Detection of a Landslide Glide Plane Using Seismic Reflection Methods: Investigation at Little Valley Landslide in Draper, Utah

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DETECTION OF A PRE-HISTORIC LANDSLIDE GLIDE SURFACE USING
SEISMIC REFLECTION METHODS AT LITTLE VALLEY LANDSLIDE IN DRAPER, UTAH

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ABSTRACT

DETECTION OF A PRE-HISTORIC LANDSLIDE GLIDE SURFACE USING SEISMIC REFLECTION METHODS AT LITTLE VALLEY LANDSLIDE IN DRAPER, UTAH

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An integration of geological and geophysical techniques has been used to characterize the internal structure of the Little Valley Landslide in Draper, Utah, USA. The Little Valley Landslide is a pre-historic landslide as old as 13ka B.P. It is found to consist of chaotic and disturbed weathered volcanic units derived from Tertiary age volcanics that comprise a great portion of the Wasatch Range. Geotechnical investigations that were integrated with the geophysical results included excavation of trenches and drilling of boreholes. Geophysical methods, in particular high-resolution seismic data, were used to provide a framework for interpreting the geotechnical observations. High-resolution seismic reflection data, seldom used in landslide investigations, were acquired and processed in order to image the basal or glide surface
of the landslide and the structure underlying the landslide. The integration of the geotechnical and geophysical investigations provided a better understanding of the geometry of a portion of the Little Valley Landslide. Trenching and drilling identified landslide material in the subsurface. The high-resolution seismic reflection data imaged the glide surface with the onset of coherent reflectivity.

A decollement or glide surface underlies the landslide indicating a large mass movement. The glide surface is observed on the seismic reflection profiles to be deepest in the center portion of the landslide. It is observed in the seismic reflection images to shallow up slope and creating a trough-like shape feature. A contour map modeling the middle of the Little Valley Landslide is derived from the seismic data. This study shows that seismic reflection techniques can be successfully used in complex alpine landslide regions. They are also efficient and cost-effective tools when compared to trenching and drilling investigations. The seismic data can (1) provide a framework to link geological data and (2) take the place of an extensive trenching and drilling program.
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# Table of Contents

Abstract ........................................................................................................................................ iv  

Table of Contents ............................................................................................................................ vii 

Table of Figures ............................................................................................................................... viii  

Introduction ................................................................................................................................... 1  

Geologic Setting .............................................................................................................................. 3  

IGES Investigations ...................................................................................................................... 5  
   - Trench Investigation ............................................................................................................. 6  
   - Radiocarbon age dating ........................................................................................................ 8  
   - Borehole 1 and Gamma log .................................................................................................. 9  

Location of Seismic Reflection Profiles ...................................................................................... 10  

Seismic Reflection Methods/Processing ........................................................................................ 11  

Observations of the profiles ......................................................................................................... 15  
   - Seismic Profile 1 .................................................................................................................. 15  
   - Seismic Profile 2 .................................................................................................................. 17  
   - Seismic Profile 3 .................................................................................................................. 17  
   - Seismic Profile 4 .................................................................................................................. 19  
   - Seismic Profile 5 .................................................................................................................. 19  
   - Geological Interpretation of the Results of the Seismic Surveys ....................................... 20  

Discussion .................................................................................................................................... 22  

Conclusions ................................................................................................................................... 25  

References ..................................................................................................................................... 27  

Figures .......................................................................................................................................... 31
# Table of Figures

Figure 1 General cross section of landslide showing toe, head, and basal plane of a landslide. ................................................................. 31

Figure 2 Index map of the Little Valley Landslide location with surrounding Salt Lake Valley, Wasatch Range, and Wasatch Fault................. 32

Figure 3 Topographic map showing boundaries of the Little Valley Landslide.......................................................................................... 33

Figure 4 Geologic map of the Little Valley Landslide showing areas investigated: toe, middle, and sag pond areas.................................................. 34

Figure 5 Enlarged view of the trench locations at the mapped toe of the Little Valley Landslide. ........................................................................ 35

Figure 6 Investigation areas in the middle portion of the Little Valley Landslide on geologic map. ................................................................. 36

Figure 7 Trench location at the head of the Little Valley Landslide near the sag pond ........................................................................... 37

Figure 8 Representative log of Trench 1 in the middle of the landslide showing landslide mass, shear zone, and volcanic block. ......................... 38

Figure 9 Representative log of Trench 2 in the middle of the landslide showing layered alluvial and colluvial deposits. ........................................... 39

Figure 10 Photograph of road cut wall showing shear contact between volcanic bedrock and landslide material......................................................... 40

Figure 11 Road cut wall photograph showing shearing and thrusting within the landslide mass. ......................................................................... 41

Figure 12 Borehole-1 Log......................................................................................................................................................... 42

Figure 13 Gamma Ray Log Borehole 1 ................................................................................................................................................... 43

Figure 14 Raw shot record with no filters (on left) and filtered shot record with AGC, Ormsby bandpass filter (50-80-240-400), and refraction statics correction of Profile 1. ................................................................. 44
Figure 15 Raw shot record with no filters (on left) and filtered shot record with AGC, Ormsby bandpass filter (50-80-240-400), and refraction statics correction of Profile 2. ................................................................. 45

Figure 16 Raw shot record with no filters (on left) and filtered shot record with AGC, Ormsby bandpass filter (50-80-240-400), and refraction statics correction of profile 3. ................................................................. 46

Figure 17 Raw shot record with no filters (on left) and filtered shot record with AGC, Ormsby bandpass filter (50-80-240-400), and refraction statics correction of Profile 4. ................................................................. 47

Figure 18 Raw shot record with no filters (on left) and filtered shot record with AGC, Ormsby bandpass filter (50-80-240-400), and refraction statics correction of Profile 5. ................................................................. 48

Figure 19 Stacks of Profile 1 with varying top mutes from 1st to 3rd zero crossing. .................................................................................................................... 49

Figure 20 Filter test........................................................................................................... 50

Figure 21 Deconvolution Test. ......................................................................................... 51

Figure 22 Seismic reflection profile 1 with interpreted and uninterpreted depth sections of Little Valley Landslide mass. The top prominent reflector is correlated with the clay layer found in Borehole 1........................................ 52

Figure 23 Seismic reflection Profile 2 with interpreted and uninterpreted depth sections of Little Valley Landslide mass. ......................................................................................... 53

