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Characteristics and Processes of Degradation on Normal Fault Scarps in Basalt, Central Oregon and Northern California

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Characteristics and processes of degradation on normal fault scarps in basalt, central Oregon and northern California

BY

Kelsay M. Davis

Accepted in Partial Completion

of the Requirements for the Degree

Master of Science

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Master's Thesis

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Characteristics and processes of degradation on normal fault scarps in basalt, central Oregon and northern California

A Thesis Presented to the Faculty of Western Washington University

In Partial Fulfillment of the Requirements for the Degree Master's of Science

By
Kelsay Davis
July 2007
Abstract

Normal faults that break the surface create scarps. Scarps in alluvium degrade predictably so that time since formation can be inferred from scarp profile, however, scarps in jointed bedrock, such as basalt, do not degrade according to previous models. Understanding the processes involved in degradation of scarps in jointed basalt may lead to the formation of a degradation model.

I use survey data and statistical analyses from 36 scarps in central Oregon and northern California, to determine the characteristics that play a dominant role in scarp degradation in jointed basalt. These data indicate that scarp facing direction, column height, and joint spacing influence degradation while vesicularity, scarp height, and scarp location play less integral roles. These data also indicate that basalt columns larger than 0.45 m diameter do not easily topple and likely need regional ground motion to induce toppling, thereby aiding scarp degradation. I use measurements from toppled columns to calculate the quasi-static peak acceleration necessary for topple. I use the acceleration to determine the Modified Mercalli intensity, which I use as a proxy for earthquake magnitude. I compare the results to published acceleration probability maps as well as to a compilation of earthquake magnitudes since 1970 in central Oregon and northern California. I also estimate the probable maximum earthquake magnitude typical faults in the region could produce based on fault segment length. The curve infers the frequency of local earthquakes with magnitudes sufficient to topple, which I use to infer a degradation rate assuming ground motion is a primary geomorphic agent. I create a geometric simulation of scarp degradation in jointed bedrock, with varying retreat increments to represent jointing, as well as variable initial scarp angles, and talus repose...
A comprehensive census of characteristics at more scarps would provide a database to improve understanding of the primary variables involved in degradation, to validate the methodology presented using toppled columns as strong motion sensors, and further refine the geometric simulation.
Acknowledgements

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Table of Contents

Abstract iv

Acknowledgements vi

List of Tables viii

List of Figures ix

Thesis

Chapter 1: Degradation characteristics of normal fault scarps in basalt, central Oregon and northern California 1

Chapter 2: Strong ground motion as a geomorphic agent in Scarp degradation. 46

Chapter 3: Geometric simulation of scarp degradation 80

Appendix A: Scarp descriptions 93

Appendix B: Scarp degradation simulation script for Matlab 125

Works Cited 131
List of Tables

Table 1: 5% level of significance values. 27
Table 2: Rock mass strength parameters. 65
Table 3: Dimensions and toppling accelerations of columns. 66
Table 4: Relationship of Modified Mercalli Intensity value to magnitude. 67
Table 5: Average number of events per year by magnitude for central Oregon and northern California, between 117° - 122° W, 40° - 45° N. 68
Table 6: Events per year for a given area in central Oregon and northern California. 69
Table A-1: Scarp characteristics, Appendix A. 101
List of Figures

Figure 1: Scarp terminology. 28
Figure 2: Scarp degradation sequence. 29
Figure 3: Limits of principal slope angle with scarp age. 30
Figure 4: Scarp location map. 32
Figure 5: Ages of scarps. 33
Figure 6: Scarp profile comparison. 34
Figure 7: Free face height versus cross sectional area of talus. 35
Figure 8: Cross sectional area of talus and normalized free face height versus scarp height and free face height. 36
Figure 9: Location of transect on scarp segment versus area of talus and free face height. 37
Figure 10: Joint spacing versus area of talus and free face height. 38
Figure 11: Column height versus area of talus and free face height. 39
Figure 12: Vesicle density versus area of talus and free face height. 40
Figure 13: Facing direction versus area of talus and free face height. 41
Figure 14: Cross sectional area of talus and free face height at scarp transects. 42
Figure 15: Differential joint size at scarps. 43
Figure 16: Basic column shape. 44
Figure 17: Vegetation at scarps. 45
Figure 18: Spalling at scarps. 46
Figure 19: Toppled columns at Boyd Skeleton Scarp, central Oregon. 70
Figure 20: Toppled column at Porcupine Jr. scarp in northeastern California. 71
Figure 21: Location of columns at Boyd Skeleton Scarp, central Oregon. 72
Figure 22: Balancing points on columns. 73
Figure 23: Peak ground acceleration for various α's. 74
| Figure A-14: Golddigger scarp photo mosaic. | 118 |
| Figure A-15: Golddigger scarp transect profiles. | 119 |
| Figure A-15: Porcupine Jr. scarp, northern California. | 121 |
| Figure A-16: Porcupine Jr. scarp photo mosaic. | 122 |
| Figure A-17: Porcupine Jr. scarp transect profiles. | 123 |
Chapter 1: Degradation Characteristics of Normal Fault Scarps in Basalt, Central Oregon and Northern California

Introduction

Normal faults that break the surface form scarps. Over time, as these scarps degrade, the scarp profile becomes rounder and the slope decreases. The evolution of normal fault scarps that break alluvium is well studied. For example, the shape of the scarps in alluvium in the Basin and Range province of the western United States is predictable and can be correlated with age (Wallace, 1977). Scarp profiles can thereby contribute to estimations of earthquake recurrence in the region, which has a sparse historical seismic record due to lack of instrumentation (Wallace, 1977; Bucknam and Anderson, 1979; Nash, 1980).

There is a common association between extensional environments and basaltic volcanism, so that many normal faults rupture basalt (e.g. Opheim and Gudmundsson, 1989; MacDonald, 1957). Basalt is commonly jointed. Little work has been done regarding characteristics of degradation for jointed bedrock, except to note that time-dependent models for degradation of scarps in alluvium are considered inappropriate for jointed bedrock (Mayer, 1984; Stewart, 1993). Refining our understanding of bedrock scarp degradation processes may ultimately lead to the formation of a model of scarp degradation in jointed bedrock, providing another tool to estimate regional earthquake recurrence.

I surveyed normal fault scarps in basalt in northern California and central Oregon and collected measurements regarding scarp characteristics. In this chapter, I use these observations in a linear correlation analysis in an attempt to describe the primary factors controlling degradation of the scarps. In Chapter 2, I discuss the use of toppled basalt columns as a means to estimate regional ground motion, earthquake recurrence and scarp
degradation rates. In Chapter 3, I present a geometric simulation similar to Nash (1981) showing the stages of degradation and retreat of a bedrock scarp.

**Scarp degradation**

Scars are formed when normal faults break the surface. Wallace (1977) was one of the first to describe scarp characteristics from inception through erosion, also establishing nomenclature for scarps in alluvium. I adopt his terminology unless otherwise stated. At inception, the scarp reflects the fault plane at an angle between 50° to 90°. The scarp face is called the free face (Figure 1). Locally, the free face may overhang the downthrown block. Wallace (1977) attributes overhanging scarps to irregularities of extension in the upper few meters of colluvium. Wallace (1977) observed that grabens and fractures commonly form at the base of scarps, particularly in bedrock, but these cracks quickly fill with material that spalls off the free face.

Three types of erosional processes may affect a scarp: gravity-controlled, debris-controlled, and wash-controlled erosion. Gravity-controlled processes involve blocks falling off the scarp free face. Debris-controlled processes involve blocks falling and rolling down the talus slope. Wash-controlled processes involve the movement of material down-slope by wind or water. Typically, wash-controlled erosion moves smaller sized material over shallower slopes (Wallace, 1977; Selby, 1993).

As a scarp erodes it undergoes several changes, termed “slope replacement” and “slope decline” (Wallace, 1977). Slope replacement begins directly after the formation of the scarp, when gravity-controlled spalling causes erosion of the free-face. The spalled material collects at the base of the scarp, forming a debris pile or talus slope. In most materials, the
angle of repose on the talus slope is about 30-40 degrees (Selby, 1993). A wash slope or colluvial wedge forms below the talus slope at a shallower angle of 5-15 degrees and typically covers the toe of the scarp. Wallace (1977) notes that the wash slope grades into the regional slope and can be very difficult to distinguish. There is commonly a crest between the top of the free face and the upper faulted surface (Figure 1) that retreats from the original scarp-line and becomes less prominent with time (Wallace, 1977). Eventually, spalling destroys the free face, and further scarp degradation is almost entirely wash-controlled (Figure 2). This signifies the beginning of the slope decline process; the slope of the scarp gradually shallows until it is once more level with the regional slope (Wallace, 1977).

Wallace (1977) noticed a direct correlation between scarp profile and age of scarps in the Basin and Range. Degradation of scarps in alluvium within the Basin and Range follows a predictable pattern. The scarp will evolve from a free face scarp with a slope between 45° and overhanging, to a wash controlled slope between 8°-25° (Figure 2). The approximate age of the scarp can be estimated by its slope and profile. Scarps in alluvium created within the past few thousand years will have a free face and a debris slope of about 35° (Figure 3). Older scarps in alluvium, between a few thousand and 12,000 years old will have a maximum total slope angle between 20°-25°, and scarps older than 12,000 years will have slopes under 20°.

The age–slope-angle curve for scarps in bedrock is significantly different than for scarps in alluvium (Figure 3). The slope-replacement stage of scarp erosion takes much longer in bedrock, so that the free face and talus slope are present for much longer periods of time. A wash slope may never develop if there is not ample sediment to wash down the
scarp face (McCalpin, 1996). The persistence of bedrock scarps provides some of the best evidence for past earthquakes. However, because their degradation characteristics have been less well documented, determining time since formation is more problematic (Zreda and Noller, 1998). Additionally, in Wallace's (1977) degradation model there is no information about the role of joints in degradation. Other works note that columnar joints in basalt likely play a significant role in the erosional characteristics and scarp evolution (MacLeod and Sherrod, 1992; Personius, 2002).

**Columnar joints in basalt**

Joints occur in a wide variety of rock types and tectonic environments, resulting from strain accommodation by brittle fracture (Pollard and Aydin, 1988). Joint-bounded columns are created by systems of interconnected tensile fractures and occur in igneous rock masses during cooling and contraction of the igneous body (DeGraff and Aydin, 1987). These columns are generally polygonal in cross section with large height/diameter aspect ratios (DeGraff and Aydin, 1987; Suppe, 1985). The columns form perpendicular to the cooling margins of the igneous body and grow incrementally inward (DeGraff and Aydin, 1987). Bands of plumose structure across the faces of columns result from propagation of the fractures and show the orientation of incremental growth, which is driven by the concentration of stress along column edges (DeGraff and Aydin, 1987). The stress induces and guides new segment growth, but as columns grow inward, joint patterns change slightly with changing maximum tensile stress, resulting in small tetragonal columns at the surface of flows, and larger, hexagonal columns at the interior of flows (Aydin and DeGraff, 1988).
Columnar joints significantly reduce the strength of a rock mass, resulting in tensile and cohesive strength values that are one or two orders of magnitude lower than corresponding values for intact rock (Schultz, 1995; Selby, 1993). The columns are strongest along the long axis (height), with strength values lowest along the diameter, reflecting the deformability of fracture faces normal to the column growth (Schultz, 1995). Columnar joints also allow for the flow of water through the rock, and the pore water pressure that results may play a role in reducing frictional resistance to failure (Selby, 1993).

Vesicles within the basalt act as additional minor partings; however, Selby (1993) suggests that major partings within a rock mass are more important to rock mass strength and frictional resistance to failure than minor partings or hairline fractures. Vesicles probably reduce strength by a negligible amount compared to the influence of columnar joints.

Geology of the Study Areas

To document characteristics of normal fault scarps in basalt, I selected scarps near Newberry Crater in eastern Oregon, at Lava Beds National Monument in northern California, and at Devil’s Garden, Modoc Plateau in northeastern California (Figure 4). The volcanic terrain of these area provides an excellent opportunity to research the processes of scarp degradation in jointed basalt. Quaternary normal faults break the surface through basalt in each of these areas (Figure 5) (Personius, 2002). In addition, each region is sparsely vegetated, providing good exposure of the scarps and easy access.
Newberry Crater National Monument

Newberry Crater National Monument is located in central Oregon, approximately 10 km south of Bend. Newberry Volcano is a large shield volcano, about 64 km by 40 km in size, with numerous basaltic lava flows covering the flanks. The volcano has been active for approximately 500,000 years, with numerous eruptive episodes in the Holocene and Pleistocene (Jensen and Chitwood, 2000).

Extensional faults commonly break the surface of the Newberry flows. These include the Sisters fault zone to the northwest of Bend, the southwest Newberry fault zone, and the Paulina Marsh-Antelope Mountain faults south of Newberry Crater (Personius, 2002). Geomorphic expression of faulting in the Sisters fault zone consists of 2-30 m high scarps in Miocene to Pleistocene volcanic rocks. Many of the scarps are through jointed basaltic bedrock. Dates of faulting are constrained by Quaternary deposits, such as by scarps in glacial outwash north of Tumalo, OR (Personius, 2002; Sherrod and Smith, 2000). Faults in the southwest Newberry fault zone form small scarps through Plio-Pleistocene lava flows (MacLeod and Sherrod, 1992; Personius, 2002). Normal faults in Paulina Marsh-Antelope Mountain are expressed as small scarps, less than 2 m high, through Holocene alluvial deposits and Miocene-Pliocene volcanic bedrock (MacLeod and Sherrod, 1992). Ages of the faults are constrained by offset on dated Quaternary deposits and possibly Mazama ash (MacLeod and Sherrod, 1992; Pezzopane and Weldon, 1993; Personius, 2002).

I selected four scarps from this region. Scarps were selected based on height and accessibility. Barr Road scarp has a maximum height of 8 m, is northwest trending and down to the northeast (Figure A-1). It is highly degraded and is the only scarp in the study that offsets non-jointed basalt. Barr Road scarp is part of the Sisters fault zone. Maximum age is
constrained by offset Miocene basalt. Offset Pleistocene outwash from trench studies on nearby scarps suggests a post-Pleistocene age for the scarp (Hemphill-Haley, 2001). Minimum age is unconstrained, but a nearby faulted deposit is overlain by a particular soil horizon (Bt horizon) that takes up to 100 ka to develop (Hemphill-Haley, 2001). There is a remarkable difference in the degradation between this scarp and scarps in jointed basalt. Barr Road scarp has shallower, soil covered talus slopes, much smaller free face heights, and pronounced wash slopes as opposed to steeper talus slopes, higher free faces and little or no wash slope.

