AN INTEGRATED STUDY OF DEEP-SEATED GRAVITATIONAL SLOPE DEFORMATIONS AT HANDCAR PEAK, SOUTHWESTERN BRITISH COLUMBIA

by

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ABSTRACT

I propose an integrated methodology for the study of deep-seated gravitational slope deformation (DSGSD) and apply the methodology to a gravitationally deforming slope at Handcar Peak in southwestern British Columbia. I mapped DSGSD-related geomorphologic features such as antislope scarps, trenches, ponds, rockslides, and rockfalls on aerial photographs and in the field, and conducted an investigation of sediments deposited behind an antislope scarp to determine the history of movement of the feature. I also characterized the structure and strength properties of the deforming rock mass through engineering geological mapping and investigated the mechanics of movement by kinematic analysis and distinct element numerical modelling.

Results suggest that the current episode of movement at Handcar Peak began during or shortly after deglaciation and is continuing. Gravitational lineaments are the surface expression of displacement on weak fault planes. Deformation in numerical models is driven by slip on these faults and on downhill-dipping joints.

Keywords: Deep-seated gravitational slope deformation, sackungen, British Columbia, Lillooet Valley, geomorphology, distinct element modelling.
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1: INTRODUCTION

Deep-seated gravitational slope deformation (DSGSD) is the slow, incremental, gravity-driven movement of a large rock slope. Gravitationally deforming slopes have been identified and studied in many alpine areas around the world (Jahn 1964, Zischinsky 1969, Tabor 1971, Radbruch-Hall et al. 1976, Bovis 1982, Gutierrez-Santolalla et al. 2005, Ambrosi and Crosta 2006). However, the phenomenon remains poorly understood due to the large size and complexity of DSGSD and an inability, in most cases, to observe deformation at depth.

This thesis contributes to an improved understanding of DSGSD by documenting a slowly deformed crystalline rock mass in the Coast Mountains of southwestern British Columbia. The goals of the research are:

1. Present an integrated methodology for DSGSD research that addresses the specific challenges of studying the phenomenon.

2. Present a case study of a gravitationally deforming rock slope at Handcar Peak, British Columbia, that demonstrates the application of the integrated methodology, using three techniques:
   a. remote- and field-based geomorphologic mapping
   b. stratigraphic and sedimentologic analysis of sediments deposited behind gravitational lineaments on the slope
c. geomechanical analysis of the deforming rock mass by engineering geologic mapping and numerical modelling.

1.1 Definition of the phenomenon

Deep-seat gravitational slope deformation is characterized by (1) an extremely to very slow rate of movement (0.5 - 80 mm/yr), (2) total displacements that are much smaller than the size of the deforming rock mass, (3) large size (volume typically > 10 million m$^3$), and (4) signature deformational surface features, such as trenches, closed depressions, ponds, uphill- and downhill-facing scarps, and bulging of the toe of the deforming slope. The phenomenon is known in the literature by several other names, notably “mass rock creep” (Chigira 1992), “lateral spreading” (Radbruch-Hall et al. 1976), and “sackung” (Zischinsky 1969). I prefer the term “deep-seated gravitational slope deformation” because of its clear and more widely inclusive meaning.

1.2 Early research

The concept of deep-seated gravitational slope deformation was first proposed by European workers to explain alpine linear surface features, referred to here as “gravitational lineaments”, that include ridge-top depressions and antislope scarps (Fig. 1). These features previously had been attributed to climatic and erosional processes such as nivation and wind scouring (Paschinger 1928). Jahn (1964) and Zischinsky (1966) were among the first to show that the lineaments were the result of gravity-driven displacement along discontinuities in bedrock underlying slopes.
Figure 1. Cross-section view of alpine lineaments associated with DSGSD.

### 1.3 Identified types and environments

Over the past several decades, researchers have presented case studies of DSGSD from a wide variety of mountain environments. Varnes et al. (1989) identify three distinct lithologic settings in which gravitational slope deformation typically occurs:

1) **Massive stiff rocks that overlie weak ductile materials.** Deep-seated deformation of this type, here termed “gravitational spreading”, occurs when soft underlying rock deforms plastically and moves outward and downward in the direction of least stress (Radbruch-Hall et al. 1976). As the overlying layer of stiff material experiences tension and begins to pull apart and sag, tension cracks and grabens develop at the surface. Radbruch-Hall et al. (1976) report examples
of gravitational spreading of ridges in the Rocky Mountains of the United States where the weak layer is composed of shale or other soft sedimentary rock. Similar cases have been reported in the Dolomite and Apennine ranges in Italy (Pasuto et al. 1994, Pellegrini and Tosati 1994).

Gravitational spreading can drive movement on a large scale when weak layers occur at depth. In Canyonlands National Park, Utah, a huge system of grabens, approximately 10 km wide by 20 km long, formed in clastic sedimentary rocks overlying a thick evaporite unit when the Colorado River incised the formation (Fig. 2). The viscoplastic behaviour of the evaporites facilitated gravitationally driven movement towards the canyon on an average slope of only 2-4° (Schultz-Ela and Walsh 2002).

2) Ridges composed of metamorphic rocks with a strong foliation, or true sackungen of the type studied by Zischinsky (1969). Chigira (1992) demonstrates that where the strike of the foliation is roughly parallel to the strike of the slope, the near-surface stress field pulling the rock mass downward will fold and deform the metamorphic layers by micro-fracturing and flexural slip. Using case studies from Japan, he provides a schematic guide to the different kinds of folds that develop with different attitudes of foliation (Fig. 3). Where foliation dips with the slope, buckling or dragging of folds forms a dispersed zone of rupture at the base of the deforming rock mass, and the attitude of the overlying metamorphic fabric is preserved (Fig. 3A).
Figure 2. Gravitational spreading over evaporites in the Grabens District, Canyonlands, USA; view northeast. Colorado River is visible at the left. The width of the spreading zone near the centre of image is approximately 10 km.

On slopes where the metamorphic fabric is sub-horizontal, the presence of steep joints or faults at the rear of the deforming mass is necessary to link the fabric-related discontinuities with low-angle shear zones that form the base of the DSGSD. Backward rotation occurs, resembling a slump (Fig. 3B). Where the metamorphic fabric is sub-vertical or dips steeply into the slope, it tends to rotate
forward and downslope over a zone of flexural toppling or drag folding. The cases presented by Zischisnky (1966) follow this scenario.

![Diagram of gravitational faults and folds](image)

**Figure 3.** Types of gravitational faults and folds in foliated metamorphic or bedded sedimentary rocks, after Chigira (1992) by permission. Dashed lines represent planes of folding or shearing.

DSGSD in bedded soft sedimentary rocks may conform to Chigira’s (1992) theory of brittle folding under gravity. In the Olympic Mountains of Washington State, Tabor (1971) found that sandstone beds are bent downslope on the side of a ridge with a pronounced ridge-top depression. His interpretive sketch depicts downslope bending of steeply dipping strata (Fig. 3C). Li et al. (2010) report a similar scenario in the mountains of southern Alaska where sandstone beds have undergone flexural toppling. Tensile stresses are concentrated at the fulcrum of the toppling beds, causing them to fracture and develop a basal shear plane.

3) **Ridges composed of fractured hard crystalline rocks.** Due to their strength, rocks of this type do not deform easily by brittle folding, faulting, or shearing of intact material under near-surface gravitational stresses. Rather, movement occurs primarily through slip on pre-existing discontinuities within the rock mass. As with DSGSD in foliated metamorphic rocks, pervasive discontinuities that strike parallel to the slope are a prerequisite for movement to
occur. Varnes et al. (1989) present case studies from several sagging ridges in crystalline rocks in Colorado and Montana. In all cases, they find a correlation between the orientation of ridge-parallel lineaments and major joint sets, and conclude that valley-ward movement typically occurs normal to the major joint sets.

Planes of weakness formed by pre-existing faults may be significant in accommodating movement, and the lineaments they form have been mistaken for active tectonic fault scarps due to their unusual length and straightness. A straight, 1.6-km-long scarp in quartz diorite that has been the subject of much debate extends along the ridgeline northeast of Mount Currie in the southern Coast Mountains of British Columbia. Bovis and Evans (1995) present movement data from survey stations around the northeastern end of the scarp showing that the hard crystalline rock mass downhill of the feature is moving towards the precipitous north face of the ridge at rates of 5-15 mm/yr. Based on field evidence and kinematic analysis, they conclude that the movement is occurring through gravity-driven toppling. Thompson et al. (1997) dug a trench across the same scarp and analyzed the sediment fill for clues about the movement history. They found evidence for several episodes of movement along a pre-existing fault, consistent with gravitational deformation.

Elsewhere in the Coast Mountains of B.C., the formation of lineaments at Hell Creek (Clague and Evans 1994) and Handcar Peak (this study) is attributed to gravitational reactivation of relict tectonic features. Pre-existing faults have also been implicated in the formation of gravitational linearments in high-grade
metamorphic rocks in the Austrian Alps (Reitner and Linner 2009) and in Italy (Ambrosi and Crosta 2006, Agliardi et al. 2009).

Weathering and progressive rock mass damage may also facilitate gravitational deformation of crystalline rock slopes, as demonstrated by studies in British Columbia that correlate rock mass quality with displacement monitoring data and modelling results. A deforming granodiorite ridge at Wahleach, east of Vancouver, displays relatively low rock mass quality near the slope surface due to weathering of joint surfaces and the presence of numerous slope-parallel shear zones (Stewart and Ripley 1999). Displacements in numerical models of the Wahleach slope are concentrated in this upper zone of low rock mass quality. Alzo’ubi (2009) examines how near-surface weathering reduces rock mass tensile strength, allowing fracturing and dilation in areas of stress concentration. His numerical models of a deforming slope at Checkerboard Creek, British Columbia, agree with field observations that tensile fracturing and related slope displacements are concentrated in the most highly weathered portion of the rock mass.

1.4 Triggering factors

As in the case of conventional large landslides, the interplay between topographic and geologic factors is the major determinant of the occurrence of DSGSD. The shaping of mountain landscapes by geomorphologic processes creates the conditions that allow or encourage slow movement of the rock mass. Cyclical processes such as seasonal groundwater fluctuations or earthquakes can drive movement by disturbing the metastable slope.
An important morphological process in the formation of DSGSD, cited in most published studies on the subject, is debuttressing of the slope (Bovis 1990, Stepanek 1992, Agliardi et al. 2001, Ambrosi and Crosta 2006, Reitner and Linner 2009). Slope debuttressing occurs at times of glacier retreat. Major ice streams deepen the mountain valleys through which they flow, steepening the toes of valley-side slopes (Fig. 4). When the glaciers retreat, support is removed from the oversteepened slopes and they relax towards the valley until equilibrium is reached between gravitational stresses and rock mass strength (Ballantyne 2002). An example is the DSGSD at Affliction Creek, British Columbia. Bovis (1990) proposes that recent retreat of Affliction Glacier has caused active gravitational deformation by debuttressing the 200-m-high slope. Debuttressing can also occur through fluvial erosion, such as incision of the Colorado Plateau by Colorado River that led to the formation of the aforementioned grabens in Canyonlands National Park (Schultz-Ela and Walsh 2002).

Figure 4. Valley deepening by glacial erosion, followed by debuttressing and slope response. 1) Valley geometry before glaciation. 2, 3). Glacier occupies the valley and deepens it through erosion. 4) Gravitational stresses cause the oversteepened slope to relax towards the valley after deglaciation.
Recently, McColl et al. (2010) have called into question the relevance of
glacial debuttressing as a significant factor in precipitating slope movement. They
point out that ice, due to its low density and ability to flow easily at low strain
rates, is not a suitable material for buttressing a heavier and stiffer rock slope. An
implication is that gravitational deformation may actually commence before
glacier retreat in response to erosion of the slope and that movement that begins
after deglaciation is more likely due to climatic factors or seismicity. Currently, no
analogue physical or numerical models have been published that test this
hypothesis.

Seasonal increases in the water table associated with heavy rains or
snowmelt can cause episodic movement of a metastable rock slope, because
higher pore water pressure lowers the effective frictional strength on joint and
fault surfaces. Measured displacements of a deep-seated rockslide at Campo
Vallemaggio, Italy, show that the movement rate fluctuates on an annual cycle,
increasing in response to seasonal increases in pore water pressures
(Bonzanigo et al. 2007). A sharp drop in velocity followed the installation of a
drainage adit at the base of the sliding mass (Eberhardt et al. 2007). Monitored
gravitationally deforming slopes in British Columbia also show a positive
correlation between precipitation and displacement rates (Stewart and Ripley
1999, Stewart and Moore 2001). Numerical modelling supports the importance of
pore water pressures in driving gravitational movement; several researchers
have simulated seasonal fluctuations in DSGSD activity by cyclically raising and
lowering a modelled water table (Bovis and Stewart 1998, Stewart and Ripley 1999, Smithyman 2010).

A potentially important climatic factor in DSGSD activity, which thus far has received little attention, is the effect of seasonal temperature variations. Watson et al. (2004) compare detailed monitoring data of displacement velocities at Checkerboard Creek, British Columbia, to temperature and groundwater levels. Interestingly, seasonal activity correlates more strongly with temperature than groundwater. Watson et al. (2004) use UDEC to model near-surface rock mass expansion and contraction in response to seasonal thermal fluctuations. Results are consistent with their hypothesis that greater activity in winter is due to a decrease in effective normal stress on discontinuity surfaces as the rock mass cools and contracts, reducing discontinuity frictional strength.

Seismic shaking is also thought to contribute to DSGSD activity in some areas. Cadoppi et al. (2007) compiled a detailed spatial database of geomorphologic features in the Susa Valley, Italy, and found a close relationship between areas of recent tectonic activity and DSGSD. Gutierrez-Santollala et al. (2005) studied sediments deposited behind antislope scarps in the Spanish Pyrenees. They conclude that movement began long after deglaciation during a dry period, and thus was most likely due to earthquakes in the region. Numerical modelling studies of DSGSD support these conclusions; seismic loading of a gravitationally deforming slope triggers additional model displacements (Stewart and Ripley 1999, Stewart and Moore 2001).
1.5 Interpretations of mechanisms

The mechanisms responsible for DSGSD movement can be broadly grouped into two categories, depending on whether the deforming rock mass is considered a continuous or discontinuous material. Continuum-based deformational mechanisms assume that a large rock mass is composed of many small individual blocks that can be represented by an equivalent continuous material. An elastoplastic Mohr-Coulomb rheology is typically assumed for large rock slopes. Failure occurs due to plastic yielding, where concentrations of stress induced by the topography overcome the internal strength of the rock mass. Continuum mechanics work well to explain lateral spreading of strong stiff blocks over weak ductile material, because the main factor controlling spreading is the ability of the underlying soft material to behave plastically (Fig. 5A).

Savage and Varnes (1987) use a closed-form mathematical solution to determine regions of plastic yield due to gravitational forces in a generalized symmetric ridge and then apply a model of plastic flow to those regions. Rupture surfaces develop where discontinuities in velocity propagate through the flowing material. Lineaments form where these discontinuities intersect the surface. Differential flow velocities indicate an extending and thinning region near the ridge top where antislope scarps and grabens form, a zone of plug flow on the middle of the slope where deformation is not expressed at the surface, and a zone of thickening and compression at the base of the slope, expressed as a toe bulge (Fig. 5B).
Figure 5. Continuum-based DSGSD mechanistic models. A: Lateral spreading of stiff blocks overlying weak ductile material (after Radbruch-Hall et al. 1976, by permission). B: Plastic flow (after Savage and Varnes 1987, by permission). C: Shearing and deformation of continuum material (grey area) in finite difference model, shown by deformation of ubiquitous joints (after Stewart 1997, by permission). No vertical exaggeration.
Researchers have used numerical modelling to explore DSGSD development from a continuum-based perspective. Stewart and Ripley (1999) use the program FLAC to model the response of a deforming slope in southwestern British Columbia to yearly fluctuations in groundwater. They simulate the strength anisotropy of the rock mass resulting from its discontinuity structure using FLAC’s ubiquitous joint Mohr-Coulomb constitutive model. Their results indicate that deformation is concentrated in the shallower part of the rock mass, which is weak from weathering and from fracturing and dilation due to stress release during tectonic and glacial unloading. In their model, movement occurs both through shear failure along steep ubiquitous joints and through plastic deformation of the rock mass (Fig. 5C).

Discontinuum-based deformational mechanisms focus on the structure of the rock mass to analyze how deformation can occur through incremental slip along discontinuities. Deep-seated movement can occur through simple kinematic mechanisms such as sliding or toppling, or through more complex failure modes.

Flexural and block-flexural toppling are among the most commonly cited mechanisms for DSGSD formation (Fig. 6A). Movements driven by toppling tend to slow or stop in the long term, because the angle of toppling discontinuities decreases to a more stable geometry as layers rotate forward. Toppling has been cited in the formation of several DSGSD in British Columbia. Bovis (1982), for example, explains how block-flexural toppling of monzonite blocks along steeply dipping joints can account for the formation of antislope scarps and
tension cracks in the upper part of a slope formed in granitic rocks at Affliction Creek. Nichol et al. (2002) report a flexural topple in metamorphic rocks at Mount Breakenridge. Weak schistosity planes dipping steeply into the slope facilitate flexural slip between layers.

Planar sliding is less commonly associated with deep-seated gravitational slope deformation, probably because rock masses moving over a discrete shear surface tend to develop a conspicuous headscarp and sidescarps and are thus classified as “landslides”, and because slow creep along a discrete sliding plane will tend to progress towards catastrophic failure as existing rock bridges are broken and the plane becomes smooth and slickensided. Nevertheless, planar sliding has been cited in cases of DSGSD where the failure plane is oriented at a low angle and movement is only likely during periods of elevated pore water pressures, such as during deglaciation or seasonally with snowmelt (Simmons and Cruden 1980, Stepanek 1992, Smithyman 2010).