Figure 24 Correlation tie between Profiles 1 and 2 .......................................................... 54

Figure 25 Seismic reflection Profile 3 with interpreted and uninterpreted depth section of Little Valley Landslide mass .......................................................... 55

Figure 26 Seismic reflection Profile 4 with interpreted and uninterpreted depth section of Little Valley Landslide mass. CDP spacing is 3 ft (1.03 m)..................................................................................................................... 56

Figure 27 Seismic reflection Profile 5 with interpreted and uninterpreted depth sections of Little Valley Landslide mass. .......................................................... 57

Figure 28 Contour map of Little Valley Landslide glide surface. ........................................ 58

Figure 29 Seismic reflection profile at the active Slumgullion Landslide in southwestern Colorado produced by the U.S.G.S.......................................................... 59
Introduction

Landslide hazards are a growing concern throughout the world. The cost of damages from landslides continues to increase as expanding communities continue to develop into areas with landslide potential. Understanding the nature of landslides in the form of internal structure and surrounding conditions is pertinent to developing an accurate and reliable slope stability analysis and mitigation strategies (Bruno and Marillier, 2000; Spiker and Gori, 2003; Godio, 2005). Boreholes and trench excavations are primary methods to investigating a landslide (Hunt, 2005), along with geomorphic observations (Bichler et al., 2004) but can be very costly and labor intensive (Bogoslovsky and Ogilvy, 1974). Geophysical methods are another method to investigating landslides (Bogoslovsky and Ogilvy, 1974; McGruffy et al., 1996) and have become increasingly more widespread in their use. Advancements in technology have allowed for more effective use of geophysical methods for investigations of landslides (Bichler et al., 2004). The geophysical methods that have been used in general landslide investigations include electrical resistivity, magnetic, electromagnetic, borehole geophysics, seismic reflection and refraction methods, ground penetrating radar, and a combination of all or some of these techniques (McCann and Forster, 1990; McGuffy, et al., 1996). Seismic reflection data, until recently, have seldom been used for landslide investigations generally due to the complexity of the internal structure and the relatively rough terrain of landslides. Seismic refraction methods are the preferred methods of geophysical investigation in order to supplement geotechnical investigations of landslides consisting of trenching and drilling (Hyland, 1996). Still, seismic refraction methods are limited when investigating complex geologic structures or interbedded low and high
velocity layers. This is because the conical head waves cannot propagate along the upper boundaries of low velocity layers (Ferrucci et al., 2000). Effectiveness of high-resolution seismic reflection data for shallow exploration has been proven various times in recently studied landslides such as the active Slumgullion Landslide in Colorado, USA (Williams and Pratt, 1996). Other recently studied landslides investigated with high-resolution seismic reflection methods are the Boup Landslide in the Swiss Alps (Bruno and Marillier, 2000), the Quesnel Forks Landslide in Canada (Bilcher et al., 2004), and the Sechilienne mass movement in France (Meric et al., 2005).

The purposes of this study are: First, to present the results of a multi-disciplinary investigation for a pre-historic landslide located in Draper, Utah, USA, previously mapped as the Little Valley Landslide (Biek, 2005a). Landslides are a growing concern along the Wasatch Front in Utah. Many recent investigations have been conducted by the Utah Geologic Survey to assess landslide hazards and their causes (Utah Geologic Survey, 2006). Landslides have been the cause of much damage to buildings and property in the mountainous areas of Utah (Ashland, 2003; Harty and Lowe, 2003; Hylland and Lowe, 1998). The Little Valley Landslide is one of many landslides under scrutiny because of expansion of residential developments into its immediate vicinity. The second purpose is to evaluate the effectiveness of the high-resolution seismic reflection technique for imaging the basal plane or glide surface of this landslide (Figure 1). It is anticipated that from the seismic data, images of the basal plane can be produced because of a velocity contrasts between the landslide-bedrock materials. The correlation of reflectors on these images across the loose network of profiles allows one to better understand the extent of portions of the landslide, especially into the depth dimension.
As it is currently mapped, the Little Valley Landslide compares in size, dimensionally, with the infamous Thistle Landslide the “most costly landslide disaster in the history of the United States” (Kaliser and Fleming, 1983). The Thistle Landslide occurred in 1983 in Spanish Fork Canyon, Utah, approximately 65 km south of the Little Valley Landslide. An estimated 22 million m³ of reactivated landslide material moved 150 meters on a 10 degree slope trough-shaped depression in bedrock.

The study of the Little Valley Landslide includes primary methods of investigation (i.e. trench excavations, borehole drilling, and radiometric dating) and geophysical methods (i.e. high-resolution seismic reflection and borehole geophysics). The integration of several geological and geophysical techniques makes this landslide one of the best constrained pre-historic landslides under investigation today. This study therefore provides a model of investigation that could be applied to landslide sites in general.

**Geologic Setting**

The Little Valley Landslide area is located along the northern flank of the Traverse Mountains that trend east-west at the southern portion of the Salt Lake Valley in Utah (Figure 2). The Traverse Mountains are located at the eastern border of the Basin and Range near the Wasatch Fault. The Wasatch Fault trends north-south through Utah and is the east-bounding normal fault separating the Basin and Range from the Colorado Plateau. The Traverse Mountains are the salient between the Salt Lake and Provo segments of the Wasatch Fault. Faults along the eastern edge of the Basin and Range Province were initiated in the late Cenozoic (Nolan, 1943) as a result of crustal extension and accompanying magmatism which followed the period of plate convergence and
crustal compression of the Sevier and Laramide Orogenies (Tooker, 2005; Hamilton, 1988). The Traverse Mountains are comprised of late Paleozoic shallow-marine rocks of the Pennsylvanian Oquirrh Group. These rocks consist of interbedded fine grained orthoquartzite and calcareous sandstone, sandy limestone, and limestone. They were uplifted during the Sevier and Laramide tectonic compressional events extending from Late Paleozoic to Late Mesozoic (Hamilton, 1988; Bryant and Nichols, 1988). The Paleozoic rocks are overlain by middle Tertiary intrusions and volcanic rocks in the form of block and ash flow tuffs, mud flow breccias, minor lava flows, and minor fluvial volcaniclastic deposits probably derived from volcanic centers in the west Traverse Mountains, including rhyolitic and latite plugs, dikes, and vents (Biek, 2003; Biek, 2005a; Biek, 2005b). These volcanic deposits are extensively hydrothermally altered and have thicknesses up to 330 m in the east Traverse Mountains near the study area. Quaternary alluvial and colluvial deposits are located in drainages around Traverse Mountain. Tertiary age normal faults parallel to the Wasatch Fault are oriented north-south through the Traverse Mountains and are not known to be presently active in the vicinity of the Little Valley Landslide (Machette, 1992; Biek, 2005a).