Boyd/Skeleton Cave scarp has a maximum height of 11 m, is northwest trending, and down to the southwest (Figure A-4). The southern end of the scarp is a monocline and there appears to be a slight change in strike where it transitions from a monocline to a free face. The basalt is jointed with 10-20 vol. % vesicles, and little to no discoloration or lichen cover. It is part of the Sisters fault zone which has late- to mid-Quaternary activity. This scarp was trench for a geotechnical study by Hawkins et al. (1988) who dated the basalt at 2.7 Ma. The minimum age is constrained by undisturbed Tumalo tuff dated at 0.29 ± 0.12 Ma (Hawkins et al., 1988).

Big Hole scarp has a maximum height of 10 m, trends to the northeast and is down to the southeast (Figure A-7). It faults jointed basalt and at some locations is expressed as a composite scarp, with a 2-3 m wide fissure between two steep faces displaying topple degradation. There are several different joint sets within the basalt. This scarp is part of the southwest Newberry fault zone which is inferred to comprise Holocene faults in Plio-Pleistocene volcanics (MacLeod and Sherrod, 1992). Weldon and others (2002) also have
mapped lineaments on Quaternary sediments. This area is much more vegetated than other
carp sites.

Oatman Flat scarp has a maximum height of 10 m, is northwest trending, down to the
southwest and faults jointed basalt (Figure A-10). It appears to be part of a series of scarps
that trend to the northwest and is probably part of the Paulina Marsh-Antelope Mountain
fault zones. The faults in the region offset Miocene to Pliocene volcanic rocks, but
Quaternary faulting is inferred from the prominence of the regional escarpments and the
presence of Quaternary sediments in the grabens. Also, in some locations, Quaternary
sediment is mapped as juxtaposed against the Miocene-Pliocene volcanics (MacLeod and
Sherrod, 1992).

**Lava Beds National Monument**

Lava Beds National Monument is located about 50 km south of Klamath Falls, Oregon on the Modoc Plateau, between the Basin and Range province to the east and the
Cascade volcanic arc to the west. The geology of the region is much like that of Newberry
Crater, with recent rhyolitic to basaltic eruptive episodes and extensional faulting through
volcanic rock (Donnelly-Nolan and Champion, 1987). Most of the lava flows in the
monument erupted from cinder cones that are flank vents of Medicine Lake Volcano to the
west. The Mammoth Crater basalt is the most volumetric and widespread of the flows and
erupted during the late Pleistocene (Donnelly-Nolan and Champion, 1987). East-west
extension has created numerous scarps, ranging in size from a few meters to several tens of
meters in height. Faults in the Gillem-Big Crack fault system have late Pleistocene east-west
extension constrained by the Mammoth Crater basalt flows, as well as outwash from glacial
events from 13 ka (Donnelly-Nolan and Champion, 1987). There are also numerous smaller
scars in the region (Wu and Crider, 2000).

I selected one scarp from this region, part of the Gillem-Big Crack fault system.
Golddigger scarp is 10-15m high, is northwest trending, down to the northeast and breaks
jointed basalt (Figure A-13). An access road takes advantage of a ramp in the scarp. North
of the access road, the fault is expressed as a series of west-stepping, echelon monoclines
with center fissures 3-4 m across. The monoclines are approximately 3 m high. The fissures
appear to be related to faulting rather than to flexural extension from folding of the
monocline because of their en echelon pattern and because of the large aperture (Wu and
Crider, 2000). The fault segment ends about 500 m north of the access road and a new
segment begins again approximately 1 km north of the road.

On the south side of access road, the scarp begins as a monocline and sharply changes
to a free face with a talus slope. There are three fault segments on this part of the scarp,
separated by broad ramps. The road goes up one north facing ramp. There is a small south
facing ramp and another north facing ramp. Golddigger scarp is part of the Gillem-Big
Crack fault system and the maximum age is constrained by the Pleistocene Mammoth Crater
flows. The minimum age is constrained by the basalt that flows over the scarp north of the
access road and is likely part of the Pleistocene eruptions in the Lava Beds region (Donnelly-
Nolan and Champion, 1987).

**East Modoc Plateau/Devil's Garden Region**

The Devil's Garden region of the Modoc Plateau is located about 50-60 km east of
Lava Beds National Monument on a plateau west of Goose Lake, California. The geology
consists of Tertiary volcanic and sedimentary rocks that are overlain by Quaternary lacustrine and alluvial deposits (Hedel, 1984). The Tertiary volcanics, consisting of rhyolites, tuffs, and basalt flows, overlie the sedimentary rocks (Hedel, 1984). Extension in the area has created north-south striking normal fault scarps through the volcanics (Hedel, 1984). The age of initiation of episodic faulting is uncertain but Holocene slip is constrained by offset glacial-lake deposits (Hedel, 1984). The Devil’s Garden area has numerous normal fault scarps through jointed basalt, such as several 2-10 m high scarps near the larger Porcupine Valley scarp (White and Crider, 2006).

I selected one scarp from this region based on accessibility. Porcupine Jr. scarp is 7-9 m high, trends to the north and is down to the west (Figure A-16). It is smaller than the main Porcupine Ridge scarp which bounds Porcupine Valley. It runs parallel to the main scarp and is also parallel to a much smaller, east facing scarp. Porcupine Jr. scarp and the smaller scarp bound a small graben, approximately 250-300 m across, striking north. In some spots, the scarp is a free face from the valley floor, while at other portions of the scarp there is a 1 m high bench before the free face. The entire scarp is covered with lichen, both the upper surface and the free face. Spalling of the rock is apparent in some areas by lack of lichen and by presence of rock flakes below the spalled area. The age of faulting has not been well constrained, and in the absence of evidence, I can only presume the latest motion post-dates the Tertiary volcanics and is part of regional Quaternary faulting.
Methodology for transect profiles and joint characteristics

Field Procedures

Prior to field work, approximately 10 prospective fault regions were selected from the USGS Quaternary fault database (Personius, 2002). Topographic maps and air photos were used to locate accessible scarps. I surveyed 35 profiles on 6 scarps between 7m and 18m high (Appendix A). Transect locations were selected for representative profiles showing different degrees of degradation.

I used a Leica TC605 total survey station (theodolite plus electronic distance meter) to collect detailed topographic profiles across the scarps. Precision of the instrument is 2.5-4 cm. I took photographs of the scarps along the surveyed sections to compile diagrams illustrating the morphology of each scarp. I measured scarp orientation in the field and on aerial photographs. I measured joint spacing and column height, and estimated vesicle density of all columns in the footwall along a 2.5 m line perpendicular to each transect. In addition, I measured the dimensions of large toppled columns to use in an analysis of toppled columns as strong motion sensors (Chapter 2).

Data Manipulation

I entered joint and scarp characteristic data gathered in the field into a spreadsheet. Averaged joint spacing data provide one variable for comparison between scarps and between transects of a single scarp. Additional characteristics were determined from field measurements, such as vesicle density, or from transect profiles, such as talus slope angle and cross-sectional area, and free face height (Table A-1). The free face height for each scarp was measured from the plotted profiles (Figure A-1). Slope angles were calculated by
taking the arctangent of the height of the talus slope divided by the width of the talus slope.
The length is measured from the toe to the horizontal coordinate of the top of the talus, while
the height is the top of the talus where it turns into a free face or a crest. The cross-sectional
area of the talus slope is used as a proxy for the 3-dimensional talus volume and is calculated
by treating the talus as a right triangle. The transects and measured characteristics were used
in statistical and visual comparisons to constrain the factors that play a significant role in
degradation. Additionally, the transects were compared to a geometric scarp degradation
model showing the stages of degradation of normal fault scarps in basalt (Chapter 3).

Scarp Profiles

The scarp profile data were projected onto a line perpendicular to the scarp using
easting and northing coordinates (Appendix A). To permit direct comparison among profiles
(Figure 6a-c), profiles were normalized vertically by scarp height and horizontally by the toe-
to-crest length. Any basal concavity in front of the toe was subtracted from each profile in
order to show only the scarp. All profiles were converted to show left facing scarps in order
to facilitate comparison between individual scarps. Photographs, surveyed profiles and
descriptions of each transect can be found in Appendix A.

Profile Comparisons

Each scarp has profiles that have one of three basic shapes: primarily a talus slope, a
small free face with a talus slope, and primarily a free face (Figures 6a-c). The profiles that
are primarily talus slopes represent the debris and wash controlled stages of Wallace's (1977)
degradation sequence (Figure 2). The profiles that have both a free face and a talus slope
represent the gravity and debris controlled stage of the sequence, while the profiles that are mostly a free face are still within the fault controlled/gravity controlled stage of the sequence. At these scarps, the primary form of degradation is from blocks loosening and falling from the free face, while in the debris and wash controlled stages, the blocks are loosened and roll down the talus slope or are transported down-slope by rain, wind, or surface water (Wallace, 1977).

The profiles that are predominantly talus slopes (Figure 6a) range in slope from 4° to 15° (Appendix A), significantly smaller than the angles of repose between 30° - 40° suggested by Selby (1993). The observed values may include wash slopes that are indistinguishable in profile form. The joint spacing at the profiles that are predominantly talus varies from no jointing up to 1 m. The column height is significantly smaller than at those profiles with free faces, ranging from no columns to 0.78 m. Total scarp height ranges from 7 to 18 m. The primary commonality is that these profiles have no free face. The profiles with no free face are contiguous on the scarp segments with profiles that show high free faces, and thus can not represent a flow margin. Boyd/Skeleton scarp is the only location lacking recurring transects with no free face. Most of the transect profiles are convex in shape; PJ1, BR2 and GD3 are concave. The talus slopes at the Barr Road profiles are all soil mantled, with a few basalt boulders, as is the first profile at Big Hole scarp. The talus slopes at all the other scarps consist predominantly of basalt boulders and blocks, commonly column shaped.

The profiles that have a talus slope and a free face (Figure 6b) range in height from 6.8 m to 14.5 m, with talus slopes similar in angle to the profiles that are predominantly talus. Joint spacing ranges from no joints to 0.93 m, while column height ranges from no columns
to 1.13 m. Many of the profiles have a toppled column near the crest which appears as a sharp point on the normalized graphs in Figure 6b. Most of the profiles at Boyd/Skeleton scarp, shown in black, have a bench between the talus and the free face.

The profiles that are predominantly free face range in height from 8 m to 12 m (Figure 6c). These profiles tend to have larger columns, with joint spacing ranging from 0.46 m to 1.59 m and column height ranging from 0.55 m to 3.93 m. This observation leads to the hypothesis that transects with larger columns tend to have less talus slope, a higher free face and therefore are less degraded. The relationship is explored in the next section.

According to Wallace’s (1977) degradation model for the Basin and Range, the gravity controlled stage (with a persistent free face) in fractured bedrock persists up to 10,000 years and the scarps may persist in the debris-controlled stage for more than a million years (Figure 3). The scarps in this study are all early to mid-Quaternary in age (Figure 5) (Personius, 2002; Pezzopane and Weldon, 1993; Hawkins et al., 1988). There are transects that show different stages of degradation along the same scarp, within a few tens of meters of each other (e.g. PJ1, PJ2, Appendix A). Individual transects cannot represent different scarp ages. Thus, Wallace’s (1977) time-dependent profile model cannot be used to estimate the time since formation of these scarps in jointed basalt. Other factors must play a role in the degree of degradation a scarp transect experiences.

**Statistical Analyses of Scarp Degradation**

**Use of correlation statistics in geoscience**

Geostatistics, the application and use of statistics in geology, has been used primarily in petroleum and mineral exploration to assess the probability of resource discovery for a
given set of samples (Davis, 2002). More recently, however, geostatistics has been applied to model complex geological structures, to test the fit of data to a pre-established curve, and to analyze the recurrence of geological events, among many applications (eg. Howarth and McArthur, 1997; Foster et al., 2000; Strebelle, 1990). Geostatistics can also be used to verify a relationship between variables observed in the field and can be used to test significant correlations between observations and processes.

**Methods**

The strength of a linear relationship between a set of values from a sample of data can be determined by using a linear correlation analysis. The analysis uses the mean values of the data sets to calculate the correlation coefficient, also called the Personian correlation coefficient (Davis 2002; Jaisingh, 2006):

$$r = \frac{n(\sum xy) - (\sum x)(\sum y)}{\sqrt{n(\sum x^2) - (\sum x)^2} \times n(\sum y^2) - (\sum y)^2}$$

(1)

where $x$ and $y$ are the dataset values and $n$ is the number of data pairs.

The Personian coefficient, $r$, is a number between 1 and -1; the closer $r$ is to positive or negative one, the stronger the linear relationship between the data sets. The sign of the coefficient denotes a positive or negative correlation, respectively. An $r$ close to zero denotes little or no linear relationship between the data sets. The $r$ can be compared among sample data sets to determine which combinations of data have the strongest relationship, and thereby, determine which factors have the strongest influence on each other or on a process (Davis, 2002; Jaisingh, 2006).
Linear correlation is useful for bivariate, small to medium data sets and is a powerful analysis because the assumptions involved are straightforward. The sample must be collected randomly, and the pairs of data must have a normal distribution (Davis, 2002; Jaisingh, 2006). The analysis does not give proof of connection but provides a measure of the correlation across the data set (Lowry, 2007).

A t-test can be used to test the significance of the correlation between two values (Lowry, 2007; Davis, 2002). For a given sample size and within two degrees of freedom, the t-test shows if the observed sample correlation is significantly different from a null hypothesis. The null hypothesis states that two variables are completely independent of each other and any nonzero $r$ value is purely coincidental (Davis, 2002). For the given sample size, $t$ must be over a given number for a significant correlation within 5% or 10% confidence (Table 1). I’ve included the t-test value, $t$, as well as the sample size, $N$, for each correlation of $r$.

I used a linear correlation analysis to determine which characteristics at the scarp transects had the strongest correlation to degradation of the scarp. These data were collected randomly at the transect locations. I compared various scarp characteristics to the degree of scarp degradation at each transect. To represent the degree of degradation at the scarps I used the cross sectional area of the talus slope (with larger areas implying greater degree of degradation) and the free face height with reference to the total scarp height (with smaller free face heights implying greater degree of degradation). I evaluated the correlation of each with a variety of characteristics: scarp height, column diameter and height, density of vesicles within the columns, position of the transect in the scarp segment, and the scarp facing direction.
Analyses

Proxies for degradation

The cross-sectional area of the talus slope is a proxy for the volume of the talus. The volume is a rough estimation of degree of degradation. The greater the area (volume) of talus, the greater amount of degradation has occurred. However, the use of the talus area as a proxy for scarp degradation may be misleading. It reflects the amount of material available to degrade, which varies between scarps and within a single scarp. The height of the scarp has a strong, positive correlation within 5% significance to the area of talus ($r = 0.64, t = 3.75, N = 22$) (Figure 7a). As the scarp height increases, the area of talus increases. Larger talus piles may simply reflect overall taller scarps that can provide ample material during degradation and more potential energy to speed degradation. The larger talus piles may not directly reflect the degree of degradation.

