

# **LANDSLIDES NEAR ASPEN, COLORADO**

**Carol P. Harden**

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LANDSLIDES NEAR ASPEN, COLORADO

by

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A contribution to the United States Unesco  
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Study of the impact of human activities on  
mountain and tundra ecosystems

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## FOREWORD

In the past few decades, population growth, affluence, increased leisure time, and improved transportation have changed the timing of man's impact on the Rocky Mountains from one of summer-oriented to yearlong. Mountain communities have grown as an alternative to urban living, but their residents, whether indigenous or transient, have little desire to function without the conveniences and amenities of the city. Consequently, the demands on the mountain ecosystems are greater than ever before, and as the pressures of recreation and development increase, the danger mounts of destroying the very characteristics for which the mountains are valued. The question then arises as to whether the mountain ecosystems can be modified and developed in such a way as to minimize the detrimental environmental impact and provide for a balanced and coordinated land-use management program.

As an outgrowth of increased global environmental awareness, the Man and the Biosphere Program (MAB) was adopted by resolution of the General Conference of Unesco in 1968. The general objectives of this international research endeavor are to improve the relationships between man and the environment and to develop the basis within the natural and social sciences for the rational use, conservation, and efficient management of the natural resources of the biosphere. One of 14 separate MAB areas of concentration, Project 6 - the study of the impact of human activities on mountain and tundra ecosystems - deals, in part, with applied scientific research related to resource development and land-use policy considerations. It is in this context that the present study has been designated as a United States Project 6 contribution to the Man and the Biosphere Program.

Expansion of the winter sports industry and the spread of second homes have made the problems associated with rapid development particularly acute in the mountain sections of Colorado. Construction is not uncommon in areas exposed to a variety of natural hazards and has thus placed man in increasing conflict with natural physical processes. Landsliding is one such process which represents a potential hazard to property and life and which may disrupt the social and economic continuum. Man's ability to induce landslides, as either an immediate or delayed response to improvident

activities, further aggravates the issue. In an attempt to define the problem, this monograph identifies and examines the origins of 62 landslides in the vicinity of Aspen, Colorado. Through the analysis of conditions associated with landslides, a descriptive model of landsliding in the area has been established. The characteristics defined by this model allow the prediction of future landslide events and may be of use in the development of a rational land-use policy for the area.

It is perhaps particularly significant that this landslide study by Carol Harden was supported by a graduate fellowship fund established in 1973 through the Institute of Arctic and Alpine Research (INSTAAR), University of Colorado, by the Aspen Skiing Corporation. The choice of research topic was left to the fellowship recipient with the one restriction that it in some way related to the Aspen vicinity. Through this foresight, the landslide hazard analysis in this monograph should enhance the ability of the Aspen Skiing Corporation to plan environmentally-sound and hazard-free development in an area whose steep slopes are especially vulnerable to human impact.

While many individuals assisted with this study in some way, several must be acknowledged by name for their special support. Dr. N. Caine, supervisor for this thesis, provided advice, encouragement and helpful criticism throughout the project. T. Marshall and L. Beidleman of the Aspen Skiing Corporation and R. Rankin and W. Johnson of the U.S.D.A. Forest Service rendered invaluable assistance in the field. Finally, the assistance of the Aspen Skiing Corporation is gratefully acknowledged.

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# ABSTRACT

Sixty-two landslides have been identified in a 68-square-mile area west of Aspen, Colorado, and mapped, primarily from air photographs. Their distribution is a clustered one which reflects the spatial control of landslide location.

Landslides are found on every major lithologic unit in the area and landslide characteristics do not differ significantly between lithologies. They are, however, located preferentially on east-facing slopes of north-trending valleys and on north-facing slopes of east-trending valleys. Nearly all landslides also involve a component vector of bedrock dip in the direction of slope. No landslides have starting zone slope angles of less than  $16^{\circ}$ .

Age determinations, based on vegetation and stratigraphy, suggest an age range of pre-Bull Lake to presently active, with most of the landslides considered to have occurred soon after 11,000 BP in association with conditions of the immediate postglacial time. A few slides have been initiated in the past 100 years.

The most critical parameters involved in landslide initiation are water and slope angle. Dry slopes and those at an angle of less than  $12^{\circ}$  are not expected to slide. Inactive old landslides and slopes of morainal deposits or Mancos Shale are exceptionally susceptible to movement. The margin of stability over the entire area is considered to be slim and caution is advised in land use, especially where a changed water use is involved.

## CHAPTER I

### INTRODUCTION

#### Landslides and Their Significance

"Landslide" is a general term covering a wide range of mass movement landforms and processes involving the moderately rapid to rapid (1 foot per year or greater) downslope transport by means of gravitational body stresses of soil and rock material en masse (Gary et al., 1972, p. 396). This study deals with rotational and planar slides which are landslide types involving the failure of an entire section of hillside. A rotational landslide is characterized by the upward rotation of a mass of material moving downslope on an arc-shaped shear surface, generally of the same material as the moving mass. The axis of rotation lies above the ground surface and is normal to the downslope direction of movement. A planar slide, on the other hand, is characterized by the downslope movement of a mass over a dipping shear plane (analogous to a sliding board) which is generally of a different, more resistant material, such as glacial till sliding on a dipping bedrock strata. The ratio of landslide length to the depth of moving material tends to be much larger for planar than for rotational landslides.

The landform produced by a rotational slide just after movement (Figure 1.1) typically has a scarp at the top which is continuous with the curved shear surface. Downslope are found segments of the moved mass with surfaces less sloped than the adjacent hillside. In cases where pore water pressure is high, a tongue may flow out onto less sloping land below. A planar landslide (Figure 1.2) leaves the sliding surface exposed, with a scarp above where the sliding material failed in tension and hummocky terrain in the deposition zone.

The stability of a hillslope is conventionally expressed in terms of the "safety factor" of the slope: the ratio of shear strength to shear stress. Shear strength,  $s$ , for mixed soils is given by the Coulomb model

$$s = c + (p - u) \tan \phi$$

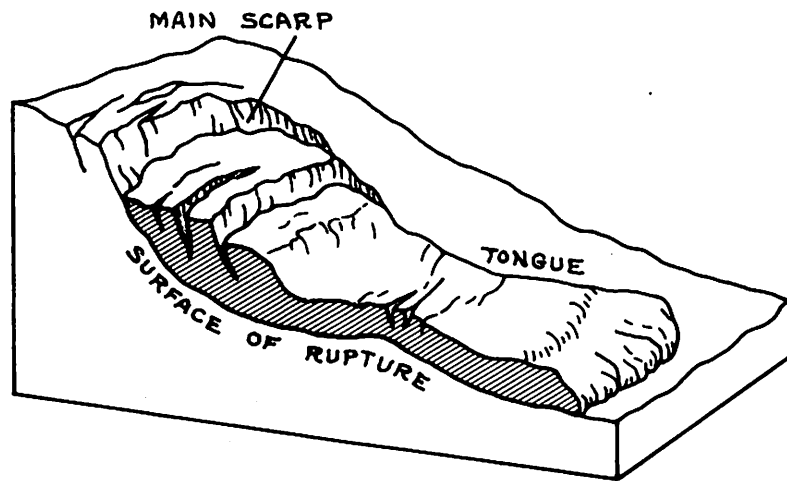


FIGURE 1.1  
Rotational Landslide

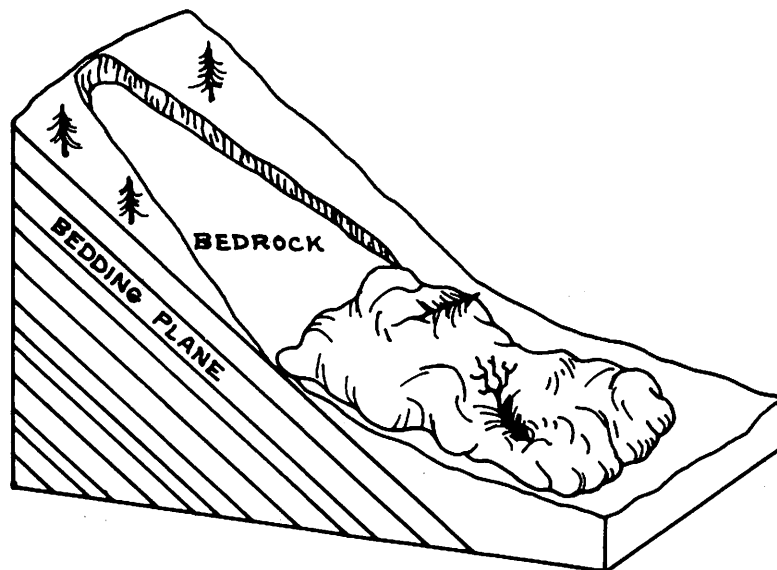


FIGURE 1.2  
Planar Landslide

where  $c$  is unit cohesion,  $p$  is the normal stress on the surface of sliding,  $u$  is the pore pressure in the normal direction, and  $\phi$  is the angle of internal friction of the material. Shear stress,  $\tau$ , is given by

$$\tau = y \cdot z \cdot \sin\theta \cos\theta$$

where  $y$  is the bulk unit weight of the soil,  $z$  is depth to the failure surface, and  $\theta$  is the slope angle (Carson and Kirkby, 1972, p. 154).

When the safety factor is 1.0 or less, shear stress is equal to or greater than shear strength (resistance) and the slope can be expected to slide.

Construction of buildings, water lines, or roads can reduce the safety factor of a hillslope by increasing the unit weight ( $y$ ) or by increasing the slope angle ( $\theta$ ). Where the safety factor of a natural slope is close to 1.0, a minor disturbance may initiate movement of the hillside.

Landsliding is an important mechanism for natural topographic adjustments. Where man is concerned, it can be annoying, environmentally destructive, and even life-threatening. An important consequence of landslides is their effect on water quality. The movement of material downslope and erosion of the resulting scarred slope can introduce debris into the streams which, in addition to detracting from their appearance, may destroy aquatic habitats, render the water unsuitable for irrigation or industrial use, and create problems of sedimentation. Landslides are life-threatening when they collapse structures whose stress thresholds are reached and when they move at speeds that preclude escape. Rates of landslide movements range from gradual (1 foot per year) to catastrophic (250 miles per hour).

Landslides intrigue man, as do other natural phenomena which threaten his life or property. Landslides also disrupt man's economy. Since landslides are induced by gravity, they are most common in hilly or mountainous areas. The demands of our growing and increasingly mobile population for resources, housing, recreation, and scenery are pushing development farther into those less stable areas; thus the recognition of past landslides and the consequences of future landslides are becoming increasingly important.

### Nature and Objectives of This Study

There are numerous examples in the engineering-geologic literature of destruction of man-made structures resulting from the failure of preliminary site investigations to identify ancient landslides (Zolotarev, 1961). Increasing usage and existing evidences of slope instability make the Highland Peak/Snowmass region an area where the investigation of previous landslides could make a contribution to the understanding of local slope conditions and thus to planning for future use and development of that region and surrounding areas.

The objectives of this study are (1) to identify and characterize the form, location, and distribution of previous landslides in the area; (2) to determine the age of those landslides, seeking clues from tectonic, glacial, or climatic events of those times to explain the causes of movement; (3) to identify any existing relationships between past and present instabilities; and (4) to synthesize this information into a descriptive model of landsliding in the area, thus establishing a basis for the prediction of future landslide events.

### Previous Study

Landslides have been studied from two complementary approaches: field study of individual cases by geologists and mathematical modeling by soils engineers. The difference in these approaches represents the slowly narrowing gap between the irregular, non-homogeneous reality of a landslide and the need to model that feature. The latter approach does not directly pertain to the present work, but both have been useful and both are necessary in considering a specific site for human activity.

There is a wide range of landslide phenomena, differing in such aspects as the nature of the material moved, the sliding surface, and the rate of movement. The first of the most notable (English language) landslide classifications was that of Howe (1909), whose classification of landslides in the San Juan Mountains was based upon the presence or absence of solid rock and was subdivided by the type of motion. Sharpe (1938) established a broader scheme, based on the type of material, and the type and rate of movement; his was the basis of the classification of Varnes (1958) which was adopted by the Committee of Landslide Investigations of the Highway Research Board. Terzaghi (1925) proposed

a classification based on the physical properties of the rocks involved. Záruba and Mencl (1969) used a scheme based upon the character of the rocks involved and the type of movement. Each system reflects the nature of the features in its author's locale. The classification in the present study appears to most resemble a Russian division done by Savarenski in 1937 (Záruba and Mencl, 1969, p. 31).

Much of what has been learned about landslide dynamics has come from detailed studies of specific landslides. Work by Shreve on the earthquake-triggered Sherman landslide in Alaska (1966); by Hadley (1964) on the Hebgen Lake, Alaska slide; by Merriman (1964) on the Portuguese Bend landslide, California; and by Helen Varnes (1949) on the Cedar Creek, Knife Edge and Ames landslides in southwestern Colorado have provided data on dimensions, rates, types of movement, nature and volume of debris moved, and other parameters and effects of those landslides.

Studies of landslides on a regional scale include those of Howe (1909) in the San Juan Mountains, Colorado; Rapp (1960) in Karkevagge, northern Scandinavia; Engelen (1967) in the Dolomites, Italy; Záruba and Mencl (1969) in Czechoslovakia; Bailey (1971) in Teton National Forest, Wyoming; and Fisher (1973) in southeastern Ohio. Comparison of numerous landslide features in the same region has made it possible to generalize about the behavior of certain geologic formations and about local controls of landsliding.

Although the present field area and its environs were not previously examined, studies have been made in other areas of Colorado. Howe, in his 1909 paper on landslides in the San Juan Mountains, pointed out that Cretaceous sediments, particularly the Mancos Shale, were largely responsible for many of the larger slides. Atwood (1918) identified landslide deposits in the San Juan Mountains so that they could be avoided in selecting reservoir sites. Helen Varnes (1949), also working in the southwestern part of the state, described several important slides, emphasizing the "lubricating" and loading roles of water and the instability of the Mancos Shale. More recent studies, instigated by specific construction problems, include the investigation by Irvine (1963) of the Ralston Creek landslide south of Boulder which required the repositioning of a section of State Highway 93; the work of Wahlstrom and Nichols (1969) on an old slide reactivated by construction of the Dillon Dam; and the

work of Barrett and Cochran (1971) on landslide problems in the Straight Creek valley associated with Interstate Route 70 and the western approach to the Eisenhower Tunnel.

