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Basal-topographic control of stationary ponds on a continuously moving landslide†

Jeffrey A. Coe,* Jonathan P. McKenna, Jonathan W. Godt and Rex L. Baum
US Geological Survey, Denver Federal Center, Denver, CO, USA

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* Correspondence to: Jeffrey A. Coe, US Geological Survey, Denver Federal Center, MS 966, Denver, CO 80225, USA. E-mail: jcoe@usgs.gov

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ABSTRACT: The Slumgullion landslide in the San Juan Mountains of southwestern Colorado has been moving for at least the last few hundred years and has multiple ponds on its surface. We have studied eight ponds during 30 trips to the landslide between July 1998 and July 2007. During each trip, we have made observations on the variability in pond locations and water levels, taken ground-based photographs to document pond water with respect to moving landslide material and vegetation, conducted Global Positioning System surveys of the elevations of water levels and mapped pond sediments on the landslide surface. Additionally, we have used stereo aerial photographs taken in October 1939, October 1940 and July 2000 to measure topographic profiles of the eight pond locations, as well as a longitudinal profile along the approximate centerline of the landslide, to examine topographic changes over a 60- to 61-year period of time.

Results from field observations, analyses of photographs, mapping and measurements indicate that all pond locations have remained spatially stationary for 60–300 years while landslide material moves through these locations. Water levels during the observation period were sensitive to changes in the local, spring-fed, stream network, and to periodic filling of pond locations by sediment from floods, hyperconcentrated flows, mud flows and debris flows. For pond locations to remain stationary, the locations must mimic depressions along the basal surface of the landslide. The existence of such depressions indicates that the topography of the basal landslide surface is irregular. These results suggest that, for translational landslides that have moved distances larger than the dimensions of the largest basal topographic irregularities (about 200 m at Slumgullion), landslide surface morphology can be used as a guide to the morphology of the basal slip surface. Because basal slip surface morphology can affect landslide stability, kinematic models and stability analyses of translational landslides should attempt to incorporate irregular basal surface topography. Additional implications for moving landslides where basal topography controls surface morphology include the following: dateable sediments or organic material from basal layers of stationary ponds will yield ages that are younger than the date of landslide initiation, and it is probable that other landslide surface features such as faults, streams, springs and sinks are also controlled by basal topography.

The longitudinal topographic profile indicated that the upper part of the Slumgullion landslide was depleted at a mean vertical lowering rate of 5.6 cm/yr between 1939 and 2000, while the toe advanced at an average rate of 1.5 m/yr during the same period. Therefore, during this 61-year period, neither the depletion of material at the head of the landslide nor continued growth of the landslide toe has decreased the overall movement rate of the landslide. Continued depletion of the upper part of the landslide, and growth of the toe, should eventually result in stabilization of the landslide. Published in 2008 by John Wiley & Sons, Ltd.

KEYWORDS: basal; slip surface; photogrammetry; morphology; aerial photographs; GPS; landslide; earth flow; hydrology; pond; time-lapse; dating; Slumgullion; San Juan Mountains; Colorado

Introduction

Ponds on landslides are often viewed as morphological indicators of recent movement (McCalpin, 1984; Crozier, 1984; Keaton and DeGraff, 1996). Because ponds accumulate fine-grained sediments, they often preserve sedimentological evidence and datable material that can be used to document initial and recurring landslide movement (Adam, 1975; Alexandrowicz and Alexandrowicz, 1999; van Den Eeckhaut *et al.*, 2007). On landslides where movement distances are short (i.e. less than a few hundred meters) and movement duration is short lived (i.e. a year or less in duration), the location of ponds is often attributed to the location of depletion zones or landslide surface structures such as grabens and scarps (Bisci *et al.*, 1996; Cruden *et al.*, 1997; Alexandrowicz and Alexandrowicz, 1999; Geertsema and Pojar, 2007), although reactivated landslides can provide exceptions to this statement (see, e.g., Baum *et al.*, 1993).

