

CONTROLS ON THE KINEMATICS OF SLOW-MOVING LANDSLIDES FROM
SATELLITE RADAR INTERFEROMETRY AND MECHANICAL MODELING

by

ALEXANDER LOUIS HANDWERGER

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Student: Alexander Louis Handwerger

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This dissertation has been accepted and approved in partial fulfillment of the requirements for the Doctor of Philosophy degree in the Department of Geological Sciences by:

Dr. Joshua J. Roering	Co-Chair
Dr. David A. Schmidt	Co-Chair
Dr. Alan W. Rempel	Core Member
Dr. Mark Fonstad	Institutional Representative

and

Scott L. Pratt	Dean of the Graduate School
----------------	-----------------------------

Original approval signatures are on file with the University of Oregon Graduate School.

Degree awarded June 2015

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DISSERTATION ABSTRACT

Alexander Louis Handwerger

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Title: Controls on the Kinematics of Slow-moving Landslides from Satellite Radar Interferometry and Mechanical Modeling

Landslides display a wide variety of behaviors ranging from slow persistent motion to rapid acceleration and catastrophic failure. Given the variety of possible behaviors, improvements to our understanding of landslide mechanics are critical for accurate predictions of landslide dynamics. Recent advances in remote sensing techniques, like satellite radar interferometry (InSAR), now enable high-resolution spatial and temporal measurements that provide insight into the mechanisms that control landslide behavior. In this dissertation, I use InSAR and high-resolution topographic data to identify 50 slow-moving landslides in the Northern California Coast Ranges and monitor their kinematics over 4 years. These landslides have similar mechanical properties and are subject to the same external forcings, which allows me to explore geometrical controls on kinematics. Each landslide displays distinct kinematic zones with different mean velocities that remain spatially fixed. Because these deformation patterns are sensitive to subsurface geometry, I employ a mathematical model to infer landslide thickness and find that these landslides exhibit a highly variable thickness and an irregular basal sliding surface. Time series

analysis reveals that each landslide displays well-defined seasonal velocity changes with a periodicity of ~ 1 year. These velocity variations are driven by precipitation-induced changes in pore-water pressure that lag the onset of rainfall by up to 40 days. Despite significant variations in geometry, I find no systematic differences in seasonal landslide behavior. To further explore how stress perturbations control landslide motion, I develop a mechanical model that reproduces both the displacement patterns observed at slow-moving landslides and the acceleration towards failure exhibited by catastrophic events. I find that catastrophic failure can only occur when the slip surface is characterized by rate-weakening friction and its spatial dimensions exceed a critical nucleation length that is shorter for higher effective stresses. These model simulations support my conclusions from the remote sensing analysis but also provide insight into the long-term evolution of landslides.

This dissertation includes both previously published and unpublished co-authored material.

CURRICULUM VITAE

NAME OF AUTHOR: Alexander Louis Handwerger

GRADUATE AND UNDERGRADUATE SCHOOLS ATTENDED:

University of Oregon, Eugene, OR
Boston University, Boston, MA

DEGREES AWARDED:

Doctor of Philosophy, Geological Sciences, 2015, University of Oregon
Bachelor of Arts, Earth Science, 2008, Boston University

AREAS OF SPECIAL INTEREST:

Landslides, Landscape Evolution, Remote Sensing (InSAR and lidar), Gas Hydrates and Submarine Landslides.

PROFESSIONAL EXPERIENCE:

Graduate Teaching and Research Fellow, Department of Geological Sciences,
University of Oregon, 2009–2015

Directed Study, Department of Earth Sciences, Boston University 2007–2008

GRANTS, AWARDS AND HONORS:

Graduate Teaching Fellowship, Department of Geological Sciences, University of Oregon, 2009–2015

Johnston Scholarship Travel Grant, Department of Geological Sciences,
University of Oregon, 2011, 2013, 2014

PUBLICATIONS:

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For my family.

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CHAPTER I

INTRODUCTION

Landslides dominate erosion and landscape evolution in mountainous areas and pose a major natural hazard that causes significant damage to infrastructure and claims thousands of lives annually. A primary objective of landslide science is to quantify the mechanisms that control slope failure and develop mechanical models that can in turn be used to predict their behavior. However, landslides occur under a wide variety of environmental conditions and consequently can display a wide variety of behaviors, which has made it difficult to establish direct relationships between the processes that drive landslide motion and the processes that control their failure mode.

To better constrain the mechanisms that control slope failure requires high-resolution spatial and temporal observations of landslide kinematics from well-chosen field sites that allow for some landslide properties (e.g., geometry) to be varied while most are held constant (e.g., lithology, climate, tectonics). Few studies have made such comparisons because most previous studies have focused on individual landslides occurring in diverse geologic and climatic settings, using point-based measurements that cannot capture a complete view of landslide kinematics. Recent advances in remote sensing techniques, like satellite radar interferometry (InSAR), now enable high-resolution spatial and temporal measurements that can quantify the kinematic history along multiple landslides subject to nearly identical environmental forcings.

In this dissertation, I seek to further our understanding of the mechanisms that control landslide motion, in particular, interactions between stress perturbations and landslide velocity, by utilizing state-of-the-art remote sensing data to quantify

the kinematics of multiple slow-moving landslides occurring under the same environmental conditions. Slow-moving landslides are ideally suited for investigation because they persist for long time periods (Bovis and Jones, 1992; Mackey et al., 2009), display distinct kinematic responses to a variety of environmental stress perturbations (Iverson, 1986b; Iverson and Major, 1987; Coe et al., 2003; Schulz et al., 2009a), and have the potential to fail catastrophically (Iverson, 2005; Schulz et al., 2009b). Furthermore, because these features move primarily by sliding along frictional faults, there are obvious connections with other frictional processes like earthquakes and glaciers (Gomberg et al., 1995), which can be much harder to instrument and monitor.

In Chapter II, coauthored with Joshua Roering, David Schmidt (University of Washington), and Alan Rempel and accepted for publication in the journal *Geomorphology*, I use satellite InSAR and develop novel processing techniques to construct an inventory of active slow-moving landslides, quantify the kinematic evolution along the bodies of individual landslides in response to seasonal variations in precipitation, and apply a mechanical model to infer subsurface landslide geometry. These landslides share the same lithologic, tectonic, and climatic conditions, which allows me to explore the relationship between landslide geometry and kinematics. I present an inventory of 50 active slow-moving landslides and a detailed velocity time series analysis of four representative landslides that vary significantly in geometry and annual displacement rate. By exploring the seasonal kinematics along the landslide body, I am able to document how individual landslides respond to changes in rainfall (i.e. effective stress) and show that these landslides display a synchronous response indicating that geometry does not strongly regulate their response to seasonal stress perturbations. Lastly, I apply a viscous-flow model to explore the subsurface controls on landslide motion. Combining this model with the continuous spatial coverage

provided by InSAR, I am able to infer slide thickness and basal topography along a landslide. I find that both thickness and basal topography are highly variable, which may have major implications for controlling the motion and subsurface hydrology of these slope failures as well as their potential for catastrophic failure. This approach has never before been coupled with InSAR data, and provides a new method for characterizing the subsurface geometry of these landslides.

In Chapter III, coauthored with Joshua Roering and David Schmidt and published in the journal *Earth and Planetary Science Letters* in 2013, I examine geometrical controls on landslide kinematics in the context of a commonly used hydrologic model for landslides. According to model predictions, diffusion of pore-water pressure from the ground surface to the basal sliding surface controls a landslide's response to precipitation such that the length of time required for landslides to respond to rainfall should strongly depend on landslide size, and specifically thickness, whereby shallow landslides will respond before deep landslides. By exploring relationships between landslide response time, i.e. the time difference between the onset of precipitation and the onset of landslide acceleration, I show that the response is remarkably similar despite a five-fold variation in thickness. These findings challenge and potentially contradict assumptions made in many landslide models that are used for landscape evolution and hazard assessment, and encourage reevaluation of the mechanical-hydrologic mechanisms that control landslide dynamics.

In Chapter IV, coauthored with Alan Rempel, Robert Skarbek, Joshua Roering, and George Hilley (Stanford University), I develop a mechanical model to describe observations from the previous chapters and further explore how stress perturbations drive landslide motion and control landslide failure mode. The model is able

to accurately reproduce the displacement patterns observed at well-documented landslides. By varying the size, mechanical properties, and effective stress along the sliding surface, the model can produce a wide range of landslide behaviors, including stable sliding, seasonal motion, and catastrophic failure. Furthermore, I describe conditions under which a slow-moving landslide can transition to runaway acceleration and catastrophic failure. The results of these numerical experiments have implications both for how landslides respond to stress perturbations, and for their long-term evolution.

In Appendix D, co-authored with Alan Rempel, I develop a numerical model to identify the conditions under which the dissociation of gas hydrates can trigger or precondition submarine slopes for failure. I focus on the dissociation of high-concentration hydrate anomalies that have prevented the normal consolidation of marine sediments. I show that stable hydrate anomalies strengthen slopes due to increased cohesion and friction angle. However, once the hydrate is destabilized, sediment consolidation can generate excess pore-water pressure, which weakens sediment and potentially triggers slope failure.

Collectively, these chapters integrate state-of-the-art remote sensing and modeling techniques to better understand landslide behavior. This work can be used to help unravel the geometrical controls on landslide behavior and improve predictive models that enable hazard mitigation, describe landslide response to climate change, and elucidate their role in landscape evolution.

CHAPTER II

KINEMATICS OF EARTHFLOWS IN THE NORTHERN CALIFORNIA COAST RANGES USING SATELLITE INTERFEROMETRY

This work is accepted for publication in *Geomorphology* and co-authored with Joshua Roering, David Schmidt (University of Washington), and Alan Rempel. The work presented in this chapter reflects major findings and techniques that I developed to identify and monitor slow-moving landslides with satellite data. I am the first author, and the paper presents my data, methods, analysis, interpretations, figures, and writing. My co-authors Roering, Schmidt, and Rempel provided editorial assistance and help with the interpretation of my results.

2.1. Introduction

Slow-moving landslides exhibit persistent downslope motion with variations in velocity driven by transient stress perturbations (Terzaghi, 1951; Iverson, 1986b; Iverson and Major, 1987; Malet et al., 2002; Schulz et al., 2009a). These stress perturbations alter both the frictional resistance and driving stress along the sliding surface, and can induce complex time-dependent motion along the landslide body. Observations from slow-moving landslides around the world show that differences in motion occur over daily (Van Genuchten and De Rijke, 1989; Schulz et al., 2009a), seasonal (Kelsey, 1978; Iverson and Major, 1987; Coe et al., 2003; Hilley et al., 2004; Calabro et al., 2010; Handwerger et al., 2013), and multi-year timescales (Bovis and Jones, 1992; Mackey and Roering, 2011; Prokešová et al., 2014). Furthermore, the duration and magnitude of motion can vary along the slide body (Baum et al., 1993; Malet et al., 2002; Coe et al., 2003). These behavioral differences reflect how slow-

moving landslides respond to changes in stress over a variety of spatial and temporal scales.

In this manuscript we focus on the kinematic evolution of slow-moving landslides in response to seasonal variations in precipitation. Slow-moving landslides regularly exhibit seasonal velocity patterns, yet it remains unclear how periods of seasonal motion evolve along a landslide body throughout the year. For instance, in some years, periods of acceleration can initiate in isolated zones and propagate along the slide body (referred to as “unsteady, nonuniform motion”), while in other years acceleration can occur simultaneously along the entire slide body (referred to as “unsteady, uniform motion”) (Iverson, 1986b; Coe et al., 2003). Given observations outlining the diverse nature of slide behavior, what is the prognosis for predicting how landslides respond to transient changes in stress? In one case, a landslide forced by local or boundary perturbations like headscarp slumping, toe erosion, or local groundwater changes may display progressive motion that travels along the landslide body (Iverson, 1986b). By contrast, spatially extensive stress perturbations resulting from changes in the seasonal groundwater level may induce spatially extensive velocity changes. To help differentiate these scenarios, Iverson (1986b) developed a quantitative framework to predict how both local and spatially extensive stress perturbations are communicated along a slow-moving landslide. According to his framework, local stress perturbations induce unsteady, nonuniform motion that spreads along the landslide body and spatially extensive stress changes produce unsteady, uniform motion. These model predictions agree with field-based observations made at the Aspen Grove landslide in Utah and the Slumgullion landslide in Colorado (Baum et al., 1993; Coe et al., 2003). Baum et al. (1993) found that the main body of the Aspen Grove landslide responded to snowmelt weeks before the toe, and the

main body continued moving for a month after motion in the toe had ceased. Similar observations were made at the Slumgullion landslide during one year; however, during other years the landslide displayed a more synchronous response to snowmelt. Coe et al. (2003) attributed these disparate behaviors to spatial variations in snowmelt and groundwater levels (i.e. local stress perturbations), which can fluctuate across a single landslide.

Landslide geometry also exerts significant control on landslide motion and dictates how stresses are communicated along the slide body (Iverson, 1986b; Baum and Fleming, 1991; Van Asch et al., 2006; Coe et al., 2009; Guerriero et al., 2014; Prokešová et al., 2014). For instance, slope and thickness can vary along an individual landslide such that there is a heterogeneous distribution of shear stress along the slide base. Along a number of landslides, surface measurements are complemented by observations from boreholes and seismic data that map irregularities in the basal slip surface (Mizuno, 1989; Baum and Johnson, 1993; Coe et al., 2009; Guerriero et al., 2014; Prokešová et al., 2014). These irregularities can be used to explain commonly observed surface deformation features and potential controls on the long-term kinematics of these slope failures (Baum and Fleming, 1991; Baum and Johnson, 1993; Coe et al., 2009; Guerriero et al., 2014; Prokešová et al., 2014). By measuring the spatial kinematics of multiple slow-moving landslides subject to the same environmental and tectonic forcing, we can characterize how stresses are communicated along landslide bodies, attempt to discriminate between the effects of local and spatially extensive stress perturbations, and better understand the geometrical controls on landslide motion, including spatial variations in the measured slope and inferred thickness.

Here, we use satellite InSAR with data from the Advanced Land Observing Satellite 1 (ALOS-1) and high-resolution topographic data, including airborne lidar, to identify slow-moving landslides in the Northern California Coast Range and monitor their kinematics between February 2007 and January 2011. Because these landslides have similar mechanical properties (i.e. lithology) and are subject to the same external forcings (i.e. climate, incision), we are able to explore interactions between landslide geometry and kinematics. This builds upon the results of previous work by Handwerger et al. (2013) in which they quantified the velocity time series and seasonal response times (i.e. time lag between rainfall and acceleration) for the fastest moving zones of 10 landslides (a subset of these slides will be discussed below) in order to test predictions of a commonly used hydrological model. In this contribution, we 1) develop and test novel InSAR processing methods that are necessary to construct time series inversions of these landslides, 2) perform a reconnaissance-level survey to identify actively moving landslides across a $\sim 14,000 \text{ km}^2$ area, 3) examine, for the first time in a regionally extensive study, the spatial and temporal evolution of discrete kinematic zones along landslides that are subject to seasonal changes in effective normal stress, and 4) infer patterns of thickness changes and basal topographic variations using a non-Newtonian viscous flow law.

2.2. Slow-moving Landslides

Slow-moving landslides commonly occur in tectonically active regions with mechanically weak, clay-rich soils, and highly seasonal precipitation. In these areas, they have a profound impact on landscape evolution because they control relief, widen valley spacing, modify drainage patterns, and set hillslope morphology (Kelsey, 1978; Booth et al., 2013b; Simoni et al., 2013). They occur in a small fraction ($< 10\%$) of the

landscape, and contribute a large fraction (up to 50%) of the regional sediment flux (Kelsey, 1978; Mackey and Roering, 2011; Simoni et al., 2013). These slope failures tend to span from hilltop to channel bottom creating kilometer-scale planar hillslopes with gentle topographic gradients (Keefer and Johnson, 1983; Booth and Roering, 2011). All of these factors combine to leave a distinct topographic signature on the landscape, which can be identified through quantitative analysis of high-resolution DEMs (Booth and Roering, 2011; Booth et al., 2013b). Additionally, slow-moving landslides are a major natural hazard, and although they seldom fail catastrophically, they cause significant damage to infrastructure.

Here, we focus on large (km-scale), deep-seated (> 5 m), slow-moving (m yr^{-1}) landslides often called earthflows (Keefer and Johnson, 1983; Hungr et al., 2001; Mackey and Roering, 2011). Although these slides display a flow-like surface morphology, most movement occurs by frictional sliding along discrete shear surfaces. In this manuscript, we use the terms landslide, slide, slow-moving landslide, and earthflow interchangeably. Earthflows exhibit a wide range of sizes and shapes, but they are most often identified by a hummocky surface, and hourglass planform (Keefer and Johnson, 1983; Mackey and Roering, 2011; Simoni et al., 2013). Their spatial dimensions can range over two-to-three orders of magnitude, with the largest slides reaching up to several kilometers in length, hundreds of meters in width, and tens of meters in thickness. Although their planform areas are relatively easy to delineate with field-observations and high-resolution DEMs, their thicknesses are often poorly constrained because most measurements of landslide thickness come from individual boreholes, hillslope-channel interfaces, or gullies incised through the landslide body, which in many cases do not intersect the basal shear zone. Recent work has been aimed at developing depth-area scaling relationships for earthflows and has found that

they tend to be slightly deeper relative to their area when compared to other types of landslides composed of unconsolidated materials (Larsen et al., 2010; Handwerger et al., 2013; Simoni et al., 2013).

Other studies have used measurements of surface deformation to infer the subsurface geometry of slow-moving landslides (Baum et al., 1993; Coe et al., 2009; Guerriero et al., 2014). Recently, Guerriero et al. (2014) used surface deformation features (e.g., cracks, faults, tilted surfaces) and the location of spatially fixed ridges and depressions (e.g., sag ponds, springs) on the ground-surface to infer basal topography. They confirmed their results with numerous borehole measurements and seismic data. Several studies have also used 2-D and 3-D mass conservation techniques to estimate changes in landslide thickness (e.g., Bishop, 1999; Booth et al., 2013a). Each of these studies has concluded that slow-moving landslides have an irregular (i.e. non-planar and rough) basal-slip surface and exhibit significant changes in thickness across the landslide body. These findings have major implications for field instrumentation design and mitigation strategies (Hutchinson, 1970), estimates of landslide volume, and for the development of slope stability models, which most often assume a smooth and listric or planar shear surface. Furthermore, the form of this basal interface may be a primary control on pore-water pressure-deformation feedbacks (Baum and Johnson, 1993; Van Asch et al., 2006).

Earthflows exhibit nearly plug-flow deformation by sliding along basal and lateral shear zones, often with minimal internal deformation. Observations from borehole inclinometers reveal that the majority of the shear deformation is accommodated along millimeter- to meter-scale shear zone thicknesses, while the near surface translates like a rigid block (Swanson and Swanson, 1977; Vulliet and Hutter, 1988; Simoni et al., 2013). This behavior can be approximated by a viscous flow model

with a non-Newtonian rheology. Using this approach, several studies have successfully modeled earthflows over both short and long timescales (Vulliet and Hutter, 1988; Booth et al., 2013b). Most recently, Booth et al. (2013b) parameterized a viscous flow model using borehole inclinometer data from multiple earthflows around the world and incorporated it into a landscape evolution model that was able to reproduce many of the key topographic features (e.g., relief, topographic slope, valley spacing) of earthflow-dominated landscapes.

Earthflows are often partitioned into three main kinematic zones, namely: the head, transport, and toe zones. Velocities tend to be highest through the transport zone and decrease towards the head and toe zones. The spatial locations of these kinematic zones remain fixed over long time periods (10^1 – 10^2 years), which suggests that these regions are set by the hillslope and landslide geometry (Coe et al., 2009; Mackey et al., 2009; Guerriero et al., 2014). Average slide velocities range from a few decimeters to several meters per year, but can be variable over daily-to-decadal timescales (Iverson and Major, 1987; Bovis and Jones, 1992; Schulz et al., 2009a; Mackey and Roering, 2011; Handwerger et al., 2013; Prokešová et al., 2014).

Seasonal velocity changes are driven by hydrologic forcing, like rainfall and snowmelt, that acts to increase pore-water pressure along the shear zone, which reduces frictional resistance and triggers acceleration (Terzaghi, 1951; Iverson and Major, 1987; Coe et al., 2003; Handwerger et al., 2013). Deceleration begins as pore-water pressures decline due to drainage, precipitation decreases, and potentially as a result of other mechanical feedbacks like granular dilation (Iverson, 2005; Schulz et al., 2009b) or frictional rate-strengthening (Wang et al., 2010). In contrast to shallow landslides, which are often triggered by short periods of intense precipitation, deep-seated earthflows typically require longer (i.e. seasonal) periods of precipitation to

increase pore-water pressures in the basal shear zone to levels that trigger detectable motion (Iverson and Major, 1987). Field-based measurements and numerical models have shown that transient variations in pore-water pressure diffuse vertically from the ground surface to the basal shear zone (Iverson and Major, 1987; Haneberg, 1991; Reid, 1994; Baum and Reid, 1995; Berti and Simoni, 2010, 2012). This theoretical framework implies that for landslides with similar precipitation, material properties, and topographic slopes, the amount of time required for pore-water pressure changes to occur in the basal shear zone should scale with the slide thickness. However, Handwerger et al. (2013) found that the transport zones for 10 slow-moving landslides (a subset of the same slides discussed in this study) that vary by a factor of 5 in estimated thickness respond to seasonal precipitation with a remarkably similar timing and magnitude. This observation either contradicts the predictions of the simple 1D diffusion model or provides evidence that the effective hydraulic parameters are outside the range of commonly assumed values for clay-rich soils. In addition, the lateral (i.e. downslope) transmission of pore-water pressure may be important in triggering landslide motion (Priest et al., 2011).

2.3. Study Area: Northern California Coast Range

Our InSAR study covers an area of $\sim 14,000$ km² in the Northern California Coast Ranges (Figure 2.1). In particular, we focus on the mainstem of the Eel River, which is well known for its earthflow activity and has been the focus of many landslide investigations in recent years (Roering et al., 2015). The Northern California Coast Range is a tectonically active mountain belt that extends northwest from the San Francisco Bay to the Oregon border. The region has a Mediterranean climate with cool, wet winters and hot, dry summers. Average annual rainfall is 1.4 m yr⁻¹, falling

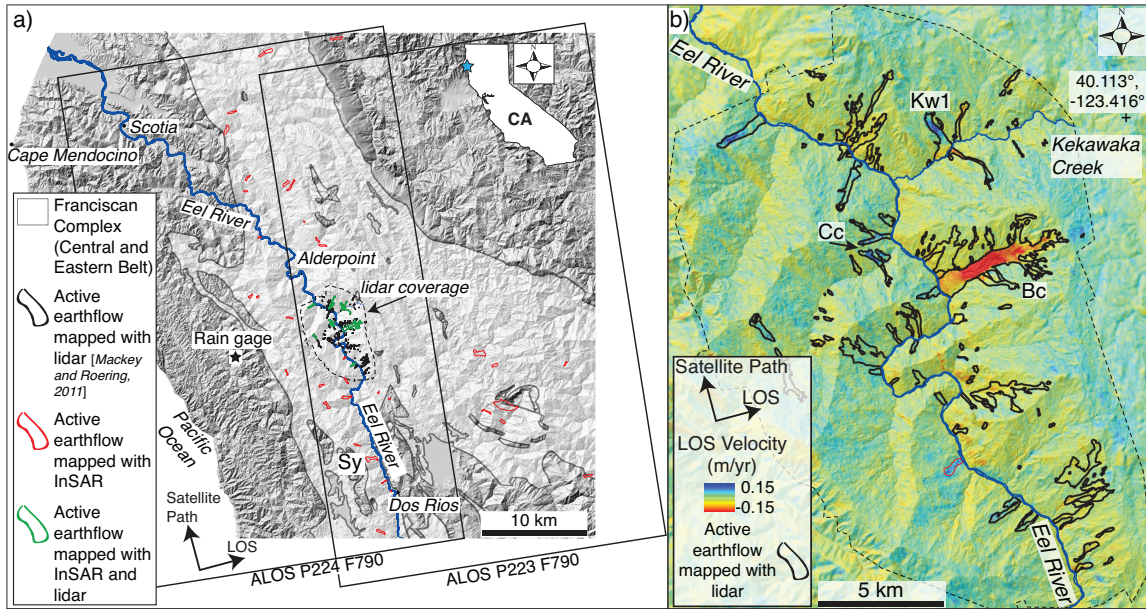


FIGURE 2.1. Landslide inventory. a) Hillshade map of Northern California Coast Range field site. Black rectangles outline ALOS PALSAR data coverage for satellite paths 223-F790 and 224-F790. Blue line highlights the mainstem of the Eel River. Light gray area outlines the Eastern and Central belts of the Franciscan Complex (Jennings et al., 1977). Colored polygons show earthflows mapped with lidar (Mackey and Roering, 2011), InSAR, and both InSAR and lidar, respectively. Inset in upper right corner shows California with blue star marking the location of the field area. Black star marks location of NOAA rain gage. b) A 46-day interferogram stack draped over lidar hillshade. Positive velocity values correspond to motion away from the satellite along the line-of-sight. Black polygons show previously mapped earthflows (Mackey and Roering, 2011). Earthflow names indicate the slides examined in our time series analysis and are abbreviated as the following: Bc = Boulder Creek, Cc = Chamise Creek, Sy = Simmerly Road, and Kw1 = Kekawaka Creek 1.

mostly between October and April and reaching peak rates between December and February. Vegetation is a mixture of open oak grassland and conifer forest.

The Coast Range is composed of the Jurassic-Cretaceous Franciscan Complex, a pervasively sheared accretionary prism that has been uplifting since the Miocene. The Franciscan Complex is separated into three distinct structural belts that decrease in age to the West, named the Coastal, Central, and Eastern belts (Jayko et al., 1989; McLaughlin et al., 2000). The Central belt, which is particularly susceptible to

earthflow-style landslides, is comprised of an argillaceous *mélange* matrix surrounding more coherent blocks of sandstone, chert, shale, and older metamorphosed units (e.g., meta-basalt, blueschist) inherited from the Eastern belt. These coherent blocks play an important role in the evolution of this landscape because they influence the local hillslope gradient, impact earthflow behavior, create river knickpoints, and have the potential to fail as large, catastrophic, river-damming landslides (Mackey et al., 2011).

Due to its high erosion rates, the Eel River catchment has been widely studied since the late 1960s. Brown and Ritter (1971) found that the Eel River had the highest average sediment yield per drainage area of any non-glacial or volcanic river in the conterminous United States. Since then, many studies have quantified erosion rates using a variety of tools, including suspended sediment and hydrologic data (Wheatcroft and Sommerfield, 2005) and cosmogenic radionuclides (Fuller et al., 2009; Balco et al., 2013; Roering et al., 2015). Each of these studies has corroborated the catchment's high erosion rates, with values from 0.3 to 1.1 mm yr⁻¹. The majority of the total erosion can be directly attributed to active landsliding. Kelsey (1978) and Mackey and Roering (2011) found that active earthflows contribute up to 50% of the regional sediment flux to channels.

Although many studies have investigated erosion rates in our field area, estimates of rock uplift are sparse and indirect (Furlong and Govers, 1999; Lock et al., 2006). The Northern California Coast Range is actively uplifting due to crustal thickening and dynamic topography resulting from the northward passage of the Mendocino Triple Junction (MTJ). Geomorphic evidence (e.g., river capture, drainage reversal) (Lock et al., 2006) and a geodynamic model (Furlong and Govers, 1999) suggest that this region is experiencing a double-humped zone of rock uplift that is riding along with the MTJ. If true, this pattern of rock uplift may have implications for the

frequency and magnitude of landslides (Roering et al., 2015). Predicted estimates of uplift near our field area, which is roughly 100 km E-SE of the MTJ, range from 0.5 to 1 mm yr⁻¹ (Lock et al., 2006). Additional estimates of rock uplift from marine terraces in the King Range, which is much closer to the MTJ, approach 5 mm yr⁻¹ (Merritts and Bull, 1989); however, the King Range is regarded as a tectonically distinct region and uplift rates are known to decay inland, coincident with a shear zone that surrounds the range (McLaughlin et al., 2000; Lock et al., 2006).

2.4. Methodology

The dynamics of slow-moving landslides are governed by complex mechanical-hydrological interactions between landslide properties and external forcings (e.g., precipitation, incision). Unraveling these interactions requires high-resolution spatial and temporal observations from well-chosen sites that allow for some landslide properties to be varied (e.g., geometry) while most are held constant (e.g., climate, tectonics, material properties). Recent advances in remote sensing techniques, like satellite radar interferometry (InSAR), now enable high-resolution spatial and temporal measurements that can quantify the kinematic history along multiple landslides subject to nearly identical environmental forcings. The resulting constraints can be used to quantify how landslide geometry influences slide behavior.

In the past two decades, InSAR has become a well-established technique for monitoring landslide displacement, taking advantage of the increased availability of SAR data from multiple satellites and the development of advanced processing techniques, such as permanent scatterers InSAR (PS-InSAR) (Ferretti et al., 2001; Colesanti and Wasowski, 2006). There are now SAR data available from more than 10 satellites, spanning over two decades, and covering almost the entire globe. In

addition, ground-based and Uninhabited Aerial Vehicle (UAV) radar interferometers are becoming more common, both of which allow for highly targeted investigations (Schulz et al., 2012; Lowry et al., 2013; Scheingross et al., 2013).

While numerous studies have used InSAR to construct regional landslide inventories (Zhao et al., 2012; Scheingross et al., 2013), assess potential hazards (Castañeda et al., 2009), and examine the kinematics of individual slides (Calabro et al., 2010), few have compared and contrasted the time-dependent displacement of multiple neighboring landslides (Hilley et al., 2004; Handwerger et al., 2013). This is likely a result of the high quantity of interferograms required to construct InSAR time series inversions (Schmidt and Bürgmann, 2003), which can be difficult to attain because InSAR studies in landslide-prone areas are often challenged by large deformation gradients, steep and vegetated topography, the one-dimensional viewing geometry of the satellite, and atmospheric noise from frequent precipitation (Colesanti and Wasowski, 2006).

Here, we use conventional 2-pass InSAR to quantify the kinematics of slow-moving earthflows in the Northern California Coast Range. Previous studies using InSAR in this region have demonstrated the capability of the method to identify and monitor multiple active landslides that span an order of magnitude in size and velocity (Roering et al., 2009; Zhao et al., 2012; Handwerger et al., 2013). The Northern California Coast Range is an ideal location for using conventional 2-pass InSAR because earthflow motion is fast enough to observe over a short time period and slow enough to avoid radar decorrelation. Although these earthflows are particularly well suited for conventional InSAR techniques, their persistent downslope motion is problematic for producing long-duration interferograms because conventional InSAR requires that the change in the phase from one pixel to the next be $< 2\pi$ radians ($<$

11.75 cm for ALOS-1). Since many of the landslides in our field area are moving on the order of m yr^{-1} , deformation associated with long-duration (> 46 days) interferograms often exceeds this phase-change threshold and the deforming area becomes decorrelated or contains unwrapping errors (Stimely, 2009). In Section 2.4.3, we describe a scalable deformation model, which allowed us to overcome these limitations so that we can produce deformation time series and explore geometrical controls on landslide kinematics.

