Characterisation of Large Catastrophic Landslides using an Integrated Field, Remote Sensing, and Numerical Modelling Approach

by
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B.Sc., Simon Fraser University, 2009

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Abstract

I apply a forensic, multidisciplinary approach that integrates engineering geology field investigations, engineering geomorphology mapping, long-range terrestrial photogrammetry, and a numerical modelling toolbox to two large rock slope failures to study their causes, initiation, kinematics, and dynamics. I demonstrate the significance of endogenic and exogenic processes, both separately and in concert, in contributing to landscape evolution and conditioning slopes for failure, and use geomorphological and geological observations to validate numerical models.

The 1963 Vajont Slide in northeast Italy involved a 270-million-m$^3$ carbonate-dominated mass that slid into the newly created Vajont Reservoir, displacing water that overtopped the Vajont Dam and killed 1910 people. Based on literature, maps and imagery, I propose that the landslide was the last phase of slow, deep-seated slope deformation that began after the valley was deglaciated in the Pleistocene. Field and air photograph observations and stream profiles provide the context of the Vajont Slide. The first long-range terrestrial digital photogrammetry models of the landslide aid in characterising the failure scar. Analysis of the failure scar emphasises the complexity of the failure surface due to faults and interference between two tectonic fold generations, influencing failure behaviour. Observations of the pre- and post-failure slope and interpretation of numerical simulations suggest a complex three-dimensional active-passive wedge-sliding mechanism, with two main landslide blocks and five sub-blocks in the west block, separated by secondary shear surfaces.

The 1959 Madison Canyon Slide in Montana, USA, was triggered by an M = 7.5 earthquake. A 20-million-m$^3$ rock mass descended from the ridge crest, killing 24 people and blocking Madison River to create Earthquake Lake. Marble at the toe of the slope acted as a buttress for weaker schist and gneiss upslope until the earthquake undermined its integrity and triggered failure. Rock mass characterisation, long-range terrestrial digital photogrammetry, and kinematic analysis indicate that the lateral, rear, and basal release surfaces formed a hexahedral wedge-biplanar failure. Dynamic
numerical modelling suggests topographic and damage amplification due to ridge geometry and pre-existing tension cracks.

Analysis of the case studies highlights the complexity of large, catastrophic rock slope failures, their causes, and their evolution from incipient failure to disaster.

**Keywords**: catastrophic rock slope failure; Vajont Slide; Madison Canyon Slide; engineering geomorphology; remote sensing; numerical modelling
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A summary of the last few years, thanks to PhD Comics:

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Chapter 1.

Dissertation Motivation

Large, catastrophic landslides are among the most devastating natural hazards. Understanding the context, or situation, of a potential landslide is critical for prediction and involves study of past events as well as the geomorphology, geology, hydrology, and climate of the site in question. Scientists have used a variety of techniques to better understand these phenomena, including aerial photograph interpretation, field mapping and observations, and, more recently, numerical modelling, photogrammetry, and laser scanning. They have completed in-depth investigations of individual events and addressed broader questions using landslide inventories.

This dissertation applies multidisciplinary, forensic investigations to two large, catastrophic rockslides. I integrate field studies, engineering geomorphology mapping, photogrammetry, and sophisticated numerical modelling to better understand the landslides. The overarching goal of my research is to improve understanding of large mass movements so that losses from similar events in the future can be reduced. The methodology that I use may be applicable in related research, such as the stability of open pit mines and existing and new urban development.

1.1. Objectives and Contributions of the Research

The principal objectives of my research are to:

• improve understanding of endogenic and exogenic factors that condition slopes to fail,
• document initiation mechanisms of large catastrophic rock slope failures,
• evaluate landslide causation in the context of landscape evolution, and
• determine relative landslide chronology based on slope and deposit morphology and evolution.

I seek to achieve these objectives by analysing two case studies, the Vajont\(^1\) Slide in northeast Italy and the Madison Canyon Slide in southwest Montana, USA. I use a multidisciplinary approach, including facets of engineering geology and engineering geomorphology, rock mechanics, and structural geology. The two case studies were chosen as they are complex rock slope failures with different triggers, namely fluctuating pore water pressures at Vajont and a large earthquake at Madison Canyon. I investigate regional and local geomorphological processes and landforms at several scales using aerial photograph interpretation, LiDAR and field mapping, terrestrial digital photogrammetry, two- and three-dimensional numerical modelling, and laboratory testing. Contributions of the project include:

• an innovative approach to landslide studies involving methods drawn from several disciplines,
• investigation of case study context, causes, triggers, and initiation at different scales and with considerations of uncertainty, and
• application of state-of-the-art techniques such as long-range terrestrial photogrammetry and two- and three-dimensional numerical modelling using new and updated software and incorporating groundwater and seismic loading effects.

1.2. Background and Influential Literature

1.2.1. Large Catastrophic Rock Slope Failures

Large bedrock failures have been studied for over two centuries in an attempt to understand and predict them. The focus of most of the earliest studies was classification. Dana (1864) presented one of the earliest landslide classifications in his description of the 1806 Goldau rockslide (Cruden, 2003). Heim (1932), the leading landslide researcher of his era, published detailed descriptions and classifications of many large landslides and commented on their mechanisms, causes, and triggers. The most widely

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\(^1\) I have chosen to use the traditional Italian spelling of "Vajont", rather than the anglicised "Vaiont". "Vajont" is now accepted when referring to the 1963 landslide, as seen in Genevois and Prestininzi (2013).
accepted classification is that of Varnes (1978), which was later revised by Cruden and Varnes (1996) and Hungr et al. (2001, 2014); it is based on material type, movement type, and velocity. In this classification system, a “rockslide” is the rapid downslope movement of rock along a discrete plane; a “rock avalanche” is an extremely rapid, streaming mass of highly fragmented rock characterized by a longer run-out than expected from a rockslide of the same volume.

Although important, classifications alone do not capture the complexity of large, catastrophic landslides. Hence, landslide research in recent decades has taken new directions, with greater emphasis on the mechanisms, kinematics, dynamics, causes, and triggers of events. Research has also broadened into a large number of disciplines, including geology, geography, rock mechanics, soil mechanics, hydrology, hydrogeology, physics, biology, and chemistry. Relatively few scientists, however, apply principles from multiple disciplines in their research.

Particularly active topics of study are those concerning the failure mechanisms affecting large rockslides, that is, the processes controlling failure behaviour. Mechanisms such as undrained loading and fragmentation forces have been proposed to account for rock avalanche mobility. Nicoletti and Sorriso-Valvo (1991) investigated how topography affects the shape and mobility of rock avalanches. Kent (1966) analysed several large rockslides and noted the possible role of compressed air in the excess transport of rock avalanches. Goguel (1978) cited pore water vapourisation as an explanation for the low frictional resistance and rapid motion of large rockslides. Masch et al. (1985) investigated frictional melting of rock and formation of frictionites, and Erismann and Abele (2001) described frictionite at Köfels, Austria as an example of the frictional melting mechanism. Abele (1994) argued that movement within rockslides is restricted to thin internal sliding planes in an otherwise coherent slide mass. Pressure concentrated along sliding planes may destroy asperities that inhibit movement and increase pore pressure. Many researchers have hypothesised possible mechanisms to explain the high terminal velocities of the Vajont Slide in Italy. From the mid-1970s to 1980s, a popular mechanism was heat-induced vapourisation (Goguel, 1978; Habib, 1975; Nonveiller, 1987). Voight and Faust (1982) doubted that vapourisation occurred, but hypothesised that frictional heat could increase fluid pressure and lead to a reduction in friction values. Vardoulakis (2002) and Kilburn and Petley (2003) supported the
vapourisation theory, but minimised its significance. Instead, Vardoulakis suggested thermo-poro-mechanical softening of the rock mass, whereas Kilburn and Petley favoured a rock-cracking model. Erismann and Abele (2001) did not believe that the velocities reached during the Vajont Slide required any exotic explanations. Rather, centre of gravity calculations demonstrate that the velocities are within the reasonable range for displacement of the centre of gravity. The slide, at estimated velocities of 20 m/s, was not very rapid and could have been predicted.

Numerous processes condition slopes for large catastrophic failure. These processes have been classified variously as causes and triggers, internal and external causes (Terzaghi, 1960), or endogenic and exogenic processes (Gerber and Scheidegger, 1969; Whalley, 1974; Leith, 2012). A trigger is the last cause that initiates failure, essentially the “straw that breaks the camel’s back”, whereas causes are underlying processes or aspects that predispose a slope to failure (Heim, 1932).

According to Terzaghi (1960), internal causes are those that reduce the shear strength of a rock mass without changing surface conditions; they include material fatigue, crack growth, alteration, decreased cohesion, and a decrease in the friction angle of the material. Conversely, external causes increase shear stresses without changing shear resistance and include slope steepening or undercutting, loading or unloading of a slope, weathering, seismicity, high precipitation, freeze-thaw cycles, snow, ice, and permafrost melting, glacial debuttressing, tectonic stress, and human activity. Although the distinction between internal and external forces is widely accepted, Erismann and Abele (2001) argued that landslide causes can be both internal and external, or neither. For example, crack growth in a rock mass caused by increased stress can be both internal and external, and gravitational acceleration does not fit either definition.

Leith (2012) provided an alternative conceptual model for processes affecting slopes. Based on earlier research by Gerber and Scheidegger (1969) and Whalley (1974), he defined “endogenic processes” as those that exert stresses on rock due to tectonics, isostasy, and volcanism; and exogenic processes as those exerting stress on rock due to gravitational or climatic forcing related to weathering, fluvial incision, and mass wasting. Leith (2012) is one of the few researchers who considered the interaction
between these two types of processes in the context of in-situ stress distributions and their relations to landforms and Earth processes.

Examples presented in Table 1.1 highlight the causes and triggers of large rock slope failures. Structural elements such as faults, folds, and fractures play a significant role in most rock slope failures. Sliding surfaces commonly are located on planes of weakness such as bedding and foliation. Brideau (2010) highlighted the importance of discontinuity sets and topography in the failures of rock slopes such as McAuley Creek and Chehalis Lake in British Columbia, and Turtle Mountain, Alberta. von Poschinger (2005, 2006) summarised new studies on the prehistoric Flims rockslide in Switzerland and emphasised geological structures as the underlying cause of the event. The Malm limestone and chalk at Flims have been metamorphosed, folded, and faulted, and are more prone to failure than would otherwise have been the case. Eisbacher and Clague (1984) linked the event to fault systems in the Rhein valley and conditioning due to previous landslides at the same location. The prehistoric Seymareh (Saidmarreh) rockslide, one of the largest landslides in the world, occurred on a dip slope in the Zagros Mountains, Iran, and might have been triggered by an earthquake (Harrison and Falcon, 1937; Roberts and Evans, 2013). Antinao and Gosse (2009) created a chronospatial inventory of large landslides in the Chilean Andes and related large landslides to geological structures and seismicity.

Table 1.1. Summary of large rock slope failures and their causes and triggers (italicised). DS = discontinuity set.

<table>
<thead>
<tr>
<th>Name</th>
<th>Date</th>
<th>Type</th>
<th>Volume (10^6 \text{ m}^3)</th>
<th>Preconditioning factors</th>
<th>References</th>
</tr>
</thead>
<tbody>
<tr>
<td>Åknes, Norway</td>
<td>&gt;1960-preset</td>
<td>DSGSD rockslide</td>
<td>35</td>
<td>gouge material, foliation, 6 DS, faults, folds, steep fjord walls, gully as lateral release</td>
<td>Ganerød et al., 2008</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>Oppikofer et al., 2009</td>
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<td></td>
<td></td>
<td></td>
<td></td>
<td>Grøneng et al., 2010</td>
</tr>
<tr>
<td>Avalanche Lake, Canada</td>
<td>&lt; 500 yr ago</td>
<td>rock avalanche</td>
<td>200</td>
<td>dolostone, regional syncline, 3 DS, snowfall, freeze-thaw</td>
<td>Evans et al., 1994</td>
</tr>
<tr>
<td>Name</td>
<td>Date</td>
<td>Type</td>
<td>Volume (10^6 m³)</td>
<td>Preconditioning factors</td>
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</tr>
<tr>
<td>Beauregard, Italy</td>
<td>1951-present</td>
<td>DSGSD</td>
<td>650</td>
<td>micaschist, paragneiss, subvertical DS, reservoir flux, snowmelt</td>
<td>Barla et al., 2010; Kalenchuk, 2010</td>
</tr>
<tr>
<td>Blackhawk, California</td>
<td>18,000 yr ago</td>
<td>rockslide, rockfall, compound slide</td>
<td>300</td>
<td>marble and gneiss, gouge, 2 faults, foliation, strain-softening</td>
<td>Johnson, 1978</td>
</tr>
<tr>
<td>Black Rapids, Alaska</td>
<td>Nov. 3, 2002 (3 events)</td>
<td>rock avalanche</td>
<td>30 in total</td>
<td>quartz diorite, multiple parallel and orthogonal DS, earthquake</td>
<td>Shugar and Clague, 2011</td>
</tr>
<tr>
<td>Campo Vallemaggia, Switzerland</td>
<td>1892 – present</td>
<td>DSGSD</td>
<td>800</td>
<td>amphibolite, schist, gneiss, folding, fault, schistosity, deep artesian pressure</td>
<td>Bonzanigo, 1999; Bonzanigo et al., 2007</td>
</tr>
<tr>
<td>Cheam, British Columbia</td>
<td>5000 yr ago</td>
<td>rock avalanche</td>
<td>175</td>
<td>volcanics, limestone, slate, folded, faulted, jointed; earthquake? high water pressures?</td>
<td>Orwin et al., 2004</td>
</tr>
<tr>
<td>Chehalis Lake, British Columbia</td>
<td>Dec. 4, 2007</td>
<td>rockslide, debris avalanche</td>
<td>3</td>
<td>quartz diorite 5 DS, fault, gully rain-on-snow</td>
<td>Brideau, 2010; Brideau et al., 2012</td>
</tr>
<tr>
<td>Downie, British Columbia</td>
<td>10,000 yr ago</td>
<td>DSGSD</td>
<td>10,000</td>
<td>micaschist, gneiss, limestone, folds, fault, 4 main DS, foliation, ponding, springs, multiple glaciations</td>
<td>Piteau et al., 1978; Kalenchuk, 2010</td>
</tr>
<tr>
<td>Eiger, Switzerland</td>
<td>July 13, 2006</td>
<td>rock collapse</td>
<td>2</td>
<td>limestone, 3 main DS, snow melt, high water pressures, glacial debuttressing</td>
<td>Oppikofer et al., 2008; Jaboyedoff et al., 2012</td>
</tr>
<tr>
<td>Name</td>
<td>Date</td>
<td>Type</td>
<td>Volume (10^6 m³)</td>
<td>Preconditioning factors</td>
<td>References</td>
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</tbody>
</table>
| Elm, Switzerland      | Sept. 11, 1881 | sturzstrom | 10               | slate, 
\textit{heavy precipitation, mining}               | Hsü, 1975                                 |
| Flims, Switzerland    | prehistoric  | bergsturz  | 8,500-12,000     | limestone, fault, fold, foliation                        | von Poschinger, 2005 
Pollet et al., 2005 |
| Frank, Alberta        | April 29, 1903 | rock      | 30               | limestone, clastics, anticline, 3 DS, faults, 
progressive deformation, weathering, 
\textit{mining, precipitation, freeze-thaw} | Benko and Stead, 1998 
Froese et al., 2009 
Jaboyedoff et al., 2009 
Brideau et al., 2011 |
| Goldau, Switzerland   | Sept. 2, 1806  | rockslide  | 40               | marl, molasse, joints, bedding, 
\textit{snow melt, high water pressures}                | Thuro and Hatem, 2010                    |
| Hope, British Columbia| Jan. 9, 1965  | rockslide  | 47               | greenstone, felsite, foliation, DS, seepage, surface runoff, 
previous event                                    | Brideau et al., 2005                    |
| Huascaran, Peru       | May 31, 1970  | rockslide/avalanche | 53-75           | granodiorite, 
\textit{glacial source of water, earthquake}        | Evans et al., 2009                      |
| Köfels, Austria       | 8700 yr ago   | rock avalanche | 2,100           | para/orthogneiss, schist, fault, 
repeated glaciation                                     | Hermanns et al., 2006                   |
| La Clapière, France   | 1900-present  | creep     | 60               | gneiss, schist, diorite, fold, faults, foliation, 
springs, river flow, weathering                        | Cappa et al., 2004 
Helmstetter et al., 2004 |
<table>
<thead>
<tr>
<th>Name</th>
<th>Date</th>
<th>Type</th>
<th>Volume ($10^6$ m$^3$)</th>
<th>Preconditioning factors</th>
<th>References</th>
</tr>
</thead>
<tbody>
<tr>
<td>Langtang, Nepal</td>
<td>25-30 ka ago</td>
<td>sturzstrom</td>
<td>15,000</td>
<td>biotite gneiss, granite, faulting, paleoseismicity, DS, high relief, glacial action</td>
<td>Heuberger et al., 1984; Schramm et al., 1998; Barnard et al., 2006; Takagi et al., 2007</td>
</tr>
<tr>
<td>Madison, Montana</td>
<td>August 17, 1959</td>
<td>rockslide</td>
<td>20</td>
<td>gneiss, schist, marble, regional faults, 3-4 DS, steep slopes, earthquake</td>
<td>Hadley 1964, 1978</td>
</tr>
<tr>
<td>McAuley Creek, British Columbia</td>
<td>May/June, 2002</td>
<td>rockslide (wedge)</td>
<td>6</td>
<td>gneiss, foliation, steeply dipping DS, alteration</td>
<td>Brideau, 2010; Brideau et al., 2012</td>
</tr>
<tr>
<td>Mt Steele, Yukon</td>
<td>July 24, 2007</td>
<td>rock/ice avalanche (5/95%)</td>
<td>27-80</td>
<td>granodiorite, diorite, gabbro, fault, 3 main DS, snow melt, ponding, high temp., earthquake</td>
<td>Lipovsky et al., 2008</td>
</tr>
<tr>
<td>Mt. Meager, British Columbia</td>
<td>August 6, 2010</td>
<td>rock/debris slide</td>
<td>48.5</td>
<td>pyroclastics, andesite, highly fractured, hydrothermal alteration, glacial debuttressing, snow melt?</td>
<td>Read 1990; Bovis and Jacob, 2000; Guthrie et al., 2012</td>
</tr>
<tr>
<td>Palliser, Alberta</td>
<td>prehistoric (2 events)</td>
<td>rockslide</td>
<td>8</td>
<td>cherty carbonates, 5 DS, regional thrust, syncline, dip slope, seepage, karst, glacial/fluvial erosion</td>
<td>Sturzenegger and Stead, 2011</td>
</tr>
<tr>
<td>Pandemonium Creek, British Columbia</td>
<td>summer 1959</td>
<td>rock avalanche</td>
<td>5</td>
<td>gneissic quartz diorite, foliation</td>
<td>Evans, 1989</td>
</tr>
<tr>
<td>Name</td>
<td>Date</td>
<td>Type</td>
<td>Volume (10^6 \text{ m}^3)</td>
<td>Preconditioning factors</td>
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<tr>
<td>Randa, Switzerland</td>
<td>April 18 and May 9, 1991 present</td>
<td>rockslide (2-stage)</td>
<td>30</td>
<td>ortho/paragneiss, 3 main DS, fault, schistosity, springs, hydraulic fracturing, previous events</td>
<td>Sartori et al., 2003 Willenberg et al., 2008</td>
</tr>
<tr>
<td>Saidmarreh, Iran</td>
<td>9000 yr ago</td>
<td>rock avalanche</td>
<td>44,000</td>
<td>limestone, shale, folds, faults, dip slope, \textit{earthquake}</td>
<td>Harrison and Falcon, 1937 Roberts and Evans, 2013</td>
</tr>
<tr>
<td>Tafjord, Norway</td>
<td>April 7, 1934</td>
<td>rock avalanche</td>
<td>1.5</td>
<td>mica gneiss, DS, foliation</td>
<td>Hermanns et al., 2006</td>
</tr>
<tr>
<td>Vajont, Italy</td>
<td>October 9, 1963</td>
<td>rockslide</td>
<td>270</td>
<td>limestone, clay, faults, folds, 9 DS, \textit{reservoir levels}, \textit{precipitation}</td>
<td>Hendron and Patton, 1985 Genevois and Prestininzi, 2013 (eds.)</td>
</tr>
<tr>
<td>Val Pola, Italy</td>
<td>July 28, 1987</td>
<td>rock avalanche</td>
<td>34</td>
<td>diorite, gabbro, quartzite, 2 faults, 4 DS, high relief, glacial erosion, \textit{rainfall, high temp., permafrost thaw}</td>
<td>Azzoni et al., 1992 Crosta et al., 2004</td>
</tr>
</tbody>
</table>

Seismic events and hydrogeological conditions commonly act in concert with lithological and structural controls to condition slopes for failure. As these two factors triggered, respectively, the Madison Canyon and Vajont slides, I examine them in more detail below.

1.2.2. **Effects of Seismicity on Slopes**

Although considerable research has been conducted on earthquake-triggered landslides, the interaction between seismic waves and slopes is not well understood
Seismicity is commonly treated as a dynamic load applied to a slope over short intervals and is thus considered a trigger of slope failure (Meunier et al., 2007). The long-term effects of multiple earthquakes, distance and travel path, directional effects, dispersion, and incoherence on slope integrity are rarely considered. Earthquakes have been mistakenly cited as triggers of some large landslides, but re-examination has shown that the landslides triggered the earthquakes, rather than being caused by them. Examples include the Hope, Frank, and Elm landslides. Strom and Stepancikova (2008) provided a discussion on how to differentiate between seismically triggered failures and landslides caused by other factors. The most definitive way to distinguish seismically triggered landslides is to link them with contemporaneous earthquake events. Other evidence includes location in an active seismic area, coincidence of the landslide with a seismogenic fault, presence of liquefaction features, and a lack of other clear triggering processes.

Keefer (1984) conducted a seminal study of seismically triggered landslides. He provided a global catalogue of 40 earthquakes that triggered landslides and investigated how specific earthquakes affect the number, type, distribution, and density of landslides. The results showed that the area affected by landslides during an earthquake is proportional to seismic magnitude. Of the 14 types of coseismic mass movements that Keefer (1984) examined, rock falls are the most common, followed by disrupted soil slides, rock slides, and rock avalanches. The materials most susceptible to failure during earthquakes are weakly cemented, weathered, fractured rocks, sand, loess, deltaic sediments, and uncompacted anthropogenic fill.

Site response to seismic ground motions is an important factor in slope stability. Seismic amplification can be caused by geology and topography. In general, high-quality rock masses resonate at higher frequencies than low-quality rock masses and soils, which are subject to low frequency excitation (Elnashai and Di Sarno, 2008). The thickness and elasticity of the soil at a given site are also important considerations in determining how the site will react to ground motions. An investigation of the San Francisco Bay area demonstrated that maximum horizontal ground velocities in unconsolidated mud and artificial fill were amplified ten times when compared to nearby bedrock (Borcherdt, 1970). As mud thickness increased, velocities increased as well. Another classic example of the effects of geology is the response of lake sediments in
Mexico City to several earthquakes, including the devastating 1985 event. Ordaz and Singh (1992) concluded that lake sediments caused 100 to 500 times amplification at spectral frequencies of 0.2 to 0.7 Hz. Areas uphill of the lake sediments produced amplifications of ten times.

Meunier et al. (2007, 2008) documented the topographic effects of ground accelerations caused by the 1999 Chi-Chi, Taiwan, 1994 Northridge, California, and 1993 Finisterre, Papua New Guinea, earthquakes, and concluded that landslides were most common on the hanging wall of the fault in each case and that landslide density is linearly proportional to peak ground acceleration. They also noted that landslides triggered by earthquakes commonly are concentrated on ridge crests and convexities in slopes; conversely, landslides triggered by rainfall are evenly distributed along the slope profile, and those triggered by storms are concentrated at the toes of slopes. Slopes facing away from earthquake epicentres have proportionally more landslides than other slopes, possibly due to asymmetric amplification of oblique seismic waves.

Del Gaudio and Wasowski (2011) investigated the effect of lithology and topography on seismic amplification in the Apennines of Italy. They concluded that slopes covered by thick colluvium (> 5 m) are prone to greater amplification of seismic ground motions than rock slopes, and that amplification is also affected by topography and structural features.

Harp and Jibson (2002) suggested that topographic amplification during the 1971 San Fernando and 1994 Northridge earthquakes explains the unusually high concentration of rock falls within Pacoima Canyon during both events. They ruled out lithological causes because rocks in surrounding areas have the same geological and geotechnical properties, but did not experience as many landslides as the canyon. Sepúlveda et al. (2005) corroborated these findings with pseudostatic and Newmark analyses and argued that discontinuity orientation and topographic amplification explain the high number of landslides in the canyon.

Havenith et al. (2003a, b) investigated the earthquake-triggered Ananevo rockslide and Suusamyr debris slump-flow in Kyrgyzstan. They used UDEC to model seismic amplification and deformation mechanisms. Both landslides initiated at ridge
crests. Amplification was highest in the thickest soft layers such as weathered rocks and sediments, particularly at low frequencies (< 2.0 Hz).

1.2.3. Hydrogeological Conditions of Rock Slopes

Water has often been cited as a cause and trigger of mass movements. Although much literature focuses on the effects of precipitation on surficial landslides involving soils, ground water affects rock slopes as well. Surface water can infiltrate rock through pores, fractures, faults, and cavities, causing hydraulic, mechanical, physical, and chemical changes in the material (Mohammed, 1997; Lu and Godt, 2013). Water can reduce the strength of the material or alter the stress conditions at the site (Figure 1.1). Pore water pressure is particularly important, as it decreases the effective normal stress acting on a slope, thus decreasing its stability. Other factors include water flow, freeze/thaw cycles, saturation, and physical and chemical alteration. Surface and subsurface flow contributes to erosion, thus degrading material strength and changing stress distributions (Mohammed, 1997). Freeze/thaw cycles work over time to stress and unload material, thus degrading material strength and contributing to weathering. Matsuoka et al. (1998) provided examples of diurnal, annual, and millennial freeze-thaw cycles in the Swiss Alps. Saturation of rock and soil changes the weight distribution in a slope, which can lead to instability. Finally, hydrothermal alteration and hydraulic fracturing contribute to the strength degradation of rock slopes, as was the case for the 2010 Mount Meager, British Columbia, landslide (Guthrie et al., 2012).

**Figure 1.1.** Summary of hydrogeological effects on rock slope stability.
1.2.4. The Role of Geomorphological Processes

Surface processes such as aeolian, fluvial, and glacial erosion, soil creep, and weathering contribute to landslides. Although large catastrophic landslides are surface processes themselves, they are influenced and conditioned by other geomorphologic processes. For example, fluvial erosion of slopes can undermine their stability, as happened, for example, at Madison Canyon.

Glacier erosion and debuttressing are important causes of landslides in glaciated and formerly glaciated areas. Ballantyne (2002), Holm et al. (2004), Cossart et al. (2008), and Ballantyne and Stone (2013), for example, have hypothesised that unloading of slopes due to glacier downwasting and retreat drastically changes stress conditions in a given rock mass, which might lead to failure. Ballantyne (2002) cited numerous examples of catastrophic rock slope failures associated with glacier oversteepening, thinning, and debuttressing, including the Sherman Glacier rock avalanche in Alaska, the Maud Glacier rock avalanches in New Zealand, the Pandemonium Creek rock avalanche in British Columbia, and the Huascaràn rockslides in Peru. Recently, however, McColl et al. (2010) disputed this conclusion on the basis that ice does not form an effective slope buttress. They instead invoked fluctuating groundwater, climate, and seismicity to explain postglacial landslides.

Another aspect of large catastrophic rock slope failures is their legacy in the landscape. Hewitt et al. (2008) examined catastrophic landslides in the Karakoram Himalaya, the Coast Mountains of North America, and the Southern Alps of New Zealand. They described their interactions with topography, substrate materials, glaciers, and other mountain processes, and emphasised the roles of large landslides in shaping the landscape and in perturbing river systems for millennia, or even longer. Hovius et al. (1997, 2000) demonstrated how landslides interact with streams in a long-term coupled system.

1.2.5. Human Impacts on Slopes

Human actions have partly caused several large rockslides. The rock slope failure at Frank, Alberta, in 1903, although mainly conditioned by tectonics, rock mass discontinuities, and fluvial erosion, may have happened, in part, due to coal mining at
the base of Turtle Mountain (Jaboyedoff et al., 2009). The 1881 Elm rockslide in Switzerland was partially caused by the extraction of slate from the Plattenberg quarry at the base of the Tschingelwald slope (Eisbacher and Clague, 1984; Heim, 1932). Several landslides have initiated on slopes above artificial reservoirs (Schuster 1979). One of the most significant of such events is the 1963 Vajont Slide, Italy, which was triggered by raising and lowering of the level of Vajont Reservoir behind the newly constructed Vajont Dam.

1.2.6. Approaches and Methods Used to Study Failing Rock Slopes

In recent years, researchers have applied a suite of new techniques to studies of landslides and unstable rock slopes, including acoustic emission monitoring, thermal imaging, satellite imagery analysis, terrestrial and aerial laser scanning, terrestrial and aerial digital photogrammetry, and numerical modelling. Study of remote and previously inaccessible areas has been possible through the use of these technologies. Nevertheless, field investigation remains the cornerstone of any project. Below, I discuss the approaches applied in this dissertation, namely field characterisation, engineering geomorphological mapping, remote sensing, and numerical modelling.

**Field Rock Mass Characterisation**

In my work, I characterised the rock masses at each site using an engineering geology approach. I used engineering geology methods such as lithological descriptions, discontinuity surveys, intact rock strength and Geological Strength Index (GSI) estimation, and block size and shape determination. Hoek and Brown (1997) and British Standards (1981) have standardised field estimation of intact rock strength, and the Geological Society Engineering Group Working Party (1977) has classified weathering grade of rock masses, ranging from I (fresh) to VI (residual soil). Hoek et al. (1995) and Marinos and Hoek (2000) presented the standard GSI chart. Cai et al. (2007) provided an updated, quantitative version based on block volume and joint condition factor. ISRM (1978) presented one approach to block size and shape, whereas Kalenchuk et al. (2006) provided another based on the shortening of the minor block axis and elongation of the major block axis.
The recognition of geomechanical domains, scale effects, and failure mechanisms in rock is also important. Pine and Harrison (2003) analysed the relationships among rock mass classification parameters using multivariate statistics. They showed that discontinuity anisotropy and scale effects are critical factors in rock masses. Brittle failure through intact rock bridges and the creation of step-paths are other elements of rock failure (Willenberg et al., 2008). Brideau (2010) also emphasised the three-dimensionality of rock mass parameters such as block shape, variation in discontinuity properties, and topography.

**Engineering Geomorphology**

The context, or situation, of an unstable slope or rock slope failure is important in understanding failure initiation and propagation. Griffiths and Whitworth (2012) applied the concept of the geomorphological process-response system to landslides, wherein mass movements are part of a larger system of processes and landforms contributing to the four-dimensional evolution of a landscape. They pointed out that this approach is similar to that used by engineering geologists and geotechnical engineers to study landslides, but with more stress on field mapping and the use of remote sensing data. The approach is part of the larger engineering geomorphology framework pioneered by Brunsden et al. (1975), Doornkamp et al. (1979), the Geological Society of London (1982), and Fookes et al. (2005, 2007). Brunsden (2002) emphasised the importance of engineering geomorphology to all stages of a project, not just the initial mapping of landforms; an understanding of landscape inheritance and evolving risk is critical to engineering works. Recently, Griffiths et al. (2012) restated the role of landscape evolution and four-dimensional ground models in engineering projects.

Engineering geomorphology is the application of geomorphological theory and methods to engineering projects (Fookes et al., 2005, 2007). It provides spatial context for engineering works and allows engineers to assess the impact of engineering on the environment and landscape and evaluate the risks and implications of landform change for the human population (Giardino and Marsten, 1999; Lee et al., 2004; Fookes et al., 2005, 2007).

In practice, engineering geomorphology involves the mapping of breaks and changes in slope and subsequent interpretation of landforms and landform processes.
The Geological Society of London (1982) recognised four types of engineering geomorphology maps: morphological, morphographic, morphochronological, and morphogenetic (Figure 1.2). Morphological maps display only landforms, without interpretation, whereas the correct shape and scale of landforms are shown on morphographic maps. Morphochronological maps emphasise the ages of landforms relative to one another, and the origin and evolution of landforms are depicted on morphogenetic maps. An engineering geomorphology map may contain elements of these four map types. Alternatively, a set of maps, each of a different type, may be generated for a project. Most researchers superimpose morphographic, morphochronological, or morphogenetic maps on morphological maps. For example, Siddle (2000) presented landslide features over time at Blaencwn in Wales using a combined morphographic and morphogenetic map, and Griffiths (2001) showed morphological and morphogenetic maps related to the Channel Tunnel project.

Doornkamp et al. (1979) illustrated the application of engineering geomorphology to five case studies. The geomorphology of each site was first interpreted using aerial photographs or topographic maps, and then mapped in more detail based on field investigations. For each case study, maps of site landforms and geology were generated at potential engineering sites in less than 200 days. Pitts (1979) focussed on slope gradient changes related to local geology at the Axmouth-Lyme Regis Undercliffs in Devon, UK, as part of an in-depth study of landslides along the cliffs. His study is an example of detailed mapping where air photograph interpretation is not feasible due to steep, densely vegetated terrain.
Figure 1.2. Four types of engineering geomorphology maps described by the Geological Society of London (1982).
Although engineering geomorphological mapping is important, additional analysis is required once maps are produced. The maps must be integrated into interpretations of processes and the evolutionary history of landforms (Bell, 1998; Giardino and Marsten, 1999). For example, Cardinali et al. (2002) produced a set of multi-temporal landslide inventory maps showing landslides that occurred between 1941 and 2001, along with zones of landslide hazard and risk, in Umbria, central Italy. Delmonaco and Margottini (2007) included geomorphological investigations and mapping in their assessment of the landslide susceptibility and hazard in the Bamiyan valley, Afghanistan. Geomorphometry – the quantitative analysis of the land surface – is another application of mapping. It uses topographic point grids such as Digital Elevation Models (DEMs) and GIS to map and model landforms and changes (Hansen and Lichti, 2002; Hengl and Reuter, 2009).

**Remote Sensing**

I used long-range terrestrial digital photogrammetry in my research on the Vajont and Madison Canyon slides to acquire data in inaccessible and dangerous areas. The methodology used in this research closely follows that of Sturzenegger (2010), who applied photogrammetry to several case studies to derive discontinuity properties and validate the technique.

Other examples of the use of photogrammetry in engineering geology studies include Brideau et al. (2012), James and Robson (2012), Westoby et al. (2012), and Clayton et al. (2013). Brideau et al. (2012) used long-range photogrammetry to characterise the rock mass that failed at Chehalis Lake, British Columbia, in 2007. Clayton et al. (2013) created three-dimensional models from air photographs of the Mitchell Creek landslide in British Columbia to map landslide evolution and lineation development over time. James and Robson (2012) and Westoby et al. (2012) compared a new open-source photogrammetric technique to laser scanning and traditional close-range photogrammetry at hand sample to kilometre scales. They concluded that comparable results can be obtained at much lower cost and with little technical knowledge using the new photogrammetric technique.

Another widely used remote sensing technique in engineering geology is laser scanning. Oppikofer (2009) used high-resolution Digital Elevation Models (DEMs) created from aerial and terrestrial laser scans to characterise and monitor slope
movements at various sites in Switzerland, Canada, and Norway. Dunning et al. (2009) extracted slide geometry, discontinuity orientations, and failure kinematics of the Namling Landslide in Bhutan from terrestrial laser scans. Metzger et al. (2009) demonstrated the application of Coltop 3D, a software for structural analysis, to terrestrial laser scans.

Analytical Techniques and Numerical Modelling

Analytical and numerical methods used to assess slope stability range from relatively simple kinematic analyses and limit equilibrium solutions to intricate continuum, discontinuum, and hybrid simulations (Stead et al., 2006; Stead and Eberhardt, 2013). With increasing computational capabilities, sophisticated numerical modelling is becoming more widely used in landslide studies. Nonetheless, slope systems are inherently data-limited, and thus modelling, especially of case studies, is restricted by the availability of field data.

Many authors have applied numerical modelling to slope stability to better comprehend underlying mechanisms, kinematics, and dynamics. Specific studies that have directly influenced this dissertation include Brideau (2010), Eberhardt et al. (2004), Kalenchuk (2010), and Leith (2012). Brideau (2010) investigated the three-dimensional effects of topography and discontinuity sets on rock slope failure kinematics using conceptual kinematic analysis, limit equilibrium, and two- and three-dimensional discontinuum simulations. He emphasised the importance of lateral confinement, as well as discontinuity persistence and block size. Eberhardt et al. (2004) used a suite of numerical modelling codes, including continuum, discontinuum, and hybrid simulations, to characterise the Randa Slide in Switzerland and to investigate progressive failure and initiation mechanisms. They outlined the benefits and limitations of each technique applied. Kalenchuk (2010) used state-of-the-art three-dimensional modelling to assess the complex geometries, shear zone parameter heterogeneities, landslide zones, and fluctuating groundwater tables of the Downie, British Columbia, and Beauregard, Italy, landslides. She validated model results with field observations and used the calibrated models to test possible future developments at each site. Leith (2012) used a two-dimensional finite difference code to survey the effects of endogenic and exogenic processes on in-situ stress distributions and slope stability in two valleys in Switzerland.
He demonstrated that geomorphic and geologic interactions are important factors in slope evolution. Ambrosi and Crosta (2011) also simulated stress distributions in mountain valleys by applying the three-dimensional finite difference FLAC3D code to slope morphology and deformation mechanisms.

Sophisticated dynamic analyses of seismic interactions with slopes are a relatively recent addition to numerical modelling literature and are one of three main methods used in assessing slope instability in earthquakes – pseudostatic analysis, permanent-displacement analysis, and stress-deformation analysis (Jibson, 2011). Pseudostatic analysis is historically the most popular method, largely because of its simplicity. It treats dynamic perturbation of a slope system as a static, permanent body force in a limit-equilibrium analysis. It simplifies dynamic motion greatly and is overly conservative. The method also does not allow for study of post-failure behaviour and is highly dependent on the choice of seismic coefficient value.

The permanent-deformation method was spearheaded by Newmark (1965). Since his pioneering work, three types of permanent-deformation analysis have evolved. The rigid-block approach is the original Newmark method. The decoupled approach separates the dynamic calculation from the rigid-body calculation and requires ground acceleration records at the ground surface. The coupled approach models the dynamic response and permanent displacement concurrently. Examples of applications of the permanent-deformation method can be found in Hsieh and Lee (2011) and Rathje and Antonakos (2011).

Stress-deformation analysis relies on the construction of sophisticated finite or discrete element models with dynamic input. It provides the most accurate representation of reality, but requires site-specific, high-quality data. Three examples of its application are in Bazgard et al. (2011) and Moore et al. (2011, 2012). Bazgard et al. (2011) derived boundary conditions for dynamic models in PFC3D and FLAC3D. Moore et al. (2011, 2012) applied Ricker wavelets and earthquake records to UDEC models of Randa and Rawilhorn in Switzerland and observed topographic and damage amplification effects related to morphology and tension cracks. In a recent study of seismic effects on large open pits, Damjanac et al. (2013) modelled two- and three-dimensional topographic amplification and stiffness contrast effects in FLAC and
FLAC3D. They also determined a dynamic factor of safety and compared results with those from pseudostatic analyses.

1.3. organisation of dissertation

Chapter 1 provides the motivation and objectives of the dissertation, as well as its background and references to influential literature. The main body of the dissertation is four chapters on two case studies: Chapters 2, 3, and 4 on the Vajont Slide and Chapter 5 on the Madison Canyon Slide. The two cases were chosen as being representative of catastrophic, structurally controlled landslides with biplanar failure surfaces, but with different triggers and failure styles. Chapter 3 has been published, and Chapters 2, 4, and 5 are to be submitted as journal papers. Some parts of Chapters 2 and 4 have been published in conference proceedings. Chapter 2 provides the regional geomorphological setting of the Vajont Slide, as well as detailed engineering geomorphology maps and their interpretations. It addresses gaps in knowledge of Vajont Valley and the geomorphic framework of the 1963 landslide. Chapter 3 focusses on the Vajont failure scar, using photogrammetry models to characterise and classify the morphology of the scar. The scar morphology has significant implications for failure mechanisms. In Chapter 4, I incorporate geomorphological and geological data and observations from Vajont into a series of numerical models. Two- and three-dimensional continuum and discontinuum models test existing hypotheses of failure geometry, the influence of tectonic and geomorphic processes on slope stability, and initiation mechanisms. Although a major topic of discussion for the Vajont Slide, an investigation of the high terminal velocity of the sliding mass is beyond the scope of this dissertation. Figure 1.3 shows the linkages among the three chapters on the Vajont Slide.

I treat the Madison Canyon Slide in one extended chapter (Chapter 5), which includes descriptions of the regional geomorphological and geological setting, laboratory, field, and photogrammetry results, detailed engineering geomorphology maps, and numerical modelling. This chapter represents the most extensive study of the landslide since Hadley (1964) reported on it shortly after the disaster.
Chapter 6 compares and contrasts the two case studies in terms of conditioning factors, damage, mechanisms, debris morphology, and initiation, and briefly discusses implications for large, catastrophic rockslides. It also discusses the new integrated approach that I applied to the two case studies, its advantages, and its disadvantages. The chapter concludes with a presentation of the main contributions and conclusions of the dissertation, as well as avenues for future work.

Figure 1.3. Summary of aspects of the Vajont Slide addressed in Chapters 2, 3, and 4.
Chapter 2.

Engineering Geomorphological Characterisation of Vajont Valley and the Vajont Slide, Italy\(^2\)

Abstract

Although the 1963 Vajont Slide in Italy has been extensively studied for over 50 years, its regional geological and geomorphological contexts have been neglected. In this chapter, I use field observations and remote sensing data to summarise the geological and geomorphological features of Vajont Valley, and comment on the interaction between endogenic and exogenic processes that conditioned the north slope of Monte Toc to fail. I present the first detailed pre- and post-failure engineering geomorphology maps of the slide area. The maps delineate two main landslide blocks and several sub-blocks, compressional and extensional zones, and secondary failures in the deposit that provide new insights into the kinematics, dynamics, and evolution of the slide. Finally, I discuss the origin of the Vajont Gorge and a prehistoric failure that occurred at the same location as the 1963 event. I propose that the prehistoric failure was a deep-seated gravitational slope deformation (sackung) that initiated during deglaciation and continued to slowly move until the catastrophic failure in 1963. The gorge was created due to the deformation of the southern slope of Vajont Valley.