Kinakin (2004) proposes a complex mode of failure to explain deep-seated gravitational deformation at Mount Mercer in southwestern British Columbia. The surface morphology is dominated by a series of benches and normal scarps, features that are most compatible with rock slumping (Fig. 6C; Kieffer 1998).
Figure 6. Discontinuum-based DSGSD mechanisms and associated landforms. A: Block-flexural toppling of quartz monzonite in response to recent retreat of a glacier and debuttressing of the valley side (after Bovis 1982, by permission). B: Sandstone blocks sliding at low angle over siltstone with low-friction kaolinite seams (after Stepanek 1992, Fig. 6 p. 235; by permission). C: Rock slumping (after Kinakin 2004, by permission).
1.6 Significance of DSGSD as a natural hazard

Creeping rock slopes can damage homes, highways, pipelines, and other permanent structures built on them. Problems of this kind are common in the Alps due to the high population density of valleys near deforming slopes. For example, slow gravitational movement has damaged small hydroelectric facilities (Ambrosi and Crosta 2006) and threatens at least one dam in the Alps (Kalenchuk 2010). Some issues of this type have also arisen in British Columbia. Slow gravitational movement of a slope at Wahleach sheared through a steel-lined power conduit that carried flow through the mountainside from a lake down to a power station in the Fraser Valley (Stewart 1997). The conduit had to be relocated deeper within the slope, below the lower limit of the deforming rock mass. The event drew attention to the active DSGSD at the site, which had not previously been identified as a hazard to the facility.

An indirect, but potentially much more dangerous hazard arises from the progressive damage to a rock mass incurred by deformation over time. In some cases, slowly deforming rock slopes weaken to the point that they fail catastrophically. Holm et al. (2004) conducted a statistical analysis of the association between the characteristics of alpine basins in the upper Lillooet River watershed, British Columbia, and the occurrence of slope instabilities. All of the catastrophic rock slope failures identified in their study area occur in areas of deep-seated gravitational slope deformation. Gradual DSGSD-type movement preceded the 1965 Hope Slide, the largest rock slope failure in Canada in the historic period. In the absence of clear hydrological or seismological triggers,
Evans and Couture (2002) conclude that rock damage that accumulated through long-term progressive displacements caused the catastrophic failure. The hazard posed by large, sudden rock-slope failures such as the Hope Slide is evident – in settled areas such as the Alps, the consequences could be catastrophic. In British Columbia, some gravitationally deforming slopes are adjacent to reservoirs (Martin Lawrence, personal communication, 2009). Sudden failure of one of these slopes could trigger a tsunami, with potentially serious local and downstream impacts. In response to this risk, BC Hydro and Power Authority has installed monitoring systems on several slopes adjacent to reservoirs and conducted regular field inspections.

Rockfall and small landslides are common within the unstable mass of some gravitationally deforming rock slopes (Fig. 7; Agliardi et al. 2001, Ambrosi and Crosta 2006). Although not as dangerous as the large failures discussed above, they are much more common and are capable of causing deaths and economic losses.

1.7 Summary

Deep-seated gravitational slope deformation (DSGSD) is the process whereby large rock slopes slowly sag under their own weight, producing surface features such as ridge-top depressions, antislope scarps, and toe bulging. Total displacements are at least an order of magnitude lower than the size of the deforming rock mass. DSGSD occurs in a variety of rock types, from stiff sedimentary rocks overlying soft evaporites to crystalline intrusive rocks.
Figure 7. Small rockslides and areas of rockfall within the DSGSD at Handcar Peak.

In weaker rock types, deformation can occur through microfracturing, brittle folding, and shearing of intact rock, whereas in strong crystalline rocks slip along discontinuity surfaces is more important for accommodating movement. Debuttressing of glacially oversteepened valley walls is the most commonly cited cause of DSGSD initiation, and in some cases seismic shaking is probably an important contributing factor. Seasonal changes in temperature and pore water pressure are important in driving ongoing movement. Gravitationally deforming slopes pose a hazard to structures built on them because of ongoing slow
movements, and some DSGSD are associated with sudden slope failures such as rockfall and rock avalanches.
2: AN INTEGRATED METHOD FOR THE STUDY OF DEEP-SEATED GRAVITATIONAL SLOPE DEFORMATION

Although deep-seated gravitational slope deformation (DSGSD) is related to conventional landslides, it differs in several important ways. The differences inform the methodology used to study DSGSD. This chapter discusses four important aspects of DSGSD: the limits and size of the deforming rock mass; the initiation and history of movement; the state of ongoing activity; and the mechanics of movement. I describe an integrated methodology that aims to characterize gravitationally deforming slopes holistically by exploring these aspects of their genesis and evolution.

2.1 Geometry, size, and depth

Conventional landslides can be mapped by the trace of the rupture surface at the head and sides of the sliding mass, by a bulging area at the toe of the landslide, and in the case of a past catastrophic failure, by the outline of the failure scar. In contrast, the limits of a DSGSD can be difficult to identify because the transition between stable and deformed rock is typically gradational (Mahr 1977). Most gravitationally deforming slopes have no recognizable rupture surface at the top or sides of the unstable rock mass, and a bulge may not be present at the toe (Savage and Varnes 1987, Reitner and Linner 2009). In many cases there are no discrete basal failure surfaces at all (Bovis 1990, Stewart and Ripley 1999, Stewart and Moore 2001). For these reasons it is important to
understand relevant aspects of the geological setting of DSGSD and to map indicators of instability on the surface of the deforming slope.

Mapping of linear surface features associated with DSGSD, such as uphill-facing (antislope) scarps and ridge-top depressions, helps to delineate the approximate edges of the unstable rock mass, especially the upper and lateral limits of deformation (Savage and Varnes 1987). Springs, displaced Quaternary sediments, or bulges may indicate deformed rock on the lower portion of a slope (Reitner and Linner 2009). Unusual drainage patterns and surficial instabilities may also be indicators of DSGSD (Ambrosi and Crosta 2006). These observations should be supplemented with information on the underlying bedrock. Faults, folds, and lithologic contacts may affect the geometry of the deforming rock mass. For example, a layer of weak metamorphic rock underlies a gravitationally deforming slope at Blaise Creek, British Columbia (Fig. 8). This layer provides a constraint for the base of the deforming rock mass.

2.1.1 Literature survey and remote sensing

The first step in the analysis of a DSGSD is to collect existing information through a library search. A review of geologic maps and relevant geologic studies in the region should be undertaken before any fieldwork is done. Geomorphologic terrain maps, if not already available, can be created from aerial photographs or satellite imagery in combination with digital elevation models (DEM). Standards for terrain mapping in British Columbia are provided in “Guidelines and Standards to Terrain Mapping in British Columbia” (Resources Inventory Committee 1996). DSGSD-related features should be mapped
Figure 8. Deep-seated gravitational slope deformation at Blaise Creek, British Columbia, showing lithological units mapped at the top of the deforming rock mass. a) gneiss, b) massive quartzites, c) weaker layers of schist, calc-silicate rocks, and marble. Arrow indicates direction of gravitational movement.

at a minimum scale of 1:20,000. High-resolution DEMs generated by airborne Light Detection and Ranging (LIDAR) surveys are recommended for identifying subtle surficial features only a few metres in height that may not be evident in aerial photographs (van Zeyl 2009). Aerial photographs, satellite images, and DEMs provide information on the surface morphology of the DSGSD and aid in planning fieldwork. In addition, good sites for trenching (Section 2.2) and engineering geological rock mass characterization (Section 2.4) can be identified at this stage in the investigation.
2.1.2 Field mapping

Geologic and geomorphologic field mapping is a vital element in the characterization of DSGSD. Site-scale geologic maps generally are unavailable, but local lithology and structure commonly control the deformation (Fig. 8). Features mapped remotely must be checked in the field to verify the height and orientation of lineaments, types and characteristics of sediments and bedrock, and evidence of surficial instability. Important features that cannot be identified through remote sensing, such as small springs, sinkholes, and subtle lineaments, can be identified in the field. Additionally, recent movement that occurred after aerial photographs or satellite images were acquired can be noted.

2.1.3 GIS

Mapping, both remotely and in the field, is greatly enhanced by the use of geographic information systems (GIS) software. Base data such as geologic maps, DEMs, and aerial photographs are stored in a georeferenced database and can be rapidly combined in different ways with mapped features (Fig. 7). Data or notes acquired in the field can also be stored in a GIS database and associated with global coordinates gathered using a handheld GPS device. Examples of such information include field measurements, qualitative descriptions of lineaments, geotechnical mapping stations, and field photographs.
2.2 Initiation and history of movement

The task of determining the causes and triggers of DSGSD is a difficult one, because the phenomenon is coeval with landscape-scale geomorphologic processes and long-term progressive weakening of the rock mass. Researchers of DSGSD directly observe the process over a short period, but it is necessary to study the history of movement over centuries or millennia to gain an understanding of the events that may have contributed to their initiation and evolution. Previous studies have correlated the initiation of movement to known paleo-events such as glacier retreat, wet climate, or earthquakes (Agliardi et al. 2001, 2009, Gutierrez-Santolalla et al. 2005).

An understanding of the history of a DSGSD serves several purposes. Past slope behaviour with respect to paleo-events may constrain slope behaviour under similar conditions in numerical models or help to predict the future behaviour of the DSGSD under a variety of conditions. DSGSD movement histories also provide key evidence as to whether lineaments are gravitational or tectonic in origin. Slow and steady movement favours a gravitational origin, whereas sudden episodic movement favours tectonism or tectonically driven gravitational movement (McCalpin 2003).

2.2.1 Historical records

Several techniques are available for reconstructing the history of gravitationally deforming slopes. In the Alps and other mountainous regions of Europe where permanent structures have existed for centuries, damaged buildings are evidence of movement (Dramis and Sorriso-Valvo 1994, Weißflog
et al. 2010). It may be possible to determine historical movement rates from the amount of displacement of damaged structures of known age. Written records may also exist, although slow gravitational deformation of slopes is far less likely to be noted in historical literature than catastrophic slope failures.

2.2.2 Displaced deposits

Displaced glacial deposits also provide a constraint on the age of DSGSD activity. Many deforming slopes are blanketed by sediments deposited during the last Pleistocene glaciation. These sediments have been displaced by antislope scarps or other gravitational lineaments. The age of these deposits provides a maximum age for initiation of the current episode of movement on the lineaments. Evidence of this type, however, does not preclude older episodes of movement because any pre-existing lineaments and displaced sediments would probably have been removed by erosion during the last glaciation. Cross-cutting relationships between lineaments and younger surficial deposits, such as rockslide rubble or Little Ice Age moraines, can provide evidence for or against more recent movement (Hippolyte et al. 2009).

2.2.3 Trench studies

The movement history of individual lineaments can be inferred from an analysis of the sediments that have accumulated in depressions behind these features (McCalpin and Irvine 1995, Thompson et al. 1997, Gutierrez-Satolalla et al. 2005, Agliardi et al. 2009). The sediment sequence is exposed in trenches dug across the lineaments. Through georeferenced photographs, detailed
sketches, and sedimentologic analysis, the researcher can determine the
direction and amount of movement on the lineament and whether the movement
was continuous or episodic (Fig. 9). Radiocarbon dating of organic material
recovered from the sediments provides ages for different stages of scarp
evolution (McCalpin and Irvine 1995, Gutierrez-Santolalla et al. 2005, Agliardi et
al. 2009). Recent studies have used cosmogenic nuclide dating of gravitational
scarp surfaces to estimate the time at which bedrock was first exposed to
sunlight as the scarp formed (El Bedoui et al. 2009, Hippolyte et al. 2009).

Trenching allows an examination of a cross-section of the shear surface
along which movement has occurred to produce a lineament. The character of
the shear surface can support either a gravitational or tectonic origin for the
feature. Gravitational shear zones formed under near-surface stresses are
asymmetric; the boundary with the downsliding upper block tends to be sharp,
whereas a gradational brecciated shear zone occurs below (Chigira 1992).
Tectonic shear zones formed at depth in the crust are expected to be more
symmetrical. Care should be taken in using this criterion, however, because
many gravitational lineaments occur along relict faults (Clague and Evans 1994,
2.3 Contemporary activity

It can be difficult to classify gravitationally deforming slopes as active or inactive because they move so slowly. Reported surface velocities extend below the detection limit of most instruments, with no known minimum threshold. Rates calculated by averaging displacement of trench sediments over hundreds or thousands of years can be less than 1 mm/yr (McCalpin and Irvine 1995). Average velocities, however, fail to represent probable scenarios in which movement occurs sporadically during periods of high pore water pressure or earthquakes. Rates of movement measured directly over short intervals range from 4 mm/yr (Bovis and Evans 1996) to 80 mm/yr (Bovis 1990).

Commonly, rates of historic movement differ spatially over a large gravitationally deforming slope. Some sections may contain active landslides, whereas others can be effectively inactive but still display lineaments or other surficial evidence of deformation (Ambrosi and Crosta 2006). The relevant problems, therefore, are to determine what parts of the slope are most active, if...
and where movements are great enough to affect structures built on the deforming slope, and whether movement is likely to slow, remain constant, or accelerate.

2.3.1 Visual inspection

Contemporary DSGSD activity can be assessed in several ways. Visual inspection of gravitational lineaments and surficial instability, either on large-scale aerial photographs or in the field, can yield information about relative activity of the deforming slope. Recently formed lineaments may be sharper, steeper, and display freshly exposed, relatively unweathered sediment or bedrock surfaces. Less active and relict lineaments gradually become weathered and their depressions fill with sediments and vegetation (Fig. 10). These criteria should be used with some caution, however, because lineaments at high elevations weather slowly and may appear fresh for a long time. Surficial instabilities such as small slumps, rockslides, and rockfall can also serve as indicators of the most recently active areas of a gravitationally deforming slope (Agliardi et al. 2001).

2.3.2 Field measurements

DSGSD activity can be assessed directly by measuring surface movements, either in the field or using remote sensing techniques. Field-based measurements are the more commonly used. With geodetic surveys, precise locations of a series of points on the surface of the deforming mass can be determined repeatedly over a period of time, providing information on the
Figure 10. A: Well vegetated, inactive antislope scarp at Handcar Peak, with person for scale (downhill direction to the right). B: Active antislope scarp at Fels Glacier in the Alaska Range with daypack for scale; note rupture of alpine vegetation mat and lack of regrowth on scarp surface.
magnitude and direction of movement (Bovis 1990, Bovis and Evans 1995, Malgot 1997). In high-risk areas, more detailed and frequent monitoring may be appropriate. For example, a 3 million m$^3$ deforming rock mass is perched on the valley side at Checkerboard Creek, adjacent to Lake Revelstoke, about 1 km above Revelstoke Dam in British Columbia (Stewart and Moore 2001). BC Hydro installed an array of instruments to monitor slope activity, including electronic distance measurement instruments, extensometers, inclinometers, piezometers, time domain reflectometry (TDR) cables, a three-dimensional strain gauge, and an automatic data acquisition system to deliver real-time updates on the slope’s behaviour. Monitoring on this scale is rarely done in studies of DSGSD due to financial constraints, but many of the techniques can be used effectively alone.

2.3.3 InSAR

Remote sensing of surface displacements is being increasingly used in many disciplines of earth science, including those concerned with landslides. Interferometric synthetic aperture radar (InSAR), a satellite-based active surface scanning technology, has proven particularly useful in studies of gravitationally deforming slopes. A good summary of InSAR and its application to landslides is provided by Colesanti and Wasowski (2006). They detail the advantages and disadvantages of traditional differential InSAR (DInSAR) and permanent scatterer InSAR (PS InSAR). DInSAR uses the phases of reflected radar waves recorded in multiple SAR images. When images of the same area taken at different times are combined, an interferogram can be produced that displays changes in the position of the ground as a phase shift (Bürgmann et al. 2000). Permanent
Scatterer InSAR works on the same principle, but uses a data-filtering technique to identify objects that produce a consistent and stable radar reflection over multiple interferograms (Colesanti and Wasowski 2006).

Both techniques can detect millimetre-scale differences in ground surface position along a one-dimensional line of sight from the track of the observing satellite. Traditional DInSAR, however, suffers from several potential sources of error that make quantification of movement difficult. Possible sources of error include satellite orbit irregularities, temporal changes in the reflectivity of the ground surface especially in vegetated areas, and atmospheric artefacts induced by water vapour in the troposphere (Colesanti and Wasowski 2006). In practice, traditional DInSAR works best in detecting slow coherent movement over a large area of lightly vegetated or bare ground. Interferograms generated in such areas have been used to map the extent and pattern of gravitational surface deformation and to differentiate these areas from areas deformed by seismicity (Saroli et al. 2005, Stramondo et al. 2005).

Permanent scatterer InSAR has many advantages over traditional DInSAR, including suppression of several sources of error, resulting in much better and quantifiable precision and the ability to correlate deformation values with individual features such as buildings or prominent rock surfaces (Colesanti and Wasowski 2006). PS InSAR has proven to be an excellent tool for mapping movement velocities of large deforming slopes in the Italian Alps. For example, relative velocities of different parts of a deforming slope on the southeastern flank of Valtellina valley, Italy, provided insight into the geometry of a potential sliding
surface (Ambrosi and Crosta 2006). At another slope, located along the east shore of Como Lake, the distribution of measured velocities was compared to that predicted by stress concentrations in a three-dimensional numerical model (Ambrosi and Crosta 2006).