2.4.1. InSAR Methodology

We use conventional 2-pass InSAR with data from the ALOS-1 satellite to identify active earthflows between February 2007 and January 2011. SAR data were acquired with the PALSAR instrument, which operates with an L-band antenna (23.6 cm wavelength) and a 46-day repeat interval. This satellite has a look direction of 075° (east of North) and incidence angle of 34.3° from vertical. Our study area is covered by overlapping satellite path 223 frame 790 (P223) and path 224 frame 790 (P224) (Figure 2.1a), whose data acquisition times are offset by a few weeks (Table A.1). Due to technical issues, the ALOS-1 satellite was decommissioned in May 2011.

We produced 81 differential interferograms using 17 scenes from P223, and 21 scenes from P224 (Table A.1). Interferograms were processed with the Repeat Orbit Interferometry Package (ROI-PAC) developed at JPL/Caltech (Rosen et al., 2004). We used a 1 arc-sec (30 m resolution) DEM from the Shuttle Radar Topography Mission (SRTM) (Farr et al., 2007) to remove topographic contributions to the phase. SAR data were processed at 4-looks in range and 20-looks in azimuth, and resampled at the resolution of the DEM upon orthorectification (~ 30 m pixel). To reduce noise, we used a power spectral filter value between 0.1 and 0.3 (Goldstein and Werner,

1998). Although other SAR satellites (e.g., ERS1/2, ENVISAT, RADARSAT) cover a longer time period and have shorter repeat intervals than ALOS-1, attempts to image landslide deformation with these C-band (5.6 cm) satellites were unsuccessful, likely because their shorter wavelength could not sustain coherence in our vegetated study area (Stimely, 2009).

2.4.2. Landslide Reconnaissance Using InSAR

To identify active landslides, we performed a reconnaissance-level investigation of both satellite paths by stacking interferograms (i.e. time-averaged deformation) and using statistical analysis to discriminate deformation from artifacts. Stacking increases the signal-to-noise ratio and highlights features that are persistent in time and space. We found that patches with high line-of-sight (LOS) velocity and high standard deviation of LOS velocity are effective criteria for identifying active slides. In addition, we overlaid the InSAR stacks onto a shaded relief map to verify that these patches correspond to landslide topography (Figure 2.1b). To directly compare landslides to each other and to previous studies, we back-projected LOS velocity onto the downslope direction using a vector projection (see Appendix A). We use both LOS velocity and downslope velocity to analyze data in the forthcoming sections. Lastly, to quantify the spatial attributes of each slide we used high-resolution DEMs generated from lidar (1 m pixel) and from the USGS National Elevation Dataset (10 m pixel) (Gesch et al., 2002).

2.4.3. Deformation Model for Processing Long-duration Interferograms

The persistent downslope motion of the Eel River slides is problematic for producing long-duration (> 46 days) interferograms. By using a deformation model,

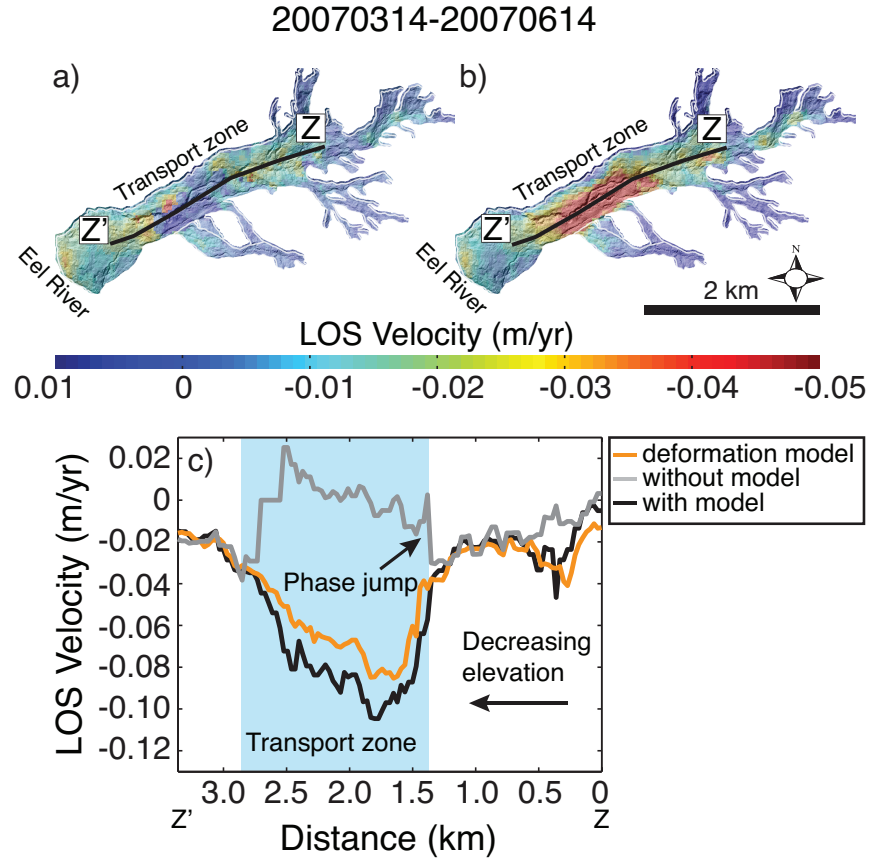


FIGURE 2.2. A 92-day interferogram of the Boulder Creek earthflow draped on lidar hillshade. Interferogram processed without the deformation model contains a phase jump due to an unwrapping error that is caused by the large deformation gradient at the margin of the transport zone. Red and blue colors indicate landslide motion towards and away from the satellite, respectively. a) LOS velocity map processed without deformation model. b) LOS velocity map processed with deformation model. c) LOS velocity profile from Z to Z' using values from a) and b) and the deformation model. Shaded blue area marks the spatial location of the transport zone.

which is subtracted from the raw interferogram, we are able to maintain coherence and minimize unwrapping errors for interferogram durations up to 276 days. The model is constructed from a stack of 46-day interferograms for each individual landslide that is scaled to the duration of the interferogram being processed. In other words, if we are processing a 92-day interferogram, we multiply the deformation model by a factor of 2. Thus, the model represents the average deformation of each landslide over the specified time period. This allows us to recover deformation beyond the scaled deformation model by maintaining coherence (Bürgmann et al., 2006; Wei et al., 2010; Tong et al., 2013). After the interferogram is filtered and unwrapped, the deformation model is added back into the interferogram and processing proceeds. This technique works well because it prevents the phase gradient from exceeding 2π radian (11.75 cm) per pixel (the threshold where coherence is lost), while allowing for the residual deformation to be processed (Figure 2.2). We have carefully compared the unwrapped interferograms to the original unfiltered wrapped interferograms to ensure that no artifacts are introduced by this process (Figure A.2).

2.4.4. InSAR Time Series

InSAR is capable of resolving complex time-dependent deformation patterns if redundant data can be used to differentiate signal from noise (Schmidt and Bürgmann, 2003). This requires a temporally rich dataset with multiple interferograms that overlap in time. To increase both the temporal resolution and number of interferograms used in the inversion, we selected four landslides that lie within the overlapping regions of satellite paths 223 and 224 (Figure 2.1a). We do not compensate for the slightly different look angle for common pixels on overlapping paths. We forward calculate the change in phase that results from differences in

the look angle and estimate that this introduces a LOS error of $< 5\%$. Using the method of Schmidt and Bürgmann (2003), we calculated the velocity time series for the four landslides by performing a linear least-squares inversion of 51 independent, unwrapped interferograms (Table A.2). In order to minimize topographic contributions to the phase and to preserve nearly complete spatial coverage of each landslide, we selected interferograms with short spatial baselines (< 1400 m) and minimal decorrelation. We are unable to process interferograms during summer 2008 because of an issue with the satellite acquisition system that produced unusually large spatial baselines and resulted in completely decorrelated interferograms.

2.4.5. Synthetic Deformation Time Series

Because we have no complementary measurements of seasonal landslide motion to compare against our InSAR time series results, we constructed a synthetic deformation time series to rigorously test our time series methodology and to identify potential sampling biases. Our synthetic time series was made using simulated SAR scenes with two populations of pixels. One population of pixels was made using a simple surface-displacement model to simulate general patterns of earthflow motion over the 4-year study period. The other population of pixels used in the simulated SAR scenes experienced no deformation and was used as a stable reference point in the time series inversion. We define the surface displacement model using

$$D = \bar{V}t + A \sin(2\pi t/\lambda) \quad (2.1)$$

where D is the cumulative displacement, $\bar{V}$ is the average annual displacement rate, t is time, A is the sinusoidal wave amplitude (i.e. magnitude of seasonal velocity changes), and λ is the sinusoidal wavelength (i.e. wavelength of seasonal velocity changes). The linear deformation term and a sinusoidal deformation term are chosen to simulate annual and seasonal displacement, respectively. We tested a wide range of annual velocities ($\bar{V} = 0.2$ to 2 m yr^{-1}), wave amplitudes ($A = 0.002$ to 0.15 m), and wavelengths ($\lambda = 0.1$ to 4 yr). We also added correlated random noise with a normal distribution (mean = 0 m , std = 0.01 m) to approximate atmospheric artifacts (Lohman and Simons, 2005). The magnitude of our synthetic error source was validated by checking the apparent deformation of areas known to be stable (e.g., hilltops) in our interferograms. The model was sampled to match the temporal distribution of the InSAR dataset and run through the time series inversion.

First, we tested the sensitivity of the inversion to the temporal distribution of interferograms. We found that by using all of the small-baseline interferograms from satellite path 223 and 224 (“P223-P224 combined”) (Table A.1), the inversion output contained high frequency oscillations that were not included in the input model (Figure A.3). This inversion misfit is a result of the high number of SAR scenes relative to the number of interferograms, which can destabilize and decrease the resolution of the inversion (Schmidt and Bürgmann, 2003).

To reduce the misfit error while maintaining high data redundancy, we manually adjusted the acquisition dates of 20 SAR scenes that were collected < 20 days apart, to the midpoint date of the acquisitions (“P223-P224 modified”) (Figure A.3; Table A.3). For example, SAR scenes with acquisition dates 20090804 and 20090821 were adjusted to 20090812. Because only minor deformation ($< 2 \text{ cm}$) occurs over the < 10 day period (i.e. midpoint of < 20 days) this does not affect the seasonal landslide signal.

In addition to the individual SAR scenes collected within 20 days of each other, there are 13 interferogram pairs that have acquisition dates < 20 days apart (Table A.1). For these interferograms, we adjusted their acquisition dates and used the average of their deformation signals. By making this adjustment to the scene acquisition dates, the inversion output did not contain high frequency oscillations and more closely matched the synthetic input signal (Figure A.3). For the remainder of our analysis we use the “P223-P224 modified” distribution of interferograms.

Lastly, we used the synthetic interferograms to determine the minimum detectable seasonal wavelength of our irregularly sampled InSAR dataset. We tested synthetic models with sinusoidal wavelengths ranging from 0.25 to 4 years and found that the misfit decreased nonlinearly with increasing wavelength such that we can resolve seasonal signals with a wavelength ≥ 0.5 years (Figure A.4).

2.4.6. Non-Newtonian Viscous Flow Model to Infer Slide Thickness

To explore subsurface controls on landslide motion we inferred landslide thickness using a viscous flow model with a non-Newtonian rheology. The viscous flow model provides an approximate relationship between surface velocity, topographic slope, and earthflow thickness at a given pixel i , such that

$$V_i = \frac{(c\rho g S_i)^p H_i^{p+1}}{p + 1} \quad (2.2)$$

where V_i is the downslope velocity, S_i is the sine of the topographic slope-angle, H_i is the earthflow thickness, c is a flow law constant, ρ is the earthflow density, g is gravitational acceleration, and p is the flow-law exponent (Booth et al., 2013b). Parameter values are listed in (Table A.4). We rearrange Eq. (2.2) to write the

thickness at each pixel as

$$H_i^{p+1} = \frac{V_i(p+1)}{(c\rho g S_i)^p}. \quad (2.3)$$

Given velocity values measured with InSAR and slope values measured along the slide axis with DEMs, Eq. (2.3) can be used to infer continuous profiles of slide thickness.

Before proceeding, it is worth noting that the viscous flow model does not explicitly describe the mechanics of frictional sliding that are believed to be primarily responsible for the observed landslide deformation. This recognition has led to considerable debate over whether it is appropriate to use viscous flow models to characterize landslide (and subglacial deforming till) motion (Hungr, 1995; Iverson, 2005, 2010). An attractive feature of the viscous flow model is the clear and simple correspondence predicted between higher driving stresses and faster surface motions. This has enabled such treatments to provide useful characterizations of landslide motion over a variety of spatial and temporal scales (e.g., Vulliet and Hutter, 1988; Booth et al., 2013b). More elaborate frictional treatments that incorporate a rate dependence through considerations of dilatancy and associated pore pressure changes yield predicted landslide behavior that is reported to strongly resemble motion resisted by viscous deformation (Iverson, 2005). Nevertheless, it should be emphasized that our viscous flow model is not intended to reveal precise soundings of flow depth, but instead only to uncover broad patterns of variability in the basal geometry.

2.5. Results

2.5.1. Landslide Inventory

We identified 50 active landslides across the $\sim 14,000$ km² area covered by the two satellite paths (Figure 2.1). The landslides have spatial dimensions ranging from 0.4

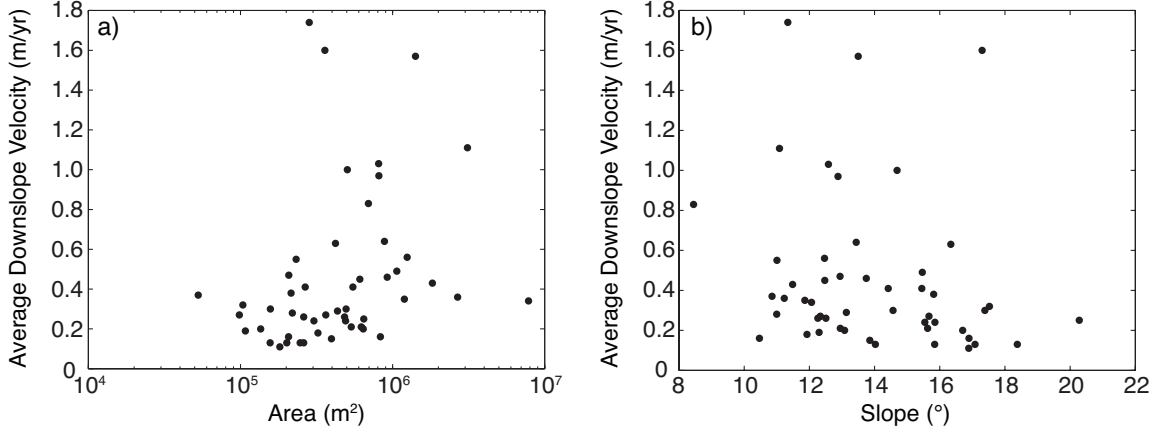


FIGURE 2.3. Projected downslope velocity for 50 slow-moving landslides as a function of a) landslide area and b) average topographic slope. Average slope is measured along the longitudinal axis of each slide. We find no clear relation between average landslide size, average topographic slope, and average slide velocity. See Table A.5 for detailed statistics.

to 5.4 km long, 0.1 to 1.4 km wide, 0.05 to 7.8 km² in planform area, and have average downslope velocities ranging from 0.1 to 1.7 m yr⁻¹ (Table A.5). All but one of the 50 earthflows are located within the Central and Eastern belt of the Franciscan Complex (Figure 2.1), which indicates a strong lithologic control on their spatial distribution, consistent with the results of Mackey and Roering (2011), who found that the majority of the earthflows in this region occur within a mechanically weak and argillaceous unit of the Franciscan Complex. There is no clear relation between landslide size, average topographic slope, and average velocity (Figure 2.3). Each slide has three main kinematic zones which we refer to as the head, transport, and toe zones. Peak velocity values occur in the transport zone and decrease towards the head and toe zones. Figure 2.4 displays annotated interferograms, highlighting each kinematic zone, for four representative earthflows. We delineate these zones based on surface velocities and deformational structures identified with the DEM. We expect these slides can be subdivided into smaller quasi-discrete zones (e.g., Guerriero et al., 2014),

but this requires detailed field investigation or multi-temporal high-resolution DEMs. For the 13 slides that occur within the area of lidar coverage (previously mapped by Mackey and Roering, 2011), we observe a similar average annual velocity and find the kinematic zones have remained spatially fixed since at least 1944, suggesting the locations of these zones are primarily set by the landslide geometry. These earthflows also display cross-slide variations in velocity, which confirms a flow-like (i.e. lateral strain) component (Figure 2.4).

It is important to note that InSAR is blind to deformation that occurs along the satellite’s flight path and any landslides with dominant motion in the NW-SE direction remain undetected (see Appendix A.1.2). In addition, there is a bias towards larger and slower landslides because smaller and faster slides (e.g., debris flows) cover only a few pixels of the InSAR data and are limited by deformation thresholds. Therefore, our landslide inventory provides a minimum estimate of the total number of active landslides in this region between 2007 and 2011.

2.5.2. Velocity Time Series

We calculated the velocity time series for four representative earthflows that lie within the overlapping regions of satellite paths 223 and 224 and have minimal decorrelation (Figure 2.4). By combining SAR data from both satellite paths we have achieved a temporal sampling ranging from 29 to 193 days with a mode of 46 days (Table A.2). These earthflows range in area from 0.16 to 3.1 km², estimated average thickness from 15 to 40 m, and average downslope velocity from 0.3 to 1.2 m yr⁻¹. To characterize general patterns of seasonal motion, we collapsed the 4-years of InSAR data from the transport zone of each slide into a single calendar year and plotted their velocity by day of the year (Figure 2.5). Consistent with the regional

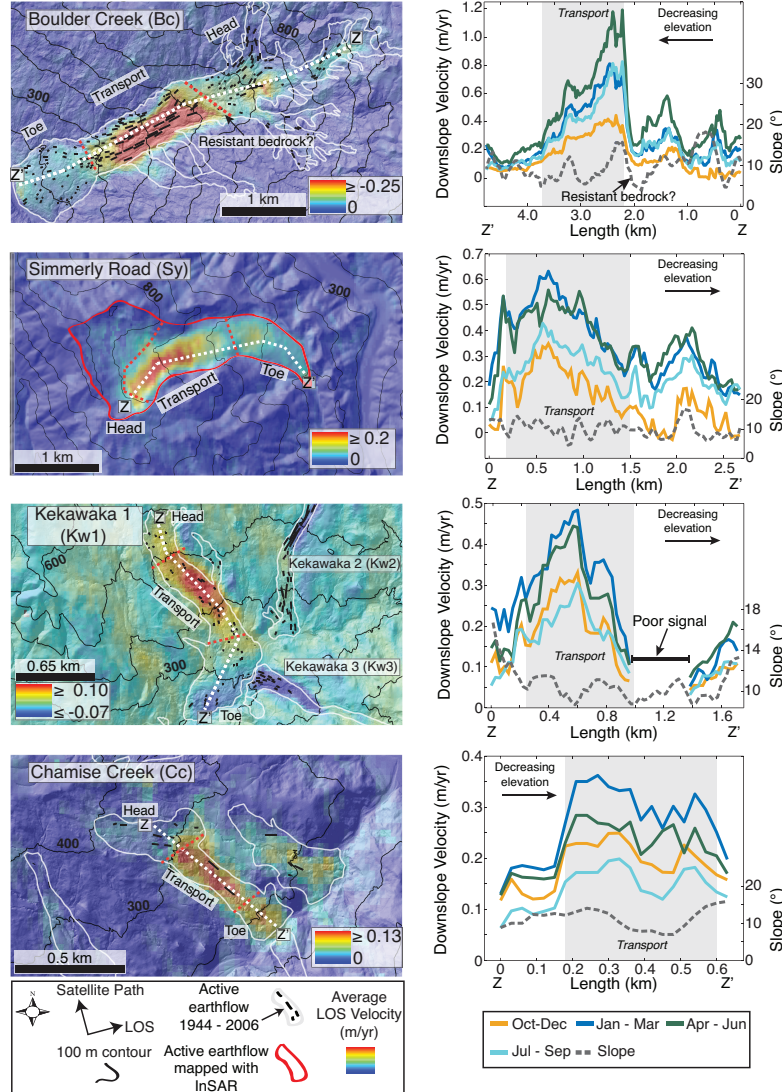


FIGURE 2.4. 4-year averaged interferograms draped over hillshade maps (1 and 10 m) for four representative earthflows modified from Handwerger et al. (2013). Velocity maps show LOS velocity values with red colors marking maximum LOS velocities. Note that the color scale is reversed for the Boulder Creek earthflow in order to highlight the transport zone with red colors. The color scales are saturated such that the actual velocity values (m yr^{-1}) exceed the maximum and minimum listed values. White polygons show landslide boundaries mapped by Mackey and Roering (2011). Short black lines show landslide motion mapped using historic aerial photos (Mackey and Roering, 2011). White dashed lines indicate locations of velocity and topographic slope profiles. Red dashed lines mark location of kinematic zones. Each slide is delineated into head, transport, and toe zones. Profiles from Z to Z' showing topographic slope and seasonally averaged velocity. Shaded gray area marks the spatial location of the transport zone. The poor signal indicated along the Kw1 slide is explained in Appendix A.1.2.

precipitation patterns, the slides exhibit well-defined seasonal velocity changes that are characterized by relatively short periods of acceleration lasting $\sim 30\%$ of the year, followed by longer periods of deceleration lasting $\sim 70\%$ of the year (Figure A.5). Each earthflow accelerates smoothly to wet-season rates, lagging the onset of rainfall by a maximum of 1 to 2 months (Handwerger et al., 2013), and then decelerates smoothly to dry-season rates. We observed no true steady-state motion over the 4-year study.

Next, we analyzed seasonal kinematics along the longitudinal axis of the four earthflows. To examine how the seasonal velocity changes occur within each kinematic zone, we binned the velocity time series data into 3-month intervals starting at the onset of seasonal rainfall in October (Figure 2.4; Figure A.6). All parts of the slide, to the sub-kinematic zone scale, display seasonal velocity changes that scale in proportion throughout the year. That is, the fastest moving region remains as such in both wet and dry seasons. We also observe distinct differences in seasonal sliding rate between each kinematic zone. Transport zone rates can exceed the head and toe zone rates by a factor of 4, while the head and toe zone rates are within a factor of 2. Although the transport zone maintains persistent downslope motion year round, during the dry season the head and toe zones either come to a halt or move too slowly ($< \text{cm yr}^{-1}$) for InSAR to detect significant motion.

We also analyzed the timing of these seasonal velocity changes in an attempt to differentiate between uniform and nonuniform motion. It appears that the majority of the seasonal velocity changes occur concurrently along the slide body (Figure 2.6), suggesting that changes in effective normal stress are spatially extensive over observed timescales. Our analysis is limited by the temporal sampling of our InSAR dataset, which is somewhat irregular (Table A.2), and we are unable to detect changes in deformation rate that occur between satellite acquisitions (mode of 46 days).

Furthermore, our analysis of the slowest moving parts of the landslides is affected by a low signal-to-noise ratio that may enhance errors. For example, the analysis appears to indicate a temporal offset in motion and even slight negative velocity values (i.e. upslope motion) along the slowest-moving parts of the toe zones for Boulder Creek during 2007 and for Kekawaka 1 during 2007 and 2010 (Figure 2.6). Given the limitations in the InSAR data, we cannot exclude the possibility that local stress perturbations are either triggering seasonal motion that propagates along the landslide body or triggering variations in motion over even shorter timescales. We calculate that if motion is triggered by local stress perturbations, the instability must travel along the landslide body at 10^2 – 10^3 m day⁻¹, which is several orders of magnitude faster than the average landslide velocity.

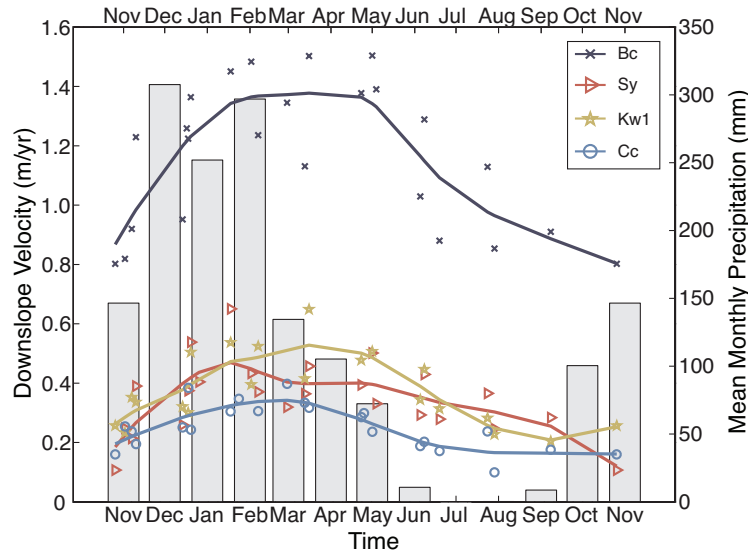


FIGURE 2.5. Velocity time series for the transport zone (spatially averaged) of four earthflows projected onto the downslope direction. Data from March 2007 to January 2011 is condensed into a single calendar year and plotted by day of the year. Symbols represent measured values and lines represent a smooth-spline fit to the data. Gray bars indicate mean monthly precipitation over the same time span. Precipitation data collected at Richardson Grove State Park, CA located ~ 30 km west of our field area (Figure 2.1). Earthflow names are listed in the caption to Figure 2.1.

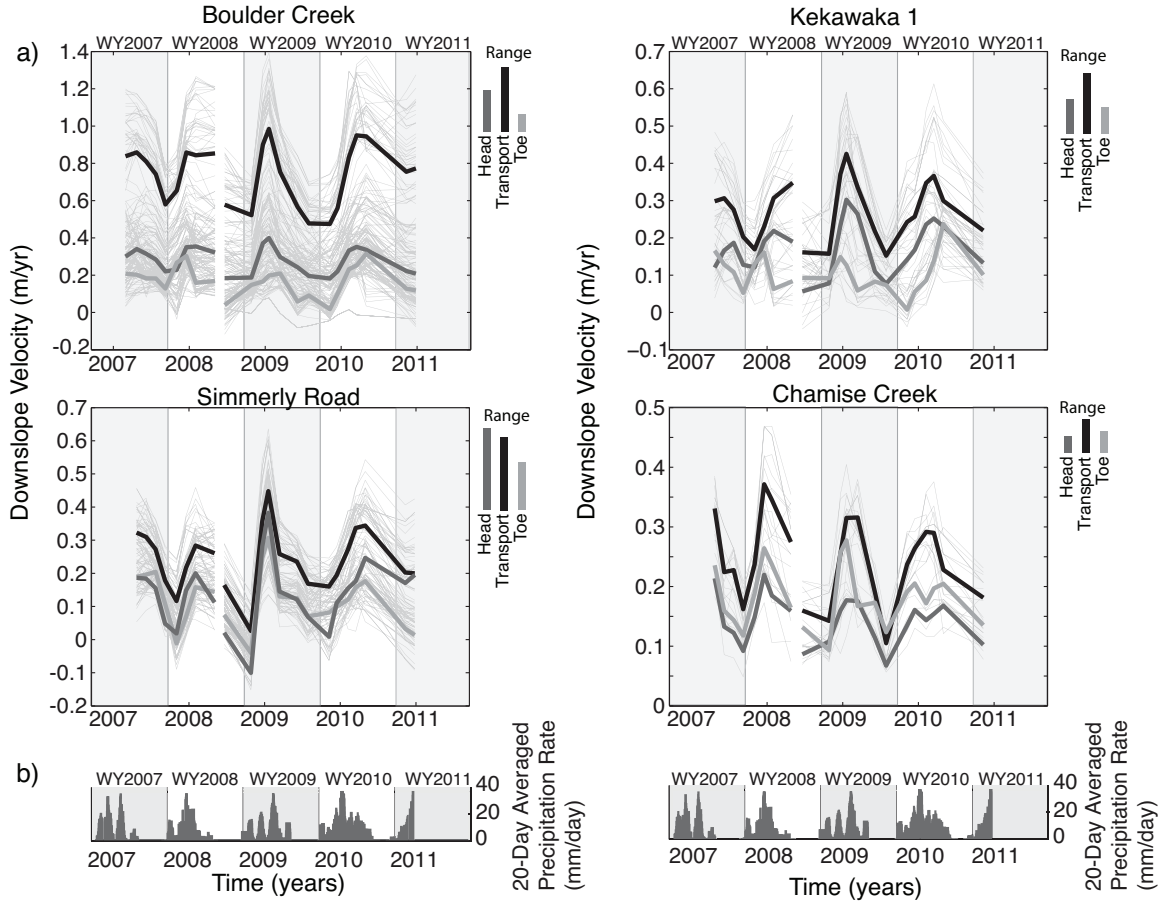


FIGURE 2.6. Velocity time series along the longitudinal axis of each landslide. a) Projected downslope velocity time series between February 2007 and January 2011. Thin gray lines represent the velocity time series for each pixel along the longitudinal profiles in Figure 2.4. Thick lines indicate spatially averaged values of each kinematic zone. Transport zone lines are modified from Handwerger et al. (2013). Range bars represent the relative range of scatter in the data for each kinematic zone. Data gap over summer 2008 results from decorrelated interferograms. b) 20-day averaged precipitation rate. Precipitation data collected at Richardson Grove State Park, CA located ~ 30 km west of our field area (Figure 2.1). Gray and white boxes and dashed lines correspond to alternating water years (WY). WY are defined as Oct 1–Sept 30. We observe that the head, transport, and toe regions accelerate in unison across the slide. Note temporal offsets and negative velocity values occur in low velocity regions of some slides. We attribute these offsets and negative velocities to inversion misfit resulting from a low signal-to-noise ratio (i.e. low velocity).

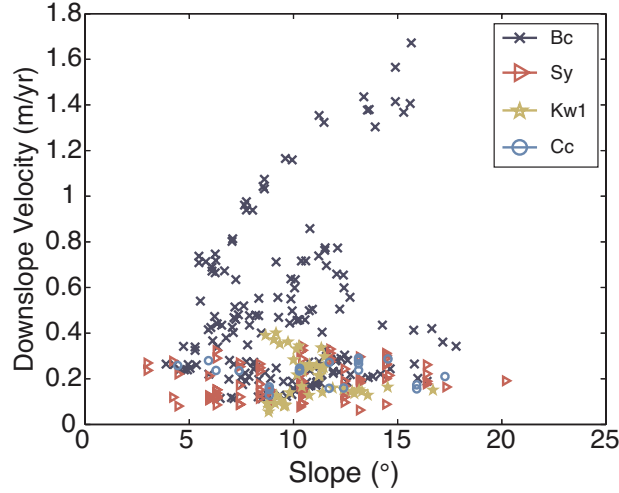


FIGURE 2.7. Topographic slope versus temporally averaged downslope velocity. Each symbol corresponds to the slope-velocity relationship at an individual pixel along the slide axis. Slope-velocity values are taken along the profiles shown in Figure 2.4. The Boulder Creek earthflow appears to have the strongest correlation between slope and velocity.

2.5.3. Topographic Slope, Surface Velocity, and Inferred Thickness

We did not find any clear correlation between average topographic slope and average slide velocity for these earthflows (Figure 2.3); however, local changes in slope and thickness (i.e. driving stress) control slide velocity. By comparing slope and velocity over small spatial scales along the slide axis (30 m pixels), we find that, in some areas, changes in topographic slope are strongly correlated to changes in slide velocity, while in other areas there is weak to no correlation (Figure 2.4). Also, for a given slope value there is a relatively wide range in velocity values (Figure 2.7).