Literature concerning the use of air photographs in landslide research is scant. Air photography has been recognized as a very useful tool for the identification of landslide features and its method of use has been summarized by Liang and Belcher (1958) and by Záruba and Mencl (1969). Gagnon (1971) used air photographs in studying landslides at St. Jean Vianney, and Bailey (1971) used air photographs to map landslides in the Tetons. The use of infrared imagery to detect landslides and monitor their movement is reported by Chandler (1972) and by Greeley and Blanchard (1974).

### Field Area

The field area of this investigation (Figure 1.3) covers about 68 square miles of Pitkin County in the Elk Range of the Rocky Mountains in western Colorado (longitude  $106^{\circ}52'30''\text{W}$  to  $107^{\circ}00'\text{W}$ , latitude  $39^{\circ}05'30''\text{N}$  to  $39^{\circ}15'\text{N}$ ). It is shown on the Highland Peak and Maroon Bells 7.5 minute U.S.G.S. topographic quadrangles. The southern two-thirds of the area is in the White River National Forest, with a strip of the southwest corner included in the Snowmass-Maroon Bells Wilderness Area.

Geologically, the area consists primarily of Pennsylvanian to Cretaceous sediments, with occasional Tertiary granodioritic intrusions: Sam's Knob, in the Snowmass Ski Area, was elevated by an intrusion of hornblende granodiorite. In the southern third of the area, the Maroon Formation, a Pennsylvanian to Permian red formation of siltstone, sandstone, mudstone, and conglomerate is predominant. The youngest sediments, those of the Mancos Shale (Cretaceous), form the low dry ridges that characterize the northern third of the area. Intermediate aged sedimentary strata are exposed in the central sector, with State Bridge Formation, Chinle Formation, Entrada Sandstone, and Morrison Formation visible on cliffs and Dakota Sandstone and the lower Mancos shale and limestone members visible in extensive outcrops. Beds in the central sector typically dip to the north at  $25^{\circ}$ . Morainal deposits, talus, and landslide debris cover about one-fourth of the surface of the region.

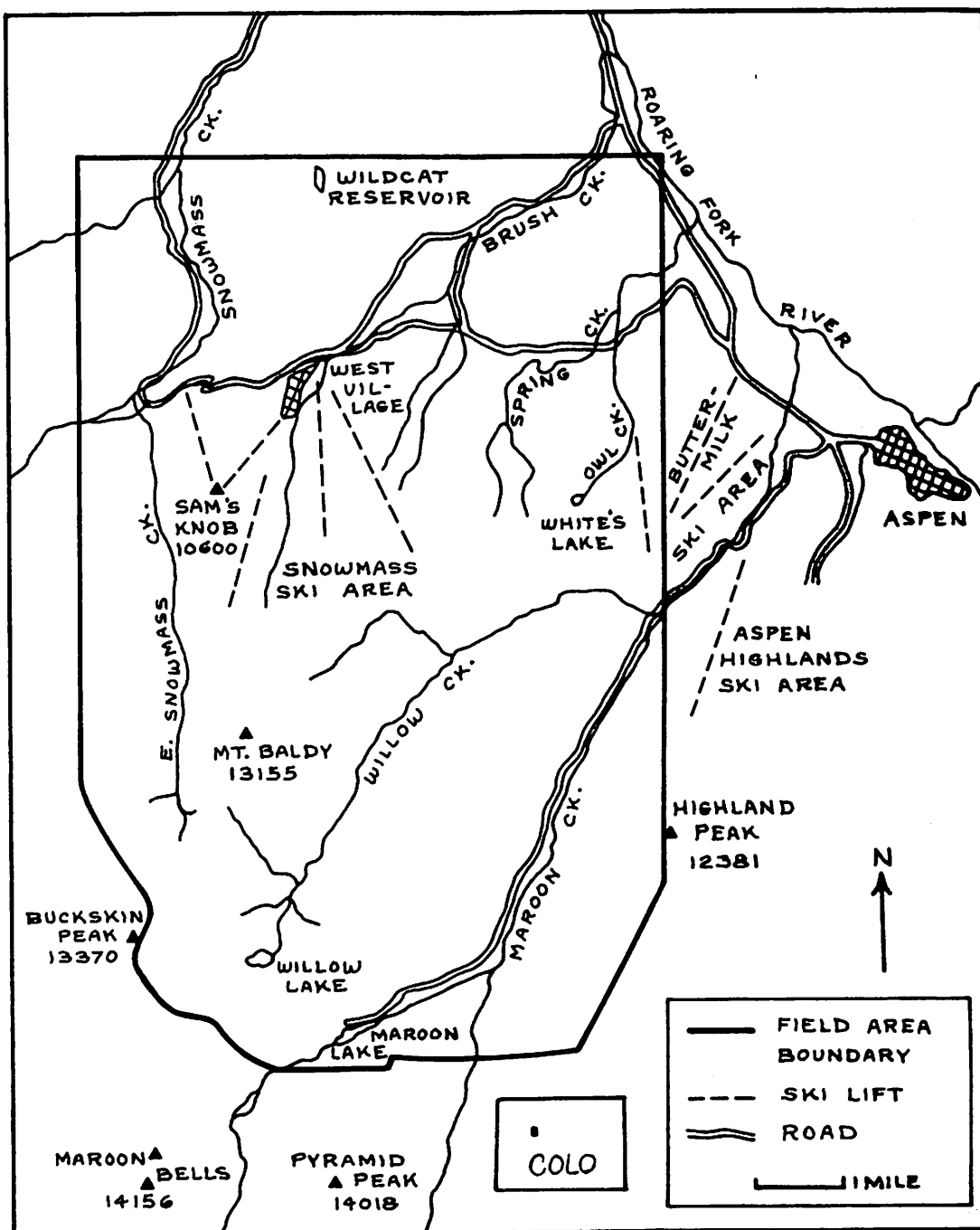


FIGURE 1.3  
Area of Study



The area includes the drainages of East Snowmass Creek, Brush Creek, Owl Creek, Willow Creek, and Maroon Creek. Elevations range from 7,600 feet to 13,336 feet. South-facing slopes are dry, characteristically vegetated by sagebrush, scrub oak, and aspen (above 9,500 feet), while north-facing slopes host forests of aspens (below 10,000 feet), spruce, and fir. Timberline is at approximately 10,800 feet, but is interrupted by large burns and numerous avalanche paths.

The area lies about two and one-half miles west of the city of Aspen, and contains Snowmass Resort (including the town of West Village, the Snowmass Ski Area, and three residential subdivisions) and the western portion of the Buttermilk Ski Area. The remaining land in the private sector is irrigated ranch land. Water is diverted to Snowmass Resort from East Snowmass Creek.

Further residential development is expected to include a 10,000 bed resort complex in the Owl Creek drainage and additional homes built in the Snowmass Creek, Brush Creek, and Owl Creek valleys. A light railroad, to connect Snowmass Resort and the new Owl Creek Resort with the city of Aspen, has been proposed and the Aspen Skiing Corporation plans to extend the Snowmass Ski Area eastward to join the Buttermilk Ski Area. Little or no development is anticipated in the valleys of Maroon, Willow, or East Snowmass creeks.

### Methods

This investigation was based on a study of air photographs supported by previous geologic mapping of the area (Bryant, 1969, 1972) and field work. The air photographs used were U.S.D.A. Forest Service black and white contact prints at a scale of 1:20,000, taken in October 1956 and September 1958. Although better quality, more recent photos would have been used had they been available, it is useful to have a set of air photographs of this area before the ski resort development took place. Ancient features are more easily identified on photographs where they are undissected by roads, ski runs, and building complexes.

The following operational definitions were established for the two types of landslides being identified: in profile, a rotational slide is expected to have a fracture zone or steep scarp at its head, an upper concave region from which material was removed, and a lower convex zone

of deposition, usually with a lobate tongue (Figure 1.1). A planar slide appears as an area of colluvial deposition downslope from bedrock with dip similar to the slope angle of the hillside (Figure 1.2).

Some features identifiable on air photographs that might be characteristic of either type of slide include a break at the scarp, arcuate ridges at terminal positions, lobate fronts, hummocky terrain, abrupt changes in vegetation between the slide and stable areas, seemingly anomalous deflection of stream drainage, streams along lateral boundaries, small ponds or bogs associated with either headward or terminal regions, lack of bedrock outcrops (in deposition zones), and a mass of material which seems to fit like a puzzle piece uphill. For recent slides, these characteristics can be striking, but for slides hundreds or thousands of years old, the scarps have been rounded off; erosion, surface movement, and vegetation of the slide have altered the appearance; the exact boundaries of the original feature have become obscure. None of the above criteria alone is decisive, but singly or in combination, they are useful in establishing areas to be investigated further. These criteria may be unrelated to rotational or planar slides. Some of the best examples of lobate tongues are those of rock glaciers, either active or relict and vegetated. Hummocky terrain may also be associated with solifluction or glacial deposits. It can be difficult to distinguish hummocky morainal features with no sliding history from landslides which might have been produced by the failure of morainal material.

The procedure for identifying landslides in this study was as follows:

- (1) Areas that showed one or more of the above characteristics, either from the field or from air photographs, were outlined as possible slides.
- (2) Field checking, an examination of the geology, and further scrutiny of the air photographs provided the information needed to decide whether the areas met the established operational definitions and, if so, where to locate their boundaries.
- (3) Boundaries were drawn on the air photograph or on a Mylar overlay and then transferred onto a topographic base map of the same scale.
- (4) Each landslide was numbered for reference.

Due to distortion away from the centers of the air photographs, it proved to be most accurate to transfer the landslide boundaries onto the base

map by drawing them according to the contours rather than by directly tracing them from the Mylar. The combination of air photograph interpretation and field reconnaissance made it possible to develop a more accurate map than would have been possible using either method alone.

Sixty-two landslides were mapped, the smallest of which was large enough (439 yards by 138 yards) to be identified on the air photographs. The completed map (Plate I) was the data source for the form, location, and spatial distribution of landslides in the area.

## CHAPTER II

### SPATIAL DISTRIBUTION

The presence or absence of spatial patterns, found by statistical analysis of the distribution of a geomorphic feature, can be indicative of parameters controlling the occurrence of that feature.

Four types of spatial data are cited by Hammond and McCullagh (1974, p. 33): (1) point distributions, (2) line distributions, (3) discrete areal distributions, and (4) continuous areal distributions. They note that whether a place is represented on a map by a point or by a discrete area is a function of the scale of the map. Although expected sampling distributions are known for a number of distribution models for points, most location measures and indices for discrete areal distributions are ignored "because nothing is known about their sampling distributions, if indeed they are derivable in mathematical terms" (King, 1969, p. 114). Deviation of a discrete areal distribution from uniformity may be computed using a Lorenz curve (Hammond and McCullagh, 1974, p. 55) and the associated Gini coefficient of concentration (King, 1969, p. 114); but any statistical method with its distribution based upon the Poisson model, e.g., randomness or nearest neighbor statistic, requires a sampling distribution of points rather than areas.

Thus, since spatial analysis of point distributions is less complicated and more meaningful than that of discrete areal distributions, it would seem to be advantageous to effectively change scales and reduce discrete areas to points. In fact, this is found to be standard practice in the geographical literature, whether it concerns urban centers (King, 1968), grocery stores (Getis, 1964), plant species (Curtis and McIntosh, 1950), or sorted patterned ground (Caine, 1972). While researchers acknowledge that some information is lost when areas are identified as points (King, 1968, p. 161; Bachi, 1962, p. 108), the validity of reducing relatively small areas to points is unchallenged.

In the present study, each landslide is represented by the point which is its center of gravity (two-dimensional). Effects of the space occupied by the landslides are considered in evaluating the statistical results. Using a polar planimeter, landslides were found to occupy

9.46% of the total field area. Some portions of the area contain no slides, while other portions contain many, some with common borders. The extent and nature of landslide contiguity in this field area must be explored before statistical tests, presuming independence of the features, can be used.

By defining a landslide as "contiguous" if any of its borders lie within one hundred feet of another landslide, 42 (68%) of the 62 slides are found to be contiguous. Twenty-eight slides are contiguous to only one other slide, while 14 slides are contiguous to two or more. The need for a greater understanding of autocorrelation effects in studies of spatial correlation has been stated (King, 1969, p. 160), but available treatments of contiguity deal with contiguous sampling units, not features (Geary, 1968; Dacey, 1968).

Spatial contiguity of the landslides could reflect a dependent relationship of slides, e.g., that one slide triggers another, and/or spatial concentration of conditions favorable to landsliding. The existence of contiguous landslides demonstrates the latter to be true. While it is possible that movement of one landslide may trigger that of another, there is no evidence for or against temporal contiguity of these features which would be needed to establish dependence. Since it is not possible to show that contiguous landslides are dependent, their independence is assumed, in order to allow the distribution to be analyzed statistically. Should this assumption prove to be incorrect, statistical results will be interpreted as having fewer degrees of freedom.

Two approaches to spatial analysis have been used in order to permit comparison of the results. Convergence of the results from both methods would tend to increase confidence in those results, while a divergence might expose an important control area, problem, or inconsistency in the method. The first approach uses point density grids superimposed on the map, with counts made of the number of points in each quadrat. The second approach used interpoint distances.

#### Quadrat Sampling Methods

After some experimentation with grids of varying shapes and sizes, it was found that quadrats of square or hexagonal cells enclosing an area of one square mile were preferred because they provided reasonable

coverage of the area within the field boundaries and because they provided numbers conducive to the use of the chi-square test. No assumptions were made about the area outside the field area; rather only those quadrats and half-quadrats which fit within the boundaries were counted. Repeated samples were taken with each grid in differing orientations. Samples were then compared with expected frequencies computed for uniform and random distributions. In order to check the validity of reducing landslides to points, each point count was accompanied by a count of the number of landslides intersecting the area of each cell.

#### *Uniform Distribution*

Points distributed uniformly over an area are located in accordance with their density. In this case, with 62 points in an area of 68 square miles, 0.917 uniformly spaced points would be expected in every square mile, or one point in every 1.09 square miles. Sampling results show that for this field area 60% of the square miles sampled contained no slides at all, 23.5% of the square miles contained two or more, and only 16.5% contained a single slide. Testing these observations against a uniform distribution model (Table 2.1) indicated that the hypothesis of a uniform distribution of landslide features in the present area can be rejected at the 1% level of significance: landslides are not uniformly distributed across the study area.

