Examining Slope Instability Dynamics on a Small Bank Slope Along the Schoharie Creek in New York State

by

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ABSTRACT

EXAMINING SLOPE INSTABILITY DYNAMICS ON A SMALL BANK SLOPE ALONG THE SCHOHARIE CREEK IN NEW YORK STATE

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Examining the slope instability dynamics on a bank slope along the Schoharie Creek near Burtonsville, New York was the goal for this research. This research aimed to examine the frequency and magnitude rates for slope instability and assess how changes in moisture relate to the instability observed at the site. Slope instability was determined from both absolute ring width differences and reaction wood presence. This research also compared precipitation and discharge with slope instability frequency and magnitude. This work concluded that slope instability frequency across the slope has shifted from episodic to more continuous instability frequency from 1969 to present, while the magnitude values remained consistent. The shift towards more continuous instability frequency coincides with a change from low to high precipitation and discharge starting in 1969. This research added to the literature as it proposed an alternate method to examine slope instability at a continuous scale.
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Chapter 1. Introduction

The global landscape is a dynamic system that is in constant change due to erosional and depositional processes. Slope processes contribute to landscape evolution through sediment removal and deposition resulting from both natural and anthropogenic factors. Focusing on river systems in the United States, erosional processes have elevated sediment in rivers to levels exceeding the streams’ natural variability (Davis and Harden, 2014; Dick et al., 2013; Verstraeten et al., 2009). Impacts associated with riverbank erosion include sediment loading, changes to river channel geometry, and changes to riparian biodiversity (Black et al., 2010, Dick et al., 2013; Ricker et al., 2008). Furthermore, increased erosion along bank slopes also results in high sediment yields at the catchment outlets, which increases silting into reservoirs and ports (Mathys et al., 2005; Saez et al., 2012; Verstraeten et al., 2006). Increased silting deteriorates storage lake functionality, such as hydro-electricity, as well as reduces dam functionality (Saez et al., 2012).

Identifying high sediment yield areas takes into account multiple factors such as: land use, slope, precipitation totals and intensity, runoff rates, soil cohesion, surface roughness, and flow volumes related to bank undercutting in high relief areas (de Vente and Poesen, 2005; Vanacker et al., 2003). On the global scale, high sediment yields act as pollution sources in rivers only if they exceed what is natural (Dick et al., 2013). In the United States, the United States Environmental Protection Agency (USEPA) has documented high sediment yields as being a primary pollutant to riverine systems (Dick et al., 2013; USEPA, 2002). River bank erosion results in approximately 80% of total sediment yield in some rivers and streams across the United States (Dick et al., 2013; USEPA, 1990). Although riverbank erosion is a main cause in high sediment yield to riverine systems, it is a natural process, which is necessary to produce
aquatic habitat and biodiversity (Florsheim et al., 2008). Therefore, understanding the occurrences and rates at which bank erosion from slope failure or instability have occurred along a bank slope are crucial to mitigating the negative impacts associated to high sediment yield in a particular system.

Slope failure contributes to erosion and downslope sediment transfer resulting in high sediment yields to nearby streams and rivers (Chorley et al., 1984; Varnes, 1978). Research on slope processes such as failure, includes studies by Aleotti and Chowdhury (1999), Anderson et al. (2007), Chorley et al. (1984), Papthoma-Kohle et al. (2007), Van Den Eeckhaut et al. (2007). Most studies examined the causes, impacts, and predicted locations that are most susceptible to failure (Baum and Godt, 2010; Corsini et al., 2006; Fiorucci et al., 2011; Gorsevski et al., 2006; Guzzetti et al., 2012). A key observation from past research is that slope instability did not result from one process alone and that processes acting on an area are mainly interconnected (Chorley et al., 1984; Guzzetti et al., 1999). The scales over which slope processes occur vary in time and space. Slope failures act as drivers changing the earth’s landscape as mass wasting events transfer sediment from sources to sinks (Guzzetti et al., 2005; Saez et al., 2012). With increased interest and awareness in slope failures due to socio-economic impacts and human pressures on the environment, understanding the spatial and temporal instability patterns as well as the factors triggering them is important to understand the conditions and slope dynamics that drive slope failure processes (Petrascheck and Kienholz, 2003; Saez et al., 2012).

In most cases, triggering factors for slope failure events are associated with hydrological inputs, changes to slope gradient, vegetation cover or removal, and substrate composition (Guzzetti et al., 2005). Since slope failure is a continuous process, it is extremely difficult to attribute a single cause to initiation. Slope instability results from interactions between the inputs
forcing the system, known as extrinsic factors or system drivers, and the system’s internal properties, known as the intrinsic factors (Harvey, 2007). The response to extrinsically induced events relates to the magnitude associated with the environmental change in conjunction with the system’s resistance to change (Harvey, 2007). Events vary greatly across temporal and spatial scales where some events displace material a few centimetres during a year, while others move material across many metres in minutes (Guzzetti et al., 2005). Furthermore, their movement typically includes flowing, sliding, toppling, falling, creep, or a combination of all these movements (Cruden and Varnes, 1996). Conventionally, geomorphic instability refers to short timescales where geomorphic response and recovery relates to individual storm events or short-term environmental change (Harvey, 2007). The temporal aspects associated with geomorphic instability relate to event frequency and magnitude (Harvey, 2007). Therefore, slope instability processes occur under numerous spatial and temporal scales with different techniques or tools applied to understand slope instability frequency and magnitude (Fall et al., 2006; Harvey, 2007).

The techniques used to study slope instability include three categories: expert evaluation, statistical methods, and mechanical approaches (Fall et al., 2006; Leroi, 1997). Each method used to assess slope stability presents multiple advantages and disadvantages (Fall et al., 2006). The success of one method over another is heavily dependent on the purpose for the research (Fall et al., 2006). Factors taken into account when selecting a specific technique to study slope instability include the study area scale, the accuracy of expected results, and the data available for analysis (Fall et al., 2006; Leroi, 1997). One such method that has advanced in dating slope instability processes in recent years is dendrogeomorphology (Guzzetti et al., 1999; Stoffel and Bollschweiler, 2008). Dendrogeomorphology provides a precise yearly tool for reconstructing
past slope instability events as well as to reconstruct time series for multiple geomorphic events (Stoffel et al., 2013). Dendrogeomorphology identifies slope instability events from tree ring records that contain mechanical disturbances related to instability (Stoffel et al., 2013). Applying methods such as dendrogeomorphology to examine slope processes provides a means to understand the temporal and spatial variability in slope dynamics that also contribute to sediment yield in riverine systems.

This thesis aimed to examine the slope dynamics on a riverine system along Schoharie Creek near Burtonsville, New York. Specifically, this work focused on the spatial and temporal occurrences in instability activity to assess the slope dynamics across the study site. The slope instability frequency and magnitude values were identified from growth anomalies in the sampled tree ring records, specifically eccentric growth and reaction wood presence, in eastern hemlock (Tsuga canadensis). This research also aimed to understand how changes in moisture affected slope instability at the site. In order to answer these questions, this study focused on the frequency and magnitude associated with slope instability events from tree ring growth anomalies attributed to slope instability. The frequency and magnitude values observed in the tree ring records identified potential patterns in slope instability for this site. Furthermore, this work compared the instability frequency and magnitude with precipitation and discharge records to examine if any changes in slope instability coincided with changes in moisture conditions on the slope. The findings serve to provide general insight into potential patterns in slope instability frequency and magnitude over time as well as understand if any changes in the frequency and magnitude patterns exist. Furthermore, the results from this research aimed to contribute to the methods used in dating and understanding slope instability using tree ring growth anomalies for future research.
Chapter 2. Literature Review

2.1 Introduction

Globally, geomorphic systems are under constant change (Chin et al., 2014; Wohl, 2014). The changes associated with geomorphic systems are a balance between internal and external forces (Guzzetti et al., 1999). The balance or equilibrium that exists on stable slopes is in constant flux where geomorphic processes have the potential to change the stability state for any system (Phillips, 2011; Harvey, 2007; Wohl, 2014). The forces driving slope instability events share both spatial and temporal dimensions (Harvey, 2007; Stefanini, 2004). Therefore, research that evaluates slope instability frequency and magnitude aims to determine patterns in slope movement and compare possible triggers with slope failures in order to understand the conditions in which slopes are most susceptible to failure (Stefanini, 2004; Wistuba et al., 2013). This chapter describes and summarizes the research literature for slope dynamics and processes associated with slope instability. This work begins with literature on landscape equilibrium with emphasis on the factors influencing slope stability. This chapter then examines specific slope instability processes and changes in sediment yield associated with slope failures. Finally, this chapter ends with an evaluation on methods used to study slope instability processes with emphasis on dendrogeomorphological approaches. Furthermore, this chapter examines some gaps in the literature with regards to methods used to examine slope instability with tree ring growth records and the anomalies in the records associated with slope instability events.

2.2 Landscape Equilibrium

Slopes remain stable if the resisting forces and driving forces acting on a surface remain in equilibrium (Chorley et al., 1984). Due to global variations in climate, vegetation, lithology, and structure, understanding slope erosional processes acting in a specific area can be quite
difficult (Chorley et al., 1984). Since it is challenging to formulate a general model for studying slope evolution and development, there have been multiple geographical approaches to examining equilibrium processes affecting slope stability on slopes (Chorley et al., 1984). Numerous studies over the past century have defined and analyzed equilibrium concepts and classifications for sloped landscapes (Ahnert, 1994; Gilbert, 1877; Harrison, 1999; Renwick, 1992; Phillips, 2009; 2011). Gilbert (1877) first introduced the term dynamic equilibrium to refer to any change in a geomorphic system that causes processes to operate in a way that minimizes the effects associated with the changes to the system. Gilbert (1877) also defined dynamic equilibrium as the system changing process rates over time in order to minimize changes within the system. The equilibrium concept has evolved over the past century and was as a general model used to explain landform pattern evolutions that appear balanced in space and/or time (Renwick, 1992). In geomorphological processes, equilibrium exists if the material removed from an areal unit per unit time is equal to the material supplied to the same areal unit during the same time (Ahnert, 1994).