Assuming these earthflows can be approximated as viscous flows with a non-Newtonian rheology, our observations suggest that the thickness of these earthflows must be highly variable to explain their slope-velocity relationship. Using Eq. (2.3), we estimated the changes in slide thickness required to explain the slope-velocity values at the Boulder Creek earthflow. We only present data from Boulder Creek,

but this analysis can be applied to each of the slides that have an InSAR signal with a high signal-to-noise ratio. Given the parameters listed in Table A.4, Eq. (2.3) predicts a highly variable thickness with values ranging from 14 to 44 m (Figure 2.8). These inferred thickness values are within the range of thickness estimates made by Mackey and Roering (2011) using lidar and field observations. However, the magnitude of our inferred thickness values depend strongly on the model parameters, which need to be calibrated to specific sets of landslides. Regardless of which flow-law parameters are used, the model predicts a highly irregular slide thickness.

We also estimated the basal topography by subtracting the thickness profile from the surface elevation data, and these predictions show that the Boulder Creek earthflow has an irregular basal-slip surface and displays variability over both short (10^2 m) and long (10^3 m) length scales (Figure 2.8). Note the basal surface closely mimics the ground-surface topography. We do not have the required borehole measurements or seismic data to test our model predictions, but our first-order analysis agrees well with predictions and observations of a variable slide thickness from other slow-moving landslides around the world (Baum et al., 1993; Coe et al., 2009; Booth et al., 2013a; Guerriero et al., 2014; Prokešová et al., 2014).

2.6. Discussion

2.6.1. Geometrical Controls on Earthflow Kinematics

The Eel River earthflows display distinct kinematic zones that can vary significantly in average velocity. By comparing the 13 slides that occur within the lidar data to Mackey and Roering (2011) landslide inventory, we found that the locations of these kinematic zones have remained spatially fixed for almost 70 years. This strongly suggests that landslide geometry exerts a primary control on long-term

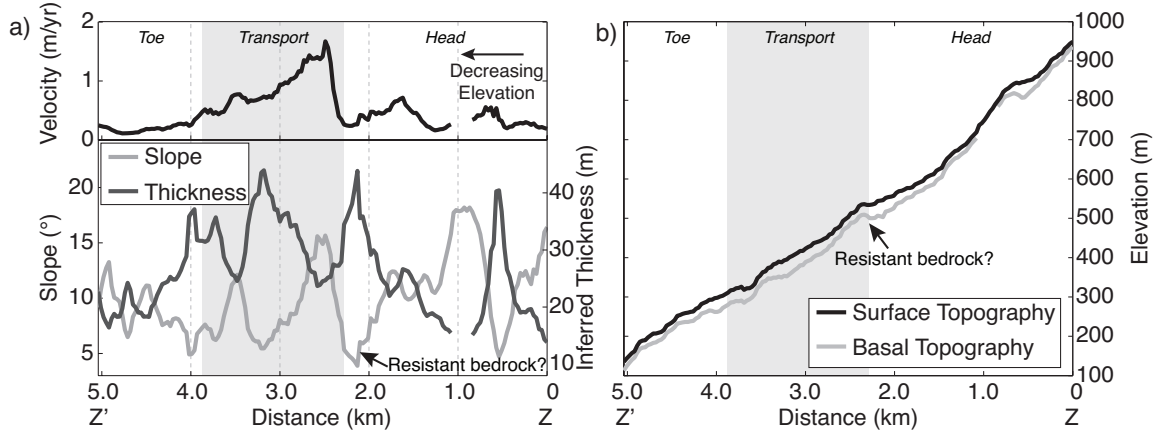


FIGURE 2.8. Topographic slope, velocity, model-inferred thickness, and elevation profiles from Z to Z' along the longitudinal axis of the Boulder Creek earthflow (see profile line in Figure 2.4). We set the earthflow density $\rho = 2145 \text{ kg m}^{-3}$, flow-law exponent $p = 3$, and flow-law constant $c = 4.6 \times 10^{-21} \text{ m kg}^{-9} \text{ yr}^{5/3}$. a) Slope, velocity, and model-inferred thickness profiles. Black line corresponds to temporally-averaged (2007–2011) velocity profile projected onto the downslope direction. Light gray line corresponds to topographic slope and dark gray line corresponds to inferred slide thickness calculated using Eq. (2.3). b) Basal and surface topography profiles. Elevation is given as meters above sea level. Basal topography is calculated by subtracting the surface elevation profile by the inferred thickness. Note the basal surface closely mimics the ground-surface topography. Both display irregularities over short (10^2 m) and large (10^3 m) spatial scales. We have indicated the location of a feature, which we have inferred to be resistant bedrock.

landslide kinematics. Our results agree well with other studies that have quantified the kinematic history of slow-moving earthflows (e.g., Coe et al., 2009; Guerriero et al., 2014). In a recent study on the Montaguto earthflow, Italy, Guerriero et al. (2014) concluded that the positions of the kinematic zones are set by the geometry of the basal-sliding surface. Given our similar results, we propose that the kinematics of the Eel River earthflows is also controlled by the geometry of the basal-sliding surface.

By modeling these earthflows with a viscous non-Newtonian rheology, we were able to infer slide thickness and basal topography along the axis of the Boulder Creek earthflow. Our model results suggest slide thickness and basal topography must vary significantly in order to explain the observed relationships between topographic slope and velocity at the ground surface. We also observed that the inferred basal topography closely mimics the ground surface topography along much of the landslide. We hypothesize that these inferred variations result from structural and lithologic features along the landslide base. It is encouraging that our results agree with studies that have used a variety of methods to infer and measure landslide thickness (Mizuno, 1989; Baum et al., 1993; Coe et al., 2009; Booth et al., 2013a; Guerriero et al., 2014; Prokešová et al., 2014); however before this method can be widely applied it requires validation through substantial field investigation.

Given that structural and lithologic features remain spatially fixed for long time periods, they can exert significant control on landslide motion as the landslide material is forced to deform in order to slide over these surfaces. These features can influence local groundwater flow and affect pore-water pressures (Keefer and Johnson, 1983; Iverson and Major, 1987; Baum and Johnson, 1993; Van Asch et al., 2006). For instance, the compression and extension of the landslide colluvium as it deforms over an irregular surface can lead to increases and decreases in pore-water pressures,

respectively. Furthermore, these features effect the stress distribution along the sliding surface and within the landslide body (Mizuno, 1989; Baum and Fleming, 1991; Guerriero et al., 2014; Prokešová et al., 2014).

Many studies have examined the effects of an irregular slip surface on controlling the motion of tectonic faults and glaciers (Weertman, 1957; Nye, 1969; Chester and Chester, 2000; Dunham et al., 2011). For faults, sliding along an irregular slip-surface produces complex behaviors resulting from variations in normal stress along the surface. These irregularities (i.e. asperities) also increase the surface roughness of the slip surface. Over time, however, the slip surface roughness decreases as subsequent sliding events break off asperities (Sagy et al., 2007). Like faults, a landslide’s slip surface may evolve over long periods of sustained motion. For instance, the preferential development and alignment of clay minerals along the shear zone may act to reduce the surface roughness and alter the mechanical properties of the slip surface (Bishop et al., 1971; Wang et al., 2010). Yet, it appears that these slope failures are unable to suppress the expression of large asperities, such as those associated with resistant bedrock (Baum et al., 1993; Coe et al., 2009; Guerriero et al., 2014). One potential example of this type of feature can be observed at the Boulder Creek earthflow near the upper edge of the transport zone (Figure 2.4; Figure 2.8). As the Boulder Creek earthflow translates over the upslope edge of this feature, there is a decrease in topographic slope, a predicted increase in slide thickness, and decrease in downslope velocity on the upslope edge; while translation over the downslope edge results in an increase in topographic slope, a predicted decrease in thickness, and an increase in velocity. Thus, this feature appears to have spatially fixed the upslope location of the transport zone.

Although landslide geometry likely controls the long-term motion of the Eel River earthflows, it does not appear to strongly regulate their response to seasonal stress perturbations. Instead, the frictional properties and pore-water pressure along the sliding surface controls the magnitude of those rates. We found that these earthflows exhibit seasonal velocity changes consistent with precipitation-induced changes in effective normal stress. For the four earthflows analyzed in detail, the velocity of the transport zone relative to the toe and head zones scales in proportion and shows no detectable phase lag through the seasonal cycle. This suggests that changes in effective normal stress are communicated along the shear zone within the timescales resolved by our InSAR dataset (i.e. < 46 days). Given that these slides are driven by precipitation-induced changes in pore-water pressure, one explanation is that these patterns of seasonal acceleration and deceleration are triggered by spatially extensive stress perturbations resulting from seasonal changes in groundwater level along the slide body (Iverson, 1986b; Iverson and Major, 1987; Berti and Simoni, 2012). However, it is possible that these earthflows exhibit differences in the timing of motion along the slide body, driven by local stress perturbations, that occur over timescales that are unresolved by the sampling frequency of our InSAR data. Iverson (1986b) concluded that the rate at which local stress perturbations travel along the landslide depends on the material properties of the landslide colluvium. If the landslide material exhibits plastic deformation, perturbations progress rapidly and diffuse along the slide body. If the landslide material exhibits viscous deformation, perturbations progress as a slow-moving kinematic wave over the course of several years. The nearby Minor Creek landslide, which likely has similar material properties to our slides, exhibits a combination of plastic- and viscous-like deformation, but is able to respond rapidly to local stress perturbations (Iverson, 1986b). Therefore, in

order to identify any potential nonuniform motion requires velocity data with a much higher temporal sampling. Future work includes plans for field monitoring and the use of data from newer satellites that have a shorter repeat intervals (< 2 weeks).

2.7. Conclusion

The Eel River catchment contains many active earthflows that display persistent long-term motion with seasonal variations in velocity. Using satellite InSAR and high-resolution DEMs, we identify 50 active slow-moving landslides over a $\sim 14,000$ km² area in the Northern California Coast Ranges. We find no clear relationship between landslide size, average slope, or average velocity. All but one of these landslides occur within the Central and Eastern belts of the Franciscan Complex indicating a strong lithologic control on their spatial distribution. Each slide displays three main kinematic zones that have remained spatially fixed over seasonal and decadal timescales, suggesting kinematics are primarily controlled by landslide geometry. Modeling earthflow motion using a viscous flow-law constrained by surface kinematics, we find that individual slides likely display significant variations in thickness and an irregular basal-slip surface. These potential variations in the basal-slip surface can be used to explain surface deformation patterns.

Time series analysis of four representative earthflows revealed that each slide, despite significant variations in geometry, exhibited seasonal velocity changes that occur along the entire landslide body. Velocities increased rapidly during the wet season and decreased gradually throughout the dry season. Although each slide displays distinct kinematic zones with different mean velocities, the timing of seasonal motion appears to occur synchronously along the landslide body. This behavior can be explained by two alternate hypotheses: 1) seasonal motion is controlled by spatially

extensive changes in effective normal stress, or 2) that local stress perturbations are communicated along the slide body over daily-to-monthly timescales, which cannot be resolved with our InSAR data. Our results suggest that slide geometry controls long-term motion and sets kinematic patterns, but does not strongly regulate their response to seasonal stress perturbations.

2.8. Bridge

In this chapter, I used satellite-based InSAR and high-resolution topographic data to construct a landslide inventory and develop techniques to perform time series inversions of multiple slow-moving landslides. By exploring the seasonal response along the landslide body, I was able to document how individual slides respond to changes in rainfall (i.e. effective stress) and show that these slides display a synchronous response both within and between landslides. I also presented a new method for characterizing the subsurface geometry of these landslides.

In the next chapter, I explore controls on seasonal landslide motion in the context of a simple, but commonly used hydrological model. This model predicts that one-dimensional vertical diffusion of pore-water pressure controls a landslide's response to precipitation, which therefore implies that deeper landslides should take longer to respond to rainfall than shallower landslides. However, as noted in the above chapter, the landslide response to precipitation is remarkably similar despite significant variations in landslide size and depth. Therefore I will explore several potential explanations that can describe this behavior.

CHAPTER III

CONTROLS ON THE SEASONAL DEFORMATION OF SLOW-MOVING LANDSLIDES

Published as: Handwerger, A. L., Roering, J. J., and Schmidt, D. A. (2013). Controls on the seasonal deformation of slow-moving landslides. *Earth and Planetary Science Letters*, 377, 239–247. As lead author, I wrote the manuscript and performed all of the analysis and interpretation, and drafted all the figures for this chapter. My co-authors, Roering and Schmidt helped me with editorial assistance and the interpretation of my results.

3.1. Introduction

Slow-moving landslides cause significant erosion, regulate hillslope angles and topographic relief, and are a principle geologic hazard that damages infrastructure. These slope failures are driven by hydrologic forcing and typically respond to precipitation (i.e., accelerate) over seasonal time scales through increased pore-water pressure and reduced effective normal stress along basal shear zones (Terzaghi, 1951). Quantifying the interaction between precipitation and landslide mobility is essential for characterizing the role of landslides in landscape evolution and hazard mitigation. Slow-moving landslides are found in diverse climatic and geological settings, and when examined in detail, do not exhibit straightforward relationships between precipitation, topographic slope, and velocity; yet, they frequently exhibit a seasonal pattern of acceleration and deceleration (Iverson and Major, 1987; Coe et al., 2003; Hilley et al., 2004; Calabro et al., 2010; Scheingross et al., 2013). These slope failures accelerate within days to months after the onset of rainfall and decelerate into the

dry season, with some coming to a halt. Periods of acceleration are correlated with increasing pore-water pressures caused by infiltration of precipitation, whereas periods of deceleration are typically correlated with decreasing pore-water pressures caused by drainage, and, potentially from dilation of the shear zone material (Iverson and Major, 1987; Coe et al., 2003; Iverson, 2005; Schulz et al., 2009b). There is ample evidence from laboratory experiments (Iverson et al., 2000; Moore and Iverson, 2002) and numerical models (Iverson, 2005) to support the occurrence of shear zone dilation, but field evidence is limited (Schulz et al., 2009b). Mechanical-hydrologic feedbacks facilitate seasonal sliding that can persist for hundreds or even thousands of years (Bovis and Jones, 1992; Mackey et al., 2009; Rutter and Green, 2011; Coe, 2012), although slow-moving landslides can sometimes fail catastrophically without warning (Petley et al., 2002).

Numerical models and field-based measurements suggest that substantial transient pore-water pressure fluctuations, which decrease effective stress and thus the frictional resistance to sliding, are transmitted from the surface to depth in the vertical direction and are well described by a one-dimensional linear diffusion equation (Iverson and Major, 1987; Haneberg, 1991; Reid, 1994; Iverson, 2000; Berti and Simoni, 2010, 2012). According to these models, the minimum time in which strong pressure change occurs is $T_D = Z^2/D_0$, where Z is landslide depth below the water table and D_0 is characteristic hydraulic diffusivity (Iverson, 2000; Coe, 2012). Calculated values of T_D can serve as a first-order approximation of the pore-water pressure response time. As a result, this simple and frequently used theoretical framework for modeling hydrologically driven changes in landslide velocity suggests that in areas where climate, hillslope angles, and regolith properties are relatively uniform the landslide response time, or the time between the onset of precipitation

and landslide acceleration, should scale with the square of landslide depth. Using field-based estimates of hydraulic diffusivity and shear zone depths typical of slow-moving landslides, this model predicts landslide response times to be on the order of days to years. And, in fact, several studies (e.g., Iverson and Major, 1987; Coe et al., 2003; Hilley et al., 2004; Calabro et al., 2010) have documented timescales that are consistent with this range of expected landslide responses.

Iverson and Major (1987) seminal study of the Minor Creek slide, Northern California, was the first to demonstrate the dominant role of pore-water pressure diffusion in controlling the response of a slow-moving landslide. They found that a simple linear diffusion model could account for the attenuation and phase lag observed in their piezometers, and that landslide acceleration was correlated with a pore-water pressure threshold located at intermediate depths within the slide. Subsequent studies have also found that application of a simple diffusion model can account for the hydrologic response of landslide colluvium (e.g., Haneberg, 1991; Reid, 1994; Berti and Simoni, 2010, 2012). However, no studies have tested how a one-dimensional diffusion model can account for the dynamics of multiple landslides in the same climatic, tectonic, and lithologic area because direct observation of landslide dynamics and hydrologic conditions are often expensive, highly localized, short-lived, and limited to individual landslides in diverse settings.

Simple, homogenous one-dimensional diffusion models provide a readily accessible, physically-based framework for understanding the hydrologic mechanisms that drive landslide dynamics, but the applicability of these models has not been broadly tested (Berti and Simoni, 2010). One-dimensional diffusion models do not account for discontinuous unsaturated zones, heterogeneous hydraulic properties, surface ponding, or spatial and temporal variations in preferential flow. Landslide

hydrology is further complicated by mechanical-hydrologic feedbacks such as shear zone dilation and the development of deformation cracks, which require significant parameterization in order to be incorporated into models (Iverson, 2005; Krzeminska et al., 2013). Deformation cracks are pervasive in clay-rich, slow-moving landslides due to stresses induced by irregular basal topography and changes in deformation (Coe et al., 2009; Krzeminska et al., 2013), yet their effect on landslide behavior is not well understood. These features may have a dual role in affecting landslide stability, in that they facilitate rapid pore pressure transmission to the shear zone as well as increase the drainage efficiency of moving slides (Krzeminska et al., 2013). Currently, there is a lack of spatially explicit data and calibrated models to illustrate how these deformation-hydrology feedbacks influence slow-moving landslides.

Remote sensing techniques have greatly improved our ability to quantify deformation of the earth's surface and provide data with high temporal and spatial resolution. Interferometric synthetic aperture radar (InSAR) can measure mm-scale surface deformation and is a powerful tool for studying a variety of phenomena such as earthquakes (Fialko et al., 2001), volcanoes (Wicks et al., 2002), glaciers (Rignot et al., 2004), land subsidence (Schmidt and Bürgmann, 2003), and landslides (Hilley et al., 2004; Roering et al., 2009; Calabro et al., 2010; Zhao et al., 2012). Previous studies using InSAR in the Eel River catchment have documented multiple active landslides that span an order of magnitude in size and velocity (Roering et al., 2009; Zhao et al., 2012), but these studies did not systematically assess potential linkages between landslide size, depth, and response time.

Here, we use InSAR to construct deformation time series between February 2007 and January 2011 for an inventory of 10 active landslides in the Eel River catchment, Northern California (Fig. 3.1). Because these slides have similar slope angles and

occur within the same tectonic, lithologic, and climatic setting, we are able to analyze the extent to which their collective behavior is consistent with a commonly used one-dimensional diffusion-driven hydrologic model. We explore the relationship between response time and landslide depth derived from field observations, DEMs, and locally-derived scaling relationships. Contrary to model predictions, our data show that slide response to seasonal precipitation is remarkably similar despite a five-fold variation in landslide depth. Our findings challenge and potentially contradict assumptions made in many landslide models that are used for landscape evolution and hazard assessment, and encourage reevaluation of mechanical-hydrologic mechanisms that control landslide dynamics.

3.2. Slow-moving Landslides, Eel River, Northern California

The Eel River catchment is notable for deep-seated, slow-moving slope failures that are induced by highly seasonal precipitation, weak mélange lithology, and high rates of rock uplift (Kelsey, 1978; Mackey and Roering, 2011). Average annual precipitation is 1.4 m, most of which falls between October and April. The Coast Range is underlain by the Franciscan *mélange*, a Jurassic-Cretaceous accretionary prism complex that is highly sheared. Our study area lies within the Central Belt of the Franciscan *mélange*, which is comprised of a *mélange* matrix surrounding blocks of coherent sandstone, meta-basalt, blueschist, meta-sandstone, and shale (McLaughlin et al., 2000).

The northern Coast Range of California has been tectonically active since the Miocene, largely shaped by the northward migration of the Mendocino Triple Junction (MTJ). Models and geomorphic evidence predict that migration of the MTJ creates a zone of rapid rock uplift, that moves north at 5 cm/yr (Furlong and Govers, 1999; Lock

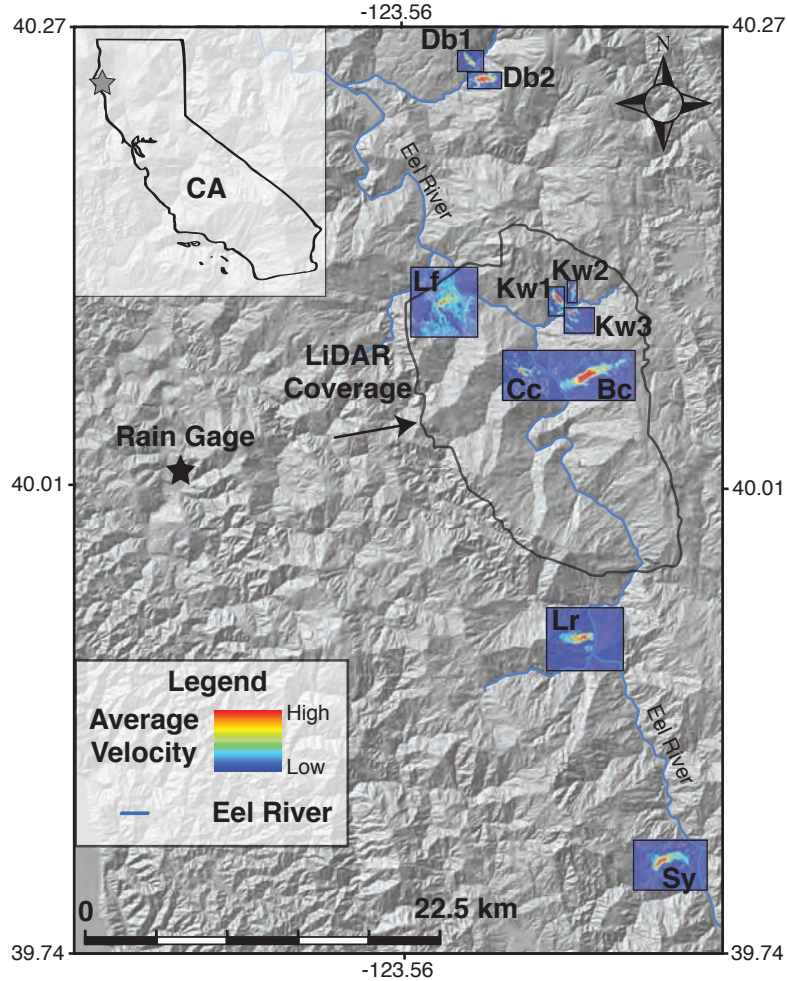


FIGURE 3.1. Eel River catchment, Northern California. Interferograms showing the time-averaged downslope velocity draped over shaded relief. The color scale indicates relative velocity values rather than the actual magnitude values because the velocity range is slightly different for each landslide. Specific landslide velocity magnitudes are reported in Fig 3.2. Black star marks the location of the NOAA rain gage. Black outline indicates lidar coverage. Main stem of the Eel River is outlined in blue. Landslide names are abbreviated as the following: Db1 = Dobbyn Creek 1, Db2 = Dobbyn Creek 2, Lf = Lauffer Road, Kw1 = Kekawaka Creek 1, Kw2 = Kekawaka Creek 2, Kw3 = Kekawaka Creek 3, Cc = Chamise Creek, Bc = Boulder Creek, Lr = Lundblade Ranch, Sy = Simmerly Road.

et al., 2006). The resulting uplift gradient influences landscape organization through drainage reversal (fish-hook streams) and river capture (Fig. 3.1) (Lock et al., 2006). Uplift rates are not well constrained in our field area, but Lock et al. (2006) estimate rates between 0.5 and 1 mm/yr, which is consistent with erosion rates between 0.6 and 1.1 mm/yr derived from suspended sediment data (Wheatcroft and Sommerfield, 2005) and in-situ cosmogenic radionuclides analyzed from stream sediments (Balco et al., 2013).

Owing to the persistent activity of slow-moving landslides, the region features relatively low relief, low gradient slopes (average slope $< 20^\circ$), and high erosion rates (0.9 mm/yr) (Wheatcroft and Sommerfield, 2005; Booth and Roering, 2011). Landslides in this region, which occupy a relatively small percentage of the total landscape ($< 10\%$), contribute a disproportionately large percentage of the regional erosion rate (Swanson and Swanston, 1977; Kelsey, 1978, 1980; Nolan et al., 1995). Mackey and Roering (2011) found that active landslides connected to channels occur in only 6% of our field site and account for $> 50\%$ of the regionally-averaged denudation rate. Although the role of slow-moving landslides in modulating topographic relief and delivering large boulders to channels has been explored (Booth and Roering, 2011; Booth et al., 2013b), less is known about what controls their seasonal dynamics or their potential for catastrophic failure.

We analyzed 10 of these slow-moving landslides, often referred to as earthflows, which exhibit plug flow deformation above narrow shear surfaces (Iverson, 1986a; Swanson and Swanston, 1977; Simoni et al., 2013). Each landslide has distinct kinematic zones (e.g., source, transport, and toe) and classical hummocky topography signaling recent movement across much of the terrain (Fig. 3.2). These failures have areas ranging from 0.16 to 3.1 km² and average slope angles of $12 \pm 3^\circ$. From historical

air photos, average landslide velocities since the 1960s are 0.2 to 4.2 m/yr and these rates do not appear to vary systematically with landslide size, depth, or average slope (Mackey and Roering, 2011). Analysis of pre-historic landslide rates using meteoric ^{10}Be shows that the Kekawaka 2 landslide has experienced movement for at least 150 years, indicating that these features can sustain perpetual movement for long periods of time (Mackey et al., 2009).

3.3. Methods

3.3.1. InSAR

To quantify the seasonal behavior of these 10 landslides, we produced 165 differential interferograms between February 2007 and January 2011 using data acquired by the PALSAR instrument on the ALOS-1 satellite. PALSAR operates with a L-band antenna (23.5 cm) and a 46-day repeat interval. There is a data gap over the summer of 2008 due to large perpendicular baselines, making surface deformation unresolvable during this time. Interferograms were processed using the Repeat Orbit Interferometry Package (ROI PAC) developed at JPL/Caltech (Rosen et al., 2004). Satellite tracks 223 F790 and 224 F790 overlap our field site, and we utilized SAR data from both frames. In contrast to typical landslide studies that require PS-InSAR techniques to quantify deformation (e.g., Hilley et al., 2004), this region is ideal for studies using conventional 2-pass interferometry because the landslides are continuously moving at a rate sufficient to observe deformation in a short time span, yet slow enough to avoid a loss in radar coherence (Roering et al., 2009). We identified active landslides over a range of length scales (< 1 to > 5 km) by stacking interferograms and then cross checking our InSAR analysis with previously published landslide inventories (Roering et al., 2009; Mackey and Roering, 2011) and

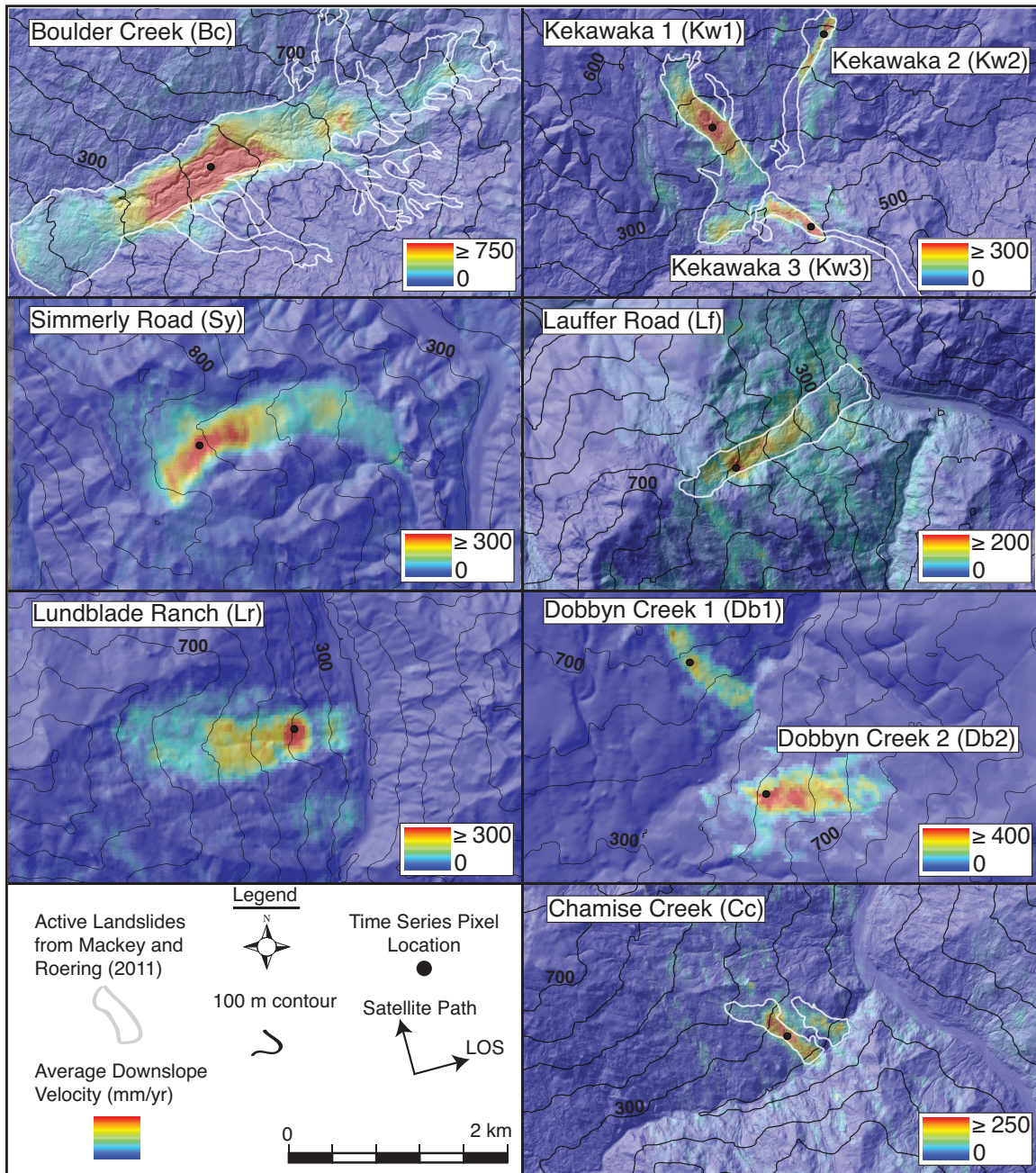


FIGURE 3.2. Time-averaged (2007–2011) downslope velocity draped over shaded relief maps derived from lidar (1 m grid) and 10 m grid. Line-of-sight velocities are projected downslope for each slide using the local slope and azimuth. Color scales were chosen to highlight kinematic zones, not to display the full extent of velocity (i.e. red pixels are $\geq$ the maximum listed value).

high resolution DEMs (1 m and 10 m grid). To allow for direct comparisons, we back-projected all line-of-sight velocity estimates to the downslope direction of each landslide using their average topographic slope and downslope azimuth inferred from the DEMs and historical air photo-derived displacement vectors (Hilley et al., 2004; Mackey and Roering, 2011).

We employed the method of Schmidt and Bürgmann (2003) to invert 51 small-baseline (< 1400 m) interferograms for a smooth time series (Table B.1). Our time series processing procedure allowed us to evaluate the translation of any coherent pixel, thus providing complete spatial coverage of each landslides deformation history (Fig. 3.2). The nominal sampling interval of our time series is set by the 46-day repeat time interval of the satellite. However, we achieved a temporal sampling as short as ~ 30 days in certain time periods because we combined SAR data from both overlapping tracks whose data acquisitions are out of phase.