Keywords: Vajont Slide; engineering geomorphology; regional geomorphology; endogenic processes; exogenic processes; sackung; rock slope damage

\(^2\) To be submitted as:
2.1. Introduction

The 1963 Vajont Slide, which resulted from catastrophic failure of the north slope of Monte Toc in the Vajont Valley in northeastern Italy (Figure 2.1), is one of the most studied landslides in the world, with over 200 publications addressing its technical and social aspects. Over the past 50 years, researchers have investigated the structure, lithology, hydrogeology, kinematics and dynamics, and effects of the slide (see Genevois and Ghirotti, 2005; Superchi et al. 2010; and Paronuzzi and Bolla, 2012, for summaries of the literature on the Vajont Slide). Notwithstanding this remarkable research effort, no scientists have examined in detail the local and regional geomorphological factors that conditioned the slope for catastrophic failure.

Semenza (2001) presented a palinspastic reconstruction of the north slope of Monte Toc (Figure 2.2), and Rossi and Semenza (1965) published two 1:5000-scale geological maps showing the lithological units and structures comprising the slope before and after the 1963 slide. Most subsequent research has referenced these maps. Recently, however, Ravagnan (2011), Superchi (2012), Massironi et al. (2013), and Wolter et al. (2014) have used field measurements, terrestrial and aerial photogrammetry, and LiDAR data to determine the orientations of discontinuities and fold axes and to trace important structures such as faults and folds. Superchi (2012), Paronuzzi and Bolla (2012), and Bistacchi et al. (2013) have provided revised geological models of the area affected by the Vajont Slide.

Giudici and Semenza (1960), Carloni and Mazzanti (1964a, 1964b), Rossi and Semenza (1964, 1965), Selli and Trevisan (1964), Broili (1967), Semenza (1967), Hendron and Patton (1985), and Guerricchio and Melidoro (1986) mention the geomorphological setting of the slide, although it was not the focus of most of these bodies of work. Broili (1967) describes the morphology of the 1963 failure surface, including irregularities related to structural features, and the role of clay in the failure. Hendron and Patton (1985) produce the first geomorphological map of the slope based on 1960 air photographs (Figure 2.3). They illustrate scarps, depressions, and areas of movement, and note that a prehistoric failure, first described by Giudici and Semenza (1960), could have been recognised on air photographs prior to 1961. An unpublished report, which describes hazards and risk in Vajont Valley, includes maps illustrating
geomorphological features (Colleselli, 2005). Only Guerricchio and Melidoro (1986), however, focus exclusively on the geomorphology of the Vajont Valley. They highlight the complexity of the geologic evolution of the valley and provide sketches and maps of processes and landforms, including prehistoric and recent mass movements, watersheds, and drainage.

Figure 2.1. Setting of Vajont Valley and the Vajont Slide in northeast Italy. Structure abbreviations are as follows: CB = Croda Bianca Thrust, CT = Col Tramontin Fault, CE = Col delle Erghene Fault, CTo = Col delle Tosatte Fault, MB = Monte Borgà Thrust, Sp = Spesse Thrust, ES = Erto Syncline, MS = Massalezza Syncline. Red line indicates location of cross-section in Figure 2.4. (Modified from Massironi et al., 2013.)
Figure 2.2. Palinspastic reconstruction of the Vajont Slide, with 1 representing the first instability in post-glacial (?) times, following through time to 8, representing the events of 1963. CV = Calcare di Vajont (Vajont Limestone), FF = Fonzaso Formation with clays, FS = thin strata of the Fonzaso and Soccher Formations, So = Soccher Formation, CE = Col delle Erghene Fault. (From Semenza, 2001.)
Figure 2.3. First geomorphological map of the Vajont area. (From Hendron and Patton, 1985.)
This chapter contextualises the Vajont Slide by identifying features in the valley that conditioned the north slope of Monte Toc for failure. I describe the mechanisms, kinematics, and dynamics of the failure using detailed engineering geomorphology maps. My approach incorporates structural geology, engineering geomorphology (following Fookes et al., 2005, 2007, and Griffiths et al., 2012), and the interactions between these disciplines. To study these interactions, I adopt an approach similar to Gerber and Scheidegger (1969), Leith (2012), and Whalley (1974), studying endogenic and exogenic processes. Leith (2012) defines endogenic processes as those stressing a rock mass due to the Earth’s geodynamic system (e.g., tectonics, isostasy, volcanism) and exogenic processes as those stressing a rock mass due to gravitational or climatic influence (e.g., weathering, landsliding, glacial and fluvial erosion). For this chapter, I include anthropic activities in exogenic processes. I contribute to the body of literature on the slide by: (1) describing the geomorphology of the Vajont Valley, (2) providing engineering geomorphology maps of the north slope of Monte Toc before and after the 1963 catastrophe, (3) discussing the interactions between endogenic and exogenic processes that conditioned the north slope of Monte Toc for failure, and (4) commenting on the mechanism of the slide.

2.2. Regional Setting of the Vajont Valley

2.2.1. Stratigraphy and Tectonics

Vajont Valley is located in the Dolomites, or eastern Southern Alps of Italy, which are dominated by Mesozoic-Cenozoic sedimentary rocks deposited on the margin of the ancient Tethys Sea (Masetti and Bianchin, 1987). The Belluno Basin, in which Vajont Valley would eventually form, originated when rifting of the Tethys basin generated an N-S-oriented fault system in carbonate rocks (Masetti and Bianchin, 1987). Units deposited in the rift basin include Mesozoic limestones, calcarenites, marls, and flysch. The 1963 Vajont Slide occurred within carbonates of the Fonzaso and Soccher formations. Of particular note in the context of this chapter are thin clay layers within the Fonzaso Formation, hypothesised to be of volcanic origin (Bernoulli and Peters, 1970), and closely associated with the sliding surface. Figure 2.4 illustrates the stratigraphic sequence found at Vajont.
Figure 2.4. Stratigraphy of the Vajont Valley area. a) N-S cross-section of the units found in Vajont Valley, located as indicated in Figure 2.1. (Modified from Semenza and Ghirotti, 2000.) DP = Triassic Dolomia Principale, SF = Jurassic Soverzene Formation, IF = Jurassic Igne Formation, CV = Jurassic Calcare di Vajont (Vajont Limestone), FF = Jurassic Fonzaso Formation, So = Jurassic-Cretaceous Soccher Formation, SR = Cretaceous Scaglia Rossa, Q = Quaternary sediments, a = Quaternary alluvial gravels. Red outline indicates where Hendron and Patton (1985) described the stratigraphy in b), which is the detail of units involved in the sliding zone of the 1963 Vajont Slide. (Modified from Hendron and Patton, 1985.)
Several compressional events preceded and accompanied the elevation of the Southern Alps. The WSW-vergent Dinaric phase of deformation, in the Paleogene, created folds with N-S-oriented hinges (Ravagnan, 2011). Uplift of the Southern Alps began in the Middle Miocene when S-SE-verging thrust faults deformed the metamorphic basement (Sauro et al., 2013) and continues today, as evidenced by strain accumulation and seismicity (D’Agostino et al., 2005). Thrusting in the region is related to high horizontal stresses and corresponding high horizontal to vertical stress ratios (see Martinetti and Ribacchi, 1980 and Montone et al., 1999 for stress maps of the area).

Vajont Valley itself follows the Erto Syncline, a recumbent fold with a hinge plunging approximately 20° E-ESE. This structure likely formed during the Miocene Neoalpine deformation phase. The refolded southern limb of the Erto Syncline coincides with the hanging wall of the Belluno Thrust and with the asymmetric Belluno Anticline, whereas the northern limb is stretched and overturned under the Monte Borgá and Spesse thrusts, fault splays that were transported passively on the Belluno Thrust (Massironi et al., 2013). The southern refolded limb of the Erto Syncline is the site of the prehistoric and 1963 failures and is responsible for the characteristic chair shape of the north slope of Monte Toc (Figure 2.5).

Another fold that has shaped the morphology of the north slope of Monte Toc is the recently mapped Massalezza Syncline, which is associated with the Dinaric deformation event. This syncline is oriented roughly N-S and is responsible for the characteristic bowl shape of the 1963 failure scar (Figure 2.5). Deformation that produced the Erto and Massalezza fold systems created complex interference patterns in the sliding zone of the Vajont scar (Ravagnan, 2011; Massironi et al., 2013; Wolter et al., 2014).

Faults were also instrumental in shaping Vajont Valley. The mouth of the valley intersects the N-S-trending Col delle Tosatte Fault, which crops out in the eastern wall of Piave Valley. This structure is a reverse fault that dips to the east (Massironi et al., 2013). Faults within the Vajont Slide area itself include the Col delle Erghene and Col Tramontin faults. The E-W-trending, normal Col delle Erghene Fault acted as a rear
release surface in the 1963 failure. The Col Tramontin Fault is the lateral release surface on the east of the slide and a minor splay of the Croda Bianca reverse fault.

Figure 2.5. a) Long-range \( f = 50 \) mm photogrammetric model of the refolded chair-like limb of the Erto Syncline taken from Longarone. b) Aspect map of the slide area, demonstrating the bowl shape of the failure scar due to the Massalezza Syncline. White curve outlines the failure scar and symbols indicate orientation of strata.
2.2.2. Geomorphology and Climate

Vajont Valley has a relief of approximately 2200 m; the highest peak, Monte Duranno, has an elevation of 2652 m asl and the valley floor has a minimum elevation of 452 m asl. The valley is flanked by 26 peaks, all above 1200 m asl, and has two main tributaries, the Mesazzo and Zémola streams (Figure 2.1).

Vajont River is a tributary of Piave River, which has a watershed area of 4130 km$^2$ and flows from the eastern Southern Alps to the Adriatic Sea near Venice. The Piave River watershed contains numerous large prehistoric and historic landslides (Eisbacher and Clague, 1984; Coppola and Bromhead, 2008; Dykes et al., 2013), of which the 1963 Vajont Slide is the most famous.

Vajont Valley is located in a humid continental climate zone, with average annual rainfall between 1200 and 2300 mm. Hendron and Patton (1985), Besio (1986), and Fabbri et al. (2013) summarise the hydrogeological conditions in Vajont Valley, with emphasis on the north slope of Monte Toc. Fabbri et al. (2013) emphasise the importance of karst in controlling groundwater flow and deep aquifers and in limiting the number of springs in Vajont Valley. Three piezometers installed in the sliding mass on the north slope of Monte Toc before the catastrophic 1963 failure indicate an overpressurised groundwater system controlled by impermeable clay layers or shear zones resulting from the prehistoric Vajont failure. Hendron and Patton (1985) suggest the presence of two water tables, one above the impermeable clays and controlled by water-level fluctuations in Vajont Reservoir and a second, deeper one in the Calcare di Vajont (Vajont Limestone). Figure 2.6 illustrates how precipitation and the Vajont Reservoir interacted to control groundwater levels within and below the slide mass.
The glacial history of the region is poorly known, as the record has been largely obscured by erosion, subsequent mass movements, and karst development, especially in Vajont Valley. Nevertheless, evidence of glaciers during the most recent Pleistocene glaciation (Marine Oxygen Isotope Stage 2) includes morainal and glaciofluvial deposits on the lower slopes of valleys, cirques, and U-shaped valleys (Colleselli, 2005). Piave Valley, for example, hosted a large glacier that extended from a 100 km$^2$ ice field south to the Po Plain (Orombelli et al., 2004). A small (<500 m thick) glacier probably occupied Vajont Valley (Castiglioni, 1940), separated from other valley glaciers by a few nunataks. Pellegrini et al. (2005) and Carton et al. (2009) suggest at least three phases of retreat of the Piave glacier, dated to between $^{14}$C 16.2±50 and 15 ka BP and documented by lateral moraines, kame terraces, and marginal moraines.
2.3. Vajont Valley Landforms and Processes

2.3.1. Longitudinal and Transverse Profiles along Vajont River

Profiles drawn parallel and perpendicular to Vajont River highlight geomorphologic anomalies in the valley. Figure 2.7 illustrates the locations of the profiles. The longitudinal stream profile (Figure 2.8) shows three knickpoints in the current stream channel, as well as the deposit of the 1963 landslide. The highest knickpoint (a) is located approximately 2 km downstream from the source of the stream, where Vajont River leaves its steep headwater tributary valley and enters the broader trunk valley. A less obvious knickpoint (b), a farther 5.5 km downstream, marks the location where Vajont River enters the residual lake behind the Vajont Slide debris. The most dramatic knickpoint (c), 12.6 km downstream from the source of the stream, is located at the downvalley limit of the slide debris, where Vajont River formerly flowed through a narrow gorge.

![Figure 2.7](image)

*Figure 2.7.* Map of profiles along (blue curve) and perpendicular to (white lines) Vajont River. Dashed golden curve outlines the 1963 slide scar. Green line indicates location of profile in Figure 2.16. (DEM from Friuli-Venezia Giulia region.)
Figure 2.8. Profile of Vajont River from its headwaters (left) to its confluence with Piave River. Labelled red points indicate knickpoints discussed in the text, and dashed line shows the average river gradient. The light grey curve shows the pre-1963 topography near the Vajont Slide, and the orange points indicate pre-1963 knickpoints.

Transverse profiles were constructed at 1-km intervals along Vajont River (Figure 2.9). They show the change of Vajont Valley from a narrow (< 2 km in width) valley with V-shaped cross-section to a much broader valley (up to 6 km in width) with a parabolic or U-shaped cross-section. This change corresponds to the major reduction in stream gradient at the first knickpoint in Figure 2.8, indicating that the wider valley has a much lower gradient than the narrow section of the valley in the headwaters. The Vajont Slide occurred in the widest part of the valley (profiles 11 and 12 in Figure 2.9) and infilled part of the gorge at the mouth of the valley. The gorge remains downstream of Vajont Dam (profile 13).

To assess the changes along the stream further, the chord-height (c/h) ratio of each perpendicular profile was calculated in a manner similar to the approach used for dams (Figure 2.10; Walter, 1962), the chord being the width and the height the depth of the valley. All c/h ratios are greater than one, indicating the valley is broader than it is deep at all points along its length. The lowest c/h ratios occur at profiles 4 to 7, where the valley is narrow and has a steep gradient. Profile 9 has the highest c/h value, as the valley is broad and shallow here. Profile 11, taken at the broadest point in Vajont Valley (approximately 5900 m wide), has a corresponding height of 1390 m, giving a c/h ratio of 4.3.
Figure 2.9. Topographic profiles perpendicular to Vajont River at 1-km intervals. All profiles are oriented such that the viewer is looking downstream. VE = vertical exaggeration.
Analysis of pre-1963 topography revealed a significant knickpoint just upstream of the Vajont Dam with 40 m of elevation change (knickpoint b’ in Figure 2.8), and another, smaller one (15 m elevation change) further upstream (knickpoint a’). The latter knickpoint occurs at a slight narrowing of the valley, which broadens downstream until it reaches knickpoint b’. At this point, the river channel narrows significantly into the Vajont Gorge. Perpendicular profiles of the pre-failure topography (Figure 2.11) illustrate the narrowing of the valley bottom downstream. Profile 4 shows the Vajont Gorge, as well as the Pian della Pozza (the flat area above the gorge on the left). The break in slope uphill of the Pian della Pozza and the bulge coincides with the limit of the 1963 failure.

Figure 2.10. Chord-height ratios for the 13 perpendicular profiles, with labels corresponding to those in Figure 2.7 and Figure 2.9.
Figure 2.11. Pre-1963 topographic profiles perpendicular to Vajont River. a) Map of the locations of the profiles. Dashed golden curve represents the future 1963 failure outline. b) Profiles with labels corresponding to those in a). All profiles are oriented such that the viewer is looking downstream. VE = vertical exaggeration.
2.3.2. Field Observations of Features in Vajont Valley

Several notable geomorphic features characterise Vajont Valley (Figure 2.12). Endogenic processes have created dip slopes on both the north and south valley walls. The south valley wall follows the refolded limb of the Erto Syncline and the Belluno Anticline at the site of the Vajont Slide, whereas the upper slope of the north wall opposite the Vajont Slide is controlled by folded strata on the hanging walls of the Spesse and Monte Borgá thrusts. These structures have contributed to several large landslides, including the Vajont failure on the south wall and the Pineda and Monte Salta failures on the north wall. The failures have substantially altered the landscape of the valley, changing the morphology of slopes and probably damming the Vajont River several times.

Another notable feature in Vajont Valley is the Palestra di Roccia below Casso (Figure 2.13). The strata dip gently to the east along the southern limb of the Erto Syncline. Tectonic and lithological settings, erosion, and weathering, however, have produced a steep wall. The steep cliff coincides with the more resistant beds of the Fonzaso and Soccher Formations in a resistant-over-recessive sequence, as noted by Francese et al. (2013). The Palestra may have been a source of instability in the past, and is a source of recent minor rockfalls (Figure 2.13).

Carbonate dissolution has altered some of the pre-existing landforms in the valley. Karst landforms such as caves and towers are especially common in the eastern part of the valley and in Zémola Valley (Figure 2.14). Aerial photographs taken before the 1963 Vajont Slide reveal several karst features on the north slope of Monte Toc. Hendron and Patton (1985) and Guerricchio and Melidoro (1986) refer to a karstic plain above the Vajont failure scar, characterised by scattered dolines, which they commented might have affected the hydrogeology of the slope. In the field, I observed a relatively flat, hummocky area above the Massalezza catchment, as well as rock pinnacles and caves consistent with karst topography. The low slope of the karst area is due to the topographic inheritance of the Belluno Anticline, which karst has exploited. The Croda Vasei (Figure 2.12d) appears to be a remnant karst tower dissected by fissures and caves. Several gullies on the slope are dry and seem to disappear abruptly, suggestive of subterranean drainage.
Figure 2.12. Geomorphic features of Vajont Valley. a) Monte Borgà, the Monte Borgà and Spesse thrusts, and the Palestra di Roccia. b) The La Pineda source area and deposit. c) The narrow Vajont Gorge seen from Piave Valley. d) Karstic plain and Croda Vasei tower above the Vajont Slide.
Figure 2.13. Photogrammetry model with a focal length of $f = 50$ mm of the Palestra di Roccia showing sources of rockfalls (white curves), bedding (gold curves), and fractures (red curves). Long-range photogrammetry was conducted from a distance of 500 m as part of field investigation. See Wolter et al. (2014) for more details.

Figure 2.14. Examples of the karst topography in the headwater area of Vajont River near Erto. “Caves” include overhanging areas.
Both ends of Vajont Valley are anomalous. On the east, the valley is bordered by the San Osvaldo pass, a saddle separating Vajont and Túana rivers from Cimoliana River (Figure 2.1). It is possible that these drainages were connected in the past. On the west, the mouth of the Vajont River flows through an extremely narrow, deep gorge. This gorge begins just upstream of the Vajont Dam site, is 1 km long and 30 m wide at its narrowest, and was exploited to build the dam. If the gorge existed in pre- or synglacial periods, glacial erosion would have focussed on this anomaly in the topography, increasing strain and damage in the area, as suggested for a similar situation at Eiger (Jaboyedoff et al., 2012). It is a young feature and has yet to be definitively explained. Nevertheless, its significant role in undercutting and oversteepening the toes of the valley slopes and contributing to the Vajont failures is clear.

The mouth of the Mesazzo Stream presents an example of the complex geomorphic palimpsest in Vajont Valley (Figure 2.15). Folded and faulted Eocene Flysch is overlain by tan-coloured, fine-grained, poorly graded, laminated glaciolacustrine silts that grade upward into tan-coloured, fine-grained, poorly graded, compacted, sandy foreset-bedded delta deposits. The delta deposits, in turn, are overlain by light brown, coarse-grained, well-graded, rounded, horizontally stratified gravels, cobbles and boulders with a sandy matrix that are probably ice-proximal outwash deposits. Above the outwash deposits is a grey to red, well-graded, compacted, matrix-supported diamicton with angular gravel to cobbles clasts. This unit is most probably a till, which in turn is covered by the pink-tan, coarse-grained, well-graded, matrix-supported debris of the large Pineda landslide. This stratigraphy suggests the following sequence of events: 1) ice advances and impounds a lake in which glaciolacustrine sediments are deposited, 2) a delta progrades into the lake, 3) ice encroaches on the area, and outwash sediments are deposited between the valley wall and ice margin, 4) the area is overridden by ice, depositing till, and 5) during or after glacial retreat, the La Pineda failure deposits material on the sequence.
Figure 2.15. Photograph (a) and sketch (b) of Quaternary sediments at the mouth of Mesazzo Stream, comprising poorly graded delta deposits overlying Flysch, possible well-graded outwash deposits; matrix-supported, well-graded till; and well-graded La Pineda landslide debris.

2.3.3. Landscape related to Rock Mass Strength (RMS)

Selby (1980) and Moon (1984) demonstrated the connection between landform and rock mass strength using a semi-quantitative classification scheme based on rock mass properties. Although I did not measure rock mass properties throughout Vajont Valley, I used slope angle as a guide for the integrity of rock masses. Figure 2.16 shows the control of rock mass quality on landform evolution, with steep cliffs forming in high quality rock masses, and gentle slopes in low quality rock masses. Superchi's (2012) GSI (Geological Strength Index; Hoek et al., 1995 and Marinos and Hoek, 2000) and UCS (Unconfined Compressive Strength) estimates of each unit corroborate the RMS assessments based on slope angle. The Calcare di Vajont (Vajont Limestone) has the highest average GSI and UCS values, 67 and 150 MPa, respectively, whereas the Fonzaso a’ unit has the lowest (27 and 49 MPa, respectively). The sliding surface of the Vajont Slide occurs at the top of the units with the lowest RMS in the valley stratigraphy. Hence, the sliding surface exploited pre-existing weak lithological layers (namely the clays in the Fonzaso Formation). The Soccher b to f units above the sliding surface are
some of the strongest rocks in the valley. This strength contrast partially explains the morphology of the slide deposit: intact blocks with high rock mass quality slid on material with poor rock mass quality.

**Figure 2.16.** Slope profile showing the relationship between average slope gradient and rock mass strength (RMS), based on Selby (1980) and Moon (1984). RMS is qualitatively assessed based on the slope angles. Formation names are as in Figure 2.4: CV = Jurassic Calcare di Vajont, FF = Jurassic Fonzaso Formation (Malm and a’ units), So = Jurassic-Cretaceous Soccher Formation (a” to f units), and SR = Cretaceous Scaglia Rossa. Profile location is shown on Figure 2.7.

### 2.4. Engineering Geomorphology Mapping of the 1963 Vajont Slide Area

Numerous authors, including Griffiths and Marsh (1986), Pasuto and Soldati (1999), Siddle (2002), Lee et al. (2004), and Griffiths et al. (2012), have demonstrated the importance of the engineering geomorphology approach in slope stability studies. Engineering geomorphologists endeavor to provide spatial context for engineering concerns, assess the impact of engineering on the environment and landscape, and evaluate the societal risks and implications of landform change (Giardino and Marsten, 1999; Fookes et al., 2005, 2007). Griffiths et al. (2012) emphasise the
consideration of geological time scales, dynamic processes, and landscape evolution to engineering.

Following guidelines presented by the Geological Society of London (1982), I created detailed air photograph-based and field maps of the Vajont Slide to study the pre- and post-1963 processes that have shaped the landscape and to better understand the 1963 event. Combined morphological and morphogenetic maps based on interpretation of air photographs flown before and after the 1963 slide are shown, respectively, in Figure 2.17 and Figure 2.18. The field map was made in 2011 by surveying with a compass and clinometer roughly N-S transects spaced 50 m apart (Figure 2.19). A sample of the field map is shown in Figure 2.20, and the entire map is presented in Appendix A.

2.4.1. 1960 Map

The 1960 map (Figure 2.17) shows the morphology of the north slope of Monte Toc before the catastrophic 1963 failure. Active processes at that time included gullying, fluvial erosion, and shallow slope failures on gully walls and at the toe of the slope where it was undercut by the newly filled Vajont Reservoir. An incipient landslide, which entered the reservoir in November 1960 after the air photographs were acquired, is evident. The eroded Pineda and Colle Isolato deposits indicate the complex history of landsliding in the valley. Gently sloping benches (areas with slope angles less than 25°), scarps, ridges, depressions, and hummocky terrain are evidence of movement of the north slope of Monte Toc. In their analysis of the 1960 photographs, Hendron and Patton (1985) suggested that these features were related to the prehistoric failure. There is no evidence of a headscarp in the photographs.

The largest features on the south wall of Vajont Valley are gently sloping, discontinuous benches (Table 2.1). Most of the benches occur in two clusters, one near the top of the slope and another at the toe of the slope. There are fewer benches in the mid-slope position. The Pian della Pozza (bench 5), the largest and most distinctive bench (excluding the Pineda deposit), with an area of 208,205 m², tilts gently into the slope. Bench 6, having an area of 131,661 m², dips downslope. A comparison of these two benches across the Massalezza Gully indicates that, although at similar elevations,
they were not connected before erosion of the gully. This observation implies that blocks on the east and west sides of the Massalezza Gully moved independently prior to the 1963 failure.

Figure 2.17. Pre-slide engineering geomorphology map based on interpretation of 1960 aerial photographs.
Steep scarps (> 45°) and extension lineations are present on both sides of the Massalezza Gully and cluster near the developing headscarp of the instability. The two most prominent scarps, each with a length greater than 200 m, are oriented WNW-ESE and are located in the upper slope area west of the gully; they are located along the trace of the Col delle Erghene Fault. A third scarp with a length of 300 m and a similar orientation is present east of the gully. These three scarps mark the 1963 headscarp, which first appeared as an M-shaped tension crack in 1960. A smaller scarp (75 m length) on the Pian della Pozza at the toe of the slope is likely related to the incipient 1960 failure (see below). Minor scarps also appear on the east side of the slope. Figure

Figure 2.18. Post-slide engineering geomorphology map based on interpretation of 1963 aerial photographs. Letters indicate blocks and sub-blocks as discussed in the text.
2.21 illustrates the orientations and lengths of the scarps mapped on the 1960 photographs.

Figure 2.19. Location of topographic transects used to document changes in slope below Monte Toc and to assist in creating a detailed field engineering geology map of the Vajont Slide. Dashed white curve delineates the area mapped in the field, and solid golden curve indicates the slide headscarp. (Modified from Wolter et al., 2013b.)
Three depressions were mapped within the unstable rock mass on the 1960 air photographs. The smallest depression (220 m in length) is located just downslope of Bench 12. The largest depression, with a length of 700 m, a width of 100 m, and a depth of roughly 15 m, is located at the rear of the Pian della Pozza and may be related to karst processes. Semenza (2001) originally thought this depression could have been the headscarp of the prehistoric failure, but, after further fieldwork, revised his interpretation and extended the prehistoric slide limit farther upslope.
Table 2.1. Average characteristics of the gently sloping benches mapped on the 1960 air photographs. The ID numbers correspond to those in Figure 2.17.

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<th>Elevation (m asl)</th>
<th>Length (m)</th>
<th>Width (m)</th>
<th>Area (m²)</th>
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<td>13,404</td>
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</tbody>
</table>

Figure 2.21. Polar plot of scarps and extension lineations mapped on the 1960 air photographs of Vajont Valley, showing lineation strikes and lengths. The three longer scarps correspond to the headscarp of the 1963 failure, whereas the smaller lineations are within the unstable rock mass.
2.4.2. 1963 and 2011 Maps

The post-slide 1963 and 2011 maps show that many of the same processes operating in 1960 remain active, but their influence has changed locations. Gullying occurs in different positions; many of the gullies that were active in 1960 are inactive in 2011. Material is still failing at the toe of the slope, but sediment resting on the exposed failure scar farther up the slope is also failing in response to the dramatic change caused by the Vajont Slide. Large amounts of material have accumulated in talus cones in the hinge zone of the refolded limb of the Erto Syncline (Chapter 3). Another area of active erosion is the Col Tramontin Fault, where a gully has formed between the fault and the slide scar. Large amounts of debris have moved along this gully into Vajont Lake. Rills have also formed in weak material in the shear zone, with intervening sharp pinnacles metres to tens of metres in height (Figure 2.22).

![Figure 2.22. Pinnacles of sediment at the Col Tramontin Fault.](image_url)
Processes connected with the 1963 slide mass include piping and ponding. Metre- to decametre-scale, circular depressions at the debris front were formed by piping. An obvious cluster of 20- to 30-m deep depressions, which Semenza (1967) named “Costa dei Crateri”, appeared near the Vajont Dam a few days after the 1963 slide (Figure 2.23). Ponds formed on and adjacent to the slide debris in three places after the failure. Reservoir water surged over most of the deposit immediately after the failure and left a shallow lake in the large depression behind the debris front at the hinge of the chair-shaped failure surface. This lake and the one remaining behind the dam after failure have since filled with sediment. Vajont Lake, whose intake is 1.5 km upstream of the slide debris, has also decreased in size due to sedimentation. The intake for the bypass tunnel constructed in 1961 is located at the downstream end of the lake.

Several features attest to the movement and emplacement history of the 1963 slide mass. The headscarp is asymmetric – it is higher next to the Col Tramontin Fault along the east boundary of the landslide than anywhere else, with the scarp attaining heights of close to 100 m. NNE-oriented lineations in the material near the Col Tramontin shear zone indicate that the failed mass sheared along the fault material to the north, rather than extending away from it to the west. Another area of the headscarp that is taller than average (approximately 25 m high) is near the west boundary of the landslide and coincides with the Col delle Erghene Fault. Most of the remainder of the headscarp has little relief (<10 m) and is subparallel to bedding.

Most of the slide deposit remained remarkably intact, preserving stratigraphy and, in some areas, even vegetation. Several authors (e.g., Superchi, 2012; Bistacchi et al., 2013) have suggested that two main blocks were involved in the 1963 event – a larger west block (block A in Figure 2.18) including the Massalezza Gully and a smaller east block (block B) that overrode the west block. These blocks are distinct on the 1963 air photographs, and have areas of 181 million m$^3$ and 43 million m$^3$, respectively, according to Superchi (2012). Upon closer inspection, however, two smaller (<100 m long) blocks that rest on the failure scar (blocks C and D) can also be seen above the east and west blocks. In addition, the west block itself is divisible into five sub-blocks (blocks A1 to A5) that moved relative to one another. Although the west block remained intact, differences in the morphology and orientation of ridges on the two sides of the
Massalezza Gully indicate that the gully marks a roughly NE-SW oriented, 1 km long intra-block boundary. Likewise, steep slopes in the southwest corner of the west block may mark boundaries between sub-blocks. Trenches observed in the field and identified on the 2011 map corroborate these air photograph interpretations.

![Figure 2.23](image)

**Figure 2.23.** Section through the largest of the craters at the front of the 1963 slide debris, close to the dam. Section location is the red line in Figure 2.20.

Ridges and shear/extensional lineations within the blocks provide information on movement directions, with ridges assumed perpendicular and lineations roughly parallel or oblique to displacement. It is difficult to differentiate between shear and extension lineations on air photographs; hence these were grouped into one feature class on the 1963 map. Figure 2.24a and b illustrate the lengths and orientations of the ridges and lineations, respectively, and Figure 2.25 shows areas of extension and compression based on the ridges and lineations. Sinuous ridges are assumed to be associated with compression and reverse faulting in the deposit, whereas linear ridges are probably related to normal faulting (Shea and van Wyk de Vries, 2008). The ridges tend to be hundreds and the lineations tens of metres in length, with the longest ridge 400 m and the longest lineation 130 m. The average trend of the ridges is E-W, whereas the average of the lineations is 230°/050°, suggesting movement directions to the N and NE, respectively. Looking at the different blocks, blocks A and C moved N to NE and blocks
B and D NNW (Figure 2.18). The movement directions agree generally with Broili’s (1967) interpretation, in which most movement vectors are oriented N.

The spatial distribution, or intensity, of the ridges and lineations indicates where the most deformation occurred in the deposit. Ridges tend to form toward the fronts of blocks and sub-blocks and in weaker areas of the debris. The ridges in block B are larger than those in block A, with amplitudes of decametres versus metres, suggesting more deformation. Lineations cluster in the Col Tramontin Fault area, block A2, and at the front of blocks A3 and B. The Col Tramontin lineations, as well as a few in sediments resting on the failure scar, appear to be dominantly related to shearing. Lineations in block A2 are oblique to movement direction, and, based on field observations of trenches, seem to be extensional.

Some of the ridges and lineations seem to have formed after initial emplacement. For example, the ENE-WSW-trending ridge on block A4 east of the Massalezza Gully and north of the residual lake indicates movement in a NNW direction, which contrasts with the general N to NNE movement direction of the west block. The ridge may thus have formed in response to collision between blocks A and B. A comparison of Figure 2.18 and Figure 2.25 further highlights the evolution of the deposit since 1963. The deposit has extended northward into the gorge, southward into the hollow behind the debris front where the residual lake is, and eastward into Vajont Lake. Block C is not present in Figure 2.25, and may have been disturbed and transported with sediment moving down the failure scar.

The debris front forms a 60 m-high, 1.5 km long cliff on the floor of the former Vajont Reservoir that is still obvious today. It follows the pre-failure gorge, with a protrusion to the north near Massalezza Gully. The debris did not impact the opposite valley wall; rather reservoir water prevented the debris front from moving farther across the valley.
Figure 2.24. Rosette diagrams of a) ridge and b) shear/extensional lineations showing the trends of the features and their lengths.
Another interaction of the debris with the reservoir is evident from the distribution of surviving 1963 vegetation on the debris. Most of the slide deposit was washed by the large displacement wave, leaving a bare surface. However, block C and most of blocks A1, A2, and B are vegetated. Figure 2.26 shows the Bosco Antico (“old forest”) preserved in the southwest corner of the debris.
Figure 2.26. Bosco Antico on blocks A1, A2, and C, indicating the displacement wave did not wash over this area.

Although the 1963 air photographs were flown shortly after the catastrophic failure, secondary adjustments are apparent in the deposit. The best example is slumping of the east block into Vajont Lake. As mentioned, lineations also suggest local adjustments after the landslide. The 2011 maps show further evolution of the slide deposit, with a new rockfall deposit from the east-central failure scar overriding part of block B.

Humans have modified the slopes of Vajont Valley for thousands of years (for example, see Hendron and Patton, 1985 and Semenza, 2001 for mention of Roman roads). Agriculture, road building, and settlement have exploited and altered the natural landscape. Since 1963, parts of the slide mass have been levelled, material moved and removed, and new roads constructed across it. The Vajont River no longer erodes the toe of the slope, as it flows through the bypass tunnel constructed in 1961.
2.5. Discussion and Conclusions

2.5.1. Interaction between Endogenic and Exogenic Processes and a New Hypothesis for the Vajont Failure

This chapter has highlighted the endogenic and exogenic processes that have shaped Vajont Valley and the north slope of Monte Toc. Interactions between these two suites of processes have created this unusual landscape. Figure 2.27 and Figure 2.28 highlight the specific causing and triggering factors of the Vajont Slide, framed in a Venn diagram and interaction matrix (Hudson, 1992). Causing factors include both endogenic and exogenic processes, which conditioned the slope for failure. Triggering factors at Vajont, heavy precipitation and reservoir level fluctuations, are exogenic in this case and were the final influences to initiate catastrophic failure. Figure 2.27 illustrates the interaction zone between endogenic and exogenic processes near and at the slope surface. The influence of past seismicity is unknown, but earthquakes undoubtedly had some effect on the Vajont slope, as it is located in an active tectonic area. Figure 2.28 further demonstrates the relationships between endogenic and exogenic processes. Each variable in the matrix (along the diagonal, with colours corresponding to those in Figure 2.27) affects other variables, and is in turn influenced by them. The interaction between pore water and the weak clays is particularly important to the triggering of the catastrophic Vajont Slide. The clays, already fatigued by erosion and slow deformation, were further softened by infiltration of precipitation and weakened by infilling and rapid drawdown of the reservoir.

The Vajont Gorge epitomises the relationship between various processes. Gorges are ephemeral landscape elements that form to bring a stream into equilibrium with downstream base level. In this case, local base level is the Piave River valley. What is unclear, however, is the cause of the disequilibrium that induced gorge formation. Was Vajont Valley a hanging valley that Vajont River eroded after Pleistocene deglaciation? Are the formations in the area of the gorge more resistant and thus more difficult to erode? Were there periods of tectonic uplift? Was Vajont River blocked at some time in the past? Has long-term, deep-seated rock slope movement below Monte Toc “pinched” the valley in the vicinity of the gorge? Or has the gorge formed due to a combination of these processes? Glaciers probably deepened Piave Valley at the expense of Vajont
Valley, leaving the latter "hanging". I demonstrated in Section 2.3.3 that Vajont Limestone, the unit in which the gorge is incised is stronger than surrounding units; thus rock in the gorge is more resistant to erosion. The mouth of Vajont Valley is also located on the west limb of the Massalezza Syncline, but there is no evidence for such large movements along faults in the gorge to suggest tectonic uplift is the dominant factor in gorge formation. Although karst features are common in the area, there is little evidence of dissolution in the gorge, supporting the argument that the rocks in the gorge are more resistant to erosion.

**Figure 2.27.** Interaction between endogenic and exogenic processes causing and triggering the Vajont Slide. The triggers are in bold.
I conclude that the most likely dominant cause in the formation and persistence of the epigenetic gorge is mass movement, as Ouimet et al. (2008) propose. The knickpoint associated with the gorge in the pre-reservoir valley is, as mentioned above, just upstream of the Vajont Dam, where Giudici and Semenza (1960) hypothesised there to be a prehistoric failure based on cataclasite, landslide debris on the north valley slope, and slope morphology. Rather than being a discrete catastrophic failure, I interpret the older Monte Toc landslide to be a sackung that has been active throughout the Holocene, if not much longer. The term sackung is defined here as slow deformation of a slope, and is not restricted to ridge crest lineations (Zischinsky, 1969). Bulging of the toe of the slope may have displaced Vajont River to the north and required it to erode a
resistant rock mass. This hypothesis also explains the lack of a distinct headscarp prior to 1963 and the hummocky terrain, scarps, depressions, and basal shear zone (cataclasite) characteristic of a sackung. The slope movement records support this hypothesis, as they show small amounts of movement immediately after instruments were installed and well before the 1963 failure. The only conflicting evidence is Colle Isolato and the old gorge it covers. According to Semenza (2001, and references therein), Colle Isolato (“isolated hill”) comprises two smaller landslide blocks that were part of the prehistoric failure, which were separated from the main deposit by fluvial incision. It is possible, however, that Colle Isolato is a separate mass that slid away from the toe of the sackung, as Broili (1967) alluded to and much like the 1960 failure into Vajont reservoir three years before the catastrophic slide. Another possibility, voiced by several authors, including Guerricchio and Melidoro (1986), is that Colle Isolato is a remnant of a landslide from the north side of the valley, possibly from the Palestra di Roccia. Semenza (2001) discredits this hypothesis, citing stratigraphy and bedding orientations as evidence. I suggest that the old river channel Giudici and Semenza (1960) mapped at the mouth of the valley was blocked by the smaller failure at the toe of the south slope of Vajont Valley, and subsequently avulsed southward for 500-600 m, eroding the landslide debris and creating the current channel. Farther east, a much smaller mass that Semenza (2001) identified as MC (“Masserella”), and similar to the Colle Isolato material, suggests other failures originated from the toe of the Monte Toc slope prior to the catastrophic 1963 failure.

2.5.2. Interpretation of the Vajont Slide Kinematics and Dynamics

I now discuss the implications of my field and air photograph observations and mapping for the kinematics and dynamics of the 1963 slide. The asymmetry of the failure scar supports Hendron and Patton’s (1985) hypothesis that the failure surface stepped up to intersect the Col Tramontin Fault to the east. The rock mass would have been stable had the sliding surface continued to follow bedding planes, which dip 20° E, in the east. The greater thickness of the sliding block on the east (approximately 200 m versus 50 m on the west) also reflects the control exerted by the Erto and Massalezza synclines. Thus brittle fracture and crack propagation through intact rock would have played an important role in the failure, particularly in the east.
The formation of the two main blocks and five sub-blocks within the west block relates to the kinematics of the slide. In order to fail, the rock mass required the kinematic freedom afforded by the creation of these blocks. Had these blocks not formed, the mass would have been more stable and might not have failed at all. Chapter 4 of this dissertation further demonstrates the importance of multi-block formation using numerical simulations.

The morphology of the deposit also has implications for the dynamics of the slide. A possible sequence of events based on the pre- and post-failure morphologies of the Vajont slope is proposed in Figure 2.29:

1. A proto-Vajont River eroded the topography caused by tectonic uplift, faulting, and folding.
2. Glaciation eroded the region, broadening the valley floor and oversteepening valley slopes (Hutchinson and Kwan, 1986).
3. Deglaciation unloaded slopes, affecting in situ stresses. Vajont River, loaded with sediment, re-established its course.
4. Movement began as slow, sackung-type deformation, probably during deglaciation. The sackung displaced Vajont River to the north, which eroded a narrow channel (ancient gorge).
5. A prehistoric toe failure occurred in the western zone of the sackung, possibly blocking Vajont River. This event may have been the source of Colle Isolato, which was identified by Semenza (Semenza, 2001). Other smaller failures such as the MC mass Semenza identified likely had similar, though less dramatic effects on the topography and river.
6. Vajont River avulsed south, eroding the new landslide deposit and incising a deep, narrow gorge.
7. After construction of the dam, movement accelerated due to reservoir filling, and another toe failure in the same location as the prehistoric landslide collapsed into the reservoir in 1960. The same year, the rear M-shaped tension crack opened.
8. After a period of lower movement rates, the rock mass accelerated in late 1962 due to another filling of the reservoir, which was followed by a rapid drawdown and filling. The western upper blocks (blocks C, A1, and A2 in Figure 2.18) acted on the lower passive material (blocks A3, A4, and A5), eventually failing and moving downslope and across the valley. Sliding of block A created kinematic freedom, allowing block B to fail and come to rest on top of the debris of the western block. Smaller blocks (blocks C and D) detached from the headscarp after the main blocks, coming to rest on the failure scar above the main deposit. Secondary failures included landsliding of material remaining
on the upper sliding surface. Drainage of the slide mass caused piping and pore water pressure dissipation.

An element of uncertainty is the exact timing of emplacement relative to the displacement wave. Did the displacement wave arrive before the east block came to rest? Or afterwards? Given the short timeframe of the sequence of events, less than a minute, either hypothesis is possible.