#### *Random Distribution*

In a random distribution of points overlaid by a grid, any point has an equal probability of occurring in any quadrat and any quadrat has the same chance of receiving a point as any other quadrat. The position of a random point is independent of the position of any other point. If the density of points is low enough that the points are distinct and independent, the Poisson distribution, a discrete frequency distribution, can be used as a test for randomness (David, 1973, p. 305). The expected number of quadrats containing a given number of points (from zero to six) was computed using the Poisson probability expansion, giving the cumulative probability of the frequency of  $n$  points per

TABLE 2.1

Comparison of Samples to a Uniform Distribution  
Using Chi-Square Test

Null hypothesis = samples are uniformly distributed.

<u>Points/Quadrat</u>	<u>Expected(E)</u> *	$\frac{(O_1 - E)^2}{E}$ (Sample 1)	$\frac{(O_2 - E)^2}{E}$ **
0	3	463.36	456.33
1	56	35.97	39.31
2 or more	3	<u>62.33</u>	<u>75.0</u>
	chi-square =	561.66	570.64

\*Expected derived from density of .917 landslides per quadrat.

\*\*Samples were obtained using differing grid orientations to eliminate any effect of grid orientation. Chi-square test shows these samples to be identical at 1% level of significance.

With two degrees of freedom, the null hypothesis is rejected at the 1% level of significance.

class as

$$e^{-\rho} (1 + \rho + \sum_{n=1}^{\infty} \rho^n / n!)$$

where  $\rho$  is the mean density of landslides over the total area.

Observed frequencies were compared to expected frequencies using the chi-square statistic to determine whether the observed samples were randomly distributed. Chi-square results, shown in Table 2.2, indicate that randomness can be rejected at the 1% level of significance for all samples. Figure 2.1 illustrates the relationship between observed samples and the Poisson model. It can be seen that the greatest deviations from the model are at zero slides per quadrat where the observed exceeds the model and at one slide per quadrat where the model exceeds the observed.

The ratio of landslide points per quadrat to the total number of landslides was compared graphically to the ratio of landslide areas intersected by each quadrat to the total count of slide intersections (Figure 2.2). Expected percentages of the landslide population for each class, derived from the Poisson distribution, are also shown. Differences between the observed histograms and the expected random histogram are generally far greater than differences between the two sets of observations. Furthermore, the sample based on landslide areas intersected tends to deviate more from the expected than does the point sample. This comparison suggests that, in this case, the area covered by landslides does not reduce the significance of the test for randomness. In fact, consideration of the space occupied by landslides tends to increase the difference between the observed population and a random distribution.

It can be concluded, then, that the distribution of landslides in this study area is significantly non-uniform and non-random. The high rate of contiguity between slides and the elimination of random or uniform distributions suggests that landslides tend to be clustered. The nearest neighbor statistic, computed without the use of quadrats, provides an index of the departure from randomness toward clustering. Extension of the nearest neighbor analysis to the  $n^{\text{th}}$  neighbor will help to define the predominant cluster sizes.



TABLE 2.2

Comparison of Samples to Random Distribution

Using Chi-Square Test

Null hypothesis = samples are randomly distributed.

$S_i = i^{\text{th}}$  sample.

Values in each column are: (observed number  
- expected number)<sup>2</sup>/(expected number).

Points per Quadrat	<u>Expected</u>	<u>S<sub>1</sub></u>	<u>S<sub>2</sub></u>	<u>S<sub>3</sub></u>	<u>S<sub>4</sub></u>	<u>S<sub>5</sub></u>	<u>S<sub>6</sub></u>
0	27.32	6.85	3.08	3.43	6.80	6.80	6.80
1	24.91	7.77	6.69	4.78	11.11	11.78	10.46
2	11.36	0.99	0.24	0.01	0.13	0.27	0.04
3 or more	<u>4.40</u>	<u>4.81</u>	<u>1.50</u>	<u>1.50</u>	<u>4.53</u>	<u>6.67</u>	<u>2.74</u>
Chi-Square =		20.42	11.55	9.72	22.57	25.62	20.05

For two degrees of freedom, critical values of chi-square are:

5.99 at 5% level of significance

9.21 at 1% level of significance

(Krumbein and Graybill, 1965, p. 418).

All samples are non-random at 1% level of significance.

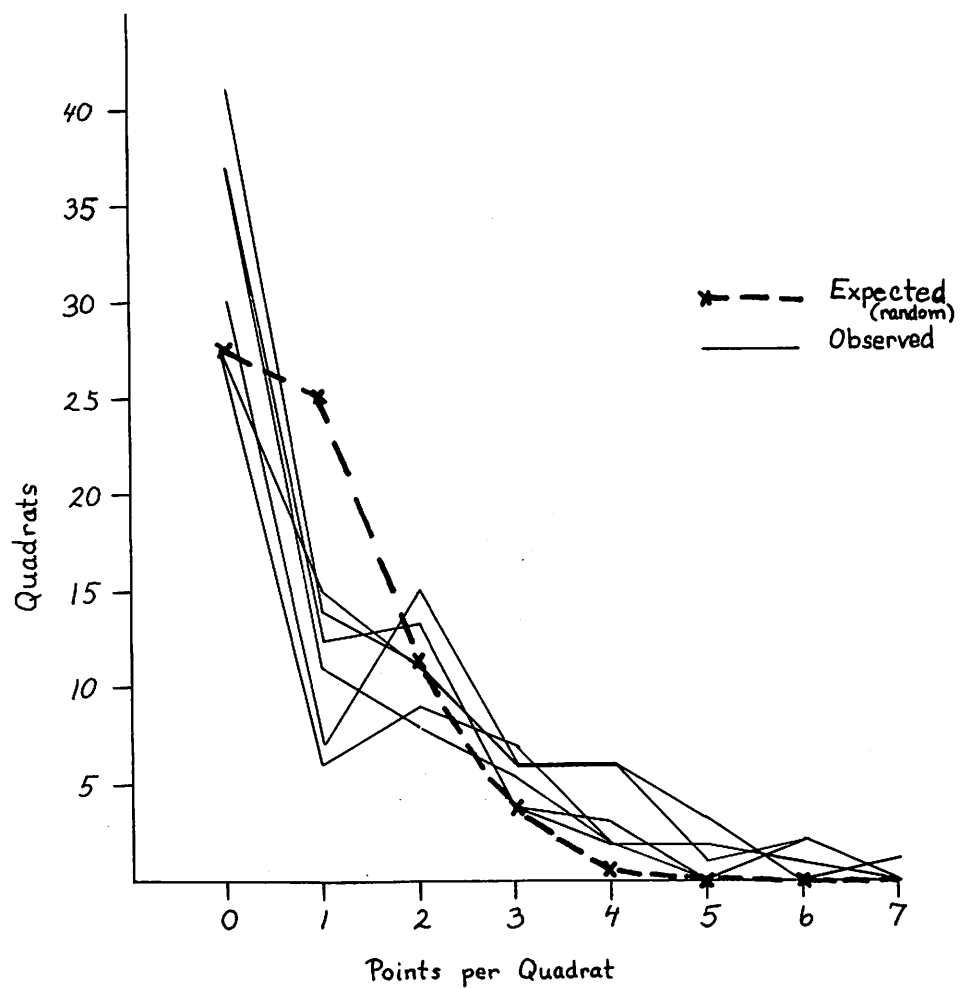


FIGURE 2.1

Quadrat Samples  
Compared to Random Distribution

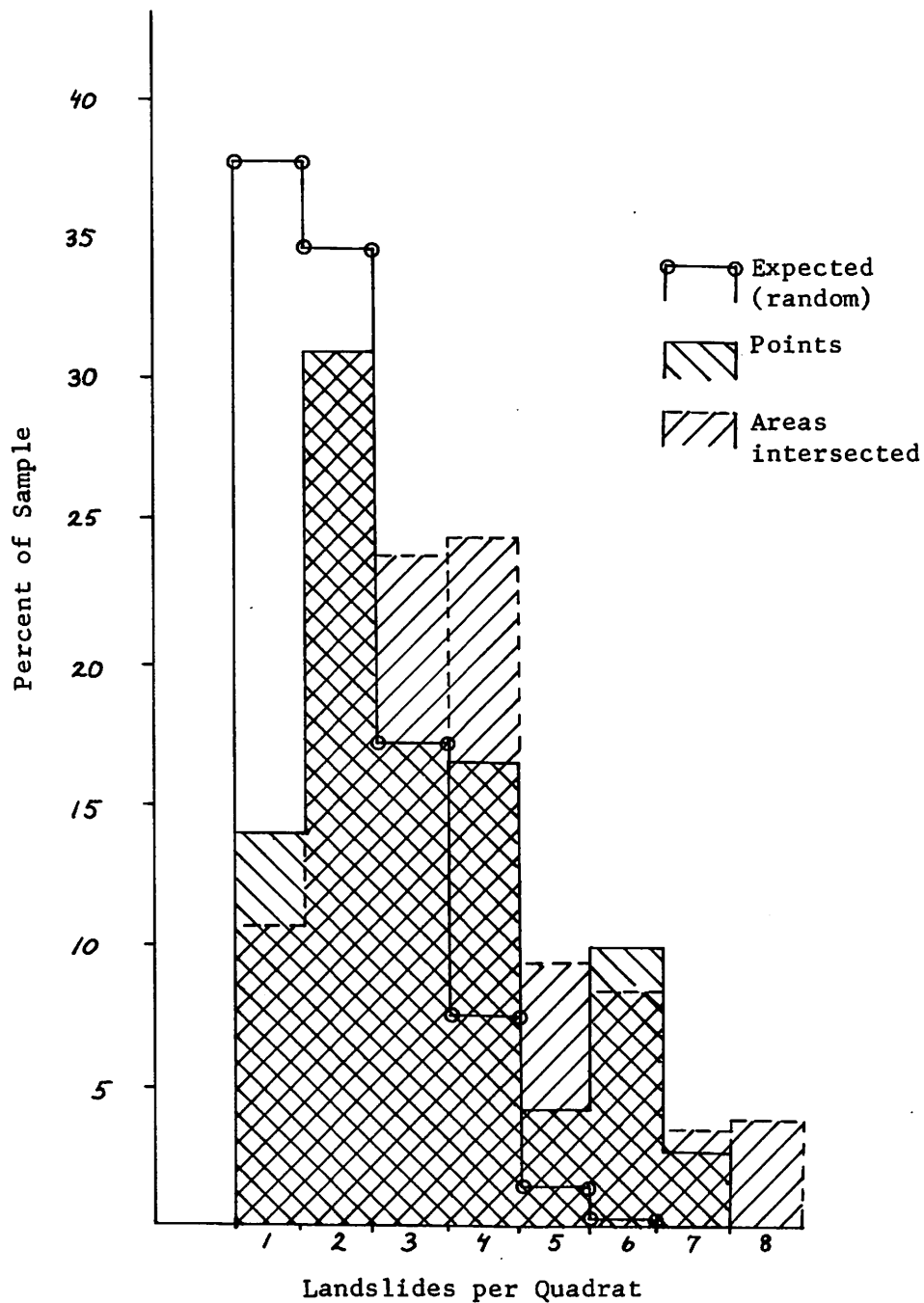


FIGURE 2.2

Comparison of Samples of Points and  
Samples of Areas Intersected

### Nearest Neighbor Analysis

Nearest neighbor analysis (Clark and Evans, 1954) provides a different test for randomness based on a Poisson distribution by avoiding the use of quadrat sampling. The mean of the distances from each point to its nearest neighbor is compared to an expected mean distance  $\frac{1}{2} \sqrt{\rho}$  which is derived from the density  $\rho$  of an infinitely large random distribution. The nearest neighbor statistic  $R$  is the ratio of the observed mean to the expected mean. A value close to unity is suggestive of a random distribution.  $R$  ranges from zero in the case of maximum clustering to 2.1491 in the case of maximum uniform spacing. The significance of a departure from randomness is tested by the normal curve.

Measurements were made from every point to its nearest neighbor, excluding those points located closer to the field area boundary than to another point. The mean observed distance between points was 0.32 miles, and  $R$  was found to be 0.668, significant at the 1% level. This statistic suggests aggregation, with nearest neighbors on the average only .668 as far apart as they would be in a random distribution (Table 2.3).

TABLE 2.3

#### First Nearest Neighbor

$$\text{Density } (\rho) = 62/67.5 = 0.9185$$

$$\text{Expected mean distance} = \frac{1}{2} \sqrt{\rho} = 0.4792$$

$$\text{Observed mean distance} = 0.32$$

$$R = \text{Observed/Expected} = 0.6678 \text{ (somewhat clustered)}$$

$$\text{Standard error Expected} = \text{S.E.} = 0.26132/\sqrt{N\rho} = 0.0361$$

$$v = \text{standard variate of normal curve} =$$

$$(\text{Expected} - \text{Observed})/\text{S.E.} = 4.41$$

$$v = 2.58 \text{ for 1\% level of significance}$$

To obtain more detail with respect to the clustering trend, the analysis was extended to include not only the first nearest neighbor, but also the second, third, fourth, and in some cases fifth or even sixth. The extension of nearest neighbor analysis is based upon the reflexive property of neighbors that occurs when the  $n^{\text{th}}$  nearest neighbor of A has A as its  $n^{\text{th}}$  neighbor. For each neighbor order, the ratio of the number of reflexive cases to the total number of points surveyed is compared with an expected value derived for the proportions of individuals in an infinite random population (Miller and Kahn, 1962, p. 385).

Results are shown in Table 2.4 and Figure 2.3. Observed values exceed the expected for the first and fifth nearest neighbors, indicating a tendency to cluster in pairs and in groups of six. The maximum deviation of observed from expected is at the third nearest neighbor, showing that groups of four landslides occur less frequently than would be expected in a random distribution. These results concur with the high incidence of contiguous landslides and with the striking visual suggestion of pairing (Plate I). While pairs of landslides are common, they tend to be isolated from other pairs such that reflexive groups of four are not common. Larger clusters occur in particularly extensive favorable

TABLE 2.4

Higher Order Nearest Neighbor Analysis

<u>Neighbor Order</u>	<u>1<sup>st</sup></u>	<u>2<sup>nd</sup></u>	<u>3<sup>rd</sup></u>	<u>4<sup>th</sup></u>	<u>5<sup>th</sup></u>	<u>6<sup>th</sup></u>
<u>Observed</u> (O) <sup>1</sup>	.7018	.3529	.0500	.0606	.1875	0
<u>Expected</u> (E)	.6215	.3863	.2401	.1492	.0927	.0576
(O - E)	+.0803	-.0334	-.1901	-.0886	+.0948	-.0576
<u>Number of Cases</u>	57	51	40	33	32	24

<sup>1</sup>Ratio of reflexive cases to total number of cases.

