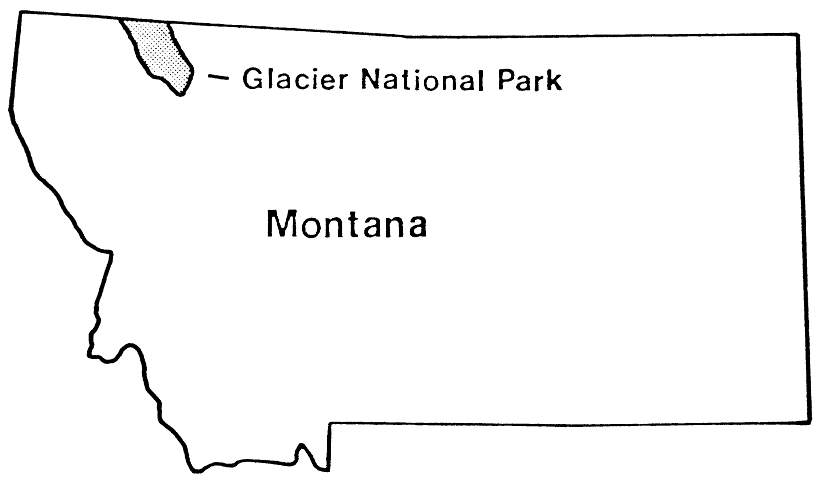


Figure 1. Location of Glacier National Park, Montana.

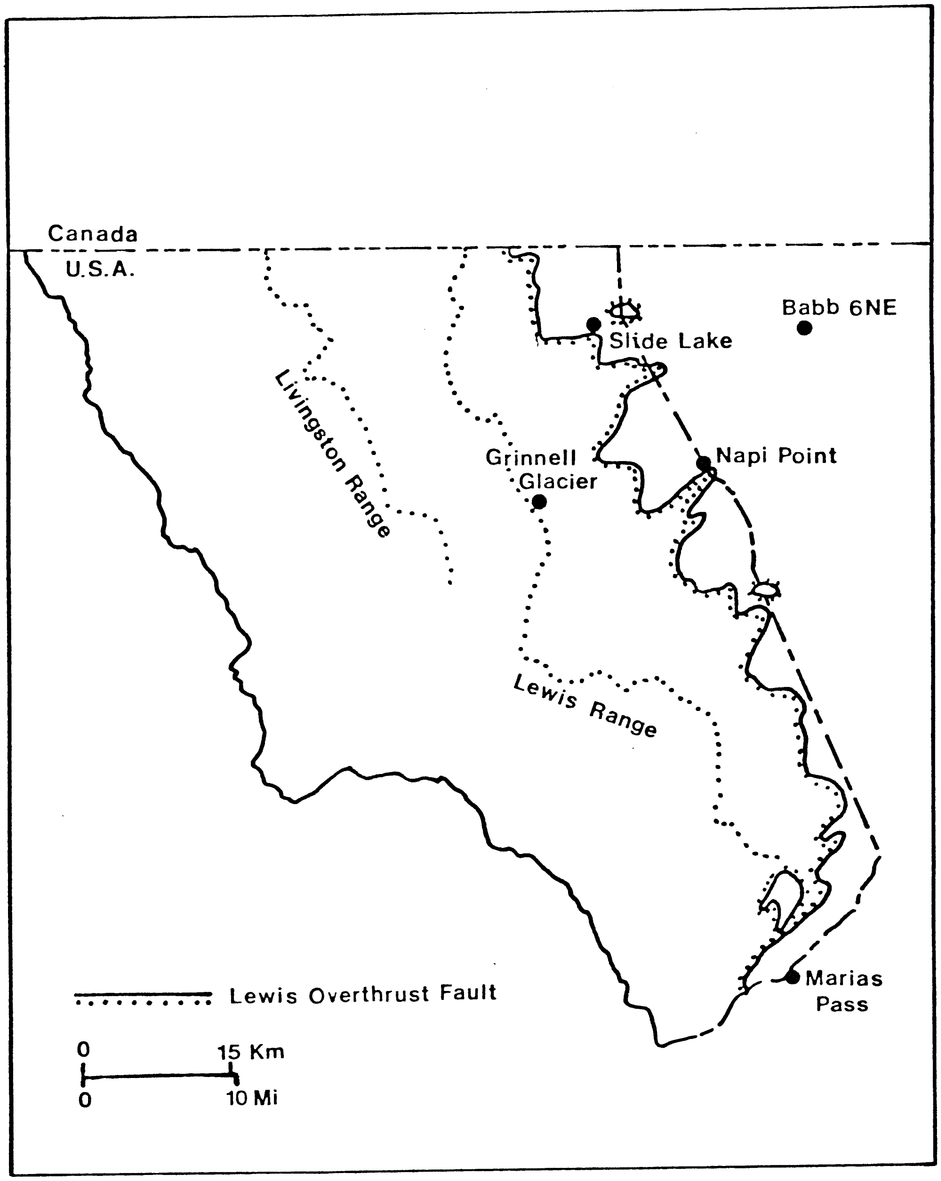


Figure 2. Glacier National Park, Including Areas of Interest in this Study.

Glacier National Park is named for its glaciated terrain, not for the 50 or more small glaciers which remain in the Park (Dyson, 1966). The Park was extensively glaciated during the Pleistocene, and retreat from the last Neoglacial advance began about 1860 (Carrara and McGimsey, 1981). Elevations in the Park range from 961 meters at Lake McDonald to well over 3,000 meters on several mountain peaks.

## Research Study Area

### Geology of Area

An understanding of the geology of the study area will aid in understanding the processes of slope movement in this area. In terms of lithology, most of the rocks in this area are of the Belt series (Ross, 1959). The sands, clays, muds, and limes that existed in shallow seas overlying the study area were deposited, eventually hardened, and formed the sedimentary rocks which are now predominant in the Park. These rocks were formed during the Precambian era, from 500 million years ago (Ross, 1959) to 1,600 million years ago (Raup *et al.*, 1983). Four main rock units are present (Alt and Hyndman, 1973). The lowest and oldest unit is the Altyn formation, consisting primarily of white Altyn limestone. Overlying this formation is the Appekunny formation, which is composed of green, gray, and red mudstones and argillites. Next is the Grinnell formation, consisting of red and green mudstones and white sandstones. The youngest of the four rock units is the Siyeh formation, a conspicuous formation of tan and gray limestone. The Shephard and Kintla formations lie above the Siyeh limestone, but only a few remnants of these formations remain in the Park (Ross, 1959). Cretaceous mudstones lie immediately at the base, and eastward, of the Belt series.

## Lewis Overthrust Fault

The dominant structural feature in the Park, and the focus of this study, is the Lewis Overthrust fault (Figure 2). Named by Willis (1902), this fault has been termed a "classical thrust fault" (Mudge and Earhart, 1980), and is a part of the "disturbed belt" in Montana. The Lewis Overthrust fault (hereafter termed "Lewis thrust fault") is an area where older Precambrian sedimentary rocks now overlie younger Cretaceous rocks - a reversal of normal geologic stratigraphic conditions. Two opposing theories exist as to how the Lewis thrust fault evolved. Dyson (1971) speculates that uplift pressure from the west pushed the Precambrian slab eastward and over the Cretaceous mudstones, a distance of some 48-56 kilometers. Alt and Hyndman (1973) believe that as regional uplift west of the Park occurred, the Precambrian slab was pulled downward and eastward, under the influence of gravity, over the Cretaceous rock. During the development of the Lewis thrust fault, extreme folding occurred in the

Cretaceous rocks beneath the Lewis thrust fault (Ross, 1959). As the Precambrian slab moved eastward, numerous faults formed, the larger of which were roughly parallel to each other and trended northwestward. The Lewis thrust fault developed in the late Paleocene to early Eocene (Ross, 1959; Alt and Hyndman, 1973).

In the Lewis thrust fault highly jointed, fractured, and faulted Precambrian Altyn limestone and Appekunny argillite overlie weak Cretaceous mudstones, and create an area prone to landslide activity. Mudge and Earhart (1980) note that everywhere along the Lewis thrust plate in Glacier National Park this reversal exists; in effect, the entire Lewis thrust fault in Glacier Park represents a landslide-prone area. Several subsidiary faults exist in conjunction with the Lewis thrust fault, adding to the instability of the area (Ross, 1959).

## Climate

Because the Continental Divide divides Glacier National Park roughly in half, two distinct climatic zones are present. The eastern section of the Park (locally termed the "East side"), within which lies the Lewis thrust fault, is a much drier, cooler, and windier area than the area west of the Divide. The weather station at Babb 6NE, Montana (Figure 2), is used in this study as a representative station for the climate for the East side. This station is located near the northeastern boundary of the Park, and has an elevation of 1,351 m. This station is the closest weather station to two landslides which were studied in detail in the field; it is located approximately 13 km from each deposit.

The variability in precipitation activity between lower and higher elevations is evident; therefore, the Lewis thrust fault probably has higher precipitation amounts than the Babb station just a few kilometers to the east. Although the Babb station receives less precipitation than the mountains immediately to the west, the monthly totals are indicative of the precipitation pattern on the East side for any one year. The average annual precipitation for this station, based on 77 years of data, is 488 mm (U.S. Department of Commerce, 1981). May and June are the wettest months, averaging 68 and 112 mm of precipitation, respectively. Winter snow accumulations are heavy, although generally less than on the West side. Important variations in precipitation exist on the East side of Glacier National Park (Dightman, 1967). Whereas the Babb station would be classified as semi-arid, the mountain ridges only a few kilometers to the west may be classed as humid, because of heavy mountain snows. A 17 year precipitation record near Grinnell Glacier (approximate elevation 2,000 m, Figure 2) averaged 2,650 mm of total annual precipitation, while an 11 year record at a nearby location averaged 3,750 mm annually (Carrara and McGimsey, 1981). These stations are both located east of the Continental Divide in the 1,800-2,100 m elevation range. Marias Pass (1,700 m, Figure 2), located on the Continental Divide at the south edge of the Park, averaged 6,500 mm of snowfall between 1935 and 1967 (Dightman, 1967).

Average temperatures from the Babb station range from $-7^{\circ}$ C in January to a high of $16^{\circ}$ C in July (U.S. Weather Bureau, no date). Chinook winds are common in the winter, at times raising temperatures as much as 6.6 C higher than the West side. These winds can achieve very high speeds, occasionally reaching 160 km/hr. During the late spring and early summer, heavy precipitation typical of May and June, combined with warm temperatures which will rapidly melt mountain snows, can produce flood situations. In 1953, 1964, and 1975, rapid snow melt and heavy rains resulted in flooding in Glacier National Park (Glacier National Park, 1964, and personal experience, 1975).

## Vegetation

A great diversity of plant life exists in Glacier National Park, both east and west of the Continental Divide. Kessell (1974) listed 1,006 vascular plant species within the Park; this brief description will center only on the major tree species present along the eastern slopes. Eastern slope tree species are similar to those found in the southern Rocky Mountains (Butler, 1979a). Because of the great differences in elevation within the Park, different "life zones" are recognized. These life zones are characterized by plants and animals which have adapted best to the many variables of climate and environment found within each zone. Ruhle (1963) and Robinson (1972) describe these life zones within Glacier Park. Much of the Lewis thrust fault lies within the Hudsonian zone, which extends from about 1,815 m to 2,425 m in the Park. Species common on the eastern slopes include subalpine fir (*Abies lasiocarpa* (Hook) Nutt.), Englemann spruce, Whitebark pine (*Pinus albicaulus* Engelm.), and the Limber pine (*Pinus flexilis* James), (Shaw and On, 1979). The last zone is the Alpine-Arctic zone; no trees grow within this zone, although some dwarfed shrubs exist. Mosses, lichens, and wildflowers typify this zone.

## The Napi Point and Slide Lake Landslides

Although the entire Lewis thrust fault is the specific area of this study, two high magnitude landslide deposits along this fault were studied in detail (Figure 2). Both occur on northerly slopes along the Lewis thrust fault. The Slide Lake landslide is located at the base of the north slope of Yellow Mountain in the northeast corner of the Park, approximately 11 km south of the Canadian border. This landslide event dammed Otatso Creek, forming Slide Lake and an unnamed pond just to the east (Butler, 1983a). The landslide deposit extends from approximately

2,180 m down to 1,820 m. The second landslide is located on the northern slope of East Flattop Mountain, approximately 23 km south of the Canadian border and 6.5 km northwest of St. Mary, Montana. It is called the Napi Point slide because it occurred off the Napi Point section of East Flattop Mountain. This deposit extends from 1,940 m down to 1,690 m. Detailed characteristics of each deposit will be provided in Chapter Four.

# CHAPTER III

# LANDSLIDES ALONG THE LEWIS THRUST FAULT

## General Statement

An inventory and description of landslide deposits along the Lewis thrust fault was completed, with two objectives in mind; 1) determine the landslide type and spatial distribution of landslide activity along the Lewis thrust fault, and 2) speculate whether slope aspect and subsidiary faulting control landslide activity along the Lewis thrust fault. Other variables, such as the amount of fracturing and jointing of the bedrock, the absence or presence of free cliff faces, or slope angle may also contribute to landslide activity. Unfortunately, no data exist from which to measure the effect of these variables.

### Landslide Classification

This study uses "landslide" as a broad term indicating the downslope movement of large masses of rock and earthen material. Several useful terms exist to more accurately describe the various mass-movement types. Sharpe (1938) defined landslides as the perceptible downward sliding or falling of a relatively dry mass of earth, rock, or mixture of the two. The slope is usually steep and the movement rapid. Sharpe (1938) divided landslides into five classes: slumps, in which the mass rotates backward, leaving the upper block surface diminished or reversed; debris-slides, which produce deposits resembling morainal topography; debris-falls, in which unconsolidated material drops from a vertical cliff; rockslides, which occur on bedding, joint, and foliation surfaces, or other planes of weakness; and rockfalls, which include blocks or plates of rock, loosened by weathering or oversteepened by rivers, valley glaciers, or wave cutting, falling under the influence of gravity.

Sharpe's (1938) work, which emphasized type of movement, was later modified by Varnes (1958). Varnes considered two variables in his classification. These variables included type of movement, which could be determined after a short period of observation or by the shape of the slide; and the type of material involved. Material involved can be determined by inspection of the slide deposit. Varnes (1958) described types of movements as falls, slides, flows, and complex; and types of material as bedrock, soils, and unconsolidated material. Varnes' (1958) classification was prepared so that features could be observed at once or with a minimum of investigation, without reference to the causes of landslides. Varnes (1975) modified his earlier classification to include topple and lateral spread movements. Varnes' (1958) landslide classification is shown in Figure 3.

| TYPE OF MOVEMENT | TYPE OF MATERIAL | | | |
|---|---|---|---|---|
| | BEDROCK | | SOILS | |
| FALLS | ROCKFALL | | SOILFALL | |
| | ROTATIONAL | PLANAR | PLANAR | ROTATIONAL |
| SLIDES — FEW UNITS | SLUMP | BLOCK GLIDE | BLOCK GLIDE | BLOCK SLUMP |
| SLIDES — MANY UNITS | | ROCKSLIDE | DEBRIS SLIDE | FAILURE BY LATERAL SPREADING |
| | ALL UNCONSOLIDATED | | | |
| | ROCK FRAGMENTS | SAND OR SILT | MIXED | MOSTLY PLASTIC |
| FLOWS — DRY | ROCK FRAGMENT FLOW | SAND RUN; LOESS FLOW | | |
| FLOWS | | RAPID EARTHFLOW | DEBRIS AVALANCHE | SLOW EARTHFLOW |
| FLOWS — WET | | SAND OR SILT FLOW | DEBRIS FLOW | MUDFLOW |
| COMPLEX | COMBINATIONS OF MATERIALS OR TYPE OF MOVEMENT | | | |

Figure 3. Varnes' (1958) Landslide Classification.

## Landslide Inventory and Mapping Methods

One of the objectives of this study is to describe the spatial distribution of landslide activity in the study area. No inventory of landslides along the entire Lewis thrust fault in Glacier National Park has been compiled; obviously it was necessary to complete this task before the objective of discussion of distribution could be fulfilled.

Varnes' (1958) classification was used in the inventory portion of the study, because it was devised to easily identify a deposit with a minimal amount of site investigation. Two landslide events in Glacier National Park, the Slide Lake and Napi Point deposits, were examined in detail (Figure 2). Much of the Lewis thrust fault area, however, is highly inaccessible. Although hiking trails penetrate most mountain valleys along the east side of Glacier National Park, travel off these trails can be extremely difficult. Dense vegetation, steep mountain slopes, and the grizzly bear (<u>Ursus arctos horribilis</u>) make bushwhacking in the study area difficult and dangerous. The sheer distances involved in tracing the Lewis thrust fault throughout the Park also precluded on-site investigation of all landslide activity in the study area. It thus became necessary to develop an inventory using other than on-site observation methods.

Large-scale aerial photographs provide a useful means to develop a landslide inventory, and their use is perhaps the most common method for developing inventories of mass-wasting deposits. A few examples of studies utilizing this technique include Nilsen and Brabb (1972), Davies (1974), Colton *et al*. (1976), Ives *et al*. (1976), Harden (1976), Ives and Bovis (1978), Butler (1979), Whitehouse (1983), and Gardner *et al*. (1983). Dishaw (1967) recommended smaller scale photos (1:50,000) to identify "massive" older landslides.

### Landslide Identification on Aerial Photos

Black and white aerial photos taken in the summer of 1968 were examined. The scale of the photos was 1:16,000. Sets of overlapping photos for the entire Lewis thrust fault were available, so stereoscopic viewing was possible. Although the photos were large scale, landslide-type identification usually was not possible, except to distinguish between rockfall/slide deposits and slump/earthflow deposits.

The geomorphic features of rockfall/rockslide avalanche deposits which Mudge (1965) described were used as indicators of those deposits on the aerial photos. They include: 1) hummocky surface; 2) relatively low relief from head to toe; 3) arcuate ridges and furrows, at least where confined; 4) lobate form, where not confined; 5) local pressure ridges where flows were impeded by barriers; 6) trough between the head of the deposit and the base of the cliff or scar; 7) movement up or over topographically high ground; and 8) volume measurable in millions of cubic meters of rock. Generally, features one (1) and four (4) provided the best indication of a rockfall/slide deposit on the aerial photos. Although most deposits do not contain millions of cubic meters of rock, the same geomorphic features are common on smaller deposits. Other features considered in landslide identification included marked vegetational differences on landslide deposits vs. surrounding areas, mounding of rock material at bases of cliffs, and large rocks and boulders present on a deposit surface.

Slumps and earthflows contained an assortment of debris material other than rock. This feature was the primary identification difference between rockfalls/slides and slumps and earthflows, and could be easily distinguished on the aerial photos. Slumps typically had hummocky surfaces; earthflows had lobate masses and possessed a fluid, flow-like morphology. Sharpe (1938) noted that the delineation between slides (i.e. slumps) and flows (i.e. earthflows) can be made according to whether a slip plane separating the moving mass from the stable ground is present. Earthflows will have no slip plane, whereas slumps will. Sharpe (1938) recognized that the delineation between the two is not always clear, and that some flows initiate with slippage. Varnes (1958) provided examples of slumps, and noted that the top surface of each deposit tilts backward toward the slope. The scarp at the head of a slump may be almost vertical, and typically is horseshoe-shaped. Varnes described earthflows as a displaced mass, with the moving material resembling the movement of viscous fluids. A ridge and furrow morphology can be present on the displaced mass. The material may consist of rock fragments, fine granular materials, mixed debris and water, or plastic clay. Earthflows in plastic or predominantly fine-grained material become mudflows at higher water content. Rapp and Nyberg (1981) described the morphology of debris flows and mudflows (finer-grained) as rapid flows of wet, lobate masses, having slide scars, erosional gullies, and outspread front fans of finer material. Delineation on the aerial photos between debris flows and mudflows was not possible.

Rockfall/slide deposits typically were found along the base of the mountains, and had individual rocks visible on the surface. Slumps and earthflows generally occurred a short distance away from the base of the mountains.