The head of the Little Valley Landslide is mapped on the Traverse Range at an elevation of 6000 ft (1828.8 m) above sea level and the toe is located down the mountain on Holocene Bonneville lacustrine sands and gravels at 5100 ft (1554.5 m). During the Pleistocene glacial interval, Lake Bonneville, an extensive body of freshwater, occupied the Salt Lake Valley and numerous other valleys between the ranges within the Basin and Range Province (Eardley et al., 1957; Crittenden, 1963). Bonneville shorelines are
preserved in places as bars, spits, and other coarse clastic shoreline features along broad terraces flanking the mountain front (Currey et al., 1984).

The local topography of the central portion of the Little Valley Landslide is characterized as a steeply sloped mountainous terrain leading into a small bowl-like valley. This small valley in the middle of the landslide is bounded to the north by a linear 150 m wide ridge trending northwest as seen on the topographic map (Figure 3). The Little Valley Landslide has been mapped 1.5 km long by 0.5 km wide by the United States Geologic Survey and identified as landslide deposits of Upper Pleistocene to Tertiary age (Machette, 1992). The Utah Geological Survey (UGS) has mapped the area as young Quaternary landslide movement from late Holocene to Late Pleistocene deposited by rotational and translational movement and characterized by the hummocky topography and chaotic bedding attitudes (Biek, 2005a). Biek (2005a) also notes that basal slip surfaces or glide surfaces most commonly form within the Tertiary volcanic rocks throughout the East Traverse Mountains. The expected direction of movement on the Little Valley Landslide is northwest.

**IGES Investigations**

Intermountain GeoEnvironmental Services (IGES), a geotechnical engineering company, has completed various amounts of work in portions of the Little Valley Landslide not accessible to the public because of private property. Data acquired and reports completed by them were accessed for further interpretation for this thesis project. The data provided by IGES is discussed in the following sections.
**Trench Investigation**

Trenches were excavated in three areas of the landslide; at the toe, in the middle section, and at the head near a present day sag pond (Figures 4-7). The trenches were excavated to identify locations of landslide masses in the shallow subsurface and associated shear zones. The length of trenching at the head of the landslide was 150 ft, (45.7 m) in the middle 513 ft (156.4 m), and in the toe 256 ft (78 m). The total trenching length was 919 ft (280.1 m).

Three trenches were excavated in the mapped toe of the landslide (Figure 5) revealing landslide material in two of them (Trenches 1 and 3). The landslide material consists of weathered volcanics with various colorations and disturbed or chaotic orientations. The volcanic clasts and matrix were observed to be weathering to clay. In Trench 1 at the toe, the landslide material was observed overlying lacustrine sand and gravel deposits from the Lake Bonneville high stand with no paleosol or organic-rich soil between them. This is a possible indication that the movement on the landslide could have coincided with Lake Bonneville (14,000 years before present). Colluvium was observed overlying the landslide material. Trench 2 had lacustrine deposits but no landslide material overlying them, an indication that the landslide may not have reached to the location of the trench in the northern extent of the landslide (Figure 5). Trench 3 had the landslide material overlain by colluvium with no lacustrine sands and gravels observed.

In the middle portion of the Little Valley Landslide four trenches were excavated where the surface geology was mapped as alluvial and colluvial deposits overlying landslide deposits of the Little Valley Landslide (Figure 6). Landslide material was encountered in trenches 1, 3, and 4; but not in Trench 2 (Figures 8 and 9). The landslide
material had characteristic shear planes, distorted contacts, and sediment filled tension fractures. It is possible that landslide material was not encountered in Trench 2 due to its location in the center of the landslide. A thick package of post-landslide alluvium and colluvium located in the center of the landslide overlies the landslide material, through which the trench was not able to penetrate. Charcoal fragments were encountered at the bottom of Trench 2 in undisturbed alluvial sediments and were sampled for radiocarbon dating to determine a bounding age of a last movement on the landslide (discussed below).

At the head of the landslide near a sag pond, one trench was excavated (Figure 7). A thick post-landslide package of alluvial sediments was encountered in the trench. The package can be divided into two parts with an unconformity separating the east dipping upper sediments from the west dipping underlying sediments. The underlying sediments are dipping cut by reverse faults with 30 cm maximum offset. The faults do not extend through the unconformity into the upper east dipping sediments. This thick package could represent a long post-landslide time interval. A paleosol encountered in the trench was sampled for radiocarbon age dating to determine a possible age of earliest deposition of sediment at the sag pond (discussion below).

A road cut in the middle portion of the site perpendicular to trenches 1 and 4 was exposed east to west. Tertiary volcanics consisting of weathered volcanic clasts ranging from small cobble to boulder sized and in a gray and brown weathered clay matrix were exposed in the western portion of the road cut (Figure 10). A near vertical shear zone with a brown clay-rich gouge layer (1 m thick) was observed in the road cut. This shear zone is a contact between volcanic rock and landslide mass (Figures 10 and 11).
Radiocarbon age dating

Radiocarbon age dating (Table 1) was performed on undeformed post-landslide alluvial sediments from trenches at the middle and the head of the landslide in order to have a constraint of the youngest possible time of landslide movement.

A sample was taken at 12 ft (3.66 m) depth from within Trench 2 at the middle of the landslide where no landslide material was encountered, but where a thick alluvial and colluvial sediment package was present. Charcoal fragments were encountered within a 1 to 3 inch thick layer at a depth of 12 feet within Trench 2 and were sampled and tested. The charcoal yielded conventional radiocarbon ages of 10780 a +/- 50 a BP (Table 1). These ages indicate the oldest post landslide deposition of the alluvium and colluvium occurred in Late Pleistocene. These dates constrain the middle of the landslide to Late Pleistocene/Early Holocene.