2.4 Mechanics of movement

A required step in understanding DSGSD, both for specific sites and for the phenomenon in general, is to analyze the physical mechanisms by which rock slopes deform slowly under gravity. Characterizing these mechanisms allows for prediction of slope behaviour and is the most important step in assessing the hazard posed by an unstable rock slope (Stead et al. 2006). Although sufficient data are rarely available for detailed predictive analysis of DSGSD behaviour, preliminary analyses can explore potential mechanisms, give a rough picture of hazard potential, and serve as a basis for deciding whether a more detailed study is needed. Chapter 1 provides a summary of published interpretations of DSGSD mechanisms. Here, I discuss techniques for performing a geomechanical analysis of DSGSD.

2.4.1 Data collection

The data mentioned earlier in this chapter should be collected before a geomechanical analysis is undertaken. The information they provide on past and current slope behaviour is crucial in a data-limited geotechnical analysis, which is typical for studies of DSGSD. Proposed mechanisms, when tested, should produce behaviour consistent with field observations of the history and style of
movement, surface expression of movement, and geometry of the deforming rock mass. Other necessary contextual information, such as topography, local geology, glacial history, and failure limits, is collected prior to geomechanical analysis.

Engineering geological field mapping is necessary to characterize the mechanical properties of the deforming rock mass. Field mapping has three major aspects:

1. **Intact rock strength characterization**: The strength properties of the intact blocks within a fractured rock mass can be assessed in several ways. Simple field methods used to estimate rock strength include the use of a geological hammer (Brown 1981), or a Schmidt Hammer, which measures the rebound of a mass impacting the rock surface with a known energy (Aydin and Basu 2005). More precise characterization of intact rock strength can be accomplished using laboratory-based testing methods, such as, in order of increasing difficulty and expense: point load, uniaxial compression, or triaxial compression (Brown 1981).

2. **Discontinuity characterization**: The geometrical structure of discontinuities that separate blocks of intact material within a rock mass should be characterized in terms of the orientation and variability of major sets of joints and the foliation or bedding where present. This characterization traditionally has been accomplished by compass-based outcrop mapping of discontinuity surfaces. Recent advances in remote sensing technology allow discontinuities to be mapped on detailed 3D digital models generated using photogrammetry or
ground-based LiDAR scanning (Sturzenegger and Stead 2009). The quality of discontinuity surfaces should be characterized by recording parameters such as roughness, weathering grade, and presence of soft infill or water (International Society of Rock Mechanics 1978).

3. Rock mass characterization: Several methods of rock mass characterization are used to evaluate the quality of a rock mass as a whole. Currently the most commonly used method is the Geological Strength Index, or GSI, which is based on estimates of average block size and quality of discontinuity surfaces (Fig. 11; Hoek et al. 2002). Estimates of rock mass quality are used to calculate physical parameters for an equivalent continuum material that approximates the overall behaviour of the rock mass (Brown 2008).

2.4.2 Types of analysis

Stead et al. (2006) define three levels of mechanical analysis that can be applied to complex rock slopes. A Level I analysis uses traditional stereograph-based kinematic techniques and limit equilibrium methods; a Level II analysis applies continuum and discontinuum numerical modelling techniques; and a Level III analysis uses hybrid continuum-discontinuum modelling programs that simulate intact rock fracturing (Table 1).

Level I analyses have limited use for the investigation of DSGSD. One problem is that they only account for geometrically simple, idealized modes of failure such as planar sliding, wedge sliding, and simple toppling. In addition,
**Figure 11. Qualitative guide for determining the GSI value of a rock mass (Marinos et al. 2005, by permission).**
they can only define the limiting conditions under which a slope will begin to fail; they do not account for progressive movement as a slope continues to deform (Lorig and Varona 2004). The latter problem is especially important for DSGSD, because progressive movement is of primary interest. The main contribution of a Level I analysis with respect to gravitationally deforming slopes is to provide preliminary tests that may guide further analysis with more sophisticated techniques. For example, a kinematic analysis might highlight the role that discontinuity structure could potentially play in accommodating rock mass deformation, or provide insight on potential release surfaces, thus serving to inform how subsequent numerical models should be constructed (Bovis and Stewart 1998).

A Level II analysis is better suited to investigating complex DSGSD mechanics than a Level I analysis, because it allows a slope failure to develop naturally from a given set of initial parameters (Lorig and Varona 2004). More complex modes of failure can develop, and some codes can model the progressive evolution of a failure for as long as desired after its onset.

There are two general approaches to Level II modelling, based on the decision of whether to represent a rock mass as a continuous or discontinuous material. The material is better represented as a continuous medium when the persistence of discontinuities in a rock mass is small compared to the size of the slope being studied, no clear kinematic modes of failure exist, or the stress field in the slope is sufficient to overcome the shear strength of the rock mass.
Table 1. Different levels of mechanical analysis and their application to DSGSD.

<table>
<thead>
<tr>
<th>Analysis type</th>
<th>Input data</th>
<th>Uses with respect to DSGSD</th>
<th>Examples in DSGSD studies</th>
</tr>
</thead>
<tbody>
<tr>
<td>Level I</td>
<td>Kinematic (stereographic)</td>
<td>Discontinuity orientations, friction angles of discontinuity sets</td>
<td>Holmes and Jarvis (1985), Bovis and Evans (1996), Reitner and Linner (2009)</td>
</tr>
<tr>
<td>Level II</td>
<td>Continuum (finite element)</td>
<td>Topographic profile, rock mass properties, major structures, glacial history, tectonic stresses, groundwater regime, expected earthquake acceleration</td>
<td>Savage and Varnes (1987), Stewart and Ripley (1999), Agliardi et al. (2001), Kinakin and Stead (2005)</td>
</tr>
<tr>
<td></td>
<td>2D</td>
<td>Same as 2D continuum but requires DEM of topography instead of profile</td>
<td>Ambrosi and Crosta (2006), Li et al. (2010)</td>
</tr>
<tr>
<td></td>
<td>3D</td>
<td>Same as 2D continuum but requires DEM of topography instead of profile</td>
<td>Kalenchuk 2010</td>
</tr>
<tr>
<td>Discontinuum</td>
<td>2D</td>
<td>Topographic profile, rock mass properties, major structures, discontinuity set orientations and strength properties, glacial history, tectonic stresses, groundwater regime, expected earthquake acceleration</td>
<td>Bovis and Stewart (1998), Nichol et al. (2002), Hurlimann et al. (2006)</td>
</tr>
<tr>
<td></td>
<td>3D</td>
<td>Same as 2D discontinuum but requires DEM of topography instead of profile</td>
<td>Kalenchuk 2010</td>
</tr>
<tr>
<td>Hybrid continuum/</td>
<td>Requires same data as 2D discontinuum models, plus intact rock properties, including time- and strain-dependent behaviour (strain-softening, progressive failure)</td>
<td>Analysis of complex failure modes involving internal deformation, strain softening and progressive failure through tensile fracturing of rock bridges</td>
<td>Alzo’ubi 2009</td>
</tr>
<tr>
<td>discontinuum, fracture simulation</td>
<td></td>
<td></td>
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</tbody>
</table>


In contrast, a discontinuum approach should be used when the relative size of discontinuous blocks in a slope is significant and their geometrical structure can be characterized, or the shear strength of the rock mass is clearly greater than the stress field in the slope (Stewart 1997, Brideau et al. 2009).

The researcher must also decide whether to model a slope in two or three dimensions. Classic two-dimensional slope analyses are based on several simplifying assumptions: (1) there are high-angle discontinuities that strike parallel to the direction of the cross-section, providing out-of-plane kinematic release; (2) the properties of the rock mass are homogenous in the out-of-plane direction; (3) the dip direction of the discontinuity sets in the model is exactly parallel or perpendicular to the slope; and (4) the potential direction of movement being analyzed is parallel to the cross-section (Brideau and Stead 2010). Natural slopes never conform to all of these conditions, of course, but may still exhibit behaviour that can reasonably be represented in two dimensions. Most published analyses of DSGSD mechanics based on numerical modelling use a two-dimensional approach (Savage and Varnes 1987, Bovis and Stewart 1998, Agliardi et al. 2001, Nichol et al. 2002, Kinakin and Stead 2005, Hurlimann et al. 2006). However, additional important insights have been gained about three-dimensional stress distribution in gravitationally deforming slopes through the application of 3D continuum modelling (Ambrosi and Crosta 2006, Li et al. 2010). Three-dimensional discontinuum modelling can also reveal complex modes of failure that are not apparent in simple stereonet analysis or 2D modelling (Kalenchuk 2010). A recent study based on 3D discontinuum modelling
demonstrates that joint sets dipping obliquely with respect to the direction of movement can exercise significant kinematic control over rock slope failures (Brideau 2010). A modeller beginning an analysis of a gravitationally deforming slope must carefully consider the slope morphology, discontinuity set orientations, and distribution of rock mass properties before deciding on the best approach.

A Level III analysis requires detailed input data about intact rock and rock mass properties, and generally should be performed by a modelling specialist. However, the unique capabilities of programs that model internal fracturing, shearing, and dilation of rock slopes can improve the analysis of DSGSD. In particular, this type of approach can shed light on the way that progressive damage in a deforming rock mass leads to its catastrophic failure and can help identify the factors that favour such failure. The Randa rockslide occurred in a slope that exhibits signs of gravitational deformation (Pedrazzini et al. 2010). Eberhardt et al. (2004) use a hybrid finite element/discrete element modelling code to demonstrate the importance of tensile fracturing in degrading rock mass strength at Randa until a shear plane could develop, precipitating catastrophic failure. Software tools for realizing Level III analyses continue to improve. For example, a new modelling code is currently under development that will improve computation speed and has the ability to couple fluid flow, pore pressure distribution, and rock deformation, including fracture of intact rock (SlopeModel – see Lorig et al. 2009).
2.5 Summary of integrated methodology

A detailed understanding of large, gravitationally deforming rock slopes requires characterization of several aspects of the phenomenon: the limits and size of the deforming rock mass; the time of initiation, conditions at initiation, and history of movement; the distribution and rate of movement of zones of current activity; and the mechanics that drive deformation of the rock mass. Table 2 summarizes the techniques discussed in this chapter according to four general attributes of DSGSD. An integrated study of a gravitationally deforming slope need not employ all these techniques, but it should aim to characterize each of the four general attributes in some way. The four attributes are interrelated: knowledge about each one contributes to a better understanding of the others. Taken together, they present a detailed picture of the phenomenon. These relationships can be demonstrated by using an interaction matrix approach (Fig. 12; Hudson 1992, Kinakin 2004).
<table>
<thead>
<tr>
<th>Attributes</th>
<th>Size and limits of the deforming rock mass</th>
<th>Initiation and history of movement</th>
<th>Distribution and velocity of contemporary movement</th>
<th>Mechanics of movement</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Activities</strong></td>
<td>Acquire existing bedrock and surficial geology maps</td>
<td>Review glacial and tectonic history</td>
<td>Inspect linears and other surface features</td>
<td>Perform engineering geological mapping: rock mass mapping, discontinuity survey, and estimate intact rock strength</td>
</tr>
<tr>
<td></td>
<td>Map geomorphic features using aerial photographs and satellite images</td>
<td>Review historical records and damaged structures</td>
<td>Perform geodetic surveys</td>
<td>Test rock strength in laboratory</td>
</tr>
<tr>
<td></td>
<td>Conduct field-based geomorphologic and bedrock mapping</td>
<td>Examine displaced surficial deposits</td>
<td>Monitor slope using instruments</td>
<td>Do kinematic and limit equilibrium analyses</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Study sediments deposited in depressions associated with lineaments</td>
<td>Perform InSAR study</td>
<td>Perform continuum and discontinuum 2D and 3D numerical modelling</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Use lichenometry, tephrochronology, cosmogenic nuclide dating, or radiocarbon dating if possible</td>
<td>Conduct hybrid modelling analysis with fracture simulation</td>
<td></td>
</tr>
</tbody>
</table>

<table>
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<tr>
<th><strong>RESOURCES</strong></th>
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### Figure 12

Interaction matrix showing the interdependency of knowledge about different attributes of a gravitationally deforming rock slope.

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<tr>
<th>Size and limits of the deforming rock mass</th>
<th>Initiation and history of movement</th>
<th>Distribution and velocity of contemporary movement</th>
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<tr>
<td>Size and limits of the deforming rock mass</td>
<td>Displaced surface deposits give maximum age for current movement and can record movement history</td>
<td>Defines outer boundaries of area that may be active</td>
<td>Distribution, extent and style of deformation indicated by mechanical analysis should agree with features mapped on slope in reality</td>
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<tr>
<td>Initiation and history of movement</td>
<td>Style of movement over time indicates whether deformational features are tectonic or gravitational</td>
<td>Movement history gives clues about current activity before direct measurements are made</td>
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<td>Distribution and velocity of contemporary movement</td>
<td>Surface movement vectors can be used to infer underground geometry of deforming rock mass</td>
<td>Currently inactive zones with surficial indicators of deformation are assumed to have been active in the past</td>
<td>Model behaviour should agree with style of past movement</td>
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<td>Mechanics of movement</td>
<td>Geometry of unseen failure surface must allow mechanically feasible modes of failure</td>
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<td>Stress distributions predict areas prone to activity Models indicate possibilities for future evolution of movement velocity and distribution</td>
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3: HANDCAR PEAK

3.1 Previous work

Numerous deep-seated gravitational slope deformations (DSGSD) have been identified in British Columbia, especially in the southwestern part of the province, but relatively few have been the subject of field investigations. Bovis and Evans (1996) report more than 20 DSGSD in the southern Coast Mountains in the vicinity of Lillooet Valley, including the Handcar Peak site (Fig. 14). They describe the geomorphic features of the DSGSDs and present the results of a kinematic analysis at four of the sites. Two other sites in the southern Coast Mountains have been studied extensively – Affliction Creek (Bovis 1982, 1990, Bovis and Stewart 1998) and Mount Currie (Evans 1987, Bovis and Evans 1995, Thompson et al. 1997).

Handcar Peak, described as “the largest known slope movement complex in the Coast Mountains” (Bovis and Evans 1996), has been the subject of reconnaissance investigations dating back to the mid-1990s. Researchers have made a kinematic interpretation of the failure (Bovis and Evans 1996) and analyzed the sedimentary fill in a trench behind one of its antislope scarps (Lund 2002). The Handcar Peak DSGSD provided a good opportunity to apply the integrated methodology discussed in Chapter 2, specifically to obtain a better understanding of the timing and cause of deformation by integrating information gained from geomorphic mapping and trench sediment analysis with engineering geologic mapping and numerical modelling.
3.2 Geologic setting

The Handcar Peak DSGSD is located on the northeast side of the Lillooet River valley, 38 km northwest of Pemberton, British Columbia (Fig. 13). This area lies within the southern Coast Belt of British Columbia, a morphogeologic region with a complex tectonic history. The local bedrock comprises Middle Jurassic to Middle Cretaceous granitic plutons that intrude a series of island arc terranes thought to have been accreted to North America in the Mesozoic (Monger and Journeay 1994). In the Late Cretaceous, northeast-directed compression caused intense deformation and metamorphism of these terranes as they were thrust to the west along a series of steeply dipping reverse faults known as the Coast Belt Thrust System. The Owl Creek Fault, which cuts through the middle of the deforming slope at Handcar Peak, is a steepened northeast-dipping ductile thrust at the west margin of this system (Fig. 13; Riddell 1992). The major lineaments at Handcar Peak are oriented parallel to the Owl Creek Fault and the strong northwest-southeast structural fabric of the region.

Three major geologic units make up the deforming rock mass at Handcar Peak (Fig. 14). Metavolcanic rocks of the Pioneer Formation of the Cadwallader Group underlie the middle and upper sections of the slope, where the linear surface features are most abundant (Monger and Journeay 1994). The Pioneer Formation consists of basaltic to rhyolitic porphyritic flows and tuffs that have been metamorphosed to greenschist facies (Riddell 1992). These rocks have been intruded by Late Cretaceous quartz diorite of the Hurley River Pluton, which forms the upper part of the eastern section of the slope. Thin (1-3 m), northwest-
striking, ductile shear zones, possibly related to the Owl Creek Fault, occur near the contact between the Pioneer Formation and the Hurley River Pluton (Fig. 15a). Late Jurassic quartz diorite of the Lillooet River Intrusion forms the lower part of the deforming slope, west of the Owl Creek Fault. Small outcrops of quartz diorite are found throughout the mapped area of the Pioneer Formation (Fig. 15b).

Figure 13. The Lillooet River valley area, showing topography, DSGSD sites, location of Handcar Peak (after Bovis and Evans 1996), and major tectonic features (after Monger and Journeay 1994).
Figure 14. Topography and geology of the area around Handcar Peak (geology after Monger and Journeay 1994). Dotted line = slope transect used in numerical models.

The landscape of the Coast Mountains has been strongly shaped by glaciation. The Coast Mountains supported alpine glaciers throughout the Quaternary Period and, at times, were covered by the Cordilleran Ice Sheet (Clague 1989). At the peak of the last glaciation (Late Wisconsinan) about 17,000 years ago (Porter and Swanson 1998), the Cordilleran Ice Sheet covered all of what is now British Columbia, southern and central Yukon, and parts of the northwestern United States. Ice was thick enough to cover all but the highest peaks in southwestern British Columbia (Clague 1989). Glacial striations near the top of Handcar Peak, found during field work, agree with previous work indicating
Figure 15. Rocks and structures at Handcar Peak. A: Ductile shear zone in Hurley River quartz diorite (hammer blade points north). B: Example of quartz diorite intruding Pioneer Formation metavolcanic rocks.
that the maximum surface of the Cordilleran Ice Sheet was above 2400 m asl (Clague 1989). The build-up of the ice sheet to its maximum occurred over a period of at least 10,000 years (Clague, 1981), with ice advancing into coastal lowlands by about 30,000-25,000 years ago (Clague 1976, 1981) and retreating at least once before the peak of the Late Wisconsinan glaciation (Lian et al. 2001). Final decay of the ice sheet was relatively rapid, but was repeatedly interrupted by glacier stand-stills and readvances between about 15,000 and 12,000 years ago (Clague et al. 1997, Friele and Clague 2002). By 10,000 years ago, the extent of glaciers in the Coast Mountains was probably similar to that of today (Clague 1981).