## Previous Description of Landslide Deposits in the Study Area

Although a vast majority of the landslide deposits were identified from the aerial photos, some were located from references in published literature. The geology of

Glacier National Park, mapped by Ross (1959), included landslide deposits only where they obscured the bedrock geology; two deposits along the Lewis thrust fault were mapped. Dyson (1971) mentioned landslide deposits along the southerly slopes of Appekunny Mountain. Raup et al. (1983) briefly mentioned a landslide originating on Red Eagle Mountain. Butler (1979a) mapped mass-movement deposits along a part of the Lewis thrust fault, and Butler (1983a) mapped and described two landslide deposits along the Lewis thrust fault in Glacier National Park.

Alt and Hyndman (1973) provided a general discussion on mass-wasting features in Glacier National Park. They stated that landslides are especially common east of the Lewis Overthrust fault, where the hills are underlain by weak Cretaceous sandstones and mudstones which are heavily plastered with glacial deposits. The authors attributed rockfall activity to the presence of steep cliffs left unsupported as ice melted at the end of the last glacial period. Mention was made of a large rockfall which slightly altered the profile of Chief Mountain during the summer of 1972. Their comments on rockfall avalanche activity in general are interesting, as they stated that such mass movements have not occurred in the Park recently, although they have occurred in nearby areas. The next few paragraphs and Chapter Four will illustrate that rockfall avalanches have indeed occurred within the Park, and within this century.

Mapping Symbology

No universal symbology for landslide mapping exists. Shroder and Putnam (1972) presented a set of symbols for mass-wasting features, based on type of movement and type of material. Gardiner and Dackombe (1983) based their slope instability mapping symbols on landform genesis; Gardner *et al*. (1983) developed symbols for mapping mass transfer features in their study area.

Crofts (1981) indicated that the map scale determines the type and extent of detail of landform mapping. Landforms on larger scale maps should be depicted by symbols which reflect their shape and areal extent; symbols on small scale maps should represent either the form or process of landforms. Salome and van Dorsser (1982) discussed six systems of geomorphological mapping. The different systems incorporate colors to indicate different aspects of geomorphological units, including morphology, morphometry, morphogenesis, and morphochronology. The authors concluded that the cost aspects of map construction and printing are important for the choice of which system is best suited for a project.

The study area for this landslide inventory was quite large, as the Lewis thrust fault extends for approximately 170 km throughout Glacier National Park (Mudge, 1977). The best available base map for mapping the landslide deposits is at a scale of 1:100,000. A small-scale map such as this precluded the use of detailed mapping symbols such as those of Shroder and Putnam (1972), so a different set of symbols was developed for mapping the landslide deposits along the Lewis thrust fault. The primary concern in developing the symbology was that the reader could easily discern between one symbol and another, and that the symbols could adequately indicate the presence of a landslide deposit.

Using the 1968 aerial photos, the entire Lewis thrust fault area was examined three times. Landslide deposits were mapped on 1:24,000 Quadrangle maps. A list of the Quadrangle maps used for mapping the landslide deposits is found in Appendix A. Information from these maps was then transferred onto a 1:100,000 map of the study area. This transfer of information from a large to small-scale map involves an inevitable loss of detail, such as boundary delineation of each deposit, but the regional pattern of landslide distribution is still preserved. Van Zuidam (1982) concluded that reducing large scale geomorphological maps to a 100,000 scale map is acceptable, with only a slight loss of geomorphic detail. The mapping symbology devised for Plate 1 still allows for identification of the smallest deposits (approximately 250 square meters). The results are shown on Plate 1. The location of the Lewis thrust fault and other subsidiary faults are from Ross (1959).

## Results and Discussion

As seen in Plate 1, landslide deposits are common landforms in the study area. On some of the massive deposits it was impossible to distinguish on the aerial photos if the area was one discrete deposit or a combination of many landslide deposits. In all likelihood, some of the deposits shown in Plate 1 are the result of many landslide events over the years. A good example is the large deposit immediately to the north of Lake Sherburne. Dyson (1971) attributed the bumpy topography of this slope to "innumerable" small landslides of Cretaceous shales. Because boundaries of individual deposits are impossible to identify, the entire area is shown as one large deposit.

With this point in mind, 78 discrete landslide deposits along the Lewis thrust fault are shown on Plate 1. The differentiation of landslide deposits was made based on type of material involved in the deposit, as could be best determined on the aerial photos. There are 55 rockfall/slide deposits, four slumps, four earthflows, and 15 undifferentiated deposits. The undifferentiated deposits are areas which typically possessed some features of landslide deposits, such as hummocky appearance and mounding of material at the base of cliffs, but lack other features usually associated with a deposit, such as a cliff scar or lobate form. These areas indicate deposition of earthen material; whether they are landslide deposits or not is uncertain.

Also shown on Plate 1 are 10 areas which may be mass-movement deposits. Classification is uncertain for two reasons: 1) as is typical of landslide deposits, the topography of the area is such that it is decidedly different than the immediately surrounding area, yet it either lacks the geomorphic features of landslides or its features are simply uninterpretable; and/or 2) the lighting on the aerial photos make these areas impossible to interpret, usually because of glare. Cloud cover was not a problem in the photo interpretation.

The high number of rockfall/slide deposits (55, or 71% of the total deposits) is expected, considering that the Lewis thrust fault is a feature associated with mountains of high relief and jointed, fractured, and weathered bedrock. Most of the eight slumps and earthflows are located a short distance from the base of the mountains, and contain "undifferentiated Cretaceous deposits" (Ross, 1959). All of the 15 undifferentiated deposits lie at the base of mountains, and appear to contain rock material. The 10 unknown deposits are a mixture; some consist of debris, some of rock.

## Possible Factors Controlling the Spatial Distribution of Landslides

### General Statement

Many studies have attempted to explain the distribution of landslides on the basis of a variety of environmental and topographical variables (Mudge, 1965: Curry, 1966; Williams and Guy, 1971; Beatty, 1972; Bogucki, 1976; Harden, 1976; Rapp and Stromquist, 1976; Clague, 1981; and Pomeroy, 1982). Examples of these variables include slope aspect, extreme precipitation events, prevailing wind direction, topography of slopes, steepness of slopes, texture of slope material, presence or absence of vegetation on slopes, and the amount of fracturing, faulting, and jointing of the bedrock. Data do not exist for most of these variables in the study area, and would have to be field-generated. The general lithology of the study area, discussed in Chapter Two, may be an important controlling factor in the distribution of landslides. Because on-site investigation of each deposit was not possible, the type of rock present in each deposit is unknown. This lack of data again precluded any effort to determine if lithology is a controlling factor in landslide distribution, although the lithology of the two field-studied sites will be discussed.

The variables of slope aspect, prevailing wind direction, and subsidiary faulting can be found on maps and in published literature. The amount of fracturing and jointing of the bedrock at the two field sites, another possible controlling factor of landslide distribution, will be discussed in Chapter Four.

Slope Orientation of Landslides

The orientation of the landslide deposits was examined to determine if aspect may be an important environmental factor in the spatial distribution of landslides in the study area. Several studies discuss the importance of aspect in spatial distributions of landslides. Beatty (1972) found that most post-glacial slumps in his study area in southern Alberta were located on northerly, northeasterly, and easterly slopes. Harden (1976) similarly found most landslides in the mountains near Aspen, Colorado were located along north- and east-facing slopes. Pomeroy (1982) found most mudflows in his study area in Pennsylvania occurred on north-northeast-facing slopes. These authors all believed the microclimatic differences present on these slopes were responsible for a majority of deposits being located on northerly, northeasterly, and easterly slopes. These slopes experience reduced energy receipt, allowing for longer periods of moisture retention on the slopes. This greater period of moisture retention can result in a greater number of landslides on slopes having these orientations, because slopes commonly fail because of saturation.

In a particularly relevant study, Gardner et al. (1983) noted that both high and low magnitude rockfall/slide deposits originated along northwest through east-facing slopes. Exposures with this northerly component tended to retain snow and ice longer than other exposures. Beatty (1972) also noted that prevailing westerly winds tended to pile snow up on northeast, east, and southeast slopes. This factor, combined with the lower levels of insolation received on north and northeast slopes, provided a strong controlling influence in the distribution of slumps. Eisbacher (1971) and Cruden and Kahn (1973) concluded that the direction and amount which rock formations dip may influence landslide activity. Landslides may concentrate along slopes where the dip parallels the slope.

The slope orientation of the 78 landslide deposits located on Plate 1 was determined by measuring azimuth from north. All deposits were placed in one of eight directional octants. Results are shown in Figure 4. Deposits are most numerous along north- and southeast-facing slopes (23 and 16, respectively). Fewest deposits are located along the south- and southwest-facing slopes (two and one, respectively). The northwest through east octants contain 49 deposits, or 63 % of the total; the southeast through west octants contain 29 deposits, or 37% of the total.

The dip of the rock formations along the Lewis thrust fault apparently exerts little influence on landslide distribution. The dip in this area is to the southwest, and generally less than 10 (Ross, 1959). Most landslide deposits are located on northwest- east-facing slopes,

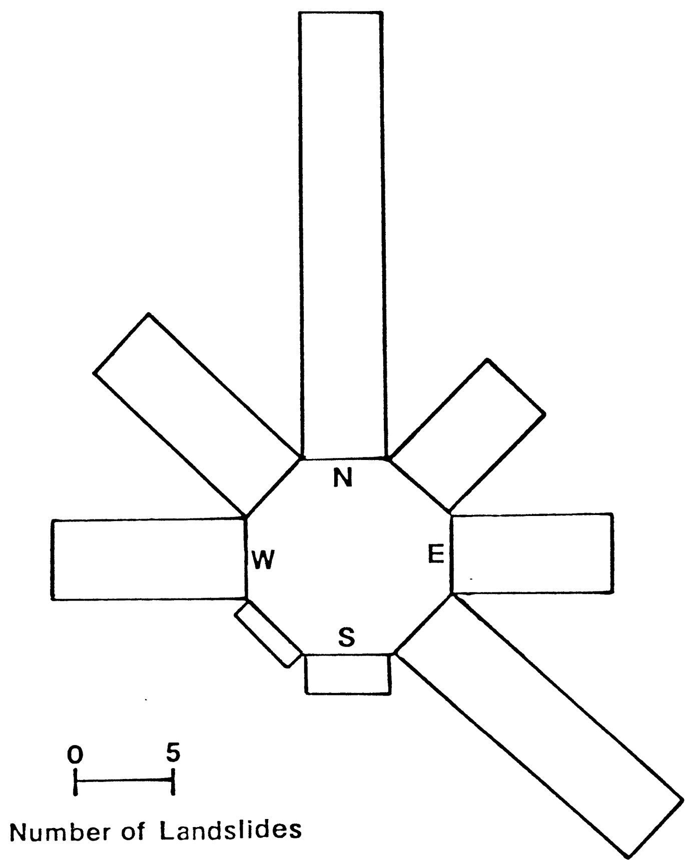


Figure 4. Slope Orientation of Landslide Deposits.

which are in the direction away from the dip of the Lewis thrust fault. These findings tend to conform to the findings of others, as discussed previously. The majority of landslide deposits on northwest- through east-facing slopes suggests possible microclimatic influences on the distribution of the deposits. Mountains in this study area receive heavy winter snowfall, and prevailing winds are from the west. These factors create a situation similar to the Beatty (1972) study. Butler's (1979a) study on the distribution of snow avalanche paths confirms this situation. Butler concluded that westerly winds produced cornices and snowpack instability on southeast-facing slopes, although shading led to more stable snowpacks on northerly exposures. Microclimatic influences may play a role in the distribution of landslides along the Lewis thrust fault, because of the orientation of slopes.

One important factor must be kept in mind, however, when considering the role of aspect in the spatial distribution of the landslide deposits. The general trend of the Lewis thrust fault is in a northwest-southeast direction. This trend would suggest a predominance of northeast-southwest-facing slopes, a bias of slope orientation. In reality, however, the Lewis thrust fault follows a very sinuous path along the eastern front of the Rockies in Glacier National Park; slopes of each octant appear to be well represented. Perhaps a more important bias is the fact that the Lewis thrust fault is exposed only along one side of the mountains of the Lewis Range in the Park, generally the eastern side. This bias would tend to reduce landslide activity in the southwest and west octants. Such a theory appears to be borne out; only three deposits are located in these octants. The southwest and west octants would be least prone to landslide activity due to microclimatic differences, also; perhaps both of these features have been important in reducing the amount of landslide activity along the southwest- and west-facing slopes.

In summary, the orientations of the landslide deposits along the Lewis thrust fault are likely a result of microclimatic influences and the built-in bias of slope orientation along much of the Lewis thrust fault. It is beyond the scope of this study to determine which factor was of primary importance.

## Subsidiary Faulting

Ross (1959) mapped the subsidiary faults associated with the Lewis thrust fault. These faults are shown on Plate 1. Some general comments concerning the role of the subsidiary faults in the spatial distribution of the landslide deposits can be made.

Generally, some type of landslide deposit is found within a relatively short distance of all subsidiary faults. Examples of areas exhibiting a close areal relationship between landsliding deposits and subsidiary faults include the northwest slopes of Curly Bear Mountain and the Gable Mountain-Yellow Mountain area (Plate 1). Both of these areas show large landslide deposits lying adjacent to subsidiary faults. The longest section of the Lewis thrust fault along which no landslide deposits are found (near the Canadian border) also has no subsidiary faults. This section is approximately 9.5 km long. Several large deposits near the southern boundary of Glacier National Park, however, are not associated with any subsidiary faulting.

Only one group of subsidiary faults is not generally associated with landslide deposits. This area is the valley of Divide Creek, between White Calf and Curly Bear Mountains (Plate 1). The farthest point from one of these faults to the nearest landslide deposit is 2.3 km; nowhere else along the Lewis thrust fault is this great a distance between subsidiary faults and a landslide deposit encountered. Perhaps part of the reason this area has no landslide deposits is the fact that the Lewis thrust fault lies away from the mountains (downslope), along the floor of much of this valley (see Plate 1).

In summary, an examination of Plate 1 indicates a general areal relationship between subsidiary faulting and landslide deposits. Subsidiary faulting may be an important spatial control of landslide activity along the Lewis thrust fault.

## Conclusions

The sheer number of landslide deposits - a minimum of 78 - along a relatively small geographical distance indicates the slopes along the Lewis thrust fault may be very active landslide areas. Mention in Chapter Two of the importance of the Lewis thrust fault in contributing to landslide activity seems to be conclusively confirmed on Plate 1. The presence of highly jointed, fractured, and faulted Precambrian rocks overlying weak Cretaceous mudstones produces an interface along which much landslide activity originates. No landslide inventory for other areas of Glacier National Park was completed to determine if landslide activity was as widespread in those areas. Butler's (1979a) mapping of mass movement deposits in the central one-third of the Park shows very few deposits, except along the Lewis thrust fault. The heavy concentration of landslide deposits along the Lewis thrust fault indicates this area is an active landslide-prone area.

# CHAPTER IV

# ANALYSIS OF THE NAPI POINT AND SLIDE LAKE DEPOSITS

## Introduction

Field work was conducted on the Slide Lake and Napi Point rockfall avalanches during the summer of 1983. The primary objectives of the field work were: 1) obtain data for absolute and relative age-dating of each deposit; and 2) examine the general topographical and lithological features of each deposit. Field work on each site included core and crosscut sampling of trees along the borders of each deposit, measuring the slope characteristics of each site, qualitatively recording the vegetation on each site, and obtaining rock weathering rind and lichen samples.

### The Slide Lake and Napi Point Rockfall Avalanches

The Slide Lake and Napi Point rockfall avalanches are quite large in terms of volume of material displaced, and have formed lobate masses. The area of the Slide Lake deposit is approximately 1.4 $km^2$; the Napi Point deposit has an area of 0.52 $km^2$ (Butler, 1983a).

Varnes (1958) discussed dry flows consisting predominantly of rock fragments, originating from large rockslides or rockfalls. He termed these "rockslide avalanche or rockfall avalanche," which are produced by the flowing movement of many millions of tons or rock material. Sharpe (1938) further distinguished rockslides from rockfalls, noting that rockslides travel or slide on bedding, joint, or fault planes, whereas rockfalls are newly detached segments of bedrock, falling freely without any slip plane.

Both the Napi Point and Slide Lake deposits qualify as "avalanches" in that both contain very large volumes of rock fragments, and apparently flowed like avalanches. They also appear to have fallen freely, not along any slip planes, and thus are considered to be "rockfall avalanches."

This chapter will initially discuss the general features of each site, followed by a literature review of the techniques and methodologies used in absolute and relative age-dating of landslide deposits. The techniques used in this study will then be presented, followed by a discussion of the results of the field work.

## The Slide Lake Rockfall Avalanche

### Morphologic Features

The Slide Lake rockfall avalanche is found on the 1968 Chief Mountain, Montana - Alberta Quadrangle (Figure

5). This deposit is located in the Otatso Creek valley (formerly the North Fork of Kennedy Creek). A rugged jeep trail across land of the Blackfeet Indian Reservation turns into a hiking trail at the National Park boundary. The deposit is first noticeable along the hiking trail approximately two km west of the Park/Reservation boundary. Three major lobes of the deposit extend downslope (north) from the base of Yellow Mountain. The straightline distance from the base of Yellow Mountain to the bottom of the lobe damming Slide Lake is approximately 1.55 km. At its widest point, near the source, the deposit is 1.58 km across. Individual lobes are up to 200 meters wide and 700 meters long. The area of the rockfall deposit is 1.4 sq. km (Butler, 1983a). A trough approximately 725 m wide extends from the headwall on Yellow Mountain northward. This headwall trough, as described by Mudge (1965), is a geomorphic feature typical of rockfall avalanches. A large headwall cliff scar on Yellow Mountain is present above the entire Slide Lake deposit (Figure 6). This scar is heavily fractured and jointed, and is comprised of Altyn limestone.

Of the three main lobes, only the easternmost lobe does not reach Otatso Creek. A small stream, fed by snowmelt, passes through this lobe and feeds into Otatso Creek. The stream is heavily silt-laden, even late in the fall (last observed in mid-October, 1983). The sediment from this small stream discolors the water in Otatso Creek

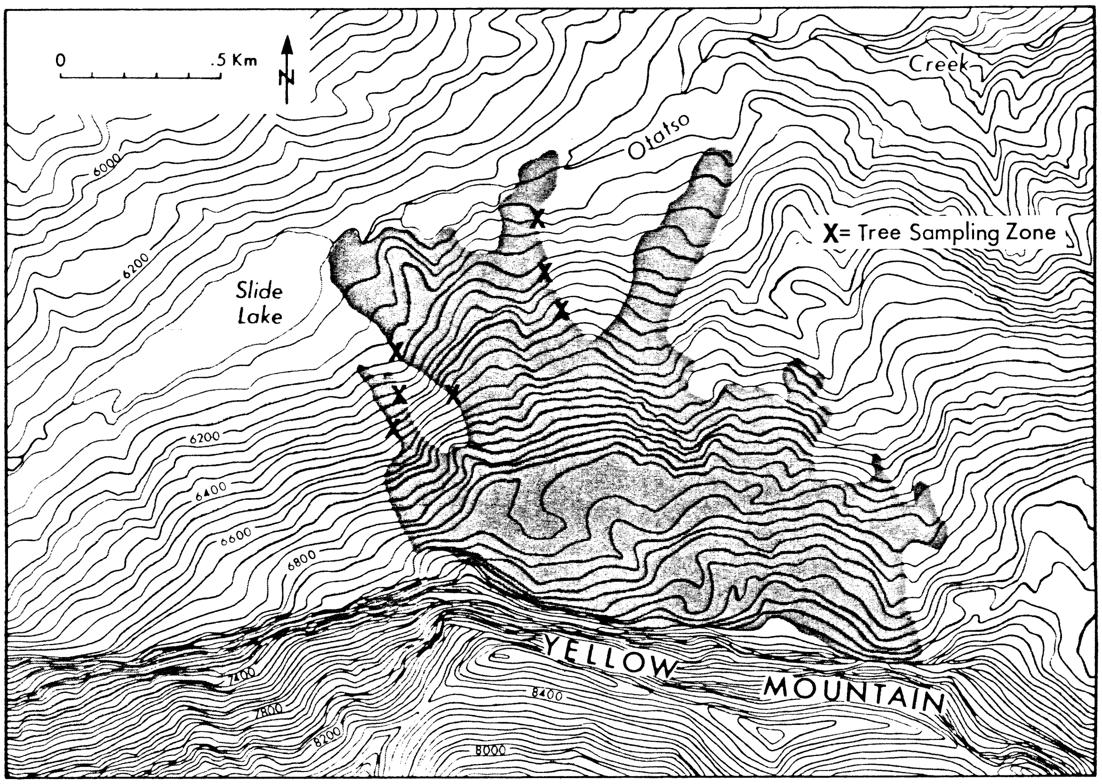


Figure 5. The Slide Lake Deposit (From Butler, 1983a).