Another sample was collected and tested at the head of the landslide from the trench near the sag pond. No landslide material was encountered in the trench but there was a thick package of alluvial and colluvial sediments. The sediment yielded radiocarbon ages of 4210 a +/- 60 a BP (Table 1). The age of the sediment is an indication of the age of movement near the head of the Little Valley Landslide. This age is younger than the age reported from the middle of the landslide. This could represent younger movement (during Holocene) upslope following an older landslide event.
Table 1: Radiocarbon samples: locations and ages

<table>
<thead>
<tr>
<th>Location</th>
<th>Material sampled</th>
<th>Laboratory no.</th>
<th>$^{13}$C/$^{12}$C ratio</th>
<th>Conventional radiocarbon age</th>
</tr>
</thead>
<tbody>
<tr>
<td>Middle of landslide Trench 2, 12 ft depth</td>
<td>Charcoal fragments</td>
<td>Beta - 183310</td>
<td>-24.1 o/oo</td>
<td>10780 +/- 50 BP</td>
</tr>
<tr>
<td>Sag Pond Trench</td>
<td>Paleosol</td>
<td>Beta - 183308</td>
<td>-25.0 o/oo</td>
<td>4210 +/- 60 BP</td>
</tr>
</tbody>
</table>

**Borehole 1 and Gamma log**

Borehole 1 was drilled in the middle portion of the landslide to identify the thickness of the landslide and locate the depth of the basal plane or glide surface. A drill rig mounted with a Becker hammer was used to break through the hard volcanic clasts including abundant cobbles and boulders within the landslide deposit. Cuttings were acquired and changes in the lithology with depth were identified. The drilling log (Figure 12) indicates the amount of fine-grained material at specific depths and the lithology of the clasts. Colluvial and alluvial sediments were observed in the shallower depths down to 11 m (36 ft). These sediments consisted of clays and silts with some gravel sized clasts. Below 11 m (36 ft) down to 55 m (180 ft) a gravel-rich unit was observed. The gravel and cobbles consisted of volcanic clasts with differing stages of weathering and ranging in color from pink and red to gray. A clay-rich zone was observed at 55- 57 m depth. Refusal was encountered at 71 m (233 ft). A clay-rich zone would be expected in a zone of deformation. Movement along the slip surface or glide surface in certain Tertiary volcanic lava flows, breccias, and lahars will create a clay-rich zone along the decollement (Shuzui, 2001).
A gamma log was completed in Borehole 1 using a wire-line method carrying a gamma ray logging sonde to record the radioactivity of the material located in the borehole. The gamma log was used to verify and to independently and more accurately identify the depths of the specific lithologies and clay content in the borehole (Figure 13). The gamma log measures counts from radioactive elements (U, Th, K). These counts are to a first approximation equal to clay content. The gamma log shows deflections (increases and decreases in counts) to the right and the left in the near-surface corresponding to the alluvial and colluvial deposits. Then the log is consistent in the range between 80 to 140 counts per second (cps) until 65 m depth with only one brief deflection to the left (decrease in counts or clay content) at 50 m depth. High counts were observed at 65 m depth possibly corresponding to the clay-rich zone encountered in the drilling of the borehole (Figure 12). This clay-rich zone was interpreted to be a possible basal slip surface of the Little Valley Landslide.

**Location of Seismic Reflection Profiles**

Five high-resolution seismic reflection profiles were acquired throughout the middle of the landslide. The large bowl-like nature of the middle of the landslide and the thick package of alluvium and colluvium encountered in the trenches at least 16 ft (4.9 m) indicated that the interior of the landslide was an optimum area to acquire reflection data. Other areas within the landslide area had very steep slopes, very rugged terrain, and private property restrictions that made them undesirable for seismic acquisition. The surveys were very adaptive and were acquired in the locations of interest. Recent road cuts were an advantage for facilitating accessibility for acquisition and reduction in acquisition noise because the vegetation had been grubbed in specific locations. Profiles
1 and 2 were located on well-developed construction roads while Profiles 4 and 5 were located in less developed road cuts. Profile 3 was acquired on native terrain and possibly on some engineered road fill on the ends of the profile causing acquisition to be more difficult. Due to the close proximity of Profile 3 to the nearby road, traffic was a contributor to noise during acquisition. A total of 1342 meters of data were collected across the area of investigation. Overall the profiles were acquired in low noise zones with adequate conditions for data acquisition. Profile 5 was located at the highest elevations. For consistency an elevation of 1680 meters was used as a datum to reference all of the profiles to the same elevation. The depth of 0 on the seismic profiles corresponds to this datum.

**Seismic Reflection Methods/Processing**

Compressional wave seismic reflection data were acquired and processed using conventional techniques (Table 2) based on the common depth point (CDP) method (Mayne, 1962; Sheriff and Geldart, 1995; Yilmaz, 1987). The data were collected using a 48-channel recording system, sourced by an accelerated 100 lb (45 kg) weight drop per mounted to an All Terrain Vehicle (ATV) along the portions of the line that were not prohibitively steep. A 10-pound (4.5 kg) sledge hammer source was used in areas of rough terrain with steep slopes not accessible to the accelerated weight drop per source. Receiver stations with 28-Hz geophones were set at 10 ft (3.05 m) spacing. The source locations were stationed with the receiver stations at 10 ft (3.05 m) spacing to produce a nominal 24-fold record. The data were acquired in three campaigns each ranging further from Borehole 1. Station locations were surveyed approximately every 12 stations in order to assign geometry to the data.
Table 2: Acquisition Parameters
Survey type: P-wave
Channels: 48
Group interval: 10 ft (3.05 m)
Geophones/array: 1
Shot interval: 10 ft (3.05 m)
Source: 100 lb (45 kg) accelerated weight drop
Alternate source: 10 lb (4.5 kg) hammer
Stack: 3
Nominal fold: 24
Sampling rate: 0.25 ms
Record length: 1.5s

The seismic reflection profiles were acquired in order to detect and locate a reflector or reflector zone that could be correlated to the landslide-bedrock contact or more specifically to the clay layer observed in Borehole 1 at 55 m (180 ft) depth (Figure 12). This correlation was expected to be present across a large portion of the mapped landslide mass.