Landslides that have moved long distances (i.e. hundreds of meters), sometimes through multiple episodes of reactivation and often over long periods of time (i.e. longer than a year), have often been the focus of research, but ponds on these landslides have rarely been studied in any systematic way. Exceptions to this statement are provided in work by Williams (1988), Baum *et al.* (1993) and Fleming *et al.* (1999). Williams (1988) studied ponds on the Manti landslide in Utah and found that they migrated downslope with landslide debris during a three-year period of landslide movement. Baum *et al.* (1993) observed ponds on the Aspen Grove landslide in Utah during their five year study of landslide kinematics. They indicated that the locations of at least two ponds appeared to result from deformation over undulations in the basal failure surface. They noted that the ponds maintained their positions while the edges and bottoms of the ponds moved downslope, and trees that started out upslope from the ponds, eventually moved into the ponds. Fleming *et al.* (1999) described pond deposits at one location on the continuously moving Slumgullion landslide in Colorado. On the basis of observed deposits, they suggested that a pond near the toe of the landslide had remained spatially fixed for at least 100 years, while landslide material moved through the pond site. Additionally, Fleming *et al.* (1999) noted the need for systematic observations of surface water on the Slumgullion landslide. The implication of previous studies at Aspen Grove and Slumgullion is that pond locations are controlled by basal slip surface topography. Basal slip surface topography can control landslide stability (Stout, 1971), and an accurate portrayal of basal topography is important for landslide modeling studies (e.g. Flageollet *et al.*, 2000).

In this paper, we evaluate the implication that pond locations are controlled by basal slip surface topography using nine years of systematic field observations of eight ponds, and Global Positioning System (GPS) observations of the spatial position and water level elevation of one pond on the Slumgullion landslide. We also examine the persistence of surface morphology at the eight pond locations using topographic profiles measured from stereo aerial photographs from 1939/1940 and 2000.

The Slumgullion Landslide

The Slumgullion landslide is located in southwestern Colorado about 6 km southeast of Lake City (Figure 1). The landslide has three main parts, well-defined inactive and active parts and an area near the head that is probably active, but is currently (April 2008) poorly defined (Figure 2). The active part of the landslide (Figure 3) is 3.9 km long, ranges in elevation from about 2950 to 3650 m and has an estimated volume of 20 million m^3 (Parise and Guzzi, 1992). The average slope of the active landslide surface is about 8º. The landslide has historically been called an earth flow because of its flowlike surface morphology (see, e.g., Varnes and Savage, 1996). We prefer the term landslide because most movement at Slumgullion appears to take place by translational sliding along shear surfaces (based on mapped surface structures by Baum and Fleming, 1996, and Fleming *et al.*, 1999).

Materials that make up the landslide are derived from Tertiary volcanic rocks that are exposed in the headscarp

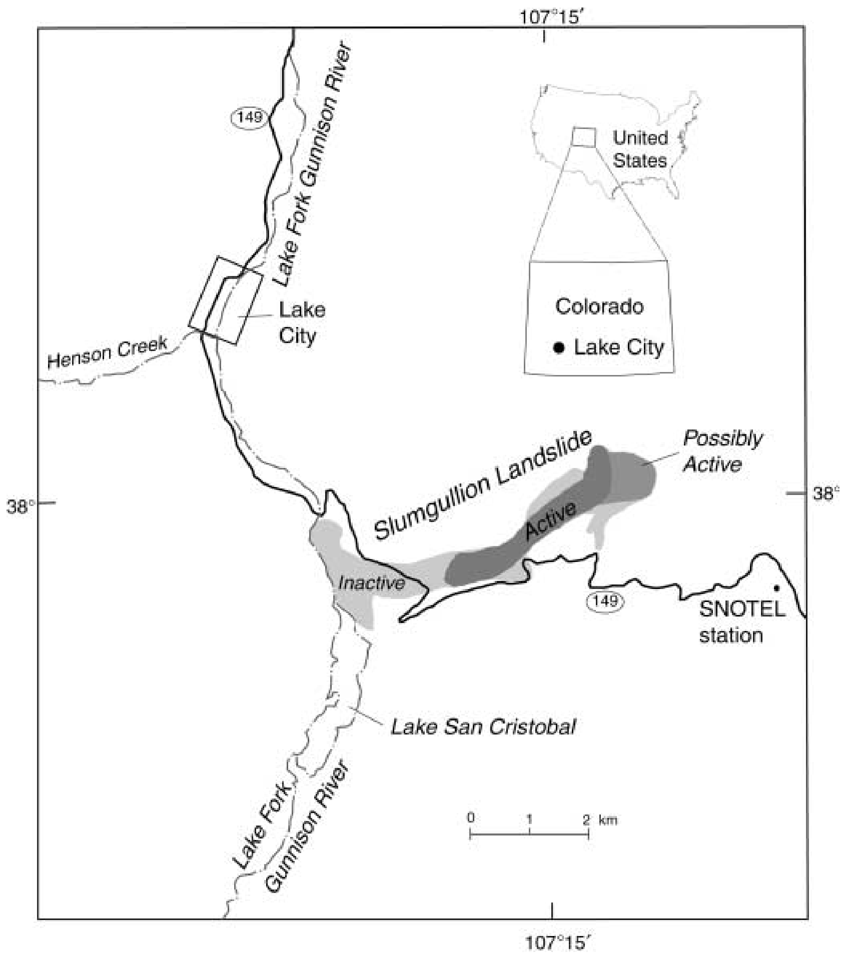

Figure 1. Map showing the location of the Slumgullion landslide southeast of Lake City in the San Juan Mountains of Colorado.