Equilibrium discussions in geomorphologic processes, such as slope failure, have led to other classifications of this concept including dynamic metastable, steady state, disequilibrium, and non-equilibrium (Chorley et al., 1984). There are numerous process-response systems acting on a slope, which affect the equilibrium keeping the system in balance. The inputs acting on a slope include solar radiation, precipitation, dissolved substances and solids, as well as weathering derived debris (Guzzetti et al., 1999; Ono et al., 2011). Outputs along a slope include evapotranspiration, percolation and water runoff, and sediment removal from erosion or incision by rivers at the slope toe (Ono et al., 2011). The inputs and outputs of a slope system combined with other factors such as geology, climate, slope gradient, aspect, gravity and vegetation, impact
slope stability and may lead to slope failure (Guzzetti et al., 1999). Understanding the complexities in geomorphic systems, including the interactions among factors and processes acting on certain landforms is dependent on the ability to comprehend the balance of these forces in a specific area.

Self-regulating forces maintain dynamic equilibriums in geomorphic processes, such as slope instability, as they act on an area over time (Ahnert, 1994). Slope failure events produce negative feedbacks between process components where a change in one process frequency causes changes in other processes that lead to counteracting effects against the initial change (Ahnert, 1994). Behaviorally, an equilibrium landform is not static or completely stable; hence, there is a tendency for the form to remain stable and to return to stable states following a threshold event, such as a slope failure (Renwick, 1992; Figure 2.1). Changes to the dynamic equilibrium model resulted in the dynamic metastable equilibrium concept (Renwick, 1992). The dynamic metastable equilibrium concept represents system dynamics over a long timescale where threshold events affect the system in question (Renwick, 1992). Therefore, dynamic metastable equilibrium is discontinuous as it jumps between different thresholds, but continues to vary around an average trend (Schumm, 1975).
Figure 2.1: Equilibrium, disequilibrium, and nonequilibrium for landform behavior over time (Renwick, 1992).
A second equilibrium type known as steady state equilibrium, occurs in landforms that may contain a constant form, however they are not necessarily static (Ahnert, 1994; Bracken and Wainwright, 2006; Chorley and Kennedy, 1971; Phillips, 2009; 2011; Renwick, 1992; Schumm and Lichty, 1965). The steady state equilibrium toward which an open system evolves is not just a stationary state, but also refers to the lowest energy state (Thorn and Welford, 1994). Therefore, dynamic equilibrium differs from steady state in the sense that the former accounts for gradual changes in landform to maintain equilibrium with varying environmental conditions (Bracken and Wainwright, 2006). In regards to topography and erosional processes, understanding the steady state equilibrium in these landforms is important as these represent the stable states towards which the dynamic tectonic-erosion system evolves (Willett and Brandon, 2002). Even if the landform never attains a steady state, the degree to which it approaches steady state equilibrium provides a measure of its age (Willet and Brandon, 2002).

Other equilibrium types, which result from geomorphological processes, include disequilibrium and non-equilibrium. Disequilibrium landforms have progressive change in both form and outputs that may result from either low land forming process rates or recent environmental change (Renwick, 1992). The processes exhibiting disequilibrium are impacted by negative feedbacks, thus they may change towards a more dynamic equilibrium system (Ahnert, 1994). Therefore, disequilibrium at a landscape feature is in constant change in an attempt to regain balance, but not yet in balance (Renwick, 1992). Non-equilibrium is a system where equilibrium is absent despite long stability records (Renwick, 1992). In geomorphological processes, non-equilibrium tendencies occur during the early phases in development after input changes, when positive feedback rather than negative feedback results in differentiation in landscape features (Ahnert, 1994; Figure 2.1). These processes typically occur during short-term
development phases where by the end, they move towards dynamic or steady state equilibrium (Ahnert, 1994). Therefore, it is important to examine geomorphological process-response systems at different spatial and temporal scales to understand the events occurring across a slope.

Recent research on steady state equilibrium has evolved to include ideas on pseudo-equilibrium or nonequilibrium and flexibility (Brierely et al., 2013; Phillips, 2011). The pseudo-equilibrium concept builds upon the notion that instead of a steady state equilibrium leading to a system’s goal or normal condition, steady states may represent an upper limit in some geomorphic systems such as slope instability (Philips, 2011). Applying the pseudo-equilibrium idea to hillslope gradients provides an example for gradient selection and threshold modulation (Phillips, 2011). The apparent steady state equilibrium achieved from the angle of repose that is developed and maintained on steep slopes is not a normal property for the system as there are many processes acting to change the slope gradient, which is achieved once a critical threshold is reached (Brierely et al., 2013; Phillips, 2011). The pseudo-equilibrium notion provides further evidence that analysis and understanding in geomorphology must include a balance between local or place based knowledge that meaningfully relate theory and practice (Brierely et al., 2013). Therefore, being able to read the landscape provides flexibility in the interpretation and discussion of results in geomorphological research, which further provides a means of combining the findings and geomorphological theory (Brierely, et al., 2013). Furthermore, the pseudo-equilibrium notion does not dispute the existence of steady state equilibrium in the real world, but instead identifies the variability in which geomorphic systems act and were not defined as a restricted state space (Brierely et al., 2013; Phillips, 2011).
2.3 Slope Stability Dynamics

In general, equilibrium concepts were predominately applied to large spatial and long temporal scales related to Quaternary climatic changes or tectonic disturbances (Chin et al., 2014; Harvey, 2007). However, equilibrium concepts need to apply to geomorphic instabilities that commonly occur over shorter timescales such as the geomorphic response to single storm events or yearly changes (Phillips, 2005). The instability principle does not imply that all geomorphic systems are unstable, rather that instability is common in most geomorphic systems and stable, steady-state equilibrium is always the norm (Phillips, 2005). Two fundamental concepts applied to geomorphic events or threshold-exceeding events are magnitude and frequency (Schumm, 1979; Wolman and Miller, 1960). The geomorphic thresholds in which, the geomorphic system abruptly changes in response to internal and external changes (Schumm, 1979). For geomorphic instability events, the issue surrounds whether the thresholds reached are reversible or irreversible (Chorley and Kennedy, 1971).

The focus on understanding geomorphic processes has shifted from examining them within conceptual models towards recognizing the complex interactions among landforms, biota, and humans (Bracken and Oughton, 2014; Demeritt, 2009; Harden, 2012; Wohl et al., 2014). Examining the interactions between system components builds upon the equilibrium concept and focuses on an alternate state relating to feedback loops (Wohl et al., 2014). Feedbacks occur when an initial perturbation triggers a system response that grows greater with time (Chin et al., 2014; Wohl et al., 2014). For example, in an area affected by a landslide deposit, the changes to the bank slope result in greater infiltration capacity, which impacts soil cohesion but also allow vegetation growth that enhances sediment deposition and aides in slope stability (Bennett et al.,
The alternate state and feedback loops associated with them play a factor in conceptualizing slope instability via a transient ratio form (Wohl et al., 2014).

Landscape system stability relates to the recurrence interval for an event causing change and the time necessary for a system to return to its equilibrium (Brunsden and Thornes, 1979; Wohl et al., 2014). The questions concerning systems where the exceeded threshold is irreversible is if a new but different equilibrium condition exists and if stability for these systems are affected by lower magnitude events inducing frequent instability events (Chorley and Kennedy, 1971). Furthermore, the temporal aspect associated with geomorphic instability combines the magnitude, frequency, and threshold principles with concepts related to the geomorphic system’s internal structure, particularly sensitivity (Chin et al., 2014; Phillips, 2005).

Geomorphic system sensitivity related to the relationship between threshold event frequency and recovery time, where the recovery time is the stabilization rate after a disturbance (Brunsden and Thornes, 1979; Harvey, 2007). Therefore, systems that recover slowly from disturbance events are sensitive systems, which exhibit considerable instability in response to disturbances, while resilient systems recover quickly (Harvey, 2007; Wohl et al., 2014). On the local scale, the sediment delivery from bank slopes into the channels predominately occurs because of instability events, such as slope failures, that have considerable influence on channel processes, morphology, and stability (Harvey, 2007). The differences in resilience between geomorphic systems relate to the magnitude and frequency associated with the events causing the changes to the system (Chin et al., 2014; Wohl et al., 2014). The magnitude and frequency concepts from Wolman and Miller (1960) introduce a key issue in understanding the resilience for a specific geomorphic system (Wohl et al., 2014). Specifically for slope stability studies, determining if instability is driven by infrequent but large threshold exceeding events or
frequent, smaller magnitude threshold exceeding events provides valuable information to aid in understanding the conditions in which the geomorphic system is unstable (Chin et al., 2014; Wohl et al., 2014).

The landscape equilibrium acting on a slope for translational landslides is a function of the safety factor or the stability ratio. In order to understand the complex slope dynamics on hillslopes, past studies focused on the balance between shear strength (resisting) and shear stress (driving) (Aleotti and Chowdhury, 1999; Alexander, 2008; Chorley et al., 1984; Martel, 2004; Turconi et al., 2010). Examining the dynamics between resisting and driving forces on a hillslope provides a means to understand the forces acting toward stability. In cases where the driving forces are more dominant than resisting forces the safety factor decreases, triggering slope failure (Phillips, 2011). Major stresses acting on a slope include gravitational force, slope angle, and the overburden mass (Bennett et al., 2013; Jakob and Weatherly, 2003; Knapen et al., 2006). The forces acting on a slope are normal stress and shear stress, where normal stress is the perpendicular component for the total stress and shear stress is the gravitational forces (Lade, 2010; Scheidegger, 1975; Figure 2.2). Opposing shear and normal stress are the forces resisting slope instability known as shear strength. Shear strength counteracts shear stress and is built upon frictional characteristics expressed as internal friction angle, effective normal stress, and cohesion (Chorley et al., 1984; Hilton, 1979; Lade, 2010; Scheidegger, 1975). Both shear strength and stress increase with depth, however shear strength decreases more rapidly with ground depth compared to shear stress (Lade, 2010). Therefore, not only will slope instability occur when shear stress exceeds shear strength, but there will also be a critical depth before instability occurs (Lade, 2010).
Figure 2.2: Effective strength parameters used to find the factor of safety for a saturated, homogenous, infinite slope (Lade, 2010). The surficial stability includes total unit weights and water pressures: (a) for forces acting on a cohesive soil block, and (b) forces parallel and perpendicular to soil slope inclination $\alpha$ (Lade, 2010). The total weight of the block, $W$, is a product of the saturated unit weight of the soil, $\gamma_{\text{sat}}$, the depth $h \cdot \cos \alpha$, and length, $b$. 
2.4 Slope Instability Processes and Products

Defining slope failures or instability events, specifically landslides, has provided many challenges for physical scientists (Guzzetti et al., 1999). One widely accepted definition encompasses magnitude, geographical location, and time recurrence concepts (Guzzetti et al., 1999; Varnes et al., 1984). In general, slope failures occur when driving forces outweigh resisting forces resulting in downward movement of sediment material (Guzzetti et al., 1999; Varnes et al., 1984). Based on these concepts, understanding the intensity, identifying the place and temporal frequency in which these natural processes occur is most important when studying slope failures. Due to the complexity and variability in landslide interactions with the environment, it is difficult to accept a single definition for landslide hazards in particular. Therefore, this section explores a few slope failure events, the triggers associated with the events, and sediment yield as a product from slope failure in riverine systems.