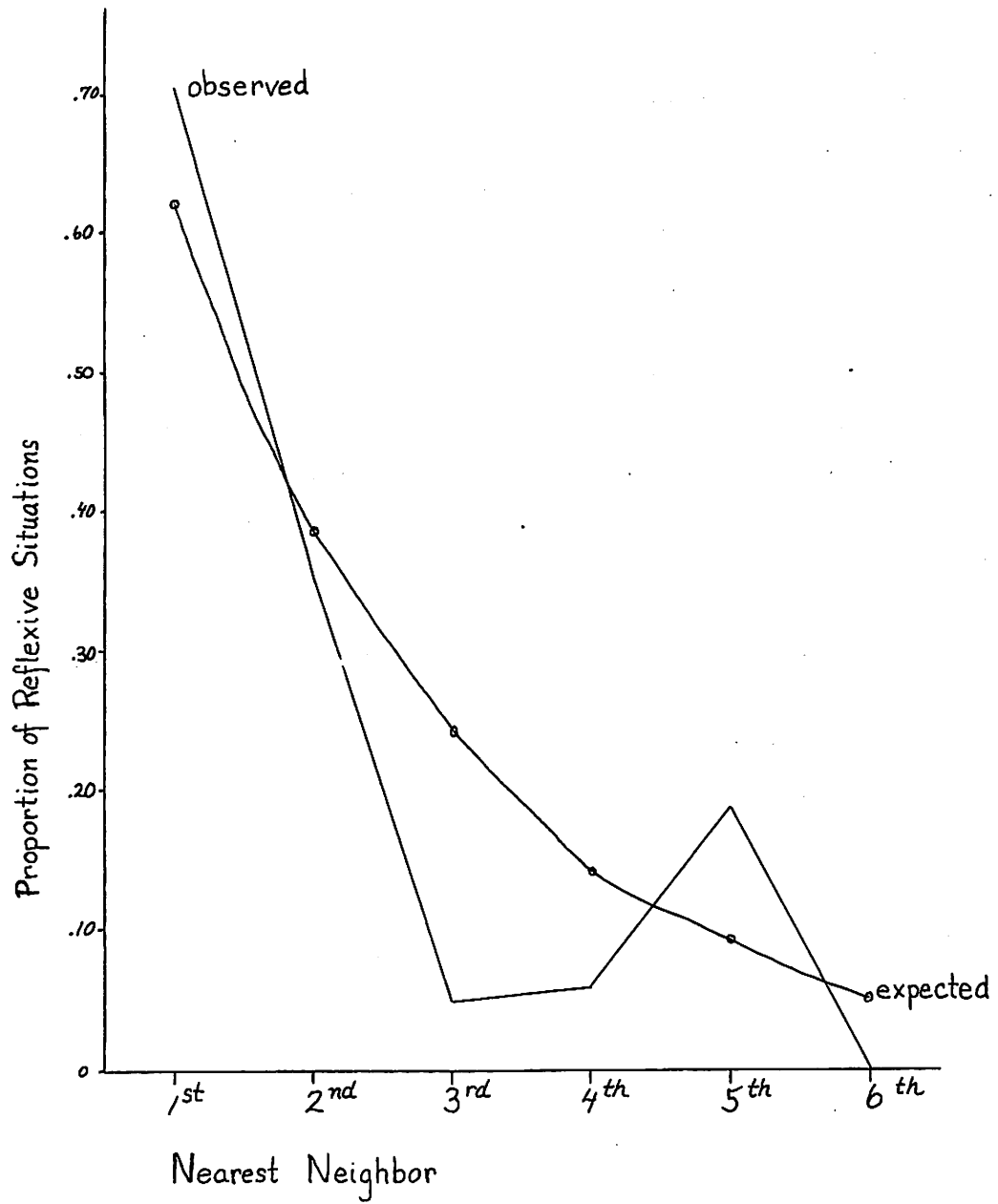


FIGURE 2.3

Nearest Neighbor Analysis

sites. No meaningful relationship between two or six times landslide size and drainage density or slope size has been found.

### Conclusion

The statistical analysis performed here shows the distribution of landslides in geographic space is neither uniform nor random but tends to be clustered. The tendency toward clustering and the high incidence of contiguity suggest either that conditons favorable to landsliding are spatially controlled or that landsliding tends to promote, or trigger, more landsliding. Although 68% of the slides are contiguous, the lack of any direct evidence for a triggering relationship between individual slides makes the former alternative more acceptable than the latter. Results obtained using both quadrat sampling methods and nearest neighbor analysis are highly significant and would still be significant if contiguous features were not completely independent, giving fewer degrees of freedom.

The treatment of the entire field area as a whole has revealed a pronounced tendency of landslides to be clustered there. It is now necessary to investigate the landslides individually in order to determine the nature of the spatial control of these features.

## CHAPTER III

### CHARACTERISTIC FORM AND LOCATION

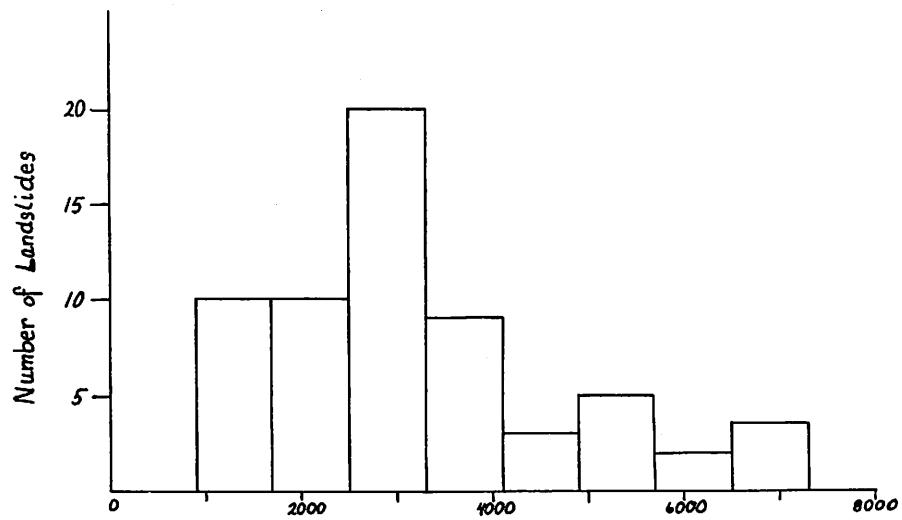
#### Form

The form, or shape, of a landslide can be an important indicator of the nature of its movement. Extensive studies of landslide forms and their relation to landsliding processes were made by Crozier (1973) who developed a series of morphometric indices for describing landslide features and found that the morphology of a landslide is closely related to its dominant genetic process. Using the parameters of overall landslide length, depth of the displaced mass, lengths and widths of the convex and concave portions of the slide form, length of the exposed surface of rupture, and length of the displaced material, he proposed indices of classification, dilation, tenuity, displacement, fluidity, flowage, and viscous flow. The lack of freshness of the landslides in the present study and the use of air photographs, however, made it impossible to measure those parameters that form Crozier's indices.

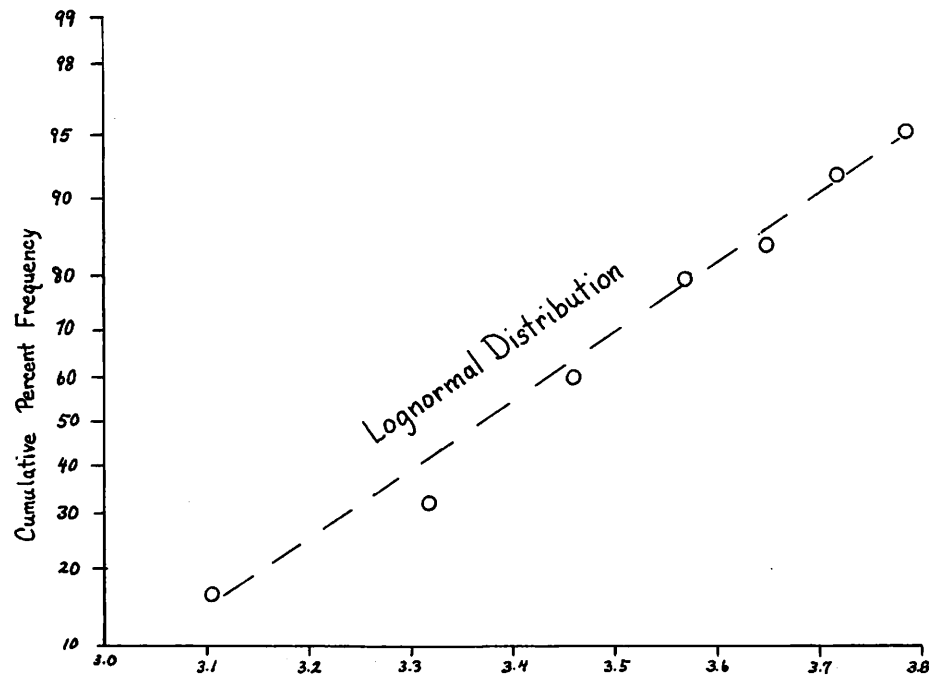
The following discussion of landslide form is included in the present study to assist in interpreting landslide process and the behavior of local materials, and also to provide characteristic forms of local slides that could be useful in predicting the extent of potential hazard areas. Form will be expressed as the ratio of the length of each feature to its width.

Map length was taken to be the longest dimension perpendicular to the slope contours from head to toe of the slide. Actual slide length was computed as the ratio of the map length to the cosine of the slope angle. Values ranged from 874 feet to 7330 feet. The frequency distribution of length is given in Figure 3.1(A). Length values do not conform to a normal distribution. The distribution of the logs of length values has a skewness coefficient of  $-.140$  and was found to fit a lognormal distribution at the 1% level of significance (Snedecor and Cochran, 1967, pp. 86, 552). This good fit with the lognormal model (Figure 3.1(B)) would seem reasonable as both the lognormal distribution and the landslide lengths are defined only for positive values, whereas a normal distribution extends infinitely into the negative. Geometric mean length





(A) Length (in feet)



(B) log Length

FIGURE 3.1

Distribution of Landslide Lengths

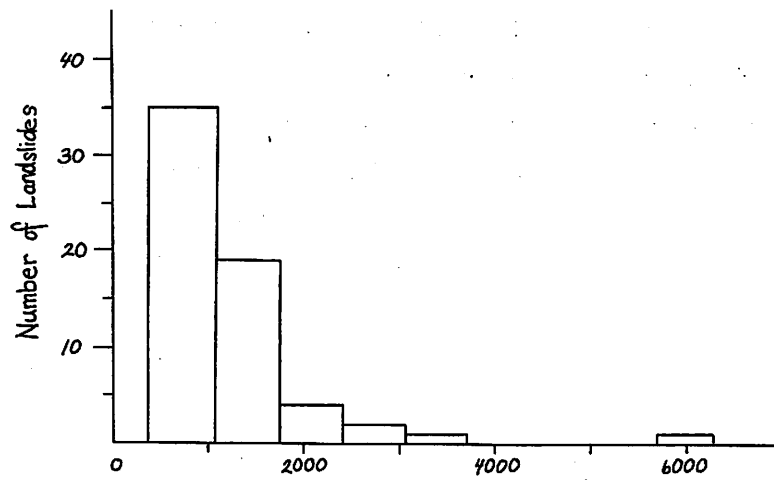
is 2860 feet. It is expected that landslide length may be controlled by slope lengths, which also follow a lognormal distribution (Speight, 1971). A discussion of topographic controls of landslide length is found in the third section of this chapter.

Width was measured as the greatest dimension perpendicular to the length. Values ranged from 414 feet to 5694 feet. The frequency distribution of widths is shown in Figure 3.2(A). Widths follow neither a normal nor lognormal distribution, with 87% of the widths measured found to be less than 1734 feet. That 87%, however, is found to conform to a lognormal distribution at the 1% level of significance (Figure 3.2(B)). Mean width of the 87% less than 1734 feet is 892 feet. It appears that widths up to 1734 feet are characteristic of landslides in this area, probably controlled by the nature of the sliding material and the topography. There does not appear to be any correlation between width classes and bedrock types, slope angles, effective dip of bedding surfaces, or slope orientation. Each of the particularly wide landslides is also particularly long.

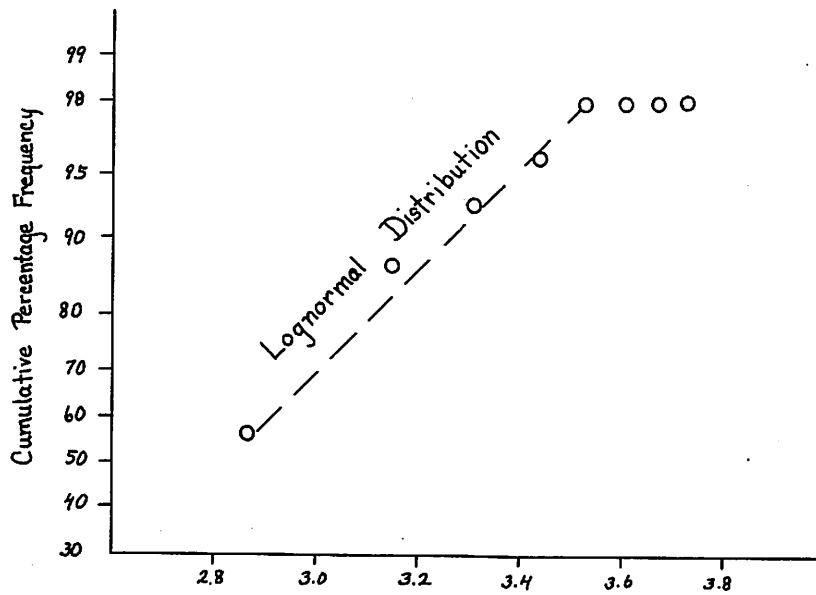
The shape ratio of length to width ( $L/W$ ) was computed for each landslide. The distribution of shape ratios, shown as a histogram in Figure 3.3(A), is not normal. The distribution of the logs of shape ratios, shown in Figure 3.3(B), has a skewness coefficient of 0.16. This distribution is sufficiently symmetrical to be considered normal at the 1% level of significance. Mean shape ratio is 2.69. Although the distribution of  $L/W$  shows a tendency toward bimodality, that trend disappears in the distribution of  $\log L/W$ . The suggestion that there are two classes of landslides, one relatively longer and narrower than the other, was not upheld by comparisons between the two "modes" with respect to slope, bedrock type, dip, location, and relation to ridge crests and valley floors.