Figure 6. Photo of the Slide Lake Headwall and Trough.

far downstream. The central lobe extends downslope and crosses Otatso Creek, extending upslope on the opposite valley wall to a height of about five m. This lobe dammed Otatso Creek, creating the pond to the east of Slide Lake; this pond will be referred to as Slide Pond. The westernmost lobe was responsible for impounding Slide Lake. The two lobes which cross Otatso Creek were studied in greatest detail; discussion will focus on them.

The general topography of the area just downstream (east) of Slide Pond, is of particular interest. Muskeg has developed in this area, with small ponds and meadows located here. Small fine-grained ridges from the rockfall avalanche extend across Otatso Creek. These ridges caused general surface-drainage disruptions and dammed some of the feeder streams into Otatso Creek. Drainage in this area is very poorly developed, and much of the ground surface has a gentle hummocky appearance.

Four important lithologic features were found on the Slide Lake deposit. Mudge (1965) described these four lithologic features commonly found on rockfall/slide avalanche deposits: 1) angularity and heterogeneity of constituents, including a mixture of relatively coarse sizes and exceptionally large blocks, with a newly-formed deposit having a very small proportion of finer grain sizes; 2) similarity of rock type to that of the source, except for rock fragments picked up enroute; 3) local imbricate structure; and 4) exceptionally high porosity, especially in newly-formed deposits. The entire deposit was exceptionally porous; running water could be heard under the rock surface in several locations on the deposit.

Longitudinal ridges extend down both lobes of the deposit; one ridge is approximately 15 m high (Figure 7). This ridge extends approximately halfway down the deposit from the lip of the headwall trough. A transverse ridge extends across the entire width of the Pond lobe (140 m) near the valley bottom. The height of this lobe is 10-15 m in some areas. Bounding ridges reach heights of 8-10 m and widths of 5-10 m.

The degrees of slope were determined by using a clinometer. The "break-in-slope" method was used; each time a distinct break in slope was visible, the degrees of slope and length of that angle was recorded. The slope characteristics for the Pond lobe and Lake lobe are presented in Figures 8 and 9.

## Lithologic Features

Both lobes consist primarily of Altyn limestone (Figure 7), which matches the rock type of the cliff scar. Rocks are angular, and the deposit is generally unstable; caution was needed when stepping from one rock to another. Most rock is relatively unweathered ("fresh looking"); little lichen growth or discoloration is found on the rock. Larger sized rocks (greater than 2 cm) dominate

Figure 7. Longitudinal Ridge on Slide Lake Deposit. Note Person for Scale. Boulders are Predominantly Altyn Limestone.

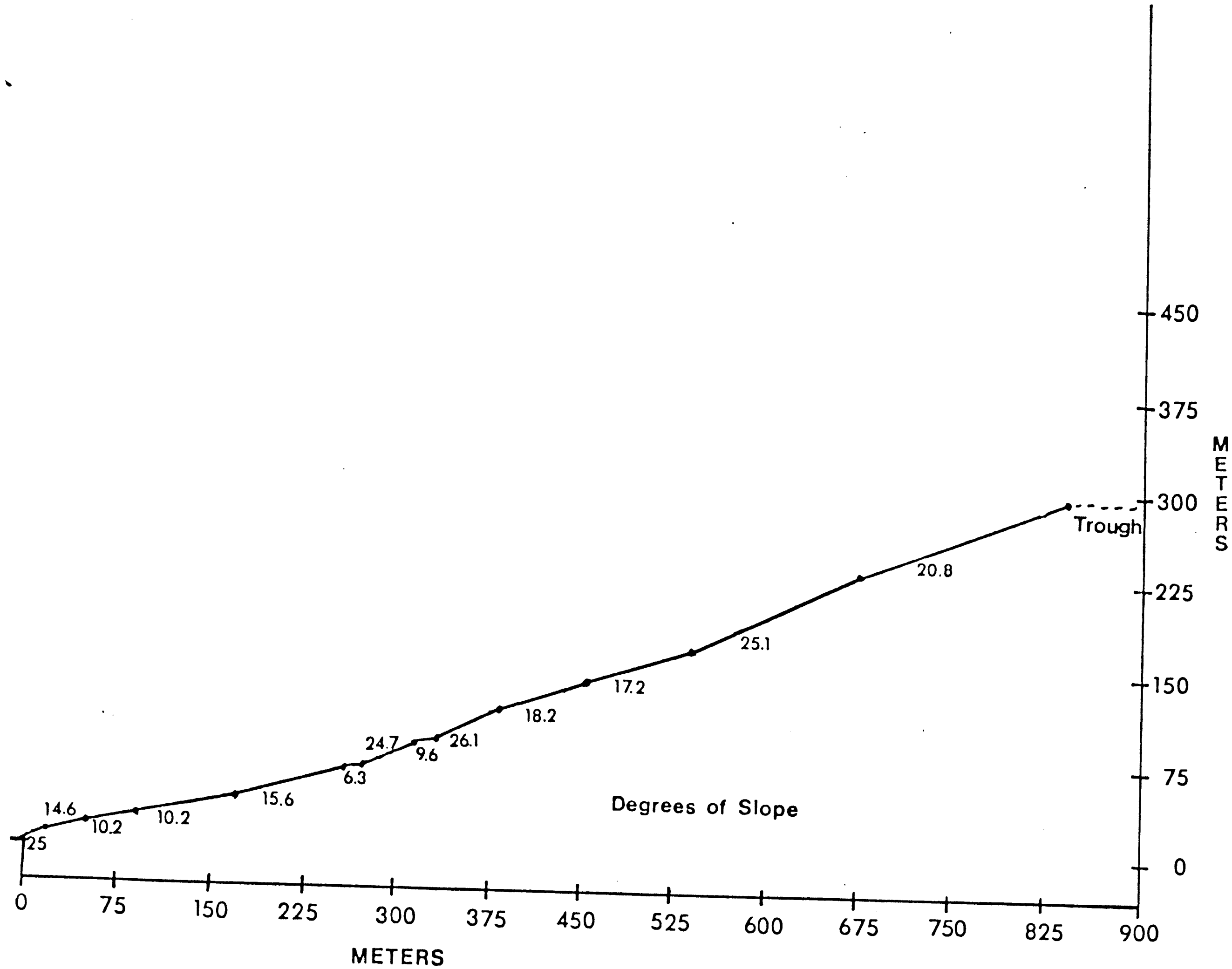


Figure 8. Slope Characteristics of the Pond Lobe.

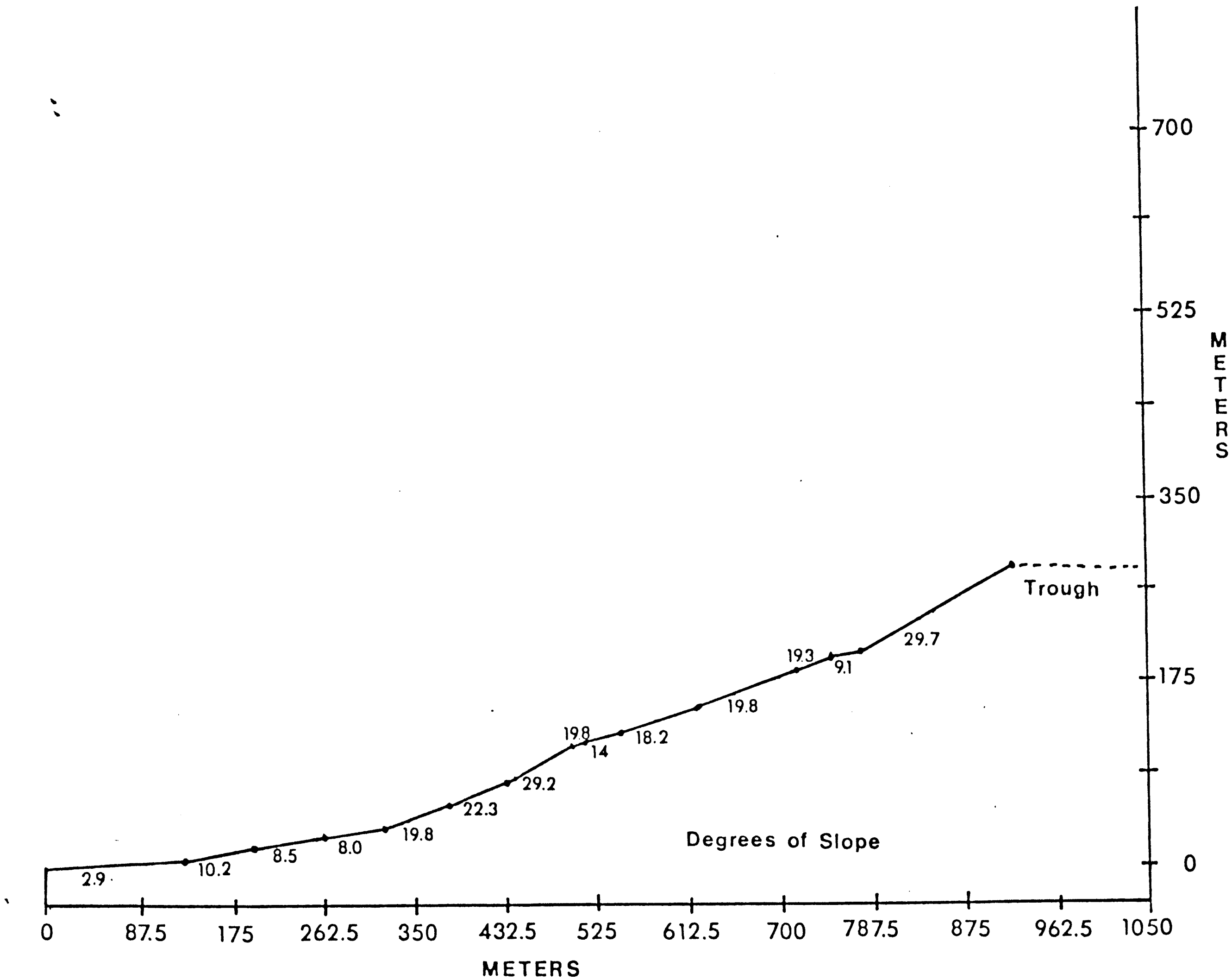


Figure 9. Slope Characteristics of the Lake Lobe.

each lobe, with exceptionally large-sized blocks present (Figure 10). The lobe which dammed Slide Pond ("Pond lobe") consists almost exclusively of coarse clasts. Only the lateral edges of this lobe have finer-grained materials. The lobe which dammed Slide Lake ("Lake lobe") consists of larger sized rocks along the upper reaches of the lobe; the lower one-third of the lobe consists of finer-grained material.

## Soils

Very little soil development exists on either lobe of the Slide Lake deposit. Coarse rocks dominate each lobe. Only the lower one-third of the Lake lobe has any noticeable soil development, with fine-grained materials located there. This area also has very gentle slopes. Soil development in this area may represent the deposition of fine-grained material from sections of the deposit above this area through water runoff. Similar soil development has not occurred on the Pond lobe because no gently-sloped areas occur on which material can be deposited.

## Vegetation

Vegetation along the northern slopes of Yellow Mountain consists primarily of mature Englemann spruce (*Picea englemannii* Parry) and subalpine fir (*Abies lasiocarpa* (Hook) Nutt.) forest. Whitebark pine (*Pinus albicaulus*

Figure 10. Photo of Large Blocks on the Slide Lake Deposit.

Englem.) and limber pine (Pinus flexilis James) are also present. Vegetation on the deposit is very scarce, except for the toe of the Lake lobe. The general lack of soil development on this deposit evidently precludes widespread vegetation establishment. On both lobes, occasional small subalpine firs were found; only the lower one-third of the Lake lobe has considerable vegetative invasion. This area has fine-grained materials and gentle slopes, and is generally an area conducive to plant establishment. Small subalpine firs, shrubs, and many annuals occupy this area.

## The Napi Point Rockfall Avalanche

### Morphologic Features

The Napi Point rockfall avalanche is found on the 1968 Babb, Montana Quadrangle (Figure 11). The deposit extends downslope (north) from the base of Napi Point, a part of East Flattop Mountain. Most of the deposit lies along the boundaries of Glacier National Park; however, both lobes cross the Park/Reservation boundary, extending north-northeast. An intermittent stream on the deposit is shown on the Babb quadrangle. The stream channel was dry when field work was conducted in August, 1983.

The Napi Point deposit extends downslope toward Boulder Creek, but does not reach the creek valley. Butler (1983a) described the Napi Point deposit as follows: the deposit reaches a maximum length of 1.32 km; at its widest point, the deposit is 0.5 km. The area of

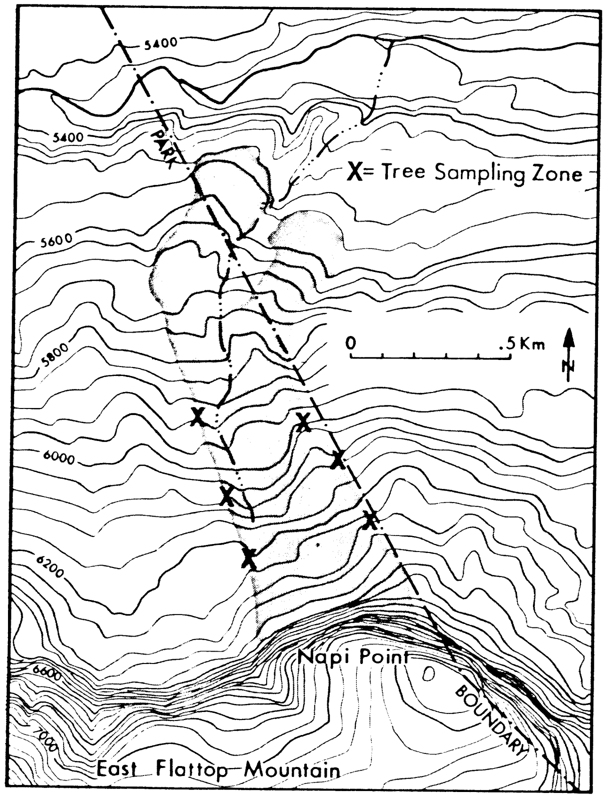


Figure 11. The Napi Point Deposit (From Butler, 1983a).

the deposit is approximately 0.52 $km^2$. A ridge and furrow morphology is present on the deposit, and several transverse ridges are found. Most of these ridges extend at least across the eastern two-thirds of the deposit. These ridges generally are less than 10 m high, although two may reach heights closer to 15-20 m. This surficial morphology resembles that of a rock glacier, but the presence of a large headwall scar and "jigsaw" boulders indicates the deposit is a rockfall deposit, and not a rock glacier.

A large headwall cliff scar on Napi Point is present above the entire Napi Point deposit. The cliff scar contains Appekunny argillite overlying Altyn limestone. This scar has a sheer vertical exposure approaching 90 degrees. Access to the cliff scar is very treacherous, because of the steep slopes and instability of the slope material; however, sampling of the number of fractures present along the cliff scar was completed in five areas. Five areas (one meter by one meter), chosen at random, were examined to determine the number of fractures present. The results, shown in Table I, indicate the cliff scar is heavily fractured. Unlike the Slide Lake deposit, there is no trough present at the base of the scar. Rather, the area at the base of the headwall is extremely steep. Slope characteristics for this deposit are shown in Figure 12. The break-in-slope method was again used to determine slope characteristics.

TABLE I

ROCK FRACTURES ON CLIFF SCAR AT NAPI POINT

| Site Number[a] | Rock Type | Number of Fractures Within Site[b] |
|---|---|---|
| 1 | Alytn Limestone | 22 |
| 2 | Altyn Limestone | 47 |
| 3 | Appekunny Argillite | 75 |
| 4 | Appekunny Argillite | 72 |
| 5 | Appekunny Argillite | 47 |

[a]Sample plots were one x one meter in size.

[b]Fractures were counted if they were a minimum of 30 cm in length.

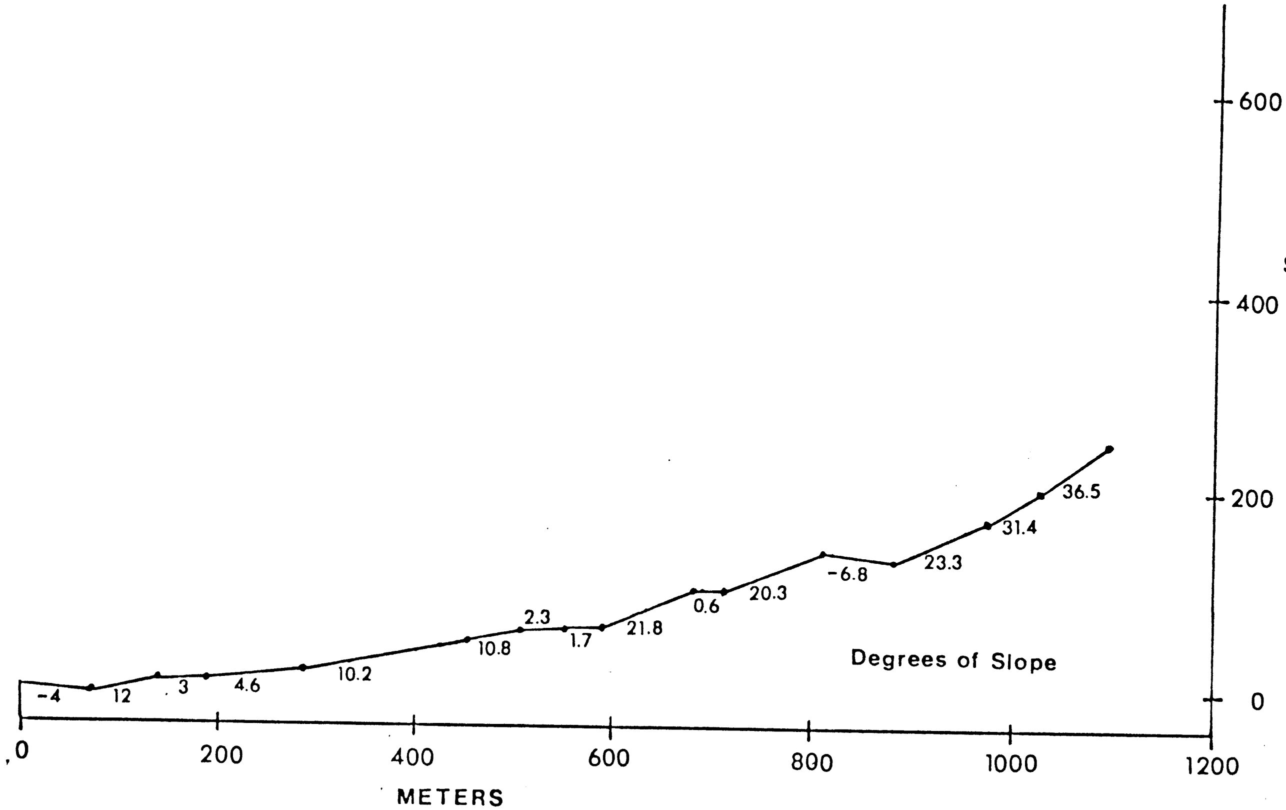


Figure 12. Slope Characteristics for Napi Point.

## Lithologic Features

The Napi Point deposit consists primarily of a mixture of Altyn limestone and Appekunny argillite, the same rock type as the cliff scar (Figure 13). Some Grinnell argillite possibly is present, although it is difficult to distinguish from the Appekunny argillite. The surface of this deposit is extremely unstable, and rocks are quite angular; caution was needed with each step taken. Rocks commonly gave way and moved when stepping on them.