2.4.1 Slope Processes

Research on landslides and soil creep occur on the global scale since they have environmental, social, and economic impacts (Ardizzone et al., 2007; Bielecki and Mueller, 2002; Haneberg et al., 2009; Jahn, 1989; Mantovani et al., 1996; Metternicht et al., 2005; Nash and Beaujon, 2006; Nichol and Wong, 2005; Rosenfeld, 1999; Van Den Eeckhaut et al., 2007; Van Westen et al., 1999). Landslides are either terrestrial or sub-aqueous and involve flowing, sliding, toppling, or falling movements (Guzzetti et al., 2012). In most translational landslides, the planar rupture surface is roughly parallel to the ground surface at shallow depths (Chorley et al., 1984). Rotational landslides occur along curved surfaces in a concave upwards motion (Ofoegbu and Ferrill, 1998; Williams and Vann, 1987). Rotational slides also tend to extend to greater depths than translational slides, thus making them less frequent (Mackey and Roering,
Studies on slumps determined two distinct feature classifications: simple and multiple rotational slumps (Varnes, 1978). Simple rotational slumps occur when the mass moves in an essentially coherent unit along a single surface (Chorley et al., 1984). Multiple rotational slumps are movements in blocks, which occur along several curved slope surfaces (Thornbury, 1969). Multiple rotational slumps follow along one planar failure, but have multiple scarps that form from these events (Varnes, 1978; Figure 2.3). Soil creep varies from slides in movement, velocity, and triggers. Soil creep is a heave movement through the expansion in the material moving downslope (Sasaki et al., 2000). Typically, soil creep occurs in the top several feet in a soil profile, and its effects decrease rapidly with depth (Kaitna et al., 2013; Lade, 2010). Soil creep has the ability to rearrange particles, thus reducing the available resistance between the materials (Kaitna et al., 2013; Lade, 2010). Since soil creep has the ability to weaken the resisting strength on a slope, much research has focused on creep as a precursor to larger slope failure events such as landslides (Jahn, 1989; Matsuoka, 1996; Sasaki et al., 2000; Xu et al., 2011; Yamada et al., 1999). The downslope movement rates for soil creep are variable due to differences in slope angle and moisture content (Sasaki et al., 2000).

Although the three aforementioned slope failure processes produce different morphological features, the causes resulting in slope failures are similar. Triggering factors for slope failures are external stimuli responsible for movement initialization (Knapen et al., 2006). Initialization factors include, but are not limited to: earthquakes, volcanoes, extreme and/or excessive rainfall events, anthropogenic disturbances such as slope excavation, river incision into the slope toe, and changes in landuse and vegetation (Knapen et al., 2006; Vanacker et al., 2003; Van Beek et al., 2008). Extreme and antecedent rainfall events are the most common and well-documented triggers associated with slope failures (Chen et al., 2006; Dahal and Hasegawa,
Water accumulation from extreme and/or prolonged rainfall events adds extra weight to a slope leading to an increase pore pressures, which adversely affect the available shear strength (Van Beek et al., 2008). In most slope failure events the moisture added to the system through intense, short-term rainfall or smaller magnitude, prolonged precipitation play a role in triggering movement (Hilton, 1979; Ono et al., 2011). The buildup in pore water pressure beyond hydrological thresholds following intense or prolonged rainfall events is a typical trigger for slope failure (Ono et al., 2011). Similarly, changes in land cover through vegetation removal causes a loss in root tensile strength, loss in canopy cover, and changes to the slope hydrology (Osman and Barakbah, 2011; Van Beek et al., 2008). Changes in vegetation cover result in erosion and diminished shear resistance across the slope, but may also aid in resisting slope failure depending on what changes in vegetation are done (Osman and Barakbah, 2011; Van Beek et al., 2008).

Another common trigger for slope failures is river incision and erosion at the slope toe (Vanacker et al., 2003). River incision into the slope toe leads to a reduction in slope stability as well as increases sediment load downstream (Vanacker et al., 2003). As the river continually erodes into the slope toe, the shear stress applied overcomes the shear resistance resulting in a translational landslide (Vanacker et al., 2003). Past studies focusing on slope failures and their triggers have had difficulty in identifying a single trigger behind failure events (Ono et al., 2011; Stefaninini, 2004). The complexity associated in geomorphic systems impacted by slope failure proves difficult to identify single triggering factors for slope failure (Chin et al., 2013; Wohl et al., 2014). Slope failure initialization from one trigger has the potential to result in positive or negative feedback loops, which may lead to future failures from different triggers (Chin et al., 2013; Wohl et al., 2014). Therefore, past studies on slope instability such as those by Silhan et
al., (2014), Stefanini (2004), and Wistuba et al., (2013) examine the connection between failure occurrence and multiple triggering events related to precipitation, seismic activity, and discharge resulting in slope undercutting. Although it may not be possible to pinpoint a sole cause for a slope failure event, being able to examine the complex relationships between these triggers will help to determine the susceptibility for a particular area as well as mitigate any environmental, societal, and/or economic impacts which may arise.
Figure 2.3: Multiple rotational landslide event (Abbott and Samson, 2009; Varnes, 1978).
2.4.2 Sediment Yield as a Result from Slope Failure

Sediment delivery into stream networks is dependent on two factors: first is the material available to be eroded from the landscape; and second is the connection to the drainage network (Reid et al., 2007). Hillslope environments have the potential to fail repeatedly, which generate sediment for long time periods until the sediment supply becomes exhausted (Reid et al., 2007). The conditions in which most slope failures occur are during rainfall events where the eroded material may supply sediment to the channel system (Lane et al., 2003; Reid et al., 2007). Hydrologic connectivity is important in hillslope systems as it determines the extent to which water and matter moves and is stored across the system (Lane et al., 2003). The sediment transfer rates from hillslope to fluvial systems are limited by the sediment delivery rates associated with hydrologic connectivity (Reid et al., 2007). The sediment delivery is a function of the connection between the displaced material and the channel (Reid et al., 2007). The hydrologic connectivity can lead to long or short term failed sediment storage and affect the sediment yield into riverine systems (Lane et al., 2003; Reid et al., 2007).

Although there are multiple slope failure classifications, a common result associated from slope failures in fluvial systems is sediment mobilization resulting in high sediment yield into the system (Ono et al., 2011). Studies by Crosta and Frattini (2008), Korup et al., (2010), and Ono et al., (2011) have focused on sediment yield impacts resulting from slope failure events. The increases in sediment yield from slope failure events result in both environmental and economic impacts (Ono et al., 2011). Erosion along riverine systems is a concern for hazard control and risk mitigation along steep sloped areas (Ballantine et al., 2009; Fontana and Marchi, 2003). Erosion and deposition processes contribute to landscape evolution in high relief areas influenced by topographic, climatological, geological, and land use conditions (Fontana and
Marchi, 2003). In humid climates with varying relief, shallow landslides and debris-flow events act as the principal sediment supply to the channel network (Fontana and Marchi, 2003). Economically, the impacts from high sediment yields induced by slope failure affect reservoirs, hydroelectric dams, and other infrastructure facilities (Kunkel et al., 1999; Ono et al., 2011). Furthermore, high sediment concentrations degrade drinking water quality, increase costs to treat water, and are harmful to aquatic organisms (Ono et al., 2011; Waters, 1995). Therefore, the interactions between these slope failure phenomena mixed with the channel network and downstream sediment routing efficiency are the main controls in watershed sediment yields (Ballantine et al., 2009).

Studies on sediment yield related to slope failure processes have focused on ranging temporal and spatial scales (Allen and Hovius, 1998; Hapke, 2005). The focus in most studies is estimating sediment yields based on lithology, local climate, land use practices and geomorphological processes (Allen and Hovius, 1998). Many recent process based studies focus on examining the strong relationship between tectonic uplift/subsidence and erosion along slopes resulting in direct or indirect sediment movement into nearby river systems (Burton and Bathurst, 1998). In areas with varying relief, sediment yield occurs when valley lowering and sediment removal from adjacent slopes is present. Fluvial incision is the principle factor-influencing valley lowering which leads to increased erosion resulting in high sediment yields during periods with peak discharge (Hapke, 2005). These valleys become incised into the slope base whose mass transfer rate is limited by weathering and anthropomorphic factors (Allen and Hovius, 1998). If the valley incision is occurring at a faster rate than weathering along the slope then instability greatly increases due to the steepening along the slope (Hapke, 2005). Also, the continuous incision coupled with slope failure triggers such as prolonged or heavy rainfall or
seismic activity greatly impacts the sediment yield in hillslope environments (Burton and Bathurst, 1998). Although slope failures do not always occur for long periods, their effect on the sediment yield budget at a specific location has the potential to surpass the average annual sediment yield for an entire watershed (Allen and Hovius, 1998; Burton and Bathurst, 1998). Therefore the long-term sediment yield averages in these unstable topographic regions is dependent upon cycles in river valley incision and short intervals with enhanced erosion from slope failure processes (Burton and Bathurst, 1998; Hapke, 2005).

2.5 Methods Used in Past Slope Instability Research

Multiple techniques have been used to study slope instability events in the past (Dai and Lee, 2002; Fall et al., 2006; Guzzetti et al., 1999; Stoffel and Bollschweiler, 2008). Three divisions for the techniques used to study unstable areas are expert evaluations, statistical methods, and mechanical approaches (Fall et al., 2006; Leroi, 1997). In slope instability studies, expert evaluation predominately focused on susceptibility mapping (Fall et al., 2006). The second type, statistical methods, determines past instability processes from the statistical determination of the variables causing instability (Dai and Lee, 2002; Fall et al., 2006). The final set of techniques, mechanical approaches, evaluates and analyzes slope stability using modeling and remote sensing techniques (Fall et al., 2006; van Westen et al., 1997). There are advantages and disadvantages to all three types (Guzzetti et al., 1999). The success in using a particular approach depends on the study area scale, the available data, and the accuracy for both the data collected and expected results (Fall et al., 2006; van Westen et al., 1997). The limitations in each method provide additional challenges in determining the best technique for a specific study. Addressing these challenges has conventionally resulted in combining multiple techniques to examine slope instability studies (Fall et al., 2006; Guzzetti et al., 1999; Silhan et al., 2014;
Stoffel and Bollschweiler, 2008). Furthermore, advancements in GIS, remote sensing, and dendrogeomorphology techniques over the past decade have increased the accuracy in analyzing slope dynamics and instability events (Fall et al., 2006; Guzzetti et al., 1999; Saez et al., 2012; Stoffel and Bollschweiler, 2008). Recent research using dendrogeomorphological methods aims to identify slope instability frequency using growth anomalies in tree ring records (Saez et al., 2012; Stoffel and Bollschweiler, 2008; Wistuba et al., 2013).