The free face height with relation to the total scarp height may provide a better proxy for the degree of degradation. Assuming the initial scarp was completely free face, the portion that is remaining represents degree of degradation. A portion of a scarp that has degraded to the debris and wash controlled stages of erosion would be represented by a lack of free face. The height of the scarp is not correlated to the height of the free face ($r = -0.19, t = -.87, N = 22$) (Figure 7b). There is a negative correlation within 5% significance between cross sectional area of talus and the normalized free face height ($r = -0.62, t = -3.53, N = 22$) (Figure 8) where the free face height is normalized by dividing by the maximum transect height. As the free face gets smaller, the talus area increases, as would be expected as more material falls from the free face, adding to the talus pile and decreasing the free face height.
The smaller the free face, the more a transect has degraded. This assumes the scarp began as 100% free face, as shown in Figure 2A and may not be a good proxy if there have been multiple faulting events. I correlate the additional variables to both talus cross sectional area and normalized free face height for comparison.

**Location of transect on scarp segment**

There are two conceptual models for fault generation and growth, each describing the location of maximum fault displacement on a fault segment. Cowie and Scholz (1992) model fault growth using a single, smooth and continuous surface of displacement in two dimensions. Their model shows that maximum slip occurs in the center of the segment, furthest from the tips. However, Willemse et al. (1995) document and model more complex slip distributions in three dimensions, considering fault segment interaction. Their model shows that the greatest slip is commonly seen near the tips of interacting segments. The location of the highest portion of the scarp is important to understanding scarp degradation. Higher portions may degrade more quickly because there is increased gravitational potential energy and ample material to erode. A correlation between the location of the scarp profile and the degree of degradation could support or refute scarp formation models in addition to better describing the varied degradation along a scarp.

The location of the transect relative to the scarp segment tip with relation to the talus area has weak negative correlation \( r = -0.34, t = -1.62, N = 22 \) (Figure 9). Segment length and transect location were measured from aerial photographs. For the analysis, transect location is represented as a fraction of the total scarp segment length; 0.10 and 0.90 are close to a segment tips, while 0.50 is located near the center of the segment. Talus cross sectional
area is greatest closer to the ends of the scarp. The location of the profile on the scarp has a very weak positive correlation to the normalized free face height ($r = 0.29$, $t = 1.4$, $N = 22$). The closer to the ends of the scarp, the smaller the free face. If I remove from consideration those transects that do not have a free face (dashed line in Figure 9b), the correlation is even weaker ($r = 0.23$, $t = 0.68$, $N = 10$). None of these values are within the 5% level of significance.

The maximum scarp height along strike was not recorded in this study, thus there is no way to confirm that the scarps are highest at the ends, thereby contributing additional material for a talus slope. The weak correlation between the normalized free face and the location of the profiles may suggest that there is a greater degree of degradation at the ends of scarps, possibly resulting from higher scarps, but again, the data show no significant relationship. Additional surveys on these scarps would be needed to fully support any fault generation models and thereby to infer degradation characteristics from fault shape.

**Joint spacing and column height**

During field work, I noted an apparent relationship between the size of columns and the degree of degradation (see also discussion of Figure 6). At profiles where there was little or no free face, columns were generally smaller than 0.45 m in diameter and height. At profiles with a larger free face, columns were generally larger than 0.45 m in height. I ran a correlation test to determine if the observed relationship was statistically meaningful.

Joint spacing is negatively correlated within 5% significance with the area of talus ($r = -0.43$, $t = -2.14$, $N = 22$). It has a strong, positive correlation of $r = 0.81$ ($t = 6.22$, $N = 22$) to the normalized free face height (Figure 10). When transects with no free face are
eliminated, the correlation is $r = 0.69$ ($t = 2.70, N = 10$). The column height is weakly correlated to the area of talus with $r = -0.35$ ($t = -1.66, N = 22$). This is not within 10% significance. Column height has a strong, positive correlation within 5% significance ($r = 0.65, t = 3.81, N = 22$) to the normalized free face height (Figure 11). When transects with no free face are eliminated, the correlation is $r = 0.74$ ($t = 3.1, N = 10$). The coefficients show that as the columns increase in size, the talus cross sectional area decreases and the free face height increase. Perhaps larger columns are more difficult to loosen, thus taking longer to topple and resulting in less voluminous talus. Columns larger than 0.45 m in diameter may topple primarily by ground shaking (Chapter 2).

To test the hypothesis that scarps with larger columns are more difficult to degrade, I ran a correlation test between joint spacing under 0.45 m and over 0.45 m with the normalized free face height. For columns with joint spacing under 0.45 m, $r = 0.49$ ($t = 1.487, N = 9$). This is not within 10% significance. For columns with joint spacing over 0.45 m, $r = 0.70$ indicating that larger columns correlate more strongly with higher free faces within 5% significance ($t = 3.40, N = 14$). In other words, although both tests produce positive correlations there is stronger correlation between less degradation and large columns.

**Vesicle density**

Selby (1993) suggests that minor partings such as vesicles are less important to the rock mass strength and resistance to failure than major partings such as joints. Selby (1993) does assume vesicles allow for an infiltration of water into the bedrock which may play a role in degradation. I ran a correlation test between average vesicle density at the scarps and the
degree of degradation. Vesicle density is not correlated with talus area \( (r = -0.037, t = -0.17, N = 22) \) nor with free face height \( (r = -0.068, t = -0.31, N = 22) \) within 10% significance (Figure 12). If I eliminate the transects with no free face, the correlation increases but not significantly \( (r = -0.46, t = 1.51, N = 10) \).

Selby (1993) suggests that the pore water pressure and frost wedging from infiltration through vesicles contributes to the loosening of the rock. However, scarps in this study are all in high altitude arid regions that receive only 22 - 35 cm of precipitation a year (Oregon Climate Service, 2006). If over 20% of the available pore space is empty, expansion of water upon freezing will not cause shattering or wedging (Ritter et al., 1995). Since the climate is so arid, there is not enough precipitation to saturate all the pore space within the columns larger than 0.45 m diameter. Thus, frost wedging or pore water pressure probably play only a minor role in toppling and degradation of the larger columns, at least in the present climate conditions of the region. However, in smaller columns this may be enough precipitation to fill up vesicles, aiding degradation.

To test the hypothesis that the vesicle density in smaller columns is correlated to degradation, I ran correlation tests for vesicle density of columns with diameters less than 0.45 m and for columns with diameters greater than 0.45 m. The vesicle density for columns greater than 0.45 m is correlated to the free face height \( (r = -0.45, t = -1.75, N = 14) \), but only within 10% significance. The vesicle density for columns less than 0.45 m is correlated within 5% significance to free face height \( (r = 0.65, t = 2.26, N = 9) \). For the smaller columns, as the vesicle density increases, the free face height increases. For the larger columns, as the vesicle density decreases, the free face height increases. This result does not appear to support the hypothesis that smaller columns should show smaller free faces with
greater vesicle density, but does lend some support to the hypothesis that vesicle density plays less of a role in degrading larger columns.

Scarp facing direction

The facing direction of a scarp may affect the degree of solar radiation the scarp receives, the amount of weathering the scarp face experiences, and the vegetation on the scarps. These features likely influence scarp degradation. I tested this by running a correlation test between the scarp facing direction (in degrees azimuth) against the cross sectional area of the talus and the normalized free face height. Lower azimuth means more easterly facing, and higher azimuth means more westerly facing. I had no scarps that faced north or south.

There is a strong positive correlation ($r = 0.58$, $t = 3.18$, $N = 22$) between the scarp facing direction and the normalized free face height (Figure 13). Facing direction is negatively correlated to the volume of talus ($r = -0.67$, $t = -4.03$, $N = 22$). Both are within the 5% significance level. This means that more easterly facing scarps tend to have smaller free faces, larger talus piles, and an overall greater degree of degradation. Northeast and east facing scarps tend to preserve snow, thus increasing water available for weathering and degradation of the scarp.

Summary

The strongest correlations are between joint spacing and column height to degree of degradation as shown by the normalized free face height. Correlations between joint spacing and column height to talus cross sectional area are also strong. The location of the profiles
on the scarp segment has no significant correlation to the degree of degradation. Vesicle
density has no correlation to the degree of degradation. Although there is a correlation
between the degree of degradation and scarp facing direction, different transects along the
same scarp at all the scarp locations have varying talus volumes and varying free face heights
with respect to the total scarp height (Figure 14). Additionally, joint spacing can vary along
a scarp face, for example, where the fault cuts a flow margin (Figure 15). There is a
correlation between joint spacing and degree of degradation. This implies that degree of
degradation on scarps through jointed basalt has little to do with the age of the scarp, at least
over the past two million years.

How do scarps in jointed basalt erode?

The characteristics of the columnar joints within the basalt appear to play a
significant role in the degree of degradation a scarp undergoes. Profile sections of the scarp
with columns of larger dimension correlate to higher free faces and smaller talus slopes,
which imply less degradation. The profile shape is therefore less dependent on time and
more dependent on bedrock characteristics.

How, then, does jointing in the basalt affect how the scarp erodes? The shape of
columns make the scarps more stable than if the bedrock were irregularly jointed. Both large
and small columns generally have a uniform block shape (Figure 16). Unless the block has
an uneven base, the block should be relatively stable; a topple force, such as ground shaking
or biogenic force, is required to make the blocks tip.

Several transects sites had vegetation growing between two columns, commonly
along the crest of the scarp (Figure 17). The roots may exert force on the columns, especially
small ones, eventually causing the columns that comprise the free face to tip. The force roots can exert is not well documented, thereby complicating any attempt to determine if the potential force from roots is enough to topple columns. Additionally, vegetation and the accumulation of organic debris between columns holds moisture which may accentuate the freeze/thaw effect on weathering and erosion of the columns.

Regional ground shaking causes otherwise stable columns to topple, speeding degradation (Chapter 2). The force needed to tip even the largest columns is relatively small and could be produced by local earthquakes between magnitudes 3.8 and 5, which are common in this area. This could also account for the relatively uniform retreat along the scarp, which sporadically spaced vegetation could not produce.

Spalling, the loosening and removal of large pieces off a rock surface, likely controls gradual degradation of the scarps. Spalling can be induced in many ways. Thermal spalling occurs as uneven heating and thermal expansion of fluids in microcracks loosens a slab. Range fires can induce thermal spalling (Bierman and Gillespie, 1991; Hettema et al., 1998). Microcracks can also form in an impact or from blasting, speeding erosion by spalling (Ahrens and Rubin, 1993). Also, clay and dust can accumulate in fissures, eventually resulting in a precipitation of carbonate which loosens and widen fissures, causing spalling (Dorn and Cerven, 2005). I saw spalled slabs at the base of scarps in the field and observed white deposits along the fronts of some joints (Figure 18).

In summary, bedrock scarps in basalt erode by the topple of columns and from spalling of small slabs off the free face. The columns topple either from biogenic force or from regional ground shaking. The accumulation of sediments or mineral deposits in microcracks likely contributes to spalling.
Conclusion

Normal faults scarps offsetting jointed basalt in central Oregon and northern California are persistent despite being Quaternary in age. The scarps displace bedrock, which Wallace (1977) shows to notably increase the amount of time the scarp remains in a gravity controlled state (Figure 3). Jointing decreases bedrock strength, but the form of columns also appears to affect the degradation rate. Because the columns are generally block shaped and therefore stable until a topple force is exerted, the free face persists longer than scarps through alluvium. Toppling may be initiated by regional ground shaking or by biogenic processes.

Additionally, joint spacing appears to control the degree of degradation a section of a scarp experiences. For transect profiles of the scarps, larger columns are positively correlated to the height of the free face and negatively correlated to the area of talus, both of which represent the degree of degradation on the scarp (Figures 10, 11). Climate of the region has not been constant over the Quaternary (Whitlock and Bartlein, 1994). Biogenic processes or freeze/thaw might have had a greater influence on scarp degradation than we can now observe.

Additional work is needed to constrain the portion(s) of scarp segments that degrades the quickest. Detailed profiles and characteristic measurements taken at many spots along one scarp segment of well constrained age, such as Boyd/Skeleton scarp, would provide a database describing the varying degradation along one scarp to use for modeling or to constrain relationships between the variables that play a role in degradation. A survey of
profiles along segment arrays, such as at Golddigger scarp, may provide insight on how the interaction of segments play a role in degradation of the scarps.

A survey of normal fault scarps through jointed basalt that are older than two million years is necessary to constrain the amount of time necessary for the entire scarp to degrade fully to a talus slope. This would refine the database by describing the stages of degradation of scarps at significantly different ages.

Constraining the time since formation of the scarp and time since topple of basalt blocks would be vital to estimating a rate of degradation, despite the degradation appearing to be episodic rather than continuous. Cosmogenic surface exposure dating may be an appropriate technique (e.g. Zreda and Noller, 1998), although sufficient sample density would be expensive.
Table 1: Critical values of $t$ for N-2 degrees of freedom within 5% and 10% (95% and 90% confidence) confidence levels of significance. From Davis (2002).