The four main lithologic and geomorphic features described by Mudge (1965) were again present on this deposit. A few exceptionally large blocks were observed (Figure 14), but most of the deposit is a mixture of rock sizes (Figure 13), ranging from 2-3 cm to several square meters. Bounding ridges are approximately 8 m high, and 5 m wide. The upper one-half of the deposit consists of larger-sized rocks, while the lower one-half contains a mixture of rock sizes, including very fine-grained material.

## Soils

Very little soil development exists on the Napi Point deposit. The lower one-third of the deposit has fine-grained materials, which may represent a situation similar to the lower Lake lobe of the Slide Lake deposit. Soil development has probably resulted from the deposition of

Figure 13. Clast and Lithologic Mixture on the Napi Point Deposit. Note Axe for Scale.

Figure 14. Photo of Large Blocks on the Napi Point Deposit.

fine-grained material into gently-sloped areas via running water.

## Vegetation

Vegetation along the northern slopes of Napi Point consists primarily of subalpine fir, limber pine, and Englemann spruce. Vegetation on the upper half of the deposit is virtually nonexistent. The finer-grained, more gently sloping bottom half has considerable vegetation, such as small subalpine firs, shrubs, and annuals. The further downslope the deposit, the more dense the tree growth becomes.

# Dendrogeomorphic Analysis of the Slide Lake and Napi Point Rockfall Avalanches

## Introduction and Literature Review

One of the primary objectives of the field work was to determine the exact year of each rockfall avalanche. Such age determinations would aid in the determination of the hazard frequency. The technique employed in gathering data to determine the year of each event is called dendrogeomorphology, or the science of tree-ring dating applied to geomorphological processes. The basic principles of dendrochronology, described by Stokes and Smiley (1968), Fritts (1971), and Burrows and Burrows (1976), are discussed only briefly. Those concepts most relevant to dendrogeomorphology are discussed in more detail.

Most tree species can be useful indicators of past environmental events because they add one new layer (or ring) of cells for each year's growth. The amount of growth for each year is visible in the width of each ring of cells. Only trees which add one ring of new cells each year can be used in dendrochronology. Each annual ring consists of two types of cells: earlywood, which consists of large, thin-walled cells; and latewood, usually a darker color, with small, thick-walled cells (Burrows and Burrows, 1976). The earlywood and latewood together comprise one years' growth, or one annual ring. The boundaries between each separate year's growth can then be distinguished based on the type of cell. It is important that the distinction between each successive annual ring be clear, for that is the only way in which the year when changes or structural damage to a ring occurred can be determined. Fritts (1971) noted that the growing tissue (or annual rings) may be damaged by fire, avalanche, landslide, an erosional agent, and an approaching ice front of a glacier. Distortions or changes in ring structure may then be used to date these events.

## Concepts and Principles of Dendrogeomorphology

The underlying principle in the technique of dendrogeomorphology is that a geomorphic process (such as a landslide) which damages a tree will cause that tree to undergo structural changes in its annual rings. The annual ring in which change begins is considered the year in which the damage occurred, i.e. the year in which a geomorphic event occurred. Examples of geomorphic processes which have been studied using dendrogeomorphic techniques include flooding (Sigafoos, 1964; Yanosky, 1982); snow avalanches (Potter, 1969; Butler, 1979a; Carrara, 1979; Butler and Malanson, 1984); landslides (Langenheim, 1956; Moore and Mathews, 1978; Butler, 1979b); and glaciation (Bray and Struik, 1963). Because this study is concerned with landslide processes, the discussion will focus on the techniques and principles most appropriate to mass movement processes.

Shroder (1978) discussed the science of dendrogeomorphology. He outlined several basic dendrogeomorphological concepts; 1) uniformitarianism: the present is the key to the past in tree growth. Present-day physical and biological controlling factors must also have been operative in the past; 2) Limiting factors and site selection: useful data cannot be extracted from a tree not limited by the particular geomorphic process in consideration. Only trees which appear most likely to yield relevant data should be sampled; 3) Sensitivity: selected trees will appear sensitive or complacent to the particular process. Sensitivity may involve ring-width or ring-density variations, or cellular and morphologic differences known as event-response phenomenon; 4) Cross-dating: sufficient samples must be closely compared within trees and between trees; 5) Replication: cross checks and careful procedures based on size and type of sample will produce internally consistent results. Comparison of the dendrogeomorphic record with documented events, such as floods, landslides, and avalanches may provide the necessary calibration for verification.

The geomorphic process of landsliding will produce a number of events affecting trees. These events will produce different responses in the trees (Shroder, 1980). This system is known as the process-event-response system. The process of landsliding can produce several events. These events include tilting of trees, as debris strikes the tree and causes it to lean; and scarring, as rocks strike a tree and remove the bark. These events then cause responses in the growth patterns of the annual rings. These responses are what are inferred from dating the landslide event.

## Growth Responses of Tree Rings to Landslide Processes

The annual rings of a tree affected by a landslide can respond in different ways, depending on the event and the nature of the damage to the tree. These responses include: 1) production of reaction wood; 2) new callous growth; 3) growth suppression; and 4) growth release

(Shroder, 1980). When coniferous trees are tilted by a landslide, the tree begins to produce "reaction wood" on the downslope side of the tree. Deciduous trees tilted by a landslide produce reaction wood on the upslope side of the tree. Reaction wood is the structural response of a tree as it attempts to bring the tree back to a vertical position and strengthen its lower trunk. Reaction wood is generally a gray-brown or red-brown color, commonly resembling latewood. The cells are short, thick-walled, and dense. Eccentric growth occurs in the annual rings, with the greatest amount occurring on the downslope side. Reaction wood production usually begins immediately after a landslide event. The distinction between normal annual rings and annual rings having reaction wood is easily made because of the differences in color and cell density. Reaction wood is one of the most common event responses, and probably the most reliable (Shroder, 1980).

Trees which have been scarred by landslide debris produce morphological changes which also can be easily dated. Scars generally become covered with callous margin, younger wood, and bark; rings of newer wood meet stained and weathered older wood at an angle. Scarring events may also cause suppression of the annual rings. Suppression occurs when the affected tree becomes severely stressed from the landslide and experiences very little growth. Growth release may also result in the annual rings of trees following a landslide, usually as a result of reduced competition around the tree.

## Field Methods in Dendrogeomorphology

Annual ring samples from trees affected by a landslide must be obtained to employ the technique of dendrogeomorphology. The two most common methods of sampling include obtaining cores and crosscut sections of damaged trees. Cores are obtained by using an increment borer. An attempt is made to hit the pith of the tree, so that all the annual rings are sampled (Stokes and Smiley, 1968). Cores should be obtained from near the base of the tree, to ensure again that all the annual rings are sampled. Burrows and Burrows (1976) recommended obtaining a minimum of two cores per tree; Shroder (1978) recommended up to eight cores per tree to ensure replication of information. Core samples shoud be placed in a protective carrier, and properly labeled. On tilted coniferous trees, cores should be obtained from the downtilt side, so that the reaction wood information is sampled.

Crosscut sections involve the felling of trees and obtaining one or more sections of the tree. At least one section should be taken near the base of the tree to ensure all the annual rings are sampled. Crosscut sections should be properly labeled.

Obtaining core samples from a tree is preferred, because the tree is not damaged, and the samples are lightweight and compact. Laboratory analysis of the cores can be extremely difficult, however. Crosscut sections involve killing the tree, and are thus undesirable for environmental reasons. Sections are bulky and weigh considerably more than cores. Analysis of the annual rings in a section is much easier, however, because the entire history of the tree can be viewed in one sample.

Shroder (1978) emphasized the need to record adequate field data to enable a laboratory reconstruction of the field site. Field data to be recorded include the species sampled, height of tree, diameter of tree at breast height, lean direction and degree of lean, orientation and height of samples above the base of the tree, and any pertinent or unusual features of the tree or surrounding topography.

Samples must be adequately prepared before analysis can be completed. Cores should first be mounted by glueing the cores into a groove in wood or cardboard (Stokes and Smiley, 1968). Cores and crosscut sections should then be sanded, starting with a coarse or medium sandpaper, and progressing to a fine sandpaper as the sample becomes smoother. When properly prepared, the distinction between annual rings of some conifers, such as subalpine fir or Englemann spruce, is easily seen. Once the samples have been prepared, they are ready for laboratory analysis.

## Laboratory Methods in Dendrogeomorphology

Laboratory analysis involves examining the tree samples to determine the dates of growth responses, such as reaction wood and scars. This examination can be done most efficiently and accurately with a binocular microscope. Stereo-microscopic viewing eases the task of counting rings. Burrows and Burrows (1976) suggested wetting the rings with water or a permanent stain, because wetting will make the rings stand out for easier viewing.

Counting from the outermost ring inward, each decade year should be pinpricked or marked with a single dot, half centuries with two dots, and centuries with three dots (Burrows and Burrows, 1976). Shroder (1978) devised symbols for plotting the different growth responses into a "modified skeleton plot." Each type of growth response, such as reaction wood, scars, suppression, and release, has a particular symbol. Each year a growth response is evident in the annual rings, the proper symbol is placed on the skeleton plot for that year. This information, plotted for a minimum of two cores or radii per tree, provides a growth response history for that particular tree.

## Dendrogeomorphological Methods of This Study

Permission to obtain core and crosscut sections from trees damaged by the Napi Point and Slide Lake rockfall avalanches was granted by the National Park Service. Core and crosscut sections from 30 trees on each deposit were taken, using an increment borer and bowsaw. A minimum of two cores or sections from each tree were taken. Only those trees which appeared to have been affected by the rockfall avalanches were sampled (Figure 15). Location of the sampled trees is illustrated in Figures 5 and 11. This selection method involved trees which were tilted or had visible scars. Cores were obtained from tilted trees, near the tree base and at heights above the base. Sections were taken primarily from scarred trees, and included the scar itself. Sections were also obtained from some tilted trees.

Each sample was labeled with an identification number. Sections were also labeled to include the downslope direction. The information recorded in a logbook for each tree sampled included species, height of tree, height of sample from base of tree, diameter of tree at breast height, orientation of tree tilt, absence or presence of scars or unusual features, general location of tree in reference to the deposit, and any unusual topographical features. This information is provided in Appendix B. The trees which were sampled were found along the border of the rockfall avalanche deposit, along the "trimline" of the deposits. Trees near small seasonal streams were not sampled, because it was believed that tilting may have been a result of water movement rather than the rockfall event.

Figure 15. A Tilted and Scarred Tree From the Napi Point Deposit. Note Quarter for Scale.

The purpose of the tree-ring dating was to clearly establish the year in which the Slide Lake and Napi Point rockfalls occurred. Bracketing dates for each rockfall event are 1902-1914 for the Slide Lake deposit, and 1936-1959 for the Napi Point deposit. These dates are based on information from Butler (1983a) and aerial photo analysis. The age of samples from the Slide Lake deposit were plotted back to 1870; the Napi Point samples were plotted back to 1900. Samples were dated this far back on each site to ensure inclusion of the years in which the disturbance of trees by the rockfall events occurred. Information from two or three separate radii per crosscut section were plotted. Radii were drawn 45-180 degrees apart from each other. Cores obviously provided only one radii for plotting.

## Relative Age-Dating of the Slide Lake and Napi Point Rockfall Avalanches

### Introduction

Relative age-dating techniques of geomorphic processes, while not capable of providing an exact date, can provide a general time-frame for geologically recent geomorphic processes. Relative age-dating techniques become especially useful when absolute age-dating techniques cannot be utilized, or do not resolve a dating question. Although samples for dendrogeomorphological dating were available at both rockfall avalanche sites, two relative age-dating techniques were also used to test their applicability to this study, and to provide information in the event that the dendrogeomorphological sampling was inconclusive. These dating techniques included lichenometry and rock weathering rind analysis.

## Lichenometry

The technique of lichenometry was developed by Beschel in 1950. A manual describing the basic principles of lichenometry, sampling procedures, and applications of lichenometry was developed by Locke, Webber, and Andrews (1979). The basic principle in this technique is that the size of lichens growing on rocks on a deposit can indicate the relative age of the deposit. Lichens grow extremely slowly, and are capable of establishment immediately after the exposure and stabilization of a rock surface, in crevices or irregularities of the rock surface (Beschel, 1973). Lichens can achieve this establishment either by means of spores or by means of fragments of older stock. At the maximum, then, lichens may be as old as the rock surface itself. Growth is slow when the lichen is very young, followed by a period of accelerated growth until the lichen thalli have a circular outline. At this point growth continues at a much slower, yet constant rate.

Several factors may affect the growth rate of lichens (Beschel, 1973; Luckman, 1977). The climatic environment exerts a major control on rates of lichen growth, especially moisture availability. Saturated conditions are essential for lichen growth. Snowmelt, dew, and rain provide moisture for lichen growth. Lichens also need a considerable amount of light. While altitude itself does not affect lichen growth, other critical factors include temperature, duration of snowcover, and characteristics of the host rock (Luckman, 1977). The host rock must have a rough surface for spores to adhere to. More importantly, the rock must be stabilized; lichens provide a measure only of how long the surface has been immobile. The chemical nature and pH of the rock are also limiting factors to lichen growth. Beschel (1973) concluded that if all these limiting factors remain approximately constant, especially moisture amount and availability, the speed of growth and increase in size of a particular lichen species will fluctuate little.

Lichenometric techniques have been used extensively in dating Holocene glacial moraines (Benedict, 1967; Beshcel, 1973; Osborn and Taylor, 1975; Luckman, 1977). Birkeland (1973) used lichenometry to date rock glacier deposits, and Porter and Orombelli (1981) used lichenometry to date large rockfall deposits.

Several lichen species have been used in lichen dating. The most commonly used species is Rhizocarpum geographicum, favored by many because of its slow growth rate and wide geographic distribution (Osborn and Taylor, 1975). Studies utilizing R. geographicum have been conducted primarily on siliceous substrates. Unfortunately, *R. geographicum* is not tolerant of calcium in carbonate rocks, which comprise much of the substrate on the Slide Lake and Napi Point deposits. The only lichen species common enough for lichen analysis on calcareous substrates in the northern Rockies is *Xanthoria elegans*. This species is bright orange, with round or oval thalli. Osborn and Taylor (1975) conclude that although *X. elegans* growth can be disrupted by several factors, the species can still be useful for dating purposes if sampling caution is followed. Possible growth-disruptive factors include: 1) growth is influenced by the presence of animal dung; 2) parts of some thalli weather away before a maximum size is attained; 3) thalli may thicken as well as grow laterally with age; and 4) the lichen is not abundant in some cases, because of unexplained environmental reasons.

## Lichen Methodology

The minimum diameters from the largest lichen thalli which are found on rocks on a deposit are measured and recorded (in millimeters). Osborn and Taylor (1975) also suggested recording the size and type of the host boulder. These authors measured only round or oval thalli, and recorded only the minimum diameters on oval thalli. An alternative to measuring individual lichen size is to measure the percent of lichen cover on a rock.

Osborn and Taylor (1975) suggested this technique may be useful on forested deposits.

Once the minimum diameter of the largest lichen thallus has been measured, this value can be located on a lichen growth curve, which is simply a graph of thallus diameter vs. time. Lichen growth curves are developed by measuring thallus diameters on deposits of known age. A curve can be extrapolated between lichen measurements from two differently-aged deposits. Lichen measurements from a deposit of unknown age can then be located on this growth curve. Debate exists whether to apply the single lichen thallus with the largest diameter to the growth curve, or to average the diameter size of several large thalli and apply that mean to a growth curve. Webber and Andrews (1973), Osborn and Taylor (1975), and Luckman (1977) recommended using the single largest lichen thallus for a growth curve, based on the belief that the largest thallus represents the lichen growing under the optimum conditions. Smaller lichens are either late colonizers or slower growing individuals. Matthews (1973) recommended averaging the size of the largest thalli, in order to exclude any older thalli which may have been transported to a moraine.

In this study, 44 Xanthoria elegans lichens were measured on the Slide Lake deposit (no lichens were found on the Napi Point deposit). Sizes were recorded in millimeters. Most lichens had an oval or elliptical growth; therefore, the minimum diameter was recorded. No sampling was done in depressions on the surfaces of the rockfall deposits, where longer snowcover may have inhibited growth. Most lichens were found toward the center of each lobe; very few were found near the edges of the lobes. Factors disrupting lichen growth mentioned by Osborn and Taylor (1975) were noted, and sampling was conducted with these factors in mind. The largest minimum diameter measured from one thallus was compared to the lichen growth curve established by Osborn and Taylor (1975) for calcareous substrates in the middle Canadian Rockies. This growth curve is the only known curve available for *X. elegans*.

## Rock Weathering Rinds

Rock weathering rinds are another relative age-dating technique which can be applied to rockfall avalanche deposits. This technique has been discussed in several studies (Thorn, 1975; Chinn, 1981; Colman, 1981). Burke and Birkeland (1979) defined a rock weathering rind as a zone of discolored weathered rock that parallels the outer surface of the rock. The discoloration results from the weathering of iron-bearing minerals. The thickness of the rind is a rough measure of the duration of weathering after exposure of the rock, such as following a rockfall (Birkeland, 1973). Rock weathering rinds can thus be used as a means to date a rock deposit. Colman and Pierce

(1981) concluded that the age of a deposit represented the most important source of variation in rind thickness. This conclusion indicates the applicability of rock weathering rinds in determining the relative age of a rock deposit.

Rock Weathering Rind Methodology

Rocks are cracked open to examine the thickness of the weathering rinds. Sample size can range from 25 (Thorn, 1975) to well over 100 (Butler, 1982). Maximum rind thicknesses are measured and recorded. Some workers use the single thickest rind as a measure of rock weathering rates (Birkeland, 1973; Carroll, 1974). Chinn (1981) cautioned that this method overemphasizes those rocks most susceptible to rind development. Thorn (1975) and Chinn (1981) used the mean maximum rind thickness as the age indicator of duration of rock weathering rind development. Whichever method is used, the rind thickness measurements are used to develop a weathering rind curve.

In this study, 30 rocks (Altyn limestone) were sampled on the Slide Lake deposit, and 60 rocks (Appekunny argillite) were sampled on the Napi Point deposit. Rocks were sampled randomly over the surface of the entire deposit on each site. Rocks were not sampled in topographic depressions, where late-lying snowcover may have affected rind development.

# Analysis of the Historical Record and Dating Information

## Historical Records of the Slide Lake and Napi Point Events

Very few historical records exist concerning the date of either deposit. No historical records were found which could document the Napi Point deposit. Bracketing dates for the Slide Lake deposit are 1902-1914, based on historical reference (Campbell, 1914). Aside from these bracketing dates, one other historical reference possibly exists, depending on the interpretation of the document. Otatso Creek, which drains the Slide Lake valley, was originally called the North Fork of Kennedy Creek in the early decades of this century. Prior to the establishment of this area as part of Glacier National Park in 1910, mining claims were allowed in the area. A mining company from Canada was working a mine two km from the present head of Slide Lake, or three km west of the Lake lobe of the deposit. Slide Lake did not exist when the mining operations were begun. In a letter dated May 29, 1912, J.H. VanPelt, owner of the mining company, requested of Glacier Park officials the permission to continue construction and improvement on a trail, in an attempt to upgrade it to a wagon road. VanPelt wrote, "...the old trail having been badly washed in the last two years and muskegs have formed by slides that are well nigh impassable." This reference ("last two years") might indicate slide activity in 1910. As noted earlier, an area similar to a muskeg still exists at the foot of the Pond lobe in the area where the road was built. In another letter, dated July 15, 1912, Van Pelt requested permission again to build bridges and fix up the road built into the mine. He wrote "The land slides of late has (sic) made the road very near impassable." This reference ("landslides of late") is less clear about the time of slide activity, but again indicates that slide activity had been occurring prior to 1912.