The above sequence compares well with Semenza’s palinspastic reconstruction (Figure 2.2). The main difference is the rate of movement of the prehistoric failure. I suggest a slow sinking rather than episodic catastrophic failure, which is supported by the slope morphology (hummocks, scarps, tension cracks, ridges) and lack of a headscarp prior to 1963. This lack of a scarp is not due to erosion, as several prehistoric failures within Vajont Valley and the broader region have clear preserved rear release surfaces. The cataclasite Semenza identified could feasibly have formed over millennia of slow deformation, as is common for creeping failures (e.g., Downie Slide; Kalenchuk, 2010). I also provide explanations for the previous river channel and Colle Isolato and Masserella masses Semenza identified.

2.5.3. Spatial and Temporal Elements of Damage

Damage can be a significant contributor to the location, size, and depth of a landslide, as well as debris fragmentation and overall landslide behaviour. It is intricately linked to the endogenic and exogenic processes that shape a slope. The spatial variation in structural damage is especially apparent at Vajont. Interpretation of Superchi’s (2012) GSI values and discontinuity sets and the discontinuity sets and roughness analysis presented in Chapter 3 indicates that the rock mass on the east side of the Vajont Slide (block B) is weaker and its failure surface has higher roughness than that on the west side (block A). Rocks in the Col Tramontin area are the most damaged in the slide area. However, the rougher surfaces and lack of kinematic freedom in the east prevented it from moving significantly before the west block failed. The centre of the failed rock mass, where the hinges of the Erto and Massalezza Synclines intersect, was highly folded. This area would have been a focus of stress in the slope and resembles a Prandtl wedge transition zone between the active and passive blocks of the slide (Mencl, 1966; Chapter 4 of this dissertation). Geomorphic damage is closely associated with
structural damage. For example, Massalezza Gully, eroding a channel down the slope, follows the damaged hinge zone of Massalezza Syncline. Vajont River has undercut the toe of the slope, and has created a narrow gorge.

Figure 2.29. Hypothesised sequence of events leading to the catastrophic failure of 1963. Stratigraphic symbols are the same as in Figure 2.4a.
Ridges and lineations are surficial indicators of damage. At Vajont, they were present both before and after the catastrophic 1963 landslide. Prior to failure, scarps and lineations clustered at the top of the incipient landslide, suggesting increased damage intensity at the future headscarp location. Post-failure, ridges formed in areas of compression. The ridges in block B are larger in amplitude than those in block A. Hence, block B deformed more and was thus more damaged than block A material. The block B material was weaker than block A material before failure, and as a result damage was amplified in this block during emplacement.

The debris also provides indications for subsurface changes in damage. Key zones of the debris are sheared and highly fragmented, whereas much material remained relatively intact during emplacement (Figure 2.30). The key zones are the basal shear zone and secondary shears separating intact blocks. The secondary shears were mapped by Rossi and Semenza (1965), and are still evident in the deposit today. If my hypothesis that the north slope of Monte Toc was a sackung before catastrophically sliding is correct, the basal shear zone accumulated damage for millennia and progressively softened before failing. Hence, both spatial and temporal damage are significant components at Vajont.

**Figure 2.30.** Focal length $f = 100$ mm photogrammetry model of the debris front of the Vajont Slide, showing intact blocks of Soccher Formation units, separated by minor (white dashed lines) and major (black dashed line) shears.
Looking at the broader setting of Vajont Valley, another area of high structural damage is the zone surrounding the Monte Borgà and Spesse thrusts. The faults have thrust Triassic-Jurassic stratigraphy over the Cretaceous Scaglia Rossa formation. At the unconformity between these two sequences, several rockfalls have occurred, including the large Monte Salta Slide (Figure 2.12).

2.5.4. Why Here?

As mentioned at the beginning of this chapter, the Piave region is a tectonically and geomorphically active area with many large mass movements. The Alleghe, Borta, Pineda, Fadalto, Borca di Cadore, and Cinque Torri failures are some examples (Coppola and Bromhead, 2008; Dykes et al., 2013). Most of these landslides initiated after retreat of late Pleistocene glaciers in response to in situ stress changes due to unloading and debuttressing of slopes. Many landslides also significantly changed the pre-existing topography, such as the Fadalto failure 15 km downstream of Vajont Valley that caused Piave River to avulse and created St. Croce Lake (Pellegrini and Surian, 1996). A cursory investigation of satellite imagery shows that many mass movements in the region have occurred on dip slopes. Several are also structurally controlled (Dykes et al., 2013). The stratigraphy at Vajont is not unique to that location, and other failures have occurred in clay-carbonate sequences. Glaciers occupied and eroded most valleys in the region, and stress changes due to unloading were ubiquitous. Vajont Gorge is very narrow and focusses in situ stresses, which increase strain and damage in the toe of the slope. Glacial and fluvial erosion of the slope have augmented damage in the valley, as unloading decreases the vertical in situ stresses and thus increases the ratio of horizontal to vertical stresses (Goodman, 1980). Vajont Valley is also located in a compressional tectonic area, leading to high horizontal stresses. Nonetheless, these in situ stress conditions are not exclusive to Vajont Valley. The linkage to changes in reservoir level is also not uncommon in the region. Many natural and artificial dams are present in the Piave watershed. Coppola and Bromhead (2008) investigate the stability of natural dams, their lasting effect on river morphology, and how anthropic modification has allowed the continued use of the dams for hydroelectric power generation. They cite the 1959 Pontesei Dam failure, when filling of the reservoir reactivated an ancient landslide, as another example of landsliding related to reservoir level changes. The
exact sequence of filling and rapid drawdown at Vajont, however, would not have been repeated elsewhere.

It is thus apparent that the factors that caused the Vajont Slide are not unique to Monte Toc. Nevertheless, their combination caused a landslide of unique and catastrophic behaviour. Structural features delimited the slide and contributed to its characteristic chair shape, creating an active-passive regime. Glacial erosion and debuttressing weakened the slope and contributed to changing stress conditions. Fluvial action undercut the toe of the slope. The changing reservoir level and weak clays triggered a catastrophic failure involving a small number of blocks that remained largely intact.

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Chapter 3.

A Morphologic Characterisation of the 1963 Vajont Slide, Italy, using Long-Range Terrestrial Photogrammetry

Abstract

The 1963 Vajont Slide in northeast Italy is an important engineering and geological event. Although the landslide has been extensively studied, new insights can be derived by applying modern techniques such as remote sensing and numerical modelling. This chapter presents the first digital terrestrial photogrammetric analyses of the failure scar, landslide deposits, and the area surrounding the failure, with a focus on the scar. I processed photogrammetric models to produce discontinuity stereonets, residual maps and profiles, and slope and aspect maps, all of which provide information on the failure scar morphology. My analyses enabled the creation of a preliminary semi-quantitative morphologic classification of the Vajont failure scar based on the large-scale tectonic folds and step-paths that define it. The analyses and morphologic classification have implications for the kinematics, dynamics, and mechanism of the slide. Metre- and decametre-scale features affected the initiation, direction, and displacement rate of sliding. The most complexly folded and stepped areas occur close to the intersection of orthogonal synclinal features related to the Dinaric and Neoalpine deformation events. My analyses also highlight, for the first time, the evolution of the Vajont failure scar from 1963 to the present.

Keywords: Vajont Slide; failure scar; structural controls; photogrammetry; morphology; roughness

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3.1. Introduction

The Vajont Slide, which happened in northeast Italy on October 9, 1963 (Figure 3.1), is one of the most famous natural disasters in modern history. Approximately 250 million m$^3$ of rock slid catastrophically from Monte Toc into the newly created Vajont Reservoir, generating a 140-m-high displacement wave that overtopped the Vajont Dam and flooded villages both upstream and downstream of the landslide. The landslide itself did not cause any injury, but the displacement wave claimed 1910 lives. Due to its social and economic impacts, as well as its complex behaviour, the Vajont Slide is the most studied landslide in the world. Currently, over 150 publications address its technical and social components (Genevois and Ghirotti, 2005; Superchi et al., 2010; Paronuzzi and Bolla, 2012).

Although extensively studied, several aspects of the event remain poorly understood. Among them are the unexpectedly high velocity attained by the sliding mass, the kinematics and dynamics of the slide, and the underlying causes of the failure. The failure mechanism has been debated for the past 50 years. Müller (1968) suggested that the north slope of Monte Toc had been moving slowly prior to the catastrophic failure in 1963, controlled by creep and material fatigue. He invoked high water pressures in discontinuities and mass thixotropy (the sudden reduction in friction caused by shocks or shear deformation) to explain the high velocity of the mass. Semenza (2010 and references therein) argued that the slide was a reactivation of an ancient failure, but this hypothesis was disputed for decades. Mantovani and Vita-Finzi (2003) theorised that the failure surface was a normal fault plane and rejected Semenza's reactivation hypothesis. However, other authors including Trollope (1980), Hendron and Patton (1985), Tika and Hutchinson (1999), and Semenza and Ghirotti (2000) supported the idea. It is now accepted that the 1963 event was indeed a reactivation of an ancient failure on the same slope.
Figure 3.1. Topography of the study area. a) Location and geomorphic features of Vajont Valley and the Vajont Slide. 1 — Sliding surface (arrows indicate general directions of movement); 2 — Vajont Slide deposits; 3 — remnants of the Pineda Slide deposits; 4 — Col Tramontin Fault; 5 — Croda Bianca Fault; 6 — Col delle Erghene Fault; 7 — Massalezza Gully, which was transposed and rotated during the 1963 event; 8 — Mesazzo Stream; and 9 — Zemola Stream. b) Window of an $f = 50$mm photogrammetric model indicating the southern limb of the Erto Syncline and a plot of average slope angles along a profile close to that of the photogrammetric model. The sliding surface follows the orientation of these beds, forming a characteristic chair shape.
The significance of clay interbeds in the landslide shear zone, another topic of discussion, was not recognized by researchers until Hendron and Patton (1985) demonstrated their relevance, both to the geological and hydrogeological settings and the mechanism of the Vajont Slide. They suggested that they had acted as aquitards, separating a perched aquifer in the Fonzaso Formation from a deeper one in the Vajont Formation. The failed rock mass appears to contain formational clays as well as clay gouge along shear zones. It is still not known, however, how widely both types of clay were distributed within the failure zone, and if and how they interacted with tectonic folds and faults on Monte Toc.

Other ideas to account for the behaviour of the slide include bedding plane slip and stress release (Skempton, 1966), heat-induced vapourisation (Habib, 1975; Goguel, 1978; Nonveiller, 1987), increased fluid pressure due to frictional heat (Voight and Faust, 1982), thermo-poro-mechanical softening (Vardoulakis, 2002), and crack coalescence (Kilburn and Petley, 2003). Based on their re-evaluation of slope displacement data, Petley and Petley (2006) argued that the deformation mechanism changed at the time of the catastrophic failure. Specifically, they suggested that ductile deformation occurred between 1960 and 1962, followed by brittle deformation immediately before the slide. They further argued that most of the deformation occurred within the clay seams, not the carbonates.

The role of rock bridges and brittle fracture, clay, folds, and dilation, and the importance of structural and geomorphological controls on the failure have not yet been investigated in detail. Most studies assume a planar, uniform, two-dimensional failure surface. A better understanding of these important aspects of the landslide can be obtained using state-of-the-art analytical techniques, including numerical modelling and remote sensing. Until recently, however, researchers have not applied these techniques. Most simulations of the slide have used a limit equilibrium approach that neglects the third dimension, the complexity of the failure zone, and the slope's evolution over time. A few exceptions include Hendron and Patton's (1985) three-dimensional limit equilibrium analysis and Ghirotti's (1992, 1994) and Sitar and MacLaughlin's (1997) two-dimensional distinct element and discontinuous deformation analysis models.
I have published a photogrammetric characterisation of the Vajont Slide (Wolter et al., 2011). Here I present the first long-range terrestrial photogrammetric models of the failure scar and deposit, and the surrounding area. These models have allowed us to evaluate previously inaccessible parts of the failure scar, as well as other areas, and have provided a more complete understanding of the structures and features, such as discontinuity sets, folds, and faults, that affected the slide. In addition to presenting the results of my photogrammetric analyses, I provide here a preliminary morphologic classification of the Vajont Slide, based on several parameters. My objectives are to: 1) describe and quantify elements of the failure scar, such as discontinuities, steps, and folds, using photogrammetric models validated by field observations; 2) determine trends in two- and three-dimensional roughness dimensions and orientations; 3) examine failure scar evolution from 1963 to the present; 4) classify areas of the scar from very smooth to very rough in an initial attempt to delineate areas of low and high effective shear resistance; and 5) integrate my structural and roughness data in order to more fully understand the failure mechanism. The rationale for the study is that improved characterisation of the failure scar affords a better understanding of the slide dynamics, kinematics, and controls, and forms an essential foundation for future three-dimensional numerical modelling of the Vajont Slide.

3.2. Background

3.2.1. Surface Characterisation

Researchers use several approaches on different scales to describe geologic and geomorphic surfaces (Figure 3.2). Tribology, or the study of surface interaction, friction, lubrication, and wear (Bhusan, 2002), pertains to anthropogenic surfaces on the scale of micrometres. It employs fractal and wavelet methods, as well as statistical parameters, to describe surface roughness, or the deviation in a surface from a reference plane (Bhusan, 2002). The tribological approach can be extended to millimetre and centimetre scales, but not to scales that cross several surfaces.
Figure 3.2. Scales of roughness. Millimetre-to-metre scale roughness can be quantified using Barton's JRC or similar measures. Metre-to-kilometre scale roughness has yet to be quantified for landslide sliding surfaces and scars.

Within the discipline of rock mechanics, methods used to quantify roughness focus on individual discontinuity surfaces at the millimeter to metre scale. The terms smoothness and waviness refer to small- and large-scale roughness on individual discontinuity planes that range in amplitude from, respectively, millimetres to centimetres and centimetres to metres (International Society for Rock Mechanics Commission on Standardisation of Laboratory and Field Tests, 1978). Discontinuity roughness, including smoothness and waviness, is an important engineering property because it affects the shear strength, shear direction, and dilation along and between discontinuity planes, which in turn influence rock mass properties (Milne, 1990; Palmstrøm, 2001; Cai et al., 2007). However, roughness is also one of the most difficult properties to measure and quantify. It is scale-dependent and spatially highly variable. One of the first proposed roughness measures is Patton's (1966) \( i \) angle, which is the angle between the average discontinuity plane and an asperity. Fecker and Rengers (1971) developed a plate method for quantifying roughness, in which plates of different sizes are placed on a joint plane and the orientations of those plates are represented on stereonets, or...
stereographic projections of a sphere onto a plane. The \( \theta \) angle and average orientation of the plane can then be determined. Another measure is the Joint Roughness Coefficient (\( JRC \)) of Barton (1973, 1976) and Barton and Choubey (1977) (Barton, 2011). They showed that the shear strength along a discontinuity is related to the \( JRC \) by the following empirical relation:

\[
\tau = \sigma_n \tan \left( \varphi_r + JRC \log_{10} \left( \frac{JCS}{\sigma_n} \right) \right)
\]  

(1)

where \( \tau \) is the shear strength along the discontinuity plane, \( \sigma_n \) the normal stress on the plane, \( \varphi_r \) the residual friction angle, and \( JCS \) the Joint Compressive Strength. \( JRC \) is determined by measuring the maximum amplitude of undulations over a two-dimensional profile length or by comparing the profile with standard profiles that are assigned \( JRC \) values. \( JRC \) is based on profile lengths of 10 cm, and, although Barton and Bandis (1982) provided a conversion to larger scales, they only extrapolated the \( JRC \) determination to 10 m. Haneberg (2007) proposed that roughness could be estimated by rotating a digital point cloud into a coordinate system defined by the dip-line, the strike-line, and the normal to the discontinuity plane after interpolating the cloud to create a surface. Two-dimensional profiles can be taken in any direction on this rotated three-dimensional discontinuity surface, and \( JRC \) and \( i \) values can be estimated from equations or charts. Most current methods utilize two-dimensional profiles of three-dimensional roughness. They include the asperity-amplitude method (Oppikofer, 2009), the fractal roughness-length method (Fardin et al., 2001; Kulatilake et al., 2006; Baker et al., 2008), and the best-fit method (Sturzenegger and Stead, 2009). Researchers have also attempted to characterize roughness using Fourier transforms, power spectral analysis (Bistacchi et al., 2011), Riemannian statistics (Rasouli and Harrison, 2010), and 2D and 3D roughness parameters (Reeves, 1985; Grasselli et al., 2002; Tatone and Grasselli, 2009, 2010; Nasseri et al., 2010; Tatone et al., 2010).

On a larger scale, researchers have described geomorphic features using a variety of qualitative and quantitative methods. They have applied geomorphometry, or quantitative land surface analysis, to several morphologic features, including valley slopes (Evans, 2012), glaciers (Herzfeld et al., 2000), and landslides (Van Den Eeckhaut et al., 2005; Martha et al., 2010). Trevisani et al. (2009, 2012), for example, employed variograms, textural analysis, and geostatistics to quantify surface roughness on a scree
slope and in an alpine basin (Figure 3.2). Grohmann et al. (2011) investigated six different surface roughness techniques and concluded that roughness is best captured by the slope standard deviation method. Kim et al. (2012) analysed the texture of a high-resolution DEM to characterise geomorphic features of landslides, and Leith (2012) determined the influence of geomorphic processes on slope stability based in part on a topographic roughness metric related to block statistics. These quantitative studies are largely limited to the deposits of landslides and do not exclusively consider their sliding surfaces or scars.

To date, characterisations of failure surface morphology have been mainly descriptive and qualitative, or have focussed on small-scale features (Willenberg et al., 2008; Dunning et al., 2009; Jaboyedoff et al., 2009; Oppikofer, 2009). Eberhardt et al. (2004) and Willenberg et al. (2008) have emphasised the role of structures such as discontinuities and rock bridges in the initiation and propagation of the Randa Slide, and Jaboyedoff et al. (2009) and Brideau et al. (2011) have done likewise for the Frank Slide. Dunning et al. (2009) qualitatively described the shape of landslides from a geomorphic perspective. Oppikofer (2009) related steps in 2D sections of landslides to their kinematics and dynamics using profiles of tens of metres in length that cross several discontinuity planes; such long profiles are outside the accepted length for use of the Barton and Bandis method.

The quantitative roughness methods mentioned above are mainly limited to the scales of smoothness and waviness, and should be extrapolated with care to features that are larger than 10 m and that cross several discontinuities. In contrast, conventional morphologic methods are mainly descriptive and qualitative, or are used on larger scales than failure scars. The morphology of landslide sliding surfaces and scars thus remains to a large extent descriptive and unquantified.

3.2.2. Geological Setting of the Vajont Slide

The Vajont Slide is located in the Vajont River valley in the Dolomites region of northeast Italy, approximately 100 km north of Venice (Figure 3.1). The steep concave valley walls are formed of carbonate rocks interbedded with clays and marls, and have roughly 2200 m of relief. Prior to the Pleistocene, the valley was narrow and occupied by
a proto-Vajont River. Subsequently, it was widened and deepened during a series of glaciations. During the Holocene, the Vajont River has created an epigenetic gorge over 250 m deep and as narrow as 30 m at the valley mouth (Figure 3.1b).

Although it is widely accepted that Vajont Valley was glaciated (Castiglioni, 1940), most of the erosional effects of glaciers on the valley have been obscured by karstic processes and deep-seated gravitational movements. Karstic processes are most obvious above the 1963 failure scar, where there is an area of dolines and piping. Prehistoric landslides dammed the Vajont River on several occasions, altering the valley floor and inducing lacustrine and deltaic deposition. Landslides in Vajont Valley are fundamentally geologically controlled, but contributing factors may include precipitation, oversteepened slopes, and long-term progressive brittle failure and creep.

The Vajont failure scar is affected by structures associated with regional Alpine deformation. The Dolomites have experienced several deformation events, of which the most pertinent to this study are the Dinaric and Neoalpine phases (Doglioni and Siorpaes, 1990; Ravagnan, 2011; Massironi et al., 2013). The Dinaric phase involved east–west compression during the Paleogene, and created the NNE-plunging Massalezza Syncline that separates the east and west halves of the failure scar (Superchi, 2012). The Miocene Neoalpine phase involved north–south compression and formed the east-plunging Erto Syncline. The refolded southern limb of the Erto Syncline gives the failure scar its characteristic chair shape (Figure 3.1b). The hinge of the chair is most apparent on the west side of the scar and becomes more arcuate eastward. The slide is bounded by the Col Tramontin Fault along its east lateral boundary and the Col delle Erghene Fault at the western headscarp (Figure 3.1a). The Col Tramontin Fault is older than the two folding events and is related to extension of the Piave basin during the Mesozoic. The Col delle Erghene Fault is relatively young and is possibly related to the Neoalpine deformation. Paronuzzi and Bolla (2012) suggested the presence of another fault, the Vajont Valley Fault, parallel to the hinge of the Erto Syncline.
3.3. Methodology

This chapter is part of an ongoing research collaboration between Simon Fraser University and the Universities of Padova and Bologna in Italy. The data presented augment the work of Ghirotti (1992, 1994), Ravagnan (2011), and Superchi (2012), which include traditional geotechnical and geological investigations and modelling.

I performed terrestrial long-range photogrammetry on the Vajont landslide scar and adjacent area in 2010 and 2011. I used a Canon 50D digital camera with 20, 50, 100, 200, and 400 mm focal length lenses to photograph the slide deposits, failure scar, and rock outcrops in the area surrounding the slide. Only my photogrammetry projects on the failure scar are presented in this chapter; they include one \( f = 20 \) mm, one \( f = 200 \) mm, and four \( f = 400 \) mm focal length models. The different focal lengths allowed us to capture features at different scales and resolutions. Distances between the photographed objects and their respective camera stations range from 100 to 2500 m. Sturzenegger (2010) provides details on the field, processing, and georeferencing methods used in this study.

I processed the photogrammetry models using CalibCam, DTM Generator, and 3DM Analyst (ADAM Technology). Each model was subsequently georeferenced in Polyworks IMAlign (Innovmetric, 2007) using 5-m-resolution airborne LiDAR images. Metre-scale and larger features are thought to affect general slide behaviour, and are thus significant in my study. The scale of the relative accuracy of the photogrammetry models remains orders of magnitude smaller than these features, and the relationship between them is unaffected by georeferencing. Smaller features, although important at the local, discontinuity scale (Barton, 1976; Barton and Choubey, 1977), cannot feasibly be incorporated into kilometre-scale models of the slope, which is the ultimate objective of my research. Wyllie and Mah (2004) discuss how scale affects the strength and kinematics of a slope failure.

I imported discontinuity orientation measurements made in 3DM Analyst on the \( f = 400 \) mm models into IMAlign to transform them into the new reference system. I then exported the transformed measurements to DIPS (Rocscience 2013a) to visualise discontinuities on equal-area stereonets. No attempt was made to examine roughness
on the discontinuity scale, as features on this scale only affect local behaviour, as
mentioned above. I mapped intersection lineations between discontinuities and the
failure scar in Polyworks by tracing them with vectors. The vectors were correlated with
known discontinuities from field and photogrammetric surveys.

Analyses of surface morphology were performed on the photogrammetry models
in Polyworks IMInspect (Innovmetric, 2007). I fit several reference shapes to the $f = 20$
mm and $f = 200$ mm model point clouds to approximate the general shape of the failure
surface and determined that a cylinder best represents the surface. I then created error,
or residual, maps by comparing the best-fit cylinder to each photogrammetric model. The
residual maps show the elevation difference between the models and the best-fit
cylinder, and highlight the large-scale, three-dimensional, concave and convex forms on
the present-day failure scar. Two-dimensional profiles based on the $f = 200$ and 400 mm
models, and taken parallel and perpendicular to the general direction of slide movement
and perpendicular to fold axes, display the smaller-scale undulations and roughness of
the failure scar. The 2D profiles provide the orientation of maximum roughness,
amplitude-to-length measurements, and the overall shape of the surface along the
profiles. I also applied Fecker and Rengers’ (1971) plate method at several locations
across the failure scar. I fitted virtual circular discs of 1, 5, 10, 15, and 20-m radius to the
$f = 200$ and 400 mm photogrammetric point clouds in IMInspect, and plotted the
orientation of each disc on a stereonet to determine the $i^*$ angle for each disc category.
The $i^*$ angle differs from the accepted $i$ angle used in discontinuity shear strength
characterisation, because the discs may traverse several discontinuity surfaces. I
believe that the extension of the method beyond a single discontinuity surface is
acceptable because I am not attempting to comment on shear strength properties. I
calculated the Fisher’s dispersion coefficient, $K$, for each $i^*$ angle distribution to
determine how much scatter is present in the data for each region of the failure scar:

$$K = \frac{N}{N - |\bar{R}|}$$  \hspace{1cm} (2)

where $N$ is the number of joint orientations considered and $\bar{R}$ is the resultant mean of the
orientation set (Goodman, 1980).
After analysing the \( f = 20, 200, \) and 400 mm point clouds in Polyworks, I exported them to ArcGIS (ESRI) and converted them to raster format. I then processed each raster, as well as the existing airborne LiDAR DEM, to produce slope and aspect maps. The angle and aspect of a slope change significantly over small areas of rough terrain. Conversely, the slope angle and aspect should be relatively constant in smooth terrain. The slope and aspect maps highlight different directions of roughness — the slope angle provides a measure of downslope variability, whereas aspect emphasises across-slope variability. To further quantify roughness, I calculated the range and standard deviations of slope and aspect over 10 × 10 m rectangular blocks using the block statistics tool in ArcMap, similar to Grohmann et al. (2011) and Leith (2012). The tool calculates statistics, here range (block maximum – block minimum) and standard deviation, over user-defined, non-overlapping neighbourhoods comprising multiple raster cells (ESRI, 2012). Null values in the photogrammetry models related to vegetation and occluded areas were not considered in the calculations of block statistics. The size of the blocks is arbitrary, but I chose a size that best illustrates roughness features on the failure scar at the scale of interest.

I used the parameters that I derived from photogrammetry models, including the profile amplitude/length (\( A/l \)) ratio, \( i^* \) angle, \( K \) value, and slope and aspect block statistics values, to create a preliminary classification of the failure scar morphology. I divided the failure scar into four roughness classes ranging from very smooth to very rough. These zones indicate areas where the resistance to sliding would have been low or high. The complete methodology is illustrated as a flowchart in Figure 3.3.
3.4. Results and Discussion

Figure 3.4 shows the photogrammetry models that I analysed, and Table 3.1 summarizes the details of the models. They show the effect of the different focal length lenses: the relative resolution, or ground pixel size, of a given model increases with increasing focal length and decreasing distance to the object of interest. The point clouds of these models served as inputs for further analysis in Polyworks and ArcMap.
Figure 3.4. Selected photogrammetric models to determine failure scar morphology; refer to Table 3.1 for details of the models. Note the different resolutions, which are related to the focal length of the lens. The overview model was produced with an $f = 20$ mm lens, whereas the other models are $f = 200$ mm (Model 2) and $f = 400$ mm windows of the failure surface. The discs in the $f = 400$ mm models represent mapped discontinuities related to the failure surface (red discs) and steps in the failure surface (blue discs).
Table 3.1. Details of photogrammetric models used in the analysis of the failure scar (model numbers correspond to those in Figure 3.4).

<table>
<thead>
<tr>
<th>Model number</th>
<th>Focal length (mm)</th>
<th>Distance (m)</th>
<th>Ground pixel size (cm)</th>
<th>Plan accuracyb (cm)</th>
<th>Depth accuracy (cm)</th>
<th>Overall accuracy (cm)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>20</td>
<td>2,000</td>
<td>46.9</td>
<td>14.1</td>
<td>80.4</td>
<td>82.9</td>
</tr>
<tr>
<td>2</td>
<td>200</td>
<td>1,800</td>
<td>4.2</td>
<td>1.5</td>
<td>17.7</td>
<td>17.9</td>
</tr>
<tr>
<td>3</td>
<td>400</td>
<td>800</td>
<td>0.9</td>
<td>0.4</td>
<td>2.8</td>
<td>2.9</td>
</tr>
<tr>
<td>4</td>
<td>400</td>
<td>1,500</td>
<td>1.8</td>
<td>0.8</td>
<td>3.7</td>
<td>3.9</td>
</tr>
<tr>
<td>5</td>
<td>400</td>
<td>1,000</td>
<td>1.2</td>
<td>0.4</td>
<td>1.7</td>
<td>1.8</td>
</tr>
<tr>
<td>6</td>
<td>400</td>
<td>1,300</td>
<td>1.5</td>
<td>0.8</td>
<td>6.1</td>
<td>6.2</td>
</tr>
</tbody>
</table>

aObject distance is the average distance from the camera station to the object of interest.
bAccuracy reflects the error in location of each point in plan, depth, and overall.

Each stage in the analysis emphasises different aspects of the failure scar, such as folds and steps, and overall roughness. I infer that structures influencing the morphology of the present-day failure scar affected the movement direction of the slide mass and the morphology of the failure surface. Although several blocks have moved since 1963 and have changed the appearance of the scar, their behaviour is controlled by the same structures that affected the 1963 failure. Thus, the recent failures provide clues to the kinematics of block movement and slope deformation prior to and in 1963.

The two dominant structures that define the failure scar are folds and steps. Folds are crenulations in the failure scar related to the principle deformation events that affected the region. Steps are discontinuities on the failure scar, characterised by scarps that separate adjacent parts of the scar. They may be parallel, perpendicular, or oblique to fold axes and are treated separately below, followed by discussions of overall roughness, the evolution of the scar since 1963, and advantages and limitations of the methods employed.

3.4.1. Fold Analysis

Folds are important features on the failure scar and are related to the Dinaric and Neoalpine deformation events. The Neoalpine event interfered with the Dinaric event to
produce complex, elongated, dome-and-basin topography on several scales. The largest scale of folding is that of the Massalezza and Erto synclines. The trend and plunge of the hinge of the Massalezza Syncline are between 018°/33° and 058°/29°, and the Erto Syncline plunges 20°E. Both structures affect the entire north slope of Monte Toc. Smaller folds are superposed on the synclines at scales of $10^1$ to $10^{-1}$ m, and are parasitic to the large structures. For example, Figure 3.5a shows steps that are subparallel and perpendicular to folds and discontinuity sets, and second- and third-order parasitic folds in the east-central area of the failure scar.

**Figure 3.5.** Examples of different scales and types of folding. a) Different orders of folds in the east-central area of the landslide scar. The dashed lines represent major steps in the failure scar, and solid lines represent discontinuity sets. b) Tight folds subparallel to the sliding direction (indicated by white arrow). c) Dome and basin folds oblique and orthogonal to the sliding direction.
Most decimetre- and metre-scale folds change amplitude, wavelength, and orientation over short distances across the failure scar. Although visible on the $f = 400$ mm models, I could not measure them reliably given this spatial variability. Nonetheless, the general trend of these features can be derived from field photographs. Two main types of fold affect the failure scar: 1) tight folds subparallel to the movement direction; and 2) dome-and-basin folds oblique and perpendicular to the movement direction (Figure 3.5b, c). The latter are more widespread, whereas the former are localised and scattered across the failure scar. Most of the folds are confined to the lower failure scar and areas near the Col delle Erghene and Col Tramontin faults. The hinges of dome-and-basin folds on the west half of the failure scar trend approximately E–W. Those on the east half of the failure scar are more complex, with fold hinges trending NE and NW.

Figure 3.6 shows the general trend of large basins and domes on the failure scar, determined from the Polyworks residual maps. The residual maps are based on the $f = 20$, 200, and 400 mm point clouds and show deviations from the best-fit cylinder mentioned in Section 3.3. The long axes of the eastern and western domes represented by white lines on Figure 3.6 plunge to the northeast and west, respectively. The eastern dome has dimensions of approximately $400 \times 650$ m (maximum lengths of the long and short axes of the dome), whereas the western dome is smaller, $130 \times 310$ m. Two of the four basins follow the west-plunging western dome, but one plunges to the east, and the fourth plunges to the northeast. All four basins are wider than 70 m and longer than 190 m. None of the high or low features relates directly to either the Dinaric or Neoalpine deformation event. Each, however, could be influenced by interference between the events, or between the folding events and reactivated faults. Elongation of the domes and basins in the east–west direction suggests that the Neoalpine event is dominant and that the folds are mainly cascading features, with hinges perpendicular to the direction of movement. The direction of maximum roughness along the failure scar is thus parallel or subparallel to the movement direction, which indicates that the folds played an important role in the initiation of the slide and in its displacement rate. The basin at the east boundary of the scar is the exception and may be related to localised erosion between the Col Tramontin Fault and the sliding mass. Different scales of folding affect the movement differently. Small-scale folds and sedimentary features, which were observed but not measured, affect the shear strength of the failing material, whereas large-scale
folds affect the initiation, kinematics, and direction of movement. Folds oriented perpendicular to the movement direction may have inhibited kinematic movement, whereas those parallel to the movement direction had little effect or facilitated movement.

**Figure 3.6.** Example of a residual map of a photogrammetry model. This map is based on the $f = 20$ mm model and is related to the best-fit cylinder. The grey areas are vegetation or talus, and the full-colour areas are the failure scar. The scale in the legend is the difference between the point cloud and the best-fit cylinder (in metres). The scar is outlined with the black line. Domes and basins determined from the residual plots are outlined in grey, and their maximum (long) and minimum (short) axes are labelled in white. Black lines represent locations of two-dimensional profiles depicted in Figure 3.11: i) perpendicular to a large step in the west corner of the slide scar; ii) parallel to the overall movement direction (Broili, 1967); iii) perpendicular to the overall movement direction; iv) parallel to the movement direction in the central area; v) perpendicular to the movement direction in the central area; vi) perpendicular to the step in the east-central area; and vii) parallel to the step in the east-central area.
Ravagnan (2011) completed a more detailed classification of fold interference at Vajont based on field measurements and modelling in MATLAB. He investigated the folds on the failure scar in terms of regional tectonic history and determined the fold interference to be a K-type in the Thiessen and Means (1980) classification scheme, which has implications for the tectonic history and structural controls of the slope.

3.4.2. Discontinuity Analysis

Rock mass discontinuities, including joints and faults, are another important failure scar feature. I mapped them both as planes on the $f = 400$ mm photogrammetry models and as lineaments in Polyworks, and compared the results with the field and LiDAR data. Lineaments represent intersections of the discontinuities and the failure scar.

All data sets highlight the change in orientation of the failure scar from east to west related to the Massalezza Syncline. The change is most apparent when comparing the $f = 400$ mm photogrammetry windows. On the east side, the mean failure scar orientation is N to NNW (46°/002° to 36°/353°, dip/dip direction). In the centre of the scar, the beds dip to the NNE (40°/008°), and on the west side, the mean orientation is NE (34°/031°). These orientations form a girdle on a stereonet of the entire failure scar (Figure 3.7). The change in orientation is also evident in Superchi's (2012) field data and in her Coltop3D analysis of the airborne LiDAR data, and suggests the presence of a synform with an axis oriented between 018°/33° and 058°/29° (trend/plunge). The former values are derived from the mean orientations of the east, central, and west portions of the failure scar, whereas the latter are based on the asymmetric girdle formed by all data points. The average of the mean orientations and the girdle is 038°/31°. The best-fit cylinder in Polyworks corroborates the presence of the syncline, which is most likely related to the E–W Dinaric compression event. Massalezza Stream has exploited the hinge zone of the syncline and runs parallel to it.
Figure 3.7. Lower hemisphere, equal-angle stereographic projection of the discontinuities mapped across the failure scar. $S_0 =$ bedding, CT = Col Tramontin Fault, CE = Col delle Erghene Fault, MS = Massalezza Syncline, ES = Erto Syncline. Diamonds represent bedding, and crosses represent steps in the failure scar. Of the 369 poles mapped, 282 are bedding and 87 are steps. The contour interval is 4% using Fisher’s method, and the maximum density is 27.6%. Contours outline the anisotropy in the bedding orientations and indicate the Massalezza Syncline.

Although the orientation of the failure scar determined from photogrammetry is similar to that derived from other data sets, the discontinuity sets (DSs) and other random discontinuities are more difficult to compare, mainly because of biases in the data sets. The field data are limited, as most of the failure scar is inaccessible. The LiDAR data were acquired from the air, and the photogrammetry data were acquired from across Vajont Valley. Photogrammetric identification of discontinuities is hampered by occlusion and the high reflectivity of the failure scar. Many discontinuities are also only represented by intersection lineations on the failure scar.
Despite these limitations, I was able to readily identify two main features on the photogrammetry models: bedding and steps oblique to the bedding. Discontinuity sets and correlative features identified by Superchi (2012) are listed in Table 3.2.

Table 3.2. Discontinuity sets (DSs) mapped on the photogrammetry models and their counterparts derived from fieldwork completed by Superchi (2012).

<table>
<thead>
<tr>
<th>Region</th>
<th>Current chapter</th>
<th>Superchi (2012)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>DS</td>
<td>Dip (°)</td>
</tr>
<tr>
<td>East</td>
<td>S0₁</td>
<td>36</td>
</tr>
<tr>
<td></td>
<td>S0₂</td>
<td>46</td>
</tr>
<tr>
<td></td>
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<tr>
<td></td>
<td>6</td>
<td>88</td>
</tr>
<tr>
<td></td>
<td>7</td>
<td>62</td>
</tr>
</tbody>
</table>

*S0 is the bedding plane.

*Blank rows indicate discontinuity sets derived from photogrammetry with no field equivalents.

Comparison of the discontinuities mapped using photogrammetry with those mapped in the field shows the benefits of each method. Several discontinuity sets and random joints determined by photogrammetry correlate with those identified in the field, validating both methods (Table 3.2). A few sets detected in the photogrammetry projects, however, were not found in the field, and vice versa. For example, the photogrammetry models in the central area, around the Massalezza Gully, indicate the presence of sets oriented S to SE (DS3 and DS6) and NW (DS4). These sets were not found in the field; instead Superchi (2012) identified discontinuity sets oriented NE (DS1) and SW (DS7) in the field.
Grouping of individual planes into sets is complicated by folding. It is unlikely that there are seven or eight discrete discontinuity sets; rather, some sets are rotated across the Massalezza Syncline. For example, DS1 and DS4 probably belong to the same set.

Examination of a map of features on the failure scar (Figure 3.8) indicates that most of the large steps are contained within DS2, DS3, and DS4, and are oriented obliquely to the general direction of movement. These three discontinuity sets therefore contribute significantly to the failure scar roughness. Four of the 18 steps measured are parallel to DS1, two are parallel to DS6, and one corresponds to each of DS5 and DS7; one step is not associated with a discontinuity set.

I plotted the length-to-height ($L/H$) ratios of several steps in the failure scar according to their position: eastern, central, and western regions (Figure 3.9). A high $L/H$ ratio indicates a relatively long shallow step, whereas a low ratio indicates a short deep step. Low and high $L/H$ ratios carry different implications for movement. The long shallow steps affected shearing resistance and displacement of the moving mass, whereas the short deep steps likely were sources of material and acted as kinematic release surfaces. The orientations of the steps suggest that blocks may have moved in different directions, rather than as a single block. The general directions of movement inferred from the morphology of the failure scar (Figure 3.8, grey arrows) correspond well with those reported in the literature, for example by Broili (1967) (Figure 3.8, black arrows).
Figure 3.8. Lineament map of the Vajont failure scar. a) $f = 20$ mm focal length photogrammetric model of the failure scar, which served as a base map for b) lineaments identified using the $f = 200$ mm and $f = 400$ mm images. Each colour represents a discontinuity set. The black arrows illustrate Broili’s (1967) hypothesised 1963 movement directions, and the grey arrows indicate movement directions perpendicular to the mapped lineations determined in this study. $S =$ bedding plane; $L =$ lineation parallel to a discontinuity set (Table 3.2); $F_1 =$ Col Tramontin Fault, which formed before the deformation events ($D_1 =$ Dinaric, $D_2 =$ Neoalpine, with compression directions as indicated); and $F_2 =$ Col delle Erghene Fault.
Figure 3.9. $L/H$ (length/height) ratios of 18 individual steps over the three regions of the failure scar. The coloured bars correlate with discontinuity sets (see Figure 3.8 for legend). Inset sketches highlight the difference between (a) a long shallow step with a high $L/H$ ratio and (b) a short deep step with a low $L/H$ ratio. The grey arrows in the insets indicate sliding directions.

Several major structures define the Vajont failure scar, notably the Erto and Massalezza synclines and the Col delle Erghene and Col Tramontin faults. The two faults have exposed dimensions of, respectively, $30 \times 485$ m and $80 \times 780$ m (height and length) over the area affected by the 1963 failure. Their full lengths are over 2 km. I determined the orientations of the Massalezza Syncline, Col delle Erghene Fault, and Col Tramontin Fault using the photogrammetry models: $038^\circ/31^\circ$ for the Massalezza Syncline (average trend and plunge of fold axis), $59^\circ/283^\circ$ for the Col Tramontin Fault (dip and dip direction), and $55^\circ/010^\circ$ for the Col delle Erghene Fault (dip and dip direction). COLTOP 3D analysis of the LiDAR data (Superchi, 2012) indicates similar orientations: $30^\circ/000^\circ$, $38^\circ/315^\circ$, and $54^\circ/010^\circ$, respectively. I could not measure the orientation of the refolded southern limb of the Erto Syncline because its axis is covered by landslide debris. The two main faults acted as lateral and rear release zones, and the Massalezza Syncline controlled the directions of movement of the eastern and western
blocks. Step-path intersections between fractures also affect the failure scar morphologies. Examples are found in the Col delle Erghene area, at the western corner of the scar, and at the eastern headscarp, where the failure scar is bounded by steps related to DS2, DS3, DS5, DS6, and DS7. DS5 discontinuities behave as the lateral release surfaces connecting rear-release DS2, DS3, DS6, and DS7 fractures (Figure 3.8 and Figure 3.10).

Figure 3.10. Block model of the main discontinuities and steps on the failure scar, and the faults (CT = Col Tramontin Fault and CE = Col delle Erghene Fault) and synclines (ES = Erto Syncline and MS = Massalezza Syncline) that influence and bound the scar.

3.4.3. Roughness Analysis

The techniques used in this chapter are not amenable to measuring millimetre- and centimetre-scale features such as chert nodules found on the failure scar. Hence, I analysed the overall roughness of the failure scar by examining the decametre-scale and larger fold and step features using residual maps and 2D profiles in Polyworks, and block statistics of slope and aspect maps in ArcGIS. A caveat of the Polyworks residual maps, for example those in Figure 3.6 and Figure 3.11, is that the failure scar is a product of the interaction between endogenic (tectonic) and exogenic (geomorphic and climatic) processes. As mentioned, the basin along the eastern boundary is an erosional
feature, as is the one beneath the large fold in the east-central area (Figure 3.5a). They cannot be interpreted as folds because they are sources of mass movements.