3.3.2. Landslide Depth, Soil Properties, and Response Time

We estimated landslide area, slope, and depth with both field and topographic data. For six of the landslides (Boulder Creek, Kekawaka Creek 1, Kekawaka Creek 2, Kekawaka Creek 3, Chamise Creek, and Lauffer Road), we measured landslide depth using lidar (1m grid) and field observations from longitudinal gullies incised through landslide bodies and from hillslope-channel interfaces where landslide toes are actively truncated by stream erosion. For the four landslides (Simmeerly Road, Lundblade Ranch, Dobbyn Creek 1, and Dobbyn Creek 2) that occur outside of the lidar coverage, we calculated depth estimates using a power-law fit to depth-area scaling data derived from 69 landslides in the Eel River catchment mapped from historic aerial photos and lidar (Fig. 3.3) (Mackey and Roering, 2011). The depth-area relationship is defined

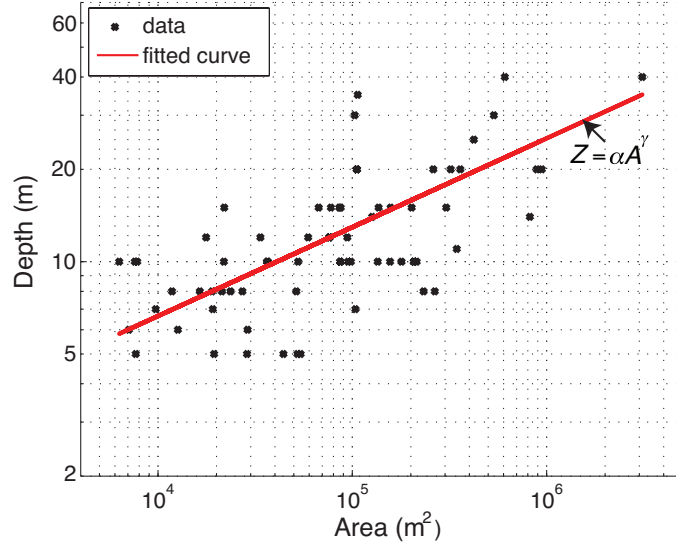


FIGURE 3.3. Depth-area scaling for 69 landslides in the Eel River catchment measured from topographic data using lidar (1m grid) (Mackey and Roering, 2011). The fitted curve shows a power-function fit with a non-linear least squares regression. The power-function is described by $Z = \alpha A^\gamma$, where Z is vertical landslide depth, α is a fit parameter, A is area, and γ is the power-law exponent. We find coefficients (with 95% confidence bound) $\alpha = 0.46$ (0.051, 0.87) and $\gamma = 0.29$ (0.22, 0.36). $R^2 = 0.45$.

by $Z = \alpha A^\gamma$, where Z is landslide depth, α is a fit parameter, A is area, and γ is the power-law exponent. Our scaling relationship ($\gamma \sim 0.29$) is in general agreement, but slightly lower than global data sets ($\gamma \sim 0.42$) (Larsen et al., 2010) and a recent earthflow scaling relationship ($\gamma \sim 0.44$) (Simoni et al., 2013). Our landslide depth estimates are not confirmed by borehole data and we likely underestimate the full range of depth values, because our estimated values come from gullies and channels that typically have not fully incised through the landslide mass.

The landslide material is poorly sorted, gravelly clay with occasional large blocks of coherent metasedimentary bedrock (Mackey et al., 2009). Hydraulic investigations at the nearby Minor Creek landslide, which occurs on the same geologic unit as our failures, yields a range of estimates for hydraulic diffusivity spanning 10^{-9}

to 10^{-4} m²/s, but 10^{-6} m²/s is adopted as the characteristic value (Fig. B.1) (Iverson and Major, 1987; Iverson, 2000, 2005). This variability in local (meter-scale) hydraulic conditions likely derives from the significant contrasts in bedrock and regolith properties associated with the Franciscan complex.

To quantify the landslide response time for the 10 landslides, we measured the time lag between the onset of seasonal precipitation and the onset of seasonal acceleration for the WY2008, WY2009, and WY2010. We exclude the WY2010 response time for the Db1 slide (which was 77 days) because that time span is poorly constrained with the InSAR data for that slide; and since Db1’s response time is identical to the 9 other slides for WY2008 and WY2009, we are confident the anomalous response is due to an error in the WY2010 data, rather than a valid slide response. Precipitation data was provided by NOAA and was collected at Richardson Grove State Park, Garberville, CA which is ~ 30 km west of our field area (Fig. 3.1). We define the onset of seasonal precipitation as the first rainfall event of the water year and the onset of acceleration is delineated as the first SAR acquisition where there is a positive change in velocity after the onset of rainfall.

3.4. Results

We analyzed 10 landslides that have areas ranging from 0.16 to 3.1 km² and estimated average depths that vary from 8 to 40 m (Fig. 3.2). Each of the landslides exhibits a zone of rapid movement in the central region of the slide, termed the transport zone. Transport zone velocities increased in the wet season and decreased in the dry season for each of the 10 landslides and the landslides exhibited continuous movement throughout the monitoring period (Fig. 3.4). In agreement with the historic record of landslide velocities in this region from air photos, our InSAR data indicate

that the average downslope velocity varied from 0.2 to 1.2 m/yr and these rates show no clear relationship with slide size, average slope, or depth (Table 3.1). The velocity time series for each slide has a well-defined periodicity of ~ 1 year and each landslide responds to seasonal precipitation within 40 days (Fig. 3.5). This implies that velocity changes, and in particular periods of acceleration following the onset of rainfall, are related to seasonal changes in effective normal stress modulated by transient pore-water pressures. The narrow range of response times belie substantial variation in landslide depth and are much shorter than would be predicted with observed values of diffusivity from the nearby Minor Creek landslide. Our analysis is limited by the ~ 40 day temporal resolution of the SAR data set, such that our response times represent maximum estimates; shorter response times elude the resolution of the current InSAR data. Nevertheless, the maximum response times reported here serve as critical constraints to test hydrologic model predictions for each landslide.

3.5. A Simple 1D Hydrologic Model for Landslide Response

Given our observations, we seek to reevaluate the applicability of the linear pore pressure diffusion model by using an analytical solution to quantify pressure changes to prescribed perturbations (i.e. rainfall). This analysis yields the magnitude of the pressure change as a function of soil diffusivity, landslide depth, and time. We assume a homogenous, fully saturated poroelastic medium (e.g., Iverson, 2000). Field investigations of the nearby Minor Creek slide have shown that hydraulic diffusivity is not correlated with depth and that the water table remains within a meter or two of the ground surface year round and approaches the ground surface during the wet season (Iverson and Major, 1987; Iverson, 2000). These observations support the use of the slide thickness as a reasonable approximation of the diffusion length scale and

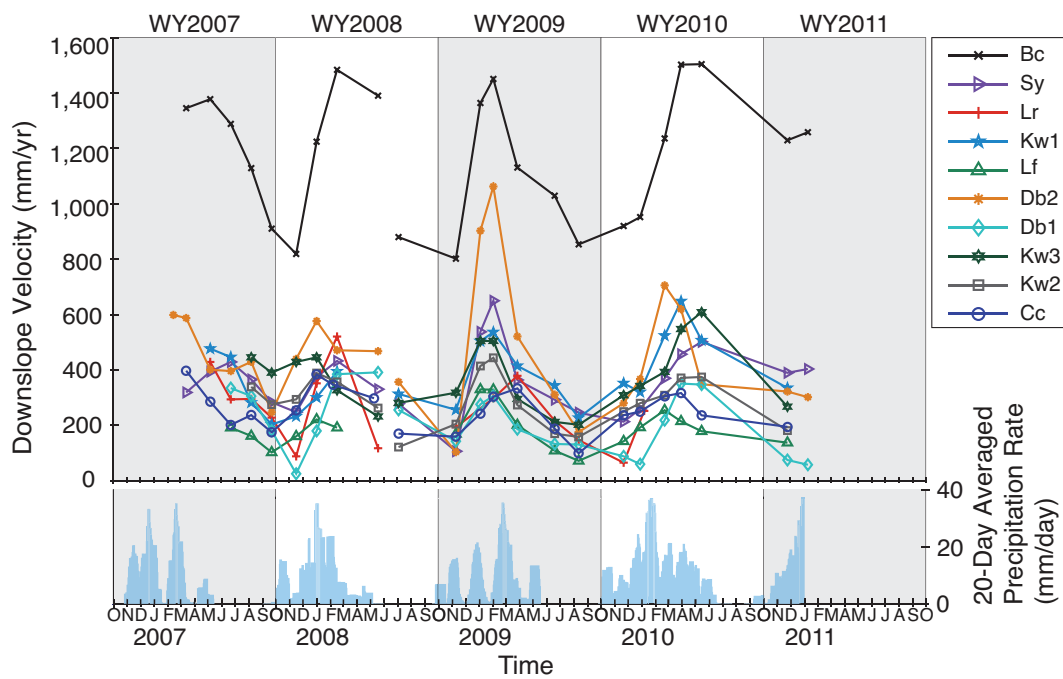


FIGURE 3.4. Downslope velocity time series for the transport zone of each landslide shown with colored lines. The data gap over Summer 2008 is due to large perpendicular baselines. Water years (starting Oct 1) are highlighted with white and gray boxes.

NAME	AREA (m ²)	LENGTH (m)	WIDTH (m)	DEPTH (m)	SLOPE (deg)	MEAN DOWNSLOPE VELOCITY (mm/yr)	RESPONSE TIME IN DAYS (WY2008, WY2009, WY2010)
Chamise Creek (Cc)	157549	782	200	15	13	260	36, 37, 40
Kekawaka 2 (Kw2)	208199	1476	140	10	10	297	36, 37, 40
Kekawaka 3 (Kw3)	232477	1921	120	8	9	374	36, 37, 40
Dobbyn Creek 1 (Db1)	374400	1090	340	19*	13	223	36, 37, 77 [^]
Dobbyn Creek 2 (Db2)	506090	1290	390	21*	15	459	36, 37, 40
Lauffer Road (Lf)	534981	1906	280	30	14	183	36, 37, 40
Kekawaka 3 (Kw3)	609348	1818	335	40	10	370	36, 37, 40
Lundblade Ranch (Lr)	1324580	2420	550	27*	15	256	36, 37, 40
Simmerly Road (Sy)	1767211	2490	710	30*	10	358	36, 37, 40
Boulder Creek (Bc)	3107283	4757	653	40	14	1,180	36, 37, 40

TABLE 3.1. Landslide metrics. Length is measured as distance from headscarp to toe. Width is calculated using area/length. See text for description of depth estimates. * indicates depth values calculated from our power-law function. Velocity is time averaged over 2007-2011. [^] corresponds to erroneous response time for Db1 (see text for description).

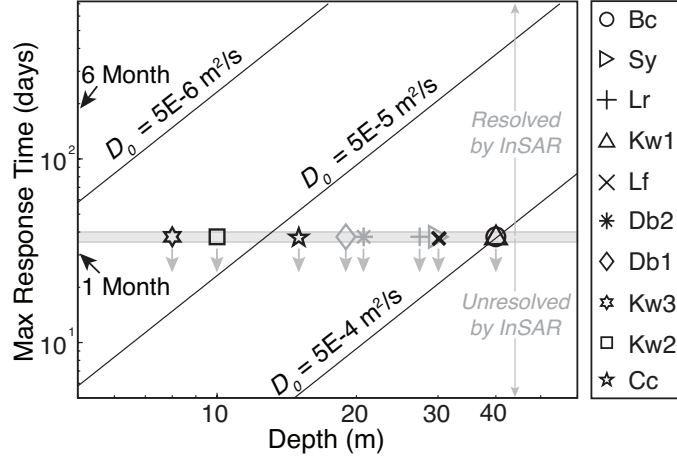


FIGURE 3.5. Maximum landslide response time. Symbols indicate the average maximum landslide response time for individual landslides. Black symbols have depth derived from lidar and gray symbols have depth derived from our scaling relationship. Gray band highlights the range of maximum response times for three water years. Downward pointing arrows emphasize that these are maximum values. Response time values are similar for each slide because the temporal sampling is identical among all of the slides. Solid diagonal lines represent T_D calculated using three diffusivities.

a relatively constant hydraulic diffusivity. However, a more realistic treatment would allow the water table height and the diffusivity to evolve throughout the year as the landslide colluvium wets, dries, and deforms. A one-dimensional pressure diffusion model is given as:

$$\frac{\partial P}{\partial t} = D_0 \frac{\partial^2 P}{\partial Z^2} \quad (3.1)$$

where P is the transient pore-water pressure imposed by a change in surface forcing (i.e., precipitation), Z is vertical depth from the water table (assumed to be at the topographic surface) to the shear zone, D_0 is the characteristic hydraulic diffusivity, and t is time. At time $t \leq 0$ we assume a background pore-water pressure has developed and then at time $t > 0$ we apply a step change at the surface such that $P(Z = 0, t > 0) = P_0$ and allow it to propagate downwards. Typically, a sinusoidal boundary condition is used at the surface to more accurately represent precipitation

events (Iverson and Major, 1987; Haneberg, 1991; Reid, 1994), but we chose to use a step change to clearly distinguish the subsurface response to the discrete onset of rainfall events. Also, rather than employ an arbitrary pressure change at the surface (P_0), we calculate the fraction of surface forcing transmitted to depth (Z) since we do not have observations of the absolute pore pressure. The analytical solution to this problem (Carslaw et al., 1959) provides the pore-water pressure change as a function of depth and time:

$$\frac{P(Z, t)}{P_0} = \text{erfc} \left(\frac{Z}{\sqrt{4D_0t}} \right) \quad (3.2)$$

where P/P_0 is the transient pressure as a percentage of the surface forcing. Using equation (3.2) we calculated the magnitude of pressure changes that occur within a year (Fig. 3.6).

Following the onset of precipitation, the diffusion model predicts that the strength of the pore-water pressure wave diffusing through the porous media attenuates with depth (Fig. 3.6). Thus, landslide thickness and diffusivity are predicted to control the strength and timing of a transient pore-water pressure signal (Iverson and Major, 1987). Studies using a diffusion model with a sinusoidal surface forcing have also found that low frequency pressure changes (i.e. seasonal precipitation) are required to penetrate deep-seated landslides, which has been used to explain why deep seated, slow-moving landslides respond to seasonal precipitation (Iverson and Major, 1987; Reid, 1994).

3.6. Potential Controls on Landslide Response

According to diffusion models with values of D_0 typical of those reported for similar soils and our range of slide depths, we expect the landslide response time to vary from days to years (Fig. 3.5). Instead, we find that landslides of highly

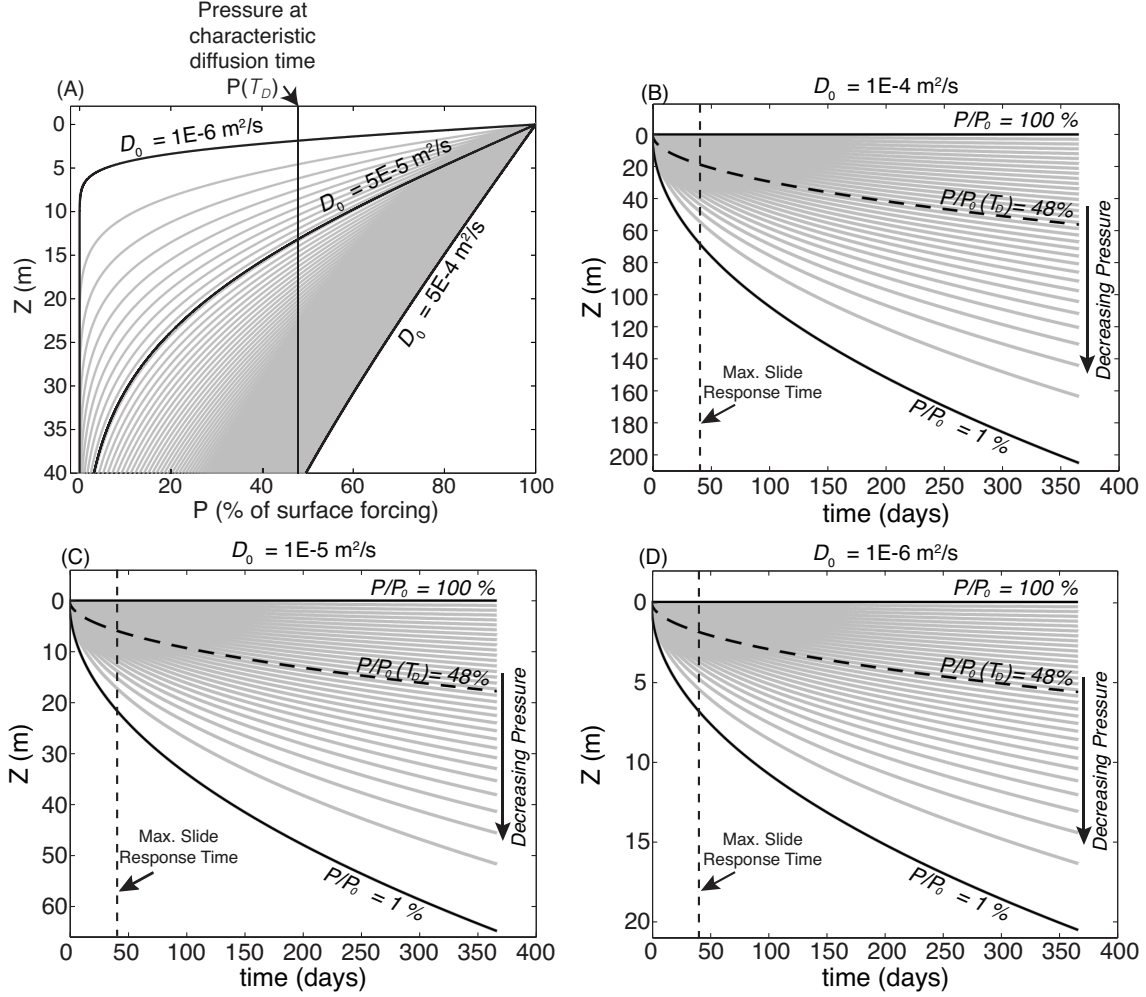


FIGURE 3.6. One-dimensional pore-water pressure diffusion model. (A) Pore-water pressure change that occurs in 40 days (maximum response time) as a function of landslide depth calculated with equation (3.2). Each curved line represents a constant diffusivity. Vertical black line marks the pressure change (48%) that occurs at the characteristic diffusion time scale $P(T_D)$. (B, C, D) Evolution of pore-water pressure over a one-year period using the three common hydraulic diffusivities. (B) $D_0 = 1E-4 \text{ m}^2/\text{s}$, (C) $D_0 = 1E-5 \text{ m}^2/\text{s}$, (D) $D_0 = 1E-6 \text{ m}^2/\text{s}$. Lines are spaced at 3% intervals and represent a constant pressure change. Vertical dashed line marks our observed maximum slide response time.

varying size and depth respond to rainfall within a narrow range of relatively rapid (≤ 40 days) timescales suggesting several possible explanations: 1) these landslides are sensitive to minor pore-water pressure changes, 2) the diffusivity of these slides is much larger than reported values, and/or 3) a simple, one-dimensional, homogeneous model cannot capture the hydraulic complexity of these slow-moving landslides.

The characteristic diffusion timescale, T_D , is commonly referenced in landslide studies as the minimum time for strong pore-water pressure response and is sometimes used as a first-order approximation of pore-water pressure response (e.g., Coe, 2012), but the exact piezometric response it predicts is rarely stated (Iverson, 2000; Montgomery and Dietrich, 2004). To quantify the pore-water pressure change that corresponds with the diffusion timescale we used equation (3.2) to find

$$\frac{P(Z, t)}{P_0} = \operatorname{erfc} \left(\frac{1}{2} \right) = 48\%. \quad (3.3)$$

Equation (3.3) shows that the diffusion time scale characterizes the time it takes for 48% of the surface forcing to be felt at depth. We find that T_D is an inappropriate predictor of the timescale for acceleration because these slow-moving slides may respond to subtle pore-water pressure changes that are much less than 48% of the surface forcing and occur well before T_D is reached (Fig. 3.6).

More importantly, our model results demonstrate that given the typical range of field-derived diffusivity values (Fig. B.1), it is unlikely that significant diffusion-driven pressure change occurs in our deepest landslides. For example, if we use the characteristic diffusivity of the Minor Creek landslide (10^{-6} m²/s), we find $< 1\%$ diffusive pressure change at depths > 7 m within the 40 day response time that can be resolved with our InSAR analysis (Fig. 3.6D). This suggests that these slides

remain close to an acceleration threshold and thus are sensitive to minor pressure changes.

Alternatively, the effective diffusivity that governs slide response could be much higher than values measured in the field. Reported values of diffusivity may underestimate the effective diffusivity that govern slide dynamics because field-based values derive from point measurements that are often selected to avoid heterogeneities (e.g., cracks) (Iverson and Major, 1987; Berti and Simoni, 2010, 2012). Furthermore, because it is common that soil diffusivity ranges over several orders of magnitude within a given landslide (Fig. B.1), previous studies have relied on point averaging (e.g., geometric mean), which can lead to a biased characterization of the hydrologic properties of a landslide as related to slide dynamics. For example, Reid (1994) found that diffusion model results accounted for the observed pressure response of the Alani-Paty landslide, Hawaii, using hydrologic parameters derived from the near surface rather than from the entire landslide. Importantly, spatially and temporally continuous measurements may be required to characterize the effective diffusivity of an entire landslide because point measurements or averages of point measurements can easily fail to capture the inherent variability of these parameters.

Preferential flow paths (such as cracks, fissures, and gullies) allow for rapid fluid transmission (up to m/hr), are common in actively deforming clay-rich landslides, and are often invoked to explain the rapid response of various hydrologic systems (Beven and Germann, 1982; Berti and Simoni, 2012; Krzeminska et al., 2013). If extensive, vertical-to-subvertical cracks (Fig. B.2) connect the near surface to the basal shear zone, they could transmit strong pressure changes much faster than we would predict using field-measured diffusivity values. Iverson and Major (1987) observed the hydrologic response of one such feature; they reported atypical behavior of a > 6

m deep well with very low hydraulic diffusivity (10^{-9} m²/s) that responded almost immediately to rainfall, suggesting proximity to a deep crack. This further exemplifies how point measurements cannot account for heterogeneities that may govern the hydraulic behavior of landslides. The role of preferential pathways in controlling the hydrologic response of soil is not well understood because they are typically overlooked in field studies given that they are difficult to map and extrapolate to depth (Berti and Simoni, 2012).

We also hypothesize that the hydrologic behavior of landslides may be controlled by an effective diffusivity whose value scales with landslide size. This would allow for the effective diffusivity to be larger than can be inferred from field measurements for large slides, while also acting to limit landslide response within a relatively narrow range. This interpretation is inspired by the scale-dependent behavior of longitudinal hydraulic dispersivity, which has been shown to increase as a power-law function of the measurement scale due to an increase in the frequency and/or magnitude of heterogeneities (e.g., high permeability zones) (Schulze-Makuch, 2005).

While the one-dimensional linear diffusion model can describe the hydrologic behavior of homogeneous soil columns (Berti and Simoni, 2010, 2012), our results indicate that the formulation may fail to capture time-dependent hydrologic changes inherent in an evolving mechanical-hydrologic system, such as a landslide. Consistent with our findings, Berti and Simoni (2012) monitored the seasonal pore-water pressure response of landslide-prone terrain and found that while the one dimensional linear diffusion model could predict shallow (< 2 m) transient pressure changes due to individual storms, it could not account for deep (> 2 m) seasonal pore-water pressure changes. Current models that predict the hydrologic response of landslides using a linear diffusion model have been invaluable in developing our understanding of these

mass movements and can be used to characterize a wide variety of landslide rates and styles (e.g., Iverson, 2000). However, these simplified models do not account for heterogeneities (e.g., spatially and temporally variable diffusivity, mechanical-hydrologic feedbacks) that may control some landslide behavior. Instead a multi-dimensional, heterogeneous diffusion model with mechanical-hydrologic feedbacks may be required to capture the full hydrologic behavior of slow-moving landslides (Krzeminska et al., 2013).

The remarkably consistent seasonal response (i.e. velocity and acceleration) of landslides in our study area suggests that these features may undergo self organization to maintain a narrow range of basal pore-water pressures, and, thus a narrow range of response times that allows for both long-term motion and seasonal oscillations. This implies that these slides remain close to an acceleration threshold (Iverson and Major, 1987), consistent with the “bathtub model” (Baum and Reid, 2000), a conceptual framework wherein landslides are hydraulically isolated from their surroundings by low permeability shear zones at the base and lateral margins. The hydrologic isolation may allow slides to maintain sufficient pore-water pressure to sustain year-round motion, but also requires that the slides have efficient drainage pathways to prevent catastrophic failure (Iverson, 2005). We hypothesize that the drainage efficiency may evolve each year through shear zone dilation and the connectivity of ephemeral preferential flow paths that help to regulate pore-water pressure (Sidle et al., 2001; Krzeminska et al., 2013). This agrees with observations of internal fractures and pervasive gully networks, which are common to these types of mass movements, and with studies that have quantified mechanical-hydrologic feedbacks such as dilative behavior and the development of deformation cracks (Iverson, 2005; Schwab et al., 2008; Schulz et al., 2009b; Roering et al., 2009; Krzeminska et al., 2013). Self-

regulatory behavior may suggest that these slides are mechanically and hydrologically adjusted to their environment such that they can accommodate daily, seasonal, and multi-year changes in effective stress without failing catastrophically.

3.7. Conclusions

Our data show that slow-moving landslides in the Eel River catchment exhibit seasonal velocity changes with a periodicity of ~ 1 year indicating that they are driven by deep seasonal pore-water pressure changes in basal shear zones. They respond within 40 days following the onset of seasonal rainfall despite significant variations in depth. This observation can be explained in several ways: 1) these slides are sensitive to minor pressure changes in basal shear zones ($\ll 48\%$ of the surface forcing), 2) the effective diffusivity that controls landslide behavior is much larger than those values typically reported (i.e. preferential flow paths) and may scale with landslide size, or 3) a simple, one-dimensional, homogenous linear diffusion model fails to capture the three-dimensional time-dependent hydrologic changes inherent in an evolving mechanical-hydrologic system, such as a slow-moving landslide. The narrow range in observed response times and persistent movement of the slides is consistent with the idea that these slides maintain a narrow range of pore-water pressure year round and are chronically close to an acceleration threshold such that hydrologic-mechanical feedbacks accommodate seasonal changes in effective stress.

3.8. Bridge

In this chapter, I showed that contrary to a widely used hydrologic landslide models, landslide response to seasonal precipitation is remarkably similar despite significant variations in landslide size, and particularly, depth. These findings

challenge and potentially contradict assumptions made in many landslide models that are used for landscape evolution and hazard assessment, and encourage reevaluation of the mechanical-hydrologic mechanisms that control landslide dynamics.

The next chapter draws directly from observations presented in Chapter II and Chapter III to explore how stress perturbations drive slow-moving landslides and also explores how stress perturbations trigger catastrophic landslides. I develop a mechanical model that captures the most salient features of landslide motion and provides a new context to understand interactions between stress perturbations and landslide failure mode.

CHAPTER IV

RATE-WEAKENING FRICTION DESCRIBES SLOW SLIDING AND CATASTROPHIC FAILURE OF LANDSLIDES

In preparation for submission to *Nature Geoscience*. Co-authored with Alan Rempel, Robert Skarbek, Joshua Roering, and George Hilley (Stanford University). The model was initially developed by Skarbek and then significantly modified by me. I drafted all the figures and text. I received editorial assistance and help with interpretations of my results from Rempel, Roering, and Skarbek.

4.1. Introduction

Catastrophic landslides cause billions of dollars in damages and claim thousands of lives annually (Schuster and Highland, 2001; Petley, 2012), while slow-moving landslides with negligible inertia dominate sediment transport on many weathered hillslopes (Mackey and Roering, 2011; Simoni et al., 2013). Surprisingly, both failure modes are displayed by nearby landslides (and single landslides in different years) subjected to almost identical environmental conditions (Iverson et al., 2000; Reid et al., 2003; Wang et al., 2010). This has motivated the search for mechanisms that can cause slow-moving landslides to transition via runaway acceleration to catastrophic failure (Iverson, 2005; Gabet and Mudd, 2006; Goren and Aharonov, 2009; Wang et al., 2010). Here, we describe landslide motion using a rate- and state-dependent frictional model adapted from standard fault mechanics treatments (Dieterich, 1979; Ruina, 1983) to incorporate a non-local stress balance that accounts for the elastic response to gradients in slip along a one-dimensional sliding surface (Segall, 2010). Our idealized model reproduces both the displacement patterns

observed at well-documented slow-moving landslides (Iverson and Major, 1987; Reid et al., 2003; Handwerger et al., 2013) and the acceleration towards failure exhibited by catastrophic events (Saito, 1969; Petley et al., 2002; Helmstetter et al., 2004). Catastrophic failure occurs only when the slip surface is characterized by rate-weakening friction and its lateral dimensions exceed a critical nucleation length h^* that is shorter for higher effective stresses (e.g., thicker sediment packages). However, landslides that are sufficiently extensive to fall within this regime can nevertheless slide slowly for months or years, during which the driving gravitational stress exceeds the steady-state frictional strength. Our results suggest that rate and state friction and elastic stress-transmission play an important role in controlling the slip rates of slow landslides and modulating the transition to catastrophic failure.

Laboratory experiments (Skempton, 1985; Tika et al., 1996; Iverson et al., 2000; Moore and Iverson, 2002; Wang et al., 2010) and numerical models (Angeli et al., 1996; Iverson, 2005) suggest that for slow-moving landslides, which persist over periods of years to centuries (Bovis and Jones, 1992; Mackey and Roering, 2011), the shear strength that resists motion increases with slip rate — a characteristic referred to as rate-strengthening — while the opposite is true for landslides that exhibit runaway acceleration and catastrophic failure. Two primary mechanisms have been frequently invoked to describe the former rate-strengthening behavior. The first characterizes landslide materials as visco-plastic (i.e. Bingham-plastic), so that an increase in velocity corresponds with an increase in strain rate and viscous resistance. These models are able to reproduce field-based measurements of velocity for seasonally active slow-moving landslides (Angeli et al., 1996; Van Asch et al., 2007). However, this constitutive behavior contradicts field and lab measurements that suggest landslide displacement is dominated by frictional sliding along basal and

lateral faults (Hutchinson, 1970; Swanson and Swanston, 1977; Keefer and Johnson, 1983; Cooper et al., 1998; Hungr et al., 2001); furthermore, such rheological models are unable to capture the transition from slow sliding to catastrophic failure. The second modeling approach applies Coulomb friction and invokes shear-zone dilatancy to regulate pore-water pressure changes so that faster sliding increases strength because of increases in effective stress associated with small porosity increases (Segall and Rice, 1995; Iverson, 2005; Gabet and Mudd, 2006; Segall et al., 2010). Such models are able to describe both slow sliding and catastrophic failure because dilatant strengthening occurs only until a critical-state porosity is reached; subsequent rapid consolidation can trigger catastrophic failure. Because slow-moving landslides commonly exhibit seasonal rate changes for many years without interruption, this mechanism requires that the shear-zone material consolidates between repeated slip events, or that displacement be accommodated along multiple shear zones. Field-based analysis has led to problematic assertions that particular slow-moving landslides should have reached their critical-state densities long ago (Schulz et al., 2009b). Furthermore, the mechanics and spatial and temporal scales of consolidation have not been addressed such that the ability of landslides to reacquire dilatant capacity remains unclear.