### 3.3 Geomorphology

Lineaments and other deformational features at Handcar Peak extend across approximately 7 km of the southwest-facing slope between Railroad Creek and North Creek (Fig. 14). The valleys of Sampson Creek and Buck Creek cut through the deforming rock mass, dividing it into three main sections, termed the western, central, and eastern areas (Fig. 14). The deforming slope rises 1850-2080 m over a distance of 5-6 km from the floor of Lilooet valley to a maximum elevation of 2340 m asl at Handcar Peak and Locomotive Mountain. The average angle of the deforming slope is 22 degrees; the steepest portions, such as the southeast face of Handcar Peak, slope 30-35 degrees. The upper part of the deforming slope is relatively planar, whereas the base has irregular undulating benches. Sampson Creek is deflected 90 degrees where it meets these benches and flows laterally along the slope for 2 km. There is also an 800
m-wide bench at about 1900 m asl on the eastern section of the deforming slope, east of Buck Creek. There, gravitational lineaments extend to the upper limit of the bench, but the slope above it to Caboose Mountain and Locomotive Mountain shows minimal evidence of deformation. The peaks above the top of the deforming slope host several small, ice-free cirques – one on the north side of Handcar Peak, and three between the peaks above the eastern section of the DSGSD.

The classic morphological features of deep-seated gravitational movement displayed at Handcar Peak include gravitational lineaments (antislope scarps, normal scarps, tension cracks, grabens, and double ridges), closed depressions and ponds, toe bulging, rockfall, and surficial instabilities within the larger deforming mass (Fig. 16). These features were mapped before the start of fieldwork using 1:15,000-scale vertical aerial photographs. Field mapping of the central and eastern parts of the DSGSD was done in the summer of 2009. The section west of Sampson Creek was not visited due to the difficulty of accessing it and the few deformation features that occur there.

Of the 77 mapped gravitational lineaments, 54 are antislope scarps, 16 are trenches or tension cracks, two are normal scarps, and five are composite forms, i.e. a combination of the other three features. Heights of antislope scarps range from approximately 0.3 m to 12 m; most are about 1-5 m high. Antislope scarps are most common on the steeper parts of the slope; trenches and tension cracks occur near the top of Handcar Peak and on the wide bench east of Buck Creek (Fig. 17). The longest and straightest lineaments are oriented northwest-
southeast, slightly oblique to the strike of the slope, and plunge to the southeast.

The lowermost of these major lineaments is an antislope scarp that appears to be continuous across the valleys of Buck Creek and Sampson Creek, with a total length of 5 km (Fig. 14). Some of the shorter lineaments, especially the trenches and tension cracks on the bench east of Buck Creek, intersect, forming a laced pattern.

Most lineaments are weathered to the same degree as the surrounding terrain, implying a lack of recent activity. Exposed bedrock surfaces are covered

Figure 16. Map of gravitational lineaments, surficial instabilities, avalanche paths, and surface drainage at Handcar Peak.
with lichens, and vegetation covers veneers of colluvium or glacial sediments draping the lineaments (Fig. 17A, B). Sections of antislope scarps on the southwestern face of Handcar Peak have been eroded or levelled where they intersect major avalanche paths, presumably by repeated impact from snow avalanches over an extended period of time. A conspicuous lineament crosses Buck Creek valley but is covered by a Little Ice Age moraine on the west side of the valley. The crest of this moraine does not appear to be offset along the trace of the lineament (Fig. 18).

Some extensional features in the upper part of the deforming rock mass appear fresh. Figure 17D shows a trench with a steep tension crack at its base where loose gravelly material sloughing in from the sides of the trench has not yet reached its long-term angle of repose. In an unpublished field investigation in 1999, John Clague (personal communication, 2010) observed fresh open cracks in the sediments on the bottom of the lake shown in Figure 18. These features were not apparent during my field investigation.

Other instability-related surface processes and features include anomalous surface water flow paths, ponds, rockfalls, and small rockslides. Anomalous water flow occurs where stream paths are routed laterally across the slope along gravitational lineaments. Locally, water is ponded behind antislope scarps. Ponds are small and relatively uncommon, perhaps due to the tendency for lineaments to plunge slightly to the southeast and thus drain effectively. However, there is a significant pond behind a long and straight antislope scarp near the upper limit of the lineaments (Fig. 16). This pond is ~200 m long, 10 m
wide, and 2 m deep. Another pond is located in a graben-type feature in Buck Creek Valley (Fig. 18).

Figure 17. Lineaments at Handcar Peak. A and B. Major antislope scarps on the main slope; view southeast. C. Minor antislope scarps at the upper east corner of the deforming rock mass; view south (average height ~1 m). D. Trench with actively spreading tension crack at its base; view northwest. E. Large snow-filled ridge-top depression near the top of Handcar Peak, looking west.
Figure 18. Late Pleistocene glacial sediments displaced by gravitational lineaments at Handcar Peak. The sediments date to the end of the last glaciation, so the movement occurred after about 14,000 years ago (Clague 1981). The highlighted Little Ice Age moraine at the top right has not been displaced, thus little to no movement has occurred along these lineaments in the past several hundred years.
Rockfall and small rockslides are common around Handcar Peak (Fig. 16). Rockfall is concentrated below the steepened upper sides of gravitational lineaments on the southeast face of Handcar Peak, beneath a steep bulging zone below the lowermost major antislope scarp, and at the foot of a steep escarpment running along the east side of Buck Creek (“Buck Creek Escarpment” in Fig. 16). Six small rockslides were identified, all in areas of locally steep topography near antislope scarps or trenches. Two of the six rockslides originated on the steep uphill walls of major trenches and four occurred just below antislope scarps. The most recent rockslide (“fresh rockslide” in Fig. 16) involved the failure of a 50-m-wide section of an antislope scarp; the rock mass slid out on discontinuities that daylight on the steepened downhill-facing side of the scarp.

3.4 Trench investigation

A 6-m-long, 2-m-deep trench was excavated through the sediment fill behind one of the major antislope scarps near the east end of the deforming area (Fig. 19). The site was chosen based on the prominence of the scarp and the presence of a sediment fill that was fine enough to be excavated. Seven units, including four sediment units and three bedrock units (Table 3), were delineated by placing pins along contacts in one of the long vertical walls of the trench. A twine grid with rectangles 25 cm by 25 cm was placed over the wall to facilitate sketching of the geology on gridded line paper. Units were distinguished based on grain size, sorting, sedimentary structures, and colour. The detailed field sketch was later digitized (Fig. 20). Bulk samples of each unit were collected for
Figure 19. Photographs of trench. A: Finished trench (see Fig. 16 for location). B: Weathered bedrock and fault gouge at base of trench. C-1: weathered diamicton. C-2: unweathered diamicton. D: Sandy deposits in upper part of section.
Table 3. Summary of the properties of units logged in the trench at Handcar Peak.

<table>
<thead>
<tr>
<th>Unit</th>
<th>Texture</th>
<th>Moist colour(^1)</th>
<th>Clasts</th>
<th>Lower boundary</th>
<th>Other characteristics</th>
<th>Interpretation</th>
</tr>
</thead>
<tbody>
<tr>
<td>G</td>
<td>Sandy gravel</td>
<td>Brown (10YR 5/3)</td>
<td>75%; subrounded to subangular, fine to medium gravel</td>
<td>Sharp</td>
<td>Less than 1 m wide (laterally)</td>
<td>Lag deposit from seasonal runoff flow behind antislip scarp</td>
</tr>
<tr>
<td>F</td>
<td>Silty sand with gravel</td>
<td>Brown (10YR 4/3); upper horizon is dark yellowish-brown (10YR 4/6)</td>
<td>&lt;10%; subrounded to subangular fine gravel</td>
<td>Sharp</td>
<td>Contains grains of volcanic glass with feldspar inclusions, glass content is ~5% near lower boundary, ~20% in dark upper horizon</td>
<td>Fluvial sand and Bridge River tephra; weak paleosol at top of the unit</td>
</tr>
<tr>
<td>E</td>
<td>Silty sand with gravel</td>
<td>Dark brown (7.5YR 3/6)</td>
<td>&lt;30%; subrounded to subangular fine to coarse gravel</td>
<td>Gradational</td>
<td></td>
<td></td>
</tr>
<tr>
<td>D</td>
<td>Matrix-supported sandy, silty diamicton</td>
<td>Weathered: light yellowish brown (2.5Y 6/4); unweathered: light brownish grey (2.5Y 6/2)</td>
<td>40%; subrounded to subangular gravel, some cobbles and boulders</td>
<td>Sharp</td>
<td>Two distinctly different levels of weathering (see text)</td>
<td>In-situ till or colluvium derived from till</td>
</tr>
<tr>
<td>C</td>
<td>Clay</td>
<td>Red (10R 4/4)</td>
<td>Rare sand-size particles</td>
<td>Sharp to slightly gradational firm (S3)(^2)</td>
<td></td>
<td>Fault gouge</td>
</tr>
<tr>
<td>B</td>
<td>Matrix-supported sandy diamicton</td>
<td>Greenish-grey</td>
<td>45%; angular, gravel-size, composition similar to Unit A</td>
<td>Not visible</td>
<td>Very weak and altered; contains thin seams of pink clay</td>
<td>Weathered bedrock, damaged by tectonic and gravitational movement</td>
</tr>
<tr>
<td>A</td>
<td>Highly fractured, crushed rock</td>
<td>Dark green, red on fracture surfaces</td>
<td>N/A</td>
<td>Not visible</td>
<td>Very weak and altered</td>
<td>Weathered bedrock, damaged by faulting</td>
</tr>
</tbody>
</table>

\(^1\) Based on UCS classification scheme (ASTM D, 2000).  
\(^2\) Munsell Colour.  
\(^3\) From International Society of Rock Mechanics (1978) classification scheme.
Figure 20. Sketch of northwest wall of trench with interpreted geologic units.

grain size analysis (See Appendix 3). In addition, small samples of sediment were collected from five different levels within a sandy unit in the upper part of the trench (Fig. 19d) to locate a tephra that occurs within the area of Handcar Peak.

3.4.1 Stratigraphy

A dipping planar surface of highly fractured and weathered bedrock underlies the antislope scarp from the southwest end of the trench to the middle of the trench (Unit A in Fig. 20). Unit A is dark green where unaltered, but dark red on fracture surfaces. In places, the weathered bedrock has been disaggregated into a loose regolith of angular sand-size particles and larger clasts (Unit B) that probably was produced by movement along the antislope scarp. The weathered bedrock is capped by a layer of red clay approximately 5-
15 cm thick (Unit C) that separates along wavy internal shear planes, some of which display slicken lines. Unit C is interpreted to be gouge that delineates a fault with up to 2 m of dip-slip displacement, corresponding to slope-parallel displacement of the antislope scarp.

A light brownish-grey, massive, matrix-supported diamicton (Unit D) covers Units A, B, and C on the antislope scarp and thickens to the northeast. The diamicton consists of subangular to subrounded clasts up to boulder size in a matrix of dense sandy silt. The clasts are derived from local rocks, mainly the Hurley River Pluton. I interpret Unit D to be either lodgement till, colluvium derived from till and deposited by solifluction, or perhaps both. Considering the environment, it is likely that at least some diamicton has moved into the depression from the antislope scarp or from the opposite hillside. This process, together with erosion of the antislope scarp, may explain why Unit D thickens substantially in the middle of the depression. Interestingly, Unit D differs in character along the length of the trench. In the middle portion of the trench, it is fissured, loose, and weathered to a yellowish brown colour, and is penetrated to its base by abundant rootlets (Fig. 19c-1). In contrast, at the ends of the trench, it is unweathered, light brownish-grey, and very compact (Fig. 19c-2). The zone of weathered diamicton is above the fault plane; it likely was originally better consolidated, but was disturbed by movement along the fault.

The diamicton is overlain by a dark brown, loose, poorly sorted, silty fine to medium sand with subrounded gravel (Unit E). This unit is 5-10 cm thick and is restricted to the middle of the trench; it pinches out both against the antislope
scarp and the rising slope at the northeastern end of the trench. Unit E is probably a paleosol that developed on the glacial diamicton shortly after deglaciation.

Unit F overlies Unit E across a sharp contact. It comprises approximately 10 cm of brown, loose, moderately well sorted, silty fine to medium sand. The uppermost 2-4 cm of the unit is a dark yellowish-brown paleosol. Abundant shards of volcanic glass were found throughout Unit F, but are especially common in the paleosol at the top of the unit. The large size of the glass shards indicates that they could not have travelled far. Some shards contain inclusions of plagioclase. These two characteristics suggest that the glass is Bridge River tephra derived from an eruption of nearby Mount Meager about 2400 years ago (Nasmith et al. 1967, Clague et al. 1995). The tephra is reworked, thus Unit F could be younger than 2400 years old. The unit as a whole may record a rapid inwash of sand into the trench after the Mount Meager eruption.

A lens of poorly sorted, subrounded to subangular gravel of mainly granitic composition (Unit G) overlies Unit F. The lens is only 75 cm long and appears to be gravel deposited by water flowing along the depression behind the antislope scarp.

3.4.2 History of movement

Units E, F, and G extend up onto the antislope scarp and have been deformed by movement along the fault that coincides with the scarp. They are warped rather than faulted, suggesting that movement was gradual and perhaps continuous. If units E, F, and G were originally horizontal, approximately 0.5 m of
vertical displacement has occurred in the past 2400 years. A pre-existing scarp composed of bedrock as weak as units A and B would have been removed by glacial erosion during the last glaciation, therefore the other 1.5 m of displacement on the fault must have occurred between about 13,000 years ago (i.e., the time of deglaciation) and 2400 years ago. The average rate of displacement during the two different periods are similar, approximately 0.2 mm/yr. The movement, however, may have been episodic, rather than occurring continuously at a low rate.

I infer that movement on the fault delineating the antislope scarp is gravitational in origin. I considered the possibility that the movement could have occurred suddenly during one or more large earthquakes. McCalpin (2003) points out that scarps of questionable origin may have formed by: 1) displacement on tectonic faults, 2) gravity failures caused by earthquake shaking, or 3) gravity failures unrelated to tectonics. Indeed, the fault that created the antislope scarp was an active tectonic feature at some time in the past, because the clay gouge could not have been produced by near-surface gravitational displacements.

Several lines of evidence support a gravitational origin for this antislope scarp and other similar lineaments at Handcar Peak. First, the lineaments are more-or-less parallel and occur in sets, whereas tectonic fault scarps are generally individual features (McCalpin 2003). Second, most gravitational lineaments, like those at Handcar Peak, occur on the upper parts of high ridges and trend approximately parallel to the slope (McCalpin 2003). Third, there is
evidence for lengthy and continuing movement that is inconsistent with sudden, episodic displacement during earthquakes. Displacements on the trenched antislope scarp occurred over the entire Holocene, with about 0.5 m of displacement since 2400 years ago, and I observed fresh cracks in some of the nearby lineaments indicative of recent activity. Handcar Peak lies within a seismically active region, so the potential influence of Holocene earthquake shaking in the development of the gravitational lineaments cannot be ruled out. However, the fault plane exposed at the base of the trench is likely a relict feature that has been reactivated recently by the DSGSD. Other lineaments at Handcar Peak probably are similar features. I observed gouge and intense rock damage along two other major lineaments, and Lund (2002) described a zone of clay gouge and damaged rock in a trench that he excavated across an antislope scarp lower on the slope.

3.5 Geomechanical characterization

3.5.1 Data collection

I collected engineering geological data at eight sites on the surface of the deforming slope over a six-day period in the summer of 2010 – two sites in the central area near the top of Handcar Peak and six sites east of Buck Creek (Fig. 16). Only a small amount of data were collected in the central area, due to lack of time and the difficulty of the terrain.

Intact rock strength and rock mass quality were quantified to estimate rock mass strength parameters for numerical models. I recorded rock mass quality
based on an updated version of the Geological Strength Index (GSI; Hoek et al. 2002), which provides an objective evaluation of block size and joint surface conditions (Fig. 21; Cai et al. 2004). I characterized discontinuity surface roughness on two spatial scales: 1-10-m scale (primary roughness) using the qualitative terminology suggested by the International Society of Rock Mechanics (1978); and a 10-cm scale (secondary roughness) using joint roughness combs to estimate a Joint Roughness Coefficient (JRC; Barton and Choubey 1977). The weathering grade was determined qualitatively using the scheme of the Geological Society Engineering Group Working Party (1977), which defines six grades of weathering between “fresh” and “residual soil”. Lithology and relevant geologic structures were also noted. I estimated intact rock strength in the field using the geological hammer test (Hoek and Brown 1997), and I collected samples for point load testing to determine the unconfined compressive strength. The detailed criteria for my qualitative field descriptions are given in Appendix 1.

I characterized orientations of major geological structures and discontinuity sets for kinematic analysis and definition of numerical model geometry. I conducted discontinuity surveys by mapping joint sets in outcrops, supplemented with photogrammetry-based mapping at three sites (Fig. 16). Photographs for photogrammetric analysis were taken with a Canon Rebel XTi digital SLR camera at a focal length of 18 mm or 55 mm. I measured relative positions and orientations of the camera stations with a metric tape and a geologic compass. These measurements were later used to register the photogrammetry models (orient and scale them in three-dimensional space)
Figure 21. GSI chart (Cai et al. 2004, by permission) with estimated ranges of GSI values mapped at Handcar Peak in a) intrusive rocks and b) metavolcanic rocks.
using the method of Sturzenegger and Stead (2009). I created and registered models with the program 3DM CalibCam and mapped discontinuities with 3DM Analyst (ADAM Technology 2007). Figure 22 shows an example of a photogrammetry model and the registration method.