## Analysis of Tree Ring Data

Plates 2-5 show the tree ring data presented in a modified skeleton plot. Although 30 trees per site were sampled, it was discovered after preparation and lab analysis that not all samples were useable. Core samples were especially troublesome. Cores were extemely hard to extract from trees, and the samples which were obtained often were broken and mangled, rendering them useless for analysis. Some crosscut sections contained rot which obliterated the tree ring record, and some had fungal growth, making tree ring interpretation difficult. Only those samples which could be confidently interpreted are shown on the modified skeleton plot. This methodological regimen ensures that only those samples with a clear annual ring history are plotted and used in analysis.

This approach reduced the tree ring sample size on each deposit. On the Slide Lake deposit, 23 separate trees provided data back until 1926; 22 trees provided data until 1914; and 21 trees provided data for the bracketing years of 1902-1914. On the Napi Point deposit, tree samples from 20-23 trees were useful for different years in the bracketing dates of 1936-1959.

Shroder (1978) emphasized the need for cross replication within trees. In his studies, as in this one, an event response was considered valid only if it occurred in two or more separated areas in a tree. Trees which had been cored needed to show replication in two separate cores to be considered valid; trees which had been sectioned needed to show replication in two separate radii of the section, or one radii each from two or more sections.

The numbers of trees exhibiting valid growth responses for the Slide Lake and Napi Point deposits are shown in Tables II and III, respectively. For each year, the number of trees providing samples, the number of valid growth responses, and the percentage of trees providing growth responses are listed. As mentioned previously, the types of growth responses plotted include reaction wood, scars, suppression, and release.

Potter (1969), Smith (1974), and Carrara (1979), agreed that the first two responses (reaction wood and scars) are the most reliable criteria for dating a mass

TABLE II

INDEX VALUES FOR THE SLIDE LAKE TREE RING SAMPLES

| Year | Number of Trees Living in Sample Year | Number of Trees With Growth Responses | Index Value | Trees With Reaction Wood and Scar Responses | Index Value |
|---|---|---|---|---|---|
| 1983 | 23 | 2 | 9 | 0 | 0 |
| 1982 | 23 | 5 | 22 | 0 | 0 |
| 1981 | 23 | 1 | 4 | 0 | 0 |
| 1980 | 23 | 0 | 0 | 0 | 0 |
| 1979 | 23 | 0 | 0 | 0 | 0 |
| 1978 | 23 | 4 | 17 | 1 | 4 |
| 1977 | 23 | 2 | 9 | 1 | 4 |
| 1976 | 23 | 6 | 26 | 1 | 4 |
| 1975 | 23 | 2 | 9 | 1 | 4 |
| 1974 | 23 | 7 | 30 | 0 | 0 |
| 1973 | 23 | 4 | 17 | 0 | 0 |
| 1972 | 23 | 4 | 17 | 3 | 13 |
| 1971 | 23 | 1 | 4 | 0 | 0 |
| 1970 | 23 | 1 | 4 | 1 | 4 |
| 1969 | 23 | 8 | 35 | 1 | 4 |
| 1968 | 23 | 4 | 17 | 1 | 4 |
| 1967 | 23 | 3 | 13 | 1 | 4 |
| 1966 | 23 | 0 | 0 | 0 | 0 |
| 1965 | 23 | 10 | 43 | 2 | 9 |
| 1964 | 23 | 6 | 26 | 3 | 13 |
| 1963 | 23 | 2 | 9 | 1 | 4 |
| 1962 | 23 | 3 | 13 | 1 | 4 |
| 1961 | 23 | 3 | 13 | 1 | 4 |
| 1960 | 23 | 2 | 9 | 1 | 4 |
| 1959 | 23 | 2 | 9 | 1 | 4 |
| 1958 | 23 | 5 | 22 | 1 | 4 |
| 1957 | 23 | 4 | 17 | 1 | 4 |
| 1956 | 23 | 3 | 13 | 1 | 4 |
| 1955 | 23 | 5 | 22 | 1 | 4 |
| 1954 | 23 | 5 | 22 | 3 | 13 |
| 1953 | 23 | 6 | 26 | 2 | 9 |
| 1952 | 23 | 5 | 22 | 1 | 4 |
| 1951 | 23 | 6 | 26 | 1 | 4 |
| 1950 | 23 | 2 | 9 | 0 | 0 |
| 1949 | 23 | 5 | 22 | 1 | 4 |
| 1948 | 23 | 3 | 13 | 1 | 4 |
| 1947 | 23 | 13 | 57 | 5 | 22 |
| 1946 | 23 | 16 | 70 | 15 | 65 |
| 1945 | 23 | 1 | 4 | 1 | 4 |
| 1944 | 23 | 2 | 9 | 1 | 4 |
| 1943 | 23 | 1 | 4 | 1 | 4 |
| 1942 | 23 | 1 | 4 | 1 | 4 |
| 1941 | 23 | 4 | 17 | 0 | 0 |
| 1940 | 23 | 4 | 17 | 1 | 4 |
| 1939 | 23 | 6 | 26 | 1 | 4 |
| 1938 | 23 | 1 | 4 | 0 | 0 |
| 1937 | 23 | 2 | 9 | 0 | 0 |
| 1936 | 23 | 3 | 13 | 0 | 0 |
| 1935 | 23 | 5 | 22 | 2 | 9 |
| 1934 | 23 | 4 | 17 | 2 | 9 |
| 1933 | 23 | 2 | 9 | 1 | 4 |
| 1932 | 23 | 2 | 9 | 1 | 4 |
| 1931 | 23 | 3 | 13 | 0 | 0 |
| 1930 | 23 | 5 | 22 | 4 | 17 |

TABLE II (CONTINUED)

| Year | Number of Trees Living in Sample Year | Number of Trees With Growth Responses | Index Value | Trees With Reaction Wood and Scar Responses | Index Value |
|---|---|---|---|---|---|
| 1929 | 23 | 0 | 0 | 0 | 0 |
| 1928 | 23 | 6 | 26 | 2 | 9 |
| 1927 | 23 | 3 | 13 | 1 | 4 |
| 1926 | 23 | 3 | 13 | 1 | 4 |
| 1925 | 22 | 4 | 18 | 2 | 9 |
| 1924 | 22 | 1 | 5 | 0 | 0 |
| 1923 | 22 | 1 | 5 | 1 | 5 |
| 1922 | 22 | 0 | 0 | 0 | 0 |
| 1921 | 22 | 4 | 18 | 0 | 0 |
| 1920 | 22 | 6 | 27 | 0 | 0 |
| 1919 | 22 | 4 | 18 | 0 | 0 |
| 1918 | 22 | 6 | 27 | 0 | 0 |
| 1917 | 22 | 6 | 27 | 0 | 0 |
| 1916 | 22 | 2 | 9 | 1 | 5 |
| 1915 | 22 | 5 | 23 | 1 | 5 |
| 1914 | 22 | 4 | 18 | 1 | 5 |
| 1913 | 21 | 6 | 29 | 5 | 24 |
| 1912 | 21 | 3 | 14 | 0 | 0 |
| 1911 | 21 | 2 | 10 | 2 | 10 |
| 1910 | 21 | 5 | 24 | 4 | 19 |
| 1909 | 21 | 5 | 24 | 2 | 10 |
| 1908 | 21 | 2 | 10 | 1 | 5 |
| 1907 | 21 | 0 | 0 | 0 | 0 |
| 1906 | 21 | 2 | 10 | 0 | 0 |
| 1905 | 21 | 1 | 5 | 1 | 5 |
| 1904 | 21 | 0 | 0 | 0 | 0 |
| 1903 | 21 | 2 | 10 | 2 | 10 |
| 1902 | 21 | 2 | 10 | 1 | 5 |
| 1901 | 21 | 1 | 5 | 1 | 5 |
| 1900 | 21 | 7 | 33 | 3 | 14 |
| 1899 | 21 | 1 | 5 | 0 | 0 |
| 1898 | 21 | 2 | 10 | 2 | 10 |
| 1897 | 21 | 6 | 29 | 2 | 10 |
| 1896 | 21 | 4 | 19 | 1 | 5 |
| 1895 | 21 | 3 | 14 | 1 | 5 |
| 1894 | 21 | 3 | 14 | 0 | 0 |
| 1893 | 21 | 1 | 5 | 1 | 5 |
| 1892 | 21 | 0 | 0 | 0 | 0 |
| 1891 | 21 | 3 | 14 | 0 | 0 |
| 1890 | 21 | 4 | 19 | 2 | 10 |
| 1889 | 21 | 1 | 5 | 0 | 0 |
| 1888 | 21 | 1 | 5 | 0 | 0 |
| 1887 | 21 | 0 | 0 | 0 | 0 |
| 1886 | 21 | 3 | 14 | 1 | 5 |
| 1885 | 21 | 0 | 0 | 0 | 0 |
| 1884 | 21 | 2 | 10 | 1 | 5 |
| 1883 | 21 | 6 | 29 | 1 | 5 |
| 1882 | 21 | 3 | 14 | 1 | 5 |
| 1881 | 21 | 0 | 0 | 0 | 0 |
| 1880 | 21 | 3 | 14 | 1 | 5 |
| 1879 | 21 | 2 | 10 | 2 | 10 |
| 1878 | 21 | 0 | 0 | 0 | 0 |
| 1877 | 21 | 0 | 0 | 0 | 0 |
| 1876 | 21 | 2 | 10 | 2 | 10 |
| 1875 | 21 | 0 | 0 | 0 | 0 |
| 1874 | 21 | 1 | 5 | 0 | 0 |
| 1873 | 21 | 2 | 10 | 0 | 0 |
| 1872 | 21 | 2 | 10 | 1 | 5 |
| 1871 | 21 | 3 | 14 | 2 | 10 |
| 1870 | 21 | 2 | 10 | 0 | 0 |

TABLE III

INDEX VALUES FOR THE NAPI POINT TREE RING SAMPLES

| Year | Number of Trees Living in Sample Year | Number of Trees With Growth Responses | Index Value | Trees With Reaction Wood and Scar Responses | Index Value |
|---|---|---|---|---|---|
| 1983 | 23 | 0 | 0 | 0 | 0 |
| 1982 | 23 | 7 | 30 | 0 | 0 |
| 1981 | 23 | 2 | 9 | 0 | 0 |
| 1980 | 23 | 2 | 9 | 0 | 0 |
| 1979 | 23 | 0 | 0 | 0 | 0 |
| 1978 | 23 | 2 | 9 | 0 | 0 |
| 1977 | 23 | 0 | 0 | 0 | 0 |
| 1976 | 23 | 1 | 4 | 0 | 0 |
| 1975 | 23 | 2 | 9 | 0 | 0 |
| 1974 | 23 | 3 | 13 | 1 | 4 |
| 1973 | 23 | 4 | 17 | 3 | 13 |
| 1972 | 23 | 7 | 30. | 4 | 17 |
| 1971 | 23 | 6 | 26 | 1 | 4 |
| 1970 | 23 | 2 | 9 | 0 | 0 |
| 1969 | 23 | 4 | 17 | 3 | 13 |
| 1968 | 23 | 1 | 4 | 0 | 0 |
| 1967 | 23 | 2 | 9 | 1 | 4 |
| 1966 | 23 | 1 | 4 | 1 | 4 |
| 1965 | 23 | 1 | 4 | 0 | 0 |
| 1964 | 23 | 5 | 22 | 1 | 4 |
| 1963 | 23 | 7 | 30 | 2 | 9 |
| 1962 | 23 | 9 | 39 | 3 | 13 |
| 1961 | 23 | 8 | 35 | 1 | 4 |
| 1960 | 23 | 3 | 13 | 0 | 0 |
| 1959 | 23 | 1 | 4 | 0 | 0 |
| 1958 | 23 | 2 | 9 | 0 | 0 |
| 1957 | 23 | 1 | 4 | 0 | 0 |
| 1956 | 23 | 4 | 17 | 3 | 13 |
| 1955 | 23 | 4 | 17 | 4 | 17 |
| 1954 | 23 | 19 | 83 | 19 | 83 |
| 1953 | 22 | 2 | 9 | 0 | 0 |
| 1952 | 22 | 2 | 9 | 0 | 0 |
| 1951 | 22 | 2 | 9 | 1 | 5 |
| 1950 | 22 | 4 | 18 | 2 | 9 |
| 1949 | 21 | 1 | 5 | 0 | 0 |
| 1948 | 21 | 2 | 10 | 2 | 10 |
| 1947 | 21 | 1 | 5 | 1 | 5 |
| 1946 | 21 | 2 | 10 | 2 | 10 |
| 1945 | 21 | 0 | 0 | 0 | 0 |
| 1944 | 20 | 0 | 0 | 0 | 0 |
| 1943 | 20 | 2 | 10 | 1 | 5 |
| 1942 | 20 | 0 | 0 | 0 | 0 |
| 1941 | 20 | 1 | 5 | 0 | 0 |
| 1940 | 20 | 1 | 5 | 0 | 0 |

TABLE III (CONTINUED)

| Year | Number of Trees Living in Sample Year | Number of Trees With Growth Responses | Index Value | Trees With Reaction Wood and Scar Responses | Index Value |
|---|---|---|---|---|---|
| 1939 | 20 | 0 | 0 | 0 | 0 |
| 1938 | 20 | 0 | 0 | 0 | 0 |
| 1937 | 20 | 3 | 15 | 1 | 5 |
| 1936 | 20 | 0 | 0 | 0 | 0 |
| 1935 | 20 | 0 | 0 | 0 | 0 |
| 1934 | 20 | 1 | 5 | 1 | 5 |
| 1933 | 20 | 0 | 0 | 0 | 0 |
| 1932 | 20 | 0 | 0 | 0 | 0 |
| 1931 | 20 | 1 | 5 | 0 | 0 |
| 1930 | 20 | 3 | 15 | 0 | 0 |
| 1929 | 19 | 0 | 0 | 0 | 0 |
| 1928 | 19 | 0 | 0 | 0 | 0 |
| 1927 | 19 | 1 | 5 | 0 | 0 |
| 1926 | 19 | 6 | 32 | 1 | 5 |
| 1925 | 19 | 1 | 5 | 0 | 0 |
| 1924 | 15 | 1 | 7 | 1 | 7 |
| 1923 | 15 | 1 | 7 | 0 | 0 |
| 1922 | 15 | 0 | 0 | 0 | 0 |
| 1921 | 15 | 1 | 7 | 0 | 0 |
| 1920 | 15 | 0 | 0 | 0 | 0 |
| 1919 | 15 | 1 | 7 | 0 | 0 |
| 1918 | 15 | 1 | 7 | 0 | 0 |
| 1917 | 15 | 0 | 0 | 0 | 0 |
| 1916 | 15 | 0 | 0 | 0 | 0 |
| 1915 | 15 | 0 | 0 | 0 | 0 |
| 1914 | 15 | 1 | 7 | 0 | 0 |
| 1913 | 15 | 2 | 13 | 0 | 0 |
| 1912 | 15 | 3 | 20 | 3 | 20 |
| 1911 | 15 | 0 | 0 | 0 | 0 |
| 1910 | 15 | 1 | 7 | 0 | 0 |
| 1909 | 14 | 3 | 21 | 0 | 0 |
| 1908 | 14 | 0 | 0 | 0 | 0 |
| 1907 | 14 | 0 | 0 | 0 | 0 |
| 1906 | 14 | 1 | 7 | 0 | 0 |
| 1905 | 14 | 1 | 7 | 0 | 0 |
| 1904 | 14 | 0 | 0 | 0 | 0 |
| 1903 | 14 | 0 | 0 | 0 | 0 |
| 1902 | 14 | 0 | 0 | 0 | 0 |
| 1901 | 14 | 0 | 0 | 0 | 0 |
| 1900 | 14 | 0 | 0 | 0 | 0 |

movement event. With this in mind, the last two columns of Tables II and III contain the number and percentage of trees which exhibited either reaction wood or scars; release and suppression responses were not included. This effort was made to compare the results of trees using all growth responses with those exhibiting only reaction wood and scars. Growth release and suppression commonly are indicators of climatic stress or change (Fritts, 1971); their use in dating landslide processes is perhaps more limited.

## Index Numbers and Critical Cutoff Values

Before the date of each deposit can be determined, a problem remains of deciding how many responses indicate a valid past occurrence of a landslide event. An index number for each year can easily be determined simply by dividing the number of trees providing a sample by the number of trees with a valid response. This percentage (or index number; Shroder, 1978) indicates the number of trees which exhibit a growth response for each year. The key problem then becomes the establishment of a minimum percentage of valid responses (the lowest index number) needed for any given year to indicate occurrence of a landslide event. This problem is unresolved in dendrogeomorphology. Moore and Mathews (1978) used scars from only two trees to date a large landslide in British Columbia, although they had some historical references to supplement this date. LaMarche and Wallace (1972) dated the movement of a fault on the San Andreas fault based on growth responses from one tree. Some snow avalanche studies, such as Potter (1969) and Carrara (1979), do not discuss a cutoff value. Years of snow avalanche activity are determined on the basis of higher index numbers. Butler and Malanson (in review, 1984) used a 40% index number to indicate snow avalanche events. This number is based on growth responses in years in which historical records clearly illustrate avalanching occurred.

Clearly no standard exists for establishing cutoff values before conclusions of geomorphic activity are made. The practice of assigning dates of geomorphic activity based on high index values is not even followed. This absence of standarization makes the determination of a cutoff value for the Slide Lake and Napi Point deposits difficult, and somewhat arbitrary.

### Calculating a Cutoff Value for This Study

Historical records for the Napi Point event do not exist; only minimal historical references exist for the Slide Lake deposit. The bracketing dates for the Slide Lake deposit (1902-1914) can be examined to determine an index cutoff value. The highest index value for any year during this period is 29%, in 1913 (24% based on reaction wood and scars). This date, however, does not match the historical reference by Campbell (1914), which states the landslide occurred "a few years ago." Two years in this time frame had index values of 24% (1909-1910). The year 1912 had an index value of 14%. Considering the index value computed from reaction wood and scar responses only, only 1910 (19%) and 1909 and 1911 (10%) show any appreciable values. In sum, although the years 1902-1914 provide the best opportunity to provide a useful cutoff value, none of the years have an index value high enough to confidently determine a cutoff value.

On both the Napi Point and Slide Lake rockfall avalanches, a definite trimline was created on both sides of all lobes. This trimline is common to snow avalanches as well. It can be inferred, then, that many trees along the trimline of each deposit may have suffered some damage from the rockfall activity, much the same as trees along snow avalanche trimlines. The 40% cutoff value which Butler and Malanson (1984) have applied to snow avalanches may be an appropriate value to apply to rockfall avalanches. For lack of any historical precedence in determining cutoff values for landslide events, 40% was tested as the critical cutoff value for indicating the year a rockfall event occurred. The results from using this value are discussed below.

Tree Ring Dating the Slide Lake Deposit

A glance at Table II indicates that most years exhibit some type of growth response. Most of these responses are growth release and suppression, however. The relationship between these responses and fluctuating climate have been mentioned; these responses may represent "climatic noise" rather than a response to a landslide event. Low-magnitude rockfall events are a common daily occurrence in mountainous areas (Luckman, 1976; Butler, 1983b; Gardner *et al.*, 1983). It seems unlikely, however, that isolated rockfalls are large enough to travel several hundred meters downslope, and are capable of inducing structural damage to as many trees as Table II indicates. Many of the growth responses, especially release and suppression, may indeed be responses to other variables, such as climate. Because of this situation, the index values for reaction wood and scar responses may be most appropriate to consider when determining landslide event dates. In this study, both index values will be discussed, although primary importance will be given to the index values obtained only from reaction wood and scar responses.

Using a 40% cutoff value, only three years have index values exceeding the cutoff, none of which occur within the bracketing years of 1902-1914. These are the years 1946, 1947, and 1965. If only reaction wood and scar responses are examined, only one of those years, 1946, exceeds the cutoff value. In 1946, 65% of the sampled trees indicate a reaction wood or scar response. The year 1947 has a relatively high index value, and probably represents a continuation of responses from 1946, or it may represent delayed responses to an event. The year 1965 has an index number (43%) exceeding the cutoff value when considering all responses; however, it has a low index value if only reaction wood and scar responses are considered (9%). Most of the growth responses involved release, a response which can be attributed to many non-mass movement variables. Three other years have index values ranging from 30-40% (1900, 1969, and 1974). The years 1965 and 1974 were years of heavy snow avalanche activity throughout Glacier National Park (D. Butler, personal communication, 1984), suggesting climatic variables may have been responsible for the high index values determined for those years. Only 1900 has an appreciable index value based on reaction wood and scars (14%).