2.5.1 Dendrogeomorphology

Dendrogeomorphology dates and quantifies earth surface processes through tree ring records (Bollschweiler et al., 2008; Hupp, 1984). The earliest natural hazard studies using dendrogeomorphology focused on flood events, as they examined tree rings to determine high magnitude flood events (Helley and LaMarche, 1968; St. George and Nielson 2003; Yanosky, 1984). In more recent years studies have applied dendrogeomorphology to sediment mobilization, specifically to slope failure events (Arbellay et al., 2010; Carrer and Urbinati, 2004; Scuderi et al., 2008). These studies designed to reconstruct past events on annual temporal scales relating to occurrence rates (Arbellay et al., 2010; Carrer and Urbinati, 2004; Scuderi et al., 2008; Silhan, 2012; Timell, 1986). Past research on identifying landslide dynamics and occurrence rates using dendrogeomorphology have successfully identified potential triggers related to meteorological data (Fantucci and Sorriso-Valvo, 1999; Stefanini, 2004). Also studies using tree ring records are able to reconstruct past events by examining tree reactions and responses resulting in tree growth abnormalities and are then able to attribute anomalies to specific geomorphic processes (Arbellay et al., 2010; Carrer and Urbinati, 2004; Scuderi et al., 2008; Silhan, 2012).
2.5.2 *Tree Ring Formation*

Tree ring growth follows a cyclical pattern, which is dependent on the seasonality and climate conditions for the respective study site (Stoffel and Bollschweiler, 2008). For example, in temperate climate zones, the tree ring-forming period lasts roughly from spring to autumn (Stoffel and Bollschweiler, 2008). Numerous internal and external factors affect tree growth rates. Internal factors include tree species ageing or sensitivity, whereas external factors influencing tree growth include water and nutrient supply, light, temperature, air and soil quality, and geomorphic processes (Schweingruber, 1996). The cyclical pattern exhibited by tree growth allows for accurate age assessment; however, the rate at which tree ring formation occurs varies based on tree species (Campbell, 1997; Rigling et al., 2002; Figure 2.4). Tree ring formation in conifers occurs in two periods: early and late vegetation period (Camarero et al., 1998; Rigling et al., 2002). During the early stage, the cambium forms large and thin-walled cells, defined as tracheids (Campelo et al., 2013; Uggla et al., 2001). However, later in the vegetation period the tracheids produce smaller and denser latewood leading to thicker cell walls that have a darker appearance (Campelo et al., 2013; Uggla et al., 2001). The way in which the vessels occur creates a pattern in the wood, resulting in wood with statistically scattered pores or wood with ring-shaped pores (Campbell, 1997).

Variations in tree ring growth are also density dependent relative to tree age and species (Fritts, 1976; Schweingruber, 1996). Abiotic features affecting growth include light, temperature, water supply, nutrient supply, and are similar for trees growing at a specific site (Stoffel and Bollschweiler, 2009). It is common for tree ring records to have variations in ring widths solely related to tree growth itself (Fritts, 1971). Conifers typically have wider tree ring production during early tree life causing a decreased growth trend (Fritts, 1971; Figure 2.4). It is also
possible to infer from the tree ring records environmental impacts and fluctuations from precipitation and seismic activity, as variation in growth is a function associated with these external environmental processes (Cook and Kairiukstis, 1990; Stoffel and Bollschweiler, 2009). Furthermore, mechanical disturbances from slope failure processes are visible in tree ring series affected from these processes (Stoffel and Bollschweiler, 2009). In studies on slope instability, the typical growth disturbance focused on is stem tilting, which results in growth eccentricity (Stoffel and Bollschweiler, 2009).
Figure 2.4: Tree ring cross section for a conifer species (Fritts, 1976).
2.5.3 Tree Ring Growth Anomalies

Studying growth disturbances related to geomorphic processes result from the process, event, and response concept first defined by Shroder (1978). This concept defines the process as a geomorphic event, such as a debris flows or flood, which produces various events impacting tree growth, as a result of tree inclination, stem burial, or root exposure for example (Shroder, 1978; Figure 2.5). The trees react to these events in the following ways: reaction wood growth, growth suppression, growth termination, or candelabra formation (Bollschweiler et al., 2008; Ruiz-Villanueva et al., 2010; Shroder, 1978; Strunk, 1997; Szymczak et al., 2010). It is possible for affected trees to show one or multiple response signs listed above for a single event (Stoffel and Bollschweiler, 2008). This section will examine the different events and associated responses affecting tree growth due to slope failure are explored in detail.
Figure 2.5: The Process-Event-Response concept in diagram form introduced by Shroder (1978).
Tree tilting is a common impact associated with slow moving slope failure processes such as soil creep. Numerous studies use dendrogeomorphology to examine tilting trees across Europe and North America as a record for dating past geomorphic processes (Braam et al., 1987; Casteller et al., 2007; Ruiz-Villanueva et al., 2010; Shroder, 1987; Strunk, 1997). Stem inclination occurs due to either sudden pressure change induced by mass movement or slow, continuous tree destabilization due to soil creep and shallow landslides (Lundstrom et al., 2007; Strunk, 1997). Trees that are experiencing tilting try to regain vertical position, thereby affecting the growth rate, which in turn affects the tree ring series (Silhan et al., 2012; Stoffel and Bollschweiler, 2008). Tilting stems results in changes from concentric to eccentric ring growth by producing dense, dark-coloured, and thicker reaction wood (Scurfield, 1973; Shroder, 1978; Figure 2.6).

In conifers, reaction wood is produced on the tilted stem side, causing considerably larger individual rings that are slightly darker in appearance compared to the opposite trunk side (Stoffel and Bollschweiler, 2009). The eccentrically shaped tree rings have stems that tilt toward the slope instability direction (Schweingruber, 1996). The increased density resulting in tightly packed rings provides a difference in colour due to the thicker, rounded cell walls in the early and latewood tracheids (Timmel, 1986). The eccentric growth present in the tree ring series allows yearly dating for the slope processes (Stoffel and Bollschweiler, 2009). Other studies suggest that trees affected by tilting respond with overall reduced growth (Bollschweiler et al., 2007). The upheaving and abrupt tilting results in root destruction that in turn reduces the annual tree ring size (Stoffel and Bollschweiler, 2008).
Figure 2.6: Stem deformation and resulting eccentric growth from coniferous trees under stress from slope instability (B in the downslope and C in the upslope direction) compared to trees with straight stems (A). D: A sampled core from an eastern hemlock with upslope growth eccentricity (Wistuba et al., 2013).
2.6 Conclusion

Past research on sediment mobilization resulting from slope failure processes reveals that high sediment yields in riverine systems affect both the environment and economy. Slope failures are complex landscape processes that result from multiple geomorphic processes acting on the landscape. Identifying slope failures moving towards a single equilibrium is quite challenging as these processes most likely exceed critical thresholds, thus changing the previous. In response to the challenges in studying slope failures, researchers aim to identify the frequency for instability events and understand the dynamics and conditions in which slopes are unstable. However, the factors controlling stability and the processes that act as triggers provide difficulty in identifying the magnitude and frequency behind slope instability. As a result, researchers strive to implement new methods that help to identify the frequency rates associated with slope failure. Dendrogeomorphology is one such method that aims to use anomalies in tree ring records to identify frequency rates associated with slope instability. This method provides an opportunity to recreate annual records for instability over centuries depending on the tree ages sampled. Furthermore, studies by Silhan et al., (2013), Stefanini (2004), and Wistuba et al., (2013), identify potential triggers, such as precipitation, discharge changing the slope toe angle, and seismic activity from comparisons with the instability years determined by the tree ring records. However, little research focuses on the magnitude associated with the growth anomalies identified from the tree ring records. The tree ring records provide a potential means to examine how the magnitude for recent slope instability events compares to past events. Thus, there is a demonstrated need to expand understanding of slope instability frequency, magnitude, and the conditions in which these events occur.
Chapter 3: Assessing slope instability dynamics on a bank slope along Schoharie Creek, Burtonsville, New York

3.1 Abstract

The main purpose for this study was to examine slope dynamics along a complex river bank slope adjacent to Schoharie Creek in New York. This research used absolute eccentric growth and reaction wood presence between the upslope and downslope direction from 111 eastern hemlock (*Tsuga canadensis*) to identify slope instability events. The variances in absolute eccentric growth for each tree revealed that the absolute growth eccentricity values were variable across the entire site. This research identified instability as years where absolute growth eccentricity exceeded a set threshold of 0.675 mm and contained reaction wood. The absolute growth eccentricity and reaction wood presence interpreted both the frequency and magnitude for slope instability events. The instability frequency appeared to shift from episodic instability to more continuous instability starting in 1969 for trees at approximately 20 to 40 m from the bank edge. The magnitude values identified from slope instability events did not change over time. Further analysis compared the instability events with precipitation and discharge data to determine the conditions during years with slope instability. The precipitation and discharge records revealed that the apparent shift to more continuous instability frequency from 1969 to present might have resulted from an exceeded threshold where there was a change from dry conditions in the 1960s to consistently wet conditions from 1969 to present. The results demonstrated that instability is disconnected across the slope and that there was a potential threshold exceeded, which changed the equilibrium state for this site in 1969 to possibly an alternate steady state equilibrium. Furthermore, the results revealed that slope instability
occurrence changed from episodic to more continuous frequency from 1969 to present, but the magnitude values associated with slope instability remained steady for the entire record.