![Photogrammetry model](image)

**Figure 22. Photogrammetry model at site P02 with mapped discontinuity planes.** The sketch depicts the camera setup and data needed to register the model in 3DM CalibCam. α - azimuth of camera line of sight; β - dip of camera line of sight; x - distance between camera stations. α and β must be measured at each camera station.

### 3.5.2 Rock mass characterization

GSI values of the rock mass, based on visual estimates of the average block volume and discontinuity surface quality, range from 50 to 70 and average 60 (Fig. 23). Estimates of the GSI blockiness index range from “blocky” to “very
blocky”. GSI joint surface conditions are generally good to fair; the roughness ranges from undulating and rough to planar and smooth, but the degree of weathering everywhere is slight (Class II – Geological Society Engineering Group Working Party 1977). Joint spacing differs from outcrop to outcrop, with block sizes in the range $10^3$ - $10^5$ cm$^3$. Block shapes range from “blocky” (approximately equidimensional) to tabular (one dimension notably shorter than the other two; International Society of Rock Mechanics 1978). Overall, the differences in rock mass quality are minor and are not clearly related to lithology or location on the deforming slope. The walls of some major antislope scarps, however, have zones of concentrated rock damage 1-2 m thick, similar to the “weathered bedrock” unit observed in the trench study (Fig. 19b).

3.5.3 Discontinuity characterization

I noted some differences in discontinuity sets between the central and eastern areas of the deforming slope, but data from the central area are limited, thus my confidence in the structures mapped there is not high. Therefore, I focus on data from the eastern area in the following sections.

Joint sets identified from spot mapping are shown in Figure 24 and summarized in Table 4. Set 1 (JS1), which is the most prevalent, strikes northwest approximately parallel to the Owl Creek Fault and dips steeply into the slope. Its pole distribution is slightly bimodal; it is steeper and more northerly striking in rocks on the west side of the Owl Creek Fault than on the east side. Some of the joints in JS1 have very high persistence of $>20$ m (International Society of Rock Mechanics 1978).
Figure 23. Photographs of engineering geological mapping sites (A) 05 and (B) 08, showing the range of GSI values mapped at Handcar Peak. Person/stick figure for scale.
Set 2 (JS2) is a sub-vertical set with high persistence. It strikes south to southwest, providing lateral release for downhill-directed movement. Set 3 (JS3) is conjugate to JS1, but appears as two separate clusters of poles on the stereonet. The set dips 60° on average in metavolcanic rocks east of Buck Creek (JS3’ in Fig. 23A); in other areas, the dip is 30-45°. In the vicinity of the lowest antislope scarps, JS3 daylight as smooth and polished sliding planes in locally
steep outcrops where small rockslides have occurred. Set 4 (JS4) dips 20-40° south to south-southeast. Its low persistence of 1-3 m and rough, undulating joint surfaces make it less likely to form a sliding surface than JS3. Joint surfaces in all sets are slightly weathered (Class II of Geological Society Engineering Group Working Party 1977).

Table 4. Summary of characteristics of major discontinuity sets mapped at Handcar Peak.

<table>
<thead>
<tr>
<th>Set</th>
<th>Dip °</th>
<th>Dip direction°</th>
<th>Spacing¹</th>
<th>Strike persistence¹</th>
<th>Dip persistence¹</th>
<th>Primary roughness¹</th>
<th>Secondary roughness²</th>
</tr>
</thead>
<tbody>
<tr>
<td>JS1</td>
<td>60-85</td>
<td>30-45</td>
<td>Close - very wide (100-3000 mm)</td>
<td>Medium - very high (3-20 m)</td>
<td>Low - very high (1-20 m)</td>
<td>Variable</td>
<td>JRC 6-12</td>
</tr>
<tr>
<td>JS2</td>
<td>70-90</td>
<td>275-300</td>
<td>Moderate - wide (500 – 2000 mm)</td>
<td>Low – medium (1-10 m)</td>
<td>Low – high (1-15 m)</td>
<td>Planar moderate</td>
<td>JRC 6-10</td>
</tr>
<tr>
<td>JS3</td>
<td>30-60</td>
<td>230-250</td>
<td>Moderate – wide (300-2000 mm)</td>
<td>Low – medium (1-10 m)</td>
<td>Medium (4-10 m)</td>
<td>Planar smooth</td>
<td>JRC 4-8</td>
</tr>
<tr>
<td>JS4</td>
<td>20-40</td>
<td>160-180</td>
<td>Moderate – very wide (500 – 3000 mm)</td>
<td>Low – medium (1-3 m)</td>
<td>Low – medium (1-3 m)</td>
<td>Planar rough</td>
<td>JRC 8-10</td>
</tr>
</tbody>
</table>

¹ Qualitative descriptions based on guidelines in International Society of Rock Mechanics (1978).
² Joint Roughness Coefficient (JRC), after Barton and Choubey (1977).

The photogrammetric discontinuity data (Figs. 24B, 24C) do not correlate precisely with spot mapping data from the same sites, which highlights the biases in both methods. Fewer discontinuities in joint set 4 were identified by photogrammetric mapping, probably because their low persistence made them difficult to identify on the photographs. In addition, all visible joints were identified by photogrammetry, including some that might be considered anomalous; that is,
not part of a persistent set (e.g., set JS4’ in Fig. 24C). Subjective spot mapping data do not suffer from this problem, although they do not represent the full natural range of joint orientations as well the photogrammetric data.

3.6 Kinematic analysis

Large-scale toppling is not kinematically feasible at Handcar Peak (Fig. 24a), even if a low discontinuity friction angle of 25° is assumed to allow for pore water pressures. Sliding is possible under these conditions, although only on the shallowest discontinuities of joint sets 3 and 4 that daylight in the slope. In 10-50-m-high cliffs, where slopes can be as steep as 60°, toppling on JS1 and sliding on JS3 and JS4 are possible. Wedge failure involving JS3 and JS4 (not shown in Fig. 24) can also occur locally in steep outcrops. Field observations indicate that planar sliding on JS3 is likely the cause of the observed surficial rockslides.

Figure 24c shows that the Buck Creek escarpment, which trends perpendicular to the strike of the main slope and lineaments at Handcar Peak, is likely stable. Field observations support this conclusion, because mass wasting from the scarp appears to be limited mainly to rockfall.

Although this kinematic analysis is useful for explaining small surficial instabilities at Handcar Peak, it has limited application to deep-seated deformation of the rock mass. At the scale of the entire deforming rock mass, the slope is not steep enough to allow large-scale toppling along rock joints. Large-scale sliding is also unlikely as a primary mechanism of deformation, because although discontinuities daylight in steep outcrops, sliding on these surfaces only provides local, surficial kinematic release.
3.7 Numerical modeling of Handcar Peak

3.7.1 Numerical model setup

Numerical models can simulate several important aspects of complex rock slope movements that traditional limit equilibrium or kinematic analyses ignore (Stead et al. 2006). Rather than assuming a pre-defined failure geometry, they allow movement to occur naturally, based on a given slope structure and rock strength properties (Lorig and Varona 2004). They also account for the effect of stress concentrations induced by slope geometry and the tectonic stress field (Kinakin and Stead 2005, Ambrosi and Crosta 2006; Li et al. 2010). Whereas simpler analytical tools can only indicate the potential of a discrete failure event, numerical models can simulate patterns of ongoing deformation.

I chose the program UDEC (Universal Distinct Element Code, Itasca Consulting Group 2004) to model the Handcar Peak DSGSD. The high intact rock strength and organized discontinuity structure at Handcar Peak dictate the use of a discontinuum approach such as UDEC that models the interactions between discrete individual blocks. In addition, UDEC’s explicit time-marching solution scheme can model progressive movement that accumulates displacements over time, but does not necessarily progress to catastrophic failure.

I created a model slope using a profile that extends southwest from the top of Caboose Mountain to the floor of Lillooet Valley (Fig. 14). To avoid boundary effects, I extended the model to the sides and below the profile according to the recommendations of Lorig and Varona (2004). The northeastern
The slope of Caboose Mountain was included in the profile to ensure that stresses in the model are realistic, because ridge shape has an important influence on stress distribution (Kinakin and Stead 2005). Figure 25 summarizes the design of the model, based on data collected at the engineering geological field mapping stations. Joint sets 1 and 3 dip approximately parallel to the azimuth of the profile and so were included in the model. The average dip of JS1 is nearly the same everywhere, but the dip of JS3 differs along the profile. Major antislope scarps mapped at the surface were represented in the model as persistent, low-strength fault planes.

![Figure 25. UDEC model structure based on locations of major gravitational lineaments and joint set orientations mapped in the field.](image)
For large slopes such as Handcar Peak, geological structure must be considerably simplified and idealized in models to accommodate limitations in processing power and data availability. Consequently, I used a wide discontinuity spacing and assigned material properties to the deformable blocks that are representative of a continuous jointed rock mass. I used an elastoplastic Mohr-Coulomb constitutive model for the rock mass and a Mohr-Coulomb area contact criterion to model slip on discontinuities (Itasca Consulting Group 2006).

Boundary conditions in the initial state of the model were set to simulate conditions at the peak of the last glaciation (Fig. 26). I covered the profile with ice to just above the top of Handcar Peak. I imposed a zero velocity condition on the lateral boundaries in the x direction and fixed the lower boundary in the y direction. Initial vertical stresses ($\sigma_{xx}$) within the model were calculated from the cover of ice and rock:

$$\sigma_{xx} = (\rho_i h_i) + (\rho_r h_r)$$

where $\rho$ and $h$ are, respectively, the density and height of the materials above a given model element. In some models, horizontal stresses ($\sigma_{yy}$) were calculated using a horizontal-to-vertical stress ratio (K) of 0.5 to explore the assumption of in situ stresses in the region. In other models, horizontal stresses were induced gravitationally based on the Poisson effect for an isotropic material:
\[
\frac{\sigma_{xx}}{\sigma_{yy}} = \frac{\nu}{1 - \nu}
\]

where \(\nu\) is Poisson’s ratio (here assumed to be 0.25), resulting in a horizontal-to-vertical stress ratio of \(\approx 0.33\).

Figure 26. UDEC model initial setup, depicted with wide joint spacing to improve visibility.

I simulated debuttressing of the slope during deglaciation by removing layers of ice from the 2000-m-deep valley in five stages, solving after each stage to equilibrate stresses (Fig. 26). Rock mass and joint strength properties were kept high during this process to prevent excessive yielding before the stresses reached equilibrium. Once the model was in equilibrium, I lowered strength properties of the rock mass and discontinuity surfaces and ran the model again to observe the resulting pattern of movement. I assumed lower discontinuity strength and rock mass properties for the upper 500 m of rock than for the rest of the model to account for near-surface weathering and rock mass damage (Fig. 26).

Because the study was very data-limited, I attempted to understand the effect that different assumptions had on model results. In addition to trying
different stress regimes, I tested a variety of configurations of model structure and strength properties (Table 5). I also compared the results of different model runs to field observations to determine the most realistic configuration.

The Mohr-Coulomb rock mass properties were obtained using the program RocLab (Rocscience 2007), which employs a statistical curve-fitting process to derive a Mohr-Coulomb envelope from known and estimated parameters of the rock mass. Table 6 summarizes the input parameters required for RocLab and their sources in this study. Table 7 shows the rock mass strength properties obtained as output. Slope height is required as an input parameter in RocLab to estimate expected average stresses in the slope, because two of the derived rock mass strength properties – friction angle and cohesion – are stress-dependent. I used two values of slope height: 500 m to derive properties for the near-surface weathered portion of the rock mass, and 2000 m for the rest of the model (Fig. 26).

I chose discontinuity strength properties based on published values for rock joints and gouge-filled fault planes in rocks of similar lithology to those at Handcar Peak (Table 8, Barton et al. 1974, Kulhawy 1975, Barton 1991). In some model runs, I assumed a very low effective friction angle to explore model response to hypothetical conditions of high pore water pressure at the time of deglaciation and during seasonal snowmelt.
Table 5. UDEC model configurations that were analysed for sensitivity to different modelling assumptions.

<table>
<thead>
<tr>
<th>Joint configuration</th>
<th>Stress ratio</th>
<th>Discontinuity frictional strength</th>
<th>GSI (weathered/competent rock mass)</th>
</tr>
</thead>
<tbody>
<tr>
<td>JS3 fully persistent</td>
<td>Gravitational</td>
<td>Moderate</td>
<td>60/70</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Low</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>Very Low</td>
<td>60/70</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>50/60</td>
</tr>
<tr>
<td></td>
<td>K=0.5</td>
<td>Moderate</td>
<td>60/70</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Low</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>Very Low</td>
<td>60/70</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>50/60</td>
</tr>
<tr>
<td>JS3 not fully persistent</td>
<td>Gravitational</td>
<td>Moderate</td>
<td>60/70</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Low</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>Very Low</td>
<td>60/70</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>50/60</td>
</tr>
<tr>
<td></td>
<td>K=0.5</td>
<td>Moderate</td>
<td>60/70</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Low</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>Very Low</td>
<td>60/70</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>50/60</td>
</tr>
</tbody>
</table>

Table 6. RocLab input properties and their sources.

<table>
<thead>
<tr>
<th>RocLab input</th>
<th>Value</th>
<th>Source of data</th>
</tr>
</thead>
<tbody>
<tr>
<td>Intact uniaxial compressive strength (sigci)</td>
<td>275 MPa</td>
<td>Averaged from point load tests</td>
</tr>
<tr>
<td>GSI</td>
<td>50/60/70</td>
<td>Field estimates</td>
</tr>
<tr>
<td>Hoek-Brown constant (m_i)</td>
<td>25</td>
<td>RocLab(^1)</td>
</tr>
<tr>
<td>Unit weight</td>
<td>2.7 g/cm(^3)</td>
<td>Measured in laboratory</td>
</tr>
<tr>
<td>Slope height</td>
<td>500/2000 m</td>
<td>Google Earth</td>
</tr>
</tbody>
</table>

\(^1\) RocLab provides typical values of these properties for different lithologies.
Table 7. Rock mass strength properties derived in RocLab.

<table>
<thead>
<tr>
<th>Rock mass properties(^2)</th>
<th>GSI 50</th>
<th>GSI 60</th>
<th>GSI 70</th>
</tr>
</thead>
<tbody>
<tr>
<td>(\rho) (g/cm(^3))</td>
<td>2.7</td>
<td>2.7</td>
<td>2.7</td>
</tr>
<tr>
<td>Bulk modulus (GPa)</td>
<td>23</td>
<td>38</td>
<td>54</td>
</tr>
<tr>
<td>Shear modulus (GPa)</td>
<td>14</td>
<td>23</td>
<td>32</td>
</tr>
<tr>
<td>Poisson’s ratio</td>
<td>0.25</td>
<td>0.25</td>
<td>0.25</td>
</tr>
<tr>
<td>Tensile strength (KPa)</td>
<td>1000</td>
<td>1500</td>
<td>2000</td>
</tr>
<tr>
<td>(\phi)(^\circ)</td>
<td>53(^1)</td>
<td>55(^1) / 46(^2)</td>
<td>48(^2)</td>
</tr>
<tr>
<td>Cohesion (MPa)</td>
<td>5.3(^1)</td>
<td>6.5(^1) / 14.5(^2)</td>
<td>17(^2)</td>
</tr>
<tr>
<td>Dilation angle ((^\circ))</td>
<td>5</td>
<td>5</td>
<td>5</td>
</tr>
</tbody>
</table>

\(^1\) Based on stresses in a 500-m-high slope.  
\(^2\) Based on stresses in a 2000-m-high slope.

Table 8. Range of discontinuity properties used in UDEC models.

<table>
<thead>
<tr>
<th>Discontinuity properties</th>
<th>Rock joints</th>
<th>Relict faults</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Moderate strength</td>
<td>Low strength</td>
</tr>
<tr>
<td>Normal stiffness (GPa/m)(^1)</td>
<td>10</td>
<td>1</td>
</tr>
<tr>
<td>Shear stiffness (GPa/m)(^1)</td>
<td>1</td>
<td>0.1</td>
</tr>
<tr>
<td>(\phi) (^\circ)(^2)</td>
<td>30</td>
<td>25</td>
</tr>
<tr>
<td>Cohesion (KPa)(^2)</td>
<td>100</td>
<td>50</td>
</tr>
<tr>
<td>Tensile strength (KPa)</td>
<td>0</td>
<td>0</td>
</tr>
</tbody>
</table>

\(^1\) Based on data published by Barton (1991).  
\(^2\) Rock joint cohesion and friction based on data published by Kulhawy (1975); fault cohesion and friction based on data from Barton et al. (1974).

3.7.2 Numerical modelling results

3.7.2.1 Stress ratios

Of the factors that I assessed, the method of calculating horizontal stresses and the resulting stress ratio have the most significant effect on model behaviour (Fig. 27). With horizontal stresses induced purely by gravity (Eq. 2),
the zone of horizontal movement reaches deep into the model, lower even than
the toe of the slope. Both sides of the ridge move southward and downward
towards Lillooet Valley. The rate of displacement decreases gradually over time,
although it does not reach zero.

When *in situ* stress with $K = 0.5$ is assumed, movement is confined to the
upper south side of the ridge, above the steepest part of the slope. The
magnitude of total displacement in this model is less than half that in the model
using gravity-induced stresses, even on the most active part of the slope. The
rate of displacement also decreases more rapidly with numerical steps and
approaches equilibrium by the end of the model run. The more limited range and
magnitude of movement in the tectonic stress model may be due to the inhibition
of slip on joint surfaces and faults by increased confining pressures, especially at
depth.