Two-dimensional profiles in IMInspect allowed us to better visualize details of the failure scar (Figure 3.11). Specifically, the profiles indicate changes in the failure scar from concave to convex, as well as changes in folds and steps in the scar. They also highlight the differences between smooth and rough areas. For example, the west corner (model 3, $f = 400$ mm) is mostly smooth, but the profiles show down-dip and oblique steps and an along-strike concavity in this area (Figure 3.11b, i–iii). The profile with the largest range between minimum and maximum heights is perpendicular to the main step in the upper part of model 3. The range is 20 m over a profile length of 240 m, giving an amplitude/length ratio of $1/12$. Conversely, in the rougher central area of the failure scar, the maximum amplitude is 12 m over a profile length of 60 m in the direction of movement, which is toward the north in this region (Figure 3.11b, iv–v). The amplitude/length ratio is thus $1/5$, indicating a rougher area, as relatively higher
amplitudes occur over a shorter profile length. The topography of the central area is more varied than that of the west corner. In the west-central area, the profiles are stepped in a direction parallel to movement, with alternating concave and convex sections. In contrast, the east-central area is uniformly convex. The profiles in both the east-central and the west-central areas are more uniform perpendicular to the direction of movement. There convex areas alternate with concave areas, but the amplitudes of the highs and lows are small compared to those along the down-dip profiles, which indicates roughness anisotropy. The roughness is greatest in the down-dip, or movement, direction. To confirm this observation, I created profiles oblique to the movement direction. As expected, these profiles had smaller amplitudes than the eastern down-dip profiles.

Analyses of plate roughness conducted in IMInspect indicate smooth and rough areas of the failure surface in three ways. First, the maximum $i^*$ angle of each area indicates the relative roughness of the area with respect to the mean orientation of the surface. If the maximum $i^*$ angle is high, the area, or part of it, is rougher than an area with a lower maximum $i^*$ angle. The area with the highest maximum $i^*$ angle (78°), given a plate radius of 1 m, is the upper section of the east-central area (EC$_2$), and the areas with the lowest maximum $i^*$ angle, with the same plate radius, are the central-west and eastern areas (C$_1$ and E$_2$; see Figure 3.12a for locations of areas). The east-central area (both EC$_1$ and EC$_2$) consistently has the highest maximum $i^*$ angles across all plate sizes (Figure 3.12). The smallest maximum $i^*$ angles occur in the W$_2$ and E$_2$ areas for a 5-m-radius plate size, and in W$_1$ for the 10, 15, and 20 m plates. The maximum $i^*$ angle does not necessarily indicate a rough area; rather it might represent an anomaly such as a step in the failure scar on an otherwise smooth surface. Hence, I used a second indicator of roughness — the dispersion coefficient, $K$ (Table 3.3). This coefficient provides a measure of the tightness of data clusters and increases as clustering becomes tighter. Areas with loose clustering or wide dispersion, that is with relatively high maximum $i^*$ angles across plate sizes, are rougher on all scales. Loose clustering indicates that the roughness is not isolated to one area, such as a step, but is characteristic of the entire region. Folded zones, for example, have wide dispersion. The areas with the lowest $K$ values across all plate sizes are C$_2$ and EC$_1$, which are also visually the roughest and most folded areas of the failure scar. Areas with the least
dispersion, or tightest clustering, are $W_1$ and $W_2$. These latter two areas have outliers, which are related to steps. The third measure of roughness related to the plate method analysis is anisotropy. Anisotropy is represented on the stereonets as oval or irregular contours around the mean plane orientation (Figure 3.12d). An isotropic distribution around the mean is indicated by circular, or concentric, contours. $W_1$, $W_2$, and $C_1$ have fairly isotropic distributions; $C_2$, $EC_1$, and $E_2$ are anisotropic, with an SW–NE direction of anisotropy; $EC_2$ shows E–W anisotropy; and $E_1$ shows SE–NW anisotropy. The SW–NE anisotropy direction may indicate the influence of interference of the Neoalpine folding event with Dinaric folds, whereas the E–W anisotropy may correspond to the Dinaric event. The SE–NW anisotropy only occurs in $E_1$, which is located in the east-central fold complex (Figure 3.5a). The isotropic areas are all located on the west side of the failure scar, illustrating that this side is smoother and less complex. The high dispersion areas are anisotropic and areas with tight clustering are isotropic.

The slope maps created in ArcGIS show the folds, steps, and release surfaces of the failure scar, as well as smaller-scale features (Figure 3.13a); they also highlight E–W-trending structures. The Col Tramontin Fault and the large step in the east-central area are conspicuous features on the maps. Subvertical steps in the failure scar are also clearly visible on the slope maps. Most of the failure scar is inclined between 25° and 40°, but parts of it are much steeper (>70°). The west half of the scar is generally less steep than the east half and has fewer deviations such as steps and other subvertical features. Comparison of the LiDAR and photogrammetry slope maps shows that the slopes are less steep overall in the former than in the latter. This difference may be due to orientation bias of both LiDAR (looking from above) and photogrammetry (looking horizontally from across Vajont Valley). The post-1963 talus apron in the transition zone between the back and the seat of the chair-shaped failure scar is easily recognized because it has a lower slope (<30°) than the failure scar. Most of the landslide deposits lie on slopes less than 20°, facilitating the distinction between the failure scar and deposit. The difference in slope angle also supports the presence of the refolded, chair-shaped southern limb of the Erto Syncline. The slope maps aid in delineating morphology domains on the sliding surface. Areas in which the slope differs considerably over a small area (for example, from 5° to 85°) are rougher, whereas areas in which the slope angle is fairly uniform are smoother.
Figure 3.12. $i^*$ angle analysis, divided into a) east (E), east-central (EC), central (C), and west (W) areas. b–c) Graphs of maximum $i^*$ angle for different plate radii (m) for the central and west areas and the east and east-central areas, respectively. d) An example of a stereonet corresponding to the E$_2$ area (498 poles); distributions of plate radii and anisotropy direction are indicated.

Table 3.3. Dispersion coefficient, $K$, of each $i^*$ angle region over all plate radii.

<table>
<thead>
<tr>
<th>Area</th>
<th>K</th>
</tr>
</thead>
<tbody>
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<td>E$_1$</td>
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<tr>
<td>E$_2$</td>
<td>107.7</td>
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<tr>
<td>EC$_1$</td>
<td>88.8</td>
</tr>
<tr>
<td>EC$_2$</td>
<td>140.7</td>
</tr>
<tr>
<td>C$_1$</td>
<td>148.9</td>
</tr>
<tr>
<td>C$_2$</td>
<td>85.1</td>
</tr>
<tr>
<td>W$_1$</td>
<td>237.4</td>
</tr>
<tr>
<td>W$_2$</td>
<td>260.9</td>
</tr>
</tbody>
</table>
Figure 3.13. Slope and aspect maps derived from the photogrammetric models and LiDAR data produced in ArcGIS. a) Slope map. The white dashed line outlines the failure scar; the grey dashed line separates talus with lower slope angles from the failure scar; and the black lines indicate steps. Also shown are rough areas, where the slope changes markedly over short distances. b) Aspect map. The white dashed line outlines the failure scar; the grey lines indicate the hinges of the Erto and Massalezza synclines; the latter corresponds to the Massalezza Stream; and the black lines indicate counterscarps and ridges. CT = Col Tramontin Fault and CE = Col delle Erghene Fault.
Aspect maps (Figure 3.13b) highlight subvertical and N–S-trending features on the sliding surface, such as metre- and decametre-scale gullies. Most of the gullies visible on the failure scar are related to sediment that originally covered the scar and has slid downhill (Section 3.4.4). Most of the colluvial fans on the west side of the failure scar face NE, whereas those on the east side face N to NW. Ridges between gullies face NE and NW. Comparison of the LiDAR and photogrammetry data beneath the large step in the east-central area illustrates the evolution of the slope since the landslide. A large talus deposit has accumulated beneath the step in the past 50 years. Also evident on the aspect maps are south-facing counterscarps and ridges in the slide deposit. The chair of the southern limb of the Erto Syncline appears as a change in aspect from NE to E, and the hinge zone of the refolded syncline limb is clear.

Morphological features of the failure scar are also apparent on the aspect maps, including the W- to NW-facing Col Tramontin Fault and part of the S-facing scarp of the Col delle Erghene Fault. Some of the lateral release steps discussed in Section 3.4.2 are discernible, but they are not as obvious on the aspect maps as on the slope maps. The aspect maps clearly distinguish the two limbs of the Massalezza Syncline. Strata constituting the west half of the failure scar, and representing the west limb of the syncline, dip N to NE, whereas those on the east side of the scar dip N to NW. The two limbs of the syncline are each approximately 1 km wide. Massalezza Stream bisects the two halves of the failure scar and thus follows the axis of the syncline. The stream and its tributaries are visible on the aspect maps as intersections between slopes facing opposite directions. Massalezza Stream, measured from its source to the toe of the deposit where the gully abruptly ends, is over 2 km long. It trends north in its upper reaches, but curves to the northeast at its mouth. Remarkably, the displaced gully is evident in the slide deposit.

I further analysed the slope and aspect maps using block statistics (Figure 3.14). The scale of roughness depends largely on the block size chosen, thus the roughness observed in this analysis is on the order of decametres (100 m$^2$ block sizes). Lineaments such as gullies, headscarps, faults, and steps produce clear curvilinear trends that are apparent on both sets of maps (see Figure 3.14 for an example of a slope map). Blocks surrounding steps in the failure scar, for example, have high to very high ranges and standard deviations, because the slope angle and aspect change over a short distance.
close to steps. Thus, stepped areas of the scar emerge as rough, linear, clearly defined areas. In contrast, folds in the failure scar are less distinct. The ranges and standard deviations are not as high in folded areas as in stepped areas, but they are still significantly higher than in smooth areas. Trends in areas with high ranges and standard deviations indicate anisotropy in roughness. For example, the east-central fold area is a rough area that trends approximately E–W. It follows the trend of the dome identified in Figure 3.6. Other rough areas, however, do not correspond to domes or basins; rather they cross dome or basin boundaries. Both the aspect and slope statistics maps show that the east half of the scar is generally much rougher than the west half. This observation supports the hypothesis that the west half moved before the east half. The east half would have required higher driving forces to overcome the resistance of the rough bedrock. Failure of the west block likely also provided the kinematic freedom required for the east block to move.

**Figure 3.14.** Examples of maps of block statistics from a) the west corner of the slide scar (Model 3) and b) the central area (Model 4), with 10 × 10 m blocks. i) Slope maps for each model (see Figure 3.13a for legend). Areas of vegetation and low point cloud density are white. ii) Range and standard deviation block statistics maps, respectively, for each model. White lines delineate the steps in the west corner, which appear as linear high-range areas. The block statistics values in the legends are in degrees (°).
3.4.4. Post-failure Scar Evolution

The morphology of the present-day failure scar cannot be attributed solely to the steps and folds described and discussed above. The scar has evolved from its form immediately after the 1963 landslide to its current state through erosion, weathering, and mass wasting (Figure 3.15). I documented significant changes to the scar by comparing photographs taken between 1964 and 2011. Between 1964 and 1985, large blocks of rock slid down the scar at several locations, and debris that covered the surface immediately after the catastrophic failure moved downslope. The blocks are defined by steps parallel to DS2, DS3, and DS4, and are controlled by the same structures that influenced the 1963 failure. Most of the scar surface is covered by debris in the 1964 photograph, and the large-scale features apparent in later photographs are not visible. Since 1985, the surface has stabilised and has lost less material. Nevertheless, several discontinuities have acted as release surfaces for small blocks. Smaller debris flows, evident in the 2002 photograph (Figure 3.15), have helped to stabilise the surface. However, observations during my two years of fieldwork suggest that the east-central area beneath the large step (labelled ‘*’ in Figure 3.15) is still prone to failure; it has sporadically released large rock masses since 1963. The differences in the height of debris between the 2002 and 2011 photographs also indicate continuing instability. Residuals between the photogrammetric data points and the LiDAR data support this observation. This area is prone to losses both slowly over decades and rapidly in catastrophic events.

Most changes to the failure scar occurred in the years immediately after the 1963 catastrophic failure. At that time, the slope was adjusting to rapid unloading due to the landslide and to erosional processes acting on the newly exposed surface. Over time, less volume was lost as the slope became more stable and approached a new equilibrium. The path to equilibrium was, however, interrupted by sporadic rockfall events such as those between 2002 and 2011.
Figure 3.15. Evolution of the failure surface from 1963 to 2011 as indicated in photographs a) to e). Asterisks mark the same location on each photograph, namely the central-east fold and step complex, which is referred to in the text. The coloured polygons represent changes in the surface between successive photographs.
3.4.5. Proposed Morphologic Classification of the Vajont Failure Scar

Each step in the analysis of the failure scar contributed new information to the characterisation of the Vajont failure. The photogrammetric models and aspect maps highlight linear features, whereas the $i^*$ angles, residual maps, profiles, and slope maps reveal different aspects of folding. In combination, the different outputs allowed us to develop a preliminary, semi-quantitative, morphologic classification of the Vajont Slide (Figure 3.16; Table 3.4). The classification is based on the two major decametre-scale elements of the failure scar – steps and folds – and their respective properties. These two features have distinct signatures and different implications for failure dynamics and kinematics, and are easily distinguished on the failure scar. Steps, as well as faults, are obvious on the photogrammetry models, have a clear $i^*$ angle plot and profile signatures (they are outliers in the former and steps in the latter), and appear as distinct, linear zones with high to very high block ranges and standard deviations in ArcGIS. The folded areas are more obscure on both the residual and block statistics maps. Slope and aspect block statistics and the dispersion coefficient for the $i^*$ angle plots are especially effective parameters for distinguishing smooth and rough areas. They correlate well with each other, and ranges of each parameter assigned to four classes – smooth and planar (class 1) to very rough and crenulated (class 4) – were chosen based on values calculated across the failure scar. The profile $A/l$ ratios are poorer indicators of surface morphology, but they can be roughly associated with the four classes. The profiles in Figure 3.11 were taken in areas categorised as class 2 and class 3 based on block statistics and $i^*$ angle $K$ values; an $A/l$ ratio of 1/12 corresponds to class 2 and one of 1/5 corresponds to class 3. These classes pertain dominantly to decametre-scale roughness. Roughness on the scale of hundreds of metres, represented by the domes and basins on the residual maps, does not directly overlap with the smaller scale roughness. The small-scale roughness is most likely parasitic to the large-scale roughness and is not concentrated in either domes or basins. The exception is the roughness in the central trough area where the hinges of the Massalezza and Erto synclines intersect. This zone is the roughest large area on the failure scar and suggests stress concentration where the two generations of folding intersect. Other rough areas are the Col Tramontin Fault and the east-central fold complex. Most of the west side of
the failure scar is smooth and planar, whereas the east half ranges from smooth and undulating to rough and folded.

**Figure 3.16.** Semi-quantitative classification of the Vajont failure scar. a) Focal length lens $f = 20$ mm model of the failure scar. b) Classification applied to the failure scar, which is outlined in black. The classification is based on the $f = 20$, 200, and 400 mm models and their respective outputs from Polyworks and ArcGIS (see Table 3.4 for details).
Table 3.4. Standard deviations (SDs) and ranges of quantitative parameters used in the morphology classification of the Vajont failure scar. \( A/l \) is the maximum amplitude/length ratio of the profiles. \( K \) is the dispersion coefficient used to quantify the scatter in \( i^* \) angles as discussed in Section 3.4.3.

<table>
<thead>
<tr>
<th>Class</th>
<th>Description</th>
<th>Profile A:l</th>
<th>( i^* ) angle K</th>
<th>Slope (°) SD</th>
<th>Slope (°) Range</th>
<th>Aspect (°) SD</th>
<th>Aspect (°) Range</th>
</tr>
</thead>
<tbody>
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<td>very smooth, planar</td>
<td>&lt;1:12</td>
<td>&gt;270</td>
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<td>0-36</td>
<td>0-45</td>
<td>336-84</td>
</tr>
<tr>
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<td>smooth, undulating</td>
<td>1:12</td>
<td>145-270</td>
<td>7-14</td>
<td>36-54</td>
<td>45-90</td>
<td>84-168</td>
</tr>
<tr>
<td>3</td>
<td>rough, folded</td>
<td>1:5</td>
<td>80-145</td>
<td>14-21</td>
<td>54-72</td>
<td>90-135</td>
<td>168-252</td>
</tr>
<tr>
<td>4</td>
<td>very rough, crenulated</td>
<td>&gt;1:5</td>
<td>&lt;80</td>
<td>21-42</td>
<td>72-90</td>
<td>135-180</td>
<td>252-336</td>
</tr>
</tbody>
</table>

Another aspect of the failure scar that is apparent in my morphologic classification is the anisotropy of the rough regions. The central rough area is elongated in an E–W direction, as are the Col delle Erghene area, the fold complex area, and the class 3 area above it. In contrast, the Col Tramontin area has an N–S orientation. Each direction is perpendicular to the direction of maximum roughness in the corresponding area. The anisotropy of the roughness influences the direction of movement on larger scales and shear strength on smaller scales. The large-scale anisotropy suggests movement in the NNE direction, except at the Col Tramontin Fault.

3.4.6. Implications of the Research for the 1963 Vajont Slide

The discontinuity, fold, and roughness analyses and my morphologic classification contribute to a better understanding of the Vajont Slide. Discontinuity sets and faults acted as important rear and lateral release surfaces. They also allowed the formation of step-paths and influenced the direction of movement. Interference folds related to two deformation events added complexity, creating a non-uniform, irregular failure zone. At the slide scale, two large synclines form a bowl with a chair-shaped cross-section downslope. Due to this configuration, the east and west halves of the failure moved toward each other. Decametre-scale and smaller folds parasitic to the large synclines are concentrated in the zone of transition between the back and seat of
the chair. These parasitic folds increased the effective shear strength of the rock mass because higher driving forces were required to either override or shear through the folds.

Another consideration is the interaction between the low-strength clays and the folds. The distribution and thicknesses of clay layers have implications for slide behaviour. The thickness of the clay interbeds influences the degree to which undulations in the failure surface affected the slide. Progressive stick-slip behaviour would be facilitated if the clay beds were sufficiently thick to allow movement until opposing undulations met, but then locked movement until the undulations sheared through. If the clay beds were thicker, the undulations would not have been as important. Conversely, if the clay layers were thin, more locking would have occurred. Another consideration is that some of the clay layers at Vajont are intra-formational, which would indicate a fairly wide distribution, whereas I would expect those associated with faults to be more localised.

Roughness elements include both discontinuities and folds. My work discriminated rough and smooth areas, including erosional basins. In addition to identifying and mapping rough areas, I documented how the failure scar has evolved since 1963. I found that the same mechanisms that controlled the catastrophic failure in 1963 were responsible for smaller, more recent failures.

My classification scheme separates the failure scar into four roughness classes and is a first step in characterising the complexity of the failure scar. Steps acted as rear and lateral release surfaces for smaller blocks within the larger slide. Folds, most of which have maximum amplitudes in the direction of sliding, inhibited movement of the rock mass until it either sheared through or overrode them; both mechanisms likely operated at Vajont. The roughest areas of the failure scar are in the hinge of the refolded southern limb of the Erto Syncline and near the Col Tramontin Fault. The difference in roughness between the east and west halves of the failure scar also has implications for slide behaviour. Work by previous authors (Broili, 1967; Hendron and Patton, 1985; Superchi, 2012; Bistacchi et al., 2013) and my field observations suggest that the west half of the slide mass failed first. The roughness of the west half of the failure scar is lower than that of the east half; thus the west slide mass had lower resisting forces and kinematic restraint. The morphology of the failure scar thus explains the observed
chronology of the landslide. More rock mass damage would be required in rougher areas to overcome asperities. The Col Tramontin Fault would have also damaged material on the east. The hinge area would have been a zone of stress concentration, not only because it is the Prandtl wedge of transition between the active upper and passive lower blocks of the slide (Mencl, 1966), but also because roughness was highest there. The failed mass would have locked there before fracture coalescence and rupture of intact rock bridges created a through-going failure surface at the moment of catastrophic failure.

3.4.7. Advantages and Limitations of Methods

The methods I have used provide valuable information, but they also are subject to uncertainties and errors. Epistemic and aleatoric uncertainties such as measurement uncertainty, bias, equipment error, and natural variability and randomness are unavoidable in any project of this type (Nadim and Lacasse, 2003). Terrestrial digital photogrammetry (TDP) enables analysis of otherwise inaccessible areas and presents data from a different perspective than is provided by airborne remote methods. My photogrammetry projects were conducted at distances ranging from 100 m to over 2 km. Projects at distances of 2 km are among the most long-range TDP projects yet completed (see Sturzenegger et al., 2009, for another example). Table 3.1 shows the relative accuracy achieved for the photogrammetry projects presented in this chapter, excluding georeferencing errors. The calculated accuracies incorporate uncertainty related to the images and the calibration of the project. The accuracies are high considering the distances to the objects of interest. The $f = 400$ mm models are significantly more accurate and produce significantly higher density point clouds than the lower focal length models, but the $f = 20$ mm model provides a good overview of the failure scar. The most significant challenges at Vajont are not the distances, but the mainly smooth, flat, and highly reflective surface of the failure scar, the variable lighting due to changing weather, and the geometry of the valley with respect to camera stations and object distances. These factors affect the number of discontinuities that can be mapped on the photogrammetry models. The models allow the orientation of the failure scar to be mapped well, but cannot detect some discontinuity planes.
Georeferencing in Polyworks was successful, but also is subject to uncertainty. The 5-m LiDAR DEM limits the absolute accuracy of the projects, although the accuracy related to the positions of features with respect to each other is high (Table 3.1). Features smaller than 5 m were visible on the photogrammetry models and metre-scale elements were mapped in some cases. As mentioned previously, metre- and decametre-scale features, which affect overall slide behaviour, are the focus of this chapter. In contrast, smaller features only affect the behaviour of localised zones.

Polyworks residual maps based on the \( f = 20, 200, \) and 400 mm models and 2D profiles derived from the \( f = 200 \) and 400 mm models highlight the large-scale basins and domes on the failure scar, as well as directions of maximum roughness. They are affected, however, by vegetation and noise in the data. Errors related to vegetation are clear and appear as spikes in an otherwise uniform surface. These anomalies can be reduced or removed by smoothing the digital terrain models in 3DM Analyst before exporting to Polyworks, or by deleting the points associated with vegetation. An example of the latter is the hole in model 2. Although noise exists in the original photogrammetry models, it is more apparent in the 2D profiles (Figure 3.11). Consideration of noise is important because it may complicate interpretation of roughness. Ideally, a minimum threshold should be used to exclude noise during the analysis. For the current work, the interpreted scales of roughness are orders of magnitude larger than the noise and the accuracy of the photogrammetry models. Poropat (2008) and Lyman et al. (2008) address noise and uncertainty in remote sensing and rock mass characterisation.

Parameters derived from each of the above methods are based on well-known variables in the rock mechanics and geomorphometry fields. The \( i^* \) angle and block statistics values were especially useful in classifying roughness and correlate well with and complement each other while emphasising different aspects of roughness. For example, slope statistics highlight downslope roughness, whereas aspect statistics highlight across-slope roughness.
3.5. Conclusions

The failure scar of the Vajont Slide is complexly folded, jointed, and stepped. It is bounded by step-paths formed from intersecting joints and faults, and is marked by steps and intersections of folds of two episodes of Alpine deformation. The surface has been modified by erosion and weathering in the 50 years since the landslide.

I present the first terrestrial digital photogrammetric models using ADAM Technology, Polyworks, and ArcGIS, and summarise endogenic and exogenic controls in a semi-quantitative, three-dimensional morphologic classification involving four roughness classes. Traditional measures of roughness proved to be inappropriate for describing features on scales of tens of metres to a kilometre or more. My classification attempts to bridge the gap between quantitative measures of small-scale features and qualitative descriptions of large-scale features. Implications for the mechanisms of the Vajont Slide include: 1) smooth areas of the failure surface had relatively low effective shear strength; 2) rough areas of the failure surface had relatively high effective shear strength, which inhibited movement; 3) discontinuities acted as rear and lateral release surfaces; 4) the sliding mass either sheared through asperities or dilated over them; 5) the clay interbeds in the sliding zone played an important role in the failure and influenced how the mass failed; and 6) the failure was structurally controlled — folds, faults, and discontinuity sets affected the failure scar and contributed to rock mass damage.

To reduce some of the limitations mentioned in the previous section, the current work is intended to be compared with results using different techniques. The University of Padova is presently investigating mesoscale folds using a UAV (unmanned aerial vehicle). Another future possibility is to use a state-of-the-art long-range laser scanner capable of mapping from distances up to 4 km.

I am currently producing preliminary, three-dimensional, discontinuum numerical models of the Vajont Slide. The effects of the four roughness classes on the mechanisms, kinematics, and dynamics of the slide will be explicitly explored within these models. I am also preparing a detailed engineering geomorphology map of the
slide, which will be combined with the photogrammetric analyses as input to the numerical models.

**Acknowledgements**

The Environment and Land division of the Friuli–Venezia Giulia region and L. Superchi of the University of Padova provided LiDAR images. I thank R. Genevois, M. Ghirotti, M. Massironi, F. Patton, L. Superchi, and L. Zorzi for sharing their knowledge of the Vajont Slide; S. Dragicevich and J. Song for discussion and assistance for the operation of ArcGIS; D. Donati and B. Ward for assistance in the field; and the two reviewers for their invaluable suggestions. This project was funded by Natural Sciences and Engineering Research Council of Canada (NSERC) Discovery Grants to D. Stead and J.J. Clague, and an NSERC scholarship to A. Wolter.
Chapter 4.

A Numerical Modelling Toolbox Approach to Investigate the 1963 Vajont Slide, Italy

Abstract

In the first study to combine detailed field investigations and numerical modelling of the 1963 Vajont Slide, I use a numerical toolbox approach to investigate its causes, kinematics, and mechanisms. Based on field evidence, I hypothesise that the catastrophic landslide was the last stage of a long-term sackung-type failure; I focus on the initiation and development of the slow, progressive phases of movement (pre-reservoir conditions). I apply five continuum and discontinuum, two- and three-dimensional numerical codes to investigate five aspects of the slide. The continuum models show the significance of in situ stress distributions and their interactions with topography in conditioning the Vajont slope for failure. Material properties such as friction angle along the sliding surface influence the stability of the slide and affect its behaviour in both continuum and discontinuum models. Discontinuum models highlight the role of major structures such as folds and faults, as well as block size, in controlling the kinematics of the Vajont Slide, and suggest that the number of landslide blocks required to approximate the actual kinematics of the slide is between 3 and 12. Finally, continuum and lattice-spring models show the importance of internal deformation of the landslide mass to allow failure and the formation of a Prandtl wedge transition zone between the active upper and passive lower slide blocks. The hydrogeological conditions at Vajont were critical to the triggering of the slide; I present preliminary hydromechanical models using a new spring-lattice code.

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4 To be submitted as:
4.1. Introduction

Prior to 2013, there were few analytical models of the 1963 Vajont Slide. The first physical model of the slide was created in 1961 at a scale of 1:200 to examine the effects of a large landslide and corresponding displacement wave on the Vajont Dam and surrounding area (Semenza, 2001). It predicted the size of the failure and its behaviour well, but the researchers conducting the modelling did not account for the high terminal velocity it achieved. Modelling after the 1963 failure showed that had the velocity of the replicated mass been increased, the model would have successfully predicted the wave height and its disastrous effects (Ghirotti et al., 2013). Subsequent modelling of the displacement wave has been done by Bosa and Petti (2011), Ward and Day (2011), and Vacondio et al. (2013). These researchers simulated the timing and spatial extent of the wave using different two- and three-dimensional numerical programs.

Numerous authors analysed the failure in two dimensions using limit equilibrium approaches (Hendron and Patton, 1985 and references therein; Lo et al., 1971). Hendron and Patton (1985) conducted the first three-dimensional limit equilibrium analyses. Ghirotti (1992, 1994) and Sitar and MacLaughlin (1997) presented two-dimensional numerical simulations using, respectively, distinct element and discontinuous deformation analysis techniques. Ghirotti (1992, 1994) analysed the mechanics and hydrogeological conditions of the failure in UDEC and illustrated the importance of hydromechanical interaction to instability. Sitar and MacLaughlin (1997), in contrast, investigated the role of block size and determined a strong correlation between friction angle and joint spacing.

Several more researchers simulated the Vajont Slide for the 50th anniversary conference held in Padova, Italy. Crosta et al. (2013) created two- and three-dimensional finite element models to simulate the rockslide and its interaction with the reservoir water. Hungr and Aaron (2013) used CLARA-W to investigate the two- and three-dimensional
limit equilibrium of the Vajont slope, and agreed with previous suggestions regarding the existence of two main blocks in the movement (Superchi, 2012; Bistacchi et al., 2013). Paparo et al. (2013) employed the minimum lithostatic deviation method, a modification of the classical limit equilibrium approach, to two sections of the Vajont Slide and highlighted the importance of friction angle and pore water pressure to slope stability. Finally, Zaniboni et al. (2013) applied the three-dimensional Lagrangian code UBO-BLOCK1, calculated velocities and basal friction coefficients of six longitudinal blocks, and demonstrated the significance of the clay beds at the failure surface to the movement of the slide.

Despite the recent advances in modelling of the Vajont Slide, most approaches remain two-dimensional and lack field constraints. Given its lithological, structural, and morphological complexity, such as the two faults bounding the east lateral and west rear scarps and the bowl-shaped failure surface, the Vajont Slide is a three-dimensional problem. Sophisticated analyses demand large data sets. Although the Vajont Slide has been studied for half a century, few authors have attempted to quantify rock mass properties, and those who have report a range of values.

This chapter presents preliminary results from five, two- and three-dimensional finite element, kinematic, distinct element, and lattice-spring codes. Each code is used to investigate a specific aspect of the Vajont Slide; I do not attempt to exactly replicate the failure, a task that is arguably not possible given the lack of site investigation data at the time of the failure. Rather, each simulation is used as a tool to test hypotheses and research questions, and thus contributes to a combined numerical modelling toolbox approach. I focus on the pre-reservoir conditions of the Vajont slope, and thus neglect groundwater effects in most simulations. The objectives of the study are to:

1. explore the effect of landscape evolution on in situ stress distributions and the conditioning of the Vajont slope for failure,
2. examine the kinematics and movement behaviour of the Vajont Slide, and
3. investigate the role of internal deformation within the rockslide.
4.2. Background

The Vajont Slide is situated in an active tectonic and geomorphological landscape (Figure 4.1). The material involved in the failure comprises Jurassic-Cretaceous limestones, micrites, marls and conglomerates of the Fonzaso and Calcare di Soccher formations (Figure 4.2). The Soccher Formation is subdivided into six units, labelled A to F, which can be grouped into three geotechnical units (Ghirotti, 1992, 1994): i) unit A and the Fonzaso Formation, ii) unit B, and iii) units C-F of the Soccher Formation. Unit B, a conglomerate, is distinguished from the other units by its higher strength.

Two faults and two sets of folds have directly affected the slope (Figure 4.3). The Col delle Erghene Fault bounds the headscarp and is oriented at 55°/010° (dip/dip direction). The Col Tramontin Fault has an orientation of 59°/283°; it is the east lateral boundary of the slide and is a splay of the Croda Bianca Reverse Fault. Interference between two deformation events, the Dinaric and Neoalpine phases, has created complex folds on the failure surface (Massironi et al., 2013). The Erto Syncline, the refolded southern limb of which gives the failure surface its characteristic chair shape in profile, is likely the result of the Neoalpine deformation event. The hinge of the Erto Syncline plunges 20° to the east (Hendron and Patton, 1985). Above the Vajont Slide, the syncline coincides with a limb of the regional Belluno Anticline, creating a relatively flat area at the top of Monte Toc. The Massalezza Syncline is oriented N-S, is associated with Dinaric compression, and gives the sliding surface its bowl shape across the slope. The failure surface is thus complex and three-dimensional.
Figure 4.1. Setting of the 1963 Vajont Slide. a) Block diagram showing controlling features of the slide, including CT = Col Tramontin Fault, CE = Col delle Erghene Fault, ES = Erto Syncline, MS = Massalezza Syncline, and discontinuity sets (DS) determined from field and photogrammetry surveys. Inset shows the location of the landslide in northeastern Italy, approximately 100 km north of Venice. b) - e) Cross-sections used for different two-dimensional analyses in Phase2 and UDEC. Sections b) and c) are based on Rossi and Semenza’s profiles (Hendron and Patton, 1985); sections d) and e) were provided by Bistacchi et al. (2013). b) also shows the stratigraphy of the slide, grouped into geotechnical units based on Ghirotti (1992, 1994), and the hydrogeological conditions hypothesised by Hendron and Patton (1985). GWT = estimated groundwater table.
Other structures that locally control the slide include steps in the failure scar, secondary shear surfaces, and the Massalezza Gully. These discontinuities create several intact blocks within the landslide body. Steps in the failure scar are oblique to the movement direction (N to NNE) and may have acted as lateral release surfaces for local movements. One large step trending SW-NE and located in the east-central area of the failure scar seems to align with the boundary between two main blocks of the landslide, the east and west blocks (Figure 4.3). The geometry of these two blocks is illustrated by Superchi (2012) and Bistacchi et al. (2013) (Figure 4.4), all of whom propose that the west block failed before the east block. Further subdivision of the west block into five sub-blocks is based on secondary shear surfaces, extensional areas, and benches mapped on pre-1963 air photographs (Chapter 2). The Massalezza Gully, although appearing continuous and apparently undisturbed, separates two areas of different morphology and is thus hypothesised to be a sub-block boundary. There are also two smaller blocks upslope of the east and west blocks.

![Figure 4.2](image-url)

**Figure 4.2.** Stratigraphy in the Vajont Slide area, looking from the headscarp across Vajont Valley. CV = Calcare di Vajont (Vajont Limestone). The Fonzaso and Soccher A – F formations were involved in the 1963 failure.
Figure 4.3. Structures controlling the 1963 Vajont Slide. Two faults (CT = Col Tramontin Fault, CE = Col delle Erghene Fault) act as lateral and rear release surfaces. The Massalezza Syncline (MS) and Erto Syncline (ES) contribute to the bowl-shaped failure surface. Also shown are two block boundaries, the E-W block boundary related to a step in the failure scar and the Massalezza Gully following the hinge of the Massalezza Syncline, are shown. See text for details.

Figure 4.4. Geometry of the east and west blocks after the 1963 Vajont Slide. From Bistacchi et al. (2013). Left figure shows the post-1963 DEM and the Rossi and Semenza map (Hendron and Patton, 1985). Right figure shows the failure scar and two main blocks, as well as contour lines (grey).
Hydrogeological conditions were crucial in triggering the catastrophic failure at Vajont. The average annual rainfall in the Vajont area ranges from 1200 to 2300 mm/year, indicating that groundwater pressures and levels could fluctuate significantly over time. In the three years leading up to the catastrophic failure, significant slope movements (> 1 cm/day) in November 1960, November 1962, and October 1963 were preceded by periods of heavy rain, indicating the effect of precipitation on the slope’s behaviour (Figure 4.5). Three piezometers installed prior to 1963 (P1, P2, and P3) demonstrate the influence of the reservoir. P1 and P3 recorded the average pore pressure conditions in the fractured rock mass above the clay-rich sliding zone and mimicked the reservoir level to within a few metres. The initial readings of P2 were 90 m above the reservoir level, but values decreased to the same levels as P1 and P3 in 1962. This discrepancy may be explained if natural sealing of the piezometer occurred due to squeezing of soft clays around the pipe. P2 seems to have recorded pore pressures at the sliding surface for a short period of two months (Hendron and Patton, 1985). Deformation rates in the slope decreased to almost zero in early 1963 with the drawdown of the reservoir to 650 m asl, but increased as the reservoir level was raised again to 710 m asl. Finally, deformation rates increased dramatically to >3 cm/day at the onset of catastrophic failure. Taking the stratigraphy (carbonates with karst formation), reservoir, precipitation, and base level flow into consideration, Hendron and Patton (1985) hypothesised two groundwater tables, one high and one low, for different conditions. In my preliminary geomechanical models, I accept and incorporate these groundwater levels. The primary focus of most of the models in this dissertation, however, is on the rock mass properties and the structural controls prior to the filling of the reservoir.
4.3. A Numerical Modelling Toolbox

I used a numerical modelling toolbox approach with five different software codes to investigate five aspects of the Vajont Slide in two and three dimensions: i) landscape evolution and in situ stress distributions, ii) material properties, iii) the kinematics of the slide and how block size affects behaviour, iv) internal deformation of the sliding mass, and v) the effects of groundwater on the slide (Table 4.1). Time-dependent behaviour such as creep is outside the scope of this study. The reader is referred to Section 4.4 for details of the methodology applied in each code, and Appendix E for sample code files.
Table 4.1. Numerical modelling toolbox used to study five specific aspects of the Vajont Slide. CT = Col Tramontin Fault, CE = Col delle Erghene Fault, ES = Erto Syncline, MS = Massalezza Syncline.

<table>
<thead>
<tr>
<th>Code</th>
<th>Input</th>
<th>Aspects analysed</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Type</td>
<td>Geometry</td>
</tr>
<tr>
<td>Phase2 v. 8.0</td>
<td>2D Finite element</td>
<td>General section</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Section 2</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Section 10A</td>
</tr>
<tr>
<td></td>
<td></td>
<td>E Section</td>
</tr>
<tr>
<td></td>
<td></td>
<td>W Section</td>
</tr>
<tr>
<td>Swedge v. 6.0</td>
<td>3D Limit equilibrium and kinematic analysis</td>
<td>CT Fault</td>
</tr>
<tr>
<td></td>
<td></td>
<td>CE Fault</td>
</tr>
<tr>
<td></td>
<td></td>
<td>ES</td>
</tr>
<tr>
<td></td>
<td></td>
<td>MS</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Basal surface</td>
</tr>
<tr>
<td>UDEC v. 5.0</td>
<td>2D Distinct element</td>
<td>Section 2</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Section 10A</td>
</tr>
<tr>
<td></td>
<td>3D Distinct element</td>
<td>Bistacchi DEM</td>
</tr>
<tr>
<td></td>
<td>3DEC v. 5.0</td>
<td>Bistacchi DEM</td>
</tr>
<tr>
<td></td>
<td>Slope Model v. 2.9</td>
<td>Bistacchi DEM</td>
</tr>
</tbody>
</table>

Preliminary kinematic, limit equilibrium investigations were conducted in Swedge (Rocscience, 2014b). The continuum code Phase2 (Rocscience, 2014a) allowed investigation of material deformation alone without consideration of discontinuities. The hypothesised sliding surface was included in the simulations with no consideration of discontinuity sets. Conversely, discontinuities were the focus of the discontinuum simulations in UDEC (Itasca, 2013c) and 3DEC (Itasca, 2013a). Blocks were assumed to be rigid or elastic; thus plastic deformation was not considered in these codes. Finally, a recently developed code, Slope Model (Itasca, 2013b), based on a lattice-spring
scheme, was used to simulate brittle fracture in the slope rock mass, allowing investigation of internal deformation of blocks prior to the development of discrete shear surfaces.

Numerical modelling simulations presented in this chapter are constrained by field and remote sensing observations. Ghirotti (1992, 1994) and Superchi (2012) provided field and laboratory data on the geomechanical properties of the rock mass involved in the Vajont Slide. Chapter 3 summarises the use of terrestrial photogrammetry in characterising the Vajont Slide failure scar and determining discontinuity sets. Chapter 2 illustrates the use of an engineering geomorphology approach to constrain the number of blocks in the landslide and determine a potential chronology of events leading to the catastrophic failure. Based on field observations, I suggest that the 1963 landslide was the last stage in a sackung-type failure that initiated after deglaciation of the area.

4.4. Modelling Methodology and Results

4.4.1. 2D Continuum Analysis of Landscape Evolution, Internal Deformation, and Material Properties

Methodology and Model Organisation

I used Phase2, a two-dimensional finite element code (Rocscience, 2014a), to investigate landscape evolution, material input parameters, and internal deformation in dry models. In the landscape evolution simulations, a representative slope profile derived from a pre-1963 Digital Elevation Model (DEM) was achieved by excavating in four main stages (Figure 4.6). After each stage, the remaining surficial material was changed from Mohr-Coulomb elastic to Mohr-Coulomb plastic to allow deformation, for a total of nine stages. A glacier was introduced after the third excavation stage to simulate glaciation of the valley, and was then removed. This approach allowed excavation of material to the final slope profile, while still accounting for slope deformation and strength degradation of the surficial material due to factors such as weathering. Kalenchuk (2010) and Leith (2012) use similar methods to simulate, respectively, the formation of a shear surface and exhumation of valley slopes. I also performed a
sensitivity analysis on the in situ stress ratio (horizontal/vertical stresses, K), using a range of values from $K = 0.33$ to $K = 3.3$. The lower values represent normal faulting conditions, whereas $K$ values > 1 represent reverse faulting. Leith (2012) used the same range of $K$ values in his study of in situ stresses. The tectonic regime in the Vajont region is reverse-strike slip, suggesting high horizontal stresses and thus $K > 1$. However, local topographic effects at shallow depths and fault slip may perturb this assumption.

In separate simulations, the effects of material inputs were examined using Rossi and Semenza’s Section 2 (Hendron and Patton, 1985; Figure 4.1b). One-material models used the properties of the Vajont limestone (Calcare di Vajont) and Fonzaso Formation, and four-material models included the Calcare di Vajont, Fonzaso, Soccher C-F, and Soccher B formations as in Ghirotti (1992) (Figure 4.7; Table 4.2 and Table 4.3). I also varied the friction angle of the failure surface from 8° to 38° to observe the effects on the stability of the sliding mass. In all simulations, the internal deformation was observed using maximum shear strain and failure state plots. Rossi and Semenza’s Section 10A and Bistacchi et al.’s (2013) east and west sections were also analysed for stability using the Shear Strength Reduction (SSR) technique (Hammah et al., 2005; Diederichs et al., 2007, and references therein) to observe the effect of topography and failure surface geometry on failure (Figure 4.8).
Figure 4.6. Stages of excavation in the Phase2 simulations of landscape evolution. a) Stages 1 to 4, showing the excavation of material. b) Stages 5 and 6, representing glaciation and deglaciation (which coincides with Excavation 3). c) Stages 7 to 9, showing the degradation of each stage material to “residual” plastic properties from elastic properties and the excavation of the final topography.
**Figure 4.7.** a) One-material and b) four-material models of Section 2 (see Figure 4.1 for location) analysed in Phase2 to investigate the effects of material properties on stability. Another set of one-material models was run with the Fonzaso and Soccher A properties.