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Figure 1 – Block diagram of a normal fault scarp in alluvium showing terminology. After Wallace (1977).
Figure 2: Scarp degradation sequence from gravity controlled (B) to wash controlled (E). After Wallace (1977).
Figure 3: Graph showing limits of principal slope angle versus the age of the fault scarp adapted from Wallace (1977). Note that fractured bedrock remains in a gravity controlled state for much longer than scarps in alluvium.
Figure 4: Quaternary faults in the Pacific Northwest and northern California. Holocene scarps in blue. Volcanoes as green stars. Location of scarps for this study in red. Location of Newberry flows, Lava Beds National Monument, and Devil's Garden on Modoc Plateau shown by red boxed. Adapted from Pezzopane and Weldon, 1993.
Figure 5: Diagram showing ages of samples examined in this study (Personius 2002; Shear and Smith, 2000; Jensen, 1987; Hedel, 1984; Duffield and McKeen, 1974; Pezzopane and White, 1993). See text for discussion of age constraints.
Figure 6: Scarp transect profiles grouped according to overall shape characteristics and normalized by height and toe-to-crest length to accommodate comparison. The profiles are also shifted so that crest is the anchor for the transects. (A) shows profiles that are predominantly talus slopes. (B) shows profiles that have both a talus slope and a free face. (C) shows profiles that are predominantly free faces. Individual profiles are shown in Appendix A. [BR= Barr Road scarp, BS= Boyd/Skeleton scarp, BH= Big Hole scarp, OF= Oatman Flat scarp, GD= Golddigger scarp, PJ= Porcupine Jr. scarp].
Figure 7: (a) Cross-sectional area of talus (m$^3$) versus the total scarp height (m). There is a positive linear correlation that shows as scarp height increases, volume of the talus pile also increases. (b) Scarp height (m) versus the cross-sectional normalized free face height. There is no correlation between the scarp height and the free face height. Normalized free face is the free face height normalized to the maximum scarp height in all following figures, unless otherwise stated.
Cross-sectional area of talus pile (m²) versus normalized free face height

Figure 8: Area of talus (m²) versus the normalized free face height. The negative linear correlation shows that as the free face decreases, the volume of talus increases. Dashed line shows correlation after eliminating transects with no free face, where N = 11.
Figure 9: (a) Location of the transect along the segment versus cross sectional area of talus (m²). (b) Location of the transect along the segment versus the normalized free face height. The dashed line represents the trend of the correlation omitting free face values of zero where $r = 0.234$, $N = 10$. In each graph, 0.0 represents the end of the scarp segment, while 0.5 represents the center of the segment.
Figure 10: (a) Joint spacing (cm) versus the cross sectional area of talus (m²). (b) Joint spacing (cm) versus the normalized free face height. The black dashed line represents the trend of the correlation omitting free face values of zero where $r = 0.688$, $N = 10$. In each graph, red dash-dot line shows cutoff between column diameter greater or less than 45 cm.
Figure 11: (a) Column height (cm) versus cross sectional area of talus (m$^2$). (b) Column height (cm) versus the normalized free face height. The dashed line represents the trend of the correlation omitting free face values of zero where $r = 0.740$, $N = 10$. 

**Cross-sectional area of talus (m$^2$) versus column height (cm)**

- $r = -0.347$
- $t = -1.66$
- $N = 22$

**Normalized free face height versus column height (cm)**

- $r = 0.648$
- $t = 3.81$
- $N = 22$
- $r = 0.747$
- $t = 3.11$
- $N = 10$
Figure 12: (a) Vesicle density (%) versus cross sectional area of talus (m²). (b) Vesicle density (%) versus the normalized free face height. The dashed line represents the trend of the correlation omitting free face values of zero where \( r = -0.475, \ N = 10 \).
Figure 13: (a) Scarp facing direction (°) versus the cross sectional area of talus (m²). There is a strong negative correlation of $r = -0.670$. (b) Scarp facing direction (°) versus the normalized free face height. There is a strong positive correlation of $r = 0.580$. More easterly facing scarps have smaller free faces and larger talus piles. Dashed line represents the trend of the correlation omitting free face values of zero where $r = 0.578$, $N = 10$. 
Figure 14: (a) Cross sectional area of talus (m^2) for each scarp. (b) Height of free face (m) for each scarp. Note that talus volume and free face height change within same scarp, suggesting that facing direction, bedrock chemistry and local climate are not the primary roles in degradation. [Scarps denoted by color: red = Barr Road, black = Boyd/Skeleton, purple = Big Hole, orange = Oatman Flat, green = Golddigger, blue = Porcupine Jr.] Blank spots are transects where talus volume or free face height is zero.
Figure 15: Joint spacing differs along strike at scarps in this study, such as at Porcupine Jr. scarp, where the fault cuts a flow margin.
Figure 16: Diagram showing the center of mass for a block. The center of mass is directly below the center of the block. If the base of the block is even and the balancing points are the edges of the block, the block is stable until some force causes toppling. The $\alpha$ is the angle between the center of mass and the balancing points of the block.
Figure 17: Vegetation and organic debris behind columns may speed degradation by biogenic force or by retaining moisture. Upper photo from Boyd/Skeleton scarp in central Oregon; lower photo at Big Hole scarp in central Oregon.
Figure 18: Slabs spalled off the front of the free face, Porcupine Jr. scarp. Top of reflector is approximately 12 cm x 12 cm. Note white deposits at exposed surface denoted by box.
Chapter 2: Strong ground motion as a geomorphic agent in scarp degradation

Introduction

In seismically active regions, precariously balanced rocks can be used as strong motion sensors (Shi et al., 1996). The size and shape of the precarious rocks can be used to estimate the minimum ground acceleration necessary to topple the rocks (Anooshehpoor et al., 2004), thereby providing constraint on the maximum ground motion that has occurred in the region. This also provides information about seismic hazard that may not be determined from recorded data due to expense of equipment and/or a small database of recorded data for large earthquakes (Anooshehpoor et al., 2004; Brune, 1996). For example, in California, there have been little data recorded for great earthquakes because they occurred before instrumentation was available (or even invented!). These great earthquakes control the maximum ground motion and seismic hazard for long recurrence times, and are therefore essential to constructing probabilistic seismic hazard maps (Brune, 1996). Without proper estimation of the recurrence interval for great earthquakes, hazard maps might over- or underestimate the potential risk, thereby impacting the cost of earthquake hazard mitigation. The toppling of precarious rocks due to a great earthquake may provide insight to the recurrence interval of these events (Brune, 1999; Anderson and Brune, 1999).

Fault-bounded basalt columns at normal fault scarps may act as strong motion sensors. Because jointed basalt is commonly associated with extensional faulting, toppled columns are readily available for earthquake constraint in many actively extending regions in the world, including the Basin and Range Province of North
America (Wallace, 1977; Bucknam and Anderson, 1979), Iceland (Opheim and Gudmundsson, 1989), and the west flank of Kilauea volcano, Hawaii (MacDonald, 1957). Estimates of the ground acceleration necessary to topple the columns can provide a basis for estimating long-term regional seismic hazard and the procedure can be validated by comparison with previously published hazard assessments. The scarps in basalt in central Oregon and northeastern California offer the opportunity to evaluate toppled columns as strong motion sensors. The approach presented here provides additional information on earthquake recurrence that is useful because the region has a short historical seismic record and a relatively sparse seismic network (Pezzopane and Weldon, 1993).

**Background**

Toppling is a form of mass movement characterized by overturning of a block through rotation or flexure (e.g. Pritchard and Savigny, 1990). Toppling occurs in jointed rock masses where the principal set of fractures strikes parallel to the rock or slope face; the fractures commonly dip steeply into the face (Norrish and Wyllie, 1996). There are several types of rock that are most susceptible to this type of failure: sedimentary units with well defined bedding and/or systematic jointing, metamorphic units with well developed foliation, and volcanic rocks with columnar joints (Norrish and Wyllie, 1996).

Toppling occurs when the center of gravity for a block falls outside the dimensions of its base. Toppling is characterized by horizontal movement at the crest of the block and rotation around a central point. Blocks have a stable factor of safety when bedding planes are horizontal and the center of gravity for the block is within the basal area (Norrish and Wyllie, 1996). The “factor of safety” is used by engineering geologists
to quantify the stability of hillslopes (Norrish and Wyllie, 1996). Toppling of precariously balanced rocks may occur due to ground shaking related to seismic events (Anooshehpoor et al., 2004). Similar to a precariously balanced rock, I expect that a joint-bounded block or column exposed at a fault scarp that is stable under normal conditions can topple under the influence of strong ground shaking.

Toppling models can be complex for precariously balanced rocks. In many cases, the precarious rock has more than one balancing point (Shi et al., 1996). When strong ground motion exceeds horizontal acceleration necessary for rocking, a precarious rock will rock back and forth on balancing points. Depending on its shape and size, the rock may topple (Shi et al., 1996). The acceleration necessary to initiate rocking to topple is called the quasi-static peak toppling acceleration (Anooshehpoor et al., 2004). At high shaking frequencies, the peak acceleration necessary to topple a rock increases with the size of the rock. Stability occurs when the rock is short, wide and uniform in shape (Shi et al., 1996). Precariously balanced rocks begin rocking but do not topple when subject to acceleration pulses less than the quasi-static peak acceleration. The rock may topple when subject to a series of forces with various frequencies while rocking. This is the dynamic toppling acceleration (Anooshehpoor et al., 2004). The columns are scarp bound on one side and therefore have limited or no freedom to rock. I therefore limit this analysis to the quasi-static peak toppling acceleration, further referred to as the peak acceleration necessary for toppling.

The necessary parameters for determining the toppling acceleration can be measured in the field: dimension of the block and the angle between the center of mass
with respect to the balancing points. The overturning force has the simple form (Anooshehpoor et al., 2004):

$$ F = mg \tan \alpha $$

(2)

where \( g \) is the acceleration due to gravity, \( m \) is the mass of the rock, and \( \alpha \) is the angle between the center of mass and a balancing point (Figure 16, Chapter 1). The quasi-static peak ground acceleration necessary to topple the rock can be used to determine the regional ground shaking (Anooshehpoor et al., 2004). From above, acceleration may be determined by:

$$ a = \frac{F}{m} = g \tan \alpha $$

(3)

If there are two rocking points, two toppling accelerations must be determined for the rocking motion of the precarious body (Anooshehpoor et al., 2004), but equation (3) may be used for rocks that are constrained to topple in only one direction, such as fault-scarp bounded blocks. In either case, the topple direction is not necessarily the direction of maximum ground motion felt at the site, so constraints on direction of maximum motion are statistically better at sites with a group of toppled rocks (Anooshehpoor et al., 2004).

The assumptions necessary to calculate the toppling force and acceleration are: 1) that the periodic rocking motion and topple are inelastic (no bouncing) during the transition from one rocking point to the other and once the column topples; 2) that there is no sliding of the precarious rock; and 3) that rocking won’t begin unless the peak
toppling acceleration is exceeded for the given rock (Shi et al., 1996; Anooshehpoor et al., 2004).

**Geology of the study areas**

During field reconnaissance, toppled basalt columns with heights greater than 1 m and diameters over 0.45 m were found three scarps described in Chapter 1 (Figures 19-21). Boyd/Skeleton scarp, northwest of Newberry Crater, Oregon offsets Pliocene basalt flows (Hawkins et al., 1988). Porcupine Jr. Scarp offsets Tertiary basalts in the Devil’s Garden region of the Modoc Plateau in northeastern California (Hedel, 1984). Oatman Flat Scarp, in the Paulina Marsh fault zone of southcentral Oregon, offsets Miocene-Pliocene basalt. The faults are Quaternary in age and are constrained by offset of dated Quaternary deposits, Mazama ash, and K-Ar dating of faulted bedrock (Pezzopane and Weldon, 1993; Personius, 2002; Jensen and Chitwood, 2000; Hawkins et al., 1988).

Boyd/Skeleton and Oatman Flat scarps trend to the northwest and are down to the southwest. Porcupine Jr. scarp trends to the north and is down to the west. Each scarp is characterized by 2-10 m vertical offset, a free face and a talus slope (Figure 1). The vertical joints or columns that make up the free face of the scarps vary in size from 0.02-2.5 m diameter and 0.05-2.5 m height. The basalt is vesiculated, with less than 10 vol. % vesicles toward the base of flows, increasing to 25-30 vol. % vesicles near the top. The increase in vesicularity toward the top of columns was used to identify the original upper surface of toppled columns in the field.
Strength characteristics of jointed basalt

Field measurements of jointed rocks can be used to estimate the rock mass strength in a classification system that weights various parameters according to previous engineering and rock mass strength studies (Selby, 1980). These parameters include joint orientation, spacing, width of partings, groundwater outflow, degree of weathering, and continuity of joints and are given a rating with is related to the rock mass strength (Table 2). Using this field classification system, I determined the rock mass strength class for the joints at the scarps. The columns are slightly weathered with joint spacing between 0.3 – 1 m. The columns are oriented vertically with respect to the scarp face, and have joint partings less than 10 mm, are continuous with no bridges of intact rock and there is no visible ground water flowing out of the scarp. This corresponds to a rock mass strength rating (RMR) of 61, which falls under the “moderately strong” strength parameter.

In the study area, scarps with smaller columns (less than 0.45 m diameter) are more degraded and have steeper, more voluminous talus slopes, implying that they degrade more readily than scarps with larger columns (Chapter 1). Under normal conditions, the blocks within the scarps should not fail because the flow layers do not dip and the columns have centers of gravity within their basal area, resulting in a stable factor of safety (Norrish and Wyllie, 1996). Freeze/thaw mechanisms and pore water pressure commonly weather and erode rock. However, if over 20% of the available pore space is empty, expansion of water upon freezing will not cause shattering or wedging (Ritter et al., 1995). The field region is relatively arid, receiving between approximately 22 - 35 cm of precipitation a year, with most of that occurring during winter months as snow (Oregon Climate Service, 2006). There is not enough precipitation to saturate all the pore
space between the larger columns (>0.45 m diameter). Thus, I expect that pore water pressure and frost wedging play a minor role in increasing the failure potential of the columns in this region. I note that columns may topple by root forcing from past localized vegetation, although there is little vegetation currently near the toppled columns. There has been little work on the force roots may exert, and this variable is therefore difficult to quantify. For this discussion, I posit that the larger basalt columns are difficult to overturn without intense ground shaking.

**Observations of toppled columns**

There are seven toppled columns at Boyd/Skeleton scarp in central Oregon ranging in height from 0.54 – 1.48 m and in diameter from 0.40 – 1.04 m. The columns lie 45 – 135° from vertical (plunge ±45°); increasing vesicularity shows the top of the column (Figure 19). Three toppled columns at Oatman Flat scarp in south-central Oregon have heights between 1.06 - 1.37 m and widths between 0.43 - 0.60 m. The columns all lie approximately 90° (plunge=0) from vertical at the base of the free face. There is one toppled column at Porcupine Jr. scarp in northeastern California (Figure 20). It has a height of 2.04 m, a width of 1.37 m and lies approximately 45° from vertical.

Columnar joints form perpendicular to the cooling surface (DeGraff and Aydin, 1987). I expect vertical columns except at flow boundaries, where the columns form perpendicular to the cooling margin. Flow margins are recognizable in the field by sinuous map traces and/or decreased joint spacing (Figure 15). The scarps have vertical columns at the free face, suggesting the free face is not a flow margin. Additionally, the
scars form long, linear segments. I am confident that these columns are not near flow margins.