Data processing followed a conventional series of steps (Table 3). Individual shot records (Figures 14-18) were examined for quality control and noisy traces were deleted. The geometry from surveyed station locations was then assigned to the data. Shot records were then analyzed for reflections and refractions, and the direct arrival to determine initial estimates of the velocities (Figures 14-18). It was observed on the shot records that profiles 1, 2, and 5 had reflections and refractions with velocities between 1100 and 1400 m/s while profiles 3 and 4 had reflections and refractions with velocities between 500 and 700 m/s. These discrepancies in velocities determined from the shot records indicate a possible lithologic character change.
Table 3: Data Processing Procedures

Trace editing and killing of bad traces
Geometry assignments
Elevation Statics
Pre-stack deconvolution for wavelet compression
Bandpass filtering: Ormsby (50-80-240-400 for profiles 1-5; 20-30-240-400 for profile 4)
Mutes (top mutes, bottom mute of the surface waves)
Refraction statics
Velocity analysis
NMO correction
Trace mixing
CDP stack
Depth conversion (from time to depth)
Display

An elevation statics correction was applied using the replacement velocities determined of 1100 m/s for Profiles 1, 2, and 5, and 700 m/s for profiles 3 and 4. With the elevation statics correction, the data were stacked in a brute stack. This stack was used to identify reflections that would stack in constructively. The data showed some dipping reflectors coherent across most of profiles 1 and 2. Profiles 3-5 lacked prominent continuous reflectors.

Surface waves were determined to be partially adding constructively in the profiles creating what appeared to be coherent reflections or parts of strong reflection in upper portions of the stacks. Due to the inherent low frequency of the source and the coarse receiver spacing specific filtering of surface waves was not effective. These waves accordingly were muted out using an aggressive bottom mute on all of the records. A top mute was also applied to all of the profiles to mute out any possible refractions that may have been adding constructively and producing apparent continuous reflections. The top mute was picked at the first zero crossing of the traces on Profile 1. A deeper mute was picked on Profiles 2-5 at the third zero crossing. Where a top mute is made at the third
zero crossing, it was attempted to ensure that any refracted first arrival are eliminated (Miller, 2002). Picking a top mute precisely in order to separate a refracted first arrival from a high-velocity reflection tracking just after it was problematic in places due to the relatively low frequency content of the data and due to the expected high velocity of the volcanic bedrock underlying the landslide material. The merging of refracted arrivals and high-velocity reflections is a common problem on shallow high-resolution seismic data especially at longer offsets and low frequencies (Steeples and Miller, 1998; Bradford and Sawyer, 2002). Positioning of the top mute is sensitive because it may eliminate possible strong reflections and/or may lead to a change in appearance of shallow reflections in the final stack of the data. A mute analysis was completed on Profile 1 to determine any possible changes in the stack due to the change on position of the top mute (Figure 19). Profile 1 was chosen because of the apparent lack of refractor. The strong reflectors were eliminated as a mute at each zero crossing (1-3) was applied (Figure 19). An Ormsby bandpass filter was applied to cut specific frequencies from being stacked (see Table 2). The application of an Ormsby bandpass (trapezoidal) frequency filter was necessary to eliminate noise from the stacked sections; however, as frequency components are filtered out from the low end of the spectrum some artificial cyclicity is set up around the onset of reflectivity on the stacked section as shown in Figure 20.

A refraction static correction was applied to correct for changes in shallow velocity and structure based on picking first arrival times on the shot records. This correction smoothed the first arrivals areas of change along the first break refraction/reflection. Along areas where a first break refraction was not clearly visible or where the refraction and reflection appear to have merged, the first break was still used to
design a refraction statics solution. The result was then compared with the elevation statics solution for consistency. A velocity analysis was performed to choose stacking velocities using constant velocity stacks (Bradford, 2002). These velocities were used in the NMO correction and were converted to interval velocities for the time to depth conversion. A CDP stack with a three-element weighted interval trace mix was applied to mitigate high-apparent velocity noise. Seismic migrations were performed on the profiles; however, due to the low reflection and short length of the profiles, migration did not produce a significantly different result. Several tests of adaptive deconvolution were applied to the post-stack data in order to examine the effects of possible reverberation (short-path multiple reflection). Various prediction times and operator lengths were tested (Figure 21). Applying a post-stack deconvolution with a short prediction time (10 ms) compresses the onset arrival into a short duration waveform and also suggests that later arrivals may simply be multiples; however, it is likely that the deconvoluted section is removing weak primary reflections further down the record. Thus, the approach of applying an adaptive deconvolution only in the pre-stack domain was taken and some reverberatory noise remains at the onset of reflectivity. This is considered to be an acceptable solution since the primary interest is in the onset of reflectivity and avoidance of eliminating weaker, less coherent primary events. The final profiles showed great improvement in the coherency of the reflections and the reduction of noise.

Observations of the profiles

Seismic Profile 1
Of the five profiles, Profile 1, which trends southwest to northeast across the site (Figure 6), is of the highest quality and represents a key profile that is correlated to a
borehole with a possible glide surface identified. Profile 1 has a CDP length of 218 m. A prominent high-amplitude reflection appears centered around 100 m below datum (1680 m) around CDP 261, which is interpreted as a high-contrast geological surface (Figure 22). This surface correlates closely with a clay-rich layer (55 m below ground surface) recognized from the lithology log from Borehole 1 drilled near CDP 261 along Profile 1 (Figure 6). The gamma ray log independently shows the top of a clay zone at this depth (Figure 13). From near the southwest end of the profile, the strong reflector dips 20 degrees toward the northeast until CDP 230 where it begins to dip toward the southwest. It climbs upward to the northeast until CDP 325. At CDP 325 there is a disturbance in the reflector. The disappearance of the strong reflector as it approaches the edge of the profile can be interpreted to be its actual disappearance in the subsurface; however, it is expected to have a decrease in the quality (signal to noise ratio) of the image at the very edges of the profile due to a gradual loss of CDP fold of cover (e.g., toward the southwestern end of the profile, where the fold of cover drops below 10 at CDP 215). The cyclicity of the strong reflection is likely due in part to short-path reflection multiple arrivals or reverberation. Post-stack predictive deconvolution was partially effective in compressing this arrival into a single, more precise wave form using a 20-ms prediction distance; however the deconvolution applied post stack tended to eliminate weaker events that were not necessarily multiples. Below the strong reflector, lower frequency reflections appear to a depth of about 160 m below datum and show a distinct change in geometry and overall character. No shallow arrivals are seen in the profile due to the bottom mute aimed at surface wave and airblast removal just above the first high apparent velocity arrivals on the shot records.
Seismic Profile 2