Models that invoke rate- and state-dependent friction provide an alternative framework for understanding changes in slip velocity in response to the changes in pore-water pressure that are commonly associated with transient precipitation over a broad range of time scales. Similar models that have been refined over the past 30 years yield successful predictions for a wide range of fault behavior, including stable sliding, slow slip and tremor, and the nucleation of earthquakes (Dieterich, 1992; Rubin, 2008; Skarbek et al., 2012). To better understand how stress perturbations drive changes in landslide motion, we formulated a mechanical model that can describe

both slow (characterized by episodic or persistent sliding at low velocities) and accelerating dynamic motion, prior to the development of significant inertial effects (details are contained in Section 4.3 and Appendix C). Here, we use our model to examine two types of landslides that are driven by precipitation-induced changes in pore-water pressure: 1) seasonally active slow-moving landslides and 2) fast-moving catastrophic landslides. For the slow-moving landslides we ran simulations with rate-weakening, rate-strengthening, and rate-neutral frictional properties. In a key advance over previous landslide treatments, we account for the non-local effects of elastic stress-transmission that accompany gradients in slip along the sliding surface (i.e. the landslide is not modeled as an isolated block on a slope) (see Appendix C). This enables us to explore interactions between sliding velocity and both localized and spatially extensive stress perturbations.

First, we define the stability conditions that ultimately determine whether a landslide governed by rate and state friction is capable of gaining significant inertia. Linear stability analysis demonstrates that landslides comprised of rate-strengthening or rate-neutral material, defined so that the frictional parameters satisfy $a > b$ and $a = b$ (see Section 4.3), respectively, will never fail catastrophically (Dieterich, 1979; Rice and Ruina, 1983; Dieterich, 1992). However, landslides comprised of rate-weakening material ($a < b$) have the potential to display both catastrophic (dynamic) and non-catastrophic (slow) motion. For a homogeneous elastic material with shear modulus G , the transition from slow to dynamic motion requires that the length L of the slipping surface exceed a critical nucleation length that is given by Dieterich (1979)

$$h^* = \frac{Gd_c}{\sigma'(b - a)} \quad (4.1)$$

where σ' is the effective normal stress (normal stress minus pore-water pressure), and d_c is the characteristic slip distance for the evolution of frictional contacts, treated here using the Aging (or Dieterich) law (see Section 4.3). A nominal range of parameter values compiled from measured properties in soil and rock mechanics experiments is given in Table C.1. When $L < h^*$, reductions in frictional strength with increased sliding velocity are compensated by elastic stresses associated with large strain gradients along the boundaries of the slipping region, whereas when $L > h^*$ the combined effects of friction and elastic stresses are ultimately unable to balance the driving stress, so unstable acceleration results. Fig. 4.1 shows the parameter space for h^* combined with a typical range of size and effective stress values for landslides. We find that h^* can range from a few centimeters to tens of kilometers (Fig. 4.1). This wide range results from the wide range of parameter values we tested and indicates that a more precise treatment requires further measurements to constrain landslide material behavior in particular instances. Nonetheless, our first-order analysis is consistent with the shows that landslides that display rate-weakening behavior can display both failure modes.

4.2. Results and Discussion

To build upon this framework, we model the motion of seasonally active slow-moving landslides driven by transient stress perturbations. These stress perturbations can be physically interpreted as seasonal fluctuations in pore-water pressure at depth. To ensure slow sliding for model runs with rate-weakening material, we set the nominal $L/h^* = 0.1$. Stress perturbations were chosen to mimic precipitation-induced variations in pore-water pressure measured at the Cleveland Corral and Minor Creek landslides in Northern California (Fig. 4.2c, Section 4.3). These landslides display

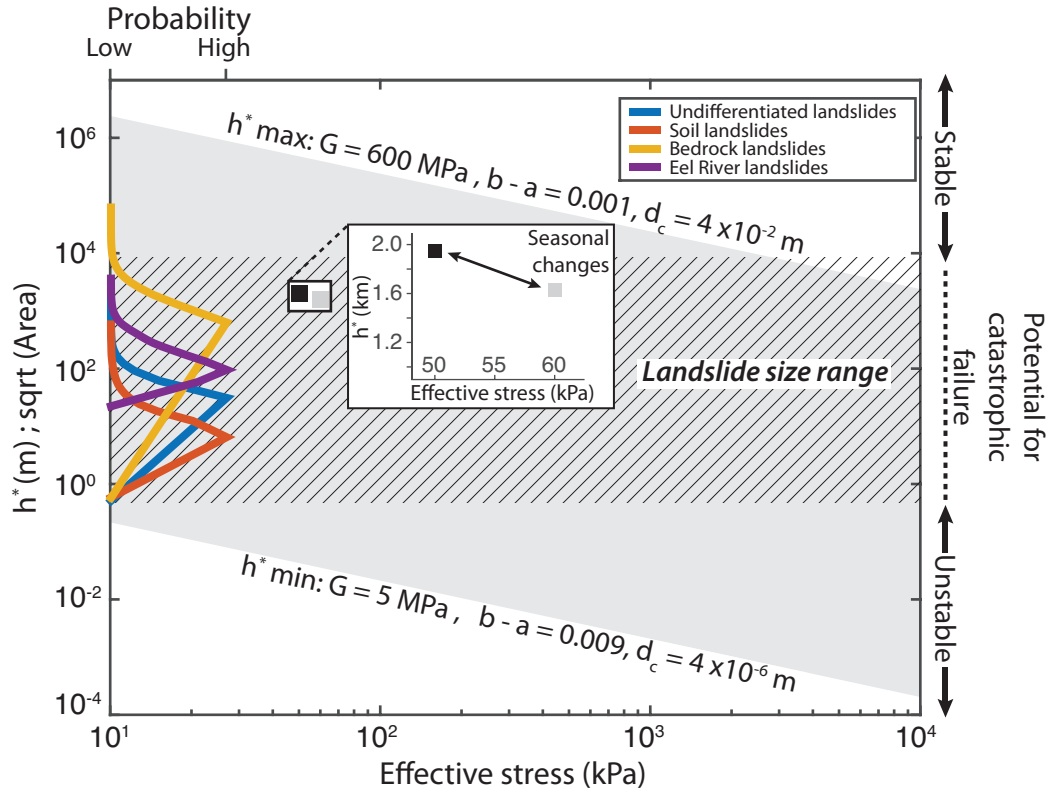


FIGURE 4.1. Critical nucleation length for landslides. Shaded gray area spans the full parameter space. Values for maximum and minimum h^* are indicated on the plot. Other parameters listed in Table C.1. Diagonal lines and kernel density probabilities correspond to measured range of size calculated as the $\sqrt{\text{landslide area}}$ for a global inventory of soil, rock, undifferentiated landslides (Larsen et al., 2010; Mackey and Roering, 2011). Sliding mode corresponds to stable (i.e. non-catastrophic) and unstable (i.e. catastrophic). Inset shows how seasonal changes in pore-water pressure affect h^* .

distinct seasonal velocity changes, with periods of increased velocity that correspond to periods of elevated pore-water pressures in the landslide body. Numerous other landslides in this region display similar displacement patterns and are likely driven by similar stress perturbations. Our model successfully reproduces the general patterns of displacement that result. Each modeled landslide displays an instantaneous velocity increase in response to the imposed decrease in effective stress (Fig. 4.2). The ensuing period of acceleration is followed by a longer period of deceleration that lasts for the majority of the year. We also find that changing the material properties results in slight differences in the timing and magnitude of the seasonal displacement. Simulations with rate-weakening material display the highest peak velocity and cumulative displacement, followed by rate-neutral and rate-strengthening materials (Fig. C.1). These relatively minor changes suggest that material controls play a secondary role in governing the overall landslide behavior in this regime. We emphasize that in this work we used the field data to determine only the timing and magnitude of realistic stress perturbations; we did not calibrate our model parameters to match the displacement data.

We also examine the time-dependent friction coefficient, defined by the rate and state friction law (equation 4.2), and resisting stress defined as effective stress multiplied by the friction coefficient, at fixed locations along the landslide base. The landslide is in a constant state of failure such that the total driving stress equals the resisting stress (equation 4.4). During periods of enhanced seasonal motion, the resisting stress and friction coefficient generally parallel the landslide velocity due to the ‘direct effect’, which describes the instantaneous strength increase that accompanies faster sliding (Fig. 4.2b). This strength increase acts to overcome the strength decrease due to the reduction in effective stress. When the effective stress

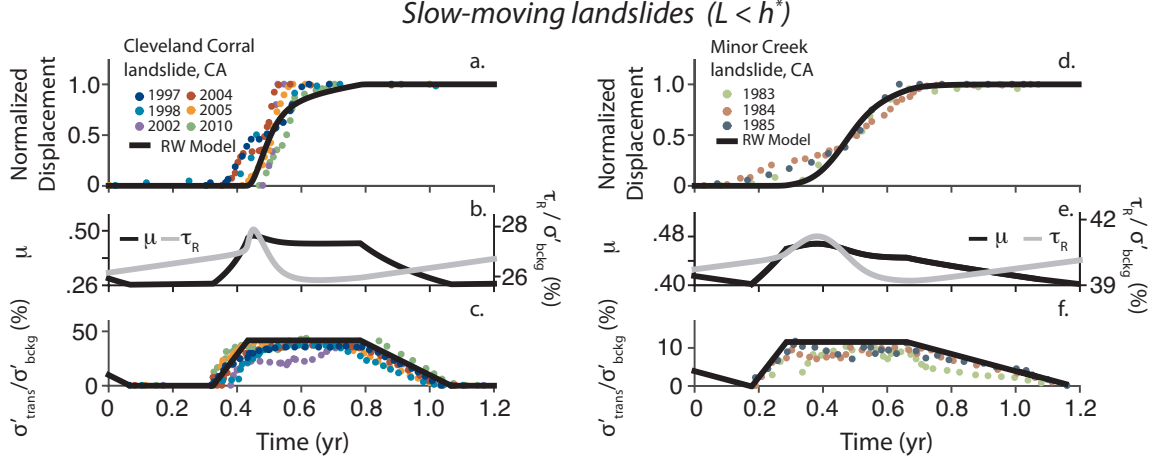


FIGURE 4.2. Model simulations for slow-moving landslides. Field data is broken into individual water years (starting Oct 1.). a., d., Normalized displacement time series for field-based measurements and rate-weakening model plotted at a representative node. Displacement is normalized by yearly maximum in order to make direct comparisons. b., e., Friction coefficient, μ , and normalized resisting stress, τ_R/σ'_{bckg} . Resisting stress is defined as friction coefficient times the effective stress. c., f., Normalized stress perturbations for each landslide. σ'_{trans} corresponds to the transient reduction in effective stress that results from an increase in pore-water pressure.

stabilizes at a constant, lower value during the wet season, the landslide velocity continues to evolve and elastic stress-transmission contributes to maintaining the force balance. There is a brief period of steady-state friction (i.e. lower friction coefficient with higher velocity) during the simulated wet season for the Cleveland Corral-like landslide, which is driven by larger magnitude stress perturbations, and during the simulated dry season for both landslides (Fig. 4.2b; Fig. 4.2e).

To explore how transient stress perturbations can trigger runaway acceleration and catastrophic failure, we ran simulations with rate-weakening material and values of L/h^* between 2 and 8. We perturbed the stress along the sliding surface using a time-dependent linear function to simulate the onset of deep pore pressure changes during rainfall events (Fig. 4.3c, Section 4.3). Acceleration begins gradually as the effective stress decreases, with slow motion persisting for several months before a rapid

transition to runaway acceleration (Fig. 4.3). During this period of slow motion, the friction coefficient and resisting stress exhibit only minor changes (Fig. 4.3b), with the former increasing modestly due to the ‘direct effect’, and the latter slowly dropping with the decreased effective stress. As the landslide approaches catastrophic failure, rapid velocity increases drive further increases to the friction coefficient until $v \sim d_c/\theta$ (equation 4.1), which subsequently causes a reduction in frictional strength that cannot balance the driving stress. The modeled displacement pattern is similar to the creep-to-failure motion that has been observed prior to several catastrophic landslides (Fig. C.2).

Finally, in an attempt to understand spatially heterogeneous stress perturbations along the landslide body, we triggered motion using localized stress perturbations. These stress perturbations can be physically interpreted as pore-water pressure changes facilitated by preferential flow paths (e.g., macropores, deformation cracks). For model runs in the slow sliding regime, local stress perturbations trigger seasonal motion that spreads along the sliding surface. The highest peak velocity and cumulative displacement occur along the section of the slip surface subjected to the stress perturbation (Fig. C.3). Similar behavior occurs for model runs in the dynamic sliding regime up until the point of catastrophic failure (Fig. C.4).

When driven by realistic stress perturbations along the sliding surface, our model reproduces deformation patterns observed at landslides and provides new insights into the mechanisms that control the mode of failure. Landslides comprised of rate-weakening material display both slow (seasonal) and dynamic (catastrophic) sliding with the character depending on the size of the sliding surface, or slip patch, relative to the critical nucleation length. This is directly analogous to behavior that is well established in fault mechanics, where a similar variety of slip modes is observed.

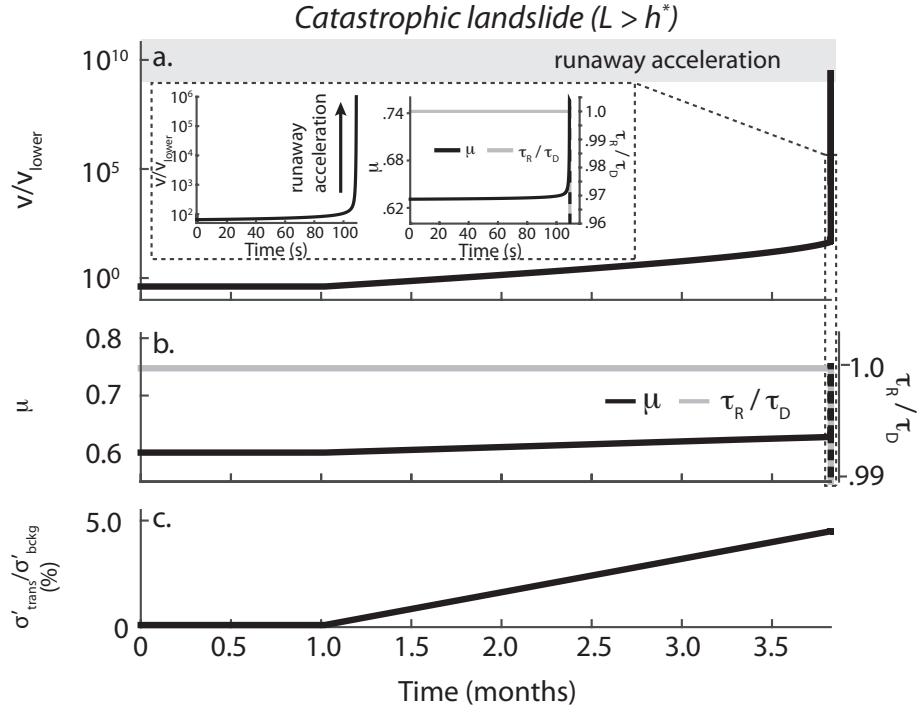


FIGURE 4.3. Model simulation for catastrophic landslide. a) Normalized velocity time series leading up to catastrophic failure plotted at a representative model node. Shaded gray area corresponds to v_{dyn} and signifies runaway acceleration. b) Friction coefficient, μ , and ratio of resisting stress to driving stress, τ_R/τ_D . Resisting stress and driving stress are calculated as the righthand side and lefthand side of equation 4.4, respectively. Friction coefficient rapidly increases and then rapidly decreases (shown with dashed line segment). c) Normalized stress perturbation where σ'_{trans} corresponds to the transient reduction in effective stress that results from increasing pore-water pressure. Inset in a) shows zoom in of normalized velocity, friction coefficient, and stress ratio. Dashed box shows location of data shown in the inset. Note the inset shows the time in seconds where 0 is the first time step corresponding to the zoomed in region.

Although equation (4.1) provides a first-order estimate for the length scale that determines failure mode, previous studies on tectonic faults have shown that multiple length scales can be relevant to earthquake nucleation, including those arising from additional physical interactions, such as dilatancy, and from geometrical and material irregularities.

Similar to tectonic faults, there are likely multiple length scales relevant to landslide failure mode and furthermore, some of these length scales can vary in time. Paradoxically, equation (4.1) predicts that landslides driven by transient stress perturbations are somewhat more stable during periods of motion since h^* increases with decreases in effective stress (Fig. 4.1 inset). This finding is particularly important for understanding the behavior of landslides that are at or near the critical nucleation length. Furthermore, landslide failure mode may also depend on the magnitude of the stress perturbation when the slip surface length is close to the critical nucleation length. Model simulations with $L/h^* = 2$ display slow sliding if the stress perturbations are $< 4\%$ of the background effective stress (Fig. C.5), or occur along a patch size that is $< 75\%$ of the sliding surface length L (Fig. C.4b). Yet, this perturbation size threshold decreases as the slip surface length becomes several times the critical nucleation length (Fig. C.6). This confirms the intuition that the size of the stress perturbation provides a further control on landslide failure mode.

Given that slow-moving landslides occur in a wide variety of lithologic units, we expect many of them to contain rate-weakening material. We hypothesize that slow-moving landslides containing rate-weakening material rarely fail catastrophically because the spatial dimensions of their most actively slipping regions are less than the critical nucleation length (Fig. 4.1). For the critical nucleation length to be larger than the typical spatial dimensions of slow-moving landslides, the landslide material must

have a high shear modulus, high characteristic slip distance, or low effective stress. While the shear modulus and characteristic slip distance can vary between landslides, particularly those occurring under different environmental conditions, we expect the effective stress to exert a primary control on landslide failure mode. In agreement with field-based measurements, our analysis suggests that landslides characterized by low effective stress have a larger critical nucleation length, and so are less likely to exhibit catastrophic failure, when compared to landslides with higher background levels of effective stress.

Although the rate and state framework provides a mechanistic interpretation for the boundaries between stable and unstable sliding regimes, the detailed processes by which landslides transition from one regime to the other are less certain. We hypothesize that the spatial dimensions of the landslide, or slip patch may evolve in time until $L > h^*$. Landslides tend to grow until they span from hilltops to channel bottoms. Furthermore, landslides can display distinct kinematic elements that differ in the timing of motion along the landslide body. These differences in motion may reflect the presence of temporary slip barriers (i.e. strong patches) that are able to contribute disproportionate resistance to the downslope translation of the landslide material. This suggests that landslides that exhibit complex motion, consisting of an ensemble of smaller slip patches, for example resulting from heterogeneity in basal strength, may be less likely to fail catastrophically when compared to landslides that exhibit more uniform motion. The Cleveland Corral landslide, which is located within 3 km of two large landslides that failed catastrophically, may provide an opportunity to test this hypothesis. In 1997 the USGS installed a long-term monitoring station in an attempt to capture the transition from slow motion to catastrophic failure (Reid et al., 2003). So far the landslide has only displayed seasonal displacement, but

when it moves it typically displays complex, nonuniform motion along its entire body. Furthermore, the landslide has developed new shear margins and kinematic elements over the past two decades. It is possible that further landslide growth, as gauged by an evolution towards more uniform motion, may presage catastrophic failure.

4.3. Methods

4.3.1. Rate and state model.

We consider an idealized one-dimensional sliding surface governed by rate- and state-dependent friction. The sliding surface has non-zero velocity at all times, although during some periods sliding proceeds at negligible rates, compared with those applied on the boundaries (see Appendix C). The friction coefficient on the landslide sliding surface is described by (Dieterich, 1979; Ruina, 1983)

$$\mu = \mu_0 + a \ln \left(\frac{v}{v_0} \right) + b \ln \left(\frac{v_0 \theta}{d_c} \right), \quad (4.2)$$

where μ_0 is the steady-state friction coefficient at reference velocity v_0 , a and b are friction parameters that control whether μ tends towards higher or lower values as slip rate v changes, and d_c is the characteristic slip distance for the evolution of frictional contacts. The state variable θ , which is conceptualized as the average life span of a set of frictional contacts, evolves according to the Dieterich (or Aging) law where

$$\frac{d\theta}{dt} = 1 - \frac{v\theta}{d_c}, \quad (4.3)$$

so that $\mu = \mu_0 + (a - b) \ln(v/v_0)$ at steady state. Stress balance requires that

$$\tau_d + \frac{G}{2\pi(1 - \nu)} \int_{-\infty}^{+\infty} \frac{d\delta/d\xi}{\xi - x} d\xi = \mu\sigma' + \frac{G}{2v_s} v, \quad (4.4)$$

where τ_d is the driving stress due to gravity, σ' is the effective normal stress, G is the shear modulus, ν is the Poisson ratio, v_s is the shear-wave speed, and the integral dependence on $d\delta/d\xi$ accounts for the effects of spatial gradients in slip along the sliding surface at local coordinate x . The first term on the right is the frictional stress and the second is the “radiation-damping” approximation for inertial effects, which become important only when the slip rate approaches $v_{dyn} \sim 2\sigma' a v_s / G$ (Rice, 1993). The second term on the left is the elastic stress, which is approximated by a rigorous treatment for dislocation motion in a whole space.

4.3.2. Stress perturbations.

Stress perturbations due to transient pore-water pressure are simulated with spatial and temporal changes in effective normal stress. For the slow-moving landslide calculations, we approximate the timing, magnitude, and duration of transient pore-water pressure using a piecewise linear function that broadly captures the patterns of pore-water pressure change measured in deep piezometers at the Cleveland Corral (Reid et al., 2003) and Minor Creek landslides (Iverson and Major, 1987), Northern California. Annual pore-water pressures vary by $\sim 40\%$ of the background (i.e. dry season) effective stress at Cleveland Corral and $\sim 10\%$ of the background effective stress at Minor Creek (Fig. 4.2c). For the catastrophic landslide calculations, we perturb the stress along the sliding surface using a linear function that increases by 5% of the background effective stress (Fig. 4.3c).

CHAPTER V

CONCLUSIONS

5.1. Dissertation Summary

This dissertation explores controls on the kinematics of slow-moving landslides using state-of-the-art remote sensing techniques and mechanical modeling. In Chapter II, I use satellite-based InSAR and high-resolution topographic data to identify 50 active landslides across a $\sim 14,000 \text{ km}^2$ area in the Northern California Coast Ranges. All but one of these landslides occur within the same lithologic unit which suggests that material properties (e.g., strength) control their spatial distribution in mountainous regions. I also develop advanced InSAR processing methods that enable me to perform time series inversions along multiple landslides. Each landslide exhibits distinct kinematic zones that vary in mean annual velocity. By comparing the InSAR data with displacement maps derived from historic aerial photos (Mackey and Roering, 2011), I find that the spatial location of the kinematic zones has remained fixed for almost 70 years. Detailed analysis of four representative landslides, which differ only in geometry, shows that each landslide displays remarkably similar velocity changes in response to transient stress perturbations from seasonal rainfall. These behavioral patterns indicate that landslide geometry sets the location of kinematic zones but does not strongly control their response to transient stress perturbations. Lastly, I use a mathematical model to explore the subsurface controls on the observed landslide motion. Using the model, I estimate the thickness and basal topography along the longitudinal axis of a single landslide. I find that both thickness and

basal topography are highly variable, which may have major implications for the mechanical-hydrologic feedbacks that govern landslide motion.

In Chapter III, I quantify the velocity time series of 10 slow-moving landslides (a subset of the landslide inventory from Chapter II) and explore geometrical controls on their kinematics in the context of a commonly used hydrologic model. According to model predictions, precipitation increases the pore-water pressure at the ground surface which then diffuses vertically to the basal sliding surface. This implies that the length of time required for landslides to respond to rainfall should strongly depend on the thickness of the landslide. However, despite a five-fold variation in thickness, I show that each landslide displays no detectable difference in their seasonal response. These findings call into question the assumptions made in many landslide models.

Finally in Chapter IV, I develop a mechanical model that can characterize landslide motion. Using the model, I explore interactions between transient stress perturbations and landslide motion. I test model predictions using field-based measurements of pore-water pressure from well-documented landslides and find that the model is able to accurately reproduce the observed patterns of landslide displacement. By changing the landslide size, effective stress, and mechanical properties, the model is able to produce a variety of displacement patterns, including seasonal motion and catastrophic failure. Furthermore, I describe the environmental conditions under which a landslide can transition from slow stable motion to runaway acceleration and catastrophic failure.

APPENDIX A

SUPPLEMENTARY MATERIAL FOR CHAPTER II

A.1. Projecting LOS Velocity onto the Downslope Sliding Direction

A.1.1. Methodology

To directly compare landslides to each other and to the historic landslide record (Mackey and Roering, 2011), we back-projected the LOS velocity onto the downslope sliding direction using a straightforward vector manipulation. First, we define the unit look vector in the direction of the satellite LOS

$$\hat{l}_{sat} = \cos(\gamma) \sin(\theta) \hat{n} + \sin(\gamma) \sin(\theta) \hat{e} + \cos(\theta) \hat{z} \quad (\text{A.1})$$

where $\hat{l}_{sat}$ is the unit look vector, γ (with units of degrees) defines the orientation of the look vector in the horizontal plane from North, θ is the look angle from nadir, and $\hat{n}$, $\hat{e}$, and $\hat{z}$, correspond to the North, East, and vertical components, respectively. These values are taken directly from the ALOS PALSAR satellite parameter file. Next, we define the unit vector for the downslope sliding direction for each landslide

$$\hat{h}_{ls} = \cos(\zeta) \sin(90 - \alpha) \hat{n} + \sin(\zeta) \sin(90 - \alpha) \hat{e} + \cos(90 - \alpha) \hat{z} \quad (\text{A.2})$$

where $\hat{h}_{ls}$ is the downslope direction unit vector, the angle ζ defines the orientation of the downslope direction (i.e., azimuth) in the horizontal plane from North, and the angle α is the average slope angle. To measure ζ and α , we use the DEM and the horizontal orientation derived from tree-vector motion (Mackey and Roering, 2011). We find that the downslope sliding direction closely matches the direction of steepest

descent along the slide axis. Then we calculate the downslope velocity at a given pixel i

$$V_i = \Delta\phi_i(\hat{l}_{sat} \cdot \hat{h}_{ls})^{-1} \quad (\text{A.3})$$

where V_i is the downslope velocity and $\Delta\phi_i$ is the LOS velocity. To avoid projection errors, we only project the LOS velocity onto the downslope direction in areas where we have high quality measurements to constrain the downslope azimuth and hillslope angle.

A.1.2. Limits of the LOS Viewing Geometry

As mentioned in Section 2.5.1, InSAR cannot detect surface deformation that occurs along the satellite’s flight path. Therefore, we are unable to measure landslide motion along hillslopes and landslides that have a poor viewing geometry, and along individual landslides that display changes in downslope direction along the length of the landslide.

To show the effects of the viewing geometry on measurements of landslide motion, we present a case study at the Kekawaka Creek 1 (Kw1) earthflow (Figure A.1). The Kw1 earthflow displays an almost 90° change in direction transitioning from the transport to toe zone. This change in direction results in a switch from ground motion away from the satellite (positive range change, red color) to motion towards the satellite (blue color), and we observe no deformation when the landslide motion crosses the orientation of the satellite’s flight path.

We also calculated a synthetic example to further emphasize how the direction of ground motion affects our ability to detect landslide displacement (Figure A.1c). Our synthetic example is constructed using an idealized landslide that has a constant slope α and velocity V . We then vary the landslide’s direction of motion ζ and rearrange

Eq. (A.3) to project displacement onto the satellite's LOS

$$\Delta\phi_i(\zeta) = V(\hat{l}_{sat} \cdot \hat{h}_{ls}(\zeta)) \quad (\text{A.4})$$

The compass plot shows how changes in hillslope aspect from 0 to 360° (where 0° is North) influence the theoretically measured LOS velocity.

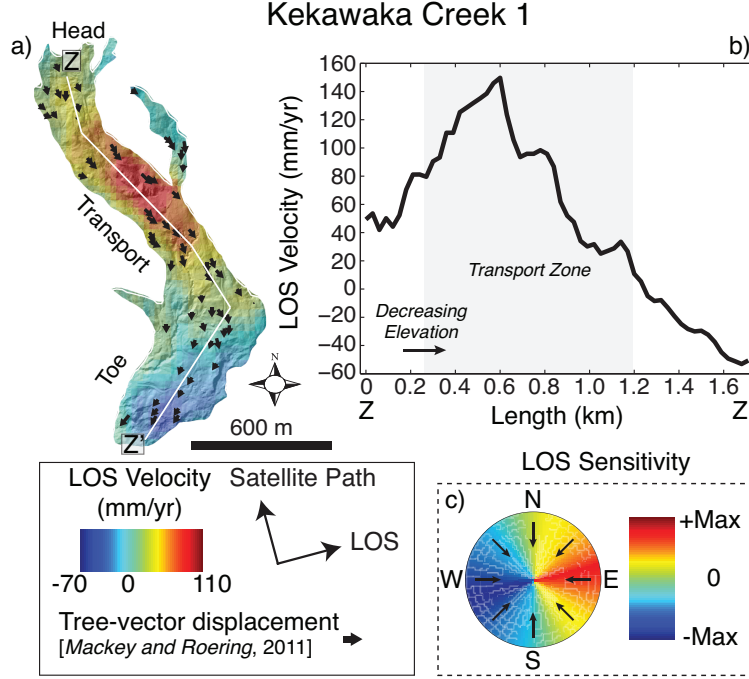


FIGURE A.1. Limits of the line-of-sight (LOS) viewing geometry. a) 4-year averaged interferogram draped over lidar hillshade map. Velocity map shows LOS velocity values with red colors marking the location of the transport zone. The color scales are saturated such that the actual velocity values (mm yr^{-1}) exceed the maximum and minimum listed values. Black vectors show landslide motion mapped using historic aerial photos (Mackey and Roering, 2011). White solid line indicates location of velocity profile shown in b). Slide is delineated into head, transport, and toe zone. b) profile from Z to Z' showing LOS velocity (mm yr^{-1}). Shaded gray area marks the spatial location of the transport zone. Note the change from negative LOS velocity (motion away from satellite), to zero velocity, to positive LOS velocity (motion towards satellite) as landslide changes downslope direction. c) synthetic example to show how the direction of ground motion affects observed landslide displacement. Compass plot shows how changes in landslide aspect from 0 to 360° (where 0° is North) influence the theoretically measured LOS velocity.