**Table 4.2.** Rock mass properties used in the Phase2 simulations, based on Ghirotti (1992).

<table>
<thead>
<tr>
<th>Property</th>
<th>Calcare di Vajont</th>
<th>Fonzaso + Soccher A</th>
<th>Soccher B</th>
<th>Soccher C - F</th>
</tr>
</thead>
<tbody>
<tr>
<td>Density (kg/m³)</td>
<td>2690</td>
<td>2700</td>
<td>2700</td>
<td>2610</td>
</tr>
<tr>
<td>Bulk modulus (GPa)</td>
<td>6.1</td>
<td>2.7</td>
<td>5.7</td>
<td>4.2</td>
</tr>
<tr>
<td>Shear modulus (GPa)</td>
<td>3.7</td>
<td>1.6</td>
<td>3.4</td>
<td>2.5</td>
</tr>
<tr>
<td>Friction angle (°)</td>
<td>n/a</td>
<td>40</td>
<td>n/a</td>
<td>40</td>
</tr>
<tr>
<td>Cohesion (MPa)</td>
<td>n/a</td>
<td>0.1</td>
<td>n/a</td>
<td>0.1</td>
</tr>
</tbody>
</table>
Table 4.3. Discontinuity properties used in the Phase2 simulations.

<table>
<thead>
<tr>
<th>Property</th>
<th>Sliding surface</th>
</tr>
</thead>
<tbody>
<tr>
<td>Tensile strength (MPa)</td>
<td>0</td>
</tr>
<tr>
<td>Normal stiffness (GPa/m)</td>
<td>5</td>
</tr>
<tr>
<td>Shear stiffness (GPa/m)</td>
<td>0.6</td>
</tr>
<tr>
<td>Friction angle (°)</td>
<td>12</td>
</tr>
<tr>
<td>Cohesion (MPa)</td>
<td>0</td>
</tr>
</tbody>
</table>

Results of Phase2 Simulations for an Assumed Dry Slope

Sensitivity analyses on the in situ stress ratio, K, indicate that damage increases as the K ratio increases. Thus, the higher the horizontal stress, the more tensile and shear failures occur in all plastic stages. Figure 4.9 illustrates the linear relationship between K and damage, with yielded elements in each plastic stage combined for each model. Only 209 elements yielded in the K = 0.33 model, mostly in tension at sharp topographic ridges, changes in slope gradient, and on the valley bottom. Most of the yield in the K = 0.33 model occurred after deglaciation in stage 7. For the K = 0.5 and K = 1 models, most elements (46% and 33% of all yielded elements for all stages for each K value, respectively) yielded in stage 4. Most elements in the K = 2 and K = 3.3 models yielded in stage 9 (40% and 37%, respectively). Clearly, the choice of K ratio affects slope instability. For the Vajont area, in a compressional-strike slip regime, K should be greater than 1, and so slope damage is expected to be fairly high. Most of the damage in the simulations occurs on the upper slopes, in areas of irregular topography, such as the area at the crest of the south slope of the valley containing dolines, and, in later stages, in the Vajont Gorge. Shear yield occurs exclusively in the gorge, whereas extension characterises the upper slopes and ridge crests.
Figure 4.8. Four different sections analysed in Phase2 to investigate the effects of topography and sliding surface geometry. Section locations are indicated in Figure 4.1. The two materials included in these models are the Vajont limestone (labelled limestone) and slide material (labelled debris) with properties equivalent to the Soccher C-F units.

Four of the nine stages in my Phase2 landscape evolution simulations were allowed to deform plastically. The deformation and yield in each of these stages indicated damage due to each successive excavation. Assuming a horizontal/vertical stress ratio of $K = 1$, the first excavation did not cause much damage, with only 115 elements, or 12% of the total number of yielded elements, failing in tension. The number of yielded elements increased markedly to 325 after the second excavation, again all failing in tension. The glaciation stage (stage 5) showed stress concentrations related to the glacier at points of high curvature in the underlying topography (Figure 4.10). After deglaciation, damage locations shifted from ridge crests to the lower slopes, indicating the effect of the simulated glacier. Of the 287 yielded elements in the post-glaciation excavation, 30 failed in shear in the area of the future gorge. After the final excavation, the Vajont Gorge influenced stress concentrations, shear strain distributions, and tensile and shear failure on the Vajont Valley slopes (Figure 4.11). Tensile failure was also
simulated farther upslope on both sides of the valley, in agreement with Leith’s (2012) suggestion that ridge crests fail in tension whereas valley bottoms are usually subject to compression and shear, depending on in situ stress conditions. Many elements yielded in tension on the north slope and north ridge crest of Vajont Valley. These extensional areas align roughly with the Monte Salta rockslide and potential extensional lineations observed in satellite imagery (Figure 4.11b). Some tensional failure was simulated near areas where the 1963 failure surface daylighted and also in small depressions in the topography at the crest of the south valley slope, correlating well with observed dolines and structures.

Figure 4.9. The relationship between the number of yielded elements and the change in K ratio from results of sensitivity analyses in Phase2.

The influence of simulated internal deformation in Phase2 on slope stability appears to be significant. A transition zone concentrating maximum shear strain and damage between the active and passive blocks was clearly simulated in the Phase2 models, in agreement with Mencl’s (1966) hypothesis of a Prandtl wedge. Shear zones were also simulated at breaks in slope (Figure 4.12). Each of the simulated shear zones has an angle of between 45° and 90° to the horizontal. The simulated shear zone located at the Prandtl wedge is 80 m thick and inclined at 70° from horizontal. These shear zone orientations are consistent with those predicted using the Mohr-Coulomb shear criterion where \( \theta = 45° \pm \varphi/2 \), and \( \theta \) is the expected angle of the failure surface.
and $\phi$ is the friction angle along that plane, and with the upper part of the failure surface dipping 20° to 40°. As weaker materials were included, another incipient, semicircular shear zone developed at the toe of the failure, oriented at 45° to the horizontal and located where the 1960 failure occurred (Figure 4.12b and c). Hence, internal deformation in a 2D continuum model allowed failure to occur without the explicit inclusion of more than one block, and indicates the development of secondary shear surfaces.

![Figure 4.10. Major principal stress, $\sigma_1$, plot for stage 5, showing the simulation of glaciation and stress concentrations on points in the underlying topography (white box). K = 1.](image)

The lithologies involved in the Vajont Slide influenced its behaviour. Ghirotti (1992, 1994) suggested that the conglomerate layer (Soccher B unit) is stronger than other units in the formation and remained relatively intact as it moved downslope. When the different units involved in the failure were incorporated into the Phase2 models, maximum shear strain concentrated in the weaker materials surrounding Unit B (Figure 4.12). However, the effect of different material properties on simulated instability was not significant; it appears that in Phase2 models the sliding surface properties dominated failure behaviour. For example, there is a direct correlation between sliding surface friction angle and critical SRF (Figure 4.13). The critical friction angle, where SRF=1, was determined to be 16°, given a dry, cohesionless failure surface.
Figure 4.11. a) Maximum shear strain and yielded elements at stage 9 of the Phase2 landscape evolution model, with $K = 1$. Inset shows maximum shear strain and principal stress tensors in Vajont Gorge. b) Satellite imagery (courtesy of Google Earth) indicating locations of the profile analysed (dashed line), extensional yield clusters on the north slope of Vajont Valley (black boxes), and possible tensional lineations found at the slope crest (red curves).
Figure 4.12. Phase2 one-material and four-material models of Section 2 showing zones of maximum shear strain within the sliding mass at the angles indicated. a) One-material Calcare di Vajont model, b) one-material Fonzaso + Soccher A model, and c) four-material model incorporating Calcare di Vajont, Fonzaso + Soccher A, Soccher B, and Soccher C-F.
Figure 4.13. Linear relationship between the friction angle ($\phi$) of the failure surface and Strength Reduction Factor (SRF) in Phase2 as the friction angle is increased. The critical friction angle, where SRF = 1, is 16°.

The four different cross-sections analysed in Phase2 – Sections 2 and 10A of Rossi and Semenza (Hendron and Patton, 1985) and the east and west sections of Bistacchi et al. (2013) – show clearly different results and indicate the complex, three-dimensional character of the Vajont Slide. Section 2 and the west section have biplanar failure surfaces with active and passive blocks. Section 10A has a circular failure surface and indicates deformation only at the toe of the slide. It has a slightly higher critical SRF than Section 2 (SRF=1.37 versus 1.28). Both of the Rossi and Semenza sections indicate stable conditions when groundwater is not considered. Bistacchi et al.’s (2013) west section has an SRF of 0.78. The east section is the least stable profile, with a linear to slightly curved failure surface and an SRF of 0.45. If only this section had been analysed, the stability of the slope would have been underestimated, especially as the east block failed after the west block. This discrepancy highlights the importance of undertaking three-dimensional analyses for complex slope failures.
4.4.2. Kinematic Analysis in Swedge

Methodology and Model Organisation

I analysed the kinematics of the east and west halves of the Vajont Slide in Swedge v. 6.0 (Rocscience, 2014b) using the major structures involved in the failure. In this version of the Swedge program, hexahedral wedges are formed by two intersecting joints, a tension crack, and a basal surface. I included the east and west limbs of the Massalezza Syncline, the Col Tramontin Fault, the Col delle Erghene Fault, the southern limb of the Erto Syncline, and a basal release plane in the analysis. Their orientations are based on field and photogrammetric observations (Chapter 3). I also included a possible block boundary, assumed to be oriented 90°/090° (dip/direction), representing the Massalezza Gully (Table 4.4). This boundary separated the Vajont Slide into east and west halves, and allowed a more complete analysis of structural controls on the landslide. I assumed slope height and length for each half to be 500 m and 1000 m, respectively, based on actual slope geometry. In a simple pore pressure analysis, I assumed a peak pressure at the toe of the slope (representing Vajont Reservoir), and a water table at 200 m above the slope toe, which is the approximate maximum elevation of the reservoir.

Table 4.4. Orientations of the major planes involved in the Vajont Slide for Swedge analysis.

<table>
<thead>
<tr>
<th>Major structure</th>
<th>East half</th>
<th>West half</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Dip (°)</td>
<td>Dip direction (°)</td>
</tr>
<tr>
<td>Slope face</td>
<td>35</td>
<td>007</td>
</tr>
<tr>
<td>Upper slope</td>
<td>34</td>
<td>007</td>
</tr>
<tr>
<td>East Massalezza Syncline limb (MSE)</td>
<td>40</td>
<td>010</td>
</tr>
<tr>
<td>West Massalezza Syncline limb (MSW)</td>
<td>n/a</td>
<td></td>
</tr>
<tr>
<td>Col Tramontin Fault (CT)</td>
<td>59</td>
<td>283</td>
</tr>
<tr>
<td>Col delle Erghene Fault (CE)</td>
<td>n/a</td>
<td></td>
</tr>
<tr>
<td>Massalezza Gully (MG)</td>
<td>90</td>
<td>090</td>
</tr>
<tr>
<td>Basal release surface (BP)</td>
<td>5</td>
<td>000</td>
</tr>
</tbody>
</table>
Figure 4.14 shows the hexahedral wedges resulting from the orientations of structures listed in Table 4.4. The basal surface friction angle and orientation determine, respectively, wedge stability and movement direction. Swedge predicts a dry critical friction angle (FS = 1) of 5° and movement direction toward the north when the basal surface is oriented 5°/000° (dip/dip direction) for both failures. When the east-plunging hinge of the Erto Syncline is included as the basal plane in the west, the dry mass is stable.

All orientations used in Swedge agree with field and photogrammetry observations, with two exceptions: i) the east limb of the Massalezza Syncline, and ii) the Erto Syncline limb. Photogrammetry measurements for the former plane average 36°/353° (dip/dip direction), but the most reasonable wedge created by Swedge includes an orientation of 40°/010° for the syncline limb. Given the roughness of the failure scar in this area, this difference is insignificant. The orientation of the Erto Syncline is commonly overlooked in stability analyses. Hendron and Patton (1985) highlighted its importance in their three-dimensional stability analysis, and approximated its orientation as 20°/090°. I find the optimal value for failure (FS = 1) to be 12°/090° in my preliminary Swedge analyses, which is in general agreement with Hendron and Patton’s (1985) estimate. Assuming a dip direction of 090°, stability increases (FS increases) as the dip of the plane increases. The maximum dip possible to form a wedge is 50°. Assuming a dip of 12°, stability increases as the dip direction of the plane increases.

Both the east and west halves of the failure require a lateral release plane to form hexahedral wedges. The plane representing the Massalezza Gully (and Massalezza Syncline hinge) served this purpose. In both analyses, the stability of the wedges decreased when the Massalezza Gully plane rotated to dip more northward (out of the slope).

The assumed persistence of the structures defining the wedges must be > 500 m in this Swedge analysis. In practice I would suggest that such high persistence structures are composite surfaces, or surfaces that have developed by the interconnectivity of less persistent discontinuities through brittle fracture of intact rock.
bridges. Structural variations such as the complex fold interference observed on the failure scar also contributed to the development of the composite surfaces.

The preliminary simple pore pressure analyses in Swedge indicated the important effect of water on slope stability. To stabilise the slope, the friction angle of the north-dipping basal plane in both failures was increased to 27°, significantly higher than the dry friction angle of 5°. In the west, where the basal plane dipped to the east (representing the Erto Syncline), a wet critical friction angle of 13° was required for failure.

![Swedge hexahedral wedges for a) the east half, and b) the west half of the Vajont Slide, assuming the structures listed in Table 4.4. The white arrows are predicted movement directions. MSE = east limb of Massalezza Syncline, MSW = west limb of Massalezza Syncline, CT = Col Tramontin Fault, CE = Col delle Erghene Fault, ES = Erto Syncline, BP = basal plane.](image)

**Figure 4.14.** Swedge hexahedral wedges for a) the east half, and b) the west half of the Vajont Slide, assuming the structures listed in Table 4.4. The white arrows are predicted movement directions. MSE = east limb of Massalezza Syncline, MSW = west limb of Massalezza Syncline, CT = Col Tramontin Fault, CE = Col delle Erghene Fault, ES = Erto Syncline, BP = basal plane.

### 4.4.3. Kinematics and Block Size in 2D Discontinuum Analyses

**Methodology and Model Organisation**

I used discontinuum codes to investigate the role of discontinuities in the kinematics of the Vajont failure, assuming dry conditions. The sliding surface geometry was varied along Rossi and Semenza’s Section 2 (Figure 4.1b) from a linear surface to circular, biplanar, and undulating surfaces in UDEC (Universal Distinct Element Code; Itasca, 2013c). Rossi and Semenza’s (Hendron and Patton, 1985) proposed failure surface was also analysed (Figure 4.15a and b). The critical friction angle for each
surface was determined to quantify the effect of sliding surface geometry on stability. Following a methodology similar to Sitar and MacLaughlin (1997), I varied the spacing of subvertical joints from 50 m to 500 m to examine the effect of block size on kinematics (Figure 4.15c). Finally, a new Trigon tessellation method (Gao, 2013) was applied to the section both to further reduce block size and to simulate brittle fracture in the rock mass. This method is an extension of the Voronoi tessellation method using triangular blocks rather than polygonal blocks and allows the development of new fractures in a UDEC model. Important bedding planes and units such as the Soccher Unit B were included in the model (Figure 4.15d; see Table 4.2 and Table 4.3 in Section 4.4.1 for material and discontinuity properties). For simplicity, I assumed that the subvertical joints in UDEC simulations have the same properties as the sliding surface in Phase2, except that the friction angle was increased to 40°.

**Figure 4.15.** Geometries analysed in UDEC simulations. a) Rossi and Semenza’s proposed failure surface, as well as linear, circular, and biplanar surfaces. b) The three undulating surfaces analysed. c) Subvertical joints with a spacing of 50 m included to investigate block size. d) Example of a Trigon model with four materials.
Results of UDEC Simulations for an Assumed Dry Slope

UDEC simulations showed the importance of failure surface configuration and block size on the Vajont failure mechanism. The morphology of the failure surface proved to be significant; results of sensitivity analyses to determine the critical friction angle for the different failure surface configurations are presented in Table 4.5. The linear and circular failure surfaces appear to be the least stable in dry simulations, with the highest critical friction angles; in contrast, the undulating surfaces are the most stable, as they do not fail at all when dry. The biplanar and Rossi and Semenza geometries affect stability in a similar way – they both require friction angles of 5° to fail, given the input parameters.

Table 4.5. Critical friction angles for the different sliding surface configurations in UDEC, assuming dry conditions.

<table>
<thead>
<tr>
<th>Model</th>
<th>Critical friction angle (°)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Linear</td>
<td>12</td>
</tr>
<tr>
<td>Circular</td>
<td>13</td>
</tr>
<tr>
<td>Biplanar</td>
<td>5</td>
</tr>
<tr>
<td>Undulating</td>
<td>stable</td>
</tr>
<tr>
<td>Rossi and Semenza</td>
<td>5</td>
</tr>
</tbody>
</table>

Adding subvertical joints to the failure mass further decreases stability, increasing the critical friction angle to 16° (one material) for the Rossi and Semenza failure surface. Varying the spacing of the subvertical joints from 50 m to 500 m altered the mode of failure. As joint spacing increased, uniform displacement vectors throughout the 50 m model changed to higher magnitude vectors in the upper blocks and lower magnitude vectors in the lower blocks, indicating a change from uniform sliding to an active-passive mechanism (Figure 4.16). They also decreased by three orders of magnitude (maximum displacement is 109 m for the 50 m spacing model versus 0.3 m for the 500 m spacing model). The simulated sliding mass thus stabilised with increasing block size, agreeing with Sitar and MacLaughlin’s (1997) observations using Discontinuous Deformation Analysis (DDA).
Figure 4.16. Change in mechanism from uniformly sliding (a and b) to active-passive (c and d) in the UDEC dry simulations with a change in joint spacing from 50 m to 500 m. The location of the section (Section 2) is shown in Figure 4.1b.

Preliminary Trigon results using Sections 2 and 10A of Rossi and Semenza (Hendron and Patton, 1985) show that most of the brittle deformation in the sliding mass coincides with the secondary shear surfaces hypothesised by Rossi and Semenza (Hendron and Patton, 1985) (Figure 4.17). Rossi and Semenza extrapolated shears from surface observations, and their subsurface interpretation may have considerable uncertainty. Simulated shear failure is constrained to the lower half of the failure mass in both sections, whereas tension dominates at the surface. Damage in Section 2 is greatest at the two lower changes in gradient of the failure surface, where shear surfaces propagate at 55° and 75° from horizontal. These shear locations and orientations agree well with those determined in Phase2. Shear surfaces in Section 10A also occur at gradient changes in the sliding surface, at orientations of 60° to 90° from horizontal. Damage is intensified in the toe of Section 2 and is higher in Section 10A in general.
Figure 4.17. Preliminary UDEC Trigon results for Rossi and Semenza’s sections 2 and 10A. Red dashed curves represent shear surfaces in all figures. c) and d) modified from Rossi and Semenza (Hendron and Patton, 1985).

4.4.4. 3D Discontinuum Analysis of Kinematics and Block Size

Methodology and Model Organisation

To investigate block kinematics and interactions, I used a simplified three-dimensional geometry in 3DEC (3 Dimensional Distinct Element Code; Itasca, 2013a) (Figure 4.18). The geometry is based on the GoCAD model of Bistacchi et al. (2013), but I excluded all topographic complexity to fully appreciate kinematic controls without unnecessary detail.
Each of the major structural planes controlling the failure – the Col Tramontin and Col delle Erghene faults, the axial plane of the refolded southern limb of the Erto Syncline, the east and west limbs of the Massalezza Syncline, and a basal release plane – were included in the models and their orientations varied systematically to determine their influence on the kinematics of the Vajont Slide. The dips and dip directions of each of the Col Tramontin Fault (east lateral boundary), the south limb of the Erto Syncline (part of the basal surface), and the east and west Massalezza Syncline (rear release) planes were varied in models assuming a basal plane dip of 5°, a subvertical discontinuity set (DS 6) spaced 200 m, and a friction angle of 5°, as indicated in Table 4.6. Average orientations were determined from photogrammetric analyses (Chapter 3). Block size within the sliding mass was decreased, and block number increased from one to 145 (Figure 4.19a to h). Model subgroups can be categorised as: I) no discontinuities
I used a rigid block constitutive model throughout, assuming that the sliding material remained intact and that discontinuity properties dominated failure. I assumed Coulomb-slip area contact constitutive behaviour for all discontinuities, and introduced two sets of joint properties, jmat 1 and jmat 2, in some models to study the effects of discontinuity properties on stability. The first set of discontinuity properties was applied and reduced over several stages to simulate strength degradation over time. Next, the second set of discontinuity properties was applied to the back of the chair-shaped failure surface (the two surfaces representing the east and west limbs of the Massalezza
Syncline) and/or to secondary discontinuities including DS2 and DS6 and the Massalezza Gully plane. I did not vary jmat 2 properties in the simulations. The reader is referred to Table 4.8 and Table 4.9 for all material and discontinuity properties used in the models.

**Table 4.7.** Number of blocks in the simple 3DEC kinematic models in relation to structures and block boundaries. See Figure 4.19 for illustration of model geometries. MG = Massalezza Gully, DS = discontinuity set, spac = discontinuity spacing.

<table>
<thead>
<tr>
<th>Block boundary</th>
<th>Number of blocks</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>1 2 3 3 12 55 145</td>
</tr>
<tr>
<td>E-W boundary</td>
<td>X X X</td>
</tr>
<tr>
<td>MG</td>
<td>X X</td>
</tr>
<tr>
<td>Secondary shear</td>
<td>X</td>
</tr>
<tr>
<td>DS6 (spac = 200 m)</td>
<td>X X X</td>
</tr>
<tr>
<td>DS2 (spac = 200 m)</td>
<td>X X</td>
</tr>
<tr>
<td>DS1 (spac = 200 m)</td>
<td>X</td>
</tr>
</tbody>
</table>
Figure 4.19. Decreasing block size with the addition of structures in dry 3DEC simulations. I to IV represent model subgroups, and a) to h) are individual models with different block numbers. DS = discontinuity set. Locations and orientations of major planes are based on field observations, and discontinuity sets are based on photogrammetric analysis. (From Wolter et al., 2013a.)
Table 4.8. Rock mass properties used for the 3DEC simulations, based on Ghirotti (1992).

<table>
<thead>
<tr>
<th>Property</th>
<th>Vajont Limestone</th>
<th>Sliding mass</th>
</tr>
</thead>
<tbody>
<tr>
<td>Density (kg/m³)</td>
<td>2690</td>
<td>2610</td>
</tr>
<tr>
<td>Bulk modulus (GPa)</td>
<td>6.1</td>
<td>2.7</td>
</tr>
<tr>
<td>Shear modulus (GPa)</td>
<td>3.7</td>
<td>1.6</td>
</tr>
</tbody>
</table>

Table 4.9. Discontinuity properties used for the 3DEC simulations.

<table>
<thead>
<tr>
<th>Property</th>
<th>Jmat 1</th>
<th>Jmat 2</th>
</tr>
</thead>
<tbody>
<tr>
<td>Tensile strength (MPa)</td>
<td>0</td>
<td>0</td>
</tr>
<tr>
<td>Normal stiffness (GPa/m)</td>
<td>5</td>
<td>50</td>
</tr>
<tr>
<td>Shear stiffness (GPa/m)</td>
<td>0.6</td>
<td>6</td>
</tr>
<tr>
<td>Friction angle (°)</td>
<td>5-60</td>
<td>30</td>
</tr>
<tr>
<td>Cohesion (MPa)</td>
<td>0-0.05</td>
<td>0.01</td>
</tr>
</tbody>
</table>

I generated a more geometrically complex model using Bistacchi et al.’s (2013) geometry and the block and sub-block boundaries determined from engineering geomorphologic mapping of the slide (Chapter 2; Figure 4.20).
Figure 4.20. Geometry and properties of the more complex 3DEC landslide block and sub-block model based on geomorphological observations.

**Results of 3DEC Simulations for an Assumed Dry Slope**

3DEC analyses were more stable than their two-dimensional counterparts. The orientations of major discontinuities defining the failure proved to be significant. For example, the dip of the basal plane connecting the planes representing the Erto Syncline and the Col Tramontin Fault had to be 5° or greater to the north for failure to occur. If it was horizontal, the sliding mass was stable. The Erto Syncline plane was most critical to failure kinematics, as Hendron and Patton (1985) suggested. Changing the dip and dip direction of this plane produced both the maximum and minimum total displacements of all simulations. The maximum total displacement of 222 m occurred in the model with the Erto Syncline oriented 10°/090° (dip/dip direction) (Figure 4.21), whereas the minimum total displacement of 0.28 m (or essentially stable) occurred when the plane was oriented 20°/110°, or into the slope. Because the Erto Syncline limb is dipping into the slope in the latter case, stability is expected. The former case indicates the
complexity of the Vajont Slide. The southern Erto Syncline limb likely contributed to a complex pentahedral or hexahedral wedge, as demonstrated in the Swedge analysis.

Figure 4.21. Displacement magnitude and vector plot of the modelled sliding mass that attained the maximum displacement (222 m), given an orientation of the Erto Syncline of 10°/090°, DS 6 spaced 200 m, a basal plane oriented 5°/000°, and dry conditions.

Another significant aspect of the Vajont Slide indicated by three-dimensional analysis is the importance of block size and kinematics. One-block models were stable (maximum displacement <0.3 m), indicating the requirement of a multi-block sliding mass, and thus an increase in kinematic freedom as suggested by block theory (Goodman and Shi, 1985), for failure to occur. Maximum displacement increased to 1.8 m when the boundary between the east and west main blocks of Superchi (2012) and Bistacchi et al. (2013) was added, creating a two-block sliding mass (Figure 4.22). When the Massalezza Gully plane was added instead, again creating two blocks in the sliding mass, maximum displacement increased by more than three times to 4.8 m for a one-joint-material model. Only the west block failed, however; the east block was relatively stable. This result indicates that, although Massalezza Gully appears to have remained intact for the duration of sliding, it may have acted as a significant release plane early
during the failure. Additional study is required to clarify the nature of the release plane in the Massaleza Gully area. When both the E-W block boundary and the Massaleza Gully are included in a one-joint-material model, creating three blocks, displacement is highest east of the gully, but failure is global, that is, all three blocks move significantly.

![Figure 4.22](image)

**Figure 4.22.** Number of blocks in the sliding mass versus maximum block displacement and maximum joint shear and normal displacement in each simulation. Letters correspond to those in Figure 4.19.

Other three-block sliding masses that included the east-west block boundary and a secondary shear surface showed further increases in displacement, with a maximum displacement of almost 10 m. Displacement increased by two to three orders of magnitude when the three discontinuity sets determined from photogrammetry (Chapter 3) were included and spaced 100 to 200 m apart. Most of the maximum displacement magnitudes in all models occurred at the front, west corner of the sliding mass, suggesting failure initiation in the west, either as uniform sliding at the deposit front or as active-passive failure along the west headscarp.

Joint normal displacement was monitored as a proxy for dilation in my analyses. As seen in Figure 4.22, maximum joint normal displacement follows similar trends as the maximum block displacement and maximum joint shear displacement when block size decreases: maximum joint normal displacement increases as block size decreases, indicating a more fragmented and dilated sliding mass. Corkum and Martin (2004) also
show an increase in dilation with an increase in block number or decrease in block size. The location of the maximum joint normal displacement is consistently at the front west corner of the sliding mass, where the Erto Syncline plane intersects the Massalezza Syncline, in the transition zone between the active and passive regions of the mass. These results indicate higher damage in this area and support the Phase2 and UDEC models.

As more discontinuities were included in the sliding mass, its behaviour changed from blocks moving in a uniform direction to blocks moving in semi-independent directions (Figure 4.23). Displacement directions diverged from mainly N and NNE to include more westward-oriented directions. Broili (1967) suggests a movement direction of N to NNE, which the low block number models corroborate. The number of grid points moving significantly (arbitrarily defined here as ≥ 0.1 m to show trends in the one-block model as well) also increased dramatically from 15 in the one-block model to 3208 in the 145-block model.

Given the relatively intact nature of the deposit in the field, it appears that there are an optimal number of blocks required in 3DEC to simulate the observed post-failure topography. Considering block size and movement direction, the models suggest this optimum to be between 3 and 12 blocks. Field mapping indicates the actual number of blocks to be two main blocks and 5 sub-blocks within the west block.

When jmat 2 (Table 4.9) was applied to subvertical joints, it stabilised the sliding mass. The increased friction angle and cohesion values reduced total displacements from 50% up to an order of magnitude. In the 55-block model, for example, maximum displacement magnitude changed from 282 m to 127 m when the second set of joint properties was used. Further analysis of stability thresholds and input parameter ranges is required for more definitive conclusions to be drawn.
Figure 4.23. Effect of block size on slide behaviour. As block size decreases (number of blocks increases), movements become more distributed and independent. See Figure 4.19 for block geometries. a) – f) Rosettes of block vector orientations. g) Rosette of Broili’s (1967) vector orientations (note agreement with simulated slope movements).

The preliminary model incorporating complex topography and blocks mapped in the field (Figure 4.20) illustrates the development of the sliding surface and secondary shear surfaces as properties of the discontinuities are degraded (Figure 4.24). Tensile
failure is more common than shear failure in the sliding mass and increases significantly when the friction angle of the sliding surface is lowered from $20^\circ$ to $15^\circ$. Tensile failure is initially more common and occurs consistently at the headscarp, as well as in areas of the toe of the failure. Shear failure does not follow a trend as discontinuity properties are degraded, but occurs on the block boundaries and on the sliding surface at later stages in the complex topography model.

![Figure 4.24](image)

**Figure 4.24.** Friction angle stage versus number of tensile and shear cracks for the five-block sliding mass model with complex topography. Stage 1: $\phi_{ss} = 90^\circ$ (initial equilibrium), stage 2: $\phi_{ss} = 36^\circ$, stage 3 = $\phi_{bb} = 40^\circ$ (from $90^\circ$), stage 4 = $\phi_{ss} = 20^\circ$, stage 5 = $\phi_{ss} = 15^\circ$. ss = sliding surface, bb = block boundaries.

4.4.5. 3D Lattice-Spring Analyses of Internal Deformation and Brittle Fracture

**Methodology and Model Organisation**

I used *Slope Model* (Itasca, 2013b), a new three-dimensional lattice-spring code, to simulate brittle fracture of the Vajont failure mass and investigate the effects of groundwater on the slope (Havaej et al., 2013, 2014; Wolter et al., 2013a). The code was developed as part of the Large Open Pit project to examine complex rock mass behaviour in open pit mines. Havaej et al. (2013; 2014) adapted the code for the Vajont
Slide by importing Digital Elevation Models (DEMs) of pre- and post-1963 topography and the sliding surface (Bistacchi et al., 2013), and adding important discontinuities such as the boundary between the east and west main blocks, discontinuity sets, and the Massalezza Gully using a Synthetic Rock Mass (SRM) approach. The properties used are based on published values for limestone, laboratory testing (Superchi, 2012) and Ghirotti (1992) (Figure 4.25). I also conducted sensitivity analyses on friction angle, ranging from 10° to 45° in dry simulations. Total displacement and crack numbers were recorded for each simulation to monitor failure (Havaej et al., 2014, for further details on the methodology applied).

![Figure 4.25. Geometry and properties of the Slope Model simulations. Models were run with various discontinuity assumptions. White dashed lines represent discontinuity sets (DS), and black dashed lines represent individual block boundaries.](image)
Results of Slope Model Simulations for an Assumed Dry Slope

Sensitivity analyses of friction angle in dry, one-block sliding mass models in *Slope Model* indicate the effects of this parameter on slope stability. Using the cracking criterion developed by Havaej et al. (2012) and Lorig et al. (2010), whereby the number of new cracks is taken as an indicator for failure, the critical friction angle for the Vajont Slide is approximately 14° in the *Slope Model* simulations (Figure 4.26). Above this value the slope is stable, and below this value displacement increases exponentially and cracks continue to generate at a constant rate, suggesting continued failure. The critical friction angle determined is between those of the Phase2 models (16°) and the UDEC (12° - 16°) and 3DEC (<15°) dry models, suggesting that the sliding masses in the preliminary *Slope Model* simulations are more stable than those in the continuum simulations, and slightly less stable than those in the discontinuum simulations.

![Graph showing sensitivity of the simulated sliding mass in *Slope Model* to the friction angle of the sliding surface assuming dry conditions.](image)

**Figure 4.26.** Sensitivity of the simulated sliding mass in *Slope Model* to the friction angle of the sliding surface assuming dry conditions. After cracking related to attaining initial equilibrium, the number of cracks either stabilises (stable model), or continues to increase, indicating instability.
The placement and alignment of crack clusters agrees well with the secondary shear surfaces of Rossi and Semenza (in Hendron and Patton, 1985) and the extensional areas mapped in the debris (Chapter 2) (Figure 4.27). Most crack clusters trend roughly E-W in both one-block and multi-block simulations. The cracks in the multi-block models are slightly more distributed across the slope than those in the one-block models, and are controlled by discontinuities that are explicitly included. The areas affected most by cracking in the models align with the toe of the failure and the transition between the steeper upper slope and the flatter lower slope of the failure mass. Hence, the cracks concentrate in the Prandtl wedge zone and where the Erto and Massalezza synclines intersect, corroborating the Phase2 and UDEC modelling and Mencl’s (1966) hypothesis (Figure 4.28). The angle of the internal shear in profile (80°) agrees well with those determined in the Phase2 simulations (65° to 90°). The concentration of cracks in the zone of transition may explain why the one-block 3DEC models remained stable whereas the Slope Model simulations failed: the models in the latter code allowed for internal deformation and damage of the failure mass, and thus progressive weakening of the material, leading to instability.
Figure 4.27. Clusters of cracks formed in the dry a) one-block and b) multi-block Slope Model simulations, compared with c) extensional areas mapped on 1960 air photographs. White line in a) indicates location of section in Figure 4.28.
4.5. A Comment on the Effects of Groundwater at Vajont

In my simulations of the Vajont Slide, I have focused on pre-reservoir conditions. I have emphasized structural, geomorphic, and mechanical controls on the development of the failure. Nevertheless, precipitation and reservoir levels clearly influenced pore water pressures and ultimately triggered the catastrophic failure in 1963. Hence, groundwater was critical to the final stages of the failure. Here, I present preliminary Slope Model results that are a part of ongoing research and will be documented in the dissertation of M. Havaej. Material and discontinuity properties for these models are the same as in Section 4.4.5. A fixed water table based on Hendron and Patton’s (1985) hypothesis for the hydrogeological conditions at Vajont was assumed.

Preliminary Results of Slope Model Simulations for an Assumed Wet Slope

The preliminary Slope Model simulations indicate that the inclusion of a groundwater table increases the critical friction angle to 17°, thus decreasing stability, but it decreases the number of cracks that form in the sliding mass (Figure 4.29). The effect of groundwater is hence to decrease the internal brittle deformation of the mass while decreasing stability of the landslide. Further studies on the effects of various groundwater levels and are the object of ongoing research.
Figure 4.29. Graph illustrating the number of cracks that form in dry and wet Slope Model simulations over the specified model run (calculation) time.

4.6. Discussion and Conclusions

In this chapter, I have analysed the Vajont Slide with a toolbox of numerical software. Each “tool” highlights a different aspect of the landslide and its settings. Table 4.10 lists the results determined using each code.
<table>
<thead>
<tr>
<th>Code</th>
<th>Output</th>
</tr>
</thead>
<tbody>
<tr>
<td>Phase2</td>
<td>• principal stress and shear strain distributions</td>
</tr>
<tr>
<td></td>
<td>• influence of K ratio on stability</td>
</tr>
<tr>
<td></td>
<td>• critical friction angle of slide</td>
</tr>
<tr>
<td></td>
<td>• Prandtl wedge damage zone</td>
</tr>
<tr>
<td></td>
<td>• angled secondary shear surfaces</td>
</tr>
<tr>
<td>Swedge</td>
<td>• importance of structural controls</td>
</tr>
<tr>
<td></td>
<td>• importance of persistence</td>
</tr>
<tr>
<td></td>
<td>• pore pressure effects</td>
</tr>
<tr>
<td>UDEC</td>
<td>• importance of sliding surface geometry</td>
</tr>
<tr>
<td></td>
<td>• importance of block size</td>
</tr>
<tr>
<td></td>
<td>• brittle fracture using Trigon method</td>
</tr>
<tr>
<td></td>
<td>• Prandtl wedge damage zone</td>
</tr>
<tr>
<td></td>
<td>• angled secondary shear surfaces</td>
</tr>
<tr>
<td>3DEC</td>
<td>• importance of controlling structures to kinematics</td>
</tr>
<tr>
<td></td>
<td>• importance of block size</td>
</tr>
<tr>
<td></td>
<td>• estimation of dilation</td>
</tr>
<tr>
<td></td>
<td>• inclusion of blocks observed in the field, separated by secondary shears</td>
</tr>
<tr>
<td></td>
<td>• simple and complex topography effects</td>
</tr>
<tr>
<td>Slope Model</td>
<td>• friction angle sensitivity</td>
</tr>
<tr>
<td></td>
<td>• Prandtl wedge damage zone</td>
</tr>
<tr>
<td></td>
<td>• angled shear surfaces</td>
</tr>
<tr>
<td></td>
<td>• brittle fracture forming secondary shears</td>
</tr>
<tr>
<td></td>
<td>• pore pressure effects</td>
</tr>
<tr>
<td></td>
<td>• complex topography effects</td>
</tr>
</tbody>
</table>

Tectonic and geomorphic processes proved to be significant in conditioning the slope for failure. The Phase2 models showed high principal stress concentrations in Vajont Gorge as Vajont Valley was exhumed and eroded. These high stress concentrations would have focussed damage at the toe of the slope, weakening the rock mass. The implications of the stress distributions may be extrapolated to the prehistoric gully that Semenza (Semenza, 2001) found. If my hypothesis that the Vajont Slide was a
slowly deforming sackung-type failure for most of its history is correct, both the prehistoric and modern gorges would have influenced its toe. An example of this influence may be the prehistoric toe failure and the November 1960 landslide that occurred at the same location. Tensile and shear failure occurred along the slopes of the gorge, while elements also failed in tension farther up the valley walls.

Generally, extension occurred at ridge crests, and compression at the valley bottom, supporting the findings of Leith (2012). Tensile failure also appeared at perturbations in the topography, such as convexities in the slopes, including the area where the Vajont Slide headscarp daylighted (Figure 4.11). As horizontal stresses increased ($K > 1$), the propensity for yield also increased, illustrating the significance of in situ stress conditions. All of these results suggest that the sustained interaction and positive feedback between geomorphic and tectonic processes are important. Principal stresses concentrate in topographically anomalous areas such as Vajont Gorge (Figure 4.30), and geomorphic processes exploit tectonic structures. An example of the latter is the karstic plain above the Vajont Slide. Karst processes have eroded towers and created dolines where the southern limb of the Erto Syncline meets the Belluno Anticline in an area of shallower slope. As seen in the Phase2 models, tension is focussed on the dolines, contributing to further damage of the rock mass.

A simulated glacier in Phase2 significantly increases damage in the staged slope models. Eberhardt et al. (2004) show the effects of including pore water pressure in glacier simulations, with a significant increase in slope damage. Considering the importance of groundwater at Vajont, this approach could be applied in future studies. The post-glaciation stage in the models consistently contained the most yielded elements along valley flanks of all stages in each simulation. Hence, both glaciation and deglaciation damaged the slope. My preliminary results applying a Mohr-Coulomb criterion agree with those of Leith (2012). In his study of the evolution of two Alpine valleys in Switzerland, he demonstrated how micro-cracking and macroscopic tensional fractures develop and damage increases in rock slopes in response to glacial loading and unloading using a trilinear criterion (Diederichs, 2003). Future work could include the investigation of the Vajont slope using the same trilinear criterion.
Another consideration in the analysis of the Vajont Slide is material properties. Lithology appears to have played a secondary role in the failure. However, all analyses indicate the importance of sliding surface friction, with critical friction angles of dry simulations below 20°, generally agreeing with results listed in Hendron and Patton (1985). For the assumed actual failure surface, the Phase2 results indicate a critical friction angle of 16°; those in UDEC are between 12° and 16°, depending on block size; those in 3DEC are less than 15°, considering block size and shape; and results in Slope Model indicate a critical friction angle of 14° (Table 4.11). The friction angle results of the three-dimensional simulations tend to be lower, suggesting that 3D models, as expected, indicate higher stability than 2D models. The 3D simulations consider the complex geometry of rear and lateral release surfaces at Vajont and indicate their importance in controlling stability. The low critical friction angles (as low as 5°) required for failure along the sliding surface in the dry simulations suggest that clays and other weak materials in the sliding shear zone, kinematics, and secondary surfaces dividing the landslide body into discrete blocks, and groundwater effects were very important.
Table 4.11. Critical friction angle results for dry analyses in each code. UDEC and 3DEC values depend on number and size of blocks included in the sliding mass.

<table>
<thead>
<tr>
<th>Code</th>
<th>Critical friction angle (°)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Phase2</td>
<td>16</td>
</tr>
<tr>
<td>UDEC</td>
<td>12-16</td>
</tr>
<tr>
<td>3DEC</td>
<td>&lt;15</td>
</tr>
<tr>
<td>Slope Model</td>
<td>14</td>
</tr>
</tbody>
</table>

The tectonic structures on the north slope of Monte Toc controlled the geometry and kinematics of the Vajont Slide (Figure 4.32I). These structures form a hexahedral wedge-biplanar sliding rock mass. The Col Tramontin Fault was the main lateral release surface for the east main block, and the Col delle Erghene Fault formed the rear release surface in the west. The limbs of the Massalezza Syncline acted as the basal release surface in the upper part of the failure. The Erto Syncline was particularly important as a basal release surface on the lower west slope, and directed movement of the west main block to the northeast. A subhorizontal basal plane was required east of the Erto Syncline plane to provide kinematic freedom for the blocks in my models. This plane represents a composite surface comprising step-paths along bedding and orthogonal discontinuities to allow stepping-up of the failure surface to the east, as Hendron and Patton (1985) hypothesised.

The geometry of the sliding surface influenced stability. In my UDEC simulations, linear and circular surfaces were least stable and undulating surfaces most stable. In reality, the west half of the failure has a biplanar sliding surface in profile with active upper and passive lower zones. This geometry focusses damage in the Prandtl wedge transition zone between the active and passive zones, as Mencl (1966) proposed. At Vajont, the Prandtl wedge coincides with the hinge line of the Erto Syncline and the intersection of the Erto and Massalezza synclines, where complex interference patterns create rough, irregular bedding surfaces. These factors damaged the rock mass in this area intensely. In my discontinuum models, secondary shear surfaces were required at the location of the Prandtl wedge for failure to occur. In the Phase2 and Slope Model simulations, damage consistently focussed in this zone, suggesting development of secondary shear surfaces here at angles between 45\(^\circ\) and 90\(^\circ\). The field evidence
supports these conclusions. The depressions at the rear of the Pian della Pozza, a flat bench west of the Massalezza Gully that existed prior to 1963, align with the transition zone (Figure 4.31). They may have formed from karst dissolution, exploiting a pre-existing area of weak, damaged rock. In reality, the Vajont Slide mass probably accommodated stress by both brittle fracture along discrete discontinuities and internal deformation of individual landslide blocks.