Profile 2, beginning just off the southwestern end of Profile 1 (Figure 7), was surveyed in order to take advantage of an available smooth dirt road. This profile was acquired to confirm and extend the observations from Profile 1. Profile 2 has a CDP length of 270 m. Profile 2 (Figure 23) shows reflections at approximately the same travel time and depth as Profile 1, including the strong reflector on Profile 1. Also observed is the deeper set of geometrically distinct reflections. A direct correlation between the two profiles cannot be made due to the fact that they do not exactly cross each other and due to the usual problem of a drop in fold of cover approaching the ends of the profiles. Instead, correlation of reflections is evident near the southwestern ends of the two profiles where data quality and fold are high (Figure 24). Examination of the correlated sections demonstrates the excellent match between the two profiles, which allows the continuation of the same strong reflector (as well as reflectors above and below it) across and along the two profiles with a high degree of confidence. This correlation thus allows the mapping in three dimensions of the strong reflector across a larger area of the site.

Seismic Profile 3

Profile 3 was surveyed from southwest to northeast, just to the north of Traverse Ridge Road, which is located north of Profile 1 (Figure 7). Profile 3 (Figure 25) was surveyed in order to determine if the same strong reflector as found in Profiles 1 and 2 could be detected further to the northwest in a down-slope position. This profile has a CDP length of 294 m. Although the quality of Profile 3 is not as high as for the previous two profiles, it does show a reflector at a maximum depth of 95 m below datum (Figure
This reflector is disrupted at CDP 240, 280 and 310. At CDP 340 there is a change in dip of the reflections from west to east. At CDP 370 and beyond the reflections appear to lose coherency. This may be due to the rougher terrain near the northeastern end of this profile as well as due to the use of a hand-held sledge hammer for the seismic source beginning at about CDP 350 (the ATV-mounted weight dropper could not be used in this area). Profile 3 was also anomalous in that the shallow P-wave rms velocities were significantly less (<1000 m/s) than observed for the other seismic profiles. This suggests that the shallow material beneath Profile 3 is not correlative with shallow materials beneath the profiles south of the road. Because this reflector does not have the same characteristic as the strong reflector on Profiles 1 and 2, despite their close proximity to Profile 3, this reflector may not represent the same surface and could possibly indicate a different geological stratum. On the other hand, its depth is close to that observed for the strong reflector on Profile 1 and 2. A reflector appears across the profile that shallows to the northeast from depths of 105 to 85 m below datum (1680 m). It was observed that the reflector has similar characteristics as observed in Profiles 1 and 2 (Figures 22 and 23). The shot records for the beginning of Profile 3 were examined for these deeper reflections and compared with shot records from Profile 1. Similarities in velocity, frequency, and amplitude suggest that the reflector maybe the same as Profiles 1 and 2. The disturbances of the reflector at CDP 250 and lower indicate that it does not represent a strong impedance contrast across the record. In summary, the correlation of the strong reflector from Profiles 1 and 2 onto Profile 3 is not completely conclusive, due to the lack of a tie point.
Seismic Profile 4

Profile 4 began near the northeastern end of Profile 1, atop the hill, and was surveyed to the southeast (Figure 7). Profile 4 has a CDP length of 332 m. It also has the similar anomalous shallow P-wave velocities as observed in Profile 3. This profile was more sensitive to different bandpass frequency filters. When using a narrower frequency filter as was done with profiles 1-3 (Ormsby bandpass filter 50-80-240-400) a thinner layer of reflective sediments is observed as compared to Profiles 1 and 2 (Figure 26). A moderate amplitude, coherent reflection dipping to the northwest appears at 80 m below datum. This reflector rests on a band of reflectors that continues down to about 130 m below datum. This reflective band loses amplitude and coherency further northwest along the profile, especially at CDP less than 330 (Figure 26). Disruptions in the reflections are observed in this zone between CDP 230 and 330. The portion of the profile between CDP 210 and 330 appears to be similar in reflectivity to Profile 3, but no tie point is located between the two. Although Profiles 4 and 2 are not exactly contiguous, the northeastern end of Profile 2 projects within 130 m of the center of Profile 4 at CDP 300 (Figure 7). At this point of projection, the elevation of the strong reflector on Profile 2 does match well with the reflector mapped on Profile 4, allowing for some change in elevation along the line of projection. The reflection changes from 90 m below datum on Profile 2 to 83 m on Profile 4. This is a good indication that the reflector is the same as seen on Profiles 1 and 2.

Seismic Profile 5

Profile 5 is located near the base of the steep slope that is 400 ft (122 m) below the sag pond further upslope and had a CDP length of 219 m (Figure 7). Profile 5 was the poorest quality in terms of reflection coherency and noise contamination. A geological
boundary varying in depth below datum between 45 and 65 m has been interpreted based on a vertical change in overall reflection character and two prominent sets of cyclic reflections (Figure 27). Deeper sub-horizontal reflectors that appear to have some continuity are interpreted as high-contrast layers within volcanic rocks. Faulting is indicated by offset in the deeper sub-horizontal reflectors located along the profile with a highly disturbed zone between stations 134 and 148. An offset 25 m was observed from the seismic reflection profile. Profile 5, which is located outside of the interior of the landslide area along the slope of the mountain, is the only one to cross an independently mapped fault and show a prominent affect (Figures 7 and 28). Existence of this fault is independently confirmed by its correlation to a fault mapped by Biek (2005a) as an inferred structure.

**Geological Interpretation of the Results of the Seismic Surveys**

A strong sub-horizontal reflector was observed on the four profiles (1-4) that cut across the middle of the landslide with varying amounts of reflective coherency in specific portions. A lower coherency sequence of reflections was observed below the strong reflector. These reflections are correlated to Borehole-1 at a clay layer observed at 55 m depth. Considering the above observations and the geological context of the site as a known pre-historic landslide, the geological expression of the strong reflector interpreted on the profiles is consistent with a glide surface at the base of the Little Valley Landslide. This interpretation is supported by the observation of a clay-rich layer in Borehole 1, corresponding to the strong reflector on Profile 1 centered around 55 m below ground surface and by the gamma ray log, which shows the beginning of a zone of increased natural radioactivity near this same depth. The higher radioactivity can readily
be interpreted as an increase in clay content, which would be expected in a zone of
deformation involving clay gouge such as a glide surface. The reflectors immediately
above the interpreted glide surface can be explained as weathered bedrock and colluvial
or alluvial sedimentary deposits. The reflectors below the interpreted glide surface are
more enigmatic, but could be interpreted as stratal layering within the volcanic rocks that
underlie the site. Volcanic rocks are exposed in the trenches excavated in the middle
section of the landslide and exposed in road cuts at the construction site. The hill on the
south side of Traverse Ridge Road consists of these volcanic rocks.