A.2. Figures

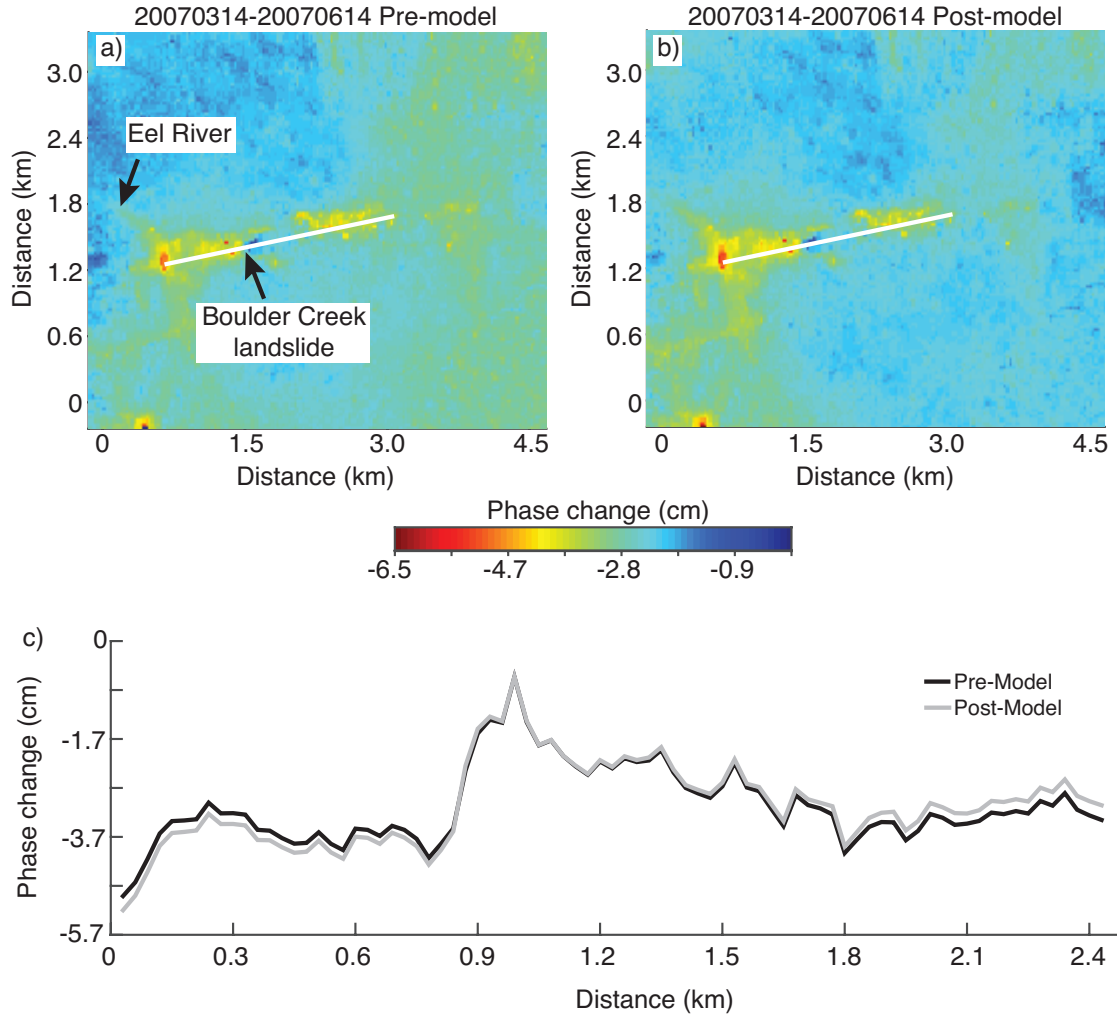


FIGURE A.2. 92-day wrapped interferograms. a) unaltered wrapped 92-day interferogram before deformation model is applied. b) optimized wrapped interferogram after using the deformation model to help with flattening and re-estimating the baselines. c) profile lines show the data are nearly identical (slight differences due to different reference points and orbital gradient) and no deformation artifacts are introduced by our methodology. See Figure 2.2 for the unwrapped interferograms and deformation model.

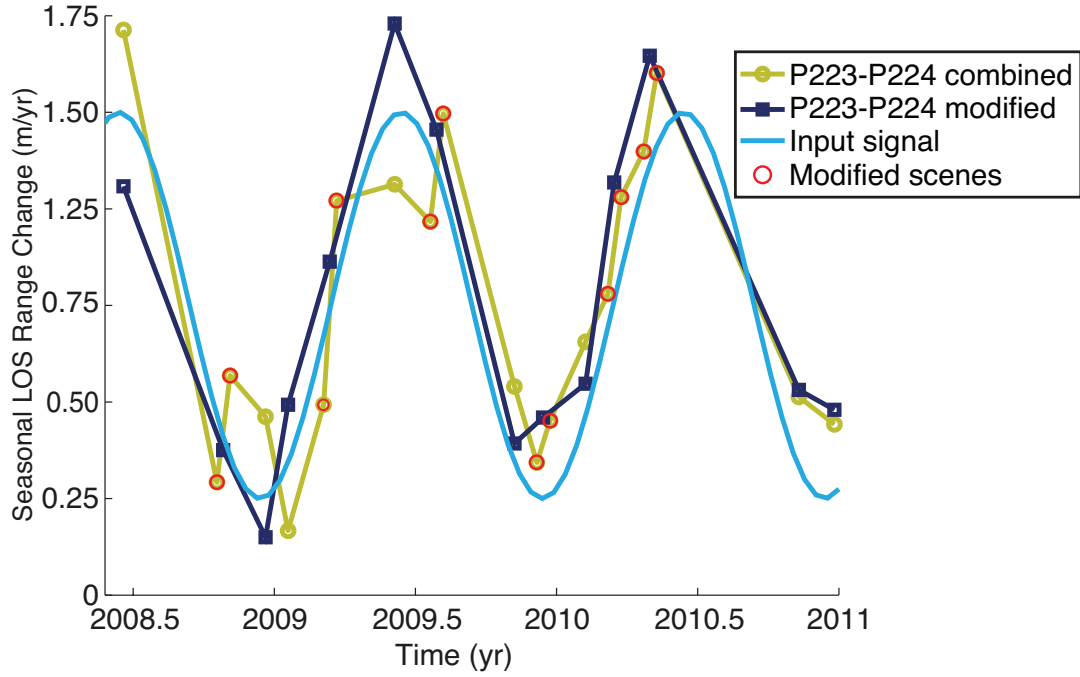


FIGURE A.3. Seasonal LOS range change for synthetic dataset. Light blue line corresponds to input signal from the synthetic model (see description in section 2.4.5). Yellow line corresponds to inversion output using the “P223-P224 combined” dataset. Red circles highlight the modified acquisition dates to construct the “P223-P224 modified” dataset. Dark blue line corresponds to inversion output using “P223-P224 modified”. Note the inversion using the “P223-P224 modified” dataset does not contain the high frequency oscillations contained in the inversion of the “P223-P224 combined” dataset.

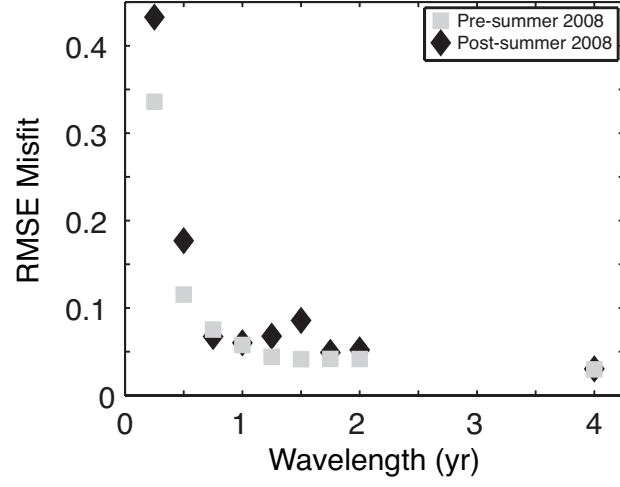


FIGURE A.4. Room mean square error (RMSE) inversion misfit as a function of seasonal wavelength, λ . Due to the lack of data spanning the Summer 2008, we separated the inversion into “Pre-summer 2008” and “Post-summer 2008”. We find that the inversion misfit decreases exponentially as the seasonal wavelength increases such that we can resolve seasonal wavelengths ≥ 0.5 years.

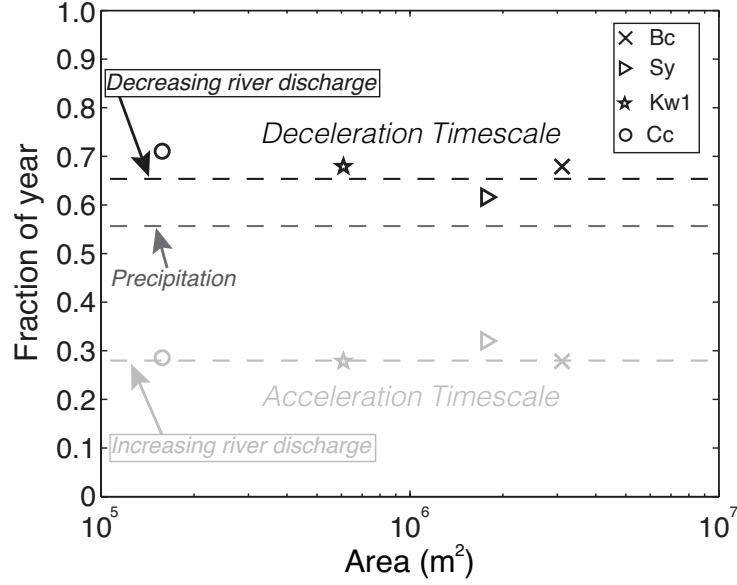


FIGURE A.5. Fraction of year that the Eel River earthflows are accelerating and decelerating calculated as (total days) / (days in a year). We calculate the acceleration (deceleration) timescales as the time difference between the onset of acceleration (deceleration) and the onset of deceleration (acceleration) each year. Symbols indicate average values for the full study period. Black symbols correspond to periods of deceleration and light gray symbols correspond to periods of acceleration. Note the asymmetric velocity pattern characterized by a short period of acceleration that lasts $\sim 30\%$ of the year and a longer period of deceleration that lasts $\sim 70\%$ of the year. Dashed lines correspond to periods of increasing (light gray) and decreasing discharge (black) in the Eel River measured in Scotia, CA and the fraction of the year with precipitation (dark gray). The coincident correlation between discharge and slide motion may indicate a tightly coupled hydrologic system.

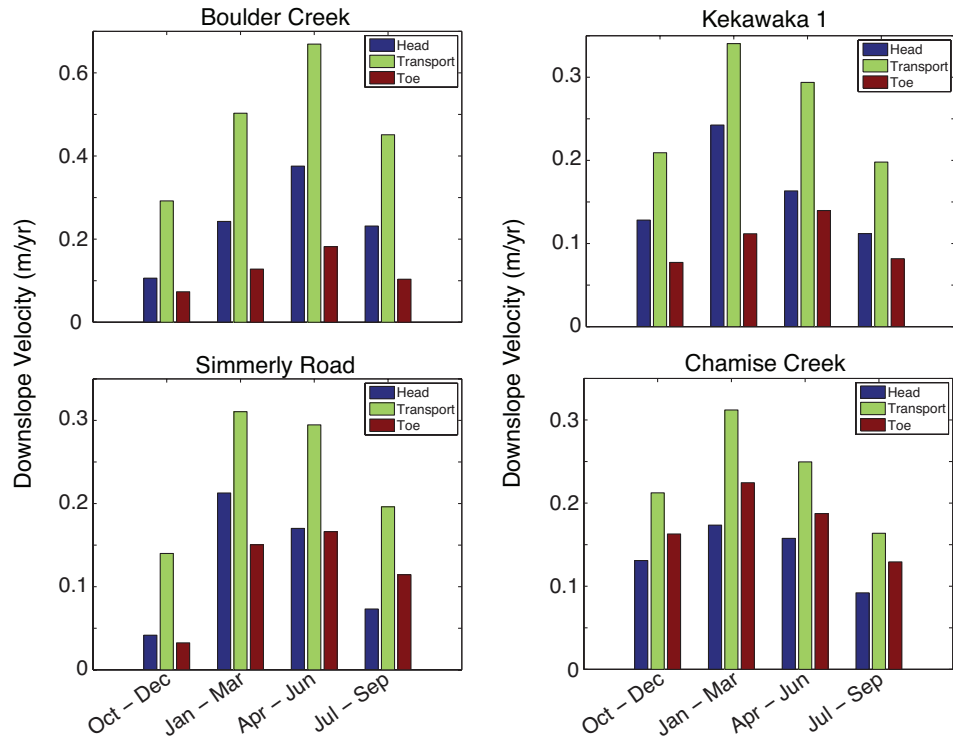


FIGURE A.6. Average downslope velocity of each kinematic zone binned into 3-month periods starting at the onset of seasonal rainfall. Wet season velocities are highest in each kinematic zone. Transport zone rates (green) can exceed the head (blue) and toe (red) zone rates by factor of 4, while the head and toe zone rates are within a factor of 2. See kinematic zone delineations in Figure 2.4.

A.3. Tables

ALOS P223 F790					ALOS P224 F790				
No.	Master	Slave	Spatial Baseline (m)	Temporal Baseline (days)	No.	Master	Slave	Spatial Baseline (m)	Temporal Baseline (days)
1	20070314	20070429	499	45	32	20070213	20070516	459	94
2	20070314	20070614	-258	91	33	20070213	20070701	634	140
3	20070314	20070730	474	138	34	20070213	20070816	921	185
4	20070314	20070914	663	139	35	20070516	20070701	175	46
5	20070429	20070614	210	46	36	20070516	20070816	461	91
6	20070429	20070730	81	92	37	20070516	20071001	720	137
7	20070429	20070914	164	138	38	20070516	20071116	1171	183
8	20070614	20070730	-129	46	39	20070516	20080101	1054	228
9	20070614	20070914	684	91	40	20070701	20070816	287	45
10	20070614	20080130	1874	229	41	20070701	20071001	545	91
11	20070730	20070914	-481	45	42	20070701	20071116	996	137
12	20070730	20080130	1378	183	43	20070701	20080101	879	183
13	20070914	20080501	2141	230	44	20070816	20071001	258	46
14	20080130	20080501	951	93	45	20070816	20071116	709	91
15	20081101	20090201	695	91	46	20070816	20080101	592	137
16	20081101	20090319	1194	137	47	20070816	20080216	1676	183
17	20081101	20090804	1119	277	48	20071001	20071116	451	45
18	20090201	20090319	499	48	49	20071001	20080101	334	91
19	20090201	20090619	846	140	50	20071116	20080101	-117	46
20	20090201	20090804	425	185	51	20071116	20080216	967	91
21	20090319	20090619	347	91	52	20071116	20080518	1354	184
22	20090319	20090804	-74	140	53	20080216	20080518	387	93
23	20090319	20091220	1337	275	54	20080703	20081118	-1430	137
24	20090619	20090804	-422	46	55	20080703	20090103	-1216	183
25	20090619	20091220	989	183	56	20080703	20090405	-193	276
26	20090804	20091220	1411	137	57	20081118	20090103	214	46
27	20091220	20100322	1006	93	58	20081118	20090405	1237	139
28	20091220	20100507	1153	138	59	20090103	20090405	1023	93
29	20100322	20100507	147	46	60	20090103	20090821	1369	231
30	20100507	20101223	966	229	61	20090405	20090821	346	138
31	20101223	20110325	1110	93	62	20090405	20091121	1026	229
					63	20090405	20100106	1264	275
					64	20090821	20091121	680	91
					65	20090821	20100106	918	137
					66	20090821	20100221	1616	183
					67	20090821	20100408	1785	230
					68	20091121	20100106	238	46
					69	20091121	20100221	936	91
					70	20091121	20100408	1105	139
					71	20091121	20100524	1146	185
					72	20100106	20100221	698	45
					73	20100106	20100408	867	93
					74	20100106	20100524	907	140
					75	20100221	20100408	169	48
					76	20100221	20100524	210	94
					77	20100408	20100524	40	46
					78	20100408	20101124	925	229
					79	20100524	20101124	884	183
					80	20100524	20110109	1327	228
					81	20101124	20110109	443	46

TABLE A.1. ALOS PALSAR interferogram pairs from satellite paths 223 F790 and 224 F790. No. corresponds to the cumulative number of interferogram pairs. Master and Slave correspond to the acquisition date (yyyymmdd) of each scene. Spatial baseline is the perpendicular baseline (m) between the SAR scene acquisitions and temporal baseline is the time period between acquisitions.

No.	Master	Slave	Spatial Baseline (m)	Temporal Baseline (days)	Path No.
1	20070314	*20070507	499	46	P223
2	20070314	*20070622	-258	91	P223
3	20070314	*20070807	474	138	P223
4	20070314	*20070922	663	182	P223
5	*20070507	*20070622	175	46	P224
6	*20070507	*20070807	461	91	P224
7	*20070507	*20070922	720	137	P224
8	*20070507	*20071116	1171	183	P224
9	*20070507	20080101	1054	229	P224
10	*20070622	*20070807	-129	46	P223
11	*20070622	*20070922	684	91	P223
12	*20070622	20071116	996	137	P224
13	*20070622	20080101	879	183	P224
14	*20070807	*20070922	258	46	P224
15	*20070807	20071116	709	91	P224
16	*20070807	20080101	592	138	P224
17	*20070922	20071116	451	46	P224
18	*20070922	20080101	334	91	P224
19	20071116	20080101	-117	46	P224
20	20071116	20080216	967	91	P224
21	20080101	20080216	1084	46	P224
Data gap					
22	20080703	*20081109	-1430	136	P224
23	20080703	*20090327	-193	275	P224
24	*20081109	20090103	-117	46	P224
25	*20081109	20090201	694	91	P223
26	*20081109	*20090327	1194	140	P223
27	*20081109	*20090812	1119	276	P223
28	20090103	*20090327	1023	93	P224
29	20090201	*20090327	499	48	P223
30	20090201	20090619	846	140	P223
31	20090201	*20090812	425	186	P223
32	*20090327	20090619	347	91	P223
33	*20090327	*20090812	-74	137	P223
34	*20090327	20091121	1026	229	P224
35	*20090327	*20091228	1264	275	P224
36	20090619	*20090812	-422	46	P223
37	*20090812	20091121	680	91	P224
38	*20090812	*20091228	918	137	P224
39	20091121	*20091228	238	46	P224
40	20091121	20100221	936	91	P224
41	20091121	*20100330	1105	139	P224
42	*20091228	20100221	698	46	P224
43	*20091228	*20100330	1006	93	P223
44	*20091228	*20100515	1153	139	P223
45	20100221	*20100330	169	48	P224
46	20100221	*20100515	210	94	P224
47	*20100330	*20100515	40	46	P224
48	*20100330	20101124	925	228	P224
49	*20100515	20101124	884	182	P224
50	*20100515	20110109	1327	228	P224
51	20101124	20110109	443	46	P224

TABLE A.2. ALOS PALSAR data used in time series inversion modified from Handwerger et al. (2013). No. corresponds to the cumulative number of interferogram pairs. Master and Slave correspond to the acquisition date (yyyymmdd) of each scene. Spatial baseline is the perpendicular baseline (m) between the SAR scene acquisitions and temporal baseline is the time period between acquisitions. Path number corresponds to satellite path P223/4 (Figure 2.1). * correspond to scenes with modified dates (see section 2.4.5 for description).

Date changed from:	Date changed to:	Path No.
20070429	20070507	P223
20070516	20070507	P224
20070614	20070622	P223
20070701	20070622	P224
20070730	20070807	P223
20070816	20070807	P224
20070914	20070922	P223
20071001	20070922	P224
20081101	20081109	P223
20081116	20081109	P224
20090319	20090327	P223
20090405	20090327	P224
20090804	20090812	P223
20090821	20090812	P224
20091220	20091228	P223
20100106	20091228	P224
20100322	20100330	P223
20100408	20100330	P224
20100507	20100515	P223
20100524	20100515	P224

TABLE A.3. Original ALOS SAR scene dates, modified SAR scene dates, and corresponding path number.

Definition	Value	Source
Wet density of earthflow colluvium, ρ [kg m^{-3}]	2145	Bill Schulz, USGS, unpublished data (2013)
Flow-law exponent, p []	3	Glen, (1955); Booth et al. (2013)
Flow-law constant, c [$\text{m kg}^{-1} \text{yr}^{5/3}$]	4.6×10^{-21}	Booth et al. (2013)

TABLE A.4. Nominal parameter values.

No.	Area (m ²)	Length (m)	Average Width (m)	Aspect Ratio ()	Estimated Thickness (m)	Estimated Volume (m ³)	Elevation Range (m)	Longitudinal Slope (°)	Landslide Aspect (°)	Average Downslope Velocity (m/yr)	
(Cc)	1	52755	400	132	3.03	11	5.80E+05	74	10	292	0.37
	2	98340	595	165	3.6	*10	9.83E+05	158	15	126	0.27
	3	103880	528	197	2.68	13	1.35E+06	163	17	31	0.32
	4	107977	592	182	3.25	13	1.40E+06	131	12	57	0.19
	5	135736	908	149	6.1	14	1.90E+06	260	16	27	0.20
	6	156395	850	184	4.62	*10	1.56E+06	281	18	65	0.13
	7	157549	980	161	6.1	*15	2.36E+06	247	14	124	0.30
	8	181444	793	229	3.46	15	2.72E+06	228	16	48	0.11
	9	201542	678	297	2.28	16	3.22E+06	181	15	48	0.13
	10	207134	1360	152	8.93	*10	2.07E+06	377	16	56	0.16
	11	208199	1502	139	10.84	*10	2.08E+06	330	12	209	0.47
	12	215306	901	239	3.77	16	3.44E+06	247	15	268	0.38
	13	219449	1111	198	5.61	16	3.51E+06	218	11	34	0.28
	14	232477	2140	109	19.7	*8	1.86E+06	398	11	302	0.55
	15	246630	1501	164	9.15	17	4.19E+06	454	17	59	0.13
(Kw1)	16	260384	1235	211	5.86	17	4.43E+06	276	12	138	0.26
	17	260533	1409	185	7.6	17	4.43E+06	339	14	73	0.13
	18	266737	2385	112	21.33	*5	1.33E+06	596	14	230	0.41
	19	283146	1426	199	7.17	18	5.10E+06	275	11	173	1.74
	20	304104	1650	184	8.95	*15	4.56E+06	434	15	108	0.24
	21	323145	1647	196	8.4	18	5.82E+06	317	11	100	0.18
	22	359329	1960	183	11	*20	7.19E+06	581	17	201	1.60
	23	363581	1735	210	8.26	19	6.91E+06	378	12	108	0.27
	24	395882	955	415	2.3	19	7.52E+06	216	13	48	0.15
	25	421965	1865	226	8.24	*25	1.05E+07	528	16	206	0.63
	26	434214	1521	285	5.34	20	8.68E+06	348	13	133	0.29
	27	483543	1736	279	6.22	20	9.67E+06	364	12	54	0.26
	28	490617	1395	352	3.96	21	1.03E+07	362	15	75	0.24
	29	494607	1726	287	6.01	21	1.04E+07	539	17	151	0.30
	30	504989	2250	224	10.04	21	1.06E+07	570	14	295	1.00
(Sy)	31	534981	1990	269	7.4	*30	1.60E+07	545	15	55	0.21
	32	549036	1938	283	6.85	21	1.15E+07	526	15	268	0.41
	33	609348	2010	303	6.63	*40	2.44E+07	413	12	141	0.45
	34	621465	1596	389	4.1	22	1.37E+07	331	12	47	0.21
	35	641852	2093	307	6.82	22	1.41E+07	489	13	74	0.20
	36	645905	1284	503	2.55	22	1.42E+07	479	20	132	0.25
	37	694295	1397	497	2.81	23	1.60E+07	203	8	246	0.83
	38	809901	3075	263	11.69	24	1.94E+07	673	12	208	1.03
	39	813667	1690	481	3.51	24	1.95E+07	374	12	157	0.97
	40	832237	2242	371	6.04	24	2.00E+07	383	10	103	0.16
	41	886265	1807	490	3.69	24	2.13E+07	412	13	227	0.64
	42	922853	1845	500	3.69	25	2.31E+07	416	13	20	0.46
	43	1066114	2366	451	5.25	26	2.77E+07	632	15	83	0.49
	44	1195537	2749	435	6.32	27	3.23E+07	542	11	81	0.35
	45	1245254	1983	628	3.16	27	3.36E+07	419	12	247	0.56
(Bc)	46	1412873	2248	547	4.11	27	3.81E+07	540	13	204	1.57
	47	1827677	3096	590	5.25	30	5.48E+07	580	11	65	0.43
	48	2674113	2640	1013	2.61	34	9.09E+07	518	11	90	0.36
	49	3107283	4850	641	7.57	*40	1.24E+08	962	11	240	1.20
	50	7843219	5480	1431	3.83	46	3.61E+08	1187	12	106	0.34

TABLE A.5. Landslide inventory. No. corresponds to the cumulative number of active landslides. Abbreviations correspond to landslide abbreviations used in Figure 2.1. Planform area (m²) of landslide measured using DEM. Length (m) of landslide calculated as the distance from the top of the headscarp to the bottom of the toe. Average width (m) of landslide calculated as Area/Length. Aspect ratio (m/m) of calculated as Length/Average Width. * correspond to estimated thickness measured using lidar and field observations. All other thickness values are calculated using scaling relationship in Handwerger et al. (2013). Change in elevation (m) measured from source to toe of landslide. Landslide aspect is the average azimuth (°) of landslide motion in the downslope direction.

APPENDIX B

SUPPLEMENTARY MATERIAL FOR CHAPTER III

B.1. Figures

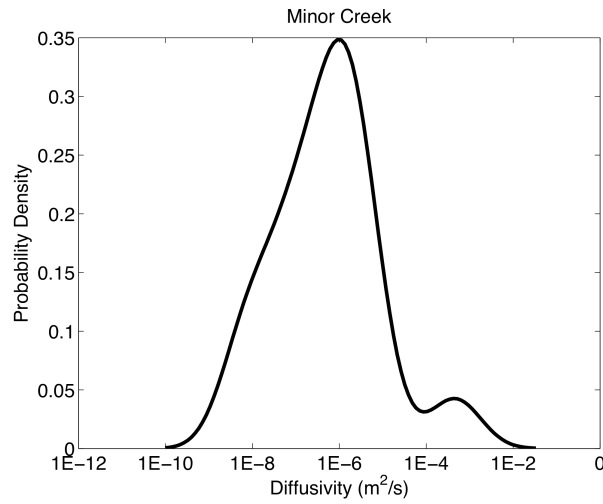


FIGURE B.1. Probability density function of hydraulic diffusivity for the Minor Creek landslide. Data from Iverson and Major (1987).

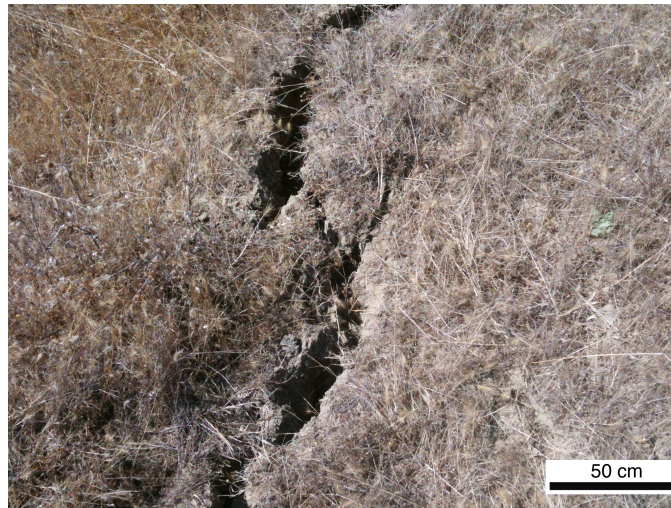


FIGURE B.2. Photograph of large open-mode surface crack at the Kekawaka 2 landslide.

B.2. Tables

No.	Master	Slave	Perpendicular baseline (m)	Days	Track No.
1	20070314	20070429	499	46	P223
2	20070314	20070614	-258	91	P223
3	20070314	20070730	474	138	P223
4	20070314	20070922	663	182	P223
5	20070516	20070701	175	46	P224
6	20070516	20070816	461	91	P224
7	20070516	20071001	720	137	P224
8	20070516	20071116	1171	183	P224
9	20070516	20080101	1054	229	P224
10	20070614	20070807	-129	46	P223
11	20070614	20070914	684	91	P223
12	20070701	20071116	996	137	P224
13	20070701	20080101	879	183	P224
14	20070816	20071001	258	46	P224
15	20070816	20071116	709	91	P224
16	20070816	20080101	592	138	P224
17	20071001	20071116	451	46	P224
18	20071001	20080101	334	91	P224
19	20071116	20080101	-117	46	P224
20	20071116	20080216	967	91	P224
21	20080101	20080216	1084	46	P224
Data Gap					
22	20080703	20081118	-1430	136	P224
23	20080703	20090405	-193	275	P224
24	20081118	20090103	-117	46	P224
25	20081101	20090201	694	91	P223
26	20081101	20090319	1194	140	P223
27	20081101	20090812	1119	276	P223
28	20090103	20090405	1023	93	P224
29	20090201	20090319	499	48	P223
30	20090201	20090619	846	140	P223
31	20090201	20090804	425	186	P223
32	20090319	20090619	347	91	P223
33	20090319	20090804	-74	137	P223
34	20090405	20091121	1026	229	P224
35	20090405	20100106	1264	275	P224
36	20090619	20090804	-422	46	P223
37	20090821	20091121	680	91	P224
38	20090821	20100106	918	137	P224
39	20091121	20100106	238	46	P224
40	20091121	20100221	936	91	P224
41	20091121	20100408	1105	139	P224
42	20100106	20100221	698	46	P224
43	20091220	20100322	1006	93	P223
44	20091220	20100507	1153	139	P223
45	20100221	20100408	169	48	P224
46	20100221	20100524	210	94	P224
47	20100408	20100524	40	46	P224
48	20100408	20101124	925	228	P224
49	20100524	20101124	884	182	P224
50	20100524	20110109	1327	228	P224
51	20101124	20110109	443	46	P224

TABLE B.1. ALOS PALSAR data used in time series inversion. No. corresponds to the cumulative number of interferogram pairs. Master and Slave correspond to the acquisition date (yyyymmdd) of each scene. Spatial baseline is the perpendicular baseline (m) between the SAR scene acquisitions and temporal baseline is the time period between acquisitions. Path number corresponds to satellite path P223/4. * correspond to scenes with modified dates.

APPENDIX C

SUPPLEMENTARY MATERIAL FOR CHAPTER IV

C.1. Model Details

Fig. C.7 displays a schematic diagram of the model domain, which parallels well-established rate and state descriptions for the behavior of tectonic faults (e.g., Rubin, 2008; Skarbek et al., 2012). We focus on how perturbations to the momentum balance cause variations in the slip velocity v . Accordingly, equation (4.2) in the main text is used to capture the empirically based logarithmic dependence of the friction coefficient μ on slip velocity v (Marone, 1998); this necessitates that modeled sliding velocities remain finite at all times. The sliding surface is forced to slip at a small rate v_{upper} for $x < 0$ and at v_{lower} for $x < L$, where we set $v_{lower} = 2v_{upper}$ for all calculations shown here (Table C.1). These boundary conditions can be interpreted to represent steady long-term motion induced by loading due to headscarp slumping at the upper boundary and toe erosion at the lower boundary. The slip-rate gradient is chosen to simulate velocity variations along the sliding surface that are commonly observed in real landslides (e.g., Iverson, 1986c; Malet et al., 2002; Coe et al., 2003; Handwerger et al., 2015). We recognize that many landslides display more complex slip gradients, but neglect such complications here in order to focus upon only the most essential underlying controls. For each model simulation, the sliding surface has constant gravitational driving stress, background effective stress, and frictional properties within the interval $0 < x < L$.