**Figure 4.31.** Relationship between mechanics in models and geomorphology. a) The highest displacement occurs in the upper east active block in the dry Slope Model simulation (see Figure 4.28 for cross-section), and the Prandtl transition zone coincides with the Pian della Pozza depression. b) 1960 aerial photograph shows topography.
The Prandtl wedge is not obvious east of the Massalezza Gully. Although a relatively flat bench existed there before the catastrophic failure, it does not match the Pian della Pozza in elevation or slope angle. Also, there are no depressions. The slope morphology agrees with the hypothesis that the east half of the failure had a more circular sliding surface and would not have developed a clear Prandtl wedge. In the analysis of east sections (Rossi and Semenza’s Section 10A and Bistacchi et al.’s E Section), damage is distributed throughout the sliding mass, not constrained to a few discrete zones. Shear strain in the toe is particularly high.

Discrete shear zones probably developed gradually through progressive failure, or the coalescence of microcracks and pre-existing macroscopic discontinuities in the rock mass, as indicated, for example, in the Slope Model simulations (Figure 4.32I). The main sliding surface followed weak clay beds, breaking through intervening carbonate beds. Based on movement and seismicity records, Petley and Petley (2006) suggested that brittle fracturing occurred predominantly in the clays at the microscopic scale, and that macroscopic behaviour was ductile from 1960 to 1962. As cracks coalesced, brittle failure in the clays became dominant. They hypothesised that brittle fracture through the carbonates was not central to the development of the shear surface, as high seismic activity would be expected in this case, which was not observed. Nevertheless, some failure through carbonates was required to connect shear planes in the weak clay layers. This mechanism was probably more important in the initial stages of the slow deformation of the slope, possibly before seismic records became available. The quality of the existing seismic records is also questionable; brittle fracture may not have been detected. Over time, cataclasites developed in reaction to the shear stresses concentrated in the weak clay beds, eventually forming the thick, brecciated units Semenza (Semenza, 2001) observed and attributed to the prehistoric failure. Paronuzzi and Bolla (2012, 2013) have reinterpreted the basal shear zone at Vajont based on borehole data and field observations. They suggest that the shear zone may have attained thicknesses of up to 80 m, and determine average sliding surface friction angles of 17° to 27°. Although my modelling results agree with their friction angle results, I cannot confirm their hypothesised shear zone thickness.

Secondary shear surfaces separating discrete blocks, such as between the east and west main blocks and the Massalezza Gully, were critical to the initiation of the
landslide, as they provided further kinematic freedom within the landslide body. They also controlled block size, which is an important aspect of rock slope failures (Sitar and MacLaughlin, 1997; Corkum and Martin, 2004). The slide comprised several large blocks on the order of hundreds of metres that remained internally intact. As my UDEC and 3DEC results demonstrated, a one-block sliding mass is stable, but multi-block models, blocks move independently of one another (Figure 4.23). Significant displacement of the sliding mass only occurred in multi-block simulations, as suggested by Martin and Kaiser (1984). Thus, the simulation of the Vajont Slide required enough blocks to allow movement, but few enough blocks for the slide to behave relatively coherently. The optimum number of blocks based on numerical simulations is between 3 and 12; geomorphological field evidence suggests two main blocks (east and west) and five sub-blocks within the west block (Figure 4.32II; Figure 4.33). The Phase2 and Slope Model results confirm the existence of secondary shear surfaces within the sliding mass. The discrete blocks remained internally intact, with little fragmentation or dilation. Dilation in the 3DEC models was highest at the Prandtl wedge zone in the west. In reality, dilation related to the fold-interference asperities would have occurred within the basal sliding zone, as the sliding mass would have had to dilate over these asperities or shear through them.

The role of the Massalezza Gully remains unclear. Although it seems to separate distinct sub-blocks in the west main block based on morphology of the debris, it appears to have remained intact during emplacement. The 3DEC results suggest that a structure subparallel and close to the Massalezza Gully is important and may have facilitated failure, with displacements in the Massalezza two-block model surpassing those in the two-block model including the east-west block boundary, thus differing from present-day field observations and interpretations. A possible explanation may be that the gully was important in the initial stages of failure, but was dominated by other, weaker zones in later stages. The block boundary may also have been slightly to one side of the gully, thus preserving it.

My modelling results confirm the hypothesis that the west main block failed first, followed by the east block (Figure 4.32II – III). Although the east block is weaker due to its proximity to the Col Tramontin Fault, and the east sliding surface is steeper and more circular, the surface is rougher in the east because of interference between the Dinaric
and Neoalpine folding events. Thus, the failure surface would have developed as intact rock bridges ruptured through asperities. In addition, the east block may have been kinematically constrained by the west block. As the west block failed, it would have provided the kinematic freedom for the east block to move. This kinematic control on slide behaviour highlights the importance of three-dimensionality at Vajont. Two-dimensional analyses neglect three-dimensional structures and commonly underestimate the stability of a slope.

Figure 4.32. Conceptual diagram of the evolution of the Vajont Slide, showing the major structural controls and initial development of secondary shear surfaces (I), slow deformation of the blocks (II), and the sequence that occurred in the catastrophic failure (III and IV), with the west sub-blocks failing first and the east block following.
The critical influence of pore water pressure at Vajont cannot be ignored. Preliminary analyses in Swedge and *Slope Model* highlight the effect of pore water in decreasing stability and the number of cracks required before catastrophic failure initiates.

### 4.6.1. Contributions and Future Work

The geomechanical modelling of the Vajont Slide presented in this chapter forms an integral part of a broader study of the geomorphological and tectonic settings, failure scar characterisation, and engineering geomorphology of the failure (Chapters 2 and 3). Highlights include demonstration of:

1. the significance of the interplay between endogenic and exogenic processes in conditioning the Vajont Slide for failure,
2. the role of material properties such as critical friction angle in the stability of the slope,
3. the effect of local structural features in controlling the geometry and kinematics of the failure,
4. the importance of pre-existing discontinuities and brittle fracture through intact rock in the development of the main sliding zone and secondary shear surfaces,
5. the impact of block size and number to the kinematics of the failure, and
6. the influence of internal deformation of the sliding mass on the initiation of failure.

Groundwater simulations are currently being completed to determine the effects of pore water pressures on stability, and sophisticated models simulating different spatial roughness of the sliding surface are being developed. Future modelling could investigate the interaction between groundwater and kinematics.

Acknowledgements

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Chapter 5.

A Forensic Engineering Geological Investigation of the Madison Canyon Slide, Montana, USA

Abstract

I used an integrated approach to investigate the Madison Canyon Slide (20 million m$^3$), which killed 24 people in southern Montana, USA, in 1959. My methods include engineering geomorphologic mapping, field surveys, laboratory testing, long-range terrestrial digital photogrammetry, kinematic analysis, and 2D numerical modelling. My objective is to determine the conditions, mechanisms, movement behaviour, and evolution of the slide. I emphasise the importance of both endogenic and exogenic processes in conditioning the slope for failure, and determine an hypothesised sequence of events based on the morphology of the deposit. A section of the slope was slowly deforming before a nearby, magnitude-7.5 earthquake triggered the catastrophic failure in August 1959. The failed rock mass rapidly fragmented as it descended the slope towards Madison River. Part of the mass remained relatively intact as it moved on a carpet of pulverized debris. The main slide was followed by several debris slides, slumps, and rockfalls. The slide debris was extensively modified soon after the disaster by the U.S. Army Corps of Engineers to provide a stable outflow channel from newly formed Earthquake Lake. My modelling and observations show that the landslide formed as a result of long-term fatigue of the slope and the short-term seismic shaking. Static models suggest the slope was fairly stable prior to the earthquake; failure would have required a significant reduction in material strength. Dynamic models indicate that

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5 To be submitted as:
seismic ground motions were critical to failure; the ridge geometry and existing tension cracks in the initiation zone amplified ground motions.

**Keywords:** Madison Canyon Slide; long-range photogrammetry; engineering geomorphologic mapping; damage; dynamic numerical modelling

### 5.1. Introduction

The Madison Canyon Slide (Figure 5.1), also known as Earthquake Slide, happened in southern Montana on August 17, 1959, when a large ($M_w$=7.5) earthquake struck only 30 km from the slope. The seismic ground motions from the earthquake released 20 million m$^3$ of schist, gneiss, and dolomitic marble from a steeply sloping ridge overlooking Madison River. The failed rock mass fragmented and accelerated rapidly into the valley on an air cushion, destroyed two campgrounds, killed 24 campers, and blocked the valley in which Earthquake Lake soon began to form, all in the span of less than 60 seconds.

Despite the magnitude of the disaster and concern over breaching of the landslide dam by overflow from Earthquake Lake, the landslide has been the subject of relatively little scientific investigation. The U.S. Army Corps of Engineers (1960) published a report on the spillway it constructed soon after the slide to prevent catastrophic dam collapse (see also Harrison, 1974). Hadley (1964, 1978) wrote two seminal reports on the geological and climatic settings, surface features, rock mass characteristics, and mechanisms of the slide. He hypothesised that the sliding surface was biplanar. Trunk et al. (1986) modelled the runout of the slide using two-dimensional profiles and viscous and bi-viscous fluids to simulate rockslide mobility. Several other papers mention the Madison Canyon Slide, including Kent (1966), who discusses fluidisation of large landslides with trapped air, and Jibson (2009), who describes the connection between seismic activity and landslides. No subsurface investigations have been completed to date. More literature exists on the earthquake than on the landslide, and includes Ryall (1962), Tocher (1962), Witkind et al. (1962), Trimble and Smith (1975), and Doser (1985).
Figure 5.1. a) U.S. Geological Survey aerial photograph of the 1959 Madison Canyon Slide just after it occurred. b) The failure scar today.
I re-examine the landslide, more than 50 years after the event, with the intent of gaining new insights into its behaviour. I use an integrated forensic approach that combines traditional and state-of-the-art techniques to better understand the factors that conditioned the slope for failure, as well as the trigger, mechanics, and evolution of the landslide. My field investigations involved the application of engineering geological techniques, notably discontinuity line surveys, block size and block shape characterisation, Geological Strength Index (GSI) estimation, and engineering geomorphologic mapping. I combine field data with information derived from laboratory sample testing, aerial photograph interpretation, and digital terrestrial photogrammetry to characterise the material that failed as well as the surrounding rock mass. This information was then incorporated in two- and three-dimensional simulations of the landslide. The main objectives of this research are to:

- understand the interaction between endogenic and exogenic processes affecting the slope, and how they conditioned the slope for failure,
- determine the effects of seismic waves on the slope, and
- establish the evolution of the slope and the landslide through time and the sequence of events leading up to and including the catastrophic failure.

### 5.2. Regional Setting

#### 5.2.1. Tectonics and Lithology

The Madison Canyon Slide occurred on a north-facing slope at the mouth of the Madison River canyon in the Madison Range of southern Montana (Figure 5.2). The canyon is 10 km long and 250-400 m wide. Highway 287 crosses the toe of the slide approximately 30 km west of West Yellowstone, Montana.
The Madison Range comprises folded and faulted Late Archaean to Early Paleoproterozoic amphibolite-grade metasedimentary and meta-igneous rocks (O’Neill and Christiansen, 2002). The Laramide orogeny created the Gallatin and Madison ranges 80-35 million years ago, including N-S-oriented faults. Late Tertiary faults and extensional basins are superimposed on the Laramide structures. The most recent, E-W faulting is associated with the development and deformation of the Snake River Plain.
The Madison Range lies within the Intermountain Seismic Belt (ISB; Doser, 1985), which extends from northwest Montana to north Arizona and is characterised by shallow (<20 km) seismicity related to the differential motion between the Juan de Fuca, Pacific, and North American plates to the west. In the Yellowstone-Hebgen Lake area, 77 seismic events were felt prior to 1959, the largest of which was the November 23, 1947 Virginia City, Montana earthquake with a modified Mercalli intensity of VIII (Doser, 1985). The fault nearest the Madison Canyon Slide is the Madison Range Fault, an active normal fault at the western foot of the range (Figure 5.2). It may have been reactivated during the 1959 earthquake. The Hebgen and Red Canyon normal faults, east of Hebgen Lake and trending SW (Figure 5.2), are the sources of the 1959 earthquake sequence. More than 6 m of vertical surface displacement occurred on these structures during the earthquake (Doser, 1985).

Two units were involved in the landslide: i) biotite quartz-rich schist and gneiss and ii) dolomitic marble interlayered with thin quartzite bands (Figure 5.3; see Appendix C for detailed petrology). The dolomitic marble crops out at the base of the slope and west of the slide and served as a buttress for the weaker metamorphic rock above it until the earthquake in 1959. Chlorite-biotite schist, mica schist, mylonite, and amphibolite dominate the valley slope opposite the landslide. For this chapter, I refer to i) above as “schist”, and ii) as “marble”.

5.2.2. Climate and Geomorphology

The Madison River watershed is within a region characterised by a subarctic climate, with extended, cold winters and short, warm summers. Average monthly temperatures in the Madison Canyon area range from -4°C to 19°C, and average annual rainfall is 25-45 cm (Hadley, 1978). Madison Canyon was relatively dry in the summer of 1959, as evidenced by the large clouds of dust the landslide generated.

de la Montagne (1960) and Shelden (1960) described the geomorphology of the Madison River watershed. de la Montagne highlights several features, including the river’s source in the rolling Yellowstone upland, the path of the river through the Madison Range, and its emergence from Madison Canyon onto the broad Missouri Flats with its distinct series of fluvial terraces. At the front of the Madison Range, the river changes
course from east to north, likely following Laramide N-trending structures. After flowing into Ennis Lake, the river again enters a canyon in the Norris Hills, before joining the Jefferson and Gallatin rivers near Three Forks, Montana. de la Montagne (1960) suggests that the river valley formed in the Pleistocene.

Figure 5.3. Lithological units involved in the Madison Canyon Slide. a) Crossed polarized image of altered schist at 2X magnification. b) Schist and gneiss units at the west end of the headscarp. c) Crossed polarized image of dolomitic marble with interlayered quartzite at 2X magnification. d) Remnant of the dolomitic marble buttress at the base of the Madison Canyon Slide slope. Ca = carbonate, Chl = chlorite, F = feldspar, Q = quartz.

Judging from the steep, V-shaped valley of Madison River canyon and narrow tributary valleys, the site of the Madison Canyon Slide was not glaciated. However, a large hummocky moraine covers the valley floor upstream of the landslide site, at the confluence of Madison River and Beaver Creek. According to de la Montagne (1960) and Shelden (1960), the Madison Range was affected by local alpine glaciers during the last two Pleistocene glaciations, locally known as the Bull Lake and Pinedale glaciations;
several of these glaciers flowed from Yellowstone ice cap. East of Hebgen Lake, a large obsidian outwash plain marks the west edge of the ice cap in the Madison River valley.

5.3. Methods

The methodological strategy that I applied to the Madison Canyon Slide and its surroundings is summarised in Figure 5.4. I integrated geomorphological observations and engineering geomorphologic mapping with field and laboratory rock mass characterisation and field and photogrammetric discontinuity surveys. These methods provided validation and control for stability analyses. I performed simple and complex simulations, starting with kinematic and limit equilibrium analysis including pseudostatic analysis in Swedge v. 6.0 (Rocscience, 2014b), and progressing to landscape evolution models in Phase2 v. 8.0 (Rocscience, 2014a) and static and dynamic analyses in UDEC (Itasca, 2013c). I simulated the slope as both continuum and discontinuum masses to analyse different aspects of the landslide. In pseudostatic analysis, a constant seismic load is applied to a slope using horizontal and vertical seismic coefficients, which are multiplied by the slope’s weight. Accepted values for seismic coefficients range from 0.1 to 0.5, but there are no clear guidelines (Jibson, 2011). As there are obvious drawbacks to pseudostatic analysis (earthquakes are not static forces acting on slopes), I also conducted dynamic stress-deformation analyses in UDEC. I applied velocity histories to the base of each model and monitored model response at a series of history points. Each section below provides more detailed descriptions of components of the integrated methodology, and Appendix E provides samples of source code files.
5.4. Geomorphological and Geotechnical Investigations

5.4.1. Geomorphology

*Madison River Valley and the Madison Canyon Slide*

Over its 240 km length (Figure 5.5), Madison River loses 900 m in elevation and has an average gradient of 0.004 or 0.2° as it flows from south to north. Three lakes – Hebgen, Earthquake, and Ennis lakes – interrupt its flow (Figure 5.6). A significant knickpoint occurs just below Earthquake Lake at the mouth of Madison Canyon, where the landslide occurred.
Figure 5.5. Map of the Madison River watershed, showing the locations of topographic profiles parallel (blue line) and orthogonal (white lines) to the river. The X shows the location of the Madison Canyon Slide. (DEM courtesy of the U.S. Geological Survey.)
I drew topographic sections orthogonal to the Madison River channel at 20-km intervals along its length to document changes in valley topography (Figure 5.7). The river valley in Yellowstone Basin and along Hebgen Lake is wide and relatively flat (sections 2 and 3 in Figure 5.7). Below Hebgen Lake, the river enters a narrow, steep-walled canyon (Madison Canyon, section 4). Valley walls here slope more than 35°. After emerging from the canyon, the river changes course and flows northward in a broad valley (sections 5, 6, 7, and 8), bounded abruptly on the east by the Madison Range and on the west by the Gravelly Range. Local relief ranges from approximately 1900 m asl in the valley bottom to over 3000 m asl in the Madison Range. A second canyon is evident in section 9, north of Ennis Lake. Farther north, the river flows across a broad plain near Three Forks, Montana (sections 10, 11, and 12). Madison Canyon is unique along the river’s length in its very narrow, V-shaped cross-profile.

To further emphasise the changes in the river valley’s morphology, I plotted the chord-height (c/h) ratios of each profile (Figure 5.8; Walter, 1962). All profiles have c/h values greater than one, suggesting that the river valley is wider than it is deep. The two open, broad terminations of the river are clear, and have c/h ratios above 50. The Madison Canyon has the smallest c/h ratio (c/h = 7), emphasising its anomalous presence in an otherwise wide valley.
Figure 5.7. Topographic profiles perpendicular to Madison River at intervals of 20 km. Each profile is viewed downstream. VE = vertical exaggeration. Locations of profiles are shown in Figure 5.5.
Figure 5.8. Chord-height ratios of the 12 perpendicular profiles along the Madison River, shown in Figure 5.5 and Figure 5.7.

Five orthogonal profiles illustrate the detail of the slide area before and after 1959 (Figure 5.9 and Figure 5.10). The pre-1959 profiles were derived from a U.S. Geological Survey 1950 topographic map and the post-1959 profiles from U.S. Geological Survey Digital Elevation Model (DEM) data. Hadley’s (1978) sections, taken at similar locations, are shown in Figure 5.11 for comparison. The profiles highlight how the Madison Canyon Slide has altered the valley: the slide’s zone of depletion is obvious, especially in profiles C and D where the slope has lost approximately 100 m in elevation; the ridge crest has receded roughly 20-30 m, and the landslide deposits raised the valley floor by 50 to 80 m. Madison River has eroded a new channel through the deposit, which was initially excavated by the U.S. Army Corps of Engineers. My observations of the extent of the marble buttress agree with Hadley’s interpretation. The marble buttress pinches out from profile E to C, and is not present in profiles A and B.
Figure 5.9. Locations of profiles in the slide area before and after the Madison Canyon Slide (Figure 5.10). The profiles are approximately the same as those of Hadley’s (1978; Figure 5.11) sections. (1950 topographic map courtesy of U.S. Geological Survey.)
Figure 5.10. Topographic profiles at the locations indicated in Figure 5.9. The pre-1959 and post-1959 topography are based on the 1950 topography map and recent satellite imagery, respectively.
**Figure 5.11.** Hadley’s (1978) sections, taken at similar locations as indicated in Figure 5.9.

Note: Reprinted from Rockslides and Avalanches, 1, Hadley, J.B., Madison Canyon rockslide, Montana, USA, 172-180, Copyright (1978).
**Engineering Geomorphologic Mapping**

I constructed engineering geomorphologic maps of the landslide based on aerial photographic analysis and field transects. Mapping was completed according to Geological Society of London (1982) guidelines. Morphological maps show breaks and changes in slope, and morphogenetic maps are the interpretations of these descriptive maps, from which I derived relations between landforms within the slide area and infer processes operating on the slope.

**1959 Aerial Photograph Map**

The 1959 morphogenetic map (Figure 5.12), based on air photographs taken immediately after the landslide, highlights several features of the landslide, including the headscarp, debris lithology, and structures in the debris related to landslide kinematics. The headscarp strikes roughly WNW-ESE, is 800 m long, and can be separated into four zones. The westernmost, 170-m-long headscarp section is steep (> 50° slope angles), roughly 30 m high, and dominated by exposed schist-gneiss bedrock (Zone 1 in Figure 5.12). East of this area, the scarp is 160 m long, lower and covered by a secondary slump (Zone 2). It is has the form of a saddle between the higher steeper scarps on both sides, and coincides with a dry gully on the south side of the ridge. The third section (Zone 3) is a wedged-shaped scarp that is 240 m long, up to 50 m high, and has > 50° slopes with overhanging sections. The easternmost section (Zone 4) is 230 m long and is the source of a large secondary slump. The lateral scarps of the landslide are subparallel and oriented approximately N-S; both dip > 45°. The east margin, however, is about 500 m long, whereas the west margin is 250 m long. Although there do not appear to be any major structures controlling the scarps, the headscarp is delineated by foliation- and discontinuity-controlled wedges, and the lateral scarps parallel the major N-S-trending normal faulting in the region (Figure 5.2).

Marble debris forms a linear band up to 100 m wide at the outer limit of the debris sheet and is absent elsewhere. Because the marble debris extends to the far east end of the debris sheet, it is apparent that marble cropped out near the base of the slope farther east than can be seen today. About 50 m south of the ribbon of marble debris, and parallel to it, is a linear band of highly crushed schist. It appears lighter on the 1959 air photographs than the surrounding schist and may be related to a more weathered and
fractured zone of schist in the in-situ rock mass upslope. The colour variations in the schist-gneiss headscarp (mauve to rusted iron) represent slight lithological changes that are preserved in the debris (Figure 5.1b).

Figure 5.12. Morphologic and morphogenetic map of the Madison Canyon Slide, based on interpretation of the 1959 air photographs. Inset shows movement vector orientations.
Transverse ridges and longitudinal and transverse lineations have important implications for the kinematics and movement behaviour of the Madison Canyon Slide. The most obvious ridges are near the northeast margin of the debris, and, given their sinuous and rounded appearance, are hypothesised to be compressional structures associated with thrust faulting (Shea and van Wyk de Vries, 2008). These ridges have amplitudes of metres to decametres, are 50 to 500 m long, and are oriented roughly NW-SE or NNW-SSE (Figure 5.13a). They are assumed to be orthogonal to the direction of movement, and thus suggest NE displacement of the debris. They occur in areas with preserved vegetation and more cohesive debris. Longitudinal lineations cluster in two locations in the distal debris – one cluster of 21 lineations is south of the lighter crushed schist band, and the second, smaller (N = 3) cluster is in the northeast corner of the deposit. They are probably associated with shearing and strike-slip displacements. The lineations appear to be parallel to flow, with NW-SE to NNE-SSW trends, and are several tens of metres long (Figure 5.13b). These lineations resemble those noted in the debris of other rock avalanches, such as the 2002 Black Rapids rock avalanche in Alaska (Shugar and Clague, 2011). Transverse lineations are dominantly tension cracks related to normal faulting and are most common in the headscarp area and at the crests of secondary slumps. Most of the tension lineations cluster in Zone 4 of the headscarp area (Figure 5.12) and are related to slow deformation of the secondary slump failure there. Another cluster occurs in Zone 2 of the headscarp, again associated with a secondary slump. Some tension cracks occur in the central-east area of the deposit and are likely related to secondary movements after the main rock avalanche event. The lineations range in length from 5 to 90 m, and in orientation from NE-SW to WSW-ENE (Figure 5.13b), suggesting movement to the NW and NNW.

The deposit contains areas of hummocky terrain, particularly in the area below the Zone 4 headscarp and in the northeast where the compressional ridges formed. No distinct hummocks were observed; rather the surface is irregular. The hummocky terrain near the headscarp appears to have formed due to extension, whereas the terrain in the NE debris is probably due to compression of the deposit.
Figure 5.13. Polar plots of feature trend and length for a) compressional ridges, and b) shear and tension lineations in the Madison Canyon Slide deposit. Locations of features are indicated in Figure 5.12.
Based on morphology such as ridges and lineations, tree cover, cross-cutting relationships of features in the debris, and differences in cohesiveness, the debris sheet can be subdivided into ten blocks or domains (labelled A to J on Figure 5.12). The most apparent division, other than the lithological boundary between the schist and marble debris, is the boundary between fragmented debris more typical of a rock avalanche in the west and more cohesive material in the east. The largest domain (A), roughly 500 m wide and 750 m long, is characterised by fragmented debris with shear flow lines at its distal end. Material in this domain moved to the NNE and NNW (Figure 5.12 inset). This domain is bordered to the east by more cohesive blocks. Domain B supported standing trees and has sinuous ridges indicating compression and movement to the ENE. Minor movements to the NNW identified by the presence of extensional lineations in the NW corner of the domain probably are secondary. Domain C is characterised by a broad circular high in its centre and a depression at its northeast margin. Apparent shear lineations cross-cut the boundary between domains C and G. Domain D is another vegetated, more cohesive zone. Ridges in this domain are sinuous, but are located at the rear of the slide domain, suggesting compression in reaction to collision with material behind it. Another ridge at the NE corner of the domain indicates compression due to collision with the leading marble front (domain G). Domain E is a smaller, secondary block that separated from domain D and moved to the east. Domain F is another smaller ridged zone that moved to the NNE. It is separated from domains A and C by shear surfaces. Domains G to J are the leading edge of the Madison Canyon Slide, consisting of blocky marble debris. The divisions between them are shear zones. Domains G, H, and I moved NE, and domain J spread NE to NW.

Some geomorphic features within the deposit may reflect pre-event topography. The broad, 600-m-long ridge that crosses the deposit from SW to NE south of Madison River may be the formerly cropping out, more resistant marble unit upon which the deposit is draped (Figure 5.14a). It extends eastward from outcropping marble at the west edge of the slide. The area of bare ground just outside the upper part of the east lateral scarp is the side wall of a pre-existing gully that rotated into the accommodation space created by the failure (Figure 5.14b). A ~2-m-high, 350-m-long scarp parallel to the main east lateral scarp and trending roughly 020° is apparent on air photographs and
in the field. The block bounded by this scarp and the main scarp is crossed obliquely by metre-high scarps that are approximately parallel to the strike of foliation (252°).

**Figure 5.14.** Features in the Madison Canyon Slide area related to pre-1959 topography. a) Broad ridge in the deposit, which may be the surficial expression of underlying marble continuing from the remnant marble cropping out in the west. b) Secondary scarp parallel to main east lateral scarp that coincides with a pre-1959 gully. (Photograph courtesy of U.S. Geological Survey.)

**Field Map**

I created a detailed morphometric field map by measuring breaks and changes in slope with a clinometer along transects oriented roughly NNE-SSW and spaced 50 m
apart (Figure 5.15). A sample of the detailed map is shown in Figure 5.16, and the entire map is included in Appendix B.

Figure 5.15. Approximate locations of transects followed to map the Madison Canyon Slide debris. The area beneath the headscarp was inaccessible in the field. The white line outlines the study area; the golden dashed curve indicates the headscarp; and points indicate field stations.

Although most of the interpretation of the slide was completed based on the 1959 maps, the field map shows how the landslide has evolved over 50 years. Modification of the debris has been extensive, particularly on the north side of Madison River. Highway 287 crosses the debris, and a visitor’s centre, parking lot, and paths have been constructed on the debris. On the south side of the river, parallel linear ridges and gravel roads were constructed to access building material. Field maps highlighted movement of material. Boulders of marble were found near the river, as the U.S. Army Corps of Engineers used these to armour the river channel to prevent erosion. The largest marble
block, near the margin of the debris sheet, was designated the “Memorial Boulder” in commemoration of those killed by the landslide.

Figure 5.16. a) Sample of the field morphometric map showing slope directions and angles, and breaks and changes in slope. b) Photograph of the mapped area, at the northeast corner of the debris.
Evolution of the Landslide and Landscape

I reconstructed the evolution of the site of the Madison Canyon Slide using a series of historical U.S. Geological Survey aerial photographs dating from 1947 to 2002. I mapped features such as gullies, scarps, and tension cracks, Madison River, marble outcrops, and mass movements on each set of georeferenced photographs taken in 1947, 1954, 1959 (both before and after the landslide), 1976, 1982, and 2002. Figure 5.17 shows an example of the linear features mapped on the 1947, 1954, and July 1959 air photographs.

![Image](image_url)

**Figure 5.17.** Example of lineations, gullies, Madison River, and marble outcrop mapped on the 1947, 1954, and pre-event 1959 U.S. Geological Survey air photographs. The white dashed line outlines the approximate location of the future headscarp of the Madison Canyon Slide. Base photograph is the 1954 image (courtesy of U.S. Geological Survey).
Ephemeral gullies aided in georeferencing and delineating the failure, especially along its east margin. Gullies undisturbed by the 1959 landslide are unchanged over the period spanned by the photographs. The gullies appear to terminate in mid-slope positions. Comparison of pre- and post-slide photographs indicates that two gullies coincided with the east lateral margin of the 1959 slide (Figure 5.17).

Several tension cracks and scarps are visible in the pre-slide photos, mainly near the ridge crest and on the south-facing slope of the ridge. The most conspicuous scarps are just east of the future slide and indicate a larger area of instability than involved in the 1959 failure. Prior to 1959, only a few scarps can be seen. As illustrated on the 1959 map (Figure 5.12), the number of extension lineations increased dramatically when the Madison Canyon Slide occurred. Since 1959, the number of lineations has not changed significantly. However, I observed fresh cracks in the field in 2011 and 2012, mainly near the west end of the headscarp (Zone 1 in Figure 5.12) and in the saddle area (Zone 2). The unstable area thus continues to expand toward the west.

Marble outcrops are distinctive on the south slope of Madison Canyon. They have the form of unvegetated, castellated ridges separated by gullies (Figure 5.18). The ridges are oriented downslope, trending approximately N-S and are several hundreds of metres in length. The contact between the marble and schist units in the slide area may have acted as part of the west lateral release surface on the upper part of the slope.

I identified several potential debris slides, slumps, and rockfalls on the pre-slide photographs, indicating a longer history of mass movements. The possible debris slide scars are located east of the 1959 slide area, but on the same slope. Debris flows travelled down Rock Creek on the opposing valley side. A large alluvial fan has been deposited where Rock Creek meets Madison River, causing the river to flow against the opposite valley wall, the future site of the Madison Canyon Slide. There is no evidence of large failures in the slide area prior to 1959.
Field Observations and Measurements

Field methods included discontinuity line surveys, examination of block size and shape both at the headscarp and in the debris field, and field estimation of rock strength and weathering. I determined the Geological Strength Index (GSI) of rock masses in the headscarp, in remnant dolomitic marble crops out adjacent to the slide, and across the valley using the standard charts of Hoek et al. (1995) and Marinos and Hoek (2000). The schist units are foliated, but are not fully controlled by these planes of weakness; hence application of the GSI was deemed appropriate. A similar metric, the Debris Strength Index (DSI), is proposed to characterise blocks in the debris based on block size and joint condition, and to quantify the differences between debris and source rock masses.

Using observations collected in the field, I separated the study area into two structural domains (I and II), corresponding to the schist and marble units. I did not observe any regional-scale folding in at outcrops that would warrant further division of...
the domains. I identified three discontinuity sets within the schist domain, including foliation. The most significant sets in terms of current instability and movement are the foliation and discontinuity set (DS) 1, which form obvious wedges in the headscarp (Figure 5.19); these are also the most persistent discontinuities. Discontinuity persistence mapped along the scanlines is less than 3 m, reflecting the blocky, highly fractured nature of the rock mass. Spacing is below 2 m for all discontinuities mapped. Most foliation planes are smooth and planar to undulating, with some slickensided, rough, wavy, and stepped surfaces. The foliation is also crenulated on the centimetre-scale. Most DS1 planes are smooth and planar to stepped, and DS2 surfaces are rough and planar to stepped. I found four discontinuity sets in the marble unit, all with persistence values less than 3 m and spacing less than 2 m. The marble is highly fractured and discontinuity measurements are scattered.

Figure 5.19. Example of wedges formed by the foliation and DS1 on the west side of the headscarp of the Madison Canyon Slide.
I estimated block size and shape at the headscarp, in the debris field, and outside the landslide area. I calculated block size by estimating the three axes of the largest, smallest, and average-sized blocks at each field station, and classified block shape based on ISRM (1978) guidelines. Average block sizes in the in-situ and transported schist are approximately 9,950 and 11,980 cm$^3$, respectively; the average in-situ marble block is 9,000 cm$^3$, whereas the average transported marble block is 19,500 cm$^3$. Transported schist and marble blocks are generally larger north of Madison River, suggesting that larger blocks were transported on a matrix of finer material at the deposit front. Secondary failures may also have transported finer material from the headscarp to the middle of the deposit south of the river. The larger block size in the debris confirms the importance of weathering. Surficial outcrops are more fractured and weathered than subsurface exposures. Tabular blocks dominate the foliated metamorphic material, whereas marble blocks are more irregular or equidimensional. The schist blocks would be classified as platy or elongated-platy, and the marble blocks as cubic or cubic-elongated, in the block shape system of Kalenchuk et al. (2006).

I made field estimates of rock strength using a Schmidt Hammer and the guidelines of Hoek and Brown (1997) and British Standards (1981). Most in-situ rock masses have unconfined compressive strengths (UCS) of 100 MPa or greater. The weakest rock mass, a schist unit outside the landslide area, has a UCS of 35-50 MPa.

Both the marble and schist are highly fractured and weathered. Weathering grade, as classified according to the scheme of the Geological Society Engineering Group Working Party (1977), ranges from II to V, or slightly weathered to completely weathered, with most material moderately to highly weathered.

GSI values range from 20 to 80. The marble outcrops have GSI values higher than 50 and an average value of 63, whereas the schist is considerably weaker, with an average GSI value of 39 and a minimum range of 20-30 (Figure 5.20). One 50-cm thick shear zone comprised of weak, soil-like material was found in a rock mass with an average GSI of 45-55. The average DSI value for the schist debris is 46, with a range of 10 to 75, and that of the marble debris is 56, with a range of 55 to 70 (Figure 5.21). The DSI values are comparable to the GSI values, but the ranges for the schist and marble are slightly larger, indicating that the debris may include a more representative range of
rock mass characteristics, from highly weathered surficial material seen in outcrops to less weathered material within the slope mass.

Figure 5.20. GSI ranges for a) a weak schist sequence at the E headscarp and b) the remnant dolomitic marble outcrop at the base of the slope. Grey polygon in the inset = Madison Canyon Slide, red curve = headscarp, blue curve = Madison River.
Figure 5.21. Histogram of DSI estimates for the schist and marble debris.

**Long-range Terrestrial Digital Photogrammetry**

To gather more information on the rock masses involved in the Madison Canyon Slide, I completed the first terrestrial digital photogrammetry projects of the headscarp and surrounding area, the outcropping marble at the base of the slope, and the landslide deposit. I employed the methodology of Sturzenegger (2010) to create and process the photogrammetry projects. The field stations used for each photomodel are shown in Figure 5.22, and the models are summarised in Table 5.1. I used existing satellite imagery and field measurements of planes to georeference the models. The georeferencing does not affect relative accuracies of the models. Figure 5.23 shows an example of the headscarp photomodels (model 6), and Figure 5.24 shows the remnant marble buttress at the base of the slope (model 9).
Figure 5.22. Field locations (red points) for the photogrammetry projects. White rectangles are areas imaged, and numbers correspond to projects.
Table 5.1. Details of the photogrammetry projects analysed in this study. Model numbers correspond to those in Figure 5.22. Accuracy is relative, not absolute.

<table>
<thead>
<tr>
<th>Model number</th>
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<th>Ground pixel size (cm)</th>
<th>Plan accuracy (cm)</th>
<th>Depth accuracy (cm)</th>
<th>Overall accuracy (cm)</th>
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<td>0.1</td>
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</table>

Figure 5.23. Focal length lens $f = 200$ mm photogrammetric model of the Madison Slide headscarp (model 6).
The three-dimensional photomodels of the outcropping schist and marble units allowed me to expand my discontinuity mapping into otherwise inaccessible areas. I determined an additional two discontinuity sets (DS 3 and 4) in the headscarp area, and another four (DS 5*, 6*, 7*, and 8*) in the marble (Table 5.2). The photogrammetry foliation sets have higher dispersion coefficients than the field measurements (K = 55
versus $K = 49$), indicating slightly less dispersion in the photogrammetric foliation set orientations. Photogrammetry discontinuities also have much higher persistence than those measured in the field, with maximum persistence lengths of 26 m. The scale effect on persistence measurements is directly correlated with camera focal length in this project. In the field, persistence of individual discontinuities was rarely recorded above 3 m. As the scale of observation increased, higher persistence features were mapped. For example, in model 5, an $f = 50$ mm model, the average persistence measured is 10 m. In model 6, an $f = 200$ mm model, the average persistence is 5 m. Although planes with a persistence of $> 20$ m were mapped in both models, the minimum persistence measured in model 5 is 3 m and in model 6 it is $< 1$ m (Figure 5.25).

Table 5.2. Summary of the discontinuity sets (DS) identified in the field and on photogrammetry models. Orientations represent averages from both sets of data. Asterisks indicate DS associated with the marble.

<table>
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<td>24</td>
</tr>
<tr>
<td>8*</td>
<td>89</td>
<td>304</td>
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</table>
A photomodel of a section through the debris near the east margin of the landslide shows the stratigraphy of the deposit in that area (Figure 5.26a). The debris in the section is more coherent than the rubbly-blocky debris at the west side of the slide (Figure 5.26b), supporting the interpretation based on air photograph analysis that large blocks in this area rode on a layer of fragmented material at depth.
Figure 5.26. a) Focal length lens $f = 50$ mm model of the section eroded by Madison River near the east end of the debris sheet (model 2 in Figure 5.22), showing displaced blocks of intact rock separated by foliation planes (yellow lines). b) Photograph of the more fragmented debris under the remnant marble buttress at the west end of the landslide.

**Laboratory Testing**

Laboratory methods used to characterise the rock masses involved in the Madison Canyon Slide included thin section descriptions, slake durability, and point load tests. Detailed thin section descriptions are found in Appendix C. The laboratory tests provide important information on the character of the materials that failed. Field
observations suggest a dry climate, and support Hadley’s (1978) statement that water was not a significant factor in the 1959 failure, justifying the exclusion of groundwater in numerical simulations.

Although the material involved in the failure is highly weathered and fractured, slake durability tests indicate that it does not fragment easily. All four analysed samples are classified as I “virtually unchanged” or II “retained specimens consist of large and small fragments”, according to ISRM (1977) standards, indicating little slaking (see Appendix D for photographs of samples).

Given the highly fractured nature of the surficial material at the site, it was difficult to collect representative samples for point load tests. Fifty-three samples were tested, of which 23 were schist and 30 were marble. Figure 5.27 shows the Uniaxial Compression Strength (UCS) calculated from the Corrected Point Load Index ($I_{s(50)}$) results. The UCS values were derived from the relationship $UCS = (K)I_{s(50)}$, where $K$ is the conversion factor and was between 22 and 23 for my analyses. The UCS strength values of the marble and schist samples range, respectively, from 4 to 180 MPa and 6 to 188 MPa. Most values (75%) are lower than 100 MPa; an outlier at 307 MPa is excluded from Figure 5.27. Low strength values are attributable to the highly weathered and fractured nature of both rock types. Weathering appears to affect the schist more than the marble; low UCS values in the schist are associated with weathering grades of III-IV (moderately to highly weathered), whereas higher UCS values correspond to weathering grades of II-III (slightly to moderately weathered) (Figure 5.27). The marble samples do not show such a clear relationship with weathering. Most samples are slightly weathered (II), but four are moderately weathered (III). Of the four moderately weathered samples, three have UCS values less than 100 MPa, and one has a UCS of 130 MPa. Eight of the 21 schist samples and 12 of the 29 marble samples shown in Figure 5.27 failed along pre-existing planes of weakness, but passed ISRM (1978) validity requirements. The highest UCS values represent intact rock strength. Anisotropy was considered during testing, but the results were not statistically different based on simple t-tests. Hence, the foliation of the schist does not seem to be significant in strength considerations, at least at the laboratory scale. The results of the point load tests, in combination with values reported in the literature (RocData (Rocscience, 2013b); Bell, 1994), guided my choice of numerical modelling property values.
Figure 5.27. Point load test results for 50 samples of marble and schist. Roman numerals represent weathering grade determined in the field. Of the 53 samples tested, two were rejected and one was an outlier.

5.5. Stability Analyses

5.5.1. Kinematic Analysis

The Madison Canyon Slide is, at the global scale, a hexahedral wedge defined by four major composite surfaces – the headscarp, the E and W lateral scarps, and the basal plane of the biplanar sliding surface – and the slope face and upper slope. I performed kinematic analyses on this wedge in Swedge v. 6.0 (Rocscience, 2014b) based on field mapping, slope and aspect maps, 2D cross-sections of the slide, and field and photogrammetric discontinuity surveys. I determined optimal orientations for the four major planes and the slope face, as shown in Figure 5.28 and listed in Table 5.3, and I conducted sensitivity analyses on the friction angles for each plane to determine critical values (i.e., when the factor of safety = 1.0). Finally, I incorporated seismic loading was
incorporated into the Swedge analysis to observe its effect on stability. I assumed a unit weight of 27 kN/m³ and the default value of zero cohesion for all analyses.

![Hexahedral wedge geometry of the Madison Canyon Slide in Swedge](image)

**Figure 5.28.** Hexahedral wedge geometry of the Madison Canyon Slide in Swedge. The seismic coefficient for this iteration is 0.4 and the critical friction angle of the basal plane is 30°.