The strong reflector interpreted as the basal plane on Profiles 1 and 2 can be
interpreted to Profiles 3 and 4, but not with a direct tie. The lower velocity reflectors seen
in Profiles 3 and 4 are interpreted to be the same reflector. Since the reflector in Profile 4
can reasonably be projected to Profile 2, it is interpreted that the reflectors observed in all
of Profiles 1-4 are the basal plane of the Little Valley Landslide.

A contour map (Figure 28) has been constructed for the depth of the interpreted
glide surface in below datum (1680 m) using Profiles 1 through 4. Profile 5 was excluded
in the formation of the contour map because of the low signal to noise ratio, no close
correlation tie to the other profiles, and to avoid introducing uncertainty to the contour
map. The structural contour map displays a generally smooth shape (black contours),
which is consistent with the reflectors representing the same geological surface. The
contour map reveals a north-plunging elongate synform (i.e., a down folded structure).
The plunge direction approximately follows the contours of the topography downslope,
as might be anticipated for a landslide glide surface (Figure 1). An irregular, undulating
scoop-shaped or concave surface is typical of landslides (Narwold and Owen, 2002).
Discussion

The geological and geophysical data collected for this investigation have revealed critical information on the subsurface of the Little Valley Landslide. Landslide material was encountered in the shallow subsurface using trenching and drilling. Specific areas in the shallow subsurface revealed the location of landslide mass, shear zones and undisturbed volcanic rocks as observed in Trench 1 and the road cut. The information gathered in the trenches was useful to understanding the nature of the materials involved in the landslide and near the shear zones. The trenches excavated at the toe of the landslide revealed deformed landslide material that extended to the Bonneville sands and gravels. In specific locations, as found in Trench 2 at the toe, the landslide may not have reached the extent as mapped. The landslide material encountered in the toe was relatively thin compared to other locations trenched in the landslide.

The trenches in the middle portion and at the head of the landslide revealed thick alluvial units deposited after landslide deformation. They also revealed only the very shallow subsurface portions of landslide. The trenching in these areas gave no indication of depth to basal plane of the landslide. Thus, no true understanding of the structural nature of the glide surface was acquired.

Various movements or events can be interpreted to have occurred as supported by the differences in the radiocarbon ages from the middle of the landslide up to the head. The radiocarbon dating on the undeformed sediments encountered in the trenches constrains movement to approximately 13,000 years before present in the middle portion of the landslide and 5,000 years before present at the sag pond near the head scarp. This
is important because the surficial nature of the pre-historic landslide area will be more complex when introducing multiple movements, erosion, and sediment deposition.

The depth to the base of the middle of the landslide was observed in Borehole 1 with the clay-rich zone at approximately 55 meter depth. The lithologic observation is supported by the gamma ray log. These observations alone are not sufficient to determine any large-scale geometry or possible extents of the landslide. This is important for supplementing the seismic reflection data. Since the Little Valley Landslide is pre-historic it is unlike other studies of active or historic landslides (Bilcher et al., 2004; Williams and Pratt, 1996) because no pre-landslide topography data is available. The information from the borehole and gamma log became critical to validating the seismic reflection profiles. The supplementary data from the seismic surveys gave critical information understanding the more fully the basal plane of the landslide.

The seismic reflection data collected were critical to our understanding the overall structure of a highly complex pre-historic landslide area. The strong reflector observed in Profiles 1-4 can be interpreted to be imaging the basal plane of the landslide. Thus, the clay-rich layer or basal glide surface zone of the landslide was successfully imaged across a large area of the middle portion of the Little Valley Landslide. The data complied into the contour map indicate that the glide surface is at a lower depth below datum, downslope to the northwest. This is expected as the landslide mass possibly moved along the pre-landslide topography.

In some areas there is uncertainty as to whether or not the glide surface is present. This uncertainty is from a lack of reflection observed on the seismic profiles or the change in reflector character. This is seen on Profiles 3 and 4 where the reflection
character is weak and velocities are different than seen on Profiles 1 and 2. This could be due to a change in characteristics of the landslide-bedrock contact. The reflector could represent the top of rigid volcanic bedrock just below the slide. Changes in the amount of clay in the damage zone may affect the reflective character of the basal plane across the landslide. The interpretation of a strong shallow reflector as a landslide glide surface is not without precedence. As mentioned before, limited but similar work has been done on landslides using seismic reflection methods. Profiles acquired on the active Slumgullion Landslide in southwest Colorado, U.S.A. (Williams and Pratt, 1996) are similar to the profiles on the Little Valley Landslide in respect to the strong reflector at the base of the landslide (Figure 29). At the Slumgullion Landslide seismic reflection and refraction methods were implemented in the study for the sole purpose of identifying the contact between slide and non-slide material. In their findings they identified a broad U-shaped feature on the profile which was interpreted to be a pre-landslide valley. It is believed that this study has found a similar scenario at locating the base of a landslide, but with more constraints to the data due to the borehole lithology and gamma ray data.

Bichler (2004) demonstrated that a suite of geophysical techniques is valuable for subsurface investigations to result in more detailed and less ambiguous interpretations of the three dimensional structure of the Quesnel Forks Landslide. It is interpreted that through this study a detailed interpretation has been made to the depth and extent of portions of the Little Valley Landslide. This is information that was not known before and will be useful to determining slope stability (not assessed in this study).
Conclusions

The integration of geological and geophysical exploration of the Little Valley Landslide has revealed a degree of structural information about a pre-historic landslide that would not be possible with more traditional approaches. An integrated investigation using multiple techniques was able to give a greater understanding to the three-dimensional geometry of a large portion of the Little Valley Landslide, which complements the information on the ages of movement ranging from $10780 \pm 50$ BP and $4210 \pm 60$ BP in the middle portion and head area, respectively.