Two main state evolution laws are used in rate and state friction models. The first, called the Aging (or Dieterich) law, is defined by

$$\frac{d\theta}{dt} = 1 - \frac{v\theta}{d_c}, \quad (\text{C.1})$$

which exhibits healing (i.e. causes μ to increase) during stationary contact. The second, called the Slip (or Ruina) Law, is defined by

$$\frac{d\theta}{dt} = -\frac{v\theta}{d_c} \ln\left(\frac{v\theta}{d_c}\right), \quad (\text{C.2})$$

which exhibits no healing during stationary contact. When sliding conditions are near steady state (i.e. $v\theta/d_c \approx 1$), these laws have the same asymptotic behavior and yield predictions with only very slight differences (Ampuero and Rubin, 2008). However, when conditions are far from steady state the predictions of these laws begin to differ significantly. Alternative treatments have been designed to account specifically for variations in frictional strength due to changes in effective stress, although relatively few laboratory studies have explored this behavior (Linker and Dieterich, 1992; Dieterich and Linker, 1992). To account for the condition of variable effective normal stress σ' , Linker and Dieterich (1992) propose that

$$\frac{d\theta}{dt} = 1 - \frac{v\theta}{d_c} - \frac{\lambda\theta}{b\sigma'} \frac{d\sigma'}{dt}, \quad (\text{C.3})$$

where the constant $\lambda \approx \mu_0/3$ (Perfettini, 2000).

Since our idealized model is primarily designed to capture the broad qualitative patterns of landslide behavior, each of the Aging, Slip, and variable effective stress (LD) evolution laws are acceptable for our purposes. To demonstrate that our results

are not biased by the choice of state evolution law, we performed several model runs using each law. As expected, we find only minor differences in predicted displacement patterns between each set of simulations (Fig. C.8). For computational efficiency, we adopt the Aging law for all calculations shown in the main paper.

We use a grid spacing $dx = L_b/10$, which provides sufficient resolution for the Aging law (Rubin, 2008), where

$$L_b = \frac{Gd_c}{\sigma' b} = h^* \left[1 - \left(\frac{a}{b} \right)_{rw} \right] \quad (\text{C.4})$$

is the nucleation length scale for perturbation-induced slip transients (Dieterich, 1992; Perfettini and Ampuero, 2008). We solve equations (4.2), (4.3), and (4.4) as coupled first order differential equations in v and θ using ODE solver routines in MATLAB. The elastic stresses are calculated in the Fourier domain where the transformed quantify satisfies

$$\mathcal{F} \left(\frac{d\tau_{el}}{dt} \right) = - \frac{G}{2(1-\nu)} |n| \mathcal{F}(v), \quad (\text{C.5})$$

where n is the wavenumber. The forced-slip section is set to a total length of $16L$ to prevent numerical error in the Fourier domain. Model simulations are run to steady state sliding before being subjected to stress perturbations. This ensures that the initial conditions have minimal influence on the model results.

C.2. Parameter Values

Numerous laboratory investigations have calibrated rate and state friction parameters for a wide range of slip rates, stresses, and materials (see refs in Marone, 1998). However, there are no published rate and state friction parameters from laboratory experiments on landslide material. Because the slip surfaces of landslides

and faults are both governed by frictional sliding (Gomberg et al., 1995), we select a nominal range of rate and state parameter values from the fault mechanics literature (Table C.1) (Linker and Dieterich, 1992; Dieterich, 1979; Saffer and Marone, 2003; Ikari et al., 2009; Niemeijer and Vissers, 2014). Since our focus here is understanding the overall patterns of landslide motion, we consider small errors introduced by uncertainty in parameter values to be acceptable. We note that Lacroix et al. (2014) back-calculated a and b using a slider block model to describe field-based measurements of a slow-moving landslide perturbed by a nearby earthquake. They found highly unusual values that are 13 orders of magnitude smaller than are typically used. It should be noted that an error in their calculations implies that the values of a and b are actually ten orders of magnitude lower than the values they published (personal comm. with Hugo Perfettini). We view these unusual parameter values with suspicion and suggest that they may be in error in part as a result of the model formulation, which did not account for variations in effective stress or for transient elastic coupling along the slip surface.

Several model parameters differ between our choices for the slow-moving landslide and catastrophic landslide calculations (Table C.1). Parameters were primarily chosen so that the models produce displacement patterns that have been observed in field and remote sensing studies of landslides (Handwerger et al., 2015; Saito, 1969; Iverson and Major, 1987; Reid et al., 2003; Helmstetter et al., 2004). For the slow-moving landslide runs, we use a larger d_c that is consistent with field-based estimates of friction parameters on faults (Marone, 1998). We select μ_0 that characterizes weak, clay-rich materials (Saffer and Marone, 2003; Ikari et al., 2009; Skempton, 1985; Tika et al., 1996) that are common to slow-moving landslides with well developed slip surfaces (Wang et al., 2010). The shear modulus and shear wave speed are based

on values measured for the Jurassic and Cretaceous Franciscan Complex, Northern California (Brocher, 2005), a region well known for slow-moving landslide activity (Handwerger et al., 2015; Iverson and Major, 1987; Kelsey, 1978; Mackey and Roering, 2011), and are notably lower than values reported for deep faults. We chose the length, effective stress, and forced-slip boundary conditions to approximate values from slow-moving landslides in Northern California (Handwerger et al., 2015; Iverson and Major, 1987; Reid et al., 2003).

For the catastrophic landslide calculations, we selected parameters that predict creep-to-failure motion similar to field observations (Saito, 1969; Helmstetter et al., 2004). This required higher effective stress (i.e. smaller h^*), lower d_c , and significantly lower forced-slip boundary conditions such that the landslide is effectively not moving until perturbed (Fig. 4.3, Table C.1). We use a μ_0 that characterizes rock and soil materials (Linker and Dieterich, 1992; Dieterich, 1979; Saffer and Marone, 2003; Skempton, 1985; Tika et al., 1996). The shear modulus and shear wave speed are based on values measured in glacial tills that are prone to catastrophic landslides (Allstadt et al., 2013), e.g., the March 2014 Oso landslide, Washington, USA, that killed 43 people (Iverson et al., 2015). Model simulations shown in Fig. 4.3 were run at a nominal scaled slip surface length $L/h^* = 2$. However, we performed model simulations with $2 \leq L/h^* \leq 8$ because the ratio L/h^* has been found to control the failure mode of faults (Rubin, 2008; Skarbek et al., 2012; Liu and Rice, 2007). Although dynamic motion occurred for $L/h^* \geq 2$, we did confirm an apparent stress perturbation size threshold that decreases with increasing L/h^* (Fig. C.6, see main text for details).

C.3. Elastic Stresses during Landslides

Significant insight has been gained into earthquake mechanics by modeling displacement along one-dimensional faults (i.e. continuous array of dislocations) with two-dimensional elasticity (Rubin and Ampuero, 2005; Segall, 2010). These models combine well-constrained laboratory measurements of friction with dislocation theory and fracture mechanics to quantify stress along a slip surface. However, elastic treatments are not commonly used in landslide models, where plastic deformation is the primary mechanism that characterizes landslide motion.

Our model approximates a landslide as a one-dimensional slip surface embedded in a homogeneous elastic medium. While it is generally accepted that the weathered rock and soil that overlies the landslide slip surface does not behave elastically over long time-scales or for large displacements, elastic deformation can be significant for small displacement and strain and over short timescales (i.e. transient motion). Furthermore, several elastic dislocation models have been used to quantify key features of landslides, including landslide depth (Aryal et al., 2015) and incipient motion (Martel, 2004). Since our model simulations describe small displacements, incorporation and use of elastic effects is reasonable.

We account for elastic stress-transmission due to nonuniform slip along the sliding surface. Fig. C.9 shows that the elastic stress is relatively small at all time steps and only becomes large leading up to runaway acceleration. For model simulations where $L/h^* < 1$, the elastic stress decreases during periods of transient slip to account for increases in velocity along the slip surface (i.e. non-local accelerations). Thus, the elastic stress is able to compensate for rate-weakening friction and prevent runaway instability. However, when $L/h^* > 1$ the elastic stress increases the total driving stress until it eventually exceeds the frictional strength and causes runaway instability. Our

findings suggest that elastic stress-transmission provides an important control on landslide failure mode.

C.4. Free Surface

The landslides we focus on are deep in comparison with the modeled displacement, but shallow in comparison with their along-strike lengths. We did not account for variations in shear and effective stress that arise in a more rigorous elastic treatment of dislocation motion near a free surface. Several studies have incorporated the free surface into their model simulations in other contexts and have found that these effects produce fairly minor differences in the overall model predictions (e.g., Rubin, 2008; Viesca and Rice, 2012). Similar to the results of these studies, we expect that our neglect of the free surface likely results in slight model inaccuracies that are of secondary importance to the overall displacement patterns predicted by our model.

C.5. Tables

Symbol	Definition	Value (slow; catastrophic)	Source
a_{rw}	friction parameter, []	0.008	Saffer and Marone (2003), Ikari et al. (2009), Niemeijer and Vissers (2014)
a_{rs}	friction parameter, []	0.012	Saffer and Marone (2003), Ikari et al. (2009), Niemeijer and Vissers (2014)
a_{rn}	friction parameter, []	0.01	Saffer and Marone (2003), Ikari et al. (2009), Niemeijer and Vissers (2014)
b	friction parameter, []	0.01	Saffer and Marone (2003), Ikari et al. (2009), Niemeijer and Vissers (2014)
d_c	characteristic slip distance, [m]	4×10^{-2} ; 4×10^{-4}	Power and Tullis (1992), Marone (1998)
μ_0	reference friction, [m]	0.45 ; 0.6	Tika et al. (1996), Saffer and Marone (2003)
v_s	shear wave velocity, [m s ⁻¹]	250 ; 600	Brocher (2005); Allstadt et al. (2013)
ν	Poisson's ratio, []	0.35	Lambe and Whitman (1969), Morris et al.(1992)
G	Shear modulus, [Pa]	4.87×10^6 ; 1.08 – 4.32×10^8 ;	Brocher (2005); Holzer et al. (2005); Allstadt et al. (2013)
v_{upper}	forced slip speed, [m s ⁻¹]	5×10^{-9} ; 5×10^{-12}	—
v_{lower}	forced slip speed, [m s ⁻¹]	1×10^{-8} ; 1×10^{-11}	—
L	Slip surface length, [m]	1000; 500	—
L/h^*	Scaled slip surface length, []	0.1 ; 2–8	—
h^*	nucleation distance, [m]	2500 ; 250	—
σ'	background effective stress, [Pa]	6×10^4 ; 5.3×10^5	—

TABLE C.1. Parameter values used in model simulations. Values on the left correspond to the slow-moving landslide models. Subscripts rw, rs, and rn correspond to rate-weakening, rate-strengthening, and rate-neutral.

C.6. Figures

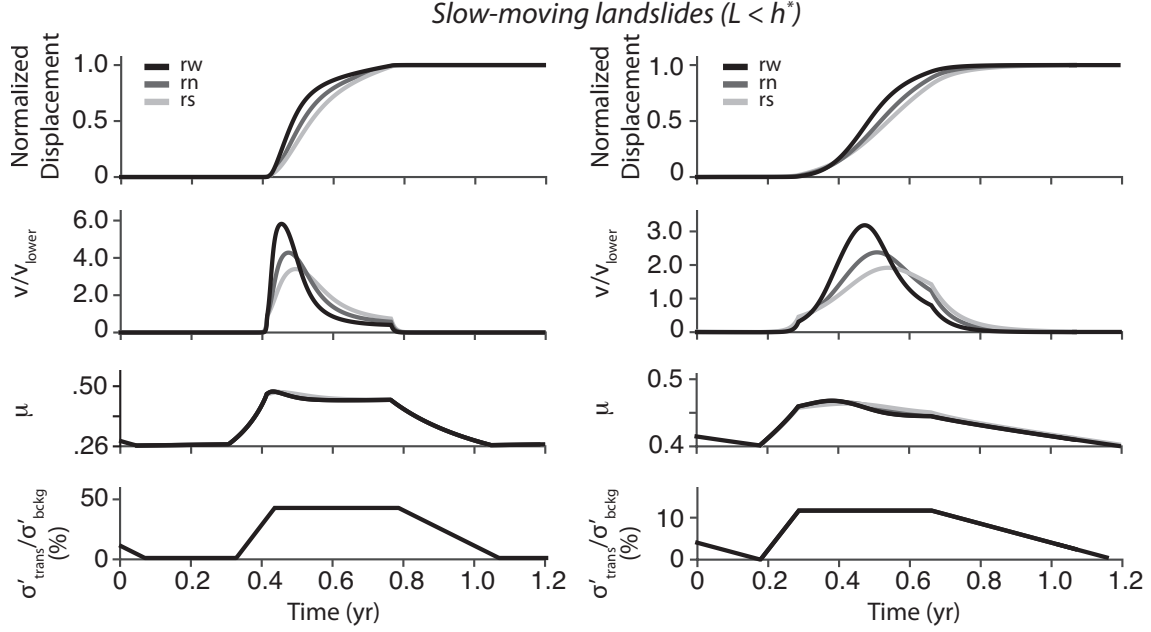


FIGURE C.1. Material controls on modeled landslide behavior. rw, rs, and rn correspond to rate-weakening, rate-strengthening, and rate-neutral materials, respectively. Model simulations performed with rate-weakening material have a nominal $L/h^* = 0.1$. Note only minor differences in behavior occur as a result of the material controls.

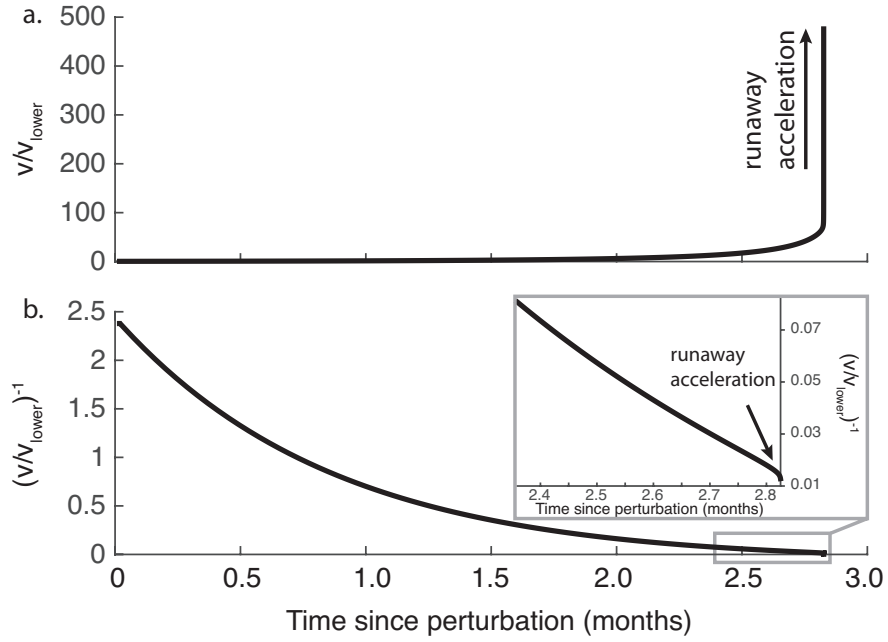


FIGURE C.2. Creep-to-failure motion for catastrophic landslides. Model simulations performed with rate-weakening material and $L/h^* = 2$. a. Normalized velocity time series shows slow acceleration for several months before a transition to rapid acceleration and catastrophic failure. b. Normalized inverse velocity time series plotted to make comparisons with field measurements of creep-to-failure behavior (e.g., Saito, 1969; Helmstetter et al., 2004).

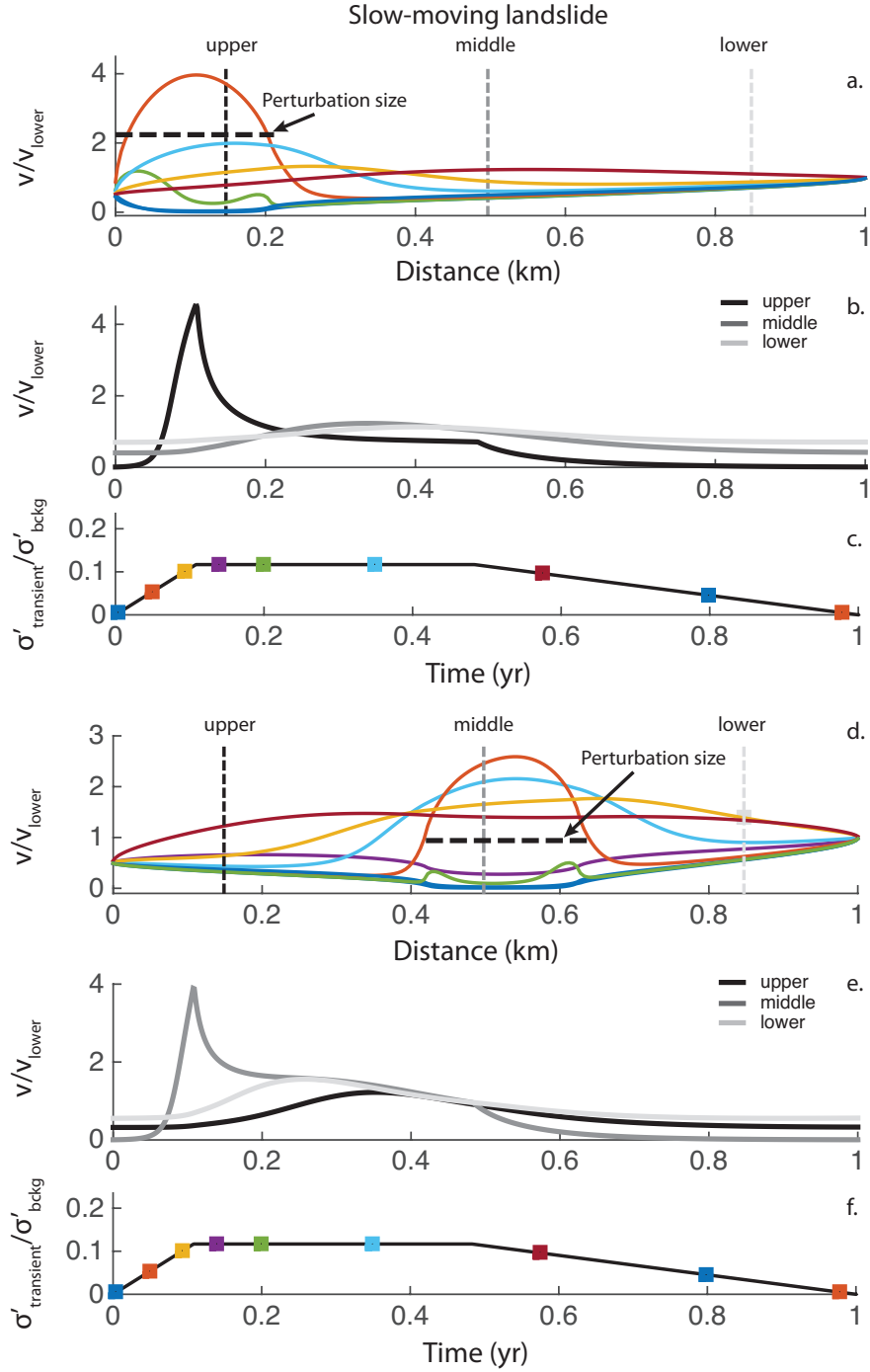


FIGURE C.3. Spatial evolution of modeled velocity in response to local stress perturbations (i.e. stresses are only perturbed on a small section of the sliding surface) for slow-moving landslides. Model simulations performed with rate-weakening material and $L/h^* = 0.1$. Velocity changes at these locations drive motion along the entire landslide due to elastic stress-transmission. Vertical dotted lines indicate location of velocity time series plotted in b. and e. Colored dots in c. and f. correspond to colored lines in a. and d.

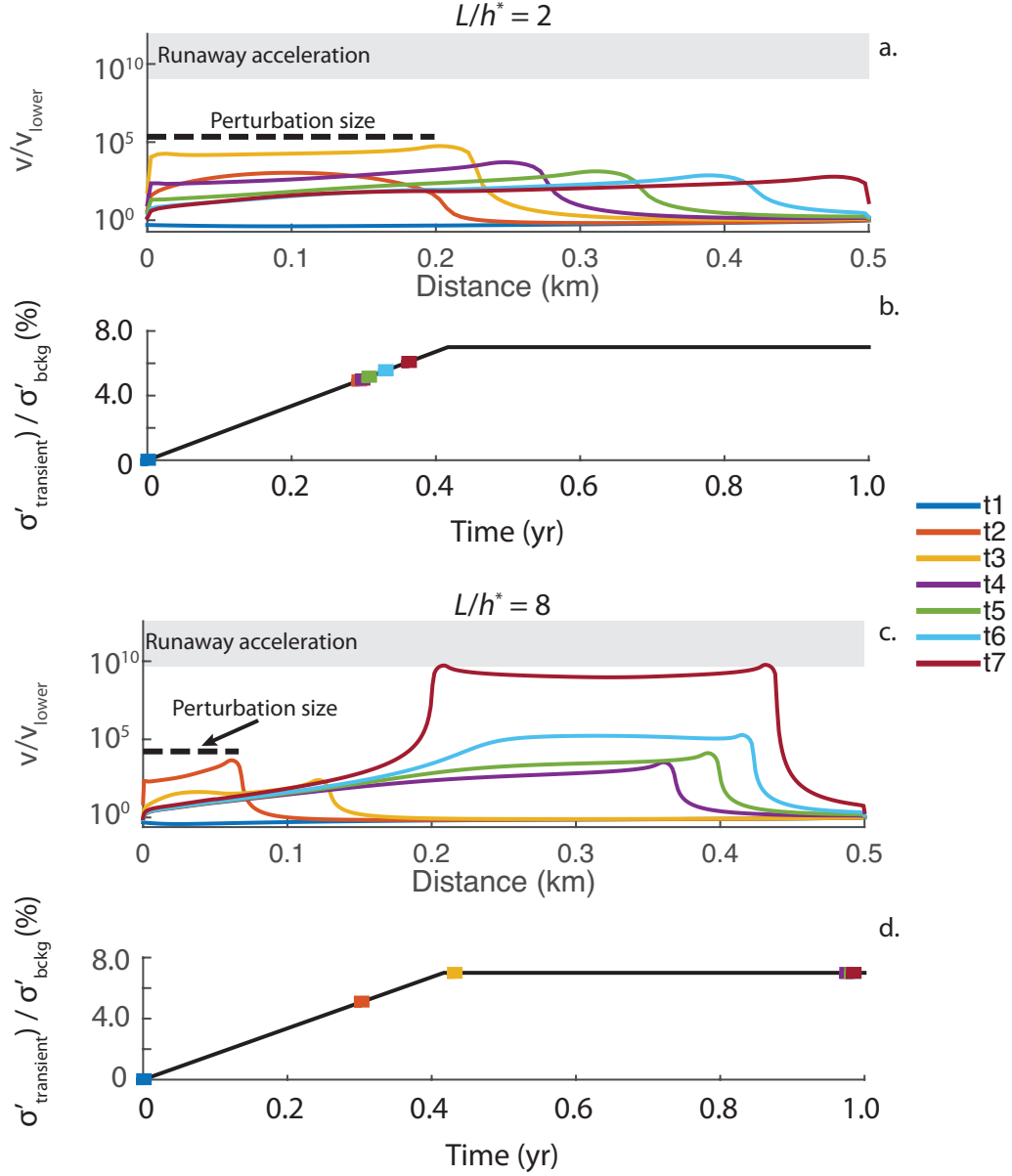


FIGURE C.4. Spatial evolution of modeled velocity in response to local stress perturbations for catastrophic landslides. For $L/h^* = 2$ there is an apparent perturbation size threshold that must be exceeded to trigger runaway acceleration. Note the perturbation size is significantly smaller for $L/h^* = 8$, but still triggers runaway acceleration. a. Velocity response when perturbation is below threshold value. c. Velocity response when perturbation is above threshold value.

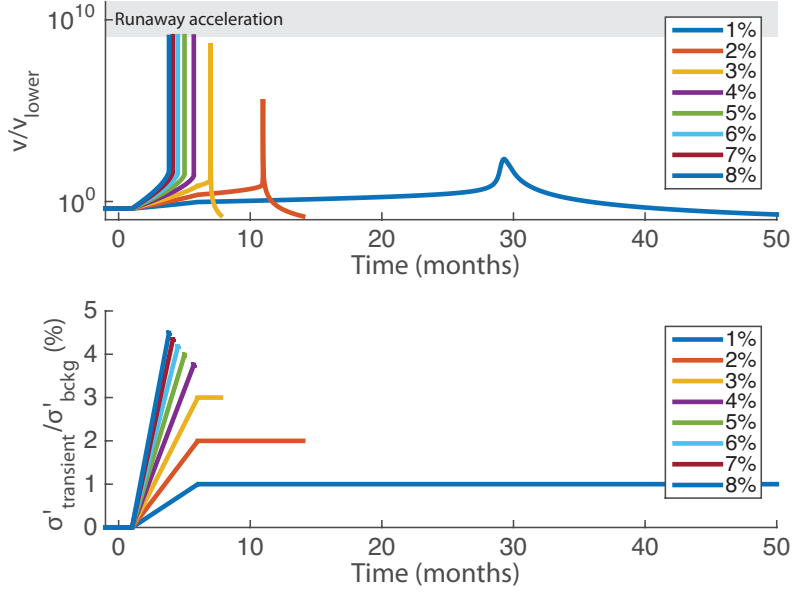


FIGURE C.5. Normalized velocity time series as function of perturbation magnitude for $L/h^* = 2$. Transient stress perturbations must exceed 4% of the background effective stress to trigger failure under these model conditions. Colored lines correspond to $\max(\sigma'_{\text{trans}}/\sigma'_{\text{bckg}})$.

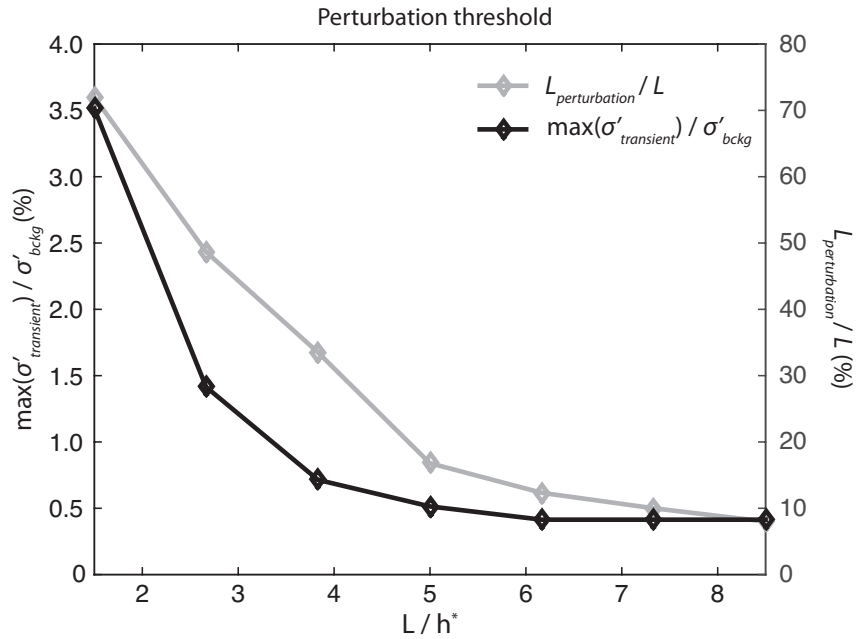


FIGURE C.6. Stress perturbation threshold as a function of L/h^* . Values correspond to the minimum perturbation size required to trigger catastrophic failure. $L_{\text{perturbation}}$ is the spatial dimension of the perturbation.

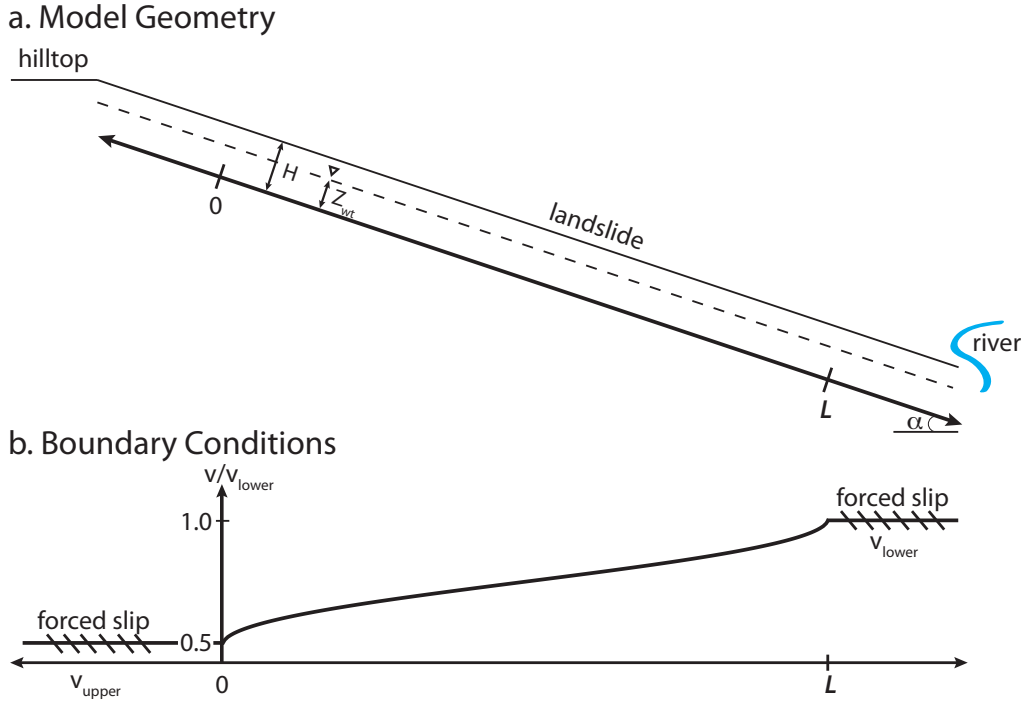


FIGURE C.7. a. Model geometry for a simplified two-dimensional landslide. The free surface is not included in the model simulations. b. Boundary conditions along the landslide.

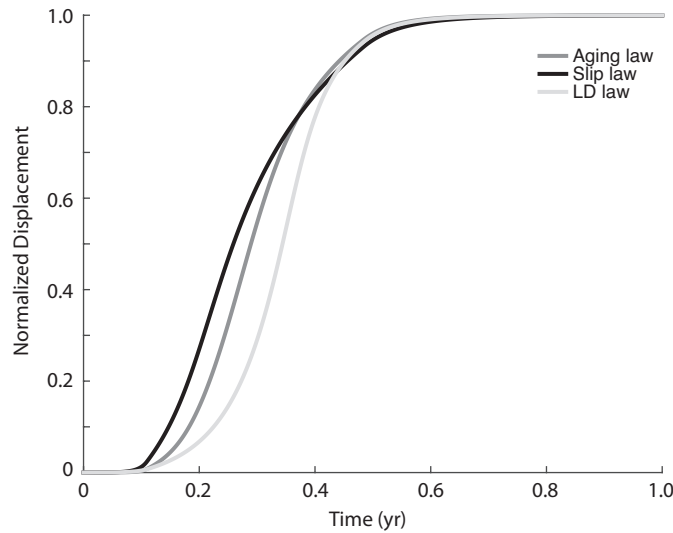


FIGURE C.8. Normalized displacement for the Aging, State, and LD evolution laws. Note there are minor differences and the timing of motion, however these are secondary to the overall displacement patterns.