3.2 Introduction

Slope instability processes as global phenomena in mountainous, escarpment, and river bank environments, which impact sediment loading to nearby rivers and channels and destroy infrastructure (Saez et al., 2012; Shroder et al., 2012). Soil erosion associated with mass wasting processes is an environmental issue worldwide with impacts on downstream sedimentation, reduced river hydraulic capacity, increasing flood risks, and reduced reservoir and dam functionality (de Vente and Poesen, 2005; Romero-Diaz et al., 2012; Scuderi et al., 2008; Sinnakaudan et al., 2003; Stoffel et al., 2013). Characterizing slope failure frequency was crucial for understanding slope instability and factors triggering slope instability processes (Silhan et al., 2014). Potential triggers associated with slope instability include changes in moisture availability from precipitation, seismic activity, change in slope geometry from undermining at the slope toe, and slope strength reduction by weathering or land use change (Corominas and Moya, 1999). Understanding the relationship between slope instability and their triggers was possible by reconstructing an occurrence record from proxies of slope change (Silhan et al., 2014; Wistuba et al., 2013). One approach to studying slope instability was to examine growth anomalies from trees impacted by slope processes (Saez et al., 2012; Malik and Wistuba, 2012; Wistuba et al., 2013; Silhan et al., 2014).

Dendrogeomorphological techniques have successfully reconstructed past slope instability events (Saez et al., 2012; Stoffel et al., 2013). Dendrogeomorphology examines the annual tree ring growth patterns for sampled trees to identify slope failure processes and the times over which these processes occur (Alestalo, 1971; Braam et al., 1987; Schweingruber,
1996; and Shroder, 1980; Silhan et al., 2012). Under normal growing conditions, tree growth on both stem sides should be nearly symmetrical for trees not impacted by slope processes (Alestalo, 1971; Braam et al., 1987; Schweingruber, 1996; and Shroder, 1980). Trees affected by slope processes exhibit asymmetrical growth, which indicate the timing and extent for the slope processes (Braam et al., 1987; Schweingruber, 1996). In coniferous trees, eccentric growth will form wider rings on the compressed tree side (Schweingruber, 2007; Timell, 1986). Furthermore, trees under stress from slope instability also tend to exhibit reaction wood (Arbellay et al., 2010; Timell, 1986). For example, in cases when the tree is tilting downslope it was typical for reaction wood to form on the downslope side (Schweingruber, 1996). Reaction wood occurs in individual rings, which were defined by the denser cell patterns formed from stress related to gravitational forces (Stoffel and Bollschweiler, 2009; Timell, 1986).

Trees reacting to slope instability result in distinct changes in growth pattern curves where there was a release in growth on only one tree side (Stoffel and Bollschweiler, 2008). Studies by Corominas and Moya (1999; 2010), Panek et al. (2011), and Stefanini (2004) used eccentric growth in combination with reaction wood to analyze slope failure timings and spatial distributions. To build on past studies, which predominately focused on identifying the exact year for a failure event, this study aimed to examine the frequency and magnitude for all observed instability events using absolute ring width differences and reaction wood presence. This study examined a century long record for slope instability from 111 eastern hemlocks (Tsuga canadensis) on a west facing bank slope along the Schoharie Creek. This study proposed that there was a connection between instability in the downslope to upslope and upstream to downstream directions on an active slope failure. This study also hypothesized that changes in moisture added to the slope, specifically in the form of precipitation and discharge, will coincide
with slope instability frequency and magnitude. This research also aimed to answer the question on what patterns exist for slope instability frequency and magnitude on the slope and how do changes in moisture coincide with the instability patterns observed at the site. The primary objectives for this study were to: (i) examine tree ring growth differences and reaction wood presence; (ii) assess the frequency and magnitude patterns for slope instability along the bank slope; and (iii) evaluate the observed slope instability events with precipitation and discharge data to determine if these processes coincide with changes in slope instability frequency and magnitude patterns.

3.3 Study Site

This research focused on an active slope along Schoharie Creek, near Burtonsville, New York (42°48’18” N, 74°15’18”W; Figure 3.1). Schoharie Creek flows north from the Catskills into the Mohawk River (Zembrzuski and Evans, 1989). The creek flows northerly for 140 km with the headwaters located at 600 m elevation in the Catskill Mountains, and the confluence at 90 m elevation at the Mohawk River (Daniels, 1998). The Schoharie watershed covers an area approximately 2500 km² (Daniels, 1998). The watershed was predominately second-growth forest and agricultural fields (Zembrzuski and Evans, 1989). In the past, milldams were located upstream from Burtonsville at a nick point in the stream (Lindner, 1991). Today there were no milldams and the river channel adjacent to Burtonsville was characterized as depositional, with imbricated cobbles along the exposed till banks.

The climate in this region was temperate (Daniels, 1998; McIntosh, 1972). Temperature and precipitation along the Schoharie are variable due to the Catskill Mountains (McIntosh, 1972). Annual temperatures average around 10°C, which varies relative to latitude and altitude.
across the watershed (Zembrzuski and Evans, 1989). Annual precipitation ranges between 900 to 1500 mm across the basin (Godwin et al., 2003; McIntosh, 1972). Most precipitation falls as rainfall in the spring and fall (Mehaffey et al., 2005). Spring snowmelt and summer rainfall events are the major contributors to runoff in this area, specifically large convective storms and hurricanes in summer and early fall (Cockburn and Garver, 2014; McIntosh, 1972). Runoff throughout the study site was approximately 500 mm/year with most runoff occurring in spring (Cockburn and Garver, 2014; Zembrzuski and Evans, 1989).

The surrounding Catskill Mountains consist of Devonian-age sandstones and shales underlain by Silurian shale, sandstone, and limestone (Zembrzuski and Evans, 1989). The glacial legacy for this area plays a role in surficial geology and sediment supply into the Schoharie Creek (Addy et al., 2014; McIntosh, 1972). Pleistocene glaciers covered the entire Catskill region during the last glacial period and are described in detail by Ridge (2003) and Rich (1935). The last glacial period resulted in the till deposits in the valleys and stony glacial till that cover the surrounding Catskill area where this study occurred (McIntosh, 1972; Rich, 1935). Specifically focused on the study site the substrate in this area was predominately unconsolidated till deposited during the last glacial period, (Addy et al., 2014; Lindner, 1991). The bank slope was susceptible to gullying and slope failures along Schoharie Creek since it was comprised of unconsolidated till (Lindner, 1991).

During early settlement along the Schoharie, hardwoods were the predominant trees in forest-covered areas (McIntosh, 1972). Nearly half (49.5%) the tree density was comprised of beech (*Fagus grandifolia*) with eastern hemlock accounting for 20.3% and sugar maple (*Acer saccharum*) for 12.8% (McIntosh, 1972). In the early 19th century, harvesting most hemlocks was popular for tanbark since the Catskills were the center of a tanning industry (McIntosh,
Due to the tanning industry and failure to reproduce effectively in secondary hardwoods forests, hemlock presence has declined (McIntosh, 1972). Most recent the eastern hemlock were predominantly found in valley bottoms, along streams, and on north facing slopes despite the extensive deforestation (Driese et al., 2004; McIntosh, 1972). Other trees present in valley forests include northern red oak (*Quercus rubra*), chestnut oak (*Quercus prinus*), and red maple (*Acer rubrum*; Driese et al., 2004; McIntosh, 1972). The forest cover at the study site included beech, white pine (*Pinus strobus*) sugar maple, and eastern hemlock. This current study focuses on eastern hemlocks since they were a coniferous tree with distinct rings as well as they measured reaction wood presence on the trunk underside (Timell, 1986). The eastern hemlocks at the study area were on the southern half, which was why this research focused on the south half and not the northern half of the slide.

The study area spans approximately 1250 m in length and 300 m wide. The relief along this slope ranges from approximately 150 m at the bank edge to 230 m asl at the furthest point upslope. This study focused on the southernmost section of the slope, in an area approximately 400 m long by 200 m parallel to Schoharie Creek due to the abundant eastern hemlock and clear evidence of slope instability (Figure 3.1). The slope had hummocky topography with multiple scarps, cracks, and many tilted trees, exposed roots, and loose soil throughout the entire area (Figure 3.2; 3.3; Figure A.1; A.2). This site was predominately forest covered with a few areas that have tall grasses with little to no trees (Figure A.4). There are multiple steep slope units from west to east across the study area. There was also a dirt ATV path ~70 m from the river that cuts through the south slope (Figure A.3).
Figure 3.1: Location of the study site relative to eastern North America (inset). Hillshade imagery for the Burtonsville bank slope extent. Each sampled tree is denoted by a red square. Focus for this research is on the southern slope half.
Figure 3.2: A: Stem deformations with head scarp highlighted in white for a slope face near the bank edge. B: Located east of image A with the white line identifying the same scarp from image A. This image shows evidence of reverse block tilting with pooled water highlighted by the blue line.
3.4 Methods

3.4.1 Field Surveying and Sampling

The study site was sub-divided into six transects ranging from 90 m to 200 m in length. Each transect was further subdivided into slope units, which represented a change in slope angle and aspect. The equipment used to survey each transect included a measuring tape, a GPS, a laser rangefinder, and a reflective target to maintain a consistent reference point for the laser rangefinder. Each transect was measured from downslope to upslope and upstream to downstream starting at the river bank edge and ending at the furthest upslope point (Figure 3.2). Surveying each slope unit and transect determine qualitative and quantitative descriptions for the entire study site.

Tree cores collected from eastern hemlock located within 10 m to the north and south along each transect. In cases where there were multiple eastern hemlocks on or within 10 m of each transect for a specific slope unit then the largest trees and presumably the oldest trees were selected for this study (Stoffel and Bollschweiler, 2008). The sampled trees had two cores extracted with an increment borer. One core sampled from the upslope side and the other core sample was from the downslope side to examine the growth anomalies.

3.4.2 Dendrogeomorphological Methods

Tree core samples were processed as described in Stoffel and Bollschweiler (2008). The reaction wood presence or absence was visually identified and recorded in addition to the ring width for a given year (Stoffel and Bollschweiler, 2008; Wistuba et al., 2013). The ring width time series for upslope and downslope cores were transformed into absolute eccentricity indicators to account for both upslope and downslope eccentricity (Equation 3.1):

\[ E_x = |D_x - U_x| \]  

(3.1)
where: D was the downslope or west facing stem side (mm); U was the upslope or east facing stem side (mm); E was the absolute eccentricity growth (mm); x was the annual ring growth for each individual tree. The absolute growth differences provided values greater than zero, with larger values represented as eccentric growth and smaller values represented as concentric growth for each year. In every tree, regardless of stem tilting direction, the upslope ring widths were subtracted from the downslope ring widths and the difference converted to its absolute value. Using the absolute differences enabled comparisons for the differences, regardless of tilt direction across the entire bank slope. This study also identified reaction wood in the upslope and downslope direction. The rings for affected tilted trees appear larger and darker in colour (Schweingruber, 2007; Stoffel et al., 2010; Timell, 1986). The difference in cell colour related to the thicker and rounded cell walls densely packed together (Stoffel et al., 2010; Timell, 1986). Therefore, identifying reaction wood focused on the densely packed and subsequent dark in colour rings.