From this study it was found that high resolution seismic reflection methods can be a very useful complimentary method for further understanding larger scale mass movements with a decollement or glide surface at depth. The initial investigations through trenching and drilling were very successful in revealing the geological characterization of the near subsurface by locating volcanic rock masses and landslide masses in specific locations (i.e. toe, sag pond), but they could not reveal a depth of the glide surface and large scale view of the structural nature of the Little Valley Landslide. This is due to the depths to the base of the landslide in the middle portion.

The seismic reflection data were able to allow a greater understanding of the geometry of the central portion of the landslide by producing an image of a reflector representing the basal plane of the landslide across a great portion of the landslide.

The acquisition technique allowed for data to be acquired quickly, even across the rough and rugged mountainous terrain with some limitations where topography was not conducive. The effectiveness of using seismic reflection methods to image the subsurface in comparison to drilling showed that more area could be covered and at less expense in
terms of resources and time. Still, the two methods compliment each other and validate one another.

The interpretation of the seismic data is strengthened from the information acquired from Borehole 1. The need for drilling various other boreholes throughout the middle portion of the site was eliminated or reduced because of the high quality and correlation of the reflectors in the seismic data.

Due to the greater desire to better understand landslides spurred by the rapid expansion of urban and suburban development worldwide, this tool can greatly enhance geotechnical investigations and slope stability modeling. Seismic reflection methods are shown again to be an accurate and precise tool to investigating landslides, even with the complexity landslides possess. The information from the seismic reflection profiles from this study will be used to create slope stability models (not assessed in this study) in the near future.
References


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Figures

![Diagram of general cross section of a landslide]

Figure 1 General cross section of landslide showing toe, head, and basal plane of a landslide.
Figure 2 Index map of the Little Valley Landslide location with surrounding Salt Lake Valley, Wasatch Range, and Wasatch Fault (modified from Personius and Scott, 1992)
Figure 3 Topographic map showing boundaries of the Little Valley Landslide. The sag pond located at the head of the landslide is seen at the southern portion of the landslide. The toe is located to the north-east.
Figure 4 Geologic map of the Little Valley Landslide showing areas investigated: toe, middle, and sag pond areas. Units: IPou-Pennsylvanian Oquirrh Group, Tv-Tertiary Volcanics, Qac- Quaternary Alluvium and Colluvium, Qmsy- young Mass Movement, Qmso- Older Mass Movement, Qlgb-Quaternary Lacustrine Bonneville Gravels, Qafy- Quaternary Young Alluvial Fan, Qf-Engineered Fill (for more descriptions of units see Biek, 2005)
Figure 5 Enlarged view of the trench locations at the mapped toe of the Little Valley Landslide.
Figure 6 Investigation areas in the middle portion of the Little Valley Landslide on geologic map. The trenches and borehole are located in the small bowl-like valley. The seismic profile locations are shown in red. (Modified from Biek, 2005)
Figure 7 Trench location at the head of the Little Valley Landslide near the sag pond (Modified from Biek, 2005).
Figure 8 Representative log of Trench 1 in the middle of the landslide showing landslide mass, shear zone, and volcanic block.
Figure 9 Representative log of Trench 2 in the middle of the landslide showing layered alluvial and colluvial deposits.
Figure 10 Photograph of road cut wall showing shear contact between volcanic bedrock and landslide material.
Figure 11 Road cut wall showing shearing and thrusting within the landslide mass.
Figure 12 Borehole-1 Log
Figure 13 Gamma Ray Log Borehole 1
Figure 14 Raw shot record with no filters (on left) and filtered shot record with AGC, Ormsby bandpass filter (50-80-240-400), and refraction statics correction of Profile 1. A-reflections, B-refraction, C-Surface waves, D-Airwave. Channels are at 10 ft (3.05 m) spacing.
Figure 15 Raw shot record with no filters (on left) and filtered shot record with AGC, Ormsby bandpass filter (50-80-240-400), and refraction statics correction of Profile 2. A-reflections, B-refraction, C-Surface waves, D-Airwave. Channels are at 10 ft (3.05 m) spacing.
Figure 16 Raw shot record with no filters (on left) and filtered shot record with AGC, Ormsby bandpass filter (50-80-240-400), and refraction statics correction of profile 3. A-reflections, B-refraction, C-Surface waves, D-Airwave. Channels are at 10 ft (3.05 m) spacing.
Figure 17 Raw shot record with no filters (on left) and filtered shot record with AGC, Ormsby bandpass filter (50-80-240-400), and refraction statics correction of Profile 4. A-reflections, B-refraction, C-Surface waves, D-Airwave. Channels are at 10 ft (3.05 m) spacing.
Figure 18 Raw shot record with no filters (on left) and filtered shot record with AGC, Ormsby bandpass filter (50-80-240-400), and refraction statics correction of Profile 5. A-reflections, B-refraction, C-Surface waves, D-Airwave. Channels are at 10 ft (3.05 m) spacing.
Figure 19 Stacks of Profile 1 with varying top mutes from 1st to 3rd zero crossing. The shallow reflectors are muted as the more aggressive (deeper) mute is applied.
Figure 20 Frequency test.
Figure 21 Deconvolution Test.
Figure 22 Seismic reflection profile 1 with interpreted and uninterpreted depth sections of Little Valley Landslide mass. The top prominent reflector is correlated with the clay layer found in Borehole 1. CDP spacing is 3 ft (1.03 m).
Figure 23 Seismic reflection Profile 2 with interpreted and uninterpreted depth sections of Little Valley Landslide mass. CDP spacing is 3 ft (1.03 m).
Figure 24 Correlation tie between profiles 1 and 2
Figure 25 Seismic reflection Profile 3 with interpreted and uninterpreted depth section of Little Valley Landslide mass. CDP spacing is 3 ft (1.03 m).
Figure 26 Seismic reflection Profile 4 with interpreted and uninterpreted depth section of Little Valley Landslide mass. CDP spacing is 3 ft (1.03 m).
Figure 27 Seismic reflection Profile 5 with interpreted and uninterpreted depth sections of Little Valley Landslide mass. A fault is interpreted offsetting prominent reflectors. CDP spacing is 3 ft (1.03 m).
Figure 28 Contour map of Little Valley Landslide glide surface (in black).
Figure 29 Seismic reflection profile at the active Slumgullion Landslide in southwestern Colorado produced by the U.S.G.S. Bedrock refraction model shown in black, bedrock reflection shown as dashed gray line (Williams and Pratt, 1996).