#### Location With Respect to Structure

Geologic structure of the region has been briefly described in Chapter I. A more thorough description is given in Table 3.1. "Bedrock types" of landslides are given as the structural unit on or within which movement took place, whether or not the moving mass was derived from that



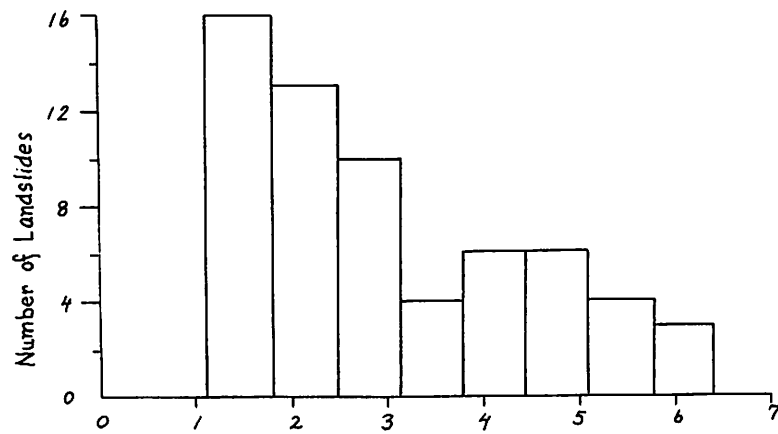
(A) Width (in feet)



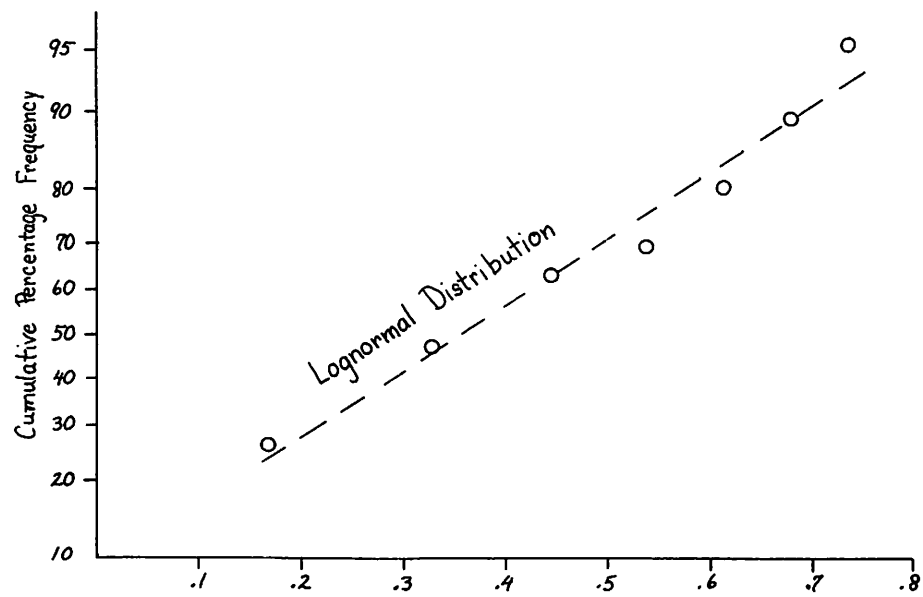
(B) log Width

FIGURE 3.2

Distribution of Landslide Widths



(A) Shape Ratio (Length/Width)



(B) log Length/Width

FIGURE 3.3

Distribution of Shape Ratios

TABLE 3.1

Stratigraphic Column

<u>Thickness</u>	<u>Formation</u> <u>Age</u> <u>Description</u>	<u>Number of</u> <u>Landslides</u>	<u>Comments</u>
	Horneblende granodiorite Tertiary Occurs as sills and dikes		Appears to have pushed up Sam's Knob.
4000 feet	Mancos Shale Cretaceous Shale and silty shale with a few sandstone beds and a limestone member.	8	So unstable and easily eroded that long steep slopes rarely exist. Contains numerous small slumps
200 feet	Dakota Sandstone and Burro Canyon Formation Cretaceous Siltstone and shale interbedded with sandstone; claystone and siltstone.	13	Landslides tend to be associated with both formations: permeable Dakota overlies less permeable Morrison.
400 to 500 feet	Morrison Formation Jurassic Siltstone and claystone grading downward into calcareous siltstone and sandstone.	10	
0 to 40 feet	Entrada Sandstone Jurassic	0	Too little exposure to map landslides
0 to 300 feet	Chinle Formation Triassic Calcareous siltstone, characterized by hackly fracture.	3	Rarely exposed.
0 to 500 feet	State Bridge Formation Permian and Triassic Calcareous and micaceous siltstone, sandstone, and polymictic conglomerate. Beds more persistent than Maroon Formation.	2	Rarely exposed.
7000 feet	Maroon Formation Pennsylvanian and Permian Interbedded calcareous and micaceous sandstone, siltstone, and conglomerate.	25	Covers majority of field area.

(Bryant, 1969, 1972)

unit. The bedrock types of slides associated with the Maroon, State Bridge, and Mancos formations were easily identifiable; whereas the bedrock types of slides associated with the Dakota, Morrison, and Chinle formations were more difficult to distinguish. This is because these latter formations have less thickness and tend to stack up in cliffs, and because there tends to be more interaction between formations, especially between the Dakota and Morrison (Lautenbach, et al, 1974), with a bed of the Morrison typically serving as the sliding surface for the Dakota.

A meaningful comparison of the frequency of landslides between different bedrock types was not possible, due to the number of complicating parameters such as irregular and unequal areas covered by each bedrock type, slope orientation of those areas, varying dip of the beds of each formation, and varying steepness of the slopes encompassed by it. More important is the fact that slides have been identified in all bedrock types that have enough exposure to be viewed from aerial photographs.

Using landslide shape as a key to the sliding process, a comparison was made between bedrock types and landslide shape ratios. Shapes of slides from the Maroon formation were compared with slide shapes from the Dakota and Morrison formations in a Mann-Whitney U test. Test results gave a  $z$  score of 0.495, showing no significant difference between the two groups. Failure to identify the characteristic differences in landslides associated with different bedrock types suggests that each type tends to respond to landslide stress in a similar manner, and/or that lithology is not as important to landsliding as some other control.

Throughout most of the area, the bedrock is composed of thin beds of sediments. The dip of these beds has a very strong influence on the topography and appears to be associated with the location of landslides. In order to compare the dip of the beds associated with each landslide, the vector of dip in the direction of the slide, was computed. This is given by the product of the magnitude of dip in the direction of dip and the cosine of the angular difference between the direction of dip and the direction of slide movement. The frequency distribution of effective dip is shown in Figure 3.4.

Of 61 landslides for which strike and dip data were available (Bryant, 1969, 1972), only two had beds dipping away from the slope. Vector mean and standard deviation were estimated from  $n$ , the number of landslides

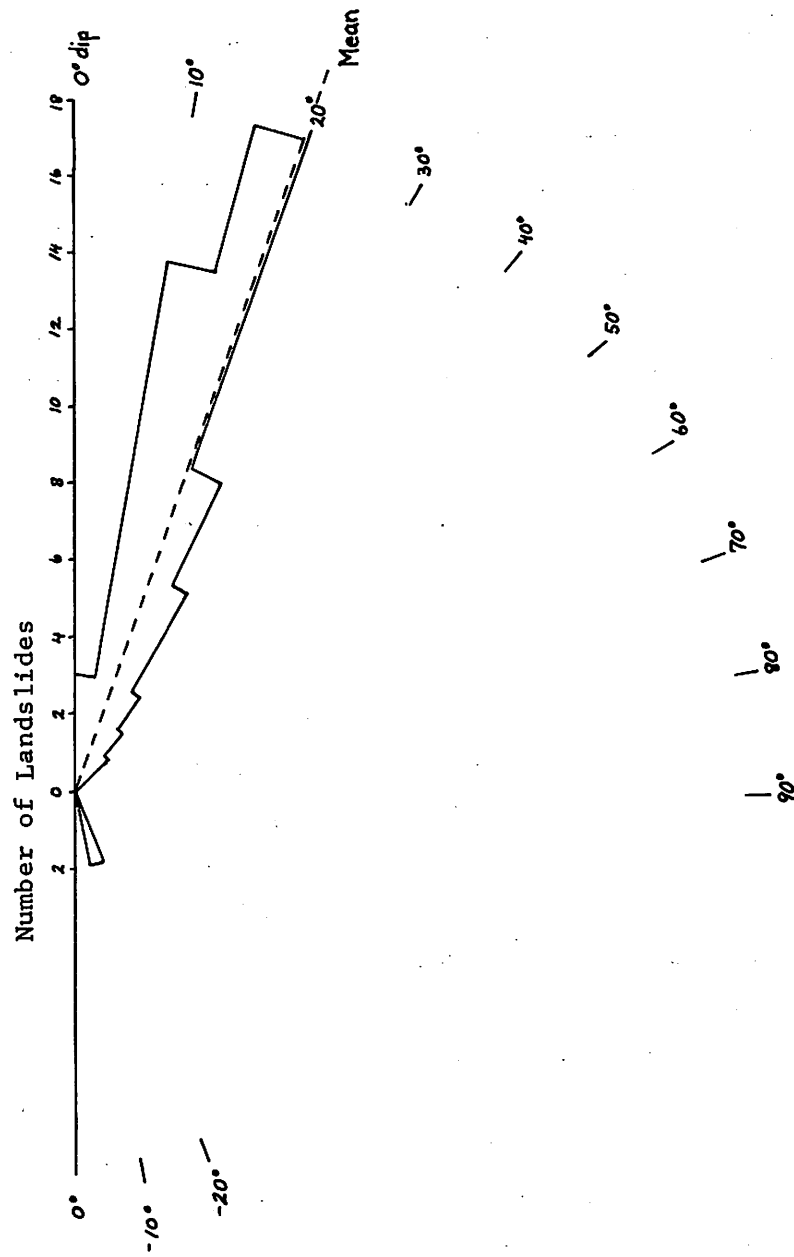


FIGURE 3.4

Distribution of Effective Dip

in the sample;  $\theta$  , the effective dip in degrees; and  $f_i$  , the frequency of  $\theta_i$  ; as:

Vector mean ( $V_m$ ) =  $\arctan \bar{S}/\bar{C}$ , where:

$$\bar{S} = 1/n (\sum f_i \sin \theta_i) \text{ and}$$

$$\bar{C} = 1/n (\sum f_i \cos \theta_i)$$

$$\text{Standard deviation } (s_o) = (-2 \log_e (\bar{S}^2 + \bar{C}^2)^{\frac{1}{2}})^{\frac{1}{2}}$$

(Mardia, 1972, p. 24, 26).

The mean effective dip is  $19.5^\circ$  below the horizontal, with standard deviation of  $14.3^\circ$ . Standard deviation is largely due to the inclusion of beds dipping in the opposite direction. The distribution of bedrock types for landslides in the group of 27 features having an effective dip greater than the mean, is compared with the distribution of bedrock types for all of the landslides in Table 3.2. Using the Kolmogorov-Smirnov test, these two distributions were found to be the same at the 1% level of significance. This suggests that the distribution of large effective dips is not preferential to any lithologic unit.

Planar slides would be expected to be directly related to the effective dip, whereas rotational slides would be expected to be independent of dip. In Figure 3.4 only two slides are seen to have no relation to dip, although the relation of slides with effective dips close to  $0^\circ$  is questionable. The seven landslides with effective dips less than one standard deviation below the mean occur on varying stratigraphic units: two from the Maroon Formation, two from the Chinle Formation, two from the upper member of Mancos Shale, and one from the lower member of Mancos Shale. With the exception of those two obvious cases of nonplanar slides, however, it is difficult to distinguish planar and rotational slides on the basis of dip. The smooth, normal (excepting the two negative values) distribution of effective dip indicates that dipping bedrock may play a more important role in nearly all slides than was initially postulated. Complicating the distinction between planar and rotational slides is the difficulty in separating the roles of dipping beds in establishing topography favorable to landsliding, and in serving as a slide surface for individual landslides.



TABLE 3.2

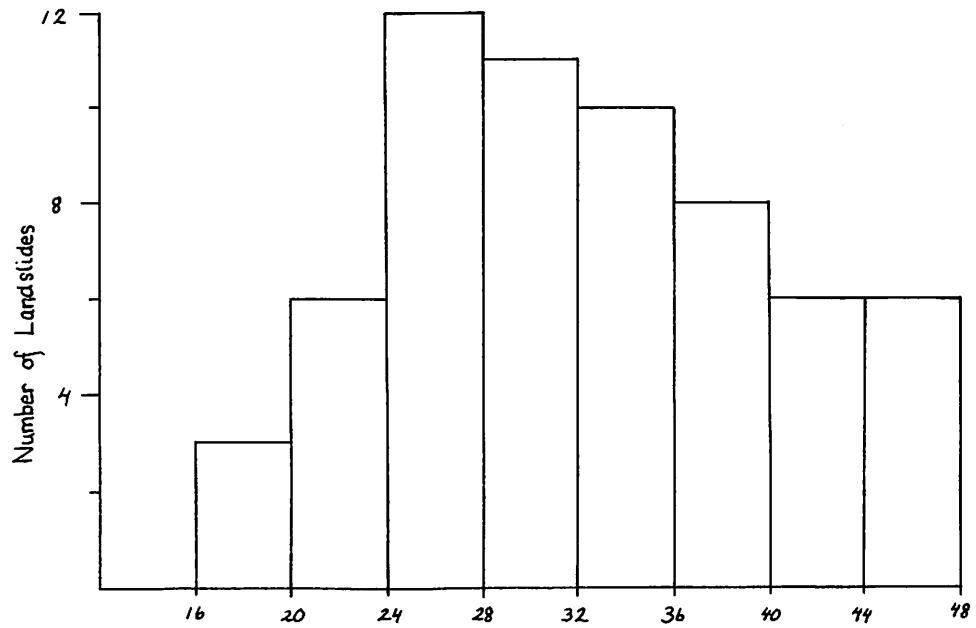
Landslide Distribution With Respect to  
Effective Dip and Lithology

<u>Bedrock Formation</u>	<u>Percent of Slides With Effective Dip Greater Than the Mean (19.5°)</u>	<u>Percent of Slides in Field Area</u>
Maroon	48%	41%
Dakota and Morrison	37	38
Mancos Shale	7	13
Chinle	4	5
State Bridge	4	2