In both models, the rock mass does not yield because the stresses
involved are low compared to rock mass strength. The movement histories of
different points on the slope surface in both models also trend towards
equilibrium, although slowly in the model using gravity-induced stresses. Of the
two models, I chose the tectonic stress model because it agrees better with the
tectonic setting of Handcar Peak: *in situ* stresses are expected in the Coast
Mountains because the region is being actively compressed (Monger and
Journeay 1994). Researchers who have conducted previous modelling studies in
the region have also assumed a stress ratio of $K = 0.5$ (Bovis and Stewart 1998,
Stewart and Ripley 1999).
Figure 27. Comparison of model response (horizontal displacement contours) for two different scenarios of horizontal stress. Both models use the persistent JS3 joint configuration and low discontinuity strength properties.

3.7.2.2 Joint configurations

Model responses with the two different joint configurations are similar (Fig. 28). If JS3 is not fully persistent (Fig. 28A), displacement of the rock mass is low due to the kinematic confinement of the joint geometry. Under this condition, only the longest fault planes on the steepest part of the slope form antislope scarps at the surface. Movement decreases gradually with depth, generally ceasing about
1500 m below the slope surface. If JS3 is assumed to be fully persistent (Fig. 28B), the pattern and depth of movement are similar, but displacement of the rock mass is greater. Surface morphology is reproduced more accurately – all of the steep fault planes form antislope scarps, although much less displacement occurs along the shorter faults. This joint configuration may more accurately simulate the ability for movement to take place by step-path linking of small-scale fractures.

Figure 28. Comparison of model results for different joint configurations using very low joint strength values and k = 0.5. A: JS3 not fully persistent; B: JS3 fully persistent.
3.7.2.3 Strength properties

Figure 29 shows model behaviour for three different combinations of discontinuity and rock mass strength properties. Assuming low discontinuity frictional strength and a GSI of 60 for the near-surface rock mass, total displacement is small (maximum 0.5 m), and the only fault plane that forms an antislope scarp at the surface is the lowermost one, just above the steepest part of the slope (Fig. 29A). If discontinuity effective frictional strength is lowered to values representative of high pore water pressure conditions, slip occurs along the other steep fault planes, additional antislope scarps form, and total displacement more than doubles (Fig. 29B). The effect of lowering rock mass strength is less pronounced: displacement magnitude increases slightly, but the distribution of movement is similar (Fig. 29C). The displacement histories of the three models show a similar pattern, trending towards equilibrium after approximately the same number of model time steps.

The value of intact unconfined compressive strength (UCS) used as input to calculate rock mass strength properties in RocLab (275 MPa), although supported by field observations and point load tests, is larger than most published values for crystalline intrusive rocks (Kulhawy 1975). A series of models was run to explore the effect of using a lower intact UCS value - 200 MPa. Model results were similar to those obtained with a UCS value of 275 MPa (Fig. 30). Total displacement was slightly higher, as would be expected, but the pattern of displacements was similar and the models also trended quickly towards equilibrium.
Figure 29. Comparison of model results using different joint and rock mass strength properties in the upper and lower rock mass. A: Low joint strength properties. B: Very low joint strength properties. C: Very low joint strength properties and lower GSI. Block displacements magnified 50x.
3.7.3 Evaluation of model behaviour

The behaviour of the UDEC model with tectonic stresses ($K = 0.5$), fully persistent joints in JS3, and very low discontinuity frictional strength is consistent with field observations at Handcar Peak. Antislope scarps form where the weak, steeply dipping fault planes intersect the slope (Fig. 31). Field observations of geomorphic features support the hypothesis that movement is concentrated in
this part of the slope; the highest antislope scarps occur there, as do most of the mapped rockfalls and small rockslides. The tendency of the model toward equilibrium is also supported by field data, including the lack of visible signs of modern activity on the tall lower antislope scarps and the low average rate of displacement over the past 2400 years (<1 mm/year) estimated from warped trench sediments.

Figure 31. Comparison of antislope scarps formed in the UDEC model to field observations (model run with $K = 0.5$, JS3 assumed to be fully persistent, and very low discontinuity strength).
The UDEC model, however, does not simulate extensional surface features such as tension cracks and trenches (Fig. 17D) on the wide upper bench of the slope. The steeply dipping faults and joints in this part of the model do not show separation; in fact, some minor thrusting is apparent on JS3 (circled area in Fig. 31). This problem also exists in the models with gravity-induced horizontal stresses. Compression on the bench may relate to the small amount of total displacement produced in the model, particularly on the steep part of the slope below the bench. This portion of the slope acts as a passive buttress, preventing the rock mass above it from relaxing towards the valley and relieving its locked-in horizontal stresses.

### 3.8 Discussion

The application of an integrated methodology based on the rationale presented in Chapter 2 provides a plausible characterization of the Handcar Peak DSGSD. The three main techniques used—geomorphic mapping, trench sediment analysis, and geomechanical analysis—complement one another and contribute to an understanding of key attributes of the DSGSD, including the limits of the deforming rock mass, the history of movement and current state of activity, and the mechanics driving rock mass deformation.

#### 3.8.1 Limits of the deforming rock mass

The Handcar Peak DSGSD lacks the typical morphologic features that are used to define the upslope and lateral limits of most landslides. The upper and lateral limits of DSGSD are best defined by surface deformational features—
lineaments, ponds, and surficial instabilities. These features delimit the presently or formerly active portion of the slope. The locations of lineaments also provide a constraint on the depth of the deforming rock mass. The antislope and normal scarps that displace late Pleistocene deposits in the valley of Buck Creek (Fig. 18) extend to the top of the 100-200-m-high scarp that forms the east margin of the valley (Buck Creek escarpment in Fig. 16). This observation indicates that movement along the fault planes forming these scarps has occurred to a minimum depth of 200 m in the rock mass. In fact, the geometry of the slope probably requires that the lower limit of the deforming rock mass extend substantially deeper, to depths of 400-500 m (Fig. 32).

The downslope limit of rock mass deformation at Handcar Peak is more difficult to define. Typical DSGSDs have a bulging area at their toe (Savage and Varnes 1987, McCalpin 2003). Bovis and Evans (1996, p. 6) comment “Below scarp 1 [the lowest antislope scarp] are two prominent taluses associated with steep, downslope-bulging masses of what appear to be dilated, disintegrating rock…” No gravitational lineaments were identified in the field or on aerial photographs below this subtly bulging area, thus it may represent the lower limit of the deforming rock mass. However, the large irregular bench at the base of the slope, which displaces Sampson Creek, also appears to bulge outward, and its location corresponds with the section of the valley side that has linear surface features near the ridge top. These observations indicate that the deforming area may extend to the valley floor, as has been documented in other studies of
DSGSD (Agliardi et al. 2001, Ambrosi and Crosta 2006, Hurlimann et al. 2006, Li et al. 2010, Weißflog et al. 2010). Figure 32 depicts both possibilities.

![Diagram](image)

**Figure 32. Hypothesized locations of the downslope limit of the deforming rock mass at Handcar Peak. A: Deformation extends to the toe of the slope. B: Deformation ends at the bulging area just below the lowest antislope scarp.**

Of these two scenarios, it is more likely that the downslope limit of deformation is just below the lowest antislope scarp than at the toe of the slope. The large irregular bench at the base of the slope that displaces Sampson Creek can be interpreted as a relict linear ridge produced by differential glacial erosion of fractured bedrock rather than a gravitational feature. Other than this bulging topography, no DSGSD surficial features were detected in the toe area of the slope. In published studies of DSGSD that do extend to the valley floor, lineaments and surficial instabilities have been noted near the toe of the
deforming slope (Agliardi et al. 2001, Hurlimann et al. 2006, Weißflog et al. 2010). However, there are rockslides and rockfalls in the upper bulging area, just below the lowest antislope scarp (Figs. 16 and 32). Modeling results support the conclusion that the deforming rock mass does not extend to the toe of the slope; the limit of the deforming area in the numerical models that use tectonic stresses (Fig. 27) corresponds well with the bulging area below the lowest antislope scarp (Fig. 32B).

3.8.2 History of movement and current activity

The trenched sediments suggest a pattern of slow and steady slope displacements throughout the Holocene. This history, however, may not apply to the entire Handcar Peak DSGSD; different portions of the slope may have moved at different times. Although a full reconstruction of movement history is not possible without more trenching, other field observations offer some insight. East of Buck Creek, the large lower antislope scarps are weathered and appear inactive. They contrast with the fresher surface features in the benched upper part of the deforming slope (Fig. 17), suggesting that movement has retrogressed up the slope over time. Moderate initial displacements on the steep middle section of the slope would remove support from the upslope rock mass, permitting a gradual relaxation of the bench that has continued to the present. Less information is available on the movement history of the middle section of the deforming slope, west of Buck Creek. The destruction of the lower antislope scarps by avalanches and rockfalls, however, implies a lack of recent activity on this part of the slope. Presumably the steeper angle and greater relief of the local
slope would allow the rock mass to adjust more rapidly along the entire slope profile.

The current episode of movement at Handcar Peak was likely triggered by glacial undercutting of the toe of the slope during the last Pleistocene glaciation and, later, by debuttressing of the slope during deglaciation (Fig. 33). The linear features, at least, postdate the last glaciation, because they would likely have been erased by glacial erosion. However, it is also possible that the slope deformed prior to the last glaciation. I found striations 3 m below the top of the wall bounding a large gravitational trench on Handcar Peak (Fig. 17E), at a location where no glaciers existed in the Holocene. That trench wall, therefore, must have already been exposed when ice overrode Handcar Peak during the last glaciation.

![Figure 33. Timeline of relative activity of Handcar Peak DSGSD. Darkness of bars is proportional to inferred relative rate of activity (darker = more active).]

3.8.3 Movement mechanism

The mechanism driving movement at Handcar Peak, as indicated by geomorphic features, kinematic analysis, and numerical modeling, does not seem to involve simple failure along a discrete basal shear surface. The lack of lateral and crown rupture surfaces suggests a gradational transition between the
deforming and stable parts of the rock mass. The base of the unstable rock mass could be a series of dispersed shear zones that grade downward into more competent rock (Stewart 1997). The increase in confining pressures with depth may also play a role in defining the lower limit of movement. Movement that occurs on low-persistence joints with a variety of orientations may die out at the depth where confining pressures mobilize sufficient frictional strength to prevent slip on discontinuities.

Steeply dipping, persistent, weak fault planes apparently play a significant role in facilitating deformation of the rock mass. Displacements in the UDEC models are concentrated on the same part of the modelled slope as the fault planes, particularly on the steepest portion of the slope below the wide upper bench (Fig. 28). The effect of the fault planes is highlighted by comparing models run with and without them (Fig. 34). The rock mass deforms without the fault planes, but the total displacement is less and lineaments do not form at the surface because slip does not occur along the steeply dipping discontinuities. Slip on the downhill-dipping discontinuity set JS3 is also clearly important in driving movement. Little displacement occurs in UDEC models where JS3 is not fully persistent (Fig. 29). Field evidence of sliding on JS3 includes the graben features in the benched upper part of the deforming slope and surficial rockslides in the steep lower area.
Figure 34. UDEC models showing the importance of weak fault planes; block deformation is shown at upper right of each model. Above: Low-friction fault planes ($\Phi = 12^\circ$). Below: Fault planes with the same properties as rock joints ($\Phi = 20^\circ/25^\circ$).

Figure 35 shows a conceptual model for a hypothesized complex mechanism of movement at Handcar Peak, in which JS3 forms the base of an active upper block in an active-passive system. The rock mass in the upper portion of the slope drives the steep portion of the slope forward, and the passive toe of the deforming area acts as a buttress. Glaciers may have formed in the cirques near the top of the slope (Fig 14) during the Late Holocene, adding mass to the active upper block. Compression and bulging of the toe, facilitated by thrust slip on JS3, provide limited kinematic freedom. The weak fault planes aid
extension in the upper part of the slope by acting as normal faults. Block-flexural toppling occurs on the steep part of the slope, where toppling columns are sections of the jointed rock mass between persistent weak fault planes. The active-passive mechanism requires some rock mass damage in the toe of the slope due to shearing or tensile fracture to provide a base for the passive block. This assumption is supported by reports of crushed or sheared rock in previous studies of DSGSD in British Columbia where borehole data are available (Stewart and Ripley 1999, Stewart and Moore 2001).

The future behaviour of the Handcar Peak DSGSD is difficult to predict based on the limited amount of data gathered in this study. Field observations and numerical modelling results, however, indicate a deformational mechanism that is currently in a state of equilibrium, even assuming conservative values of rock mass and discontinuity strength. Previous studies have shown that massive landslides associated with DSGSD exploit existing structures that allow kinematic modes of movement. The Hope and Randa landslides, for example, occurred on slopes with daylighting discontinuities that helped to form a basal sliding plane (Evans and Couture 2002, Eberhardt et al. 2004). The slow-moving La Clapiere landslide is apparently driven by toppling on sub-vertical discontinuities (Guglielmi et al. 2005). At Handcar Peak, simple sliding and toppling do not appear feasible on a large scale, given the slope geometry and overall joint structure. In addition, the rock mass strength is relatively high in comparison to near-surface stresses. A strong rock mass can degrade through progressive microfracturing and destruction of rock bridges (Kemeny 2003, Eberhardt et al.)
2004), but in such cases slope failure is generally preceded by an increase in activity and fresh signs of deformation such as tension cracks at the top of the slope (Moss et al. 2006, Petley and Petley 2006, Glastonbury and Fell 2010). Field observations indicate that the Handcar Peak DSGSD is mostly inactive at present.

![Figure 35. Conceptual model of the complex deformation mechanism at Handcar Peak.](image)

### 3.9 Conclusions

The extensive network of linear surface features at Handcar Peak is characteristic of deep-seated gravitational slope deformation. Long and straight antislip scarps are the surface expression of displacement along weak, persistent fault planes. The maximum extent of the deforming rock mass corresponds to the area in which DSGSD-related geomorphic features –
antislope scarps, trenches, rockfalls, and small rockslides – occur. The thickness of the deforming rock mass is more poorly constrained, but geomorphic evidence and numerical model results indicate values in the range of 200-500 m.

The current episode of movement probably began during Late Pleistocene deglaciation when the oversteepened valley sides were debuttressed. Earlier phases of deep-seated gravitational slope deformation, however, are likely. Geomorphic evidence indicates that the displacement has been regressive; the tall lower antislope scarps are now apparently inactive, whereas the upper part of the slope continues to deform. Movement of the upper portion of the DSGSD has been slow and relatively steady throughout the Holocene, with 0.5 m of movement occurring on one lineament in the past 2400 years. Field evidence is consistent with numerical modelling results in indicating that the DSGSD was probably most active soon after deglaciation and is presently near equilibrium.

The existing joint network is unfavourable for simple sliding or toppling failure. Distinct element modelling with the program UDEC highlights the importance of in situ tectonic stresses, joint configuration, and the relict fault planes in determining the extent, magnitude, and temporal evolution of displacement. A complex active-passive mechanism is indicated, one involving extension by normal slip on discontinuities in the active upper portion of the deforming rock mass and compression in the passive toe. Displacements are preferentially accommodated along the weak faults.

The conclusions reached in this study should be considered preliminary for several reasons. Numerical modelling results are based on sparse data,
particularly with respect to the subsurface deforming rock mass. Subsurface joint and fault orientations in numerical models are extrapolated from surface observations, and differences in this structure can have a large impact on slope kinematics. Other unseen structures such as shear zones or dikes could potentially have a large impact on slope stability. Groundwater conditions at the site also remain unknown. The potential effect of pore water pressures on slope stability could only be simulated by proxy in numerical models, by using low values of discontinuity shear strength. Seismic loading was not considered in the numerical analysis; nor was potential for progressive weakening of the rock mass over time due to strain softening.

The reliance on mostly qualitative data to interpret the state of activity at Handcar Peak is another significant limitation of this study, both for assessing the current hazard and for predicting the future evolution of the slope. Surface displacement monitoring using remote sensing techniques such as InSAR, which could quantify movement across the entire deforming slope, would aid in defining the limits of the unstable rock mass and in constraining the behaviour of numerical models. Finally, I did not evaluate the mechanics of movement in the central portion of the Handcar Peak DSGSD in this study. Available field data indicate that this section of the slope is similar in many ways to the eastern section, which was studied; however, it is both steeper and higher.
4: SYNTHESIS

4.1 Comparison of Handcar Peak study to other work

Deep-seated gravitational slope deformation is a complex process that occurs in many different lithological and structural environments. Comparison of individual case studies thus is difficult. The Handcar Peak DSGSD shares some attributes with other studied DSGSDs, but it is different in many important ways. The methodology I used in this study addressed the site-specific challenges at Handcar Peak that were evident from a review of previous work and air photo interpretation. A combination of geomorphic mapping, study of trench sediments, and numerical modelling proved effective for characterizing the DSGSD. A literature survey revealed techniques that could improve understanding of the Handcar Peak DSGSD and, more generally, the phenomenon.