An examination of index values for scars and reaction wood only indicates that three other years have appreciable values. These years include 1910 (19%), 1913 (24%), and 1930 (17%). It is interesting that two of these years fall within the bracketing dates of 1902-1914.

It appears that on the Slide Lake deposit a major rockfall event occurred prior to, or in the early stages of, the 1946 growing season. This would probably include the time period from September 1945 through June 1946. Of the 23 trees providing samples, 15 (65%) had growth responses of either reaction wood or scars. In 1947, five of 23 trees (22%) had similar responses, probably a continuation of, or delayed response to, the 1946 event. Unfortunately, no other years exceeded the 40% threshold value, including the years 1902-1914. The years 1909-1913 show a cluster of reaction wood and scar responses (0-24%). Such a clustering is not found in any other period between 1870-1983, and perhaps indicates responses to some rockfall activity.

It is puzzling why the years from 1902-1914 do not show higher index values. Campbell (1914) stated a great mass of shale slid on this site between 1902-1914. Definite trimlines on each side of three lobes were formed, and many tilted trees are found along these trimlines. Yet with this time frame to work from, very few trees exhibit growth responses. Age data (Plates 2-3) reveal that these trees were living prior to 1900; therefore, these trees apparently did not respond to the rockfall event. Only those trees which appeared to have suffered some type of structural damage were sampled; either the trees with a useful tree ring record were not sampled, or the major rockfall event between 1902-1914 does not show up well in the annual rings of the sampled trees. This last possibility is especially puzzling, considering that the 1946 event does show up often in the tree ring record. The possibility exists that debris from the 1946 rockfall event covered most of the trees which were damaged by the 1910 rockfall event. If this is the case, I may have been sampling trees which were slightly away from the trimline created by the 1910 rockfall, and was then sampling trees along the trimline created by the 1946 rockfall event.

Although none of the index values in the 1902-1914 time frame exceed 40%, 1910 (with an index value of 19%) probably represents the major rockfall event. The year 1913 has a higher index value (24%), but is discounted for reasons based on historical records (discussed previously). The 19% index value for 1910 is almost double that of all other years in this time frame. Based on this tree ring information, plus historical documentation, 1910 is the most likely date of the rockfall event at Slide Lake.

Many years not in the 1902-1914 period have small index values, which may indicate small rockfall events, although this cannot be confirmed. Many of these years had scars, which are usually a more reliable indication of a rockfall event. Customs officials a few km north of the Slide Lake deposit reported a large dust cloud in this area in 1969; this was later interpreted to be the result of a rockfall event (Ed Harp, personal communication, 1984). The index value for 1969, although high for all growth responses, is quite low for reaction wood and scar responses (4%).

### Tree Ring Dating the Napi Point Deposit

Table III lists the index values for samples obtained along the trimlines of the Napi Point deposit. Again using the 40% cutoff value, only one year exceeds the cutoff value. This year is 1954, when 19 of 23 trees (83%) exhibited growth responses. All 19 trees exhibited reaction wood and scar responses. This large index value, in conjunction with the fact that no other years have index values exceeding 40%, indicates that 1954 is the probable date of origin of the Napi Point rockfall avalanche. This date would include the time prior to the 1954 growing season (September 1953 through May 1954). The year 1955 has a 17% index value, probably indicating a continuation or delayed response to the 1954 event. Considering all growth responses, several years have index values exceeding 30% (1926, 1961-1963, 1972, and 1982).

Heavy snow avalanching occurred throughout Glacier Park in 1963, 1972, and 1982, again suggesting "climatic noise" was responsible for the high index values determined for those years. Only two of those years, 1962 (13%) and 1972 (17%) showed appreciable index values for reaction wood and scar responses. Aside from the high index value for 1954, only one year had an index value for reaction wood and scar responses of 20% or more. This year was 1912, with a value of 20%.

Considering index values for reaction wood and scar responses, only nine of the 83 years have index values of 10% or more, including the 1954-1955 years. Three separate trees in 1962 have scar responses, possibly indicating renewed rockfall activity (although the index value for 1962 is only 13%). The situation is similar in 1972 (17%), with two separate trees bearing scar responses.

Except for the bracketing dates of 1936-1959 for this deposit, no historical evidence exists for dating the major rockfall event. Retired Park ranger Bob Frausen recalls possible activity in 1959 and in 1961 or 1962 (Bob Frausen, personal communication, 1983; Ed Harp, personal communication, 1984). The tree ring record shows no indication of growth responses in 1959. One sampled tree was scarred in 1961, whereas three trees were scarred in 1962. A large rockfall event may have occurred in 1962, although affecting only a small area.

## Lichenometric Analysis on the Slide Lake Deposit

Forty-four *Xanthoria elegans* were sampled on the Slide Lake deposit. Results are shown in Table IV. A typical lichen sample is shown in Figure 16. Most lichens were oval in shape; the minimum diameter was recorded as the size of the lichen thalli (Osborn and Taylor, 1975). Of the 44 lichens measured, the largest minimum lichen

TABLE IV

LICHEN MEASUREMENTS ON THE SLIDE LAKE DEPOSIT

| Minimum Diameter Size (in Millimeters) | | | |
|---|---|---|---|
| 21 | 13 | 10 | 9 |
| 17 | 12 | 10 | 9 |
| 16 | 12 | 10 | 8 |
| 16 | 12 | 10 | 8 |
| 15 | 11 | 10 | 8 |
| 15 | 11 | 10 | 8 |
| 14 | 11 | 9 | 8 |
| 14 | 10 | 9 | 7 |
| 13 | 10 | 9 | 7 |
| 13 | 10 | 9 | 7 |
| 13 | 10 | 9 | 7 |

$\bar{x}$ Minimum = 10.9

Figure 16. Photo of Typical Lichen Sample on the Slide Lake Deposit.

diameter found was 21 mm. Larger lichen diameters were found on older, weathered rock, but could not be considered valid. Although movement of rocks generally disrupts lichen growth (Beschel, 1973), lichens can survive catastrophic rock movements (Shreve, 1966). It is assumed the large lichens (46 mm) found on the more weathered rock survived the rockfall event, and have continued to grow.

The largest minimum lichen diameter found on this site (21 mm) was applied to the _Xanthoria_ _elegans_ growth curve established by Osborn and Taylor (1975) for the Banff/Jasper National Park, Alberta, area (Figure 17). According to this curve, the age of the Slide Lake deposit is approximately 63 years old. This age would place the rockfall event at 1920. This date is slightly younger than the bracketing dates of 1902-1914. A short time period of stabilization is necessary before colonization can occur, and this ecesis period can vary from a few to several years. Osborn and Taylor (1975) established their lichen growth curve for an area ranging from 300-600 km northwest of this study site. Colonization and the variability of growth curves between sites may explain the difference between the bracketing dates and the date indicated by lichenometry.

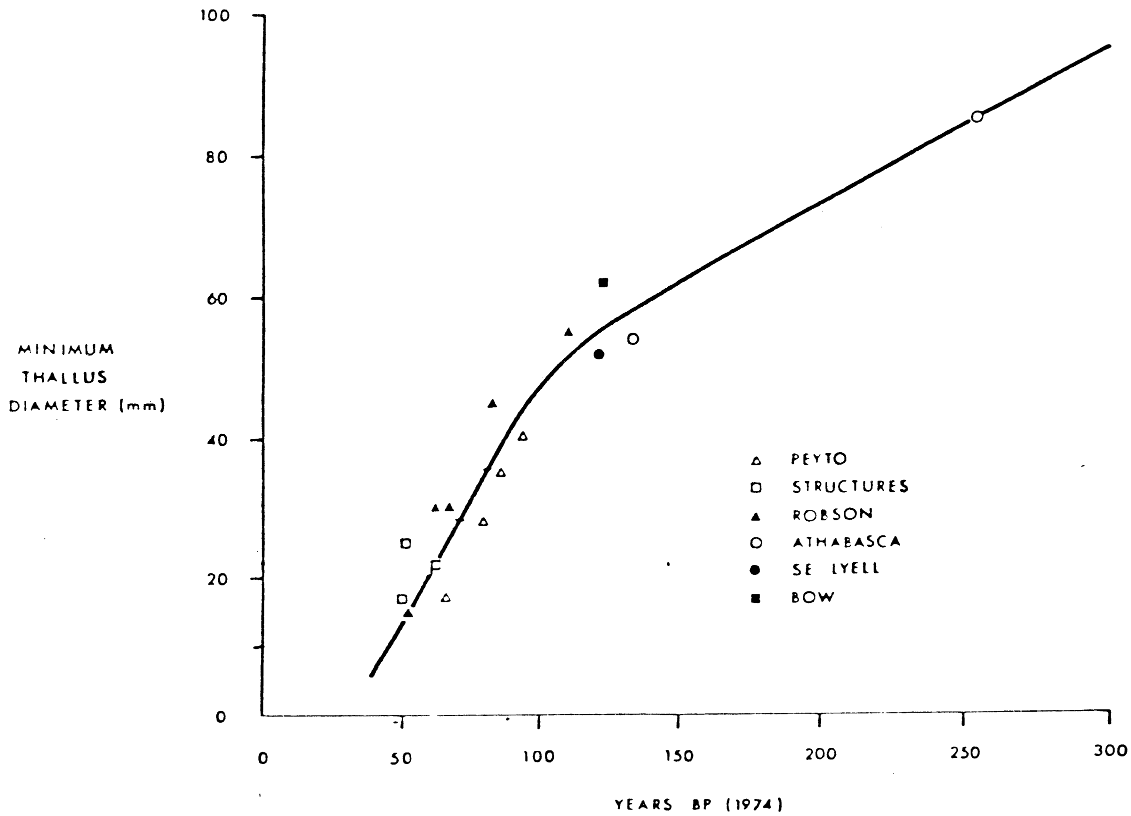


Figure 17. The Osborn and Taylor (1975) Lichen Growth Curve for Xanthoria Elegans.

## Rock Weathering Rind Analysis on the Napi Point and Slide Lake Deposits

Rocks on each deposit were cracked open to determine the thickness of the weathering rind. Table V shows the results. Almost no rind development has occurred in the sampled rocks. Only three of the 30 Slide Lake samples showed a measurable rind; on the younger Napi Point deposit, no rocks showed measurable rind development. A measurement of 0.1 mm was assigned to the three Slide Lake samples. The accuracy of the measurement of these three samples is questionable, as field measurement with a ruler having smallest units as millimeters may lead to erroneous measurements. Nonetheless, these three samples indicate that some rock weathering rind development is proceeding in the Altyn limestone rocks.

The size of rock weathering rinds from unknown deposits can be compared to a weathering rind curve. The deposits in this study were formed too recently, however, and the rind thicknesses were too small to compare to a weathering curve. This technique could perhaps be applied quite successfully on some of the older deposits shown in Plate 1.

TABLE V

ROCK WEATHERING RIND DATA

| Location | Number of Rocks Sampled | Number of Rocks With Measurable Rinds |
|---|---|---|
| Slide Lake | 33 | 3[a] |
| Napi Point | 60 | 0 |

[a]Size of rinds was 0.1 millimeters.

## Conclusions

The date of the Slide Lake rockfall avalanche is probably 1910, based on the tree ring record and historical information. The index value for 1910 falls short of the 40% index value used in this study, but is the largest index value of any in the 1902-1914 period. Historical references tend to support this date. Lichen dating of the deposit suggests a more recent age, and along with rock weathering rind data, should be considered only as a means of relative age-dating for this study. Further movement of the deposit occurred in 1946. Earthquake and precipitation data for the 1902-1914 and 1946 time periods will be considered in the next chapter in an attempt to further postulate the actual date of the Slide Lake rockfall avalanche occurrence.

The date of the Napi Point deposit appears to be 1954. The tree ring data for this deposit strongly suggest this date. No lichens were found on this deposit, nor had any rock weathering rind development occurred. This information suggests the deposit is of recent origin. Earthquake and precipitation data will also be examined for this site.

# CHAPTER V

# CAUSAL MECHANISMS OF ROCKFALL AVALANCHES

Examination and discussion of the causal mechanisms of rockfall avalanches are important with regard to the hazard implications of rockfall avalanches. Although other types of landslides (such as rockslides, slumps, and earthflows) are located near and along the Lewis thrust fault, only the causal mechanisms of rockfall avalanches will be discussed, for the two rockfall avalanches are the only landslide types for which dates of occurrence are known. Once the years of landslide activity are known, potential causal factors (such as seismic and precipitation data) can be examined to determine what factors may have triggered the landslide event.

## Literature Review

Several factors exist which may lead to rockfall avalanche activity. Terzaghi (1950) provided a discussion on the mechanisms of landslides. He states that landslides will occur if the average shearing stress (those forces working to destabilize a slope) on the potential surface of sliding equal the average shearing resistance (those forces working to maintain the stability of a slope). In short, a threshold exists at which stress forces are greater than resisting forces, and landslide activity occurs.

Sharpe (1938) distinguished the causes of landslides as: 1) the basic or passive conditions favoring landslides, and 2) the active or initiating causes of landslides. Examples of basic conditions include the presence of inherently weak rock formations, and badly fractured and jointed rock. Examples of active or initiating causes include the removal of support through oversteepening of cliffs by glacial activity, overloading of slopes through moisture saturation, reduction of friction along the slip plane through moisture saturation, the softening of weak rock masses by percolation of water, earth vibrations caused by earthquakes, freeze-thaw processes, and production of overly-steep constructional slopes by thrust faulting.

Several authors have concluded that the amount of fracturing and jointing within a rock formation is a very important factor leading to the initiation of rockfall/slide activity. Crandell and Fahnstock (1965) and Cooper (1980) noted that the amount of fracturing and jointing in rocks increases the instability of the rock. Mudge (1965) concluded that most rockfalls in the northern Rocky Mountains originate in carbonate rock, primarily because of the widely spaced joints in this rock. Clague (1981) similarly noted the importance of fractured bedrock in landslide activity. Gardner _et al_. (1983) noted that the existence of bedding planes and joints are factors contributing to rockfall events.

Freeze-thaw processes can increase the spaces between fractures and joints, further weakening rock formations. Cruden (1976) stated that freeze-thaw mechanisms are probably the most active means of enlarging surficial fractures in the mountain environment. Douglas (1980) noted that freeze-thaw action produced many microfractures in well-jointed rock in Ireland, which reduced the stability of the rock. Crandell and Fahnstock (1965), Cooper (1980), and Butler (1983b) similarly concluded that freeze-thaw action can contribute to the destabilization of rock formations.

Alt and Hyndman (1973), Clague (1981), Porter and Orombelli (1981), and Gardner et al. (1983) stated that oversteepening of slopes by glacial activity may remove the underlying support of rock cliffs, eventually contributing to rockfall/slide activity. Campbell (1914) and Butler (1983b) discussed the importance of precipitation in rockfall activity. Excess precipitation is more commonly associated with the production of mud- and debris-flows (Curry, 1966; Bogucki, 1976; and Rapp and Stromquist, 1976). Sharpe (1938), however, stated that moisture saturation can soften weak rock masses by percolation of water, leading to rock failure.

Earthquakes are frequently cited as the trigger mechanism initiating rockfall/slide avalanching events. A catastrophic rockslide in Peru, triggered by an earthquake, killed 20,000 people in 1973 (Browning, 1973). In 1959 a large earthquake triggered several landslides and rockfalls in southern Montana near Hebgen Lake, killing 28 people (Hadley, 1964). Several other authors have attributed rockfall/slide events to seismic activity (Mudge, 1965; Moore and Mathews, 1978; and Clague, 1981). Whitehouse (1981) concluded that most historical rock avalanches have a close association with high-magnitude earthquakes. Hewitt (1983) stated that rapid mass movements are a feature of so many earthquake disasters that they must be regarded as primary rather than secondary aspects of the earthquake hazard. In many cases, damage from the mass movement is more intensive and widespread than from the shaking of the earth.

Several causal mechanisms exist which lead to rockfall activity, but seldom, if ever, can a rockfall/slide event be attributed to a single cause (Varnes, 1958). Several factors lead to the development of a landslide, from the beginning of the rock formation through the various weathering processes affecting the rock, until some action sets the rock mass in motion downhill. The final factor often is nothing more than the trigger which sets in motion a rock mass already on the verge of failure. Earthquakes, while a common trigger mechanism, are usually only the final factor which initiates rockfall events.

## Causative Factors and Trigger Mechanisms of the Slide Lake Deposit

Undoubtedly several causal mechanisms led to the Slide Lake rockfall avalanche. The presence of the Lewis thrust fault was a primary factor. Highly jointed, fractured, and faulted Precambriam Altyn limestone and Appekunny argillite overlie weak Cretaceous mudstones. In the warm season, higher elevations can have cool evenings and warm days, while occasional Chinook winds in the cold season allow for warm days. These climatic features can produce extensive freeze-thaw activity. This freeze-thaw activity probably widened the fractures and joints, further destablizing the already highly fractured and jointed rock mass. The Otatso Creek valley was extensively glaciated during the Wisconsin stage of the Pleistocene, probably resulting in oversteepened slopes and rock faces (Alden, 1932). These several factors together contributed to the general instability of the rock mass prior to its release as a rockfall avalanche in 1910.

Campbell (1914) indicated that the trigger mechanism which led to the release of the Slide Lake rockfall avalanche was excessive precipitation. He wrote of the cloud burst activity of "a few years ago" on the east side of the mountains which caused all the streams to do great damage to their banks and floodplains. The water softened the shale on the slope south of Otatso Creek, and a great mass slid down, blocking the valley and forming the lake. Table VI shows the precipitation data for the Babb station for 1909-1910. As concluded in Chapter Four, 1910 is the accepted year for the Slide Lake rockfall avalanche. In 1910, excessive precipitation occurred in February (69.6 mm), March (42.9 mm), and September (109.7 mm). It is not known when in September the excessive rainfall occurred. If intense rainfall occurred in early September and triggered the rockfall avalanche, growth responses in trees may still have shown up in the 1910 annual rings. Generally, the summer months at Babb were well below the normal monthly precipitation levels. The Slide Lake deposit, approximately 13 km west of the Babb station, is located in the mountains, where precipitation probably exceeds that of the Babb station. Although the precipitation record for the Babb station in 1910 does not indicate excessive moisture in the summer months, the mountains to the west of the station (near Slide Lake) may have received intense precipitation while the Babb station did not. Campbell (1914) used the term "cloudbursts" to describe the precipition event, suggesting a more localized excessive precipitation event.