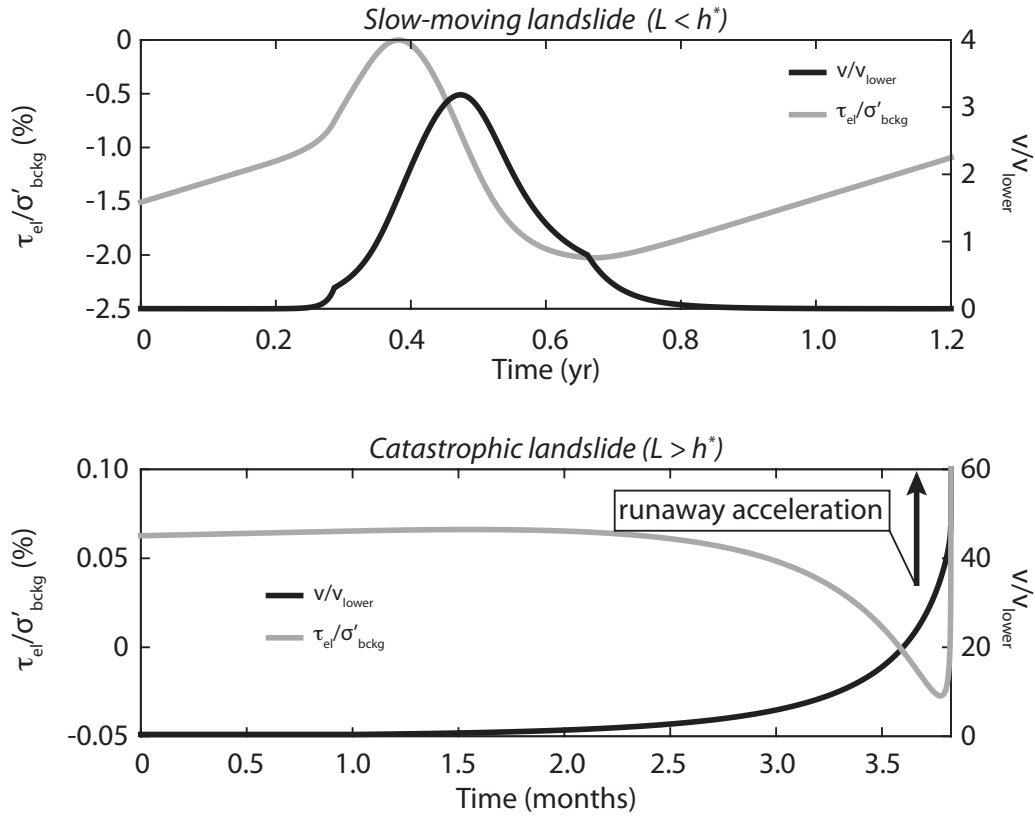


FIGURE C.9. Normalized elastic stress and normalized velocity for slow and catastrophic landslide model simulations. For the slow-moving landslide calculations, the elastic stress is small at all time steps and increases during transient slip periods to balance the resisting stress. For the catastrophic landslide calculations, the elastic stress is negligible until runaway acceleration at which point the resisting stress cannot balance the elastic stress.

APPENDIX D

ENVIRONMENTAL CHANGE, HYDRATE DISSOCIATION, AND SUBMARINE SLOPE FAILURE ALONG CONTINENTAL MARGINS: THE ROLE OF SATURATION ANOMALIES IN LANDSLIDE TRIGGERING

This work is co-authored with Alan Rempel and is to be published online in the *Proceedings of the 8th International Conference on Gas Hydrates*, (ICGH8-2014), Beijing, China, 28 July–1 August, 2014. The proceedings are not peer-reviewed.

D.1. Nomenclature

Symbol	Definition
c	Athys constant []
C	Cohesion [Pa]
C'	Effective cohesion [Pa]
FS	Factor of Safety []
g	Gravity [$\text{m}^2 \text{s}^{-1}$]
G	Shear modulus [Pa]
k	Permeability [m^2]
n	Porosity []
P_{tot}	Pore fluid pressure [Pa]
P_{hydro}	Hydrostatic pore fluid pressure [Pa]
P_{ex}	Excess pore fluid pressure [Pa]
S_h	Hydrate saturation []
t	time [s]
u	Darcy velocity [m s^{-1}]
W	Lamberts function []
x	Lateral dimension of hydrate [m]
z	Depth [m]
Δ	Change in
ϕ	Friction angle [deg]
ϕ'	Effective friction angle [deg]
δ	Displacement for rate and state [m]
ξ	Lateral dimension of hydrate [m]
ρ	Density [kg m^{-3}]
σ'	Effective stress [Pa]
σ'_h	Stress supported by hydrate [Pa]

D.2. Introduction

Submarine landslides are a natural part of the marine sedimentary system and are ubiquitous on both active and passive continental margins (Locat and Lee, 2002; Chaytor et al., 2009; Masson et al., 2010). Clear evidence is recorded in coastal sediments worldwide that past changes in environmental conditions have caused hydrates to become unstable and potentially trigger submarine landslides (Sultan et al., 2004). The direct hazards posed by such events can generate tsunamis, damage infrastructure (e.g., oil platforms, telecommunication lines), and if they occur within or beneath the methane hydrate stability zone (MHSZ) cause catastrophic release of methane to the ocean and atmosphere (Nisbet and Piper, 1998).

Numerous factors can lead to slope failure by either increasing the shear stress or decreasing the shear strength typically by reducing the effective stress, but also potentially by reducing cohesion or the effective coefficient of friction. Factors that contribute to slope failure are often separated into two categories; 1) preconditioning factors and 2) triggers (Masson et al., 2010; Mosher, 2009; Mosher et al., 2012). Sediments may be preconditioned for failure from their inherent characteristics (e.g., porosity, pore pressure, permeability structure) and stress histories. There are a variety of potential landslide triggers but the best recognized are earthquakes, tectonic uplift, rapid sedimentation, storm waves, anthropogenic influence, and potentially the dissociation of methane hydrates (Masson et al., 2010; Mosher, 2009; Mosher et al., 2012; Grozic, 2010; Owen et al., 2007).

A prerequisite for slope failure is the development and maintenance of elevated pore pressures. Several past studies have quantified excess pore pressures generated from methane hydrate dissociation (Kwon et al., 2008; Lee et al., 2010; Xu and Germanovich, 2006) and a number of authors have gone further and applied standard

slope stability analysis techniques (Sultan et al., 2004; Kwon and Cho, 2012; Nixon and Grozic, 2007). Nixon and Grozic (2007) used a numerical model to calculate pore pressures generated by hydrate dissociation and combined their results with the infinite slope stability criterion. Their results indicate significant reductions in slope stability that can trigger failure under many conditions. Although these model calculations are instructive and point to the importance of hydrate reservoirs for slope stability issues, they are limited in predictive capability because they follow from several limiting assumptions. In particular, they assume that 1) dissociation is complete and instantaneous, 2) hydrates are located at a uniform depth along the failure plane, 3) the failure plane is at the base of the MHSZ, 4) the soil parameters are uniform along the slope dip, and 5) the friction coefficient is independent of rate and state effects.

Xu and Germanovich (2006) quantified excess pore pressures in marine sediments caused by volume expansion during methane hydrate dissociation. To initiate dissociation, they assumed changes in environmental conditions such as sea level drop, tectonic uplift, and heating of hydrate-bearing sediments. Using their theoretical approach, they found that in confined sediments hydrate dissociation generates several tens of megapascals of excess pressure, while in well-connected sediments excess pore pressure development is lower, but still significant.

Sultan et al. (2004) developed a model to simulate the distribution of hydrates in sediment while taking into account the pressure, temperature, pore-water chemistry, and aspects of the pore-size distribution in the MHSZ. Their model quantifies the excess pore pressure during hydrate dissociation and, unlike Nixon and Grozic (2007), does not assume dissociation and slope failure necessarily occur at the bottom of the MHSZ. They use a limit-equilibrium slope-stability analysis combined with empirical

data from the Storegga Slide, offshore Norway — one of the largest submarine landslides ever identified. Although a number of alternative mechanisms have been suggested (earthquakes, glacial sedimentation), they conclude that the dissociation of hydrates was the likely cause of the Storegga event.

Recently, Viesca and Rice (2010) and Viesca and Rice (2012) introduced models for submarine slope behavior that treat slide nucleation as a frictional instability that operates according to the same physical principles as identified in laboratory rock-friction experiments that are designed to probe the nucleation of earthquakes (e.g., Marone, 1998). In either geologic setting, whether along a fault or a landslide surface, for nucleation to be possible the frictional resistance must decrease as the rate of sliding (or the slip distance) increases — behavior commonly referred to as rate-weakening (or slip-weakening). With rate-weakening friction, infinitesimal perturbations to the slip rate grow to the point where inertial effects are important (i.e. landslides) only if the slipping patch exceeds a finite and predictable size that depends on the frictional properties and effective stress. Once rapid sliding does take place, significant heat can be generated, with further feedbacks on the resisting strength as a result of the pressurization and generation of new pore fluids (Brantut et al., 2011; Rempel and Rice, 2006).

Here, we outline a 1D numerical model that we are developing to calculate hydrate dissociation and changes in effective stress within the MHSZ. We perform slope stability analyses using the infinite-slope approximation as a prelude to development of a rate and state friction approximation. We use parameters consistent with those measured and inferred from contemporary natural hydrate reservoirs (e.g. Hydrate Ridge, offshore Oregon, USA). First, we generate a heterogeneous distribution (i.e. hydrate anomalies) of gas hydrates using a model from Rempel

(2011). Rempel (2011) model predicts high-concentration hydrate anomalies that occur along stratigraphic boundaries due to changes in solubility. We induce hydrate dissociation by perturbing the equilibrium conditions via changes in the temperature and/or the pressure at the seafloor. The dissociation of high-concentration hydrate anomalies at fine to coarse-grained stratigraphic boundaries are of particular interest because they involve abrupt changes in saturation level over short distances where segregated, lens-like deposits and nodules push the sediment particles apart in order to grow, and also because they are associated with large changes in sediment permeability and strength. Our numerical models track the evolution of the pore-water pressure as solid hydrate anomalies decay to lower the cohesion of sediments, decrease the friction angle, consolidate, and alter the average density of the fluids that they contain. We hypothesize that rapid consolidation of sediment will generate excess pore pressure that will significantly reduce material strength and that enhanced fluid flow resulting from changes in the average density of the pore fluids will also weaken sediment.

D.3. Modeling Stable Hydrates in a 1D Sediment Column

D.3.1. Hydrate Distribution

Hydrates typically occupy 1 to 50% of the pore space in marine sediments (Rempel, 2011; Trehu et al., 2004; Bohrmann and Torres, 2006). The distribution of hydrates is controlled by the supply of gas and the local phase equilibrium conditions, both of which are modulated by sediment properties such as permeability and pore size. Empirical data (i.e. acoustic imaging and chemical analysis) and numerical modeling show that the hydrate distribution within a sediment column is heterogeneous and has “spikes” (high concentrations) and “holes” (low concentrations)

at stratigraphic boundaries (Figure D.1) (Rempel, 2011; Trehu et al., 2004; Bohrmann and Torres, 2006). In order to properly quantify the effects of hydrate dissociation, the distribution of hydrates within the pore space must be known.

Within most of the hydrate stability zone, two-phase equilibrium is achieved between solid hydrate and an aqueous solution that contains dissolved gas. Where hydrate is present, the dissolved gas concentration is assumed equivalent to its solubility. The surface energy of the hydrate-liquid interface gives rise to “capillary effects” that result in an increased solubility where the interface is highly curved, as in the throats of small pores. The “wetting” properties of hydrate and mineral particles also enable thin aqueous films to separate these solids along pore walls. Because of these wetting and surface-energy interactions, the gas solubility is higher in more fine-grained sediments and it also increases in a given sediment as the hydrate saturation level increases (i.e. the fraction of the pore space that contains solid hydrate), which reduces the thickness of aqueous films on pore walls and increases the radii of curvature of hydrate crystal surfaces that are comparatively distant from mineral particles. At boundaries that separate sediments with different pore-size distributions, these microscale perturbations to the phase behavior enable hydrate to grow in the more coarse-grained fraction while the gas concentration is still lower than its equilibrium solubility in the adjacent fine-grained sediments. The net result is the development of large-amplitude, narrow (meter-scale or less) spikes in hydrate saturation level adjacent to hydrate-free holes. Gas transport supplies the growth of spikes by advection with the moving pore fluid and dispersive transport down the concentration gradient (Rempel, 2011).

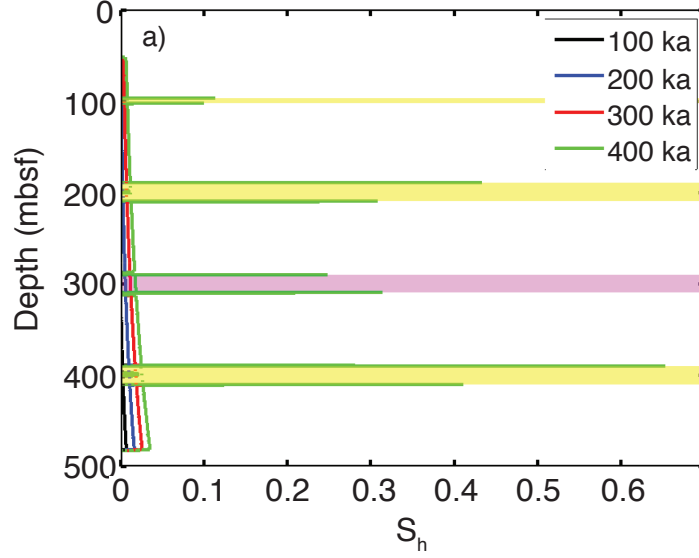


FIGURE D.1. Predicted hydrate saturation in a sediment column. Modified from Rempel (2011).

D.3.2. Effective Stress and Porosity

Once the hydrate spikes interconnect to distances greater than the characteristic pore size, the hydrate skeleton is able to transmit the overburden load, which reduces the local effective stress and prevents normal consolidation of the sediment (Rempel, 2011). For hydrate-bearing sediment the effective stress increases with depth according to

$$\frac{\partial \sigma'}{\partial z} = (1 - n)(\rho_s - \rho_f)g + (1 - n)\frac{\mu}{k}u + \frac{\partial \sigma_h}{\partial z}, \quad (\text{D.1})$$

where the first term on the right accounts for buoyancy, the second term accounts for the fluid flow (i.e. nonhydrostatic), and the third term accounts for the stress gradient transmitted between the hydrate and pore-matrix. The magnitude of the load transmitted by hydrate depends on the hydrate saturation, S_h , and the thermomolecular coefficient, Π , which is a measure of the hydrate-particle repulsive

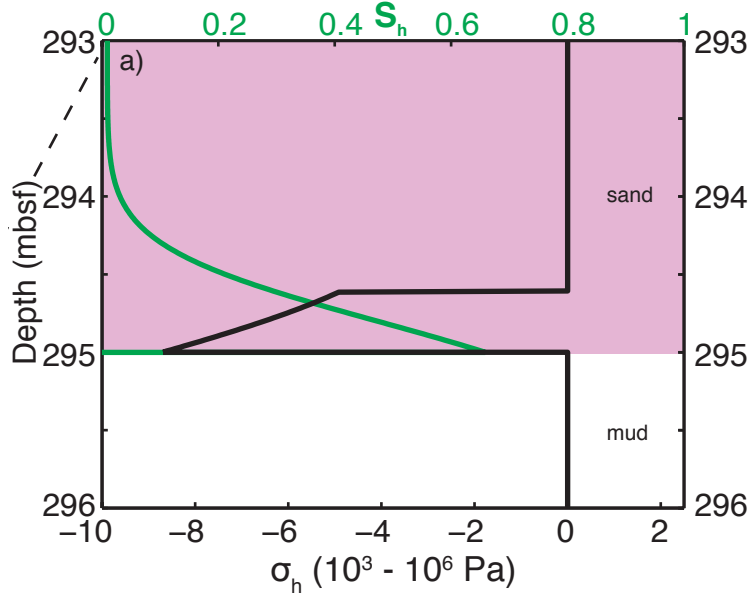


FIGURE D.2. Close up view of hydrate anomaly. Hydrate is able to transmit load and reduce effective stress. Pink shaded areas correspond to the location of hydrate anomalies in Figure D.1. Modified from Rempel (2011).

interactions (Rempel, 2011). Figure D.2 shows an example of the calculated reduction in effective stress due to a hydrate spike.

Next we define sediment porosity as a function of effective stress (Skarbek and Saffer, 2009)

$$n = -W \left\{ -\exp \left[\frac{-b\sigma'}{g(\rho_s - \rho_f)} \right] + \ln n_0 - n_0 \right\} \quad (\text{D.2})$$

Combining equations (D.1) and (D.2) we can predict evolution of the sediment porosity in a hydrate bearing sediment column. We solve for effective stress and porosity for a sediment column with and without hydrates.

D.3.3. Excess Pore Pressure

In our model, excess pore pressure (i.e. nonhydrostatic) is generated due to consolidation and from fluid flow driven by changes in the average pore fluid density as hydrates dissociate. We use consolidation parameterizations fit to empirical data

from consolidation tests (e.g. Tan et al. (2006)) and a 1D consolidation model. The excess pore pressure will evolve according to

$$\frac{\partial P_{ex}}{\partial t} = D_0 \frac{\partial^2 P_{ex}}{\partial Z^2} \quad (\text{D.3})$$

which is the linear pore pressure diffusion equation and the total pore pressure $P_{tot} = P_{hydro} + P_{ex}$. P_{ex} is driven by changes in σ' as hydrate spikes decay. We also account for changes in pore fluid density.

D.3.4. Infinite Slope Model for Slope Stability

We begin by calculating the slope stability in a 1D sediment column that contains hydrate anomalies using the infinite slope model. The infinite slope stability model is applicable to translational landslides with a planar shear surface that is much longer than it is deep, which can be a reasonable approximation for some submarine slopes. The analysis leads to

$$FS = \frac{C' + \sigma' \tan \phi'}{\tau}, \quad (\text{D.4})$$

where FS describes the ratio of shear strength to the shear stress. Because C' and ϕ' are functions of the hydrate distribution and effective stress, we use functional relationships (e.g. $C' = f(S_h, \sigma')$) to describe the evolution of these parameters as hydrates dissociate.

D.3.5. Rate and State Model for Slope Stability

We are continuing to develop our rate and state formulation and presently only outline how we will proceed in future work. Treating the bulk sediment as a linear-elastic material, the perturbation to the stress along an incipient rupture surface due

to displacement $\delta(\xi, t)$ between end-points $\xi = a \pm (t)$ (as set, for example, by lateral dimensions of hydrate anomalies) that are much closer together than the depth below the seafloor is (Viesca and Rice, 2012)

$$\Delta\tau(x, t) \approx -\frac{G}{2\pi(1-\nu)} \int_{-a_+(t)}^{+a_+(t)} \frac{d\delta/d\xi}{\xi - x} d\xi \quad (\text{D.5})$$

We are modifying the model developed in Chapter IV to include pore pressure changes that are predicted by our hydrate dissociation models and examine the necessary conditions for slope failure.

D.4. Results and Discussion

D.4.1. Effective Stress and Porosity

Figure D.3 shows the effective stress profile for hydrate-bearing sediments adjacent to hydrate-free sediments at a stratigraphic boundary. Calculations demonstrate that under typical conditions, the presence of interconnected hydrates lowers the effective stress by between 10^3 and 10^6 Pa, depending on model parameters (Rempel, 2011). This suggests that hydrates can support overburden ranging from a small fraction of the grain-to-grain contact to completely unloading the sediment.

Using equations (D.1) and (D.2) we find porosity increased up to 6% from normally consolidated sediments due to load-bearing hydrate anomalies (Figure D.4). This calculation assumes the largest reduction in σ' , which approaches a few megapascals. Deviations from the normal consolidation line (NCL) decrease with depth as the overburden increases.

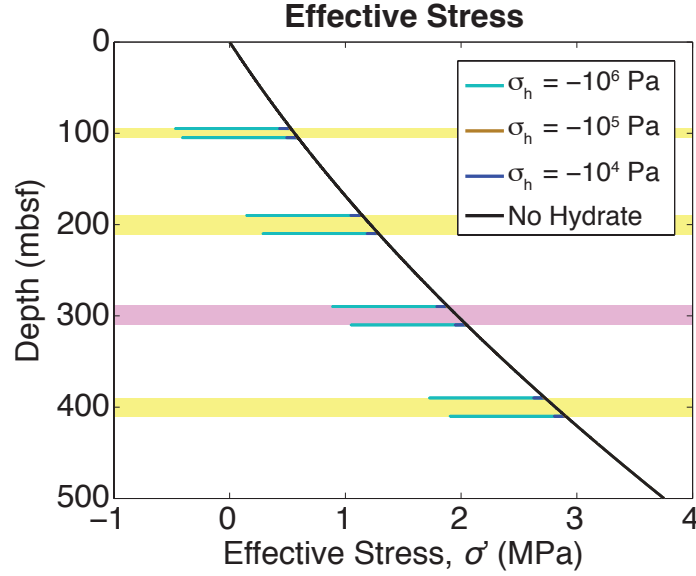


FIGURE D.3. Effective stress along a sediment column. We allow the reduction in effective stress from hydrate anomalies to range from 10^4 to 10^6 Pa. Colored lines correspond to the stress reductions resulting from the load-bearing hydrate. Black line corresponds to the NCL. Yellow and pink shaded areas correspond to the location of hydrate anomalies in Figure D.1.

D.4.2. Slope Stability

We quantify the slope stability for marine sediments with stable hydrates and those with dissociated hydrates. We have not yet fully coupled Rempel (2011) hydrate model with the slope stability models, so here we only describe the instantaneous dissociation of hydrates. Using the infinite slope model described by equations (D.4), we find that slopes are more stable if they contain stable hydrates (Figure D.5). The increase in sediment strength is due to the increased cohesion and friction angle of hydrate-bearing sediments (Waite et al., 2009) that counteracts the reductions in effective stress. However, once hydrates dissociate the cohesion and friction angle return to their intrinsic values and the sediment will begin to collapse back to the NCL.

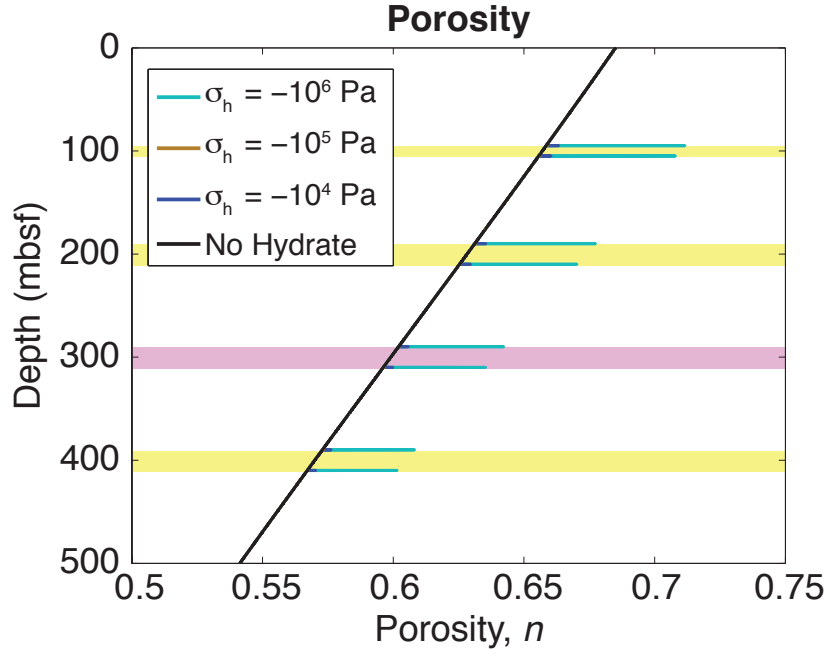


FIGURE D.4. Predicted porosity for an under-consolidated 1D sediment column. Colored lines correspond to the stress reductions resulting from the load-bearing hydrate. Black line corresponds to the NCL. Yellow and pink shaded areas correspond to the location of hydrate anomalies in Figure D.1.

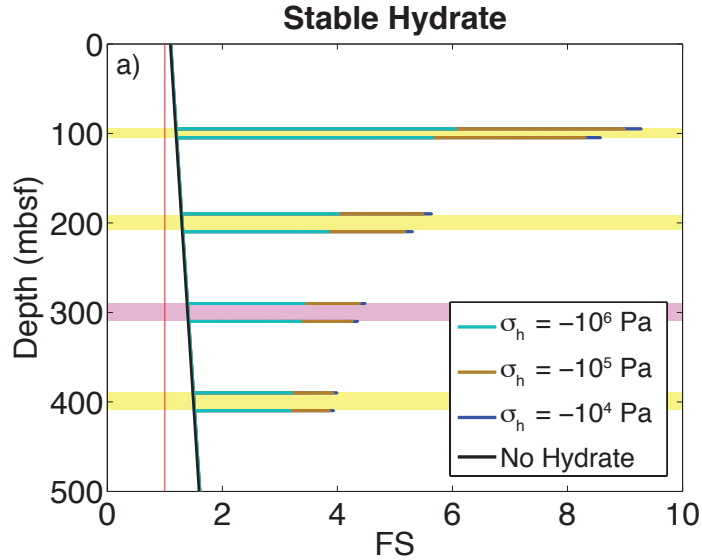


FIGURE D.5. Factor of safety in a hydrate-bearing sediment. Colored lines correspond to the stress reductions resulting from the load-bearing hydrate. Black line corresponds to NCL. Yellow and pink shaded areas correspond to location of hydrate anomalies in Figure D.1.

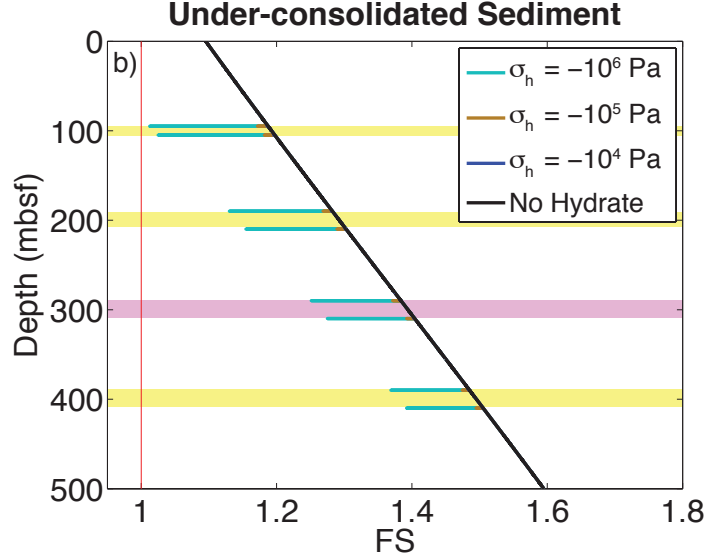


FIGURE D.6. Factor of safety in under-consolidated sediment that no longer contains hydrate. Colored lines correspond to the stress reductions resulting from the load-bearing hydrate. Black line corresponds to NCL. Yellow and pink shaded areas correspond to location of hydrate anomalies in Figure D.1.

In addition, we calculate the slope stability of the sediments assuming instantaneous dissociation of hydrate occurs before consolidation ensues (Figure D.6). Due to the increased porosity, the marine sediment is significantly weaker compared to both normally consolidated sediment and sediment with stable hydrate. However, we find that this is not enough to trigger failure and the generation of excess pore pressure is still required.

We rearrange equations (D.3) to calculate the P_{ex} required to initiate failure. We find that P_{ex} ranges from 10^3 to 10^6 Pa (Figure D.7).

In order to generate enough P_{ex} to trigger failure the dissociation must occur before the pressure can diffuse away from the location of the former hydrate anomaly. In other words, slope failure is expected when the timescale for consolidation following the loss of hydrate-supported pore rigidity is less than

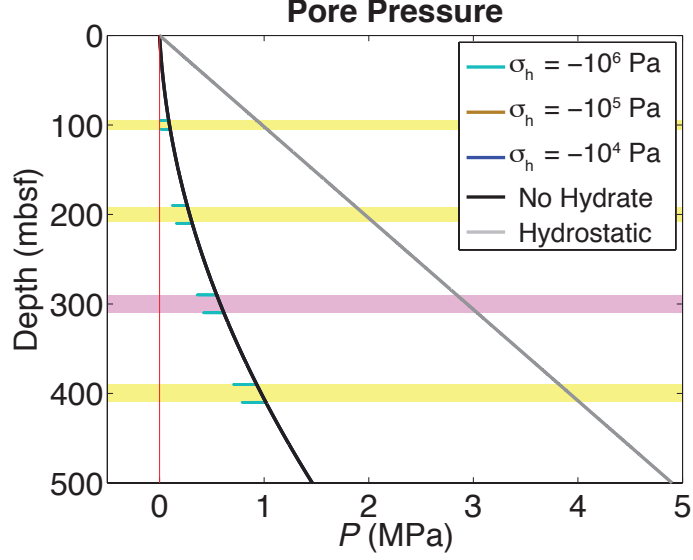


FIGURE D.7. Black and colored lines shows the excess pore pressure required to trigger failure. Gray line indicates hydrostatic pore pressure. Yellow and pink shaded areas correspond to the location of hydrate anomalies from Figure D.1.

$$\Delta t \sim \frac{z_{spike}^2 \Delta^2 \eta}{k \beta P_{ex}^2} \quad (D.6)$$

Using nominal parameters listed in Table D.1 we find that Δt can range up to 10^5 years, which is much longer than the expected timescale for the dissociation of hydrates driven by a thermal or pressure perturbation.

D.5. Concluding Remarks and Future Work

Our calculations suggest that high-concentration hydrate anomalies reduce the effective stress and prevent normal consolidation. Such hydrate spikes are expected to form at stratigraphic boundaries, where contrasts in sediment permeability and the potential for rapid consolidation increase the potential for slope instability. Using constitutive relationships between effective stress and porosity, we show that marine sediments with stable hydrates have porosity up to 6% greater than normally

Parameters	Value
Cohesion	
Sediment, C [Pa]	$0.03\sigma'$
Hydrate, C_h [Pa]	1.0×10^6
Friction angle	
Sediment, ϕ [°]	29
Sediment with hydrate, ϕ_h [°]	40
Sediment density, ρ_s [kg m ⁻³]	2670
Water density, ρ_l [kg m ⁻³]	1000
Slope angle, θ [°]	3 - 10
Sediment depth, z [m]	500
Surface porosity, n_o []	0.685
Athy's constant, b	4.0×10^{-4}
Lambert's function, W	$We^W = x$
Hydraulic diffusivity, D [m ² s ⁻¹]	$10^{-9} - 10^{-6}$
Anomaly thickness, z_a [m]	$10^{-1} - 10^0$
Permeability, k [m ²]	$10^{-18} - 10^{-14}$
Pore compressibility, β [Pa ⁻¹]	10^{-10}
Dynamic viscosity, η [Pa s]	10^{-3}

TABLE D.1. Nominal parameter values.

consolidated sediments. Although hydrates significantly reduce the effective stress, they tend to strengthen slopes due to increased cohesion and friction angle. Once hydrate is destabilized, the sediment strength is significantly reduced due to the increased porosity. Consolidation of the sediment will generate excess pore pressure that will further weaken the marine sediment and potentially trigger slope failure. Further, quantitative predictions will be possible with the completion and coupling of a pore-consolidation model. The infinite-slope model calculations presented here are instructive and point to the importance of hydrate reservoirs for slope stability issues, but they are limited in predictive capability because they do not account for the heterogeneous distribution of strength on the slip surface that is produced by the dissociation of hydrate anomalies. Ongoing work entails the use of a rate-and-state friction formulation, which should give a more refined view by accounting for changes in frictional resistance along finite slipping patches (i.e. the

size of dissociating hydrate anomalies) and the triggering produced by heterogeneous slipping as hydrate anomalies decay. This new formulation holds promise for examining how the dissociation of gas hydrates along stratigraphic boundaries can trigger and/or precondition slope failure along both passive and active continental margins.

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