The types of linear surface features found at Handcar Peak (antislope scarps and trenches) are typical of those of most deep-seated gravitationally deforming slopes. The length and arrangement of the scarps at Handcar Peak, however, are exceptional. Lineaments formed by gravitational processes are typically short – tens to hundreds of metres in length – and discontinuous (McCalpin 2003). Much longer, straight lineaments have formed by gravitationally driven movement along relict fault planes, such as at Mount Currie (Evans 1987, Bovis and Evans 1995, Thompson et al. 1997) and Hell Creek (Clague and Evans 1994). However, only one lineament is present at Mount Currie and the
Hell Creek slope is dominated by a single long lineament, whereas Handcar Peak displays an array of lineaments that are 1-3 km long. Lineaments 1-2 km length have been reported in the Alps (Ambrosi and Crosta 2006, Reitner and Linner 2009), but do not occur in such straight parallel sets. Two site-specific factors may explain the unusual length and pattern of lineaments at Handcar Peak. First, a pervasive structural fabric associated with the Owl Creek Fault controls the orientation of the fault planes that form the antislope scarps. Because the Owl Creek Fault is a dispersed ductile shear zone (Riddell 1992), it affects the rock fabric across the entire 2-km width of the deforming slope, rather than just along one discrete fault plane such as at Mount Currie. Second, the Handcar Peak slope is planar, which emphasizes the apparent straightness of the lineaments. Discontinuity planes that form antislope scarps generally dip and thus curve where they cross irregular topography. For example, antislope scarps on a gravitationally deforming slope in the Austrian Alps are not straight in plan view because they cross an irregular surface (Reitner and Linner 2009).

The history of movement at Handcar Peak is similar to that inferred for other DSGSD in glaciated mountain landscapes. A major similarity is activity during or shortly after deglaciation. Factors that favour slope deformation at that time include glacier erosion, debuttressing, frost wedging, and a high water table (Simmons and Cruden 1980, Bovis 1982, Stepanek 1992). Carbon dating of organic material in trench sediments indicates that a period of 2-4 ka separates deglaciation and the initiation of movement on some slopes, which has been

Unfortunately, no suitable material for carbon dating was found in the trench fill excavated in this study, and the time of the initiation of movement is only constrained to the interval between deglaciation and deposition of the Bridge River tephra 2400 years ago. Numerical modelling, however, indicates that high pore water pressures, like those that would have been present during deglaciation, were probably necessary to trigger slope movement at Handcar Peak by lowering the effective frictional strength of discontinuity surfaces. This interpretation implies that movement began during deglaciation rather than several thousand years later. Gravitational scarps that apparently existed prior to the peak of the last glaciation have also been noted in both this study and that conducted at Mount Currie by Thompson et al. (1997), suggesting some pre-Holocene activity in the Lillooet Valley.

The low rate of activity inferred at Handcar Peak is similar to that documented at a site in Colorado by McCalpin and Irvine (1995). They found that the rate of slip on an antislope scarp decreased over a period of 11,000 years. Other reconstructions of DSGSD activity indicate a higher and steady rate of movement (Thompson et al. 1997, Gutierrez-Santolalla et al. 2005) or increasing rate of movement (El Bedoui et al. 2009).

Many DSGSDs, including that at Handcar Peak, contain zones of greater or lesser activity. Ambrosi and Crosta (2006), for example, documented a spatially variable pattern of movement on two deforming slopes in Italy using PS-
InSAR. In some cases, discrete landslides develop within a gravitationally deforming slope; these can be slow, as in the case of the Ruinon (Agliardi et al. 2001) and La Clapiere (Bedouin et al. 2009) landslides, or rapid, as in the case of the Randa rockslide (Jaboyedoff et al. 2004). However, these three landslides occurred at the oversteepened toe of their deforming slopes. The movement pattern at Handcar Peak, where the flatter upper part of the deforming slope is apparently the most active, is more likely an effect of the gradual upslope migration of stress release. In the case of previous large rockslides on slopes with a similar structure to that at Handcar Peak, kinematic release at the toe of was necessary before a large landslide could occur, and slope failure was preceded by rock noise, surficial instabilities, and cracking over a period of months to several years (Glastonbury and Fell 2010).

Geomechanical interpretations of DSGSD behaviour differ, as do the techniques used to evaluate hypothesized movement mechanisms. Finite element or finite difference continuum modelling is typically used to analyze sites where no clear kinematic mode of failure exists (Stewart and Ripley 1999, Agliardi et al. 2001) or where deformation is thought to be due to plastic behaviour of weak geologic materials (Radbruch-Hall et al. 1976, Schultz-Ela and Walsh 2002). Continuum modelling has been used to predict deformation of a large slope based on the zones of stress concentration. I used a hybrid approach in this study – a discontinuum-based, distinct element model with plastic deformable blocks. This approach has been used to model discrete kinematic modes of failure, such as rock slumping (Kinakin 2004), flexural toppling
(Pritchard and Savigny 1991, Nichol et al. 2002), and block-flexural toppling (Bovis and Stewart 1998, Tosney et al. 2004). In this study, I used the distinct element method primarily because the geological materials have high strength relative to near-surface stresses and because the method is capable of reproducing gravitational lineaments at the surface.

Toppling has been cited as the primary mode of movement in previous studies of slopes with a discontinuity structure similar to that at Handcar Peak and that exhibit prominent antislope scarps (Bovis 1982, Holmes and Jarvis 1985, Tosney 2004, Reitner and Linner 2009). Rather than simple toppling, I propose a complex active-passive mechanism of movement at Handcar Peak, involving slip on downhill-dipping joints and steeply dipping weak fault planes, as well as tensile fracturing and shearing near the toe of the deforming rock mass. Antislope scarps form on weak fault planes, both through normal slip and block-flexural toppling. Many of these characteristics have been noted in the literature at other DSGSD sites. The importance of pre-existing fault planes in forming gravitational lineaments is emphasized by Clague and Evans (1994), Thompson et al. (1997), Ambrosi and Crosta (2006), Agliardi et al. (2009), and Reitner and Linner (2009). Alzo'ubi (2010) demonstrates that, when downhill-dipping discontinuities dip more steeply than the slope and do not daylight, as at Handcar Peak, tensile fracturing and bulging in a toe buttress are important in facilitating deformation of the rock mass above.

Advanced numerical modelling techniques enable the simulation of pore water pressures, seismicity, and intact rock fracturing. In most cases, and
certainly at Handcar Peak, these factors are potentially significant in driving
gravitational movement. Some previous modelling studies of DSGSD have
included explicit loading by pore water pressures and seismic forces (Stewart
simulations of intact rock fracture has been useful to investigate the transition
from slow DSGSD-type movement to landsliding through progressive damage to
the rock mass by tensile fracturing (Eberhardt et al. 2004, Alzo’ubi 2009). In this
study, I approximated the effect of pore water pressures indirectly by reducing
discontinuity effective frictional strength, but I did not consider seismic
disturbance or intact rock fracturing.

A difficulty in applying these advanced techniques in modelling DSGSD is
that each layer of added complexity and uncertainty requires more data, time,
modeller experience, and consequently more resources (Morgenstern 1995).
Subsurface data on rock structure and groundwater conditions are essential, and
a record of monitored slope behaviour is needed to constrain modelling results.
Due to their considerable cost, the most sophisticated modelling studies of rock
slopes are done at high-risk sites where significant economic damage or loss of
life is possible or has already occurred. Simple numerical modelling studies can
still provide useful insight on slope deformation mechanisms, especially when
done in tandem with other studies using an integrated methodology (Fig. 13).
Standard discrete element modelling can yield results similar to those provided
by advanced modelling techniques when attempting to simulate slope
deformation without exploring the effect of progressive rock damage (Alzo’ubi
2009). The integrated methodology used in this study is suitable for a preliminary investigation of possible mechanisms of deformation and can help to determine if a more detailed study is required.

4.2 Summary and conclusions

Deep-seated gravitational slope deformation (DSGSD) affects large rock slopes with diverse topographic, lithological, and structural characteristics. Typical geologic conditions reported in previous DSGSD studies include stiff blocky rocks overlying ductile sedimentary rocks, ridges composed of foliated metamorphic rocks, and crystalline rocks with significant weakness planes that strike parallel to the slope. Gravitationally deforming slopes move slowly, from less than 1 mm to several centimetres per year, and their total displacements are small compared to the scale of the slope. They commonly lack a discrete, through-going basal shear plane and therefore generally do not exhibit the morphological characteristics of landslides, such as a well-defined headscarp and sidescars. Instead, DSGSD is characterized by surficial features such as linear scarps and trenches that are parallel to the strike of the slope, closed depressions, ponds, and rockfalls and other small landslides. DSGSDs near human settlements pose a hazard because slow, but persistent movement can damage infrastructure and, in rare cases, is a precursor to catastrophic slope failure.

An integrated methodology is required in DSGSD studies, with the overall objectives of determining the limits and size of the deforming rock mass, the time of initiation and history of movement, the current state of activity, and the
mechanism that drives rock mass deformation. Important techniques available to study DSGSD include geomorphic mapping, study of trench sediments, monitoring of surface movement, and numerical modelling.

An integrated methodological approach was used in the study of the DSGSD at Handcar Peak. Geomorphic mapping of antislope scarps, trenches, and small rockslides indicates that the unstable rock mass extends at least 7 km along the side of the valley and 2 km downslope from the ridge top. Current activity is restricted to the upper part of the slope; other areas appear to be stable. The unusually long and straight lineaments at Handcar Peak are the surface expression of slip on weak relict fault planes striking parallel to the shear fabric of the Owl Creek Fault, a steeply dipping, dispersed ductile thrust fault that cuts across the deforming slope.

Some evidence exists for pre-Holocene slope movements at Handcar Peak, but the current episode of activity must have begun after the last major glaciation, because pre-existing surface features would have been removed by glacial erosion. Numerical modelling results suggest that gravitational movements may have begun during or soon after deglaciation, because high pore water pressures such as those that would have been present during deglaciation are required to trigger slip on the fault planes that form antislope scarps. Study of trench sediments behind one antislope scarp indicates that most of the displacement on that scarp happened before 2400 years ago, although some movement occurred afterward. Some geomorphic evidence of recent activity was observed in the upper part of the deforming slope, but the evidence
on balance points to a lack of significant displacements in the past few hundred years.

I used data obtained from engineering geologic mapping of the deforming rock mass to explore the mechanism driving movement at Handcar Peak. Kinematic analysis of discontinuity sets indicates that simple sliding or toppling is not a feasible mechanism for the deformation. Distinct element numerical modelling with the program UDEC suggests a complex active-passive mechanism, in which slip on downhill-dipping joints and relict fault planes plays a critical role in driving rock mass deformation. Preferential slip along the fault planes creates the antislope scarps evident at the surface; scarp height appears to be related to fault plane persistence. Model displacement histories are consistent with qualitative observations that slope activity is trending towards equilibrium. However, I did not consider strain softening by ongoing slope movement in my modelling. Advanced modelling techniques capable of simulating progressive rock mass damage by tensile fracturing and shearing are needed to predict the stability of the slope in the future. Such sophisticated predictive modelling must be supported by more comprehensive and detailed data on the characteristics of the subsurface rock mass, groundwater conditions, and displacement rates at the surface as well as within the deforming rock mass.

4.3 Recommendations for future work

Based on the results of this project, I have several suggestions for future work to improve knowledge of DSGSD in general and, specifically, for understanding the phenomenon in the Coast Mountains.
1. A conceptual framework for a database of deep-seated gravitational slope deformation in British Columbia should be established and the database populated with published data. Modern geospatial tools such as Google Earth should be used to identify and record previously unrecognized gravitationally deforming slopes.

2. Detailed three-dimensional digital models of deforming slopes should be created with airborne LiDAR or by using aerial photographs and digital photogrammetry. The discontinuity planes that form gravitational lineaments should be mapped on these digital models.

3. Permanent scatterer InSAR (PS-InSAR) should be used to map the distribution of surface movements at selected areas in the southern Coast Mountains. Displacements at sites that exhibit gravitational lineaments can be compared to the slopes around them to determine the hazard indicated by the presence of lineaments.

4. More trenching of sediment fills behind antislope scarps, both at Handcar Peak and elsewhere in the Coast Mountains, should be done to refine the temporal relationship between deglaciation and the onset of slope deformation.

5. Conceptual numerical modelling can be done to explore the relations among slope debuttressing during deglaciation, toe erosion, elevated pore water pressures, and deep-seated gravitational slope deformation to determine the relative importance of these factors in triggering gravitational displacements.
6. DSGSD mechanics should be explored at a site where subsurface data are available, using advanced numerical models that simulate intact rock fracturing and progressive rock mass damage. Appropriate models should be employed to back-analyze the transition to catastrophic failure of a large deforming slope through progressive rock mass damage. These models should account for destabilizing factors such as pore water pressures, earthquakes, and annual temperature cycles. Signs of the transition to failure should be determined, including the trend of modelled surface displacements and the development of associated morphologic features.
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5: APPENDIX 1 – DETAIL OF FIELD METHODS

5.1 Geomorphic mapping

I mapped DSGSD-related geomorphic features onto aerial photographs in the field and later digitized them in GIS. I studied many of these features in the field, recording the following information at each site:

Station number
UTM coordinates and elevation (taken with handheld GPS)
Time
Short identifying description of the feature being recorded
Dimensions of the feature
Orientation of the feature
Type of surficial sediments filling or covering the feature
Type of vegetative cover on the feature
Bedrock lithology if outcrops are present
Sketches
Photographs and samples taken
Other observations (interesting sediments or bedrock structures, water, signs of recent activity)
5.1.1 Example of notes taken at a geomorphologic mapping field station

Station M23
UTM 494028/5602581 el. 1956 m asl
16:54
- Fresh-looking trench in bedrock

Dimensions (trench – see Fig. 36):

<table>
<thead>
<tr>
<th>W</th>
<th>H</th>
<th>UHA</th>
<th>ASA</th>
<th>DHA</th>
</tr>
</thead>
<tbody>
<tr>
<td>4.5 m</td>
<td>2.5 m</td>
<td>15°</td>
<td>35°</td>
<td>14°</td>
</tr>
</tbody>
</table>

Orientation: 099°, plunging 02° E

Fill: Fresh, angular, tabular to blocky boulders. They appear to mainly come from steep downhill trench wall.

Vegetation: Grass and heather

Lithology: Meta-volcanics and intrusives are both present.

Photograph 825/29: The trench, looking east

Photographs 825/30-31: Contact between meta-volcanic and intrusive rocks in downhill wall of trench

- There is a deep crack at the base of this trench, ~0.3 m wide, at least 2 m deep. Its edges are fresh, jagged, and rough, indicating tensional opening here rather than the fault-slip that would create an antislope scarp.

![Figure 36](image)

Figure 36. Dimensions of gravitational lineaments recorded in the field. W = width, H = height, DHA = downhill angle, ASA = antislope angle, UHA = uphill angle. Angles were measured with an inclinometer; distances were estimated visually.
5.2 Engineering geological mapping

I performed engineering geological mapping at eight outcrops in the central and eastern sections of the Handcar Peak DSGSD, making a detailed qualitative description of the rock mass at each station. Following a system developed by Brideau (2010), I recorded the following information at each outcrop:

**Basic**
- Station number
- UTM coordinates and elevation
- Time
- Description of outcrop site
- Sketch of outcrop
- Numbers and descriptions of photographs taken and samples collected

**Lithologic description**
- Rock type/s
- Colour, fresh and weathered

**Structures where present**
- Folds (fold axis, axial plane)
- Fault (width, trace, gouge, orientation)
- Shear zone (width, trace, orientation)

**Rock mass description**
- GSI estimate (recorded as a range of 10)
- GSI structure (Fig. 21)
- GSI surface condition (Fig. 21)
- Rock strength estimate using the geological hammer test (Table 9)
- Block shape (Table 10)
- Weathering class (Table 11)
Seepage (presence/absence and description)

Discontinuity description
Orientation (3-5 measurements of each set; recorded as dip and dip direction)
Spacing (Table 12)
Dip persistence (Table 13)
Strike persistence (Table 13)
Primary roughness (Table 14)
Secondary roughness (Fig. 37)

The following tables explain the criteria that I used when making qualitative field descriptions of intact rock strength, block shape, degree of weathering, and the spacing, persistence, and roughness of discontinuities. I assessed the discontinuity properties visually, expressing them as an average for each discontinuity set identified.

Table 9. Criteria for qualitative description of intact rock strength using the geological hammer test (Hoek and Brown 1997)

<table>
<thead>
<tr>
<th>Description</th>
<th>UCS (MPa)</th>
<th>Grade</th>
</tr>
</thead>
<tbody>
<tr>
<td>Indented by thumbnail</td>
<td>&lt; 1.25</td>
<td>R0</td>
</tr>
<tr>
<td>Crumbles under firm blows with point of a geological hammer, can be peeled by a pocket knife</td>
<td>1.25 – 5</td>
<td>R1</td>
</tr>
<tr>
<td>Can be scraped or peeled with a knife; specimen can fracture with a single blow from a geological hammer</td>
<td>5 – 25</td>
<td>R2</td>
</tr>
<tr>
<td>Cannot be scraped or peeled with a knife; specimen can fracture with a single blow from a geological hammer</td>
<td>25 – 50</td>
<td>R3</td>
</tr>
<tr>
<td>Specimen requires more than one blow from geological hammer to fracture it</td>
<td>50 – 100</td>
<td>R4</td>
</tr>
<tr>
<td>Specimen requires many blows of a geological hammer to fracture it</td>
<td>100 – 200</td>
<td>R5</td>
</tr>
<tr>
<td>Specimen can only be chipped with geological hammer</td>
<td>&gt; 200</td>
<td>R6</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Description</th>
<th>Shape</th>
</tr>
</thead>
<tbody>
<tr>
<td>Few joints or very wide spacing</td>
<td>Massive</td>
</tr>
<tr>
<td>Approximately equidimensional</td>
<td>Blocky</td>
</tr>
<tr>
<td>One dimension considerably smaller than the other two</td>
<td>Tabular</td>
</tr>
<tr>
<td>One dimension considerably larger than the other two</td>
<td>Columnar</td>
</tr>
<tr>
<td>Wide variations in block size and shape</td>
<td>Irregular</td>
</tr>
<tr>
<td>Heavily jointed to “sugar cube” texture</td>
<td>Crushed</td>
</tr>
</tbody>
</table>

Table 11. Qualitative criteria for determining weathering class of uniform material, according to Geological Society Engineering Group Working Party (1977).