3.5 Results

3.5.1 Field Analysis

Field observations and the surveyed transects identified multiple steep slope units across the entire south bank slope. The bank edge was steep with exposed banks adjacent to Schoharie Creek (Figure 3.2; 3.3). Slope angles observed at all transects nearest the slope edge range from 21.3° to 34.3°, and were approximately 20 m in length for each sampled transect (Figure 3.2; Figure 3.3). A second steep slope observed in transects C to E, occurred approximately 25 m from the bank edge with angles ranging from 17.8° to 26.3° and lengths ranging from 20 to 40 m (Figure 3.3). The steep slopes observed on transects C to E were separated by a change in slope angle and aspect suggesting there was evidence for reverse block tilting (Figure 3.2; 3.3).
Furthermore, the slopes 20 to 40 m from the bank edge show visible slope instability such as tilted trees, 1 to 2 m scarps, tension cracks, and exposed roots (Figure 3.2; 3.3). The steep slope on transects C to E was not similar to the slopes on transects A to B and F (Figure 3.3).

There were multiple steep slopes located on each transect upslope beyond 70 m from the bank edge (Figure 3.3; Figure A.5 and A.6). At transects C and E, there were a few steep slopes measuring approximately 20 to 40 m in length with slope angles ranging from 12.8° to 22.9° beyond 70 m from the bank edge (Figure 3.3). Transect B also had two steep slopes with inclinations of 18.6° and 27.5° at 90 and 120 m from the bank edge (Figure 3.3; Figure A.5 and A.6). Transect A had only one steep slope farthest from the bank edge starting at 112 m with slope inclination of 30° (Figure 3.3). The remaining slope units had changes in slope inclinations and alternate in slope aspect from east to west facing.

A common feature at these reverse block tilts was the change in slope aspect and ponded water (Figure 3.2). The surveyed transects indicated that reverse block tilting occurred across the slope. The first area on the slope closest to the Schoharie where this occurred was approximately 20 m from the bank edge on transects C, D, and E (Figure 3.3). Further upslope on transects C to E, at approximately 50 m from the bank edge there was again of reverse block tilting (Figure 3.3). The reverse block tilting was not limited to transects C, D, and E as evidence for these features occurred on transects A and B. Specifically at 30 m to 40 m from the bank edge on transects A and B there was a change in slope angle and aspect where the reverse block tilting was observed (Figure 3.3).
Figure 3.3: Cross-sections for surveyed transects. The elevation of the Schoharie adjacent to the slope is approximately 150 m asl.
Slope facies, were identified from the 1 m LiDAR derived Digital Terrain Model (DTM; Figure 3.4). The slope facies are the areal distributions for the topography at the site defined by features such as slope angle and/or slope aspect, identified from the 1 m LiDAR derived DTM to understand the variability in slope topography. The mapped slope facies identified that this slope was not one continuous face, but instead there were multiple slope facies across each transect, which resembled step blocks with reverse tilt blocking occurring in all transects (Figure 3.3; 3.4). The mapping reveals multiple slope facies across the entire south slope side and were parallel to the Schoharie Creek (Figure 3.4). Aerial photographs also mapped changes to the medial bar in the Schoharie Creek (Figure 3.5). This study used eight aerial photographs (May 7, 1999; March 31, 2001; March 31, 2004; April 30, 2007; May 23, 2008; May 3, 2009; May 26, 2011; October 7, 2011) to map the medial bar location and covered low flow, high flow, pre-hurricane Irene, and post-hurricane Irene conditions (Figure 3.5). Each aerial photograph was georeferenced, digitized, and mapped on the DTM to show the size and location for the medial bar relative to the flow conditions at the time of each photo (Figure 3.5). Two major storm events that occurred in this area were hurricane Irene and tropical storm Lee (Lumia et al., 2014). Hurricane Irene tracked through the New York state area from August 28, 2011 to August 29, 2011 (Lumia et al., 2014). The areas surrounding the Catskill Mountains had storm rainfall exceed 45 cm in a 24-hour period (Lumia et al., 2014). Although less severe than hurricane Irene, tropical storm Lee hit New York State from September 7 to 11, 2011, which was less than weeks after hurricane Irene (Lumia et al., 2014). The approximate total rainfall observed in south-central New York in August and September 2011 was over 50 cm, which was the greatest rainfall on record for this two-month period (Lumia et al., 2014). The aerial photographs included images for the medial bar before and after these two events occurred (Figure 3.5).
Figure 3.4: Slope facies at the study site. Each square represents a sampled tree. The head scarps for each slope face is mapped with downslope direction indicated.
Figure 3.5: The medial bar location digitized from aerial photographs. A: the island bar size and location during high flow conditions. B: The medial bar size and location during low flow conditions.
3.5.2 Dendrogeomorphology Analysis

The oldest tree sampled dated back to 1731, and by 1902 there were 57 live trees or ~51% of all the trees sampled; while the shortest time series began in 1976. The oldest trees were along the central transects (transects C, D, and E, Figure 3.1) with the youngest trees in the northern and southern extents in the study area (Transects A and F, Figure 3.1). From the tree ring records, we were able to identify the absolute growth differences and reaction wood presence for all sampled cores (Figure 3.6). Through observation and comparisons between trees and the average absolute differences in ring widths, it was determined that an absolute difference in ring widths of 0.675 mm represented a minimum threshold for slope instability (Figure 3.6B). Years with differences greater 0.675 mm indicated substantial differences in the upslope and downslope growth patterns in that tree, for that year. This research identified reaction wood as either present or absent from the sampled cores (Figure 3.6C). Years with large absolute eccentric ring widths (> 0.675 mm) typically coincided with reaction wood presence and years with small (< 0.675 mm) absolute eccentricity values often had no reaction wood present (Figure 3.6C). Years with large ring width differences or eccentric growth and reaction wood presence combined to represent years with slope instability (Figure 3.6D). From the 111 trees sampled, 105 trees (95%) had at least one year with reaction wood present and absolute eccentric growth above the 0.675 mm threshold.

The dating for instability events for each tree were based on two criteria modified from Wistuba et al., (2013). Firstly, the absolute eccentric growth must have been above the minimum difference threshold set at 0.675 mm. Secondly, the years with absolute eccentric growth above the threshold must have also contained reaction wood (Figures 3.6B and 3.6C). Previous dendrogeomorphology studies (Saez et al., 2012; Silhan et al., 2014; Stefanini, 2004; Stoffel et
al., 2013; Wistuba et al., 2013) focused on dating single, large disturbances in tree growth (e.g., landslides). The modifications presented in this study evaluated slower or cumulative processes. The two set criteria identified the unstable years from each core sampled (Figure 3.6D). The two interpretations inferred from the absolute growth eccentricity and reaction wood presence were the frequency and magnitude associated with each slope instability event (Figure 3.6D; 3.7; 3.8). The analysis for slope instability frequency and magnitude were on an annual basis with emphasis on years where at least 50% of the sampled trees are living (Stoffel et al., 2013).

In order to evaluate the spatial patterns in slope instability across the study site, this research used the F-distribution to compare the absolute growth differences for all years. The F distribution for absolute growth differences for all 111 sampled trees resulted in a 29.29 calculated F statistic, which was greater than the critical F value of 1.37. This indicated that the variances in absolute growth differences between every sampled tree were significantly different and that there were no common or clustered areas within the sampled trees. Therefore, the comparisons between trees were relative to their position from the bank edge (Figure 3.7).
Figure 3.6: A: The upslope and downslope tree ring widths for a single tree. B: The absolute difference between upslope and downslope tree ring widths, this becomes growth eccentricity. The horizontal line indicates the minimum difference threshold of 0.675 mm. Values exceeding the threshold are shaded. C: Years where reaction wood is present. D: Years with exceeded absolute eccentricity values and reaction wood presence, and therefore inferred instability.
Figure 3.7: Instability periods at each sampled tree. Black squares represent a year with instability. Each tree's distance from the bank edge in 2013 was represented on the y-axis and for the purpose of this figure remains constant. The grey bar indicates tree age, while white space represents no trees alive in that area at that time. Appendix B details the number and age of each tree sampled.
Figure 3.8: Ring widths greater than 0.675 mm at the lower and upper transect regions. The black squares indicate years where reaction wood occurred. The percent of trees alive in each transect is represented by the black line. White space in years before the black line represents no trees alive during that time at that specific site. The grey shaded area under the black line represents no instability observed in trees at that time for that specific site.
This research also compared the total annual precipitation and discharge records with the slope instability events. Previous studies had demonstrated strong relationships between slope instability, total precipitation, and change in slope geometry due to undermining of the slope toe by rivers (e.g., Corominas and Moya, 1999; Silhan et al., 2013; Stefanini, 2004; Wistuba et al., 2013). This study compared the slope instability frequencies with the precipitation and discharge records from nearby stations to compare the affects that changes in moisture and changes in the slope angle from toe erosion may have on slope instability (Figure 3.9). Specifically, the comparison between slope instability with the precipitation and discharge records focused on instability frequency and location across the slope. The monthly precipitation data was from Albany, NY, located approximately 50 km east of Burtonsville provided an approximate 160-year record for precipitation to compare with slope instability. The discharge was recorded from the USGS gauge station at Burtonsville, NY (latitude 42°48'00" and longitude 74°15'46" NAD83) located less than 1 km upstream from the study site and included monthly data from 1940 to 2012. The results from the precipitation and discharge comparisons with the instability frequencies focused on understanding if slope instability frequency and magnitude patterns coincide with changes in stream power, which may possibly erode the slope toe.
Figure 3.9: Total calendar annual precipitation and discharge with the five year average for both compared to the frequency plots for transects A to E.
3.6 Discussion

3.6.1 Slope Instability Frequency and Magnitude

In this study, the mapping analysis examined the complex slope topography along the southern slope. The topography observed at the south bank slope was steep with exposed banks and fallen trees near the bank edge. The exposed tree roots, cracks, scarps, and abrupt changes in slope aspect suggested that the southern bank slope was susceptible to slope instability (Figure 3.2; 3.3; Figure A.2). The surveys and LiDAR imagery revealed that slope instability at this site was quite complex (Figure 3.3; 3.4; 3.5). It was difficult to classify the entire slope as a single distinct mass wasting process as it was a complex combination of translational and rotational landslides as well soil creep (e.g., Dikau et al., 1996; Migori et al., 2010). Surface features identified the complex processes acting on the slope included stepped slope profiles with back tilted blocks and cracks with pooled water (Figure 3.2; 3.3).