**Table 5.3.** Optimal orientations and properties of the major planes involved in the Madison Canyon Slide and the slope face used in Swedge analyses.

<table>
<thead>
<tr>
<th>Structure</th>
<th>Dip (°)</th>
<th>Dip direction (°)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Slope face</td>
<td>45</td>
<td>010</td>
</tr>
<tr>
<td>Upper slope</td>
<td>15</td>
<td>010</td>
</tr>
<tr>
<td>Headscarp</td>
<td>60</td>
<td>010</td>
</tr>
<tr>
<td>E lateral scarp</td>
<td>65</td>
<td>70</td>
</tr>
<tr>
<td>W lateral scarp</td>
<td>65</td>
<td>315</td>
</tr>
<tr>
<td>Basal surface</td>
<td>15</td>
<td>010</td>
</tr>
</tbody>
</table>
Assuming a slope height of 400 m, a slope length of 500 m, and an upper face width of 100 m, Swedge determines a wedge volume of 20.8 million m$^3$, very close to Hadley’s (1964) initial estimate of 20 million m$^3$. The required persistence of each lateral scarp is roughly 700 m, which agrees with field measurements of lateral scarp length. The sliding direction follows the basal plane, with a trend of 010° and plunge of 15°, again in agreement with field and remote sensing observations.

Given the geometry determined from field observations, the wedge is stable (FS > 1) under static conditions when the basal plane friction angle is greater than 5°. This is a very low value and explains why the slope was relatively stable prior to the 1959 earthquake. The friction angles of the lateral scarps do not affect stability.

When seismic loading is applied to the slope, the critical friction angle required for failure increases and stability decreases. Using the accepted range of seismic coefficient values (0.1 to 0.5; Jibson, 2011), I determined a linear relationship between seismic coefficient and friction angle, suggesting a strong effect of seismicity on the stability of the slope.
5.5.2. Landscape Evolution in 2D Finite Element Simulations

I used Phase2 v. 8.0, an elasto-plastic 2D finite element code (Rocscience, 2014a), to examine the effects of landscape evolution on the stress distribution in the Madison slope for profiles B, C, and D in Figure 5.10. For most analyses, I assumed a horizontal/vertical stress ratio (K) of 1, but I also conducted a parameter study, changing K from 0.5 to 3.3. These values represent a spectrum of normal (low values) to thrust (high values) faulting conditions, as Leith (2012) illustrated. Rock mass and discontinuity properties are based on maximum Geological Strength Index (GSI) values determined for the schist and marble units in the field, as well as values in the literature (RocData v. 4.0, Rocscience, 2013b; Table 5.4 and 5.6). The modulus ratios (MR) are within the ranges for schists and marbles. Rock masses 1 and 2 represent less weathered, less fractured material at depth, with properties based on higher GSI values than surface
values. Each cross-section slope was first excavated in four stages in a procedure similar to that of Leith (2012), but using a different constitutive criterion (Mohr-Coulomb versus trilinear; Diederichs, 2003). After each excavation, the upper remaining material was changed from a Mohr-Coulomb elastic (rock mass 1) to a Mohr-Coulomb plastic (rock mass 2) material to allow plastic deformation (Figure 5.30). Excavation of material in the models simulated the combined effects of tectonic uplift and fluvial erosion on the topography of the valley where the landslide occurred. Then, to simulate weathering, I degraded the near-surface rock mass again after the final excavation from the rock mass 2 to the rock mass 3 properties in Table 5.4. Finally, the sliding surface properties were reduced and a Shear Strength Reduction (SSR) analysis completed.

**Table 5.4.** Rock mass properties used in the Phase2 simulations.

<table>
<thead>
<tr>
<th>Property</th>
<th>Rock mass 1</th>
<th>Rock mass 2</th>
<th>Rock mass 3</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Schist</td>
<td>Marble</td>
<td>Schist</td>
</tr>
<tr>
<td>Unit weight (kN/m³)</td>
<td>27</td>
<td>27</td>
<td>27</td>
</tr>
<tr>
<td>GSI</td>
<td>75</td>
<td>95</td>
<td>55</td>
</tr>
<tr>
<td>UCS (MPa)</td>
<td>250</td>
<td>300</td>
<td>100</td>
</tr>
<tr>
<td>MR</td>
<td>1100</td>
<td>500</td>
<td>1100</td>
</tr>
<tr>
<td>Poisson's ratio</td>
<td>0.3</td>
<td>0.3</td>
<td>0.3</td>
</tr>
<tr>
<td>Tensile strength (MPa)</td>
<td>n/a</td>
<td>n/a</td>
<td>0.3</td>
</tr>
<tr>
<td>Friction angle (°)</td>
<td>n/a</td>
<td>n/a</td>
<td>43</td>
</tr>
<tr>
<td>Cohesion (MPa)</td>
<td>n/a</td>
<td>n/a</td>
<td>2.4</td>
</tr>
</tbody>
</table>

**Table 5.5.** Discontinuity properties used in the Phase2 simulations.

<table>
<thead>
<tr>
<th>Property</th>
<th>Joints</th>
</tr>
</thead>
<tbody>
<tr>
<td>Tensile strength (MPa)</td>
<td>0</td>
</tr>
<tr>
<td>Normal stiffness (GPa/m)</td>
<td>20</td>
</tr>
<tr>
<td>Shear stiffness (GPa/m)</td>
<td>2</td>
</tr>
<tr>
<td>Friction angle (°)</td>
<td>10-35</td>
</tr>
<tr>
<td>Cohesion (MPa)</td>
<td>0</td>
</tr>
</tbody>
</table>
Maximum ($\sigma_1$) and minimum ($\sigma_3$) principal stress values ranged from 0 MPa at the surface to 43 MPa at the base of the $K = 1$ models. As excavation progressed, maximum principal stresses concentrated on the floor of the Madison River valley, with $\sigma_1$ values as high as 25 MPa (Figure 5.31a). In the profile C and D simulations, the contacts between the schist and marble units were also locations of stress concentrations. The marble unit concentrated maximum shear strain.
Figure 5.31. Examples of the landscape evolution simulation results for profile C. a) Elastic maximum principal stress ($\sigma_1$) results for stage 8, showing maximum principal stress concentrations at the bottom of the Madison River valley. The subvertical green lines represent the contacts between the schist and marble. b) Plastic results for stage 11, showing $\sigma_1$ contours, principal stress tensors, and elements yielding in tension.

Changing the stress ratio from 0.5 (normal faulting conditions) to 3.3 (reverse faulting conditions) affects stresses in the slope differently. The results of the $K = 0.5$ simulations did not differ significantly from the $K = 1$ results, as Hoek et al. (2011) discussed. High horizontal stresses in the $K = 3.3$ models, however, caused high shear strains in the slope. As the Madison Canyon Slide occurred in a region dominated by active normal faulting, the $K = 0.5$ and $K = 1$ results are hypothesised to be most appropriate.
Degradation of the properties of the upper layers in the models focused shear strain in the weaker materials. Shear yield was concentrated in the valley bottom, whereas tensile yield was focussed at ridge crests, agreeing with Leith (2012). For all three cross-sections analysed, extensional yielded elements appear mainly at the ridge crest of the failed slope (Figure 5.31b). The extensional yielded elements align well with tension cracks observed in the field and on the pre-1959 air photographs. Section D showed additional extension at both contacts between the schist and marble. Yield in section B suggested extension at each change in gradient along the sliding surface (Figure 5.32). Shear also occurred near the sliding surface in this simulation. Section B thus indicates the development of active-passive block transition zones that are absent in profiles C and D. The angles of the upper and lower secondary shear planes are 80° and 105° from horizontal, respectively.

**Figure 5.32.** Maximum shear strain and yield plot for profile B in plastic stage 11, showing two areas of strain related to changes in sliding surface gradient, with orientations from horizontal as indicated.

Shear Strength Reduction (SSR) analysis in Phase2 indicates that, provided a slip surface with no cohesion existed in the slope, the slope could have failed statically without seismic loading (Table 5.6). The critical friction angle for each profile was determined by a parametric study of the friction angle. Profile B is more stable than profiles C and D, as it requires the largest reduction in friction angle to reach failure (from 35° to 16°). The difference in stability across the slope is likely related to two
factors: i) the changing slope and sliding surface geometry, and ii) the changing extent and location of the marble buttress. Profile B has a gentler, straighter slope and a less inclined sliding surface than the other two sections. The lower half of the slope in profile D is dominated by marble, whereas marble is lower on the slope in profile C. Hence, profile D requires more reduction in friction angle to fail than profile C, as the marble stabilises the slope.

Table 5.6. Shear Strength Reduction (SSR) results for profiles B, C, and D in Madison Canyon. The Strength Reduction Factor (SRF) is equivalent to the factor of safety.

<table>
<thead>
<tr>
<th>Profile</th>
<th>Critical friction angle (°)</th>
<th>Critical SRF</th>
</tr>
</thead>
<tbody>
<tr>
<td>B</td>
<td>16</td>
<td>1.0</td>
</tr>
<tr>
<td>C</td>
<td>23</td>
<td>1.0</td>
</tr>
<tr>
<td>D</td>
<td>20</td>
<td>1.0</td>
</tr>
</tbody>
</table>

5.5.3. Influence of Discontinuities and Sliding Surface Properties on Kinematics in 2D Distinct Element Simulations

I used elastic UDEC v. 5.0 (Universal Distinct Element Code; Itasca, 2013c) models to investigate the effects of discontinuity orientation and spacing and sliding surface properties on the Madison Canyon slope, and to conduct sensitivity analyses on friction angle and cohesion along these discontinuities. I applied geometry based on profiles C and E, and schist and marble material properties based on accepted values in the literature (Bell, 1994; RocData (Rocscience, 2013b); Table 5.7 and 5.9). Zero velocity boundary conditions were applied to all models. For the discontinuity model set, I investigated three subsets of models (Figure 5.33): i) two discontinuity sets (DS) + schist, ii) two DS + schist + two faults representing the contacts between the schist and marble, and iii) two DS + schist + two faults + marble. The dip of the first of the two discontinuity sets included in the models, representing foliation, changed in orientation from dipping out of the slope to vertical to dipping into the slope, and the second subhorizontal set, acting as basal release surfaces, dipped gently out of the slope. In parametric studies on the friction angle and cohesion of the joints, I changed their values, respectively, from 30° to 5° and 100 kPa to 0 kPa.
Table 5.7. Material properties used in the static UDEC simulations.

<table>
<thead>
<tr>
<th>Property</th>
<th>Schist</th>
<th>Marble</th>
</tr>
</thead>
<tbody>
<tr>
<td>Density ($\text{kg/m}^3$)</td>
<td>2700</td>
<td>2700</td>
</tr>
<tr>
<td>Bulk modulus (GPa)</td>
<td>16.7</td>
<td>30.0</td>
</tr>
<tr>
<td>Shear modulus (GPa)</td>
<td>10.0</td>
<td>20.0</td>
</tr>
</tbody>
</table>

Table 5.8. Joint properties used in the static UDEC simulations.

<table>
<thead>
<tr>
<th>Property</th>
<th>Joints</th>
</tr>
</thead>
<tbody>
<tr>
<td>Tensile strength (MPa)</td>
<td>0</td>
</tr>
<tr>
<td>Normal stiffness (GPa/m)</td>
<td>20</td>
</tr>
<tr>
<td>Shear stiffness (GPa/m)</td>
<td>2</td>
</tr>
<tr>
<td>Friction angle (°)</td>
<td>5-30</td>
</tr>
<tr>
<td>Cohesion (kPa)</td>
<td>0-100</td>
</tr>
</tbody>
</table>

Several observations of the DS model subsets highlight the kinematics at Madison. First, changing the orientation of the foliation altered the failure mechanism from sliding (DS out of slope) to toppling (DS into slope) (Figure 5.34). Hence, the modelling suggests that the actual orientation of the foliation (50°/342°, dip/dip direction) favours sliding. Second, when the faults were included in the models, displacements increased, as additional release surfaces existed. This effect is particularly apparent in the profile E simulations, as the faults propagated to mid-slope, rather than to the base of the slope, as in profile C. The faults also allowed the base of the profile E slope to fail first, without much interaction with the upper slope blocks. The addition of marble material between the two faults stabilised the slope, as it has a higher rock mass quality. Third, the critical friction angle is 15° when cohesion = 0 MPa in all models. Significant displacement (> 1 m) only occurred with friction angles of 5° and no cohesion. These low values, even for fully persistent structures, suggest that most of the slope was stable prior to the 1959 earthquake.
Figure 5.33. Geometries analysed in the UDEC static discontinuity model set. a) Profile C with two discontinuity sets (DS), two faults, and the schist and marble; the foliation DS is oriented for toppling. b) Profile E with two DS and the schist; the foliation DS is oriented for sliding.

Only the biplanar sliding surface and contacts between the marble and schist were included in the second set of static simulations. In profile C, the models indicate the development of an active-passive failure, with a transition zone at the break in the sliding surface slope (Figure 5.35; Mencl, 1966). Conversely, profile E shows the marble buttress failing first, followed by the weaker schist (Figure 5.36). High horizontal and vertical stresses develop in the zone surrounding the south contact between the schist and marble – horizontal stresses reach 3 MPa and vertical stresses 12 MPa. The marble, which dominates the lower slope in profile E, must fail before creating the kinematic freedom for the schist to slide.
Figure 5.34. Block rotation plot of a toppling simulation in UDEC for the profile E geometry. Rotations are shown in green at block centres. Maximum rotation displacement is 32 m.

Figure 5.35. XX stress contour plot of a profile C sliding surface model. High stresses develop in the Prandtl wedge transition zone (red outline) between the active and passive blocks.
5.5.4. Effects of Seismicity in 2D Distinct Element Simulations

I studied three geometries – straight, concave, and convex and based on profile C – in UDEC (Itasca, 2013c) to investigate the effects of morphology on seismic wave propagation (Figure 5.33; Table 5.10 and Table 5.11). Simulations were run using elastic, one-material, unjointed and jointed models and elastic, two-material jointed models with the assumed failure surface (based on Figure 5.10 and Figure 5.11) and tension cracks (based on geomorphological observations) included. I completed static runs with zero velocity boundaries to achieve initial equilibrium before each dynamic run, whereupon I changed boundary conditions to free field and viscous boundaries at, respectively, the lateral and basal borders of each model, to prevent reflection of waves at artificial boundaries. In the preliminary simulations of seismic waves, I applied Ricker wavelets with dominant frequencies ranging from 0.1 to 10 Hz converted to transient stress boundary conditions at the base of each model; more advanced simulations incorporated a record from the 2001 Nisqually, WA, USA, earthquake as a natural earthquake record. The methodology used in the jointed models is similar to that of Moore et al. (2011, 2012): I used elastic material properties and high initial joint properties to create a base case for comparison to simulations with degraded discontinuity properties.
Figure 5.37. a) Straight, b) concave, and c) convex geometries analysed in UDEC dynamic simulations. I also analysed the convex geometry by including the marble unit, the sliding surface, and tension cracks (TCs) at the ridge crest. Boundary conditions for the dynamic analyses are: ff – free field, visc – viscous (quiet), and transient stress boundaries.
Table 5.9. Material properties used for the elastic dynamic UDEC simulations.

<table>
<thead>
<tr>
<th>Property</th>
<th>Schist</th>
<th>Marble</th>
</tr>
</thead>
<tbody>
<tr>
<td>Density (kg/m³)</td>
<td>2700</td>
<td>2700</td>
</tr>
<tr>
<td>Bulk modulus (GPa)</td>
<td>11</td>
<td>48</td>
</tr>
<tr>
<td>Shear modulus (GPa)</td>
<td>6</td>
<td>22</td>
</tr>
</tbody>
</table>

Table 5.10. Discontinuity properties used for the elastic dynamic UDEC simulations. High properties are equivalent to intact rock, and low properties open fractures.

<table>
<thead>
<tr>
<th>Property</th>
<th>Joints</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>High</td>
</tr>
<tr>
<td>Tensile strength (MPa)</td>
<td>17</td>
</tr>
<tr>
<td>Normal stiffness (GPa/m)</td>
<td>100</td>
</tr>
<tr>
<td>Shear stiffness(GPa/m)</td>
<td>10</td>
</tr>
<tr>
<td>Friction angle (°)</td>
<td>40</td>
</tr>
<tr>
<td>Cohesion (MPa)</td>
<td>50</td>
</tr>
</tbody>
</table>

The unjointed models indicate that the highest topographic amplification due to interaction of the model and applied Ricker wavelets occurred at the ridge crest, regardless of slope geometry (Figure 5.38). The first peak in Figure 5.38 is interpreted as the eigenmode of the ridge in each system (straight, concave, and convex). The second peak in the figure relates to the focussing effect of the ridge compared to the left (south) and right (north) valley bottoms, respectively. The observed amplification pattern is the result of complex wave interaction that includes wave focussing as well as interference between incident, reflected, and Rayleigh waves generated at the inclined slope surface. Between 2 and 4 Hz, the convex geometry has the highest amplification, roughly 2.8, and the concave geometry has the lowest amplification, approximately 1.9. This result confirms the findings of Meunier et al. (2008), who determined that seismically triggered landslides initiate most often at the crests of ridges and at convexitities in slopes. The amplification factors agree well with the results of Moore et al. (2011, 2012) and Damjanac et al. (2013) for rock slopes and open pits. Figure 5.39 provides an example of the reaction of the unjointed model to seismic input analogous to the 2001 Nisqually earthquake (Mw 6.9). The ridge-crest x velocity is consistently at least
0.5 times to 2.5 times higher than those at the left (south) and right (north) valley bottoms until the motion is damped. I determined similar trends for the concave and straight geometries.

**Figure 5.38.** Dominant frequency versus amplification for the averaged Ricker wavelets (0.1 to 10 Hz) at a) the ridge crest versus the left (south) valley bottom and b) the ridge crest versus the right (north) valley bottom, applied to the straight, concave, and convex geometries.
Preliminary results for the jointed models show the effect of discontinuities on the models. Inclusion of discontinuities suggests failure after an input motion is applied (Figure 5.40). X velocity histories show that discontinuities decrease the amplification effect of the topography, but also create a distributed velocity peak that remains higher at the ridge crest than at the valley bottoms in each model. Hence, discontinuities absorb some of the seismic energy while creating a lower but sustained x velocity peak (Figures 5.45 and 5.46, Table 5.11). The orientations of the discontinuities seem to affect the x...
velocity peak. For example, including only the marble contacts reduces the peak x
velocity significantly (from 3.6 in the unjointed model to 2.29), but does not affect the
time interval over which the central Ricker peak occurs at the ridge crest. Conversely,
the sliding surface discontinuity does not affect the peak velocity, but doubles the time
interval. Including all discontinuities significantly reduces the peak and increases the
time interval.

Figure 5.40. Elastic displacement plot of a jointed model including the biplanar failure
surface and tension cracks at the ridge crest with a Ricker wavelet of 5 Hz
applied at the base.
Figure 5.45. Investigation of the effect of discontinuities on the seismic input in UDEC. a) Geometry of the section analysed and the discontinuities included. The colours of the three history points (monitoring points for data) correspond to the records presented in b), x velocity histories for a Ricker wavelet of 5 Hz. I monitored the peak and the time interval of the central peak of the Ricker wavelet input at the ridge crest to compare the models.
5.6. Discussion and Conclusion

5.6.1. Endogenic and Exogenic Processes

The south slope of Madison Canyon was conditioned for failure by endogenic and exogenic factors including tectonic uplift of the Madison Range, previous earthquakes, erosion of the valley and undercutting of the slope by Madison River, gullyng and other erosional processes on the slope, and weathering. The final trigger for
catastrophic failure was the 1959 earthquake. The effects of past earthquakes are apparent in offset fluvial terraces and exposed fault scarps. Madison River cuts through the Madison Range, keeping pace with tectonic uplift. Although glaciers did not directly affect the slopes in the slide area, their increased sediment loads affected Madison River, and thus its capacity to erode the canyon walls in the slide area.

These processes, in combination with rock mass fatigue and damage, have progressively weakened the slopes in Madison Canyon and set the stage for the 1959 failure. Their interaction over-stressed the slope, causing microscopic and macroscopic cracking of the rock masses and degrading their strength, as Leith (2012) suggested for Alpine valleys. My landscape evolution modelling indicates that extension occurs at ridge crests and compression at the valley bottom. In a reverse faulting regime, the simulated slopes yield significantly more than under normal faulting conditions. The surface geometry of slopes and sliding planes also affect stress distributions and instability. The straighter, gentler slope of profile B was more stable than the convex, steeper slopes of profiles C and D. Extensional and shear yielding also was focussed at changes in the gradient of the hypothesised failure surface, rather than being more distributed in the upper slope. For all profiles, extension at the ridge crest aligns well with tension cracks observed in the field. Hence, the interaction between in situ stresses and exhumation and erosion of the landscape was significant in the development of the Madison Canyon Slide.

The role of exogenic processes in conditioning the Madison Canyon slope for failure is clear when considering the location of the landslide. It occurred on a ridge at the west front of the Madison Range, in one of the narrowest reaches of the valley and at the most significant knickpoint along the channel. River incision and undercutting of the south slope of Madison Canyon were important factors in undermining the strength of the slope toe. Bigi et al. (2006) investigated the relation between knickpoint migration and landsliding, and suggest that landslides occur downstream of knickpoints.

Rock mass damage due to the processes mentioned above is recognisable from morphology, differing GSI values, and block size. The area of highest tectonic and gravitational damage is the ridge crest, as indicated by the clusters of tension cracks, highly fractured rock, and thin shear zones observed in the field. The contact between
the schist and marble probably concentrated tectonic damage and contributed to the west lateral release surface of the Madison Canyon Slide. Perhaps most importantly, the site of the failure was one where seismic energy was focussed, as revealed by the topographic amplification in the dynamic models.

Effects of Seismicity on Slope Stability

As mentioned, seismicity is an important component of the history of the southern Madison Canyon slope. The 1959 earthquake triggered the catastrophic failure, allowing coalescence of discontinuities to form continuous sliding surfaces and zones and fracturing the resistant marble buttress at the toe of the slope. However, other earthquakes also damaged and weakened rock masses, possibly initiating extension at the ridge crest. Long-term seismic conditioning resulted in cumulative fatigue damage both in the rock mass and along discontinuities.

Through numerical simulations, I determined that the interaction of the slope with seismic waves is critical to slope stability. Two-dimensional dynamic deformation analyses using UDEC suggested an explanation for why the Madison Canyon Slide occurred on the south slope of Madison Canyon. Two aspects of the slope contribute to its uniqueness in the area: i) slope geometry (leading to topographic amplification), and ii) pre-existing tension cracks and fractures (leading to structural amplification) (Figure 5.41). The convex slope is located 30 km almost directly west of the 1959 earthquake epicenter and faces NNW. The ridge changes orientation from a NE-SW trend to a WNW-ESE trend at the location of the Madison Canyon Slide headscarp, and then tapers to meet the wide, open Missouri Flats to the west. These geometrical characteristics amplified seismic shaking, focussing energy on the slope, and thereby exacerbating the effects of the earthquake at this location.

The pre-existing tension cracks, which may have been part of a sackung, also altered seismic energy in the slope. In an hypothesised positive feedback system, previous earthquakes might have opened fractures at the ridge crest, which then focussed the 1959 seismic energy, contributing to the catastrophic failure of the slope. A similar hypothesis has been proposed for the Randa and Rawilhorn slopes (Moore et al., 2011, 2012).
5.6.2. Chronology of Events at Madison Canyon

From the geomorphological and structural evidence, a potential chronology may be constructed as follows (Figure 5.42):

I. Due perhaps to past seismicity, tension cracks appeared in the south wall of Madison Canyon, indicating incipient instability. These tension cracks formed in the schist-gneiss units, highlighting their low strength relative to the more resistant marble. They also signify the slow fatigue of the slope over time.

II. The 1959 $M_w = 7.5$ Hebgen Lake earthquake triggered the catastrophic Madison Canyon Slide. The rock mass on the west side of the slope appears to have failed first, probably at the base of the marble unit that had formed a buttress on the lower part of the slope. Fragmenting marble acted as a carpet on which the much larger mass of schist moved. The schistose debris also pushed the marble debris ahead to the distal limit of the landslide north of Madison River. The ribbon of marble debris that marks the outermost portion of the debris sheet was deformed by, and in places turned back into, the schist debris trailing it.
III. Shortly after the marble ribbon came to rest, the more coherent large blocks to the east, riding on highly comminuted debris, compressed as they encountered resistance from the rising slope north of Madison River and then expanded into the valley. These blocks still have trees on them, although most were toppled during transport. Late during the landslide, or possibly even later, a large secondary failure occurred from the headscarp on the east side of the slide. This slump regressed to the crest of the ridge. Several smaller rockfalls, slumps, and slides occurred in the days, months, and years following the main slide, particularly along the western lateral margin and below the headscarp. The catastrophic landslide dammed Madison River and formed Earthquake Lake, significantly altering the canyon mouth.

VI. Shortly after the 1959 landslide, the U.S. Army Corps of Engineers constructed a channel through the slide debris to prevent collapse or overtopping of the landslide dam. The construction of the Earthquake Lake visitor’s centre further changed the deposit morphology. The deposit and headscarp have continued to evolve over the past 50 years, with processes such as rockfalls and slides transforming the slope. Deformation of the ridge crest continues, as evidenced by fresh tension cracks.

The above chronology highlights the complex evolution of the Madison Canyon Slide: in the west it behaved like a rock avalanche with highly fragmented debris; farther east, large coherent blocks were transported on a carpet of debris; a large slump followed the main event at the southeast corner of the landslide and several smaller failures occurred in response to the sudden change in the slope. The marble unit at the base of the slope, which had buttressed the weaker schist and gneiss above, failed during the 1959 earthquake. The highly fractured state of the remaining outcrops indicates damage and fatigue due to weathering, erosion, and previous seismicity that conditioned the slope for failure.
Figure 5.42. Hypothesised sequence of events culminating in the catastrophic 1959 Madison Canyon Slide.

The Madison Canyon Slide can be considered a hexahedral wedge comprising the headscarp, the east and west lateral scarps, a basal release surface, the slope face, and the upper slope. The plane that parallels the headscarp acted as a rear release surface and as a sliding surface. It is a composite surface, representing the average of smaller wedges created by the foliation, DS 1, and other discontinuities. The sliding surface likely formed from brittle failure through intact rock bridges and coalescence of pre-existing discontinuities such as the narrow foliation-parallel shear zone observed in the headscarp. Discontinuity roughness, persistence, and spacing would have controlled this coalescence. Roughness would have inhibited development of the sliding surface, and contributed to dilation in the rock mass once sliding began. The relatively low persistence and spacing of most discontinuities, and resulting small block sizes, partially explain the high degree of fragmentation in the debris.
Two-dimensional stability analyses highlight different aspects of the failure. Phase2 simulations suggest a static critical friction angle of 16° to 23°, depending on the cross-section analysed, and static UDEC simulations show the importance of discontinuity orientations for failure kinematics. Further evidence of seismic effects is provided by the dynamic UDEC simulations. The optimal geometry for seismic amplification is a convex slope, and ridge-crest amplification is consistently > 1.5. Tension cracks at the top of the slope also amplify seismic energy; models required only one seismic event to fail when the sliding surface was included.

5.6.3. Conclusions

I have applied a range of techniques to analyse the Madison Canyon Slide. I constructed the first long-range terrestrial photogrammetry models of the failure, conducted engineering geological fieldwork and characterised the rock masses involved in the landslide, created detailed field and air photograph engineering geomorphologic maps, and simulated the event with three numerical programmes.

The chapter contributes to an improved understanding of the Madison Canyon Slide as follows: 1) endogenic and exogenic processes, including tectonic uplift, fluvial erosion, and weathering, were significant individually and in combination in bringing the slope to failure; 2) the earthquake was an especially effective trigger; 3) the site of the landslide was subject to high topographic and discontinuity amplification, further increasing the efficacy of the earthquake as a trigger; 4) failure mechanisms include combined wedge-planar failure and secondary rotational failures; 5) a rapidly fragmenting rock mass streamed across Madison River, while largely intact blocks rode to the northeast on a carpet of finely crushed material; 6) marble debris derived from rock outcrops low on the slope rode in front of and beneath schist sourced higher on the slope; and 7) slumps, debris slides, and rockfalls subsequently occurred after the main slide. Future areas of research may include subsurface investigations, structural domain mapping, and three-dimensional numerical modelling.
Acknowledgements

I gratefully acknowledge the help of the U.S. Forest Service, and particularly Joanne Girvin. Diane Doser supplied seismogram records. Camille Christiansen and Jeanette Klassen assisted in the field. The U.S. Geological Survey provided imagery. Research was funded through an NSERC scholarship to A. Wolter and NSERC Discovery Grants to D. Stead and J.J. Clague.
Chapter 6.

Discussion and Conclusions

Throughout this dissertation, I have emphasised the complexities of the Vajont and Madison Canyon landslides in the context of their geology, geomorphology, mechanisms, and evolution. Here, I compare and contrast the two landslides, discuss implications of the research for rock slope instabilities in general, and highlight the advantages and disadvantages of each method used to analyse the two events. I conclude with research highlights and recommendations for future work.

6.1. Comparison of Two Large Catastrophic Rock Slope Failures

Both the Madison Canyon and Vajont failures are large catastrophic landslides. The mechanisms of the Madison Canyon Slide and west block of the Vajont Slide are similar. Both have active and passive regions separated by a Prandtl wedge transition zone. The Prandtl wedge zone is accentuated at Vajont by the geometry of the southern limb of the Erto Syncline. It is not expressed as clearly at Madison Canyon.

Kinematically, both events appear to be complex wedge-biplanar failures. At Vajont, the two limbs of the Massalezza Syncline form part of a large hexahedral wedge that directs movement to the NNE. The east-plunging southern limb of the Erto Syncline forms a basal release surface in the west, and the sliding surface steps up to the Col Tramontin Fault lateral release surface in the east (Hendron and Patton, 1985). The Col Tramontin Fault and east Massalezza Syncline form another asymmetrical wedge. The Madison Canyon Slide was a hexahedral wedge comprising the slope face and upper slope, and the composite surfaces of the headscarp at the ridge crest, lateral release surfaces, and basal plane formed by brittle fracture through intact rock bridges and pre-
existing discontinuities such as foliation planes. Gullies acted as lateral release surfaces in the east. Smaller wedges formed by the foliation and discontinuity sets probably contributed to the fragmentation of the debris. The principal movement direction of the slide was to the NNE. The contact between the marble and schist-gneiss may have influenced movement at the mid-slope position. Part of the Madison Canyon landslide behaved as a rock avalanche – highly fragmented debris streaming across Madison River, accompanied by development of multiple local shear planes within the failed rock mass.

The sliding zone at Vajont is controlled by major structural features, whereas the sliding surfaces at Madison Canyon comprise meso- and micro-scale features. The Vajont failure scar is dominated by bedding planes and is characterised by steps and complex zones of interference between two sets of tectonic folds. The catastrophic failure apparently was the final stage of slow, deep-seated gravitational slope deformation (sackung) operating over a period of many thousands of years during which the shear zone slowly formed and cracks coalesced, breaking through carbonate asperities to connect the weaker clay layers. Rapidly changing reservoir levels and high precipitation triggered catastrophic failure. At Madison Canyon, multiple shear zones at the base of and within the failing mass contributed to fragmentation of the debris. The basal shear zone appears to have been fairly planar, despite the existence of centimetre-scale crenulations in the foliation of the metamorphic units (Figure 5.11). No regional-scale folds or faults were observed at the site. The highly weathered nature of the rock masses suggests that, although discontinuities guided development of the failure surface, they were not as dominant in terms of kinematics as at Vajont. Rather, the sliding surfaces at Madison developed by coalescence of cracks and discontinuities, part following existing foliation and discontinuity planes, part fracturing through intact rock. The brittle fracture of the marble buttress was critical to the development of the large failure.

Both the Vajont and Madison Canyon slides involved secondary failures that occurred within seconds of the main events. At Vajont, the east block collapsed after the larger west block. It was weaker and slid on a circular-to-planar sliding surface, but was buttressed by the west block until the latter failed, thus providing kinematic freedom for the east block to move. Similarly, the upper east portion of the Madison Canyon Slide
seems to have failed after the west portion. Again, the material on the west provided kinematic freedom. The area of the secondary failure was unstable before the 1959 slide, as evidenced by tension cracks in pre-1959 photographs. It also seems to have been weaker than the rock to the west, as it failed by rotation on retrogressive circular sliding planes. Both the Vajont and Madison Canyon slides were the final stages of longer-term gravitational slope deformation, and in both cases secondary failures are responses of the slope to catastrophic failure, indicating the continuous evolution of the landscape and reaction to abrupt changes in slope equilibrium.

Rock mass damage at different scales played an important role in both failures. At Vajont, micro-scale damage accumulated along the basal shear zone as rock bridges progressively fractured. Due to its proximity to the Col Tramontin Fault, the east block was more damaged on the macro-scale and thus weaker than the west block. On the meso-scale, the west block was subject to deformation in the Prandtl wedge zone. Yet, most of the failed rock mass at Vajont remained relatively intact. Damage was concentrated in the basal shear zone and a few discrete secondary shears that separated landslide blocks. The Madison Canyon Slide was different in this aspect. The rock mass was weaker, foliated, more highly weathered, and very fractured prior to failure. The marble buttress was more resistant than the schist-gneiss until the 1959 earthquake destroyed its integrity. Unlike Vajont, part of the rock mass rapidly fragmented. Areas of the deposit include more intact blocks, but these do not compare to the huge blocks at Vajont. The difference can be explained by the clay layers at Vajont. These acted as a constrained weaker unit, concentrating most of the strain, so that the material above them slid intact downslope.

The contexts of the two case studies are complex (Table 6.1). Most of the factors that condition slopes for failure are found at both sites, with the exception of groundwater and glacial erosion at Madison and a seismic trigger at Vajont. However, as at Madison, Vajont Valley is located in an active tectonic environment, and strong earthquakes have certainly affected the Monte Toc slope numerous times in the past.

The stratigraphy at the two sites is different. Vajont is dominated by carbonate rocks. The most significant layers in terms of stability, however, are the smectite-rich clay layers. These layers have very low friction angles and focussed most of the strain in
The cataclasites that Ferri et al. (2011) and Semenza (2010) reported were likely the result of accumulation of strain and damage in these clay layers and the units surrounding them as the slope deformed gradually. The stratigraphy at Madison is dominated by two main units: a schist-gneiss sequence that I treated as one geotechnical unit and a more restricted, dolomitic marble unit. The marble served as a buttress for the weaker foliated rocks above until it failed in the 1959 earthquake. Although there are small shear zones in the headscarp, there do not appear to be anomalously weak units along the basal sliding zone. Thus, the damage related to lithology is more pervasive at Madison Canyon than at Vajont.

Table 6.1. Factors contributing to the complex contexts of the Vajont and Madison Canyon rock slope failures.

<table>
<thead>
<tr>
<th></th>
<th>Vajont</th>
<th>Madison Canyon</th>
</tr>
</thead>
<tbody>
<tr>
<td>Weak lithology</td>
<td>✓</td>
<td>✓</td>
</tr>
<tr>
<td>Structural controls</td>
<td>✓</td>
<td>✓</td>
</tr>
<tr>
<td>Steep slopes</td>
<td>✓</td>
<td>✓</td>
</tr>
<tr>
<td>High pore pressures</td>
<td>✓</td>
<td>×</td>
</tr>
<tr>
<td>Seismicity</td>
<td>×</td>
<td>✓</td>
</tr>
<tr>
<td>Glacial action</td>
<td>✓</td>
<td>×</td>
</tr>
<tr>
<td>Fluvial erosion</td>
<td>✓</td>
<td>✓</td>
</tr>
</tbody>
</table>

Structural factors were significant to failure at both sites, but particularly at Vajont. The Vajont Slide was constrained by two faults that formed the east lateral and rear release surfaces and two synclines that formed the basal release surface. These structures contributed to the three-dimensional complexity of the slide. The structural controls at Madison are not as apparent. Small wedges in the headscarp have been discussed, but there are no mapped regional structures bounding the failure. The lateral release surfaces are suspiciously parallel, and might be fault-controlled; their orientation is roughly consistent with the N-S trending regional stress regime. Gullies acted as lateral release surfaces in the east.

Both failures occurred on slopes ranging from 20° to 50°, within the common range for large landslides. The Madison Canyon Slide occurred on a convex slope, whereas Vajont failed on a straight to concave slope. Both slopes were vegetated; the
forest at Vajont, however, had been partially removed for agriculture long before the slide.

The hydrogeological conditions at the two sites were different. The Madison Canyon slope was reported dry when it failed (Hadley, 1978). In contrast, the Vajont Slide was triggered by changes in groundwater in the slope. Elevated pore pressures would have buoyed the sliding mass, facilitating failure along already weak shear zones. Other factors such as karst formation also may have affected rock mass strength at Vajont.

An earthquake triggered the Madison Canyon failure. The slope was optimally oriented for amplification of seismic energy, and tension cracks contributed to seismic amplification. The long-term effects of recurrent seismicity are less obvious, both at Vajont and Madison. Although the quantification of this effect is difficult, previous seismicity would have contributed to the gradual degradation of strength and increase of damage at each location.

The geomorphological heritage of Vajont includes glacial and fluvial erosion. Repeated Pleistocene glaciation broadened the valley, steepened slopes, and contributed to rock damage. The role of debuttressing is unclear. Several landslides in the valley, however, occurred soon after deglaciation, including the hypothesised initiation of the sackung on the north slope of Monte Toc. The erosion of the toe of the slope by Vajont River concentrated stress in the epigenetic Vajont Gorge. Glaciers did not reach the site of the Madison Canyon Slide; however, large sediment loads from glaciers upstream would have been carried past the slope, altering river gradient and energy. An alluvial fan forced Madison River against the south slope of the canyon at the site of the failure, undercutting the marble buttress. The final triggers of failure at Vajont and Madison Canyon, although different, were both short-lived and created disequilibrium in the two slopes.
6.2. Implications of Research

Both the Vajont and Madison Canyon failures began as sackung-type, slowly deforming instabilities that suddenly transitioned to catastrophic landslides when external forces triggered them, causing dramatic losses in basal shear strength. Must there be a distinct trigger for catastrophic failure to occur? What are the common factors for catastrophic failures? Do all large catastrophic rock slope failures evolve from creeping slopes? I briefly investigate these questions using the Eiger, Frank, Hope, and Randa sites as case studies to provide a broader context for my research.

On July 13, 2006, a 2-million-m³ rock collapse occurred in massive limestone at Eiger in central Switzerland (Jaboyedoff et al., 2012). The landslide happened on a prominent rock spur that had been recently (150 years ago) deglaciated; the spur is adjacent to a deep narrow gorge. There was no obvious trigger, as no exceptional precipitation or seismic events were recorded at the time of failure. However, snow melt and related high water pressures may have played a role. The first movements of the slope were noticed in 2005 after heavy precipitation when large rockfalls occurred and two tension cracks opened. Movement of the unstable mass then accelerated to up to 80 cm/day before final collapse. Jaboyedoff et al. (2012) suggest that strains related to the gorge geometry, glacial unloading, and rock slope fatigue probably brought the slope to the point of failure. The basal shear plane would have developed progressively parallel to pre-existing discontinuities, affected by a subsiding rear block, until ultimate shear strength was lost.

The 1903 30-million-m³ Frank Slide, one of the largest landslides in Canada, highlights the role of structural controls on rock slope failures. Its source is in the hinge zone of the Turtle Mountain anticline. The rock mass failed by sliding and toppling along bedding and is dissected by two orthogonal joint sets related to folding (Jaboyedoff et al., 2009; Brideau et al., 2011). Movement had been observed several years before the catastrophic failure, which appears to have been triggered by precipitation, freeze-thaw action, and mining activities in the toe of the slope. Present monitoring of the slope indicates further instability of the South Peak, suggesting a progressive failure mechanism.
The 1965 Hope Slide (47 million m$^3$) in British Columbia, Canada, occurred in greenstone and felsite (Brideau et al., 2005). Structural controls included faults in the headscarp area, foliation, and several discontinuity sets. Originally thought to have been triggered by earthquakes, it was later determined that the landslide itself produced seismic activity. No other trigger has been found. Brideau et al. (2005) observed seepage and surface runoff. An old landslide occurred at the same location, and the 1965 event may have been a reactivation of the same slope.

Like Turtle Mountain, the Randa slope is an example of continued deformation after catastrophic failure. A two-stage, 30-million-m$^3$ rockslide occurred in orthogneiss and paragneiss in 1991, controlled by three main discontinuity sets, a fault, and schistosity (Sartori et al., 2003). Movement continues at present at rates of up to 14 mm/day (Löw et al., 2012). Instability was apparent on the slope prior to 1991, indicating a long-term unstable slope system. Although no definitive trigger has been identified, hydraulic fracturing due to pore water may contribute to continuing instability.

The brief synopses of the four case studies above highlight several aspects of large, catastrophic rockslides. First, a distinctive trigger is not always apparent or necessary for catastrophic failure. The context of the slope, such as structures, geomorphic processes (especially glacial action), hydrogeological conditions, and slope morphology may be sufficient to damage the rock slope and bring it to the point of failure. At this stage, any small change in equilibrium may cause catastrophic failure. Second, a common thread in all the cases examined is structure. Faults, folds, and discontinuities are significant in conditioning slopes for failure. Finally, all of the cases investigated showed some sign of movement prior to collapse, and some slopes continued to slowly deform after failure. Glastonbury and Fell (2010) made a similar statement based on their examination of large rapid rockslides. They suggested that rupture surface strength loss is a key element in the initiation of rapid failures. Not all slowly deforming slopes, however, have evolved into rapid landslides; the La Clapière and Downie complexes are classic examples. The Åknes slope in western Norway is also deforming at a constant rate of 6 cm/day (Grøneng et al., 2010), but is being closely monitored given the high risk.
6.3. Advantages and Limitations of Methods Used

The methods used at Vajont and Madison Canyon form an integrated approach to the study of rock slope stability. They include traditional field techniques, laboratory testing, long-range photogrammetry, and numerical modelling. Table 6.2 lists the main advantages and disadvantages of each of these methods.

Table 6.2. Advantages and limitations of methods that I used to characterise complex large catastrophic landslides, specific to this dissertation.