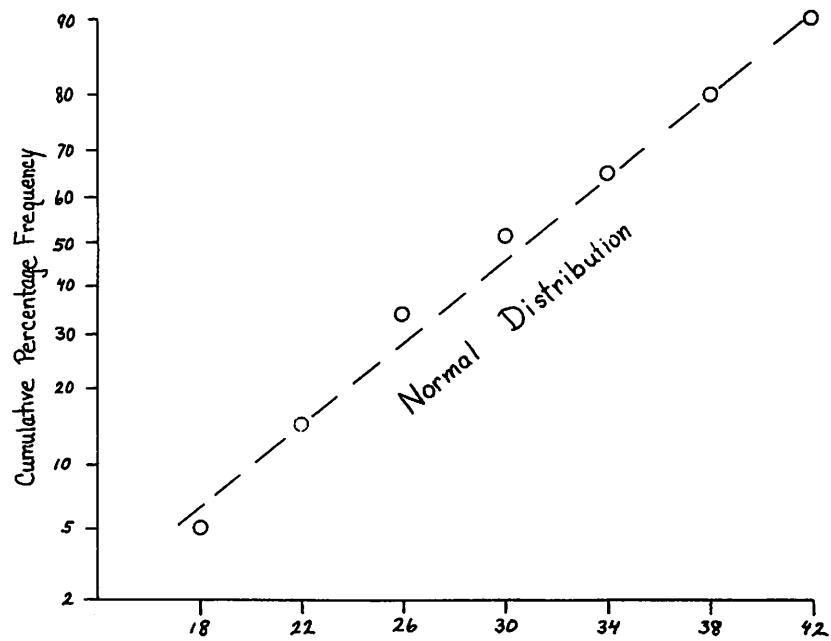
Location With Respect to Topography

The topography of the area has been briefly described in Chapter I. Except in the northern sector where the easily eroded Mancos Shale predominates, the nature of the slopes appears to be a direct function of the structure. In the direction of dip, slopes tend to follow the angle of dip, whereas cliffs are the rule in the direction opposing the dipping beds. The distribution of slope angle, measured at the starting zone of each landslide, is found to be normal at the 1% level of significance (Figure 3.5). Values range from  $16^{\circ}$  to  $44^{\circ}$ , with a mean of  $32^{\circ}$  and standard deviation of  $7^{\circ}$ . As no landslides in the area have a starting zone slope of less than  $16^{\circ}$ , that could be a useful figure for land use planning. Landslides with the lowest starting zone slope angles occur on the Dakota Formation. Table 3.3 gives the minimum slope in each bedrock that was found at the starting zone of a landslide.

It would seem that planar slides would be more likely to occur than rotational slides on slopes where the effective dip is equal to (within  $2^{\circ}$ ) or greater than the slope angle. Ten slides (16%) were found to have an effective dip exceeding the slope angle, 8% were equal, and 76% had effective dip less than the slope angle. All of the slides considered to be nonplanar due to low effective dips were among the 76%. Field research identified only four landslides as having unquestionably failed on a bedrock sliding surface. Their locations, effective dips, slope angles, and stratigraphy



(A) Slope Angle (degrees above horizontal)



(B) Slope Angle

FIGURE 3.5

Distribution of Landslide Slope Angles

TABLE 3.3

Minimum Slope Angles for Landslides  
on Different Bedrock Formations

<u>Bedrock Formation</u>	<u>Slope Angle</u>	<u>Number of Landslides</u>
Chinle	34°	3
Dakota	16.2	13
Mancos (lower member)	19.9	5
Mancos (upper member)	21.1	3
Maroon	22.3	25
Morrison	21.1	10
State Bridge	25.8	2

TABLE 3.4

Planar Slides

<u>Map Identification Number</u> <sup>1</sup>	<u>Location</u>	<u>Slope</u>	<u>Effective Dip</u>	<u>Bedrock</u>
51	Willow Ck.	26°	19.85°	State Bridge
33	Brush Ck.	26	19.99	Dakota
34	Brush Ck.	16	14.89	Dakota
10	Sam's Knob	21	30.0	upper member, Mancos Shale

<sup>1</sup>Landslides are identified on Plate I

are described in Table 3.4. Only two of these slides have an effective dip equal to or greater than the slope angle, as do one-half of the landslides with slopes less than the mean. All landslides with slope gradients greater than the mean had a slope angle greater than the effective dip. While it is possible that the relationship of effective dip to slope angle is critical for low-angled slopes, the fact that it is only important in half of the cases suggests that the mere existence of effective dip and other structural or environmental factors may be more significant controls. Table 3.4 demonstrates that identifying slides with effective dip greater than or equal to the slope angle does not provide adequate information for the differentiation of planar and rotational slides. Whether or not slides are considered planar seems to depend on whether dipping bedrock is exposed above and adjacent to the slide path.

Figure 3.6 illustrates the distribution of downstream orientations for the major valley drainages in the study area. The vector mean is  $34^{\circ}$  and standard deviation  $44^{\circ}$ . The distribution, however, appears to be bimodal, with modes generally to the north and to the east. Breaking the distribution at the population vector mean orientation makes it possible to analyze the two modes by standard parametric methods. Using the notation given on page 31, the vector mean  $V_m$  and standard deviation  $s_o$  of a  $180^{\circ}$  distribution may be estimated as:

$$V_m = \arctan \frac{(1/n \sum f_i \sin 2\theta_i)/2}{(1/n \sum f_i \cos 2\theta_i)}$$

$$\text{or } \arctan (\bar{S}_2/\bar{C}_2)/2.$$

$$s_o = 57.3 (-2 \log_e (\bar{S}_2^2 + \bar{C}_2^2)^{1/8})^{1/2}/2.$$

(Mardia, 1972, p. 26)

The results of this analysis are shown on Table 3.5.

Landslide orientations were obtained from the map and plotted in a rose histogram (Figure 3.7). Orientations range from  $292^{\circ}$  (west) to  $102^{\circ}$  (east), with one south-facing ( $177^{\circ}$ ) landslide. Using orientations of all landslides, the mean was found to be  $25.9^{\circ}$ , with standard deviation of  $54^{\circ}$ . The large standard deviation and the lack of points in the vicinity of the mean made it seem reasonable to analyze this as a bimodal distribution, in the manner described above. Results are given in Table 3.6.

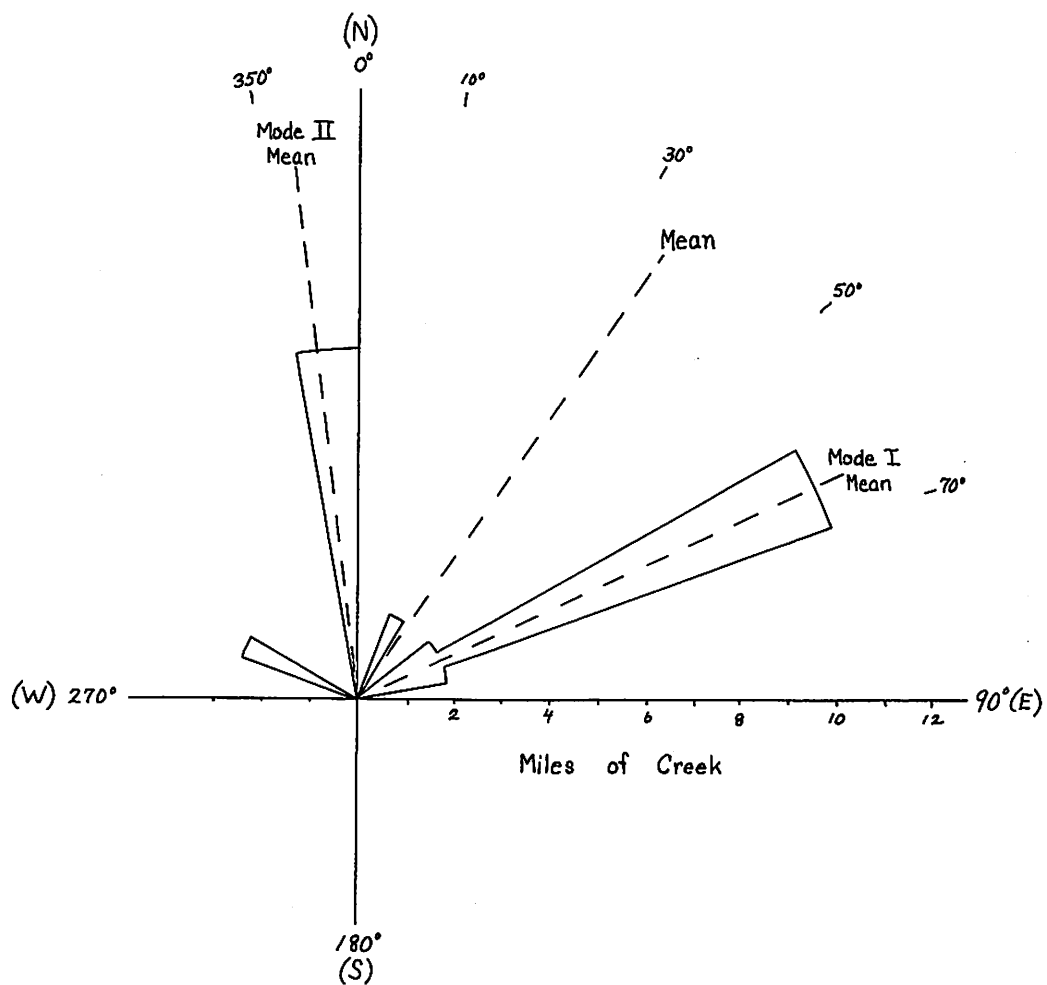


FIGURE 3.6

Valley Orientations

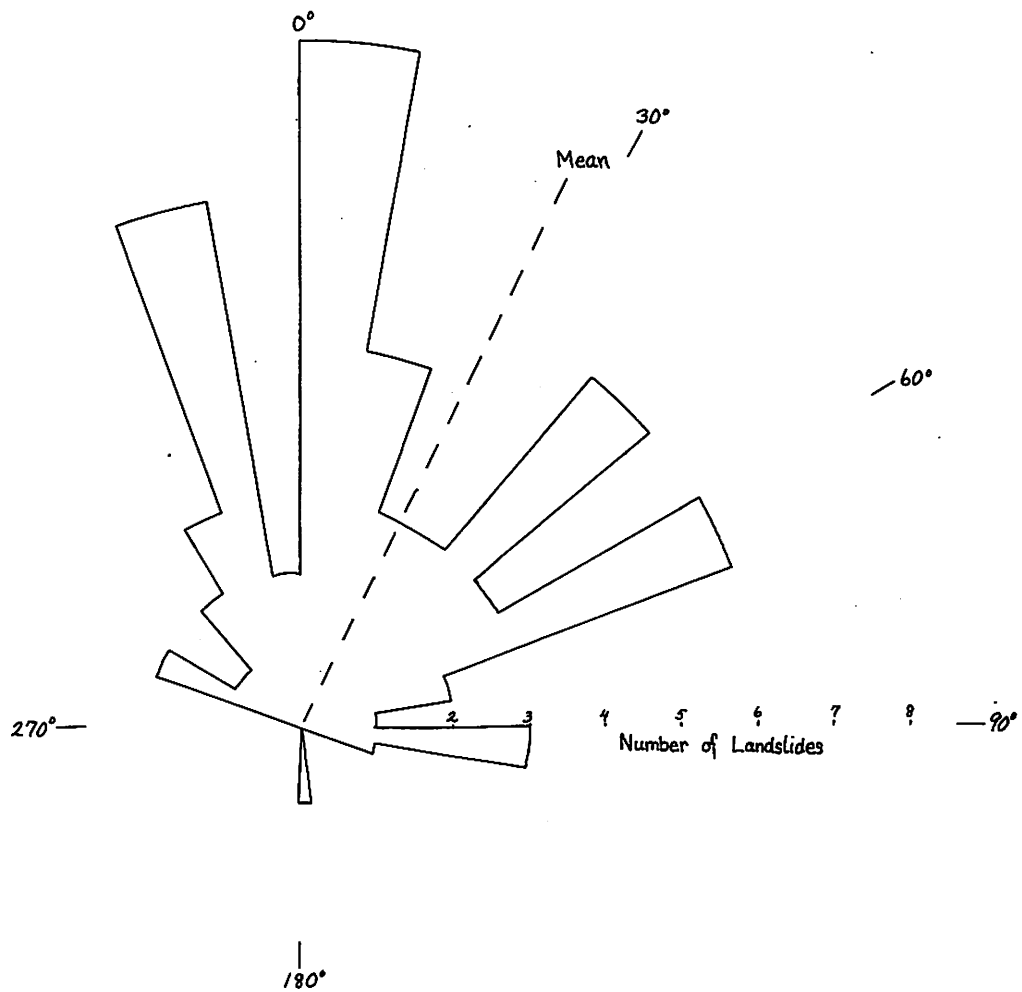


FIGURE 3.7

Landslide Orientations

TABLE 3.5

Valley Orientations		
	<u>Mode I</u> (east-facing)	<u>Mode II</u> (north-facing)
$V_m \pm s_o$	$65 \pm 2^\circ$	$353 \pm 29^\circ$
Valley Names	Brush Ck.	Snowmass Ck.
	Willow Ck.	E. Snowmass Ck.
	W. Willow Ck.	Maroon Ck.
	Owl Ck.	
	W. Snowmass Ck.	
	W. Maroon Ck.	

TABLE 3.6

Landslide Orientations		
	<u>Mode I</u> (east-facing)	<u>Mode II</u> (north-facing)
$V_m \pm s_o$	$60 \pm 13^\circ$	$353 \pm 12^\circ$
Number of Slides	28	34
Valley Names	Snowmass (3/3)*	Willow (5/5)
	E. Snowmass (13/16)	Brush (11/17)
	Maroon (2/3)	W. Willow (3/3)
	lower Willow (1/1)	Owl (4/6)
		W. Snowmass (6/6)
		W. Maroon (2/2)

(\*Ratio of number of landslides in listed mode to total landslides in that valley)

A comparison of Tables 3.5 and 3.6 indicates that the vector mean of Mode II of the distribution of valley orientation coincides with that of Mode II of landslide orientation, and that the vector means of Mode I (valleys) and Mode II (landslides) are also similar. An analysis of the valleys involved in each mode shows that, for the most part, Mode I landslides occur in Mode II valleys and Mode II slides in Mode I valleys. This supports the expectation that landslides tend to occur in a direction perpendicular to the valley orientation, i.e., on the valley sides. Landslides occurring in a valley of the same mode of orientation are not exceptions; rather they can be explained as occurring at valley heads or at opposite extreme values of the mode, or as being influenced strongly by local topography.