The possibility of earthquake activity contributing to the instability of the rock mass at Slide Lake should also be considered. Limited earthquake data exist for this general area for the early 1900's (Qamar and Stickney, 1983). Major earthquakes occurred in eastern

## TABLE VI

## PRECIPITATION DATA FOR BABB 6NE, MONTANA

| Year | Jan | Feb | Mar | Apr | May | Jun | Jul | Aug | Sep | Oct | Nov | Dec | Annual |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| 1909 | 68.6 | 19.0 | 14.5 | 121.4 | 53.1 | 91.2 | 154.0 | 0.0 | 40.6 | T | 24.6 | 31.7 | 618.7 |
| | +45.0 | -3.1 | -12.9 | +79.7 | -10.4 | -1.0 | +106.0 | -46.2 | -12.5 | -30.5 | 0.0 | +8.6 | +122.7 |
| 1910 | 17.8 | 69.6 | 42.9 | 39.9 | 43.7 | 41.9 | 15.7 | 38.6 | 109.7 | 16.8 | 19.8 | 0.2 | 456.7 |
| | -5.8 | +47.5 | +23.1 | -1.8 | -19.8 | -50.3 | -32.3 | -7.6 | +53.6 | -13.7 | -4.8 | -22.9 | -39.3 |
| 1945 | 7.6 | 23.9 | 39.9 | 66.5 | 82.5 | 139.7 | 17.0 | 31.5 | 72.1 | 13.7 | 15.7 | 2.8 | 525.8 |
| | -16.0 | +1.8 | +12.5 | -24.8 | +19.0 | +47.5 | -31.0 | -14.7 | +19.0 | -16.8 | -8.9 | -20.3 | +29.8 |
| 1946 | 20.6 | 12.7 | 46.7 | 16.8 | 77.0 | 55.4 | 52.8 | 43.4 | 47.0 | 117.1 | 35.1 | 22.6 | 549.7 |
| | -3.0 | -9.4 | +19.3 | -24.9 | +13.5 | -36.8 | +4.8 | -2.8 | -6.1 | +86.6 | +11.5 | -0.5 | +53.7 |
| 1953 | 56.9 | 38.9 | 38.6 | 90.4 | 103.6 | 181.6 | 14.9 | 15.2 | 40.9 | 8.4 | 11.2 | 41.1 | 641.8 |
| | +33.3 | +16.8 | +16.5 | +48.7 | +40.1 | +89.4 | -33.1 | -31.0 | -12.2 | -22.1 | -13.4 | +18.0 | +145.8 |
| 1954 | 56.4 | 32.0 | 43.2 | 69.8 | 12.4 | 30.5 | 8.1 | 87.4 | 81.3 | 7.6 | 3.0 | 4.3 | 461.5 |
| | +32.8 | +9.9 | +15.8 | +28.1 | -51.1 | -61.7 | -39.9 | +41.2 | +28.2 | -22.9 | -21.6 | -18.8 | -34.5 |
| Average Monthly Precipitation | 23.6 | 22.1 | 27.4 | 41.7 | 63.5 | 92.2 | 48.0 | 46.2 | 53.1 | 30.5 | 24.6 | 23.1 | 496.0 |

Precipitation measured in millimeters.
Top number is the measured amount.
Bottom number is the deviation from normal.

Sources: U.S. Department of Agriculture Weather Bureau, 1926.
U.S. Weather Bureau, unpublished weather statistics.
Butler, D.R., 1979a.

Montana in May, 1909, and in April, 1910, in southcentral Montana. These earthquakes, a considerable distance from the deposit, may have had effects in the Slide Lake area. While not triggering the Slide Lake rockfall avalanche, earthquakes may have contributed to the instability of the rock mass.

Precipitation and seismic data can also be examined to postulate the trigger mechanisms for the 1946 rockfall event at Slide Lake. Precipitation data from 1945-1946 (Table VI) generally indicate no excessive precipitation at the Babb station which might have triggered the rockfall event. A situation similar to the 1910 event, when cloudbursts in the mountains triggered the rockfall avalanche, may have occurred.

The seismic data for 1945-1946 are interesting. A major earthquake occurred approximately 115 km west of Slide Lake on September 23, 1945 (Qamar and Stickney, 1983). The effects of this quake were detected over an area of 93,000 $km^2$, and the Slide Lake area is well within the region affected by this quake. If this earthquake triggered a major rockfall event, the growth responses in the damaged tree rings would probably not have shown up until the 1946 growing season, because late-September 1945 is beyond the 1945 growing season. A strong possibility exists that this late-September 1945 earthquake may have been the trigger mechanism which initiated the "1946" rockfall event at Slide Lake.

## Causative Factors and Trigger Mechanisms of the Napi Point Deposit

The same causal mechanisms which led to the release of the Slide Lake rockfall avalanche probably were important mechanisms leading to the release of the Napi Point rockfall avalanche. These factors include the presence of the Lewis thrust fault, with highly jointed and fractured bedrock; freeze-thaw processes operating to further weaken the rock mass; and oversteepening of the slope and rockface by glacial activity during the Pleistocene (Alden, 1932). Table I lists the number of fractures in the bedrock found on the cliff scar immediately above the deposit, indicating the large amount of fracturing and jointing found there.

The possibility of excessive precipitation or earthquake activity as trigger mechanisms to the Napi Point rockfall avalanche again can be examined, considering that 1954 has been accepted as the year the rockfall avalanche occurred. Table VI lists the monthly precipitation data for 1953-1954 for the Babb station. During the time period from late 1953 through the 1954 growing season, only August and September, 1954 show excesses of precipitation, although no extreme amounts of precipitation fell during these months. The Napi Point deposit may have received excessive precipitation or cloudburst activity which does not clearly show up in the Babb precipitation data. In short, excessive precipitation may have triggered the Napi Point rockfall avalanche, although this conclusion cannot be confirmed. Excessive precipitation in August, 1954, could have triggered the rockfall avalanche, and the resulting growth responses in damaged trees would be visible in the 1954 annual rings.

Three earthquakes occurred in July and August 1953 whose effects may have been felt in the Napi Point area (Qamar and Stickney, 1983). One earthquake occurred on July 8, and the other two on August 8. The area experiencing the effects of these earthquakes is unknown, although one of the August 8 quakes was a high intensity quake, based on the Modified Mercalli Intensity Scale of 1931 (Qamar and Stickney, 1983). These earthquakes probably did not immediately trigger the rockfall avalanche, because growth responses in damaged trees would have been visible in the 1953 annual rings, and this was not the case. These earthquakes may have led to further instability of the rock mass which eventually gave way, and should be considered as important causal mechanisms which led to the eventual release of the Napi Point rockfall avalanche.

## Conclusion

Rockfall avalanches result from many causal mechanisms, and are not a result of just one factor. Trigger mechanisms of rockfall avalanches, however, may be a single action or factor. An historical reference indicates the 1910 Slide Lake rockfall avalanche was triggered by excessive precipitation. The 1946 Slide Lake rockfall event may have been triggered by seismic activity in late-September 1953. The trigger mechanism for the 1954 Napi Point rockfall avalanche is less evident, although excessive precipitation in August, 1954, may have been responsible.

Information concerning the trigger mechanisms of these rockfall events is important, for it enables one to speculate on the hazards of rockfall avalanches, and the potential for future rockfall avalanche activity along the Lewis thrust fault area in Montana.

# CHAPTER VI

# CONCLUSIONS OF THIS STUDY

Two of the three original objectives set forth in Chapter One have been achieved and discussed. They include the mapping of the spatial distribution of landslide activity along the Lewis thrust fault, and the determination of the dates and trigger mechanisms of two high magnitude rockfall avalanches. Discussion of the third objective, the placement of the landslide activity along the Lewis thrust fault into a natural hazards context, will be presented in this chapter.

## Hazard Threats of Landslide Phenomena

Threats from natural hazards to human activity have been previously mentioned in this study. The Mount Huascaran rockslide in Peru in 1970 killed 20,000 people (Browning, 1973). The Hebgen Lake , Montana, landslides and rockfalls in 1959 killed 28 people (Hadley, 1964), and the rockfall/slide at Frank, Alberta, in 1903 killed 70 people (Sharpe, 1938). Porter and Orombelli (1981) described similar rockfall hazards in the western Italian Alps.

Varnes (1975) concluded that the most destructive slope failures in the western United States, in terms of lives lost and property damage, are rapid debris-flows and mud-flows. Many residential developments have been destroyed by these slope failures. In 1969, twenty lives were lost in the Los Angeles metropolitan area. Spreading failures, which involve flowage of several distinct units of earth and rock material, have also caused considerable damage in California and Alaska (Varnes, 1975).

Landslide phenomena can cause extensive property damage and great loss of life. Other hazard threats exist from landslide activity, however, which are also of great importance. Several authors cite the role landslides play in increasing sediment loads in streams. Whitehouse (1983) concluded that rock avalanche deposits are a significant sediment source for rivers in mountain environments. Gonsior and Gardner (1971) noted that slope failures in central Idaho will greatly accelerate sedimentation, adversely affecting aquatic habitats. Increased silting of downstream reservoirs may also result.

## Landslide Hazards in the Study Area

As described in Chapter Three, 78 post-glacial landslide deposits are located along the Lewis thrust fault in Glacier National Park (Plate 1). This large number of landslide deposits indicates landslide activity in this area poses a natural hazard threat. These landslide hazards include the hazards to humans and human structures within the Park; the hazards to gas and oil companies which are extracting mineral resources along the National Park/Reservation boundary; and the hazards associated with increased sediment loads from landslide activity to streams and reservoirs in the study area.

Several roads, trails, campgrounds, and buildings are located near the Lewis thrust fault in Glacier National Park (Plate 1). Some of these structures lie close enough to the Lewis thrust fault so that a high-magnitude landslide event, such as a rockfall avalanche, could cause considerable damage to these structures.

There has been considerable interest shown the last few years by gas and oil exploration companies in the area along the National Park/Blackfeet Reservation border. Several wells are producing gas or oil on the Reservation, most of which were drilled in the 1950's and between 1978-1982 (Glacier National Park, 1983). Plate 1 shows that several landslides, including rockfall avalanches, rockfall/slides, slumps, and earthflows, extend beyond the Park Service boundary onto Reservation land. The Chevron Oil Company has recently applied for leases to conduct oil and gas exploration in creek drainages extending from Otatso Creek south to St. Mary Lake (Hungry Horse News, 1984). Plate 1 shows this area has extensive rockfall avalanche, rockfall/slide, and slump deposits. The structures associated with oil and gas extraction could suffer considerable damage from landslide activity.

A potential significant impact of landslide activity in the study area is increased stream sediment load. As mentioned previously, the easternmost lobe of the 1910 Slide Lake rockfall avalanche deposit continues to provide a heavy sediment load into Otatso Creek.

Three major negative effects of increased stream sediment loads from landslide activity can be identified: 1) adverse impacts on aquatic habitats, 2) increased siltation of downstream reservoirs, and 3) a reduction in water quality which could affect downstream water-use purposes.

Two reservoirs are located along the National Park/Reservation boundary. The Sherburne Dam was constructed in approximately 1914-1915, and impounds Swiftcurrent Creek to form Lake Sherburne. Lower Two Medicine Lake was formed by the dam which impounded Two Medicine Creek. Plate 1 shows that the slopes on either side of Lake Sherburne have extensive landslide deposits, as do the slopes north and south of Lower Two Medicine Lake. The potential exists for future landslide activity to cause siltation of these reservoirs.

Water from the rivers flowing eastward out of the mountains in Glacier National Park is used extensively for irrigation purposes (Bill Conrod, written communication, 1984). Alfalfa, wheat, and sugar beet crops east of the mountains are irrigated with mountain streamwater. Landslides along the Lewis thrust fault could greatly increase the turbidity of the streamwater, thereby lowering the quality of the water used for irrigation purposes. By a joint Canadian/United States treaty, three-fourths of the St. Mary River water is reserved for Canadian use, and one-fourth for American use (Bill Conrod, written communication, 1984). Water from the Sherburne Reservoir flows into the St. Mary River. A reduction in the water quality in streams affected by landslide activity thus has international implications. Aquatic habitats can be severely impacted through sediment load increases from landslide activity (Gonsior and Gardner, 1971); landslide activity could disrupt many aquatic life forms in the study area.

An interesting situation could arise should a high-magnitude landslide produce significant water quality deterioration in a stream, because of the differing philosophies between the National Park Service and other agencies. Glacier National Park is managed to perpetuate the natural processes operating there, and thus a landslide causing turbidity in a natural water system would be viewed as a natural process. Agencies and individuals utilizing these natural water systems downstream may have different perceptions and philosophies.

The potential for development of structures for natural resource extraction on the Blackfeet Reservation near the Park Service border is high. Landslide hazard threats exist on Park Service and Blackfeet Reservation lands to existing human structures, and to future development of Reservation land near the Park Service boundary. Development on Park Service and Reservation lands near the Lewis thrust fault in the future should be kept to a minimum, however, given the landslide hazard of the area.

## Potential for Future Landslide Activity Along the Lewis Thrust Fault

No evidence exists to indicate that landslide activity along the Lewis thrust fault is a relict geomorphic process; indeed, this study has shown the opposite is true. Excessive precipitation and cloudburst activity in this area are natural meteorologic phenomena; such events have triggered landslide activity in the past. The highly jointed, fractured, and faulted rock formations of the Lewis thrust fault can only become more unstable, as natural freeze-thaw mechanisms and weathering processes continue to create and amplify rock mass instability.

The general area around Glacier National Park continues to be seismically active. The region just west-southwest of the Park (the Flathead Lake Region) remains very seismically active (Qamar and Stickney, 1983). The immediate Lewis thrust fault area (East Flathead Lake Region) has experienced low seismicity recently, but seismic potential remains high. Many of the earthquakes in the Flathead Lake Region are also felt in the Lewis thrust fault area. Between 1901-1979, 309 earthquakes were recorded or documented for these two regions (Qamar and Stickney, 1983); many smaller earthquakes, especially in the early 1900's, may have gone unrecorded in less populous areas (such as Glacier National Park). In October, 1983, a strong earthquake, centered near Challis, Idaho, was felt in the Park (personal experience, 1983), and measured 3.0 on the Richter scale in the Park area. A concrete bridge and trees were swaying, and walls in Park Service offices were moving. Another large earthquake occurred on February 12, 1984, which shook buildings in Idaho, Montana, and Alberta (Tulsa World, February 12, 1984). These facts indicate the general Glacier Park area is very seismically active, and earthquakes may be an important trigger mechanism of landslides in the future.

In summary, it would appear that the same causal mechanisms which have led to landslide activity in the past along the Lewis thrust fault should continue to produce landslide activity in the future. Landslide hazard threats certainly exist along the Lewis thrust fault zone in Glacier National Park.

## Future Research Possibilities

Several potential future research topics can be associated with this study. The Slide Lake and Napi Point deposits could be evaluated with plant successional studies. The rates of succession and types of species inhabiting these initially bare deposits would make a useful study. The relative age-dating techniques of lichenometry and rock weathering rinds could be applied to older landslide deposits along the Lewis thrust to determine the dates of prehistoric mass movement.

It would also be interesting to visit the two study sites in the near future to determine if the strong earthquakes of October, 1983, and February, 1984, have triggered further rockfall activity. Monitoring the Lewis thrust fault area after the occurrence of potential trigger mechanisms could aid in the understanding of causal mechanisms of landslide acivity in this area.

## Concluding Statement

The fact that two high-magnitude rockfall avalanches have occurred within this century, one as recently as 1954, indicates that the slopes along the Lewis thrust fault in Glacier National Park are geomorphically active. Clague (1981) and Gardner et al. (1983) concluded that in their study areas in the Canadian Rockies, rockfall/slide activity peaked soon after the Pleistocene glaciation, and that little large-scale rockfall/slide activity still occurs today. Similar to the Cruden (1982) study in Alberta, this study clearly indicates that high-magnitude slope failures continue to occur in the northern Rockies, long after the cessation of Pleistocene glacial activity.

Mudge and Earhart (1980) stated that the known length of the Lewis thrust fault is 452 km, extending from Steamboat Mountain, Montana northward into Canada for 225 km. Although little development will occur on National Park Service lands, other areas along the Lewis Thrust fault may experience development. Any future development in these mountain areas should consider the threats which landslide activity could impose on that development.

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# APPENDIX A

## NAMES AND LOCATIONS OF QUADRANGLES IN THE STUDY AREA

Babb: 48°45'00" N - 48°52'30"N
113°30'00"W - 113°37'30"W

Blacktail: 48°15'00" N - 48°22'30"N
113°22'30"W - 113°30'00"W

Chief Mountain: 48°52'30"N - 49°00'00"N
113°30'00"W - 113°37'30"W

Cut Bank Pass: 48°30'00"N - 48°37'30"N
113°22'30"W - 113°30'00"W

Gable Mountain: 48°52'30"N - 49°00'00"N
113°37'30"W - 113°45'00"W

Kiowa: 48°30'00"N - 48°37'30"N
113°15'00"W - 113°22'30"W

Lake Sherburne: 48°45'00"N - 48 52'30"N
113°30'00"W - 113°37'30"W

Many Glacier: 48°45'00"N - 48°52'30"N
113°37'30"W - 113°45'00"W

Mount Cleveland: 48°52'30"N - 49°00'00"N
113°45'00"W - 113°52'30"W

Mount Rockwell: 48°22'30"N - 48°30'00"N
113°22'30"W - 113°30'00"W

Rising Sun: 48°37'30"N - 48°45'00"N
113°30'00"W - 113°37'30"W

APPENDIX A (CONTINUED)

St. Mary: $48^{\circ}37'30''$N - $48^{\circ}45'00''$N
$113^{\circ}22'30''$W - $113^{\circ}30'00''$W

Squaw Mountain: $48^{\circ}22'30''$N - $48^{\circ}30'00''$N
$113^{\circ}15'00''$W - $113^{\circ}22'30''$W

Summit: $48^{\circ}15'00''$N - $48^{\circ}22'30''$N
$113^{\circ}15'00''$W - $113^{\circ}22'30''$W

## APPENDIX B

## FIELD DATA FROM TREE SAMPLING
## - SLIDE LAKE -

| Sample Number | Tree Species | Tree Height (in Meters) | Diameter at Breast Height (cm) | Orientation of Tilt (Azimuth) | Height at Which Sample was Taken (cm) | Comments |
|---|---|---|---|---|---|---|
| SL 1 | *Picea englemannii* Parry (Englemann spruce) | 6 | 18 | 13 | 61<br>150 | Scarred and tilted tree |
| SL 2 | Englemann spruce | 4.5 | 16 | 12 | 43 | Tilted tree |
| SL 3 | Englemann spruce | 4.5 | 16 | 328 | 7 | Tilted tree |
| SL 4 | Englemann spruce | 3 | 10 | 310 | 9<br>23<br>81 | Scarred and tilted tree |
| SL 5 | Englemann spruce | 6 | 18 | 67 | 93<br>186 | Scarred and tilted tree |
| SL 6 | Englemann spruce | 4.2 | 12 | -- | 145 | Scarred tree |
| SL 7 | Englemann spruce | 4.2 | 12 | 25 | 26 | Scarred and tilted tree |
| SL 8 | Englemann spruce | 6 | 17 | 37 | 42<br>128 | Tilted tree |
| SL 9 | Englemann spruce | 6.4 | 18 | 15 | 91<br>151 | Scarred and tilted tree |

APPENDIX B (CONTINUED)

| Sample Number | Tree Species | Tree Height (in Meters) | Diameter at Breast Height (cm) | Orientation of Tilt (Azimuth) | Height at Which Sample was Taken (cm) | Comments |
|---|---|---|---|---|---|---|
| SL 10 | Englemann spruce | 3.6 | 13 | 22 | 41 | Tilted tree |
| SL 11 | Englemann spruce | 3.3 | 10 | 325 | 21<br>71 | Scarred and tilted tree |
| SL 12 | Englemann spruce | 2.2 | 11 | 294 | 20<br>170 | Scarred and tilted tree |
| SL 13 | Englemann spruce | 6 | 19 | 57 | 30<br>275 | Tilted tree |
| SL 14 | Englemann spruce | 4 | 16 | 42 | 23 | Scarred and tilted tree |
| SL 15 | Englemann spruce | 3.3 | 9 | 7 | 23 | Scarred and tilted tree |
| SL 16 | Englemann spruce | 6 | 13 | -- | 25 | Scarred tree |
| SL 17 | Englemann spruce | 4.5 | 11 | -- | 50<br>87 | Scarred tree |
| SL 18 | Englemann spruce | 5.5 | 14 | 6 | 50<br>85 | Scarred and tilted tree |
| SL 19 | Englemann spruce | 4.5 | 16 | -- | 59 | Scarred tree |
| SL 20 | Englemann spruce | 4.5 | 15 | 36 | 31<br>136 | Tilted tree |

## APPENDIX B (CONTINUED)

| Sample Number | Tree Species | Tree Height (in Meters) | Diameter at Breast Height (cm) | Orientation of Tilt (Azimuth) | Height at Which Sample was Taken (cm) | Comments |
|---|---|---|---|---|---|---|
| SL 21 | Englemann spruce | 4.5 | 10 | 66 | 24 | Scarred and tilted tree |
| SL 22 | *Abies lasiocarpa* (Hook) Nutt. (Subalpine fir) | 7.5 | 30 | -- | 80 | Scarred tree |
| SL 23 | Subalpine fir | 7.3 | 21 | 29 | 54 | Tilted tree |
| SL 24 | Subalpine fir | 9.1 | 16 | 350 | 153<br>474 | Tilted tree |
| SL 25 | Subalpine fir | 5.5 | 12 | 70 | 39<br>99 | Scarred and tilted tree |
| SL 26 | Englemann spruce | 7.6 | 16 | 302 | 41<br>214 | Tilted tree |
| SL 27 | Englemann spruce | 4.5 | 13 | 310 | 47<br>144 | Scarred and tilted tree |
| SL 28 | Englemann spruce | 4.5 | 13 | 332 | 36 | Tilted tree |
| SL 29 | Englemann spruce | 3.5 | 14 | 50 | 41<br>128 | Tilted tree |
| SL 30 | Englemann spruce | 6 | 18 | 251 | 33<br>89 | Tilted tree |