Other workers have reported tilted columns due to monoclinal folding during or preceding normal faulting (White and Crider, 2006; Grant and Kattenhorn, 2004). Under normal conditions, the columns should not fail by folding or toppling if the center of gravity is within the base area. When the stability is disturbed, a column topplies by disconnecting from other columns and rotating away from the slope (Norrish and Wyllie, 1996). At the scarps studied, each toppled column is disconnected at the base from the columns below and is not joined to other toppled columns on any side, which suggests the columns have toppled independently, and that tilting of the columns is not due to folding. At each scarp, the toppled columns are lying on a talus or wash slope below the free face of the scarp (Figure 21). This suggests the columns did not topple with initial scarp formation because there had to be time for material to be eroded off the scarp to form the talus or wash slope.

Methodology

In the field, I measured the dimensions of the toppled columns with a tape (Table 3). The angle, \( \alpha \), between the center of mass and the toppling point of the precarious object, can be measured directly in the field for precariously balanced objects with erratic shape. However, basalt columns are essentially rectangular blocks or cylinders for which the balancing points are the edges of the column. For a rectangle or cylinder in a homogenous material, \( \alpha \) equals the inverse tangent of half the width or diameter divided by half the height which simplifies to width over height (Figure 16). Since the width or
diameter are approximately the same in the columns, \( \alpha \) is the same whether we treat the columns as rectangular blocks or cylinders:

\[
\alpha = \tan^{-1}(w/h)
\]  \hspace{1cm} (4)

Substituting into equation (3):

\[
a = g \frac{w}{h}
\]  \hspace{1cm} (5)

It may not be appropriate to assume the base of the column is perfectly flat, especially if the base represents the top of the underlying basalt flow (Figure 22). An uneven base would create balancing points other than the edges of the column. Equation (4) assumes the edges of the block are the balancing points, but balancing points away from the edges effectively decreases the width, thereby decreasing \( \alpha \) and changing the peak ground acceleration necessary to topple the column. Tall, slender columns topple more readily. As \( \alpha \) increases, it takes a larger ground acceleration to topple the column (Figure 23). Thus, assuming the edges of blocks are balancing points yields the largest acceleration required for toppling and is the most conservative estimate.

I used equations (4) and (5) to determine the peak ground acceleration necessary to topple each column (Table 3). Since most hazard maps present ground motion as a percent of gravity, I converted the accelerations to better compare my results with published estimations of ground shaking hazards. Previous studies typically address the maximum ground shaking necessary to topple currently upright precariously balanced rocks (Anooshehpoor et al., 2004; Shi et al., 1996; Brune, 1996). The basalt columns that topple expose new columns as the scarp face, so determining the minimum acceleration
necessary to topple an already toppled column is analogous to determining the maximum acceleration to topple the next, standing column, assuming they are of similar size.

Peak ground acceleration can be used to calculate the Modified Mercalli intensity according to the equation (Murphy and O'Brien, 1977):

\[
\log a = 0.25 I_{mm} + 0.25
\]  

(6)

where \(a\) is the acceleration and \(I_{mm}\) is the intensity. This equation is based on empirical observations that as acceleration increases, intensity increases. The constants were determined empirically by analyzing over 900 worldwide seismic events for which acceleration and intensity data were available (Murphy and O'Brien, 1977). I used equation (6) to convert the peak ground acceleration calculations to intensity (Table 3).

Modified Mercalli intensity is a qualitative description of ground shaking during an earthquake which is calculated from the degree of damage to structures, the degree of shaking felt by observers, and the amount of secondary damage from landslides or liquifaction. There are many intensities for a given earthquake, usually with the highest intensities nearer the maximum fault displacement (Yeats et al., 1997). In this sense, intensity is similar to ground acceleration. Both decrease further from the maximum displacement. Intensity is not directly proportional to magnitude since it depends on local geology and distance from the maximum displacement. However, a general scale relating the two has been produced through many observations comparing intensity to magnitude (Table 4) (Earthquake Hazards Program, 2006). It is important to note that a given intensity could be caused by a close, small event or a large event further away.
Results and Discussion

The calculations indicate accelerations of 0.39-0.87g were required to topple the large columns measured for this study (Table 3). The columns with the largest dimensions do not necessarily need the largest accelerations to topple. The columns with the ratio, w/h, approximately 1, such as the column at Boyd/Skeleton scarp (topple E, 0.54 x 0.47 m), need the largest accelerations to topple because they are more stable.

I used the USGS National Seismic Hazards (Frankel et al. 2000) peak ground acceleration probability maps to compare the ground accelerations I calculated to regional ground shaking hazard. The probability maps show contours of the maximum acceleration for which a region has a 10% probability of exceedence in 50 years. A site within the 10% g contour will be more likely to feel small events or distant large events, while a site within the 90% g contour will be more likely to feel larger events, even though each site has only a 10% chance of feeling that ground motion in 50 years.

The probability maps are estimations of ground motion calculated from historical seismic and geologic data concerning the recurrence rate of fault ruptures (Earthquake Hazards Program, 2006). For a given area, all potentially active faults are identified, typically by aerial photographs. The probability hazard maps also use geologic data obtained from trenching faults, measuring slip from scarps, and coring offset stream deposits to support the seismic record (Frankel et al. 2000; Frankel et al., 2002). These data are used to construct a frequency-magnitude plot. The recurrence curves are combined with attenuation data describing the decrease in acceleration with distance from the scarp to estimate the likelihood a region will experience a given acceleration (Yeats et al., 1997). The probability curve is then transferred to a map.
The seismic hazards maps suggest that in the next 50 years there is a 10% chance that the region near Boyd/Skeleton scarp in central Oregon will feel ground shaking between 0.08-0.10 g. Oatman Flat scarp in southcentral Oregon will have a 10% chance of feeling between 0.09-0.10 g. Porcupine Jr. scarp in northeastern California will have a 10% chance of feeling between 0.10-0.15 g (Figures 24).

According to my calculations, it takes between 0.39 and 0.87 g to topple the columns at these scarps. While consistent with ground acceleration estimates from precarious balanced rocks in other studies (Weichert, 1994; Brune, 2001), these accelerations are much higher than the 10% probability maps predict the scarps will feel in the next 50 years. The maps are estimations of future hazard, but are based on data from historical events and geologic information, so a low probability of future hazard corresponds to low event recurrence. According to these seismic maps, it will take more than 50 years for the region to feel strong accelerations.

The probability maps suggest the regions will not feel the ground shaking necessary to topple such large columns in the near future, and likewise, hasn’t felt those accelerations in the recent past. What is the regional seismic recurrence on a longer time scale? The acceleration necessary to topple the columns can be related to intensity of ground shaking and event magnitude (Gutenberg and Richter, 1942; Murphy and O’Brien, 1977; Stein and Wysession, 2003). Earthquake statistics, such as the distribution of earthquake magnitude in a region, can be used to predict the recurrence interval of large earthquakes (Stein and Wysession, 2003).

The distribution of earthquakes worldwide follows the Gutenberg-Richter relation. This maintains that the number of earthquakes occurring around the world
varies with magnitude and that successively smaller events are more common (Stein and Wysession, 2003; Yeats et al., 1997). This relation, also called the frequency-magnitude relation, is invariant with scale, so that it applies worldwide or to smaller regions. The relationship is linear in semi-log space and is shown by (Gutenberg and Richter, 1954; Stein and Wysession, 2003):

$$\log N = a_1 - bM$$  \hspace{1cm} (7)

where $N$ is the number of earthquakes with magnitude greater than $M$ for a given time period (Figure 25). The intercept, $a_1$, depends on the number of events in the given time and region, but the slope of the line of best fit, $b$, is generally about 1. This means there is approximately a tenfold increase in occurrence for successively smaller events. This relationship is used to construct recurrence curves, such as are used in creating probabilistic hazard maps (Yeats et al., 1997).

This relation is generally utilized for large sample areas and populations, but can be applicable to smaller regions with a large earthquake catalog (Stein and Wysession, 2003). The relation will deviate from the expected curve if the number of large earthquakes in the given region is small, if the sample population is small, or if there are earthquake swarms associated with volcanic activity (Stein and Wysession, 2003). These variations will skew the curve, because the lack of evenly distributed samples will make it appear that large events are less frequent than a more comprehensive population would predict (Stein and Wysession, 2003).

Using the Advanced National Seismic System earthquake catalog (ANSS, 2006), I extracted data about earthquakes greater than $M=3$ in the region during the period 1970-
2006 (Table 5). Data before 1970 were deemed incomplete due to the lack of a seismic network and the sparse population of the region, and were therefore not used. In total, there were 354 events between magnitudes 3 and 6. I created a graph showing the log of the frequency of events per magnitude between 1970 and 2005 in the region between 40-45° N, 117-122° W (Figures 26, 27). This region represents the northern portion of Basin and Range extension without including seismic events related to the volcanic arc or offshore transform boundaries. Although for a relatively small region and timespan, the graph approximately follows the Gutenberg-Richter relationship, with the slope of the trendline $b = 0.8$ (Figure 27).

The intensity range calculated from the ground accelerations necessary to topple the columns is between III and V (Table 3). Assuming the columns toppled from ground shaking caused by local earthquakes and the shaking is not a result of a large magnitude event a long distance away, the intensity values can be loosely correlated to magnitude. Therefore, the minimum moment magnitude to topple the columns is between approximately 3.8 and 5.0 (Table 4). The seismic catalog and recurrence graph suggest that less than one event between magnitude 4.0 and 5.0 occur yearly, but up to 52 events of this size between 1970 and 2005 (Table 5).

I calculated the number of events for a given magnitude per year from the frequency-magnitude relationship in Figure 27 and divided this interval by the cataloged region area (236,745 km²) to get the number of events, per year, per km² (Table 6). I also determined the number of events per year within 10 km² of a scarp (Table 6) to eliminate the uncertainty caused by distance from source on the ground acceleration and intensity a column feels. This gives the expected average number of local earthquakes with
magnitudes necessary to topple columns near any given scarp in a year. The rate is very small but adds up quickly over larger time spans. For example, while there are no expected local events in any given year, there are approximately 9 local earthquakes in 10 ky, approximately 400 events every 500 ky, and about 850 events every 1 my. This suggests that there are events frequently enough through time to account for toppled columns at scarps.

Another, more traditional method for estimating the maximum magnitude a fault or fault zone can produce is by using the fault length to estimate the rupture area of a fault, and thereby to estimate the moment and moment magnitude possible if the entire fault ruptures (Hanks and Kanamori, 1979). This is shown by the equation:

$$M_w = \frac{2}{3} \log M_o - 10.7$$

where $M_w$ is the moment magnitude and $M_o$ is the moment of the event shown by the equation:

$$M_o = \mu AD$$

where $\mu$ is the shear modulus ($\mu = 3 \times 10^{11}$ dyne/cm$^2$), $A$ is the rupture area and $D$ is the amount of slip. Using equations (8) and (9), I calculated the maximum moment magnitude for varying rupture areas and amounts of slip. Scarp segments in central Oregon tend to be under 2.5 km in length (Hawkins et al., 1988) so I chose lengths varying in size from 0.5 km to 2.5 km. In the absence of information about the depth of faulting, I choose to keep the depth of the rupture plane constant at 1 km for simplicity in calculations. Amount of slip ranged from 1-5 m as the range of single-event scarp heights commonly seen in the region. The resulting possible magnitudes range from 4.7 to 5.7 (Figure 28). These values equal or exceed the estimated magnitude required to
topple the observed fallen columns. Therefore, given the assumptions previously discussed, faults in this region can produce events of magnitude sufficient to topple large basalt columns. Frequency calculations suggest that between two and six M 5.5 – 6 earthquakes occur every 500,000 years in the region (Table 6).

It is possible to roughly estimate scarp degradation rates by combining estimates of magnitude required for column topple with the regional frequency-magnitude relationship. Assuming a column diameter of 0.50m, there are 2000 columns along the face of a 1 km long scarp. Only a few toppled columns were observed at each scarp, so assuming that only 10 columns topple during a local earthquake of sufficient magnitude, then approximately 200 events are necessary to topple all the columns along the face of the scarp. Thus, 200 events are necessary for the scarp to retreat 0.50m. Following the frequency-magnitude relationship for the region, it will take approximately 250 ky for 200 events between M3.5-6.0 to occur within 10 km\(^2\) of a scarp. This gives a retreat rate of 10\(^{-6}\) mm/yr. This estimate is based on many assumptions and could be improved by a more comprehensive census of toppled columns at a larger number of scarps.

Wallace (1977) predicts that fractured bedrock will persist through the gravity controlled stage of degradation (i.e. maintain a free face) for approximately 10\(^4\) – 10\(^5\) years (Figure 3). Wallace’s (1977) model infers degradation rates two to three orders of magnitude higher than my estimated degradation rates. This may be because the columns are vertical and therefore stable until a force is exerted to induce toppling. This estimation also assumes that ground motion is the only factor to induce toppling of columns and discounts ground motion from large, distant earthquakes. These values are
speculative, but represent the first estimate of degradation rates for scarps on the high lava plains.

**Time since toppling of columns**

Absolute ages of toppling are required to test hypotheses about scarp degradation rates. Other studies (e.g. Brune, 1992) use rock varnish to estimate the time since topple. No rock varnish was observed on the columns in this study, but it is possible that toppled columns with varnish may be located in additional studies. Lichenometry is another relative dating technique which uses the diameters of lichen growing on toppled rocks to estimate the time since exposure. This technique has been used successfully in dating regional rockfall related to seismic events (Bull and Brandon, 1998). However, thousands of lichen must be measured for this method to be statistically sound. While the toppled columns have some lichen growth, not only are there not enough columns, there are not enough lichen for this to be a viable method in this instance.

Radiometric techniques, such as cosmogenic surface exposure dating, may be useful for precisely dating the toppled columns. Cosmogenic dating measures the accumulation of isotopes that are created with the interaction of cosmic rays with the Earth’s surface. The production rate of these isotopes is well constrained and surfaces that have been exposed to the atmosphere can be numerically dated (Pavich, 1987). This may be the best option for dating the toppled columns. Prior to topple, one or more surfaces should be adjacent to the scarp face and not exposed to the atmosphere. After toppling, however, these surfaces will be exposed. Discrepancies may arise if the columns are partially exposed prior to toppling due to wide joint partings or partial tilting.
from a prior seismic event. These discrepancies could be reduced if nearby columns, the hanging wall, and the foot wall are also dated for more constrained values for erosion and toppling. The dates from the hanging wall and foot wall provide the maximum exposure time of all the basalt. The freshly exposed side of the toppled column and of the column exposed at the free face, which was at one time behind the toppled column, should be similar if the column toppled all at once, exposing both surfaces. Many samples would need to be dated to reduce error. This may be impractical due to the expense of cosmogenic dating.