<table>
<thead>
<tr>
<th>Method</th>
<th>Advantages</th>
<th>Limitations</th>
</tr>
</thead>
<tbody>
<tr>
<td>Field rock mass surveys</td>
<td>• cost-effective</td>
<td>• limited spatial extent, local scale</td>
</tr>
<tr>
<td></td>
<td>• allows statistical analysis of fracture sets</td>
<td>• subject to orientation bias</td>
</tr>
<tr>
<td></td>
<td>• can derive rock mass properties</td>
<td>• requires experience</td>
</tr>
<tr>
<td>Laboratory testing</td>
<td>• provides rock strength estimates</td>
<td>• extrapolated to rock mass strength</td>
</tr>
<tr>
<td></td>
<td>• time efficient (point load tests)</td>
<td>• requires several samples for significance</td>
</tr>
<tr>
<td></td>
<td>• simple, standardised procedures</td>
<td>• scale effect</td>
</tr>
<tr>
<td>Long-range photogrammetry</td>
<td>• allows characterisation of remote areas</td>
<td>• orientation bias, occlusion</td>
</tr>
<tr>
<td></td>
<td>• 3D models of object of interest</td>
<td>• requires specific field conditions</td>
</tr>
<tr>
<td></td>
<td>• cost-effective</td>
<td>• can be limited by geometry of field site</td>
</tr>
<tr>
<td></td>
<td>• high relative accuracy</td>
<td>• parameter uncertainty</td>
</tr>
<tr>
<td>Engineering geomorphology</td>
<td>• detailed observation of geomorphic features</td>
<td>• interpretation requires experience and engineering judgment</td>
</tr>
<tr>
<td>mapping</td>
<td>• slope change mapping easy to complete</td>
<td>• time-consuming, especially digitising</td>
</tr>
<tr>
<td></td>
<td>• integrates geology, landscape evolution, landform processes</td>
<td>• does not provide quantitative data</td>
</tr>
<tr>
<td></td>
<td>• tailored to each investigation</td>
<td>• requires experience</td>
</tr>
<tr>
<td>Numerical modelling</td>
<td>• allows testing of concepts and hypotheses</td>
<td>• computational inefficiency</td>
</tr>
<tr>
<td></td>
<td>• powerful analysis tool</td>
<td>• experience required</td>
</tr>
<tr>
<td></td>
<td>• tailored to research question</td>
<td>• model and parameter uncertainties</td>
</tr>
<tr>
<td></td>
<td>• many codes available</td>
<td>• difficult to constrain</td>
</tr>
</tbody>
</table>
Field investigation should be an integral part of any rock slope stability study. It provides validation for numerical models and remote sensing analyses. Scales of observation range from sub-outcrop to regional scale. I conducted field line surveys at a variety of locations in the Madison Slide headscarp area and outside the landslide. These surveys allowed measurement of all discontinuities along the survey lines. Where possible, I minimised orientation bias by surveying two or three lines oblique to each other at each site. Nevertheless, uncertainties and errors are inherent in any discontinuity characterisation method (Einstein and Baecher, 1982; Palmstrøm, 2001). I further characterised the rock mass by estimating block size and shape, GSI and DSI values, weathering grades, and rock mass strength; I also described source rock lithologies. Each of these techniques involves uncertainties and errors, including measurement error, inherent spatial variability of properties, and subjectivity. I minimised inconsistencies by using the same equipment throughout the field phase of the project, and the same observer to record information. Much of the geotechnical fieldwork at Vajont was completed by Ghirotti (1992) and Superchi (2012), who graciously provided their data.

Laboratory testing of Madison Canyon rocks yielded estimates of intact and rock mass strength and slake durability, which were used in numerical modelling. These results provide the first estimates of the geotechnical properties of the Madison Canyon rock masses, and are significant in validating numerical models. Limitations of the laboratory point load test data include small sample size, scale effects, and measurement and equipment errors. Intact rock samples in the laboratory should be considered upper bounds of strength. Several of my samples failed along pre-existing discontinuities, whereas others failed through intact rock. Hence, strength ranged widely, from 10 to 300 MPa. The rock masses at Madison are highly fractured and weathered, which contributes to the large range in their strength. I evaluated the sensitivity of model results to important parameters and attempted to assign reasonable values. Vajont material was tested by Superchi (2012).

Long-range photogrammetry was a useful complement to field observations, particularly as parts of both slopes are inaccessible. Photogrammetric discontinuity mapping yielded results consistent with field measurements. I used three-dimensional photogrammetric models as tools to map lineations and morphological features and to
make general observations. High relative accuracy was achieved in most photogrammetric models. The limitations of photogrammetry, as Poropat (2008) and Lyman et al. (2008) summarise, are noise, orientation bias, occlusion, and field conditions such as lighting, visibility, and weather. I endeavored to minimise occlusion and orientation bias. The limiting factors at Vajont are the smoothness and high reflectivity of the failure scar, which contribute to possible overexposure. At Madison Canyon, the line-of-sight from the camera to the headscarp was steep, possibly contributing to occlusion. Another factor at both sites was the distance between the camera and the scarps. Photomodels of each sliding surface were created from stations at least 800 m away. Although the distances did not appear to affect accuracy, they did prevent the use of laser scanning technology. The long ranges used in the photogrammetry methodology, some of the highest at up to 2.5 km, demonstrate the application of the technique at large distances from an area of interest.

Engineering geomorphology has found little application in North America, particularly in landslide research. It was, however, an integral part of my work, as it aided in the interpretation of the two landslides and highlighted the importance of geomorphic processes to landslide conditioning and causation. It also provided detailed knowledge of the two sites, as the mapping involved traverses in lines 50 m apart. Engineering geomorphology mapping, however, requires experience and is at best, semi-quantitative. For example, placement of symbols representing convex or concave breaks in slope is subjective and depends on mapper preferences. Engineering judgment is also a crucial component of interpretation, adding to subjectivity. Although morphometric mapping allows measurement of individual landforms and features such as slope angle, maps are still qualitative.

Slope stability analysis and numerical modelling, including kinematic and limit equilibrium analyses and numerical simulations, were performed at both sites. Stead et al. (2006) provide a comprehensive synopsis of the critical input parameters, advantages, and limitations of these methods, thus here I highlight particularly relevant aspects. Kinematic analysis emphasised the three-dimensional, complex nature of each landslide. Both slides involved combined wedge-biplanar sliding modes of failure. Kinematic and limit equilibrium analyses at Vajont suggest groundwater was necessary for failure; they underestimate the stability of the failure when only the east profile was
analysed. At Madison, the simulations required dramatic decreases of strength parameters to fail under static conditions. These results suggested that more sophisticated simulations were required. Continuum and discontinuum numerical modelling provided more reasonable and realistic results. One of the major limitations of these analyses is input parameter and model uncertainty. The subsurface geology is the main uncertainty for input into numerical models at Vajont and Madison Canyon. Two facets of this issue were apparent when choosing input values: i) degradation of laboratory- and field-estimated properties to rock mass scale, and ii) assignment of properties to different depths in the kilometre-scale models. Although much research is currently focussed on variability of properties and uncertainties (see, for example, Oberkampf et al., 2002; Wiles, 2006; Jefferies et al., 2008; Hammah et al., 2009), no definitive approach has been accepted. The second issue is also largely ignored, with uniform properties applied throughout models. Exceptions include Leith (2012) and Woo et al. (in press).

Contemplation of landscape evolution is important in applying the above techniques to the Vajont and Madison Canyon landslides, as they both occurred 50 years ago. I considered the changing slopes by examining photographs of each slide area from before, just after, and in the decades after the failures, carefully noting changes. I also explicitly investigated landscape evolution at each site, noting secondary failures, vegetation changes, and anthropic modifications. Field observations confirmed relative ages of features.

6.4. Highlights and Recommendations for Future Work

This dissertation presents an innovative, integrated approach to the study of large, catastrophic rock slope failures. This approach allowed me to achieve the four main objectives outlined in Chapter 1. The interaction between endogenic and exogenic processes and their effect on the landscape and specifically slope stability were determined through investigation of imagery and engineering geomorphological mapping. Analysis and integration of field, remote sensing, and numerical modelling results suggested initiation mechanisms at each of the two study sites. Landscape
evolution proved to be important in conditioning slopes and causing them to fail. Finally, a relative sequence of events was determined for each case study.

Highlights of my research include:

• one of the first studies on the geomorphology of Vajont Valley, providing context for the Vajont Slide, including the pre-1963 topography and gorge,
• the most extensive study of the Madison Canyon Slide since Hadley (1978),
• the first detailed field and remote engineering geomorphology mapping at both sites, contributing to interpretations of local landscape and landslide evolution and behaviour, and the determination of landslide blocks or zones (two main blocks at Vajont and five sub-blocks, ten zones at Madison Canyon),
• the first high-accuracy, long-range photogrammetry models at each site used to characterise rock masses and debris and determine discontinuity sets,
• a preliminary morphology classification of the failure scar at Vajont, illustrating areas of roughness with implications for failure mechanisms,
• kinematic analyses of the complex, three-dimensional failures, highlighting the importance of the Erto Syncline at Vajont and the composite release surfaces at Madison Canyon,
• agreement of simulated block movement directions in the 3DEC models of Vajont with Broili’s (1967) interpretations,
• preliminary groundwater simulations in limit equilibrium analyses of Vajont, showing reduction in strength due to pore water pressures,
• further analysis of the kinematics of Vajont using two- and three-dimensional codes, showing that the optimum number of blocks in the Vajont sliding mass was between three and 12, and that the Erto Syncline and Massalezza Gully may have played important roles in the development of the failure,
• successful simulation of brittle fracture and shear zone development at an average angle of 75°-80° from horizontal in the Prandtl transition zone of the Vajont Slide using UDEC Trigon and Slope Model, supporting Rossi and Semenza’s hypotheses,
• illustration of the importance of weak materials and groundwater to failure at Vajont,
• simulations showing the importance of discontinuity and sliding surface geometry at Madison Canyon,
• sophisticated two-dimensional dynamic modelling of the effect of seismicity on the Madison Canyon slope, illustrating topographic and structural amplification,
• demonstration of the importance of the interaction between endogenic and exogenic processes at each site and its role in conditioning each slope for failure, and
• determination of a detailed sequence of events from slope conditioning to post-catastrophic failure for both the Vajont and Madison Canyon landslides.

Contributions to the study of large catastrophic rock slope failures include:

• successful and highly effective combination of field, remote sensing, engineering geomorphology mapping, and numerical modelling to understand complex rock slope failure and geological and geomorphological contexts and conditions,
• creation of high-accuracy, long-range terrestrial digital photogrammetry projects up to 2.5 km from the object of interest, highlighting the range and application of this versatile technique,
• illustration of engineering geomorphological mapping and consideration of landscape evolution as a significant tool in any landslide study,
• development of a Debris Strength Index (DSI) to relate landslide debris to in situ source material, and
• the application of a toolbox of sophisticated numerical modelling codes, adjusting the software to the problem and validating modelling results against field and imagery observations.

The techniques that I have applied to the Vajont and Madison Canyon landslides can be used effectively in other, potentially more data-rich, back analyses and forensic investigations of past mass movements. They can also be used to contextualise, discover, and monitor future instabilities.

Several areas of further research are planned and recommended:

• integration of the proposed methodology with subsurface investigations at landslides to allow improved rock mass characterisation and understanding of failure mechanisms,
• further investigation of the Quaternary history of Vajont Valley,
• monitoring of the tension cracks at Madison Canyon and the secondary landslide source areas at Vajont,
• development and application of the Debris Strength Index (DSI) to other large catastrophic failures,
• analysis of debris photogrammetry models, with block size and shape quantification,
• use of Unmanned Aerial Vehicles (UAVs) in creating photogrammetry projects, thus increasing accessibility to failures,
• application of thermal imaging techniques to rock slope failures (e.g., Vivas-Becerra et al., 2013),
• creation of more numerical simulations, paying attention to parameter input, the level of sophistication required, and aspects such as seismicity and groundwater,

• application of the trilinear failure criterion and strain-softening (data permitting) to model landscape evolution and landslide behaviour,

• examination of the interaction between groundwater conditions and failure kinematics,

• increased integration of block theory with three-dimensional numerical modelling of landslide mechanisms, and

• a more rigorous appraisal of parameter and model uncertainty using a probabilistic approach.
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Appendix A.

Vajont Slide Engineering Geomorphology Field Map

Note: A larger image of the map can be found as a supplementary file named AppendixA_VajontMap.pdf
Appendix B.

Madison Canyon Slide Engineering Geomorphology Field Map

Note: A larger image of the map can be found as a supplementary file named AppendixB_MadisonCanyonMap.pdf
Appendix C.

Madison Canyon Thin Section Descriptions

MC-AWG-11-2-07A

Rock Name: K-feldspar gneiss

<table>
<thead>
<tr>
<th>Mineral Constituents</th>
<th>%</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Primary</strong></td>
<td></td>
</tr>
<tr>
<td>Quartz</td>
<td>15</td>
</tr>
<tr>
<td>Plagioclase</td>
<td>20</td>
</tr>
<tr>
<td>Potassium-Feldspar</td>
<td>40</td>
</tr>
<tr>
<td>Muscovite</td>
<td>5</td>
</tr>
<tr>
<td>Biotite</td>
<td>10</td>
</tr>
<tr>
<td>Chlorite-after-biotite/hornblende</td>
<td>10</td>
</tr>
<tr>
<td><strong>Secondary</strong></td>
<td></td>
</tr>
<tr>
<td>Chlorite after biotite/hornblende</td>
<td></td>
</tr>
<tr>
<td>Sericite in feldspars</td>
<td></td>
</tr>
<tr>
<td>Introduced: opaques (iron oxides),</td>
<td></td>
</tr>
<tr>
<td>carbonates</td>
<td></td>
</tr>
</tbody>
</table>

**lots of open fractures, voids ~25% of view

Textures:
- <0.1 – 2 mm grain size
- feldspar coarse
- highly fractured (inter and intragranular)
- xenoblastic to subidioblastic
- feldspar twinning: some deformation, some growth
- muscovite plates
- weak foliation
- porphyroblastic/poikiloblastic; feldspars (and some quartz) much larger than micas, etc.
- micrographic/myrmekitic
Photograph:

![Photograph of sample MC-AWG-11-2-07A under 2X magnification and crossed polars.]

**Figure C.1.** Photograph of sample MC-AWG-11-2-07A under 2X magnification and crossed polars.
**Rock Name: feldspar schist**

### Mineral Constituents

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<thead>
<tr>
<th></th>
<th>%</th>
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<tbody>
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</tr>
<tr>
<td>Quartz</td>
<td>45</td>
</tr>
<tr>
<td>Plagioclase</td>
<td>15</td>
</tr>
<tr>
<td>Potassium-Feldspar</td>
<td>20</td>
</tr>
<tr>
<td>Muscovite</td>
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<tr>
<td>Chlorite-after-biotite/hornblende</td>
<td>10</td>
</tr>
<tr>
<td><strong>Secondary</strong></td>
<td></td>
</tr>
<tr>
<td>Chlorite after biotite/hornblende</td>
<td></td>
</tr>
<tr>
<td>Sericite in feldspars</td>
<td></td>
</tr>
<tr>
<td>Introduced: opaques (haematite, pyrite)</td>
<td></td>
</tr>
</tbody>
</table>

**Textures:**

- <0.01 – 0.8 mm grain size
- feldspar coarsest
- feldspars subidioblastic
- quartz xenoblastic
- quartz: undulose extinction, sutured
- muscovite needle-like
- moderately foliated (pressure shadows)
- feldspar porphyroblasts, poikiloblasts, exsolution laminae, perthitic, some grain boundaries crushed
- micrographic/granophyric/myrmekitic feldspars, quartz (crystallised at same time)
Figure C.2. Photograph of sample AW-MCS-12-4-05A under 2X magnification and crossed polars.
Rock Name: quartz-K-feldspar-plagioclase gneiss

<table>
<thead>
<tr>
<th>Mineral Constituents</th>
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<tbody>
<tr>
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<td>40</td>
</tr>
<tr>
<td>Plagioclase</td>
<td>15</td>
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<tr>
<td>Potassium-Feldspar</td>
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<td>Muscovite</td>
<td>2</td>
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<tr>
<td>Chlorite-after-biotite/hornblende</td>
<td>18</td>
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<td><strong>Secondary</strong></td>
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<tr>
<td>Chlorite after biotite</td>
<td></td>
</tr>
<tr>
<td>Slight iron staining along some microfractures</td>
<td></td>
</tr>
<tr>
<td>Sericite in feldspars</td>
<td></td>
</tr>
<tr>
<td>Introduced: opaques</td>
<td></td>
</tr>
</tbody>
</table>

Textures:
- <0.1 – 1 mm grain size
- feldspar and quartz coarser than mafics
- xenoblastic
- some feldspars zoned, most twinning growth
- quartz sutured, some undulose extinction
- weakly foliated
- graphic/granophytic/myrmekitic feldspar and quartz very common
- long, open fractures common (inter- and intragranular)
Figure C.3. Photograph of sample AW-MCS-12-5-05A under 2X magnification and crossed polars.
Rock Name: K-spar-quartz-chlorite gneiss

<table>
<thead>
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<th>Primary %</th>
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<tbody>
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<tr>
<td>Plagioclase</td>
<td>10</td>
</tr>
<tr>
<td>Potassium-Feldspar</td>
<td>40</td>
</tr>
<tr>
<td>Muscovite</td>
<td>2</td>
</tr>
<tr>
<td>Chlorite-after-biotite/hornblende</td>
<td>15</td>
</tr>
</tbody>
</table>

Secondary

- Chlorite after biotite
- Sericite in feldspars
- Introduced: opaques (haematite) in clusters 3

Textures:

- <0.1 – 1 mm grain size
- feldspar coarser, quartz smaller, muscovite tiny plates
- muscovite plates
- anhedral-subhedral quartz and feldspar, rims broken
- twinning in feldspars: some growth, some deformation, minor bending and kinking
- opaques: some rhombs and cubes, mostly infill/irregular
- micrographic/ granophyric/myrmekitic quartz in feldspar
- opaques superimposed on quartz and feldspars
- **foliation not apparent**
Photograph:

Figure C.4. Photograph of sample AW-MCS-12-6-03A under 2X magnification and crossed polars.
Rock Name: quartz schist

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<tr>
<td>Quartz</td>
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<td>Plagioclase</td>
<td>15</td>
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<td>Potassium-Feldspar</td>
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<td>Muscovite</td>
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<tr>
<td>Chlorite-after-biotite/hornblende</td>
<td>8</td>
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<tr>
<td><strong>Secondary</strong></td>
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</tr>
<tr>
<td>Chlorite after biotite</td>
<td></td>
</tr>
<tr>
<td>Sericite in feldspars</td>
<td></td>
</tr>
<tr>
<td>Introduced: opaques (haematite) in clusters, calcite</td>
<td></td>
</tr>
</tbody>
</table>

Textures:

- <0.1 – 0.2 mm grain size, some feldspar ~1 mm
- feldspar hypidioblastic
- quartz, micas, and chlorite xenoblastic
- quartz: undulose extinction, sutured
- foliated micas, chlorite, some needle-like
- some porphyroblastic/poikiloblastic feldspar (intergrown)
- very little micrographic/granophyric feldspar/quartz
- calcite superimposed on quartz
- suggests sinistral shear direction
Figure C.5. Photograph of sample AW-MCS-12-6-08A under 10X magnification and crossed polars.
AW-MCS-12-6-11B

Rock Name: quartz schist

<table>
<thead>
<tr>
<th>Mineral Constituents</th>
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<td>Plagioclase</td>
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</tr>
<tr>
<td>Potassium-Feldspar</td>
<td>25</td>
</tr>
<tr>
<td>Muscovite</td>
<td>3</td>
</tr>
<tr>
<td>Chlorite-after-biotite/hornblende</td>
<td>7</td>
</tr>
<tr>
<td><strong>Secondary</strong></td>
<td></td>
</tr>
<tr>
<td>Chlorite after biotite</td>
<td></td>
</tr>
<tr>
<td>Sericite in feldspars</td>
<td></td>
</tr>
<tr>
<td>Introduced: opaques (haematite) in fractures, quartz veinlets</td>
<td></td>
</tr>
</tbody>
</table>

Textures:

- <0.1 – 1 mm grain size
- feldspar coarse
- vein quartz coarser than primary quartz
- micas fine (muscovite coarser than chlorite)
- feldspars hypidioblastic
- quartz, micas xenoblastic
- quartz: undulose extinction, sutured
- moderately strongly foliated
- feldspar porphyroblasts, poikiloblasts, twinning kinked (some deformation, growth twins)
- micrographic/granophyric/myrmekitic feldspars, quartz (crystallised at same time)
- suggests dextral shear direction
Figure C.6. Photograph of sample AW-MCS-12-5-11B under 2X magnification and plane light.
MC-AWG-11-4-02A

Rock Name: banded quartz marble

<table>
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<tr>
<th>Mineral Constituents</th>
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<tbody>
<tr>
<td><strong>Primary</strong></td>
<td>%</td>
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<tr>
<td>Carbonate</td>
<td>70</td>
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<tr>
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</tr>
<tr>
<td>Minor iron oxide staining</td>
<td></td>
</tr>
<tr>
<td>Introduced: quartz veinlets</td>
<td></td>
</tr>
</tbody>
</table>

Textures:

- <0.01 – 0.1 mm grain size
- irregular (mostly anhedral, some subhedral)
- undulose extinction in quartz
- twinning in calcite (mostly growth twins)
- sharp contact between quartz and calcite, some mixing
- calcite fractured (conjugate fractures), infilled with cc? (v. small grains)
Figure C.7. Photograph of sample MC-AWG-11-4-02A under 5X magnification and crossed polars.
Rock Name: banded quartz marble

<table>
<thead>
<tr>
<th>Mineral Constituents</th>
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</thead>
<tbody>
<tr>
<td><strong>Primary</strong></td>
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</tr>
<tr>
<td>Carbonate</td>
<td>60</td>
</tr>
<tr>
<td>Quartz</td>
<td>40</td>
</tr>
<tr>
<td><strong>Secondary</strong></td>
<td></td>
</tr>
<tr>
<td>Minor iron oxide staining (ankerite?), pyrite</td>
<td></td>
</tr>
<tr>
<td>Introduced: quartz veinlets</td>
<td></td>
</tr>
</tbody>
</table>

Textures:
- <0.01 – 0.1 mm grain size
- carbonate grains mostly smaller than quartz, but also larger
- irregular (mostly anhedral, some subhedral cubes and rhombs)
- some quartz grains in carbonate
- veinlets oriented NW in thin section, parallel
- sharp contact between quartz and calcite
Figure C.8. Photograph of sample AW-MCS-12-1-06A under 2X magnification and crossed polars.
Appendix D.

Madison Canyon Slake Durability Results

D.1. Results

Table D.1. Results of slake durability tests on four samples from Madison Canyon.

<table>
<thead>
<tr>
<th>Project Name: AW-MCS-12-2-01B</th>
<th></th>
<th>Project Name: AW-MCS-12-2-02A</th>
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<td>Mass of Drum + samples pre test (g)</td>
<td>1325.7</td>
<td>Mass of Drum + samples pre test (g)</td>
</tr>
<tr>
<td>Date/Time in Oven: 05/10/2012 17:20</td>
<td></td>
<td>Date/Time in Oven: 06/10/2012 8:33</td>
</tr>
<tr>
<td>Date/Time out of Oven: 05/10/2012 22:30</td>
<td></td>
<td>Date/Time out of Oven: 06/10/2012 14:00</td>
</tr>
<tr>
<td>Mass of Drum + samples after first drying (g)</td>
<td>1322.3</td>
<td>Mass of Drum + samples after first drying (g)</td>
</tr>
<tr>
<td>Date/Time in Oven: 05/10/2012 22:55</td>
<td></td>
<td>Date/Time in Oven: 06/10/2012 14:32</td>
</tr>
<tr>
<td>Date/Time out of Oven: 06/10/2012 8:00</td>
<td></td>
<td>Date/Time out of Oven: 06/10/2012 22:30</td>
</tr>
<tr>
<td>Mass of Drum + samples after final drying (g)</td>
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<td>Mass of Drum + samples after final drying (g)</td>
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<tr>
<td>Slake Type</td>
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<td>Slake Type</td>
</tr>
<tr>
<td>Pre Water Temperature I (°C)</td>
<td>23</td>
<td>Pre Water Temperature I (°C)</td>
</tr>
<tr>
<td>Post Water Temperature I (°C)</td>
<td>23.5</td>
<td>Post Water Temperature I (°C)</td>
</tr>
<tr>
<td>Pre Water Temperature II (°C)</td>
<td>22.5</td>
<td>Pre Water Temperature II (°C)</td>
</tr>
<tr>
<td>Post Water Temperature II (°C)</td>
<td>24</td>
<td>Post Water Temperature II (°C)</td>
</tr>
<tr>
<td>Oven Temperature (°C)</td>
<td>110</td>
<td>Oven Temperature (°C)</td>
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<table>
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<tr>
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<th>Project Name: AW-MCS-12-4-03B</th>
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</tr>
<tr>
<td>Date/Time out of Oven: 05/10/2012 22:30</td>
<td></td>
<td>Date/Time out of Oven: 06/10/2012 14:00</td>
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<tr>
<td>Mass of Drum + samples after first drying (g)</td>
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<td>Mass of Drum + samples after first drying (g)</td>
</tr>
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<td>Date/Time in Oven: 06/10/2012 14:32</td>
</tr>
<tr>
<td>Date/Time out of Oven: 06/10/2012 8:00</td>
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<td>Date/Time out of Oven: 06/10/2012 22:30</td>
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<td>Mass of Drum + samples after final drying (g)</td>
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<td>Mass of Drum + samples after final drying (g)</td>
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<td>Slake Type</td>
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<td>Pre Water Temperature I (°C)</td>
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<td>Pre Water Temperature II (°C)</td>
<td>22.5</td>
<td>Pre Water Temperature II (°C)</td>
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<td>Post Water Temperature II (°C)</td>
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<tr>
<td>Oven Temperature (°C)</td>
<td>110</td>
<td>Oven Temperature (°C)</td>
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</table>
D.2. Photographs

Figure D.1. Prepared blocks of the AW-MCS-12-2-01B sample before slake durability test.

Figure D.2. Prepared blocks of the AW-MCS-12-2-01B sample after slake durability test.
Figure D.3. Prepared blocks of the AW-MCS-12-2-02A sample before slake durability test.

Figure D.4. Prepared blocks of the AW-MCS-12-2-02A sample after slake durability test.

Figure D.5. Prepared blocks of the AW-MCS-12-2-03A sample before slake durability test.
Figure D.6. Prepared blocks of the AW-MCS-12-2-03A sample after slake durability test.

Figure D.7. Prepared blocks of the AW-MCS-12-4-03A sample before slake durability test.

Figure D.8. Prepared blocks of the AW-MCS-12-4-03A sample after slake durability test.
### Appendix E.

## Methods Used and Numerical Modelling Code

### E.1. Summary of Techniques Used

Table E.1. Summary of methods, output, and software used in each chapter.

<table>
<thead>
<tr>
<th>Chapter</th>
<th>Scale</th>
<th>Method</th>
<th>Output</th>
<th>Software</th>
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<tbody>
<tr>
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<td>regional features, knickpoints, gradient</td>
<td>ArcGIS 3D Analyst</td>
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<td>transverse profiles</td>
<td>rock mass strength</td>
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<td></td>
<td>Selby (1980) RMS</td>
<td>c/h ratio</td>
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<td></td>
<td>long-range TDP</td>
<td>3D photomodels</td>
<td>AdamTech suite</td>
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<tr>
<td></td>
<td></td>
<td>Vajont Slide</td>
<td>bench, lineation mapping</td>
<td>Adobe Illustrator, ArcGIS/Illustrator</td>
</tr>
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<td>engineering geomorphology mapping</td>
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<td>Polyworks</td>
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<td>lineation mapping</td>
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<td>Illustrator/Excel, ArcGIS 3D Analyst</td>
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289
<table>
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<th>Software</th>
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<td>distinct element complex geometry</td>
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<td>lattice-spring (DEM model)</td>
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<tr>
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<td>Madison Valley/</td>
<td>field observation stream profile</td>
<td>regional features</td>
<td>ArcGIS 3D Analyst</td>
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<tr>
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<td>Madison Slide</td>
<td>transverse profiles c/h ratio</td>
<td>knickpoints, gradient</td>
<td>ArcGIS 3D Analyst</td>
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<tr>
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<td></td>
<td>long-range TDP</td>
<td></td>
<td>ArcGIS/Excel</td>
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</table>

**E.2. Codes**

**E.2.1. Chapter 4: UDEC**

*Sliding Surface Geometries: Rossi and Semenza Section 2 example*

;;--GEOMETRY--;

291
block -615.6,-1000 -615.6,910 -300,920 -250,915 -125,835 0,744.2 33.6,689.3 207.4,646.6 402.6,573.4 573.4,463.6 677.1,414.8 750.3,353.8 902.8,366 994.3,366 1274.9,140.3 1274.9,24.4 1293.2,115.9 1342,122 1476.2,256.2 1543.3,280.6 1543.3,-1000

;Ghirotti slip surface and boundaries

;crack 33.6,689.3 36.6,658.8
;crack 36.6,658.8 207.4,561.2
;crack 207.4,561.2 402.6,439.2
;crack 402.6,439.2 573.4,317.2
;crack 573.4,317.2 677.1,219.6
;crack 677.1,219.6 750.3,183
;crack 750.3,183 902.8,140.3
;crack 902.8,140.3 1274.9,140.3

fix range -620,1550 -1005,0

;--MATERIAL PROPERTIES--

;properties of Dogger, linear-elastic (mat=1,cons=1, default)

set degrees on

prop mat=1 d=2690 k=6.14e9 g=3.68e9 f=45

change mat=1 cons=1

joint model area jkn=5e9 jks=6e8 jfric=20

set gravity 0 -9.81

hist unbal

hist xdis 1000,300 ;centre of lower block

hist ydis 1000,300

hist xvel 1000,300

hist yvel 1000,300

hist xdis 750,350 ;near top of intersection between chair back and seat

hist ydis 750,350

hist xvel 750,350

hist yvel 750,350

292
hist xdis 300,600 ;centre of upper block
hist ydis 300,600
hist xvel 300,600
hist yvel 300,600
solve

;--REDUCED PROPERTIES--
joint model area jkn=5e9 jks=6e8 jfric=12 ; jfric subsequently decreased to 10° and 5°
solve
; jfric decreased to 10° and 5° in subsequent models

**E.2.2. Chapter 4: 3DEC**

*Simple Block Model Sample with DS6 spaced at 200*

;--GEOMETRY--
poly brick (0,4000) (-100,2600) (-500,1750)
pl bl
jset dip 0 dd 0 origin 0,0,500 ;valley bottom
jset dip 5 dd 0 origin 0,1940,500 ;base of sliding surface
;Monte Toc slope
hide range z -500,500
jset dip 50 dd 0 origin 0,0,1750
jset dip 30 dd 0 origin 2000,550,1270
jset dip 5 dd 0 origin 2000,1350,850
delete range y 1180,2600 z 800,1750
;Vajont Gorge
jset dip 67 dd 0 origin 0,1885,630
jset dip 67 dd 180 origin 0,1995,630
delete block 8549
hide range y 2000,2600
;Col Tramontin Fault
jset dip 59 dd 283 origin 2940,670,1245
;Erto Syncline seat
jset dip 20 dd 90 origin 1030,1260,860
;Massalezza Syncline
jset dip 34 dd 31 origin 1030,1260,860
jset dip 36 dd 353 origin 2950,270,1430
seek
hide block 217
hide block 6263
hide block 7251
join on
mark region 1
seek
hide region 1
join on
mark region 2
jset dip 40 dd 90 origin 1950,800,1110
jset dip 40 dd 270 origin 1950,800,1110
delete block 6263
join
jset dip 88 dd 180 num 100 origin 0,1885,630 spacing 200
seek
jset dip 90 dd 0 origin 0,0,0
jset dip 90 dd 0 origin 0,2500,0
jset dip 90 dd 90 origin 100,0,0
jset dip 90 dd 90 origin 3900,0,0
jset dip 0 dd 0 origin 0,0,-400
;--INITIAL PROPERTIES AND EQUILIBRIUM--
prop mat=1 density=2690 bulk=6.14e9 shear=3.68e9
prop mat=2 density=2610 bulk=2.66e9 shear=1.6e9
prop jmat=1 jkn=5e9 jks=6e8 jfric=60 jcoh=5e4 ;high properties
change region 1 mat=1
change region 2 mat=2
gravity 0,0,-9.81
fix range y -100,0
fix range y 2500,2600
fix range x -10,100
fix range x 3900,4010
fix range z -510,-400
history unbal
history ydisp 1720,510,1290
history xdisp 1720,510,1290
history zdisp 1720,510,1290
history ydisp 2170,510,1290
history xdisp 2170,510,1290
history zdisp 2170,510,1290
history ydisp 2100,1880,640
history xdisp 2100,1880,640
history zdisp 2100,1880,640
solve
;--PROPERTIES REDUCED--
prop jmat=1 jfric=30 jcoh=2.5e4
solve
prop jmat=1 jfric=20 jcoh=0
solve
prop jmat=1 jfric=10
solve
prop jmat=1 jfric=5
solve

**Complex Topography Sample**

;--GEOMETRY

new
call k1.3dec ;block model created in Rhino/Kubrix
hide range region 2
join on
seek
hide range region 1
join on
;
save "Vajont_lsblocks_oneblockini.sav"
;landslide blocks
jset dip 90 dd 137 origin 2470,923,1166 ; east block
jset dip 90 dd 119 origin 2006,1524,838 ; lower Massalezza
jset dip 90 dd 85 origin 2006,1524,838 ; upper Massalezza
hide dip 90 dd 119 origin 2006,1524,838 above
join on
seek dip 90 dd 85 origin 2006,1524,838 below
hide range region 1
jset dip 90 dd 12 origin 2006,1524,838
jset dip 90 dd 16 origin 1980,1116,1000
save "Vajont_lsblocks_geomini.sav"
;--INITIAL PROPERTIES
config array 6000
seek
prop mat=1 density=2690 bulk=6.14e9 shear=3.68e9
prop jmat=1 jkn=100e9 jks=10e9 jfric=90 jcoh=5e4 ;high properties to start
gravity 0,0,-9.81
insitu topo kox 1 koy 1 koz 1 zup
gen edge 100
fix xrange 50,150 ;xvel 0
fix xrange 4100,4200 ;xvel 0
fix yrange 250,350 ;yvel 0
fix yrange 2800,2900 ;yvel 0
fix zrange -1050,-900 ;xvel 0 yvel 0 zvel 0
history unbal
;
def Get_out
loop n (1,10)
savfile = 'Vajont_propinistep' + string(n)
command
step 1000
save @savfile
endcommand
endloop
end
@Get_out
;
solve
save "Vajont_lsblocks_propini.sav"
;--DOWNGRADED PROPERTIES--
prop jmat=2 jkn=5e9 jks=0.6e9 jfric=36 jcoh=1e4 jten=400
change jmat=2 range joint 1
;
def Get_out
loop n (1,10)
savfile = 'Vajont_propssstep' + string(n)
command
step 1000
save @savfile
endcommand
endloop
end
@Get_out
;
solve
save "Vajont_lsblocks_propss.sav"

**E.2.3 Chapter 5: UDEC**

*Static Modelling: Sliding Surface Properties Sample*

;MADISON CANYON Cross-Section E-E'
;Back analysis with failure surface, no joints
;
round 0.5
block 0,0 0,1225 173,1277 383,1216 398,1216 758,902 1028,951 1223,1079 1223,0
;
;marble wedge
crack 398,1216 1223,498
crack 383,1216 1055,0
;
;add failure surface
crack 87,1251 421,1073
crack 421,1073 633,1012
;
;MATERIAL PROPERTIES
; weathered gneiss/schist
prop mat=1 d=2700 k=1.1e10 g=0.6e10 f=32 c=9e6
; marble buttress
prop mat=2 d=2900 k=3.6e10 g=2.2e10 f=35 c=12e6
; failure surface
prop jmat=1 jkn=1e10 jks=1e9 jfr=20
;
; assign properties to blocks
change 0,450 0,1250 mat=1 cons=3
change 1000,1230 700,1000 mat=1 cons=1
change 450,1000 400,1150 mat=2 cons=3
change jmat=1 jcons=2
;
; create deformable block mesh
gen edge 15 range 100,700 1000,1250
;
set gravity 0 -9.81
;
; fix boundaries
fix -1,1250 -1,900
;
; damp local
;
hist unbal
hist vmax
hist xdis 296,1185
hist ydis 296,1185
solve
Dynamic Modelling: Convex Geometry with all Discontinuities Sample

new
set memory 500

; This code was developed with much help from Valentin Gischig and Fuqiang Gao.

;--GEOMETRY--
block -400,1000 -400,2045 0,2045 660,2310 1050,2150 1335,1930 1800,1930 2000,2045 2200,2045 2200,1000

table 100 -400,2045
table 100 0,2045
table 100 660,2310
table 100 1050,2150
table 100 1335,1930
table 100 1800,1930
table 100 2000,2045
table 100 2200,2045

crack 1095,2130 1550,1200 ; marble buttress
crack 1335,1930 1710,1200

crack 1500,1201 1750,1201

crack 620,2300 620,2280 ; ss and tension cracks
crack 660,2310 660,2200
crack 700,2300 700,2200
crack 620,2280 1150,1950
crack 1150,1950 1335,1930
jdelete
gen quad 15 ; max frequency modelled accurately f=9.9 (f=C_s/lambda=C_s/10deltal)
gen edge 15
save 'MadisonC_ridge_convex_ssdol_geomini.sav'

; --PROPERTIES--
def set_param
    ; Metamorphics properties
    M_den=2700
    M_k=11e9
    M_g=6e9
    ;
    ; Marble properties
    D_den=2700
    D_k=47.7e9
    D_g=22e9
    ;
    ; Sliding surface properties
    s_fri=40
    s_coh=5000
    s_ten=1732
    s_kn=100e9
    s_ks=10e9
    ;
    ; Tension crack properties
    t_fri=30
    t_coh=5000000
    t_ten=1732
    t_kn=1e9
    t_ks=0.1e9
end
set_param
;
set ov=0.1
;
;assign properties
prop mat=1 dens=M_den bulk=M_k shear=M_g
prop mat=2 dens=D_den bulk=D_k shear=D_g
;
change mat=2 range x 1200,1600
;
prop jmat=1 jkn=s_kn jks=s_ks jfric=s_fri jcoh=5000000 jten=1732000
prop jmat=2 jkn=s_kn jks=s_ks jfric=s_fri jcoh=5000000 jten=1732000
change jmat=2 range angle 80,100
save 'MadisonC_ridge_convex_ssdol_propini.sav'

;--INITIAL CONDITIONS, STATIC EQUILIBRIUM
insitu stress -3.47e7,0,-3.47e7 ygrad 2.65e4,0,2.65E4 ;k=1, h=1310 (peak surface height)

def stress_state
ko = 1.0
s_yy = -2045*M_den*9.81 ;Pa
s_xx = s_yy*ko
s_zz = s_yy*ko
syg=M_den*9.81
sxg=syg*ko
szg=syg*ko
end
stress_state
bound stress \( (s_{xx}, 0, s_{yy}) \) ygrad \((sxg, 0, syg)\) range xrange -400.1,-399.9
bound stress \( (s_{xx}, 0, s_{yy}) \) ygrad \((sxg, 0, syg)\) range xrange 2199,2201
boundary yvel 0 range yrange 999.9,1000.1

set gravity 0,-9.81

cal hist_routine.txt ; creates a "history field", i.e., histories spaced at 20 m intervals along surface.
+
history unbal
hist vmax

solve ratio 1e-10
save 'MadisonC_ridge_convex_ssdol_runstatic.sav'

;--NISQUALLY DYNAMIC RUN--
mscale off
table 1 read seis179.seis
;
def convert
\[
c_p = \sqrt{(M_k + (4.0*M_g/3.0))/M_{den}}
\]
\[
c_s = \sqrt{M_g/M_{den}}
\]
norm_str = 1.0*M_{den}*c_p ; vs and vn=1.0
shear_str = 1.0*M_{den}*c_s
end
set M_k=11e9 M_g=6e9 M_{den}=2700
convert
;
boundary ffield

boundary xvisc ff_bulk=M_k ff_shear=M_g ff_density=M_den range yrange 999.9,1000.1
boundary yvisc ff_bulk=M_k ff_shear=M_g ff_density=M_den range yrange 999.9,1000.1
boundary stress 0.0,0.0,0.0 range yrange 999.9,1000.1
boundary stress 0.0,shear_str,0.0 hist=table 1 range -400.1,1800.1 999.9,1000.1
reset vel time disp
damp 0.001 2
;
def get_out
  star=1
  n cyc=40
  each=5
  loop icyc(star, n cyc)
    snam = 'convex_ssdol_Nisqually179_shear_' + string(icyc) + 's' + '.sav'
    command
cyc time 5
  sav snam
  endcommand
endloop
end
get_out
save 'MadisonC_ridge_convex_ssdol_rundyn_Nisqually179_shear.sav'

;--REDUCED PROPERTIES ALONG TENSION CRACKS--
;original properties
prop jmat=2 jkn=s_kn jks=s_ks jfric=s_fri jcoh=5000000 jten=1732000
; properties for failed discontinuities
prop jmat 3 jfric=t_fri jcoh=0 jten=0 jkn=t_kn jks=t_ks
set echo off

def fracturedjoints
  ci = contact_head
  loop while ci # 0
    n_f = c_nforce(ci)
    s_f = c_sforce(ci)
    if n_f = 0
      if s_f = 0
        if c_mat(ci) = 2
          c_mat(ci) = 3
        end_if
      end_if
    end_if
    ci = c_next(ci)
  endloop
end
fracturedjoints
set echo on
solve rat 1e-6
save 'MadisonC_ridge_convex_ss dol_lowTCprop.sav'

--;--RICKER DYNAMIC RUN
mscale off
table 2 read Ricker01Hz.txt ; subsequently run for 0.1, 0.5, 1, 2, 5, 8, and 10 Hz
;
def convert
c_p = sqrt((M_k + (4.0*M_g/3.0))/M_den)
c_s = sqrt(M_g/M_den)
norm_str = -2.0*M_den*c_p ;vs and vn=1.0
shear_str = -2.0*M_den*c_s
end
set M_k=11e9 M_g=6e9 M_den=2700
convert ;
boundary stress 0.0,0.0,0.0 range -400.1,1800.1 999.9,1000.1
boundary stress 0.0,0.0,0.0 range -400.1,1800.1 999.9,1000.1
reset vel time disp
damp 0.001 0.1

hist norm_str
hist shear_str
;
def get_out
    star=1
    ncy=40
    each=5
    loop icyc(star, ncy)
        snam = 'convex_ssdol_R0Hz_shear_' + string(icyc) + 's' + '.sav'
        command
cyc time 5
sav snam
endcommand
endloop
endloop
end
get_out
save 'MadisonC_ridge_convex_ssdol_rundyn_R0Hz_shear.sav'