It is thus apparent that the orientation of valleys in this area controls the orientation of landslides by establishing the orientation of slopes. The interesting factor resulting from this comparison is that even though valleys provide two slope orientations, landslides occur preferentially on the north- and east-facing slopes and not on south- or west-facing slopes. No landslides are found in the study area in the southwest ( $180^{\circ}$  to  $270^{\circ}$ ) quadrant, and only four are found in the two quadrants from  $134^{\circ}$  to  $315^{\circ}$ . This phenomenon is in part due to the structure, described in the preceding section of this chapter, and in part due to soil moisture conditions.

At elevations of 7,600 to 13,336 feet at  $39^{\circ}$  N latitude, the significant difference in solar radiation received by northeast- and southwest-facing slopes produces a striking difference in moisture-retaining capacities of the respective slopes. Average annual precipitation in the Aspen area is only 37.5 inches, more than one-third of which is snow. Sublimation and wind contribute to the loss of snow from southwest-facing slopes, exposing the soil to water loss from evaporation two or three months before northeast-facing slopes are snow free. The presence of moisture in the soil and its role in overloading, lubricating, or activating clay constituents of the soil mass appears to be a key factor in causing Colorado landslides (Varnes, 1949; Irvine, 1963; Wahlstrom and Nichols, 1969; Barrett and Cochran, 1971). The contribution of water is favored by the larger and longer-lasting snowpack and the protection from evaporation offered by northeast-facing slopes.

Field observations show typical mass movements on southwest-facing slopes to be not landslides, but rockfalls and mudflows. On these dry slopes, low rates of weathering and soil production appear to barely keep pace with



erosion rates, thus preventing the accumulation of a thick debris mantle. Therefore, slopes that have not developed or retained such a debris mantle cannot produce the types of landslides dealt with in this study which require the displacement of unconsolidated material.

The orientation of each landslide is plotted with its starting zone slope angle in Figure 3.8. This distribution resembles the "dipping girdle" fabric pattern of river gravels and sands (Pettijohn, 1957, p. 77). An axis of soil moisture (the inverse of solar radiation) in the present example could be considered analogous to stream current, the axis of the fabric pattern. The lack of points in south-facing sectors indicates the control exerted by orientation, i.e., moisture and solar radiation, while the girdle pattern of points indicates the control exerted by slope angles. The least ( $16^{\circ}$ ) starting zone slope angles are found on north- and northeast-facing slopes. Minimum starting zone slope angles gradually increase to the east and west as slopes receive more radiation and retain less moisture.

Another effect of topography is to limit the dimensions of landslides by providing a finite length slope. This was not a limiting factor in this study. Only 52% of the total number of landslides reached the valley floor (or a major break in slope), with only 39% of the slide covering the entire length of the slope (Table 3.7). Seventy-two percent of slides shorter than 2302 feet did not reach the bottom of the slope. None of the slides reaching the valley floor extend beyond the low point or up the opposite side, and none traveled significantly in a downvalley direction. This suggests that slides reaching the valley bottom characteristically do not have more energy than was required to move down the valley-side slope. If the valleys were lower, but with less slope approaching the valley floors, the slides might not have had sufficient energy to move the full length of the slope. The lack of excess energy of the landslides studied also suggests that movement was more likely to have been gradual than catastrophic.

#### Summary

Landslides in this study area have occurred on every bedrock type that has sufficient exposure for slides to be identified on aerial photographs. Landslide lengths fit a lognormal distribution and are not controlled by the slope length. Most of the landslide widths and the shape ratios follow a lognormal distribution. Mean length is 2680 feet; mean width is 892 feet;

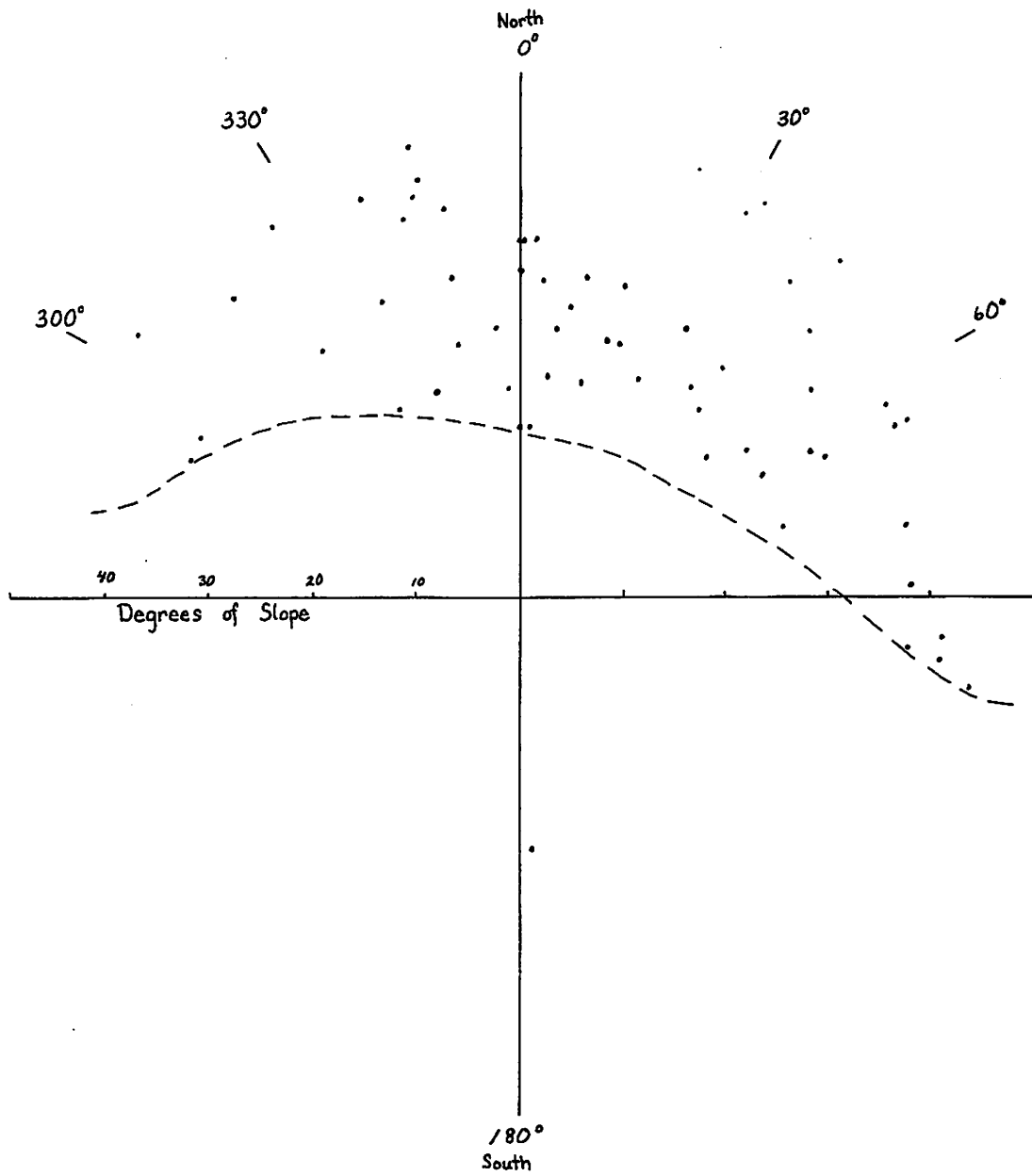


FIGURE 3.8

Landslide Orientations and Slopes

TABLE 3.7

Relation of Landslide Length to Slope Length

Group I Slide starts at top of hillside, extends to bottom.

Group II Slide neither starts at top nor extends to bottom.

Group III Slide starts at top but does not reach bottom.

Group IV Slide starts midslope and extends to bottom.

<u>Length Class</u>	<u>Number of Slides</u>	<u>Group I</u>	<u>Group II</u>	<u>Group III</u>	<u>Group IV</u>
All slides	62	.39	.21	.27	.13
Less than 2302 feet	25	.20	.28	.44	.08
2302 feet to 3880 feet	28	.43	.18	.18	.21
Greater than 3880 feet	9	.78	.11	.11	0

The Kolmogorov-Smirnov test indicates at the 1% level of significance that slides of 2,302-3,880 feet length have the same group distribution as all slides examined.

and mean shape ratio is 3.69. Different shape ratios are not correlated with different lithologies.

The effective dip, not related to the type of bedrock, appears to play an important role in establishing topography favorable to landsliding and in serving as a sliding surface for planar slides. The distinction between planar and rotational landslides is difficult to make and is not associated with a lithologic control.

Both rotational and planar landslides occur preferentially on north- and east-facing slopes, with orientations ranging from 292° to 177°. The lack of southwest-facing slides seems to be due to the lack of favorable structural conditions and soil moisture regimes on those slopes. The Wildcat Slide (#48) is proof that a south-facing slope with favorable conditions can slide.

The minimum starting zone slope angle is found to be 16°. Relationships between landslide locations, ridge crests, and valley floors suggest that the features studied were not high energy (catastrophic) slides.

## CHAPTER IV

### AGE, ENVIRONMENT OF ORIGIN, STABILITY

#### Age

Landslides in this study area appear to be old and inactive. They are typically vegetated and eroded, with boundaries and surface features lacking freshness. An estimation of the ages and environments of origin of these features is important to the understanding of landslide causation in this area and to any land use of the affected areas. Given the conditions associated with landslide initiation, it should be possible to estimate the probability of landslide reactivation.

#### *Minimum Age*

Minimum ages of these features is estimated from the vegetation. Development of a pine or aspen forest does not require a stable substrate, but the establishment of a mature spruce forest continuous with that of adjacent slopes would require a minimum of 300 to 500 years of stability (P. V. Krebs, J. W. Marr, personal communications). The minimum age of unforested landslides is unknown. Thirty of the 62 landslides studied (including five in a burned area) support spruce forests and are considered to be at least 300 years old. Thirteen landslides support aspen and pine forests and 19 have no tree cover. No minimum age of stability is given to the latter 32 features.

#### *Maximum Age*

Maximum ages of landslides in this area are estimated from stratigraphic relationships of landslide deposits to glacial deposits. In each valley, landslides override glacial deposits that are considered to be Pinedale (11,000 years) or late Pinedale (9,000 years) in age (B. Bryant, personal communication). The following discussion is intended to present the available information concerning maximum ages as well as a few thoughts on possible conditions of origin.

The landslides in the Snowmass Creek valley appear to have involved Pinedale morainal material, failing on bedrock in five of six cases. These

landslides must have taken place within the last 9,000 to 11,000 years.

Very little morainal material is found in the valleys of either East Snowmass or Willow creeks. Characteristic U-shaped valley sides of Willow Creek suggest that landsliding there may have occurred in adjustment to slopes oversteepened by glacial activity. So much of the east-facing side of East Snowmass Creek (Figure 4.1; Plate I) has moved, however, that it is difficult to estimate the angle of the lower slopes before movement occurred, and thus the relation of East Snowmass Creek landslides to glacial events is largely unknown.

Many of the landslides in the Brush Creek valley occur in cirque-like areas but it is not known whether glacially initiated cirques were subsequently altered by landsliding, or if ancient landslide depressions harbored ice in glacial times. However initiated, Brush Creek landslides override moraines which vary in age from Bull Lake to Pinedale, from 42,000 years B.P. to 9,000 years B.P. The large landslide (#10 on Plate I) on Sam's Knob in this drainage is overridden by Bull Lake deposits, thus opening the possibility of a pre-late Pleistocene origin. Slide #48 is not located near any glacial evidence: its age is unknown. Although its form is well preserved, the dry climate of its south-facing environment might favor preservation of a form of Wisconsin age, such as those reported by Smith (1936) in a similar environment in New Mexico. The large size of this feature, its small starting zone, and unique exposure suggest that it is a product of a time when environmental conditions were much different, perhaps more moist.

Three of the five landslides in the Maroon Creek drainage postdate morainal material and thus appear to be more recent than the last glacial time. Hillslopes tend to be quite steep in this valley, typically 35 to 40°, but downslope bedrock vectors range only to 19.6°. Maroon Creek landslides have been eroded and vegetated (three by spruce, one by aspen) and, except for drainage anomalies, appear continuous to adjacent slopes. Ages of landslides in this drainage then, may vary from more than 11,000 years to the present.

#### Environment of Origin

It appears that most of the landslides in this study area are between 11,000 and 300 years old. Several further arguments favor the earlier part of this range as the period of greatest landslide activity. Since large-



PLATE II

Southward view into East Snowmass Creek, showing landslide tongues. So much of the east-facing (right) side of the valley has slid that it is difficult to determine whether valley sides were glacially oversteepened.

scale landsliding in this area is not now an active process, it probably requires different environmental conditions than those of the present (i.e., the last 300 years).

Landsliding should be favored by an absence of trees and by higher moisture levels, both in the ground and in eroding streams. Those conditions are generally associated with postglacial or periglacial environments when glacial oversteepening of slopes and loading of slopes with morainal material could also be expected to contribute to the potential for movement. Given the angle of internal friction  $\phi$  of slope material, and bulk unit weights of soil  $y$ , and water  $y_w$ , the stable slope angle  $\theta$  can be estimated from

$$\tan \theta = \left( \frac{y - y_w}{y} \right) \tan \phi$$

(Carson and Kirkby, 1972, p. 156). If the angle of internal friction of natural soil materials varies from  $34^\circ$  to  $46^\circ$  (Bailey, 1971, p. 82), minimum stable slope angles will vary from  $18.6^\circ$  to  $27^\circ$ . Allowing for irregularities in materials, a conservative definition of an oversteepened slope in this area is one of more than  $30^\circ$ . Except on cliffs with no soil mantle,  $30^\circ$  slopes are rare in the present topography, possibly due to the work of landslides in reducing oversteepened slopes left by glaciation. A contributing factor to landslide initiation could have been changes in stress distributions (reduction of normal stress) brought about by the removal of glacial ice, analogous to unloading structures observed in bedrock (Ollier, 1969, pp. 5-9).

Other research in the Rocky Mountain region favors the association of landsliding with the period at the end of or immediately after a glacial epoch. Watson and Wright (1963) suggest that Pleistocene treeline may have been as low as 8,500 feet in northwestern New Mexico. Flint and Denny (1958) discuss the relationship of landslides to meltwater percolation in non-glaciated sites of a Utah glaciated region; Smith (1936) relates landslide initiation to the intense frost action of periglacial environments in New Mexico; and Yeend (1969) correlates the development of the slumps of Grand Mesa, Colorado, with ice-sapping and the subsequent removal of ice support at the mesa sides following glaciation.