## APPENDIX B

## FIELD DATA FROM TREE SAMPLING
## - NAPI POINT -

| Sample Number | Tree Species | Tree Height (in Meters) | Diameter at Breast Height (cm) | Orientation of Tilt (Azimuth) | Height at Which Sample was Taken (cm) | Comments |
|---|---|---|---|---|---|---|
| NP 1 | Abies lasiocarpa<br>(Hook) Nutt.<br>(Subalpine fir) | 5.5 | 13 | 219 | 23 | Tilted tree |
| NP 2 | Pinus flexilis<br>James<br>(Limber pine) | 7.6 | 15 | 191 | 64 | Tilted tree |
| NP 3 | Subalpine fir | 7.6 | 15 | 105 | 81<br>162 | Tilted tree |
| NP 4 | Subalpine fir | 6 | 13 | -- | 36<br>61 | Scarred tree |
| NP 5 | Subalpine fir | 9.1 | 15 | 177 | 60<br>255 | Scarred and tilted tree |
| NP 6 | Limber pine | 4.5 | 15 | 3 | 43<br>59 | Tilted tree |
| NP 7 | Limber pine | 4.5 | 15 | 25 | 42<br>75 | Tilted tree |
| NP 8 | Limber pine | 4.2 | 14 | 226 | 46<br>152 | Tilted tree |
| NP 9 | Subalpine fir | 9.1 | 15 | 119 | 37<br>227 | Tilted tree |

APPENDIX B (CONTINUED)

| Sample Number | Tree Species | Tree Height (in Meters) | Diameter at Breast Height (cm) | Orientation of Tilt (Azimuth) | Height at Which Sample was Taken (cm) | Comments |
|---|---|---|---|---|---|---|
| NP 10 | Subalpine fir | 5.5 | 11 | 95 | 29<br>65 | Tilted tree |
| NP 11 | Limber pine | 10 | 15 | 170 | 60 | Tilted tree |
| NP 12 | Subalpine fir | 7.6 | 13 | 105 | 61<br>156<br>198 | Scarred and tilted tree |
| NP 13 | Subalpine fir | 8.2 | 14 | 204 | 52<br>206 | Scarred and tilted tree |
| NP 14 | Subalpine fir | 3.6 | 9 | 273 | 16<br>86 | Scarred and tilted tree |
| NP 15 | Subalpine fir | 3.6 | 10 | 65 | 32 | Scarred and tilted tree |
| NP 16 | Subalpine fir | 7.3 | 17 | 76 | 41<br>276 | Tilted tree |
| NP 17 | Subalpine fir | 7.6 | 14 | 9 | 46<br>105 | Scarred and tilted tree |
| NP 18 | Subalpine fir | 4.2 | 11 | 87 | 45<br>115 | Scarred and tilted tree |
| NP 19 | Subalpine fir | 4.2 | 11 | 248 | 48<br>85 | Tilted tree |
| NP 20 | Subalpine fir | 3.9 | 10 | 44 | 59<br>143 | Scarred and tilted tree |

APPENDIX B (CONTINUED)

| Sample Number | Tree Species | Tree Height (in Meters) | Diameter at Breast Height (cm) | Orientation of Tilt (Azimuth) | Height at Which Sample was Taken (cm) | Comments |
|---|---|---|---|---|---|---|
| NP 21 | Subalpine fir | 7.6 | 15 | 7 | 37<br>58<br>92 | Tilted tree |
| NP 22 | Subalpine fir | 7.6 | 11 | 146 | 27<br>84 | Tilted tree |
| NP 23 | Subalpine fir | 5.5 | 10 | 164 | 38<br>88 | Tilted tree |
| NP 24 | Subalpine fir | 4.2 | 11 | 329 | 24<br>109 | Tilted tree |
| NP 25 | Subalpine fir | 6.1 | 12 | 267 | 46<br>274 | Scarred and tilted tree |
| NP 26 | Subalpine fir | 3.9 | 13 | -- | 36<br>75 | Scarred tree |
| NP 27 | Subalpine fir | 4.8 | 16 | 90 | 80<br>242 | Scarred and tilted tree |
| NP 28 | Subalpine fir | 6 | 13 | 12 | 60<br>190 | Tilted tree |
| NP 29 | Subalpine fir | 3.6 | 10 | 28 | 52<br>95 | Tilted tree |
| NP 30 | Subalpine fir | 3.6 | 8 | 59 | 9 | Tilted tree |

VITA

Jack Graham Oelfke

Candidate for the Degree of

Master of Science

Thesis: THE LOCATION AND ANALYSIS OF LANDSLIDES ALONG THE LEWIS OVERTHRUST FAULT, GLACIER NATIONAL PARK, MONTANA

Major Field: Geography

Biographical:

Personal Data: Born in Detroit Lakes, Minnesota, December 21, 1956, the son of John F. and Marjorie K. Oelfke.

Education: Graduated from Frazee High School, Frazee, Minnesota, June, 1975; received Bachelor of Arts degree at Moorhead State University, Moorhead, Minnesota, March, 1982; completed requirements for the Master of Science degree at Oklahoma State University, Stillwater, Oklahoma, May, 1984.

Professional Experience: Resource Management Intern at Glacier National Park, October, 1983 to January, 1984. Graduate Teaching Assistant for the Department of Geography, Oklahoma State University, August, 1982 to May, 1984. Maintenance Worker Leader, Glacier National Park, Montana, 1979-1984 (seasonally).

Plate 1

Quadrangles Studied
INDEX TO 7.5 MINUTE MAPS
EXPLANATION
International boundary
County line
Park boundary
Reservation boundary
Roads and buildings
Trails
Campground
Picnic area
Lake or pond
Intermittent lake
Perennial streams
Intermittent streams
Cascades or falls
Spring
Glacier or snowfield
Contours
Triangulation station
Bench mark
Elevation above mean sea level
Township and section lines
ROAD CLASSIFICATION
Primary highway, hard surface
Light-duty road, hard or improved surface
Secondary highway, hard surface
Unimproved road
Interstate Route
U. S. Route
State Route
Lewis Thrust Fault
Subsidiary Fault
Boundaries drawn on larger deposits; dashed lines where boundaries are unknown.
Rockfall/Rockslide
Rockfall avalanche
Slump
Earthflow
Undifferentiated mass movement deposit
Unknown deposit
LANDSLIDES ALONG THE LEWIS THRUST FAULT
GLACIER NATIONAL PARK
PART OF
WATERTON-GLACIER INTERNATIONAL PEACE PARK
MONTANA
1968
N4813—W11310/47X80
SCALE 1:100,000
CONTOUR INTERVAL 80 FEET
NATIONAL GEODETIC VERTICAL DATUM OF 1929
Mapped, edited, and published by the Geological Survey
Control by USGS and USC&GS
Compiled from 1:24,000-scale, 7.5 minute maps, surveyed 1958, 1962, 1964, 1966, 1968. Supersedes 1:125,000-scale map dated 1938
Polyconic projection. 1927 North American datum
50,000-foot grid based on Montana coordinate system, north zone
Where omitted, land lines have not been established or are not shown because of insufficient data
Metric elevations to the nearest half meter are shown in parentheses (Meters = feet × .3048)
MONTANA
MAP LOCATION
UTM GRIDS AND 1968 MAGNETIC NORTH DECLINATION AT CENTER OF SHEET
THIS MAP COMPLIES WITH NATIONAL MAP ACCURACY STANDARDS
FOR SALE BY U. S. GEOLOGICAL SURVEY, DENVER, COLORADO 80225, OR RESTON, VIRGINIA 22092
A FOLDER DESCRIBING TOPOGRAPHIC MAPS AND SYMBOLS IS AVAILABLE ON REQUEST
LAKE SHERBURNE
SAINT MARY LAKE
LOWER ST MARY LAKE
Saint Mary
Two Medicine Lake
LEWIS AND CLARK NATIONAL FOREST

Plate 2

# MODIFIED SKELETON PLOT - SLIDE LAKE 1-15



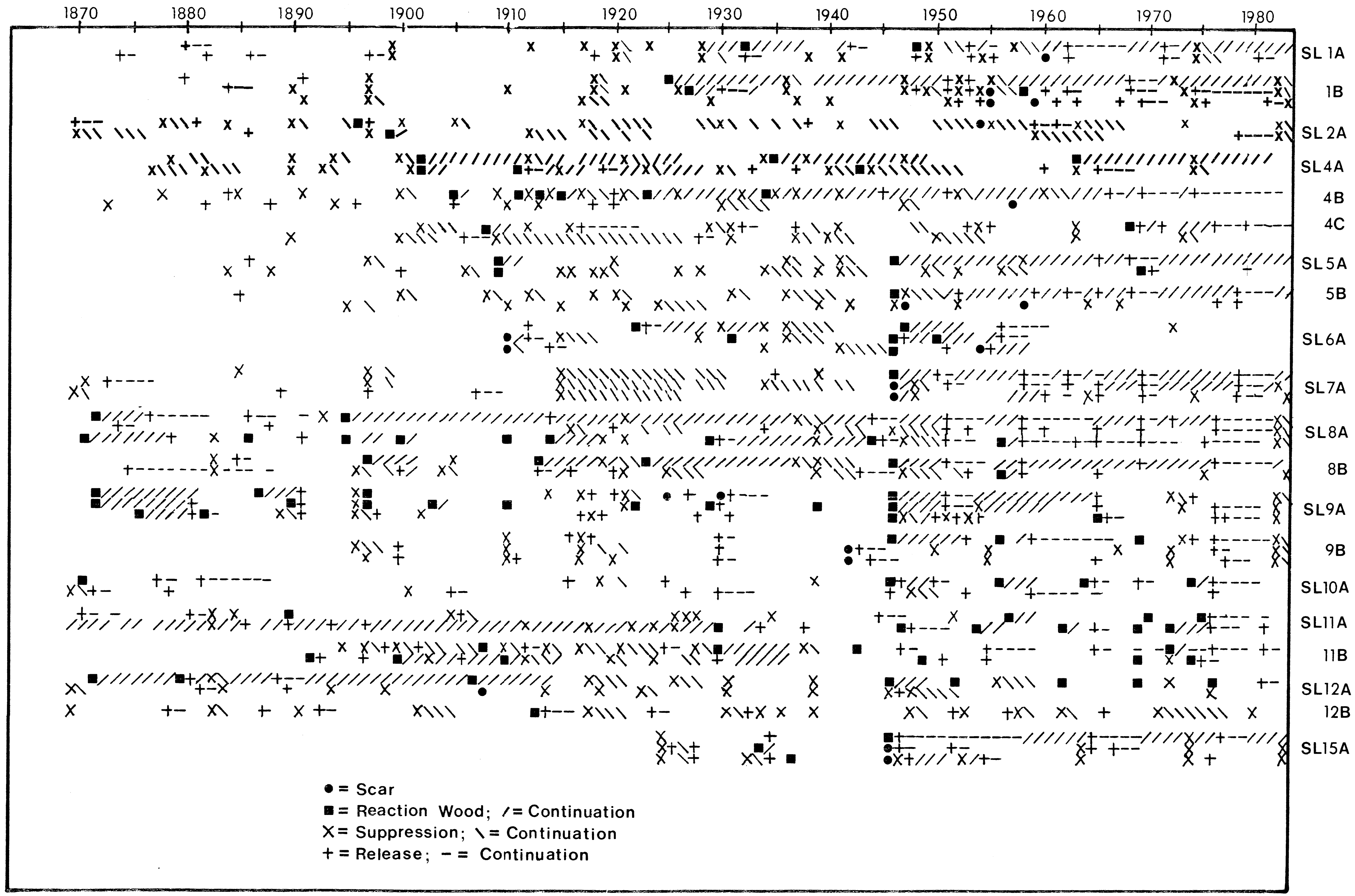

Plate 3

# MODIFIED SKELETON PLOT - SLIDE LAKE 17-30

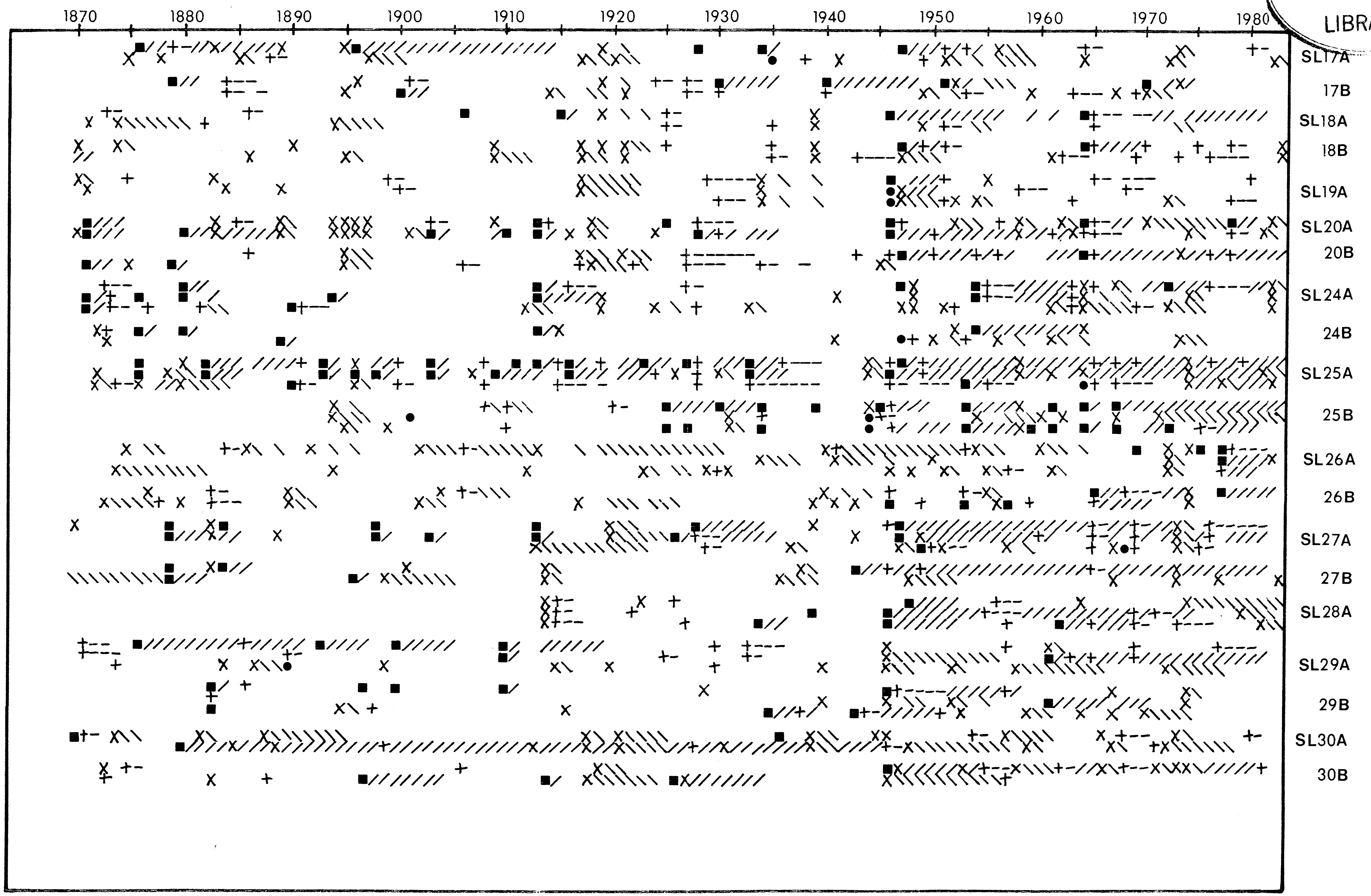

Plate 4 MODIFIED SKELETON PLOT - NAPI POINT 1-17

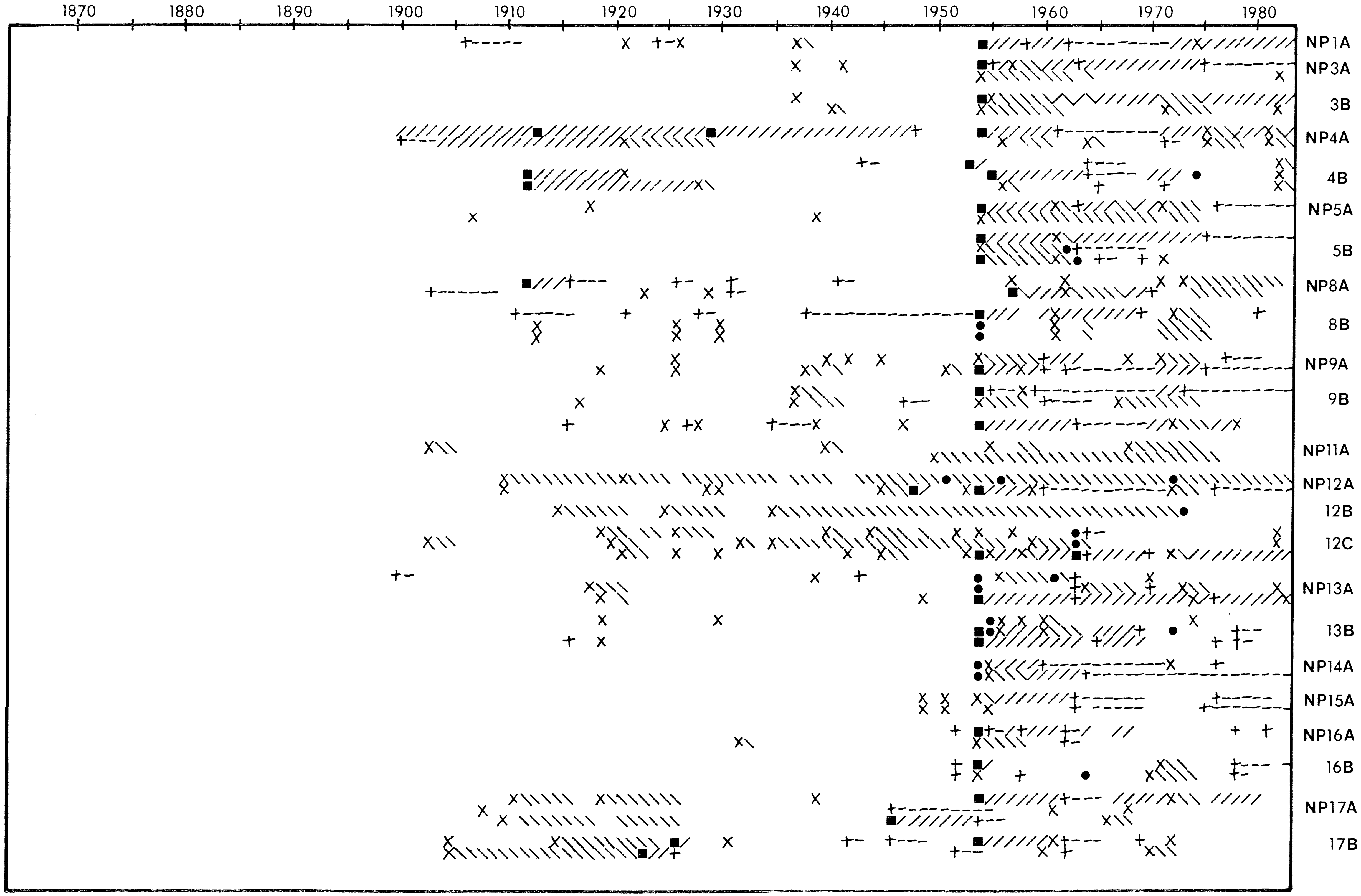

Plate 5

# MODIFIED SKELETON PLOT - NAPI POINT 18-30