Conclusions

Basalt columns over 0.45 m in diameter should topple only with strong ground shaking (>0.39g) and therefore may be used as strong motion sensors. For columns found at Quaternary normal fault scarps in central Oregon and northeastern California, I calculated ground acceleration necessary for topple to be between .39 and .97g. This corresponds to minimum local earthquake magnitudes between 3.8 and 5.0. The probabilistic seismic hazard maps and a recurrence curve constructed from historical seismicity suggest a recurrence interval in the region for this size events every 10 to 100 years. The number of events necessary to topple between 10 ky to 1my yrs within 10 km² of a scarp is enough to account for the toppled columns. Based on the lengths of fault segments, with a rupture surface of 1 km depth and various magnitudes slip for a single event, I estimate a regional maximum magnitude to be 5.7.
The recurrence interval of events necessary to topple columns may account for the persistence of scarps through jointed basalt if strong ground motion plays a primary role in scarp degradation. A speculative estimate suggests degradation rates of $10^6$ mm/yr ± one order of magnitude.

In regions of low historical seismicity, especially in regions of active extensional tectonics where faulting and basaltic volcanism are commonly associated, toppled basalt columns may also be used to estimate strong ground motion. Used with relative or cosmogenic dating techniques, the toppled columns could be powerful paleoseismological tools. Additional work to apply dating methods and to potentially identify events which cause topples must be done to refine the approach before use in earthquake recurrence studies or seismic hazard assessments.
Table 2: Parameters for estimating rock mass strength from field measurements. Reproduced from Selby (1980).

<table>
<thead>
<tr>
<th></th>
<th>Field Rating</th>
<th>Very Weak</th>
<th>Weak</th>
<th>Moderately Weak</th>
<th>Moderately Strong</th>
<th>Strong</th>
<th>Very Strong</th>
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</thead>
<tbody>
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<td></td>
<td></td>
<td>0.0-2.6</td>
<td>2.0-5</td>
<td>5.0-10</td>
<td>10.0-20</td>
<td>20.0-40</td>
<td>40.0-60</td>
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<td>Grain size</td>
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<td>1.25-2</td>
<td>2.0-3.0</td>
<td>3.0-4.0</td>
<td>4.0-5.0</td>
<td>5.0-6.0</td>
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<td>1.5-2.0</td>
<td>2.0-2.5</td>
<td>2.5-3.0</td>
<td>3.0-3.5</td>
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<td>0.5-1.0</td>
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<td>1.5-2.0</td>
<td>2.0-2.5</td>
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<td>Joint orientation</td>
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<td>180-270</td>
<td>270-360</td>
<td>360-450</td>
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<td>6-9</td>
<td>9-12</td>
<td>12-15</td>
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<td>10-15</td>
<td>15-20</td>
<td>20-25</td>
<td>25-30</td>
</tr>
<tr>
<td>location</td>
<td>height (m)</td>
<td>width (m)</td>
<td>w/h</td>
<td>$g$ (°)</td>
<td>$a$ (m/s^2)</td>
<td>$g$</td>
<td>Intensity (mod.merc)</td>
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<tr>
<td>------------------------------</td>
<td>------------</td>
<td>-----------</td>
<td>-----</td>
<td>---------</td>
<td>-------------</td>
<td>-----</td>
<td>---------------------</td>
</tr>
<tr>
<td>Boyd/Skeleton scarp A</td>
<td>1.48</td>
<td>0.76</td>
<td>0.52</td>
<td>27.35</td>
<td>5.07</td>
<td>0.52</td>
<td>IV</td>
</tr>
<tr>
<td>Boyd/Skeleton scarp B</td>
<td>1.50</td>
<td>1.04</td>
<td>0.69</td>
<td>34.80</td>
<td>6.81</td>
<td>0.69</td>
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<tr>
<td>Boyd/Skeleton scarp C</td>
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<td>0.41</td>
<td>0.43</td>
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<td>4.18</td>
<td>0.43</td>
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<td>Boyd/Skeleton scarp D</td>
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<td>0.40</td>
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<td>8.53</td>
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<td>6.92</td>
<td>0.71</td>
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</tr>
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<td>Boyd/Skeleton scarp G</td>
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<td>0.62</td>
<td>31.81</td>
<td>6.08</td>
<td>0.62</td>
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<td>0.43</td>
<td>0.41</td>
<td>22.08</td>
<td>3.98</td>
<td>0.41</td>
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<tr>
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<td>0.39</td>
<td>21.48</td>
<td>3.86</td>
<td>0.39</td>
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<td>Porcupine Jr. scarp A</td>
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<td>1.37</td>
<td>0.67</td>
<td>33.88</td>
<td>6.58</td>
<td>0.67</td>
<td>IV - V</td>
</tr>
</tbody>
</table>

Table 3: Dimensions of toppled columns with calculated ground acceleration necessary to topple and Modified Mercalli Intensity. $a=\tanog$ where $g=9.81 \text{ m/s}^2$. Intensity was calculated to two decimal places using equation (56). For decimal places 0.25 and lower, the intensity was rounded down. For decimal places 0.75 and higher, the intensity was rounded up. For decimal places between 0.25 – 0.75, the range of intensity was reported.
Table 4: Relationship of Modified Mercalli Intensity value at the epicenter to earthquake magnitude. Adapted from U.S. Geological Survey Earthquake Hazards Program (2006).

<table>
<thead>
<tr>
<th>Magnitude</th>
<th>Modified Mercalli Intensity</th>
</tr>
</thead>
<tbody>
<tr>
<td>1.0-3.0</td>
<td>I</td>
</tr>
<tr>
<td>3.0-3.9</td>
<td>II-III</td>
</tr>
<tr>
<td>4.0-4.9</td>
<td>IV-V</td>
</tr>
<tr>
<td>5.0-5.9</td>
<td>VI-VII</td>
</tr>
<tr>
<td>6.0-6.9</td>
<td>VII-IX</td>
</tr>
<tr>
<td>&gt;7.0</td>
<td>&gt;VIII</td>
</tr>
<tr>
<td>Magnitude</td>
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Table 5: Average number of events per year by magnitude for central Oregon and northern California between 117° - 122° W. 40° - 45° N. Data from the Advanced National Seismic System catalog (ANSS, 2006). Total number of events, N = 354.
Table 6: The number of events of a given magnitude per year in cataloged region (236,745 km²) and within 10 km² of a scarp. The number of events within 10 km of a scarp in 10,000, 500,000, and 1,000,000 years. Events within 10 km of the scarp can account for observed number of toppled columns. While there are no events any given year, there are approximately 9 events between M3.5-6 every 10 ky. Approximately 400 events every 500 ky. and approximately 850 events every 1000 ky.
Figure 19: Toppled columns at Boyd Skeleton Scarp in central Oregon. Note that vesicles increase in density toward top of column. The columns are separated at the base from other columns and lie on top of talus, implying that they toppled after erosion of the scarp had progressed. Tape measure case is 6.5 cm x 8 cm.
Figure 20: Toppled column at Porcupine Jr. scarp in northeastern California. Reflector is 1.5 m tall.
Figure 21: Boyd/Skeleton scarp in eastern Oregon. Note that the toppled columns lie on a talus slope, which suggests they toppled sometime after the scarp formed and eroded.
Figure 22: The base of basaltic columns may not be perfectly flat, especially if it represents the upper surface of an underlying flow. If the base is not flat and there are balancing points other than the edges of the column, $\alpha$ will be decreasingly smaller as the width becomes decreasingly smaller, as shown by $\hat{\alpha}$. 
Figure 2.3: Graph showing peak ground acceleration for various alpha angles. As alpha increases, it implies a larger acceleration.
Figure 24: Ground shaking probability map for the Pacific Northwest showing the study region. Approximate location of scarps shown by red lines (not to scale). [BS = Boyd/Skeleton scarp, GD = Golddigger scarp, PJ = Porcupine Jr. scarp.] Generalized using on-line custom map tool from the 2002 National Seismic Hazard Map (Frankel et. al., 2000; National Seismic Hazard Maps, 2006).
Figure 25: Graph showing frequency vs. magnitude for all earthquakes greater than M=5.0 during 1968-97 from the National Earthquake Information Center catalog. The values are shown cumulatively and as incremental values in 0.1 magnitude units. From Stein and Wysession, 2003.
Figure 26: Pacific Northwest Quaternary fault map showing the region between 40-45° N, 117-122° W for the frequency-magnitude relationship in the red box. Adapted from Pezzopane and Weldon (1993). See Figure 4 for explanation of fault colors.
Figure 27: Graph showing frequency vs. magnitude for all earthquakes greater than M = 3.0 during 1970-2005 between lat. 40-45° N, long. 117-122° W. Total number of earthquakes N = 354. Data from the Advanced National Seismic System (ANSS, 2006).
Figure 28: Graph showing the moment magnitude of earthquake possible from various scarp lengths and slip amounts common for central Oregon/northeastern California, using 1 km depth for rupture surface.

Scarp length (m) | Moment magnitude
---|---
5 m | 5.5
10 m | 5.6
20 m | 5.8
40 m | 6.0
80 m | 6.2
160 m | 6.4
320 m | 6.6
640 m | 6.8
1280 m | 7.0
Chapter 3: Geometric simulation of scarp degradation

Introduction

A model of scarp degradation in basalt could be used to estimate the ages of normal faults in many extensional environments. Previous models of scarp degradation in alluvium are inappropriate to describe the evolution of scarps in bedrock, particularly jointed bedrock such as columnar basalt (Mayer, 1984; Stewart, 1993). The purpose of this chapter is to present an idealized geometric simulation of scarp degradation in bedrock in order to improve intuition about degradation characteristics. A fully analytical model would require substantially more field measurements and observations.

Background

Bucknam and Anderson (1979) attempted to quantify scarp degradation and age with an empirical scarp-height – slope-angle model. After surveying many colluvial scarps of a given age in western Utah, they found that the slope angle is directly proportional to the logarithm of the scarp height, expressed by

\[ \theta = -8.5 \log T + 52.5 \]  

where \( \theta \) is the scarp-slope angle and \( T \) is time in years. Also, for a given height, the slope angle of the scarp decreases with age (Figure 29). The change in slope angle over a given period of time can be expressed as

\[ \Delta \theta = 8.5 (\log T_1 - \log T_2) \]  

where \( T_1 \) and \( T_2 \) are the initial and final times, respectively.

Although this model is entirely empirical, it can be used as a proxy for estimating scarp age from height and slope angle of scarps in alluvium found in similar climates to
the Utah scarps. However, Bucknam and Anderson's (1979) approach would not work for this study because scarp form, and thus, apparent degradation, varies within a single scarp.

Nash (1980) further refines the scarp degradation-age relationship with a model in which the rate of change of elevation on a slope is proportional to the curvature of the profile at that same point. Quantitatively, his model is expressed as

\[ \frac{\partial y}{\partial t} = c \frac{\partial^2 y}{\partial x^2} \]  

(12)

where y and x are the coordinates of a point on the profile, t is time in years and c is the degradation rate.

According to this diffusion-type model, where the profile of the scarp is concave up (positive curvature), it will increase in elevation with time; where the profile is convex up (negative curvature), it will decrease in elevation with time. This model allows for variations in degradation on the same scarp, while Bucknam and Anderson (1979) assume that profiles are smooth and degradation is constant over the entire scarp.

These models are relevant to scarps formed in alluvium. There have been very few models of scarp degradation in bedrock. Bedrock-scarp degradation models primarily have been qualitative or empirical (e.g. Wallace, 1977; Sonmez et al., 1998). Although scarps in bedrock are similar in form to scarps formed in alluvium, erosional processes on bedrock and alluvium differ (Nash 1981). Since slope-replacement takes much longer in bedrock, the free face and talus slope are present for longer periods of time. A wash slope may never develop due to lack of sediment to wash down the scarp face (McCalpin 1996). Thus, a diffusion-type degradation model may be inappropriate for bedrock-scarps (Mayer, 1984; Avouac, 1993).
Geometric simulation of bedrock scarp retreat

I have produced a geometric model (or simulation) for bedrock-scarp retreat. It recreates the scarp profile shapes and may be useful to improve understanding of the processes involved. Eventually, such an exercise may lead to a fully analytical model of bedrock scarp degradation through time. This initial simulation addresses only the gravity controlled stage of scarp degradation (Figure 2A-C). Debris slope and wash slope processes are not included.

Nash (1981) produced a FORTRAN model of scarp retreat in bedrock called FAULT. I have recreated this model in a more user-friendly format using Matlab (The Mathworks, Inc., 2006). This recreated simulation allows users to model scarp retreat with flexible retreat increments, various initial scarp angles, and various angles of repose for talus. In the future, it possibly could be modified to include variable bedrock characteristics.

A diagram of a scarp can be thought of as four distinct and connected lines: a base or hanging wall, a talus slope, a free face, and a top or footwall (Figure 30). There are two primary angles relevant to the scarp: the initial angle of the scarp, which is equivalent to the fault dip, designated \( \theta \); and the angle of repose for the eroded bedrock (talus pile), designated \( \alpha \). I use angle of repose values more closely reflective of the angles observed in the field (4°-15°) rather than higher values suggested by Selby (1993).

These lines are defined on Matlab using the three points of intersection between the lines. \textit{Toe} is the intersection point between the hanging wall and the talus slope; \textit{top} is the point between the talus slope and the free face; and \textit{crest} is the point between the
free face and the footwall. These intersection points are defined using trigonometry and assuming a maximum height of the footwall at 1 unit. The points mark the ends of the lines, with the elevation along the lines determined through trigonometry (Figure 30).

Retreat of the scarp can be thought of as an area being removed from the free face and transferred to the talus slope, with an overall conservation of area (two dimensional “volume”). The initial retreat can be considered as the entire fresh fault face retreating with that entire area being formed into a talus pile. In successive iterations of retreat, only the remaining free face retreats, with the area being added to the talus pile. With additional intervals of scarp retreat, the talus pile increases in size until the entire front of the scarp is a talus pile and at the angle of repose (Figure 31).

The slopes of the lines do not change with retreat, but the positions of the intersection points change as the talus slope gets larger and the free face retreats. Toe moves horizontally toward 0 (the front of the model); top moves vertically closer to 1 (the height of the scarp) and horizontally closer to crest; crest moves horizontally away from 0 (toward the back of the model) until top and crest are the same point.