Additional morphological features included hummocky topography, exposed head scarps, reverse block tilting, and steep slopes with inclinations greater than 20° throughout the study area, which were characteristic features in rotational and translational slides (Buma and van Asch, 1996; Migori et al., 2010). Based on the LiDAR derived imagery, the steep slope facies resemble curved or listric faults that resembled a concave shape (Ofegbu and Ferrill, 1998; Williams and Vann, 1987; Figure 3.3). It appeared that the faulting near the bank edge for most transects relate to the unconsolidated material and the combination between deep-seated and shallow landsliding, which were likely triggered from changes in moisture as described by Stefanini (2004). Particularly, the cracks observed within 50 m from the bank edge likely feed water to the slip plane. This suggested that this area may remain unstable as water continues to act as a lubricant for the potentially deep-seated failure (Saez et al., 2012; Figure 3.2; 3.3).
The spatial variability in slope instability between upslope and downslope ring widths identified a disconnect in slope instability events between the bank edge and upslope areas. Instability varied dependent on the slope location; with little evidence, to suggest that instability at one area drove instability in another (e.g., Stefanini, 2004; Figure 3.7; Appendix B). Since the absolute growth differences for all sampled trees had such variability across the slope, this suggested that a single movement slide or fault line does not dominate the entire slope but instead there are multiple blocks, which move at different times and rates (e.g., Stefanini, 2004; Figure 3.7; Appendix B). For example, trees located near the bank edge on transects A and B experienced many years with slope instability, especially since the late, which were not observed in the trees furthest upslope on these two transects during the same period (Figure 3.7). Furthermore, there was no evidence to suggest that the trees furthest downslope on transects C, D, and E experienced instability during or before slope instability events in the trees furthest upslope (Figure 3.7; Appendix B).

The slope instability records reveal that instability was most frequent at transects A to E near the bank edge over approximately the last 60 years with more continuous instability frequency occurred in the late 1960s to the 1970s, the late 1980s, and the mid-2000s (Figure 3.7; Appendix B). This implied that conditions acting on the slope during these years heavily affected slope instability across the southern slope half, which followed similar findings in studies by Stefanini (2004) and Silhan et al. (2014). One similarity observed between most transects was that instability frequency was episodic before 1969 with distinct periods where slope instability was prevalent followed by years with stable or very little observed instability (Figure 3.7; Appendix B). For instance, interpreted absolute eccentric growth records for transects C to E near the bank edge in the early 1900s were followed by nearly forty years with stable growth
until the 1940s when at least one instability event was recorded in each decade until the record’s end (Figure 3.7; Appendix B). For the trees near the bank edge, slope instability frequency was more continuous rather than episodic starting in 1969 to present. The more continuous frequency for instability events started in 1969 appeared to shift from episodic instability frequency in the earliest records (Figure 3.7). However, the more continuous slope instability frequency observed for the trees near the bank edge were not similar for the trees furthest upslope (Figure 3.7; Appendix B). The trees furthest upslope revealed that the instability observed was less frequent compared to trees near the bank edge (Figure 3.7). The dynamics in slope instability frequency between the trees located furthest downslope and upslope do not coincide over time. The variability in slope instability frequency within transects for this study was similar to previous studies by Bollschweiler et al., (2007), Migori et al., 2010, Silhan et al., (2014), and Stefanini (2004) who found that the disconnection in slope instability events was driven by the complex slope dynamics.

The overall patterns in slope instability frequency revealed that there was an apparent shift from episodic instability pre 1969 to more continuous instability starting in 1969 to present. This shift from episodic to more continuous instability frequency was best observed for trees near the bank edge, at approximately the 20 to 40 m from the bank edge slope units on transects A through E (Figure 3.7). The recent years appear to have more continuous instability frequency, where most years from 1969 to present record instability events for most trees at the 20 to 40 m from the bank edge distance in transects A to E (Figure 3.7). The change to more continuous instability frequency from episodic instability frequency did not occur for trees located furthest upslope. The shift in frequency suggested that potential external factors or threshold events might have exceeded, causing the shift towards more continuous instability frequency.
The magnitude values were determined from both the absolute difference between the downslope and upslope ring widths and reaction wood presence for each sampled (Figure 3.8). The slope instability magnitude indicated the degree of stem tilting stress for each tree and therefore severity in slope instability (Figure 3.8). The magnitude values for trees near the bank edge on transect A through E remained consistent over time (Figure 3.8). In the early 1900s until 1920 there were few large magnitude values for the entire record, especially on transects C to E (Figure 3.8). Trees located farthest upslope on transects A and B had very few slope instability frequency and magnitude values for their entire records. The trees on transects C through E and trees near the bank edge on transects A and B had consistently high magnitude values throughout their entire record. The one major change in slope instability magnitude observed on the slope was that the frequency for these events had become more continuous rather than episodic starting in approximately 1969. Therefore, although these magnitude events did not increase in size, they occurred more frequently and affected the slope units near the bank edge.

The shift from episodic to more continuous slope instability frequency near the bank edge starting in 1969 suggests there may be a change in equilibrium for this site to a potential alternate steady state. Furthermore, the slope instability magnitude from 1969 to present were not impacted by external forces since the magnitude values remained consistent over time; instead, the change to more continuous instability frequency implied that frequency was most likely impacted by external factors such as precipitation, discharge, and medial bar location. Therefore, the slope instability frequency and magnitude interpretations revealed that instability at this site has shifted to more continuous frequency but the magnitude for the instability events has remained as large as the events from before 1969.
3.6.2 Precipitation and stream flow controls on slope instability

The slope instability reconstruction revealed that slope failure across the entire site shifted from episodic instability frequency to more continuous instability frequency from 1969 to present. Combining all instability events into a single table or graph did not address the role in which added moisture from precipitation affected instability. A comparison between instability frequency and both precipitation and discharge was necessary to assess any coincidences between patterns in slope instability frequency and changes in moisture conditions (Saez et al., 2012; Stefanini, 2004). The instability observed on transects C and E beyond 70 m from the bank edge, occurred under dry conditions from 1907 to 1916 (Figure 3.9). The instability at this time had some of the driest conditions during and before these years for the entire precipitation record (Figure 3.9). This suggested that instability observed in the trees at C and E furthest upslope were more sensitive to drier conditions, similar to findings by Saez et al., (2012). However, most recent instability events observed furthest upslope occurred in years with high precipitation (Figure 3.9). The constant magnitude values for the trees upslope during entire record revealed that the trees furthest upslope are recently more sensitive to wet conditions rather than the drier conditions observed in the early 1900s (Silhan et al., 2014; Wistuba et al., 2013).

The comparison between precipitation and slope instability for trees near the bank edge revealed that the shift to more continuous instability frequency from 1969 to present occurred during a shift from low precipitation in the 1960s to consistently high precipitation from 1969 to present (Figure 3.9). The comparison also revealed that the magnitude for instability in most recent years was similar to those before 1969, which suggested that slope instability frequency rather than magnitude was more sensitive to changes in precipitation (Figure 3.7; 3.8; 3.9). This
suggested that the slopes near the bank edge experienced more continuous instability frequency during continuously high moisture conditions (Stefanini, 2004; Wistuba et al., 2013; Figure 3.9).

The comparison between discharge and instability frequency focused primarily on the trees located near the bank edge as slope toe undercutting from high stream flow was a possible trigger for instability (Stefanini, 2004). Although there was a coincidence between high precipitation and high discharge, both processes remained separately compared since precipitation addressed moisture availability to the slope, while discharge examined the stream power available for slope toe removal. Furthermore, there were instances where high discharge did not directly coincide with high precipitation. For example, in 1996 the second highest annual total discharge occurs but had moderate total annual precipitation compared to other years (Figure 3.9). This particular year was also one that did not have high instability frequencies for the slope facies near the bank edge (Figure 3.7). Therefore, from the 1996 example, we see that high discharges did not always coincide with high precipitation totals, suggesting the need to examine each process separately when comparing them to instability frequencies.

Similar to the precipitation records, there was an apparent change in the discharge record from low-recorded discharge in the 1960s to higher discharge recorded from 1969 to present (Figure 3.9). More continuous instability frequency near the bank edge associated with the discharge records may relate to the medial bar, which was located near the slope toe between transects A and E (Figure 3.5; Figure 3.9). The medial bar directs Schoharie Creek into the slope toe (Figure 3.5). The digitized medial bar revealed changes in area during high and low flow conditions as well as the fact that the medial bar’s position might have changed after Hurricane Irene and Tropical Storm Lee events from 2011 (Figure 3.5). The digitized medial bar shows changes in position, area, and width for varying flow conditions (Figure 3.5). In periods with low
flow conditions, the medial bar was much larger than during high flow conditions, with water directed toward the slope during both conditions (Figure 3.5). This suggested that during high flow conditions, the medial bar appears to direct most of the water towards the bank edge, which may account for the high instability frequencies observed for the trees near the bank edge (Figure 3.7; Appendix B). Since the events from 2011, it appeared that the channel has migrated further downstream (Figure 3.5). The medial bar migration might now lead to instability observed further upslope for areas near transect F, which in the past appeared stable compared to the transects located further south. These findings are similar to past studies on slope instability using tree ring records by Fantucci and Sorriso Valvo (1999), Saez et al., (2012), Silhan et al., (2014), Stefanini (2004), and Wistuba et al., (2013).