<table>
<thead>
<tr>
<th>Characteristics</th>
<th>Classifier</th>
<th>Grade</th>
</tr>
</thead>
<tbody>
<tr>
<td>Unchanged from original state; perhaps slight discolouration on major discontinuity surfaces.</td>
<td>Fresh</td>
<td>I</td>
</tr>
<tr>
<td>Slight discolouration of rock and discontinuity surfaces; slight weakening.</td>
<td>Slightly weathered</td>
<td>II</td>
</tr>
<tr>
<td>Considerably weakened; penetrative discolouration; large piece cannot be broken by hand.</td>
<td>Moderately weathered</td>
<td>III</td>
</tr>
<tr>
<td>Large pieces can be broken by hand. Does not readily disaggregate (slake) when dry sample immersed in water.</td>
<td>Highly weathered</td>
<td>IV</td>
</tr>
<tr>
<td>Considerably weakened; slakes; original texture apparent.</td>
<td>Completely weathered</td>
<td>V</td>
</tr>
<tr>
<td>Soil derived by in situ weathering but retaining none of original texture or fabric.</td>
<td>Residual soil</td>
<td>VI</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Description</th>
<th>Spacing</th>
</tr>
</thead>
<tbody>
<tr>
<td>Extremely close</td>
<td>&lt; 20 mm</td>
</tr>
<tr>
<td>Very close</td>
<td>20 – 60 mm</td>
</tr>
<tr>
<td>Close</td>
<td>60 – 200 mm</td>
</tr>
<tr>
<td>Moderate</td>
<td>200 – 600 mm</td>
</tr>
<tr>
<td>Wide</td>
<td>600 – 2000 mm</td>
</tr>
<tr>
<td>Very wide</td>
<td>2000 – 6000 mm</td>
</tr>
<tr>
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Figure 37. Qualitative guide for estimating the secondary (10 cm-scale) roughness of discontinuities using a joint roughness coefficient (JRC) (Barton and Choubey 1977).
This appendix contains examples of the code used in my UDEC models. The code in the first section builds the geometry of a model, sets up initial properties and boundary conditions, and gradually brings model stresses to equilibrium by removing 500-m-thick layers of glacier ice above the slope profile one at a time, solving after each layer. A copy of the current state is saved when the model is equilibrated. The code in the second section of the appendix recalls the saved state of the model, lowers rock mass and joint properties, and runs the model for a large number of steps. Additional states of the model are saved for later examination and several plots are created for a quick assessment of the results.

**6.1 Model setup code**

```
round 3.0

;-------------------------------------------------------
;Geometry
;-------------------------------------------------------

;BLOCK CORNERS ENTERED CLOCKWISE
;block x1,y1 x2,y2 x3,y3 x4,y4
block (0 0) (0 4790) (14000 4790) (14000 0)

change cons 1

;CREATING PROFILE
Table 1 0 3752
Table 1 1041 3752
Table 1 1768 3885
Table 1 3246 4460
Table 1 3512 4493
Table 1 3706 4583
Table 1 3779 4641
Table 1 3876 4667
Table 1 3997 4609
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Crack Table 1

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Crack Table 13

Hide block 6878 5955 1798

jregion id 1 (0 0) (0 4790) (14000 4790) (14000 0)

;CREATING JOINTS IN MODEL OUTSIDE AREA OF INTEREST
jset 75,0 600,0 0,0 200,0 range jregion 1
jset -45,0 1200,0 0,0 400,0 range jregion 1
jdelete
show block 6878

;CREATING WIDE BASE JOINTS
jregion id 2 (4215 1760) (4215 4790) (12000 4790) (12000 1760)
jset 75,0 600,0 0,0 100,0 range jregion 2
jset -45,0 1200,0 0,0 200,0 range jregion 2
jdelete

show block 5955

;CREATING MORE JOINT REGIONS
Table 3
| 4715 | 4672 |
| 4715 | 4172 |
| 5659.1 | 4075.1 |
| 6755.8 | 3899.6 |
| 7438.7 | 3576.7 |
| 8819 | 2780 |
| 11378 | 2260 |
| 14000 | 2260 |

| 4990 | 4782 |
| 4990 | 4532 |
| 5712.9 | 4275.7 |
| 6815.5 | 4122.5 |
| 7498.4 | 3799.5 |
| 8819 | 3030 |
| 11128 | 2510 |
| 14000 | 2510 |

crack table 3
crack table 4
crack 4990 4532 4715 4172
crack 4715 4172 4215 3672
crack 8819 3366 8819 2366

;CREATING MAJOR FAULTS

crack 5766 4474 5248.4 2542.1
crack 6880 4363 6362 2431
crack 7558 4022 7040 2090

crack 6292 2617 6537 2372

;CREATING MINOR FAULTS

crack 5935.1 4453.5 5741 3729
crack 6371 4440 5996 3790.5
crack 6685.2 4431.6 6310.2 3782.1
crack 7074 4243 6879.9 3518.6
crack 7293.2 4142.9 7099.1 3418.5

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;CREATING JOINT SETS
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;DENSE JOINTS 0-250 M DEPTH
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jset 75,0 120,0 0,0 30,0 range jregion 3

jset 75,0 1000,0 0,0 25,0 range jregion 4
jset -60,0 125,0 0,0 50,0 range jregion 4

jset 75,0 1000,0 0,0 25,0 range jregion 5
jset -30,0 100,0 0,0 50,0 range jregion 5

jset 75,0 2000,0 0,0 25,0 range jregion 6
jset -44,0 100,0 0,0 50,0 range jregion 6

jset 75,0 2000,0 0,0 25,0 range jregion 7
jset -44,0 100,0 0,0 50,0 range jregion 7

;DENSE JOINTS 250-500 M DEPTH
jset 75,0 200,0 0,0 30,0 range jregion 8
jset -35,0 240,0 0,0 30,0 range jregion 8

jset 75,0 200,0 0,0 30,0 range jregion 9
jset -35,0 240,0 0,0 30,0 range jregion 9

jset 75,0 1000,0 0,0 25,0 range jregion 10
jset -60,0 250,0 0,0 50,0 range jregion 10

jset 75,0 1000,0 0,0 25,0 range jregion 11
jset -30,0 200,0 0,0 50,0 range jregion 11

jset 75,0 2000,0 0,0 25,0 range jregion 12
jset -44.0 200.0 0.0 50.0 range jregion 12
jset 75.0 2000.0 0.0 25.0 range jregion 13
jset -44.0 200.0 0.0 50.0 range jregion 13
jset 75.0 2000.0 0.0 25.0 range jregion 14
jset -44.0 200.0 0.0 50.0 range jregion 14

; MEDIUM JOINTS 500-1000 M DEPTH
jset 75.0 200.0 0.0 60.0 range jregion 15
jset -35.0 480.0 0.0 60.0 range jregion 15
jset 75.0 2000.0 0.0 50.0 range jregion 16
jset -35.0 480.0 0.0 60.0 range jregion 16
jset 75.0 1000.0 0.0 50.0 range jregion 17
jset -60.0 500.0 0.0 100.0 range jregion 17
jset 75.0 1000.0 0.0 50.0 range jregion 18
jset -30.0 400.0 0.0 100.0 range jregion 18
jset 75.0 2000.0 0.0 50.0 range jregion 19
jset -44.0 400.0 0.0 100.0 range jregion 19
jset 75.0 2000.0 0.0 50.0 range jregion 20
jset -44.0 400.0 0.0 100.0 range jregion 20
jset 75.0 2000.0 0.0 25.0 range jregion 21
jset -44.0 400.0 0.0 50.0 range jregion 21
jset 75.0 2000.0 0.0 50.0 range jregion 22
jset -44.0 400.0 0.0 100.0 range jregion 22
jdelete
show block 1798

; DEFINING AND CREATING ICE LAYERS

Table 5
7874 3766
Table 5 8479 3474
Table 5 8819 3366
Table 5 9182 3284
Table 5 9376 3261
Table 5 9545 3279
Table 5 9885 3188
Table 5 14000 3188

Table 6
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Table 6 7874 3766
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Table 6 14000 3590

Table 7
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;ZONING MODEL

gen edge 25

; Boundary conditions

; FIXING BOUNDARIES
; fix range Xi,xu,yi,yu OR
;bound xvel & yvel (lower upper)
bound xvel 0 range x -5 5
bound xvel 0 range x 13995 14005
bound yvel 0 range y -5 5

;Material properties
;-------------------------------------

;PROPERTIES FOR DEFORMABLE BLOCKS
prop mat 1 dens 2800 fric 46 coh 1.45e7 tens 2e6 dil 5 bulk 38.1e9 shear 22.9e9
prop mat 2 dens 2800 fric 46 coh 1.45e7 tens 2e6 dil 5 bulk 38.1e9 shear 22.9e9
prop mat 3 dens 2800 fric 46 coh 1.45e7 tens 2e6 dil 5 bulk 38.1e9 shear 22.9e9
prop mat 4 dens 2800 fric 46 coh 1.45e7 tens 2e6 dil 5 bulk 38.1e9 shear 22.9e9

;PROPERTIES FOR ICE
prop mat 5 dens 980 fric 5 coh 1e9 tens 1e6 dil 5 bulk 23.8e9 shear 3.5e9

;PROPERTIES FOR JOINTS
prop jmat 1 jkn 1e10 jks 1e9 jfric 36 jcoh 1e6 jtens 1e5
prop jmat 2 jkn 1e10 jks 1e9 jfric 36 jcoh 1e6 jtens 1e5
prop jmat 3 jkn 1e10 jks 1e9 jfric 36 jcoh 1e6 jtens 1e5
prop jmat 4 jkn 1e10 jks 1e9 jfric 36 jcoh 1e6 jtens 1e5
prop jmat 5 jkn 1e10 jks 1e9 jfric 36 jcoh 1e6 jtens 1e5
prop jmat 6 jkn 1e10 jks 1e9 jfric 36 jcoh 1e6 jtens 1e5
prop jmat 7 jkn 1e10 jks 1e9 jfric 36 jcoh 1e6 jtens 1e5
prop jmat 8 jkn 1e10 jks 1e9 jfric 36 jcoh 1e6 jtens 1e5

;ASSIGNING MATERIALS
change mat 2 range above table 13
change mat 2 range above table 2
change mat 3 range above table 3
change mat 4 range above table 4
change mat 5 range above table 1

;ASSIGNING JOINT MATERIALS
;Locked joint region borders
change jmat 8 range id 2 8 9 10 11 12

;Ice
change jmat 5 range id 1
change jmat 5 range above table 1

;Upper faults
change jmat 6 range region (5765 4475) (5767 4475) (5660.1 4074.1) (5658.1 4074.1)
change jmat 6 range region (6879 4364) (6881 4364) (6756.8 3898.6) (6754.8 3898.6)
change jmat 6 range region (7557 4023) (7559 4023) (7439.7 3575.7) (7437.7 3575.7)
change jmat 6 range region (5934.1 4454.5) (5936.1 4454.5) (5827.5 4047.3) (5825.5 4047.3)
change jmat 6 range region (6370 4441) (6372 4441) (6118.9 4000.7) (6116.9 4000.7)
change jmat 6 range region (6684.2 4432.6) (6686.2 4432.6) (6411 3953.9) (6409 3953.9)
change jmat 6 range region (7073 4244) (7075 4244) (6957.5 3803.7) (6955.7 3803.7)
change jmat 6 range region (7292.2 4143.9) (7294.2 4143.9) (7175.9 3700.4) (7173.9 3700.4)

;Lower faults
change jmat 7 range region (5658.1 4074.1) (5660.1 4074.1) (5249.4 2541.1) (5247.4 2541.1)
change jmat 7 range region (6754.8 3898.6) (6756.8 3898.6) (6363 2430) (6361 2430)
change jmat 7 range region (7437.7 3577.7) (7439.7 3577.7) (7041 2089) (7039 2089)

change jmat 7 range region (5825.5 4047.3) (5827.5 4047.3) (5742 3728) (5740 3728)
change jmat 7 range region (6116.9 4000.7) (6118.9 4000.7) (5997 3789.5) (5995 3789.5)
change jmat 7 range region (6409 3953.9) (6411 3953.9) (6311.2 3781.1) (6309.2 3781.1)
change jmat 7 range region (6955.7 3803.7) (6957.5 3803.7) (6880.9 3517.6) (6878.9 3517.6)
change jmat 7 range region (7173.9 3700.4) (7175.9 3700.4) (7100.1 3417.5) (7098.1 3417.5)

;-------------------------------------
;Stresses
;-------------

set grav 0 -9.81

insitu stress -6.579e7 0 -13.157e7 ygrad 1.35e4 0 2.7e4

;RECORD HISTORIES
hist xdisp 5740 4465
hist xdisp 5785 4435
hist xdisp 6630 4420
hist xdisp 7515 4010
hist xdisp 7585 3965
hist xdisp 8765 3300
hist xvel 6880 4363
hist unbal

;;MELTING ICE AND SOLVING

solve

delete above table 9

solve

delete above table 7
delete above table 8

solve

delete above table 10
delete above table 6

solve

delete above table 5
6.2 Model running code

restore c:\handcar\j8ice\j8i3bk5p.sav
reset disp jdis

;===================================================================
;------------------------------------
; ROCK PROPERTIES
;------------------------------------
;Mat 1: Generic rock mass
;------------------------------------
;Lower rock mass, GSI=70
prop mat 1 dens 2800 fric 48 coh 1.7e7 tens 2e6 dil 5 bulk 53.7e9 shear 32.2e9
prop mat 2 dens 2800 fric 48 coh 1.7e7 tens 2e6 dil 5 bulk 53.7e9 shear 32.2e9

;Upper rock mass, GSI=60
prop mat 3 dens 2800 fric 55 coh 6.5e6 tens 1.5e6 dil 5 bulk 38.1e9 shear 22.2e9
prop mat 4 dens 2800 fric 55 coh 6.5e6 tens 1.5e6 dil 5 bulk 38.1e9 shear 22.2e9

change cons 3

;------------------------------------
;JOINT PROPERTIES
;------------------------------------
;Jmat 1: Generic rock joints
;------------------------------------
;Lower joints
prop jmat 1 jkn 1e9 jks 1e8 jfric 25 jcoh 5e4 jtens 0
prop jmat 2 jkn 1e9 jks 1e8 jfric 25 jcoh 5e4 jtens 0

;Upper joints
prop jmat 3 jkn 1e9 jks 1e8 jfric 20 jcoh 0 jtens 0
prop jmat 4 jkn 1e9 jks 1e8 jfric 20 jcoh 0 jtens 0

;------------------------------------
;Jmat 2: Faults
;------------------------------------
;Lower faults
prop jmat 7 jkn 5e8 jks 5e7 jfric 12 jcoh 0 jtens 0
;Upper faults
prop jmat 6 jkn 5e8 jks 5e7 jfric 12 jcoh 0 jtens 0

title
Joints 8i3b model, GSI 70/60, k=0.5, very lo fric

step 100000

set output c:/handcar/final/bk5/bk5_xlofricLC_100kpl.jpg
set plot jpg size 1200 900
plot block pl pen

step 100000

save c:/handcar/final/bk5/bk5_xlofricLC_200k.sav

set output c:/handcar/final/bk5/bk5_xlofricLC_200kpl.jpg
set plot jpg size 1200 900
plot block pl pen

step 100000

set output c:/handcar/final/bk5/bk5_xlofricLC_300kpl.jpg
set plot jpg size 1200 900
plot block pl pen

step 100000

set output c:/handcar/final/bk5/bk5_xlofricLC_400kpl.jpg
set plot jpg size 1200 900
plot block pl pen

step 100000

save c:/handcar/final/bk5/bk5_xlofricLC_500k.sav

set color black
set back iwhite
set output c:/handcar/final/bk5/bk5_xlofricLC_500k.jpg
set plot jpg size 1200 900
plot bl pen

window 5500 9500 1750 5750
set output c:/handcar/final/bk5/bk5_xlofricLC_500kd1.jpg
set plot jpg size 1200 900
plot block disp pen

set output c:/handcar/final/bk5/bk5_xlofricLC_500kbm.jpg
set plot jpg size 1200 900
plot block mag 30 pen

window 4100 12100 1800 9800
set output c:/handcar/final/bk5/bk5_xlofricLC_500kd2.jpg
set plot jpg size 1200 900
plot block disp pen
window 0 14000 0 14000

set output c:/handcar/final/bk5/bk5_xlofricLC_500ksxx.jpg
set plot jpg size 1200 900
plot sxx proj fill pen

set output c:/handcar/final/bk5/bk5_xlofricLC_500ksyy.jpg
set plot jpg size 1200 900
plot syy proj fill pen

set output c:/handcar/final/bk5/bk5_xlofricLC_500kpl.jpg
set plot jpg size 1200 900
plot block pl pen
7: APPENDIX 3 – DIGITAL DATA

The attached electronic data and files form part of this work. They include:

1. `particlesize\particle size data.xls (24 kb)`: An Excel file with data and plots from the laboratory analysis of trench sediments, including particle size analytical data and cone penetrometer liquid limit test results on the gouge clay for UCS classification.

2. `eng geo mapping\eng geo mapping data.xls (52 kB)`: An Excel file with field notes from each engineering geological mapping station and a summary data sheet of all the discontinuities recorded by spot mapping and photogrammetry.

3. `gisdata (5.29 MB)`: A database of GIS shape files. The data include:
   a. Mapped surficial features (gravitational lineaments, surficial instabilities, ponds, and avalanche chutes)
   b. Geological contacts within the study area, revised slightly from Monger and Journeay (1994)
   c. Locations of geomorphological and engineering geological field mapping stations
   d. Detailed information about individual gravitational lineaments.

4. `photos (3.23 GB)`: Photographs and videos taken at Handcar Peak.