To account for the role of columnar jointing in the shape of the scarp as it degrades I adjusted the basic simulation script. The initial scarp face represents the fault and is at angle \( \theta \). The first increment of degradation occurs when a column topples. The columns are block shaped, so the new free face has an angle of 90° (Figure 32).

There are several assumptions inherent for the model to work: 1) all material eroded during retreat must be added to the talus pile so that no material is carried away or out of the model; 2) after the initial retreat, all material being removed from the scarp is removed only from the free face; 3) after the initial scarp forming event, there are no
further events which would increase the height of the free face or change the angle theta.
The full Matlab script is reproduced in Appendix B.

Discussion of model

The simulation shows the gradual degradation of a scarp in bedrock and is similar in shape to transect profiles. The simulation starts out entirely free face, as the scarps in Figure 6c. As the scarp degrades by increments, the talus pile increases, the free face decreases and the profiles begin to look more like the scarps in the debris-controlled stage of erosion (Figure 6b). The basic simulation provides users the opportunity to adjust fault angle, angle of repose, fault height and retreat increment.

Previous models for scarp degradation in both alluvium and fractured bedrock have assumed a continuous degradation rate that is consistent along and across the entire scarp (Wallace, 1977; Bucknam and Anderson, 1979; Nash, 1980, 1981; Arrowsmith et al., 1996;). These assumptions are not appropriate for scarps in jointed basalt because joint characteristics influence the degree of degradation. Since joint characteristics vary along a scarp, degradation rates also vary along a scarp.

The simulation I produced addresses the role of joints with a variable retreat increment. Assuming that the topple of columns is the primary process at the gravity-controlled stage, the retreat increment can be set to the column diameter. I varied the retreat increment to produce two different simulations (Figure 33). Using a simulation scarp height of 1, I gave the first simulation a retreat increment of 0.1 which would be equivalent to 1 m joint spacing in a 10 m high scarp. The second simulation had a retreat increment of 0.04 which would be equivalent to a 0.4 m joint spacing in a 10 m high
scarp. Within ten iterations, the model with the larger increment degrades further, as
would be expected, although both simulations show a similar profile shape throughout.
The model assumes both simulations degrade at the same rate. Field observations and the
statistical analysis both suggest the joint size affects the rate of degradation. Earthquake
frequency-magnitude estimates suggest it will take approximately 10X as many years to
experience a sufficient magnitude from a local event to topple a 1.0 m column than a
0.40m column (Figure 27).

As the scarp degrades, the talus pile grows and the free face height decreases. As
the free face height decreases, the height of column that is available to topple decreases,
as well. Shorter columns are more stable than taller columns and take a larger amount of
force to topple. I added a formula to calculate the quasi-static acceleration necessary to
topple a column of free face height and increment width at each degradation iteration.
The acceleration necessary for toppling the columns increases as the free face, and hence,
the column height, decreases (Figure 34). Larger accelerations correspond to more
intense ground shaking from larger magnitude events which occur less frequently than
smaller magnitude events (Figures 25, 27). As columns become more stable (decreasing
height or increasing base), the acceleration necessary to topple increases, and the rate of
toppling decreases. Thus, the rate of scarp degradation due to ground motion should
decrease through time. This may account for the persistence of scarps with large or
equant columns.

Additional work to refine the simulation is needed so that the model more closely
resembles nature. Simulating more than one joint set, a variable climate, and multiple
faulting events could refine understanding of the form of bedrock scarps and how they
Conclusion

A geometric simulation of scarp degradation produces profiles similar to those observed in the field at normal fault scarps in jointed basalt. Variable initial scarp angles, talus repose angles, scarp heights, and retreat increments can be used with the simulation. The retreat increment represents the joint spacing for scarps in jointed basalt, and the simulation shows that larger retreat increments produces overall greater retreat. The simulation doesn’t show variable degradation rates. A version of the simulation shows the acceleration necessary to topple columns with spacing shown by retreat increment and height represented by scarp free face. The columns become more equant with time, needing larger accelerations to topple. If ground acceleration plays a primary role in column topple, then degradation rates may vary with time and overall scarp degradation.

A more comprehensive census of scarps and scarp characteristics would enhance the simulation by providing a larger database of profiles for validation, as well as a more complete understanding of degradation processes at work on normal fault scarps in basalt.
Figure 29: (A) Slope angle vs. scarp height for selected scarps in Utah. The slope angle is proportional to the scarp height. (B) Decrease of slope angle – scarp height with time for three scarps. Fish Springs scarp is the youngest and Panguitch scarp is the oldest. From Bucknam and Anderson (1979).
Figure 30: Geometry and trigonometric functions related to the scarp shape and degradation.
Figure 31: Scarp retreat process using degradation simulation. (A) shows scarp after formation. (B) shows scarp and talus slope with first retreat. (C) shows scarp and talus after several iterations of retreat. Notice that the crest moves to the right, the toe moves left, and the free face becomes smaller with each iteration. Horizontal distance and height are in arbitrary units simply to show scarp shape through the degradation simulation.
Figure 32: Simulation of 90° free face occurring when basalt columns topple. Horizontal distance and height in arbitrary units.
Figure 33: (A) shows degradation of a scarp of unit height 1 with retreat increment 0.1 after ten iterations. (B) shows degradation of a scarp of unit height 1 with retreat increment 0.04 after ten iterations. The simulation with larger retreat increment has degraded further for the given iterations, but the simulation doesn’t take into account degradation rate differences with changes in retreat increment or joint spacing. Horizontal distance and height in arbitrary units.
Figure 34: As the simulation free face decreases in height, the column height above talus decreases and the acceleration necessary to topple the column increases. Example using column width of 1.0m.
Appendix A: Scarp descriptions

*Barr Road Scarp*

I surveyed four transects at Barr Road scarp (Figures A-1, A-2, A-3). The first profile (BR1, Figure A-3a) shows the northerly-most transect along a soil-mantled talus slope. It is 102 m long with 9.75 m of relief. The entire transect line is smooth with no free face.

The second transect (BR2, Figure A-3b) is approximately 25 m south of BR1. It is 50 m long with a little over 7 m of relief. There are a few boulders on this slope although the slope is all soil-mantled until 25 m from the start of the transect. There is a free face 30 m from the transect start that crests at 4 m. The upper surface slopes steeply from the free face until it flattens at 7 m height.

The third transect (BR3, Figure A-3c) is about 20 m south of BR2 and is 58 m long with 7.75 m relief. At 35 m from the start of the transect the talus slope ends at 3.25 m height. The first sharp change in slope represents a large basalt boulder. At 4.5 m height, 40 m from the transect start is a free face which crests at 6.5 m; the upper surface of the scarp slopes steeply until 7.75 m height when it flattens.

The fourth transect (BR4, Figure A-3d) is 15-20 m south of BR3 and is about 50 m long with 6.8 m relief. The talus slope ends 20 m from the start of the transect; the first change in slope represents a large basalt boulder. The wash-slope/talus-slope continues to 30 m from the start of the transect where the free face begins, cresting at 6.5 m. The upper surface slopes to 6.8 m before flattening.
Boyd/Skeleton Scarp

I surveyed five transects across Boyd/Skeleton scarp (Figures A-4, A-5, A-6). The first profile (BS1, Figure A-6a) shows the most easterly transect on this scarp and is approximately 120 m long. There is a shallow depression at the base of the scarp which grades into a wash-slope until 3.75 m height, approximately 60 m from the transect start. There is a toppled column which creates an almost vertical slope to 6 m, when the slope shallows to a crest at 7 m. There is a trough behind the topple at 6.5 m height which abruptly steepens to a free face that crests at 9.5 m, 90 m from the transect start. Directly behind this crest is a trough with a juniper tree growing in it. The upper surface behind the crest and trough has a gentle slope to about 11 m height.

The second transect (BS2, Figure A-6b) is about 100 m northwest of BS1 and is about 110 m long with 11 m relief. Again, there is a shallow depression at the base of the scarp, but at this transect there is no wash-slope. The slope steepens from 0 m to 6.75 m in 20 m; the columns are horizontal and plunge perpendicular to the scarp at the base of this slope and are vertical at the crest. There is a 20 m wide bench directly behind this crest which steepens into a free face from 6 to 11 m high. The upper surface behind the crest of the free face is horizontal.

Once again, there is a shallow depression at the base of the scarp in transect three (BS3, Figure A-6c), which is about 30 m northwest of BS2, is approximately 90 m long, and has 11 m relief. From the depression, the slope rises fairly steeply to 5 m height where it steepens into a free face 70 m from the transect start. The slope shallows about a meter further north and the finally crests at 11 m.
Transect four (BS4, Figure A-6d) is about 20 m northwest of BS3 and stretches 70 m with 10.5 m of relief. The slope is smooth from the start of the transect to 50 m, where the free face starts at a little over 7 m height. The free face crests at a little over 10 m and the upper surface gently slopes to 10.5 m until it flattens.

The final transect (BS5, Figure A-6e) is 30 m northwest of BS4 and is 80 m long with 11 m of relief. There is a shallow depression at the base of the scarp which grades into a smooth slope that rises to a little over 8 m height at 80 m from the transect start. The slope steepens to a free face which crests at just under 11 m. The upper surface gently slopes to 11 m before flattening.

**Big Hole Scarp**

I ran three transects on this scarp (Figures A-7, A-8, A-9). The first, southernmost transect (BH1, Figure A-9a) is approximately 90 meters long and has approximately 10.5 meters of relief. It is a relatively smooth profile with a steepening in slope at 2 m and 8.5 m in height.

The second transect (BH2, Figure A-9b) is about 60 m long, with 10.5-10.75 m of relief. A wash-slope stretches approximately 25 m from the start of the transect to a height of 1.8 m; there is a sharp break in slope and the profile steepens into a free face which breaks again at a little over 6 m in height. The profile slope shallows a little and elevation drops at 35 m from the transect start at a height of 7 m. There is a trough approximately 7-8 m across directly behind the first free face. At approximately 48-50 m from the transect starting point the slope once again steepens to a second free face, which ends on the upper surface of the scarp at 10.75 m height.
The third transect (OF3, Figure A-9c) is the most northerly transect for Big Hole Scarp and is about 50 m long, with 12 m of relief. The wash-slope extends 30 m from the transect start to a height of about 2.25 m where the slope steepens sharply to a rounded face with a 1:2 slope which ends in a free face at a height of 8 m, about 40 m from the transect start. The free face ends at about 11 m when the slope shallows until the scarp crest at 12 m height.

**Oatman Flat Scarp**

I ran five transects along this fault, from south to north (Figures A-10, A-11, A-12). The first profile (OF1, Figure A-12a) shows the southern-most transect and is 43 m long with 10.5 m of relief. The initial slope is smooth and represents the wash slope on the west side of the scarp. At approximately 30 m from the start of the transect and at 6 m height the free face begins, cresting at about 12.5 m. The upper surface of the scarp then curves and slopes up to about 13 m when it flattens and then begins to lose elevation.

The second transect (OF2, Figure A-12b) is approximately 50-75 m north of OF1, is a little over 55 m long and has just over 10 m of relief. The wash slope ends approximately 22 m from the start of the transect and at a height of 5 m. The talus slope ends 36 m from the start of the transect at a height of about 9 m where the base of the free face begins. The free face crests at just at approximately 13 m. There is a shallow depression directly behind the crest on the upper surface before the slope gently rises and flattens at just under 13 m height.
The third transect (OF3, Figure A-12c) is about 100 m north of OF2, is 44 m long and has 10.5 m of relief. The wash slope ends approximately 16 m from the start of the transect, at a height of 4 m. The talus slope ends approximately 30 m from the start of the transect at a height of 7 m. The free face crests a little over 11 m and there is a gradual increase in slope on the foot wall to 13.6 m.

The fourth transect (OF4, Figure A-12d) is about 75 m north of OF3, is 49 m long and has 10.7 m of relief. The wash slope ends approximately 23 m from the start of the transect, at a height of 5.1 m. There is a large block of basalt which is partially separated from the scarp face, creating a small chasm behind the block. The block is approximately 2 m wide and peaks at a height of 9.1 m. The free face begins about 34 m from the start of the transect, cresting at 12.2 m height. The transect ends when the slope flattens at about 13.7 m height.

The fifth transect (OF5, Figure A-12e) is 50 m north of OF4, is 56 m long with 10.1 m of relief. There is no free face. The transect starts at a height of 3.7 m and gradually increases to 13.8 m height over a distance of 56 m.

**Golddigger Scarp**

I surveyed six profiles at Golddigger scarp (Figures A-13, A-14, A-15). The first profile represents the most southerly transect along the northwest striking scarp. All the transects are between 50 and 200 m apart. Transect one (GD1, Figure A-15a) is 90 m long with almost 16 m of relief. There is a shallow depression at the base of the scarp; the profile then rises in a smooth slope to 16 m height. This transect consisted of a talus pile with 0.5-1 m wide toppled columns.
The second transect (GD2, Figure A-15b) is 65 m long with 11.5 m of relief. There is also a shallow depression at the base of the scarp. The profile steepens to a fairly smooth slope approximately 30 m from the beginning of the transect. The profile has some roughness resulting from large toppled columns but none of the sharp changes in slope represent a free face.

Transect GD3 (Figure A-15c) is 115 m long with 18.25 m of relief. The slope is smooth along the entire profile with the few spots of roughness representing toppled columns.

Transect GD4 (Figure A-15d) is a little over 70 m long with 14.25 m of relief. A wash-slope can be interpreted from the shallow slope approximately 17 m from the start of the transect. At 23 m from the start the profile changes slope rapidly, representing toppled columns. At 5 m height and 40 m from the start of the transect the slope smooths and steepens to almost 1:1 slope. At 10 m height and 50 m from the start of the transect the slope abruptly changes to a free face that crests at 14 m. There is a depression resulting from a toppled column at the crest and the upper surface gently slopes to 14.25 m height.

The fifth transect (GD5, Figure A-15e) is 115 m long with 16.2 m relief. At the scarp base, 40 m from the start of the transect is a large toppled column which crests at 1.2 m. The scarp slope is fairly smooth to 16.2 m height and represents a talus slope.

Transect GD6 (Figure A-15f) is 30 m long and has 3.75 m relief. This part of the scarp is a monoclinal structure with no free face. The entire profile is smooth.
Porcupine Jr. Scarp

I surveyed 5 transects across Porcupine Jr. scarp (Figures A-16, A-17, A-18). The first profile (PJ1, Figure A-18a) shows the northerly most transect on this scarp. It is a little over 70 m long with 7.75 m relief. The entire transect is over a talus slope; there is no free face.