Further refinement of landslide age estimates would probably require correlation of landslide deposits with pollen records or more precise methods such as radiocarbon dating of organic matter found in landslide deposits.

Pollen samples could be obtained from Maroon Lake, (at 9,580 feet), Willow Lake (11,795 feet), the pond near the head of landslide #48 (8,720 feet), bogs on Sam's Knob at 9,300 feet, or the bog known as Ziegler Pond (8,920 feet) just north of the #10 landslide. Results of such an investigation could provide a better minimum age for slides #10 and #48 as well as a clearer picture of climatic trends since glacial time.

## Stability

### *Natural Stability*

At present, all except two (#8 and #25 on Plate I) of the landslides are considered to be stable. Evidence for this conclusion includes: no discontinuity in vegetation, no tension cracks, and no tilted trees. In the 100 years of recorded history of the area, there have been no new natural slides and no cases of natural slide reactivation (with the possible exception of #25 in Plate III, which is a natural slide of unknown history. Although a landslide is presumably an adjustment toward increasing stability, old landslides are reported (Bailey, 1971, pp. 29, 114) to be "barely" or "conditionally" stable in the natural state. There are numerous examples of natural reactivation of old landslides due to stream undercutting, the impoundment of a lake by a new landslide, earthquakes, or the accumulation of stresses, such as from slumping at the head region (Bailey, 1971, pp. 41, 66, 92).

### *Stability Affected by Man*

New landslides in the present study area have been man-induced. The largest, slide #1 (Plates I, IV, V), occurred in the summer of 1956, two weeks following the installation of a new, unlined irrigation ditch in the Snowmass Creek valley. The new ditch ran across the top of a lateral moraine on a 26° slope. Without previous indication, failure occurred during the night along a sliding surface within the moraine. Similarly constructed ditches in the area have crossed moraines and other deposits of questionable stability for ten to thirty years without incident.

The construction of Snowmass Resort in the Brush Creek valley began in 1967. The West Fork of Brush Creek was diverted and West Village, a cluster of condominiums, lodges, shops, swimming pools, and parking lots, was built on and around the original stream course. Trees were cleared for the Snowmass





PLATE III

Upper section of slide #25 (circled). Note the tension cracks. This is the only slide in the area known to be moving without man's influence.

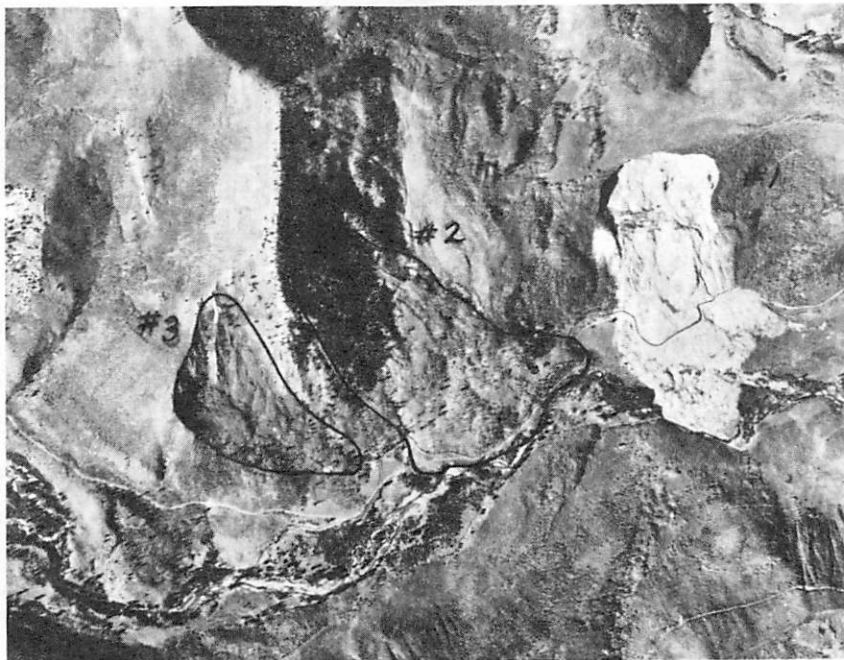


PLATE IV

Landslides #1, #2, and #3, from  
U.S.F.S. air photo taken in 1956.



PLATE V

Landslide #1, September 1974. Slope failure occurred  
along the line of a two-week-old ditch, June 1956.

Ski Area and ski lift towers installed. Construction of West Village and expansion of the ski area have been taking place continually since that time. Even though West Village has been built on an ancient landslide (#10), no stability problems have yet been reported. Local Soil Conservation Service officers, Forest Service officials and the Pitkin County Building Inspector's Office acknowledge, however, that the stability of West Village has been favored by a series of years with relatively light spring runoff and that a wet year could produce movement for the first time (personal communications). Out of awareness of the landslide hazard, West Village subdivisions have been given strict building codes which include special foundation restrictions. One of the subdivisions is located on south-facing slopes of Mancos Shale, while the others are located on north-facing slopes of Bull Lake and Pinedale deposits. According to the Pitkin County Building Inspector, only one case of downslope building movement has been reported in the entire Aspen area in recent years: the 10-inch movement of a new group of condominiums located on a 20° sloping morainal deposit at the base of Aspen Mountain. Facing slightly northeast, the slope that failed lay within the critical envelope of Figure 3.8.

In the course of ski area development around Aspen, there has been remarkable little landsliding. An annual check of the power lines serving the Snowmass Ski Area has shown no movement, and there have been only two instances of lift tower movements. One of these, at Snowmass, was due to a slope washout caused by the rupture of a clogged irrigation ditch. The other, just west of the study area in the Buttermilk-Tiehack Ski Area, was an eight-foot downslope movement of two lift towers in the spring of 1973. The towers had been in place since 1968. Movement was attributed to high pore water pressure created by the collection of snowmelt runoff in an old, "dry" watercourse upslope from the towers. Bedrock for both incidents was Mancos Shale. Also on Mancos Shale is a slumping road switchback in the Snowmass Ski Area. Initial movement of this "Slump Corner" coincided with an attempt to build a scenic pond on glacial deposits overlying Mancos Shale, one-quarter of a mile upslope. The pond has not held water, but there has been no indication of where the water goes.

The most significant current movement problem in the area occurs where a culverted irrigation ditch and a smaller pipe cross a steep cut in old

landslide (#8) debris to bring irrigation and drinking water to West Village from East Snowmass Creek. During the spring runoff of 1974, soil movement pressure caused the culvert to rotate outward and rupture, producing saturation and flowage of the entire hillside. Replacement of the culvert required an extensive cut into the already steep ( $36^{\circ}$ ) slope, thus leaving the site more unstable than before. Although this is the only site of a serious landslide problem located on a former slide, four old landslides have been impinged upon by man's activities in the area.

Cases of man-caused reactivation of old landslides are reported in the literature to be due to excavation at the toe and loading upslope (Wahlstrom and Nichols, 1969) or to the addition of water (Varnes, 1949). Old landslides generally stabilize with the draining of high pore pressures, the reduction of the slope angle, and/or the lowering of the disturbing moment (or conversely, increasing the stabilizing moment). Reactivation could then be anticipated as a consequence of increasing at least one of these parameters.

#### *Future Stability*

New and renewed landslide movements in this study area are typically associated with the disturbance of slopes by the removal of a supporting part of the slope and by the use of local, insufficiently compacted and poorly drained landfill. The introduction of water, from springtime snowmelt or from seepages or ruptures of ditch systems, appears to be a necessary trigger of movement. Development in the area is very new; the possibility of future movement due to existing structures should not be overlooked.

Stream undercutting is the only available condition required for the natural initiation or reactivation of movement in the area, and would be expected to produce movement at predictable locations. Man, however, has the potential to induce landslides almost anywhere in the area, either as an immediate or delayed result of his activities. There are numerous small rotational landslides along roads and streams in the area that are too small to be recognized on air photographs. These are indicative of the narrow margin of natural stability, especially on old landslide deposits, morainal deposits, and soils derived from the Mancos Shale. Proposed expansion of the Snowmass Ski Area, construction of a light railway, and development of the Owl Creek Resort involve all of the three particularly unstable substrates mentioned above. No slopes of less than  $16^{\circ}$  have been known to fail in this area. Figure 3.8 indicated that while south-, east-, or west-facing slopes

of  $25^{\circ}$  are probably free of landslide hazard, a north- or northeast-facing slope of  $16^{\circ}$  or greater would be expected to slide when disturbed.

### Conclusion

Although landslides in this study range in age from pre-Bull Lake to the currently active, most of them appear to have occurred since the Pinedale but before 300 years B.P. Ages within this period are unknown, but it is suggested that they tend toward the earlier half, when the general environment would have been very different from that of today and more favorable to movement.

The present inactivity of the features studied is seen to be a conditional state: increases in moisture, slope angle, unit weight, or decreased supportive strength could be expected to bring about reactivation. The reactivation of old landslides has been seen to occur in this area and is known in other areas as well. New slides have been initiated here in glacial deposits (slide #1) and in the Mancos Shale (Slump Corner), with both cases involving water introduced by man.

It is concluded, then, that the margin of stability in old landslides, morainal material, and Mancos Shale is very narrow, such that relatively small disturbances have the potential to initiate further movement. The chances of naturally-induced movement in the future are limited; the chances of man-induced movement are far greater. Only those slopes that lie below the critical envelope given in Figure 3.8 are considered to be free of landslide hazard.

## CHAPTER V

### CONCLUSION

#### Attainment of Goals

In Chapter I, the following objectives were outlined for this study:

- 1) to identify and characterize the form, location, and distribution of previous landslides in the area;
- 2) to determine the age of those landslides, seeking clues from tectonic, glacial, or climatic events of those times to explain the causes of movement;
- 3) to identify any existing relationships between past and present instabilities;
- 4) to synthesize this information into a descriptive model of land-sliding in the area, thus establishing a basis for the prediction of future landslide events.

Although these objectives have been reached within the limitations of this study, further study is required to satisfactorily treat the second and fourth. The age of landslides in the area and conditions associated with their initiation are still unknown. A more extensive dating effort will be necessary to deal with this objective. Measurement of soil movement, on old landslide deposits as well as on "stable" substrates, would help to assess present stability and would thereby facilitate a more site-specific model than is possible at this time.

#### Landslide Distribution and Form

Landforms of 62 landslides were identified and mapped in the study area. The distribution of these features, shown in Plate I, is significantly clustered. Clustering represents spatial controls of landslide distribution, i.e., that landslides are more likely to occur on some slopes than on others.

Lengths of the landslides studied are found to fit a lognormal model, as do landslide shape ratios and all but the particularly large widths. Geometric mean length is 2,860 feet. As only 54% of the landslides reached the valley bottom, and only 39% originated at ridge crests, it is concluded

that landslide length is controlled, not by slope length, but rather by the balance of forces acting on the moving mass.

Landslides occur over all of the major lithologic units represented in the study area. Fewer landslides have occurred on the Mancos Shale than would have been expected from the literature, probably due to the low slope angles and dryness of Mancos exposures in the area. Bedrock lithology is seen to have no effect on landslide shape; thus no differences in landsliding process are attributed to lithology.

#### Conditions Favoring Landsliding

Landslides are located on east-facing slopes of north-trending valleys and on north-facing slopes of east-trending valleys. Preferred orientations are correlated with bedrock having a vector of dip in the direction of the slope and with more favorable, i.e., more moist, climatic conditions on slopes receiving the least solar radiation. Minimum starting zone slopes are  $16^{\circ}$  on slopes facing north and northeast, with angles increasing to  $30^{\circ}$  and  $40^{\circ}$  as orientations become more eastward, westward, and southward.

Although the landslides range in age from pre-Bull Lake to presently active, most of them are thought to have originated sometime between the end of the Pinedale (11,000 years B.P.) and 300 years B.P. Periglacial or postglacial environments would have been conducive to landslide initiation due to increased moisture, decreased tree cover, recently oversteepened valley sides, loading of slopes with morainal material, and changes in stress resulting from the removal of an ice load.

The safety factor model, described in Chapter I, gives the following conditions for slope failure: decreased cohesion of the material, increased pore water pressure, decreased angle of internal friction, increased slope angle, and increased unit weight of the material. Recent development in the area has produced remarkably little slope movement, but landsliding has occurred in cases where water has been introduced, where slopes have been oversteepened by excavation, or where insufficiently compacted and poorly drained landfill has been used. These alterations by man are explained by the model, with each of them contributing to the reduction of the safety factor.

Moisture and slope angle are the most critical parameters in this area as well as in other areas. Historic slope failures in the area are all

associated with snowmelt runoff and initiation of landslide movement appears to require a minimum slope of  $16^{\circ}$ , with even steeper slopes being needed on east- and northwest-facing slopes. The lower starting zone slope angles of north- and northeast-facing slopes and the lack of landslides on southwest slopes are attributed to soil moisture regimes of the respective slopes.

### Prediction

The only predictable instances of naturally-initiated new landslides would involve stream undercutting at some time in the future and would require stream courses to shift laterally from their present channels. Otherwise, new and renewed landslide activity will be the work of man. Evidence in the present area and reports from other areas demonstrate a reactivation when old landslides are disturbed, either naturally or by man. Initiation of new landslides, large and small, is seen to be a real danger of man's increased use of the area. Poor drainage, allowing saturation of material when sufficient water is available, is characteristic of old landslide deposits. Morainal material, due to poor drainage, and the Mancos Shale, due to its high clay content, are also particularly susceptible to landsliding. Disturbance of these substrates should be avoided, or should be carried out as if they were already moving, especially when water is involved.

This study, of a regional nature, provides an overview of the landslide hazard and should be of practical value in land-use planning. Conditions favorable to landsliding have been delineated, with Figure 3.8 and Plate I summarizing the information in readily applicable form. Statements regarding the stability of a specific site, however, still require a detailed investigation of the soil material properties. Until further landslide research is done in nearby areas, the validity of extrapolating the conclusions of this thesis beyond the boundaries of its study area will be undetermined.



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APPENDIX

SYMBOLS USED

$c$  = unit cohesion

$N$  = number of observations

$n$  = number of points per class

$p$  = normal stress on sliding surface

$R$  = nearest neighbor statistic

$s$  = shear strength

S.E. = standard error of Expected

$u$  = pore pressure

$v$  = standard variate of normal curve

$y$  = bulk unit weight of soil

$y_w$  = unit weight of water

$z$  = depth to failure surface

$\theta$  = slope angle

$\rho$  = density

$\tau$  = shear stress

$\phi$  = angle of internal friction

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