Further comparison between slope instability frequency and magnitude with the precipitation and discharge records revealed that the shift from episodic to more continuous instability frequency that started in approximately 1969 appeared to coincide with a change from very low precipitation and discharge in the 1960s to high and continuously wet conditions from 1969 to present. The change in precipitation and discharge at the site starting in 1969 to present are two possible external factors that potentially explain the shift to more continuous instability frequency during the past 40 years. Furthermore, the increased discharge at the site also played a factor in directing more water towards the slope due to the medial bar located adjacent to transects A to D. If the discharge at the site remains high, then there may be a greater chance for more water directed towards the slope by the medial bar, which then may lead to increased slope toe erosion and may explain the more continuous instability frequency for trees near the bank edge.
3.7 Conclusion

Analysis from this research focused on slope dynamics pertaining to instability frequency and magnitude variability on a bank slope along Schoharie Creek. The field surveys conducted revealed that this slope was extremely complex with numerous steep slope facies. The complexity in the slope dynamics proved difficult to attribute a single mass movement to the entire slope, as it appears to be a combination between rotational and translational landslides and soil creep. The complexity in slope movement was confirmed from the interpreted absolute growth differences and reaction wood presence, which determined the frequency and magnitude for slope instability. The frequency data identified instability across the entire site as disconnected in both the upslope to downslope direction and the upstream to downstream direction for the entire slope. In addition, the tree ring records revealed that instability was episodic but after 1969, there was a shift to more continuous instability frequency. This suggested that there were multiple slope facies, which initiate slope instability at different periods across the entire site. Furthermore, the slope instability magnitude revealed that slope instability magnitude values remain consistently high for the entire record. Therefore, although slope instability frequency is becoming more frequency, the magnitude associated with these events remains consistently large.

This research also compared the frequency and magnitude results from the identified instability events with the precipitation and discharge records for the Schoharie region to determine any coincidences between instability and the hydrometeorological processes. Evidence from the medial bar mapping revealed that this feature may play a role in directing the water towards the bank edge, thus increasing the stream power eroding and destabilizing the slope bank. The shift to more continuous instability frequency near the bank edge occurred in years
with continuously high moisture. The precipitation and discharge records coincided with the apparent shift in slope instability frequency starting in 1969 to present. The shift in 1969 recorded a change from low precipitation and discharge in the 1960s to consistently high precipitation and discharge from 1969 to present. The coincidence between increases in precipitation and discharge with more continuous instability frequency suggested that an exceeded threshold occurred in 1969 where there was a shift to an alternate steady state on the site resulting in continuous instability frequency and similar magnitude values for the instability events.

Therefore, this research provided a means to interpret slope instability from tree ring growth differences and reaction wood for over a century of data. In addition, this research provided an example for understanding a potential alternate steady state occurring on a landscape system. Instability recreated from each tree as described by this research may help to recreate patterns in slope instability frequency and magnitude, which when compared to external factors such as precipitation and discharge may aid in understanding the role that changes in moisture may have in initializing slope instability. Future research at this site may continue to monitor instability frequency and magnitude in order to determine if another threshold may have exceeded following the two major storms in 2011. If a threshold was exceeded in 2011, it may have resulted in another alternate steady state equilibrium for this landscape system. This may provide future concerns at this site as patterns in instability frequency and magnitude may become altered compared to those identified in this research. The change to an alternate steady state equilibrium might result in areas observed as stable in this research, such as transect F and areas furthest upslope, to become more unstable in the future.
Chapter 4: Conclusion

This thesis examined the slope dynamics on a riverine bank slope in upstate New York. The frequency and magnitude of slope failure events over a 100-year period were evaluated from eccentric tree ring growth and reaction wood presence. This work compared observed instability events with both precipitation and discharge records to link moisture conditions along the slope during years with instability. The study and findings have important implications for future research on slope instability. The major results, discussion points, and contributions are summarized below:

- Slope instability frequency across the study area was variable and disconnected in the upslope to downslope and upstream to downstream directions. More continuous instability was observed for the trees located near the bank edge, approximately 0 to 70 m from the bank edge, for transects A to E. When comparing the frequencies between trees near the bank edge to those furthest upslope on all transects there were no connections to suggest that instability near the bank edge does not lead or lag instability furthest upslope. Furthermore, there were few unstable years observed furthest upslope on all transects. Comparing the instability frequencies in the upstream to downstream direction revealed similar years with instability for the trees near the bank edge and few similarities in slope instability frequency for the trees located beyond 70 m from the bank edge.

- This research provides an alternate approach to examine slope instability frequency by comparing absolute growth eccentricity and reaction wood presence. Identifying instability events by using both growth anomalies built upon past studies by Silhan et al., (2014) and Wistuba et al., (2013) who date instability frequency using only growth eccentricity. Furthermore, the magnitude values interpreted from the absolute growth
eccentricity provided an alternate method not commonly seen in past dendrogeomorphic studies to assess the instability events (e.g., Silhan et al., 2014; Stefanini, 2004; Wistuba et al., 2013). The frequency and magnitude data provided a means to examine potential shifts in equilibrium for this unstable bank slope. The more continuous instability frequency and consistently large magnitude values observed near the bank slope in most recent years suggested that the slope was unstable and had not yet reached a steady state. Furthermore, from 1969 to present the frequencies appeared to shift from episodic instability before 1969 to instability frequency that is more continuous from 1969 to present. The changes observed in the frequency records revealed that external processes likely affected slope instability frequency since there was the shift from episodic to instability frequency that is more continuous in 1969. Given that the slope units near the bank edge appeared to be most unstable, there was a higher likelihood for sediment deposition into the Schoharie, which might have affected aquatic life and damage reservoirs downslope.

- The magnitude values calculated from the absolute growth eccentricity for each tree was proportional to the stress induced due to instability. The patterns observed in this study for magnitude values revealed that trees located furthest upslope on transect A and B had low magnitude for slope instability over the entire record. The magnitude values for trees on transect C through E and trees near the bank edge on transect A and B remained consistently high throughout their entire record. This suggested that although the instability magnitudes did not change over time, slope instability frequency became more continuous from 1969 to present. Therefore, slope instability frequency is more
continuous than years before 1969, while the magnitude for events after 1969 remain consistently large compared to the past instability events.

- The precipitation and discharge records compared with the instability events revealed that high moisture conditions are common with instability events. Although precipitation and discharge are connected processes, analysis on both records revealed that high total yearly discharge did not always coincide with above average total yearly precipitation totals. Therefore, this work compared the precipitation and discharge records separately with the observed slope instability frequency and magnitude. The analysis that compared the precipitation and discharge records with slope instability frequency and magnitude revealed a shift from episodic instability frequency to more continuous instability frequency starting in 1969. The apparent shift to more continuous instability frequency appeared to coincide with a change from very low moisture conditions in the 1960s to consistently high precipitation and discharge after 1969. Therefore, the comparison between instability and changes in moisture availability revealed that an apparent threshold might have been exceeded at this landscape system, which resulted in an alternate steady state equilibrium at this site. It appeared that high precipitation and discharge totals are two external factors that potentially have led to the continuous instability frequency observed for this area while magnitude values for slope instability remain consistently large. Furthermore, with consistently high discharge records, it appeared that from the medial bar mapping more water was available to direct towards the slope, which may generate greater stream power available to erode the slope toe. The directed water towards the slope toe may be an external factor explaining why instability
was most frequent for the slope units near the bank edge, particularly at the 20 to 40 m from the bank edge location.

The knowledge gained from this study provided an alternate method to study slope instability using tree ring records as well as potentially identified a shift towards an alternate steady state equilibrium for this landscape system. This research identified multiple areas where current knowledge needs enhancement by future researchers for slope instability studies. Firstly, this research built upon past dendrogeomorphological methods used to examine and date slope instability events. Specifically, this work focused on the comparison between reaction wood presence and absolute growth eccentricity above a set threshold in order to identify instability frequency. This built on past research that predominately aimed to identify instability initialization and less emphasis on the years following the initialization event. Examining the frequency and magnitude for instability across the entire slope provides insight into understanding if current events are lasting for longer or shorter durations compared to past-observed instability. Secondly, this research demonstrated a way to interpret the magnitude for instability events for each tree related once again to the absolute eccentric growth values and reaction wood presence. Few studies that interpret slope instability events from tree ring growth anomalies examined the magnitude values from the growth eccentricity. The magnitude values from the interpreted absolute growth eccentricity provided an idea on the size associated with instability events. The magnitude and frequency records built upon the ideas from Wolman and Miller (1960) tested if large magnitude but infrequent events play a greater role in affecting geomorphic systems. This work also found that areas where instability frequency became more continuous over time had magnitude values remain consistently large. This suggested that the instability frequency at this site was more continuous and remained as large compared to
earlier years where instability frequency was more episodic. Finally, this study compared the moisture conditions on the slope with observed instability frequencies. This study demonstrated that the more continuous instability frequency coincided with a shift from low to consistently high precipitation and discharge records as demonstrated in numerous past studies (e.g. Fantucci and Sorriso Valvo, 1999; Saez et al., 2012; Silhan et al., 2014; Stefanini, 2004; and Wistuba et al., 2013). It was believed that the medial bar location plays a major role in directing water towards the slope toe, which generates higher stream power available to erode the slope toe. If the discharge records on the Schoharie remain consistently high or increase in the future, there may be the potential that instability frequency may remain continuous at the site, especially near the bank edge. Furthermore, if the medial bar migrates further upstream because of the two major storms observed in 2011, then areas further upstream which appeared to be more stable, particularly transect F, may have more frequent instability in future years. Also, the two major storm events from 2011 may have resulted in another threshold exceeded at the site resulting in an alternate steady state, which future research monitoring this potential shift at the site will provide beneficial in understanding how instability frequency and magnitude may have changed across the slope following 2011.
References


Appendix A

This appendix provides images taken across the study area, focusing on the southern slope half. Each image is referenced throughout the chapter and provides a visual representation of the features discussed throughout this study.

Figure A.1: Slope face near the bank edge for transect A (A) and transect B (B). Both images are taken from the slope nearest the bank edge with A facing towards the Schoharie and B taken parallel to the slope.
Figure A.2: The head scarp located on transects C, D, and E approximately 25 to 30 m from the bank edge. The view is looking away from the Schoharie into the slope.
Figure A.3: The road cutting through Transect B. This road is found across the entire southern slope.
Figure A.4: An area located on transect E with few trees and instead mainly consists of small vegetation. This image is located upslope of the road that cuts through the entire study site.
Figure A.5: Furthest steep slope on transect A. This image is taken looking upslope away from the creek.
Figure A.6: A and B: Located on transect C approximately 80 and 140 m from the bank edge respectively. These are the two steepest slope blocks furthest upslope on transect C.
Appendix B

This appendix provides the percent of slope instability frequency for all sampled trees. The frequency distributions were divided based on the tree’s location on the slope with the percentages of trees showing instability for a particular zone.
Figure B.1: Slope instability frequency percentages for all sampled trees. The percent of instability is a factor of how many trees are recording instability to how many trees are alive for each year at each site.