Transect PJ2 (Figure A-18b) is about 10 m south of the first, is about 50 m long and with just under 8 m relief. There is a shallow depression at the base of the scarp with a wash-slope rising half a meter, 30 meters from the transect start. The slope steepens until 40 m from the start at a height of 3 m at the base of the free face. The free face crests at 7.25 m with the upper surface sloping gently to 7.75 m when it flattens.

Transect PJ3 (Figure A-18c) is 10 m south of PJ2 and is a little over 80 m long with 8.3 m relief. There is no slope until 43 m from the start of the transect where the free face rises to a height of 3.5 m where the slope shallows and broadens to a shelf approximately 8-10 m wide. A second free face rises at 50 m from the start to a crest at 7.75 m height. The upper surface gently slopes to 8.3 m before flattening.

The fourth transect (PJ4, Figure A-18d), 10 m south of PJ3, has no wash or talus-slope. The profile is flat until about 47 m from the transect start where it abruptly rises in a free face until it crests at 7.75 m height. The upper surface gently slopes to 8.3 m height when it flattens, approximately 70 m from the transect start.

The final transect (PJ5, Figure A-18e) is 10 m south of PJ4, is approximately 60 m long with 8.75 m relief. There is a broad slope beginning 8 m from the start and half a meter high until a topple 30 m from the start and cresting just above 2 m. The edge of the toppled column shallows to under 2 m about 6 m from the crest of the topple where it
 touches the edge of a bench 10 m wide and with a gentle slope. The base of the free face is about 48 m from the start and crests at 8.25 m. The upper face gradually rises to 8.75 m before it flattens.

Observations from all transects are summarized in Table A-1.
Table A-1: Scarp characteristics for transects at studied scarps.

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Figure A-1: Barr Road scarp in red, central Oregon. It can be accessed by turning south on Barr Road off the Redmond-Sisters Hwy 126, at 121°20'38"W, 44°10'40"N. Blue shows traces of other local scarps. Box shows transect region and location of Barr Road photo mosaic from Figure A-2. Transect location on photo mosaic. Portion of aerial photo 7096-95, 5/8/1994 from USDA Forest Service.
northeast.

part of the Sherry Fault Zone and faults uninformed basalt. The rod is 1.0 m high at the center of the reflector. Snap ends 3.2 m and bases

Figure A.2: Photo mosaic of Bank Road Scarp in central Oregon. Lower photos are a continuation of N side of upper photos. The scarp is
Figure A-3: Scarp profiles for Barr Road Scarp, central Oregon. Profiles start at theodolite and all heights are measured with regard to theodolite elevation. Each profile has 2X vertical exaggeration with east to the left. All subsequent profiles are presented according to the same coordinates, with facing directions represented accordingly.
Figure A-3 continued.
Figure A-4: Boyd/Skeleton scarp, central Oregon. It can be accessed by turning east off Oregon SR97 onto China Hat Road/FS road #18, then turning north at the sign for Skeleton Cave, at 121°09'42"W, 43°55'40"N. Box shows transect region and location of photo mosaic from Figure A-5. Transect location on photo mosaic. Portion of aerial photo 7095-167, 5/8/1994 from USDA Forest Service.
Figure A-5: Photo mosaic of Boy/Skidden Scarp in central Oregon. The lower photos are a continuation of the 5 edge of the upper photos. The scarp is part of the Sisters Fault Zone. Field assaull is approximately 1.6 m high. Scarp trends NW and faces southwest.
Figure A-6: Scarp profiles for Boyd/Skeleton Scarp, central Oregon. All profiles have 2X vertical exaggeration with west on the left.
Figure A-6 continued.
Figure A-7: Big Hole scarp in central Oregon. It is located within the Big Hole 7.5' quadrangle in central Oregon, north off OR 47 on FS Road #4805, at 12115' W, 4327' N. Box shows transect area and location of photo mosaic from Figure A-8. Aerial photographs from TerraServer, [http://terraserver-usa.com](http://terraserver-usa.com) by entering latitude/longitude data into database search.
Figure A-8: Photo mosaic of Big Hole Scarp in central Oregon. Field assistance is approximately 0.6 in high.
Figure A-9: Scarp profiles for Big Hole Scarp, central Oregon. Each profile has 2X vertical exaggeration with southeast to the left.
Figure A-10: Oatman Flat scarp and larger, graben-bounding scarp in red, central Oregon. It can be accessed by turning north off Oregon SR297 onto an unmarked gravel road approximately 10 miles west of Silver Lake at 121°09'45"W, 43°06'55"N. Blue shows traces of other local scarps in Antelope Mountain fault zone. Box shows transect region and location of photo mosaic from Figure A-11. Transect locations on mosaic. Portion of aerial photo 12251-29, 7/25/2000 from USDA Forest Service.
center of reflection

Figure A-1: Photo mosaic of Oatman Peak Sheep in central Oregon. Lower photo starts at SE edge of upper photo. Road is 1.6 m high in 10
Figure A-12: Scarp profiles for Oatman Flat Scarp, central Oregon. Each profile has 2X vertical exaggeration, with southeast to the left.
Figure A-12 continued.
Figure A-13: Golddigger scarp, Lava Beds N.M., northern California. The fault may be accessed by turning west onto road # 46N21 for the western boundary road of Lava Beds National Monument, at 121°32'19"W, 41°49'20"N. Box shows transect region and location of photo mosaic from Figure A-14. Transect location on mosaic. Portion of aerial photo 6216-18, 7/30/1993 from USDA Forest Service.
Figure A-14: Photo mosaic of Golddigger scarp in northern California. Middle photo starts at northern edge of upper photo. Lower photo starts at northern edge of second photo. Red box indicates assistant and surveying rod, which is 1.6 m high at center of reflector.
Figure A-15: Scarp profiles for Golddigger Scarp, northern California. Each profile has 2X vertical exaggeration, with east to the left of the page.
Figure A-15 continued.
Figure A-16: Porcupine Jr. Scarp, Devil's Garden, Modoc Plateau, northern California. Porcupine Jr. scarp can be accessed by turning north off California SR299 to Alturas at road # 73 toward Goose Lake then west on road 43N18 to road 43N13 south into Porcupine Valley at 120°43'45"W, 41°36'13". The scarp is approximately 10 m high. Pink box shows area of transect surveys and location of photo mosaic from Figure A-17. Yellow scarp trace represents the larger Porcupine Valley scarp, approximately 50 m high. Portion of aerial photo 10497-232, 8/3/1998 from USDA Forest Service.
Figure A-17: Close-up image of the saline ponding at the vent, showing the chemical composition of the pond water.
Figure A-18: Scarp profiles for Porcupine Jr. Scarp, northern California. Each profile has 2X vertical exaggeration with west on the left.
Figure A-18 continued.
Appendix B: Scarp degradation simulation script for Matlab

Basic simulation

clear all
theta=40;
theta=theta*pi/180;%slope of fault and free face
alpha=20;
alpha=alpha*pi/180; % angle of repose of talus
beta=(pi-theta);
epsilon=(pi-alpha-beta);
gamma=((pi/2) -theta);

bp=l ; % retreat increment

h=l; % scarp height

base=2; % horizontal position of the base of scarp (no talus)

last=7; % horizontal position of the last point plotted

% initial scarp after formation without talus%

crest=[base+(h/(tan(theta)))];% first point on foot wall, horizontal position of top of scarp

x = [0, base, crest, last];
z = [0, 0, h, h];

plot (x, z, 'r')
axis equal
pause

%% calulations relevant to first talus slope%%

crest=[crest+bp];% horizontal position of top of scarp through degradation

baset=[base+bp]; % theoretical base of free face

Ap=(bp)*(h); % area removed from first scarp retreat

hyp=[sqrt((Ap*2*sin(beta))/(sin(alpha)*sin(epsilon)))]; % slope distance of first talus

top=[sin(alpha)*hyp]; % height of top of first talus

x=(top)/tan(theta); % horizontal increment for top of first talus from base of bedrock

btx=top/tan(alpha); % horizontal distance from first talus toe to talus top

btm=[(sin(epsilon))/hyp]; % base length of first talus


toe=base-btm-bp; % toe of first talus

topx=[baset + x]; % x position for top of first talus

x2=[0, toe, topx, crest, last];
z2=[0, 0, top, h, h];

plot(x2, z2, 'g')
axis equal
pause

%% calcs for later retreats%%
topb=top;
for i=1:1:5;
    crest=[crest+bp]; %horizontal position of top of scarp through degradation
    baset=[baset+bp]; %theoretical base of free face
    hf=h-topb;
    Apb=bp*hf; %area removed from scarp retreat
    Atotal=Ap+Apb;
    hypo=[sqrt((Atotal*2*sin(pi-theta))/(sin(alpha)*sin(pi-alpha-(pi-theta))))]; %slope
distance of talus
    topb=[sin(alpha)*(hypo)]; % height of top of first talus
    x=(topb)/tan(theta); % horizontal increment for top of first talus from base of bedrock
    btx=topb/tan(alpha); % horizontal distance from first talus toe to talus top
    %btm=[(sin(epsilon))/hyp]; % base length of first talus
    b=btx-x; %base length of talus
    toe=baset-b; % toe of first talus
    topx=[baset+x]; % x position for top of first talus
    Ap=Atotal;
end
x3=[0, toe, topx, crest, last];
z3=[0, 0, topb, h, h];

plot(x3, z3, 'bo-')
axis equal;
pause
end
**Simulation with a free face at 90°**

```matlab
clear all
theta=40;
theta=theta*pi/180; % slope of fault and free face
alpha=20;
alpha=alpha*pi/180; % angle of repose of talus
beta=(pi-theta);
epsilon=(pi-alpha-beta);
gamma=((pi/2)-theta);
bp=.1; % retreat increment
h=1; % scarp height
base=2; % horizontal position of the base of scarp (no talus)
last=7; % horizontal position of the last point plotted

% initial scarp after formation without talus%
crest=[base+(h/(tan(theta)))] ; % first point on foot wall, horizontal position of top of scarp
x = [0, base, crest, last];
z = [0, 0, h, h];

plot(x, z, 'r')
axis equal
pause

%% calculations relevant to first talus slope%%
crest=[crest+bp]; % horizontal position of top of scarp through degradation
baset=[base+bp]; % theoretical base of free face
Ap=(bp)*(h); % area removed from first scarp retreat
hyp=[sqrt((Ap*2*sin(beta))/(sin(alpha)*sin(epsilon)))] ; % slope distance of first talus
top=[sin(alpha)*hyp]; % height of top of first talus
x=(top)/tan(theta); % horizontal increment for top of first talus from base of bedrock
btx=top/tan(alpha); % horizontal distance from first talus toe to talus top
btm=[sin(epsilon)/hyp]; % base length of first talus
toe=base-btm-bp; % toe of first talus
topx=crest; % x position for top of first talus

x2=[0, toe, topx, crest, last];
z2=[0, 0, top, h, h];

plot(x2, z2, 'g')
axis equal
pause

%% calcs for later retreats%%
topb=top;
```
for i=1:1:10;
    crest=[crest+bp]; % horizontal position of top of scarp through degradation
    baset=[baset+bp]; % theoretical base of free face
    hf=h-topb;
    Apb=bp*hf; % area removed from scarp retreat
    Atotal=Ap+Apb;
    hypo=[sqrt((Atotal*2*sin(pi-theta))/(sin(alpha)*sin(pi-alpha-(pi-theta))))]; % slope
    distance of talus
    topb=[sin(alpha)*(hypo)]; % height of top of first talus
    x=(topb)/tan(theta); % horizontal increment for top of first talus from base of bedrock
    btx=topb/tan(alpha); % horizontal distance from first talus toe to talus top
    %btm=[(sin(epsilon))/hyp]; % base length of first talus
    b=btx-x; % base length of talus
    toe=base-b; % toe of first talus
    topx=crest; % x position for top of first talus
    Ap=Atotal;

    x3=[0, toe, topx, crest, last];
    z3=[0, 0, topb, h, h];

    plot(x3, z3, 'bo-')
    axis equal;
    pause
end
Simulation with toppling acceleration calculation

%simulation with toppling acceleration
clear all
theta=40;
theta=theta*pi/180; % slope of fault and free face
alpha=20;
alpha=alpha*pi/180; % angle of repose of talus
beta=(pi-theta);
epsilon=(pi-alpha-beta);
gamma=((pi/2)-theta);
bp=.1; % retreat increment
h=1; % scarp height
base=2; % horizontal position of the base of scarp (no talus)
last=7; % horizontal position of the last point plotted
g=9.8

%initial scarp after formation without talus%
crest=[base+(h/(tan(theta)))] % first point on foot wall, horizontal position of top of scarp
x = [0, base, crest, last];
z = [0, 0, h, h];

plot (x, z, 'r')
axis equal
pause

%%% calulations relevant to first talus slope %%
crest=[crest+bp]; % horizontal position of top of scarp through degradation
baset=[base+bp]; % theoretical base of free face
Ap=(bp)*(h); % area removed from first scarp retreat
hyp=[sqrt((Ap*2*sin(beta))/(sin(alpha)*sin(epsilon)))] % slope distance of first talus
top=[sin(alpha)*hyp]; % height of top of first talus
x=(top)/tan(theta); % horizontal increment for top of first talus from base of bedrock
btx=top/tan(alpha); % horizontal distance from first talus toe to talus top
btm=[(sin(epsilon))/hyp]; % base length of first talus
toe=base-btm-bp; % toe of first talus
topx=crest; % x position for top of first talus

x2=[0, toe, topx, crest, last];
z2=[0, 0, top, h, h];

plot(x2, z2, 'g')
axis equal
pause

%%% calcs for later retreats%%%
topb=top;
for i=1:1:10;
crest=[crest+bp]; %horizontal position of top of scarp through degradation
baset=[baset+bp]; %theoretical base of free face
hf=h-topb;
Apb=bp*hf; %area removed from scarp retreat
Atotal=Ap+Apb;
hypo=[sqrt((Atotal*2*sin(pi-theta))/(sin(alpha)*sin(pi-alpha-(pi-theta))))]; %slope
distance of talus
topb=[sin(alpha)*(hypo)]; % height of top of first talus
x=(topb)/tan(theta); % horizontal increment for top of first talus from base of bedrock
btx=topb/tan(alpha); % horizontal distance from first talus toe to talus top
b=btx-x; %base length of talus
toe=base-b; % toe of first talus
topx=crest; % x position for top of first talus
ja=bp/hf;
accel(i)=ja*g
Ap=Atotal;

x3=[0, toe, topx, crest, last];
z3=[0, 0, topb, h, h];
display accel;

plot(x3, z3, 'bo-')
axis equal;
pause
end

plot(accel)
Works Cited