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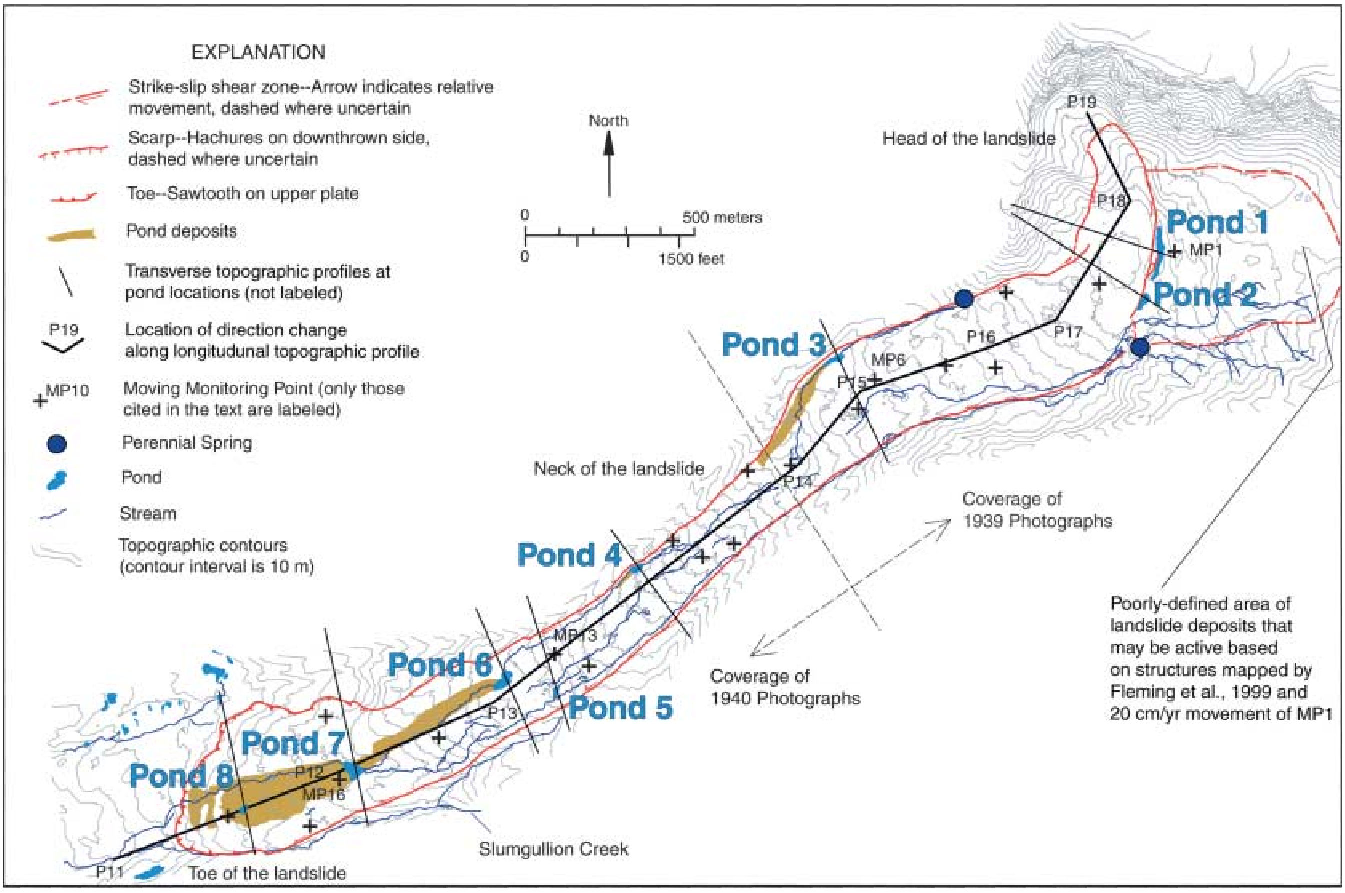

Figure 2. Map of the active part of the Slumgullion landslide. Pond locations are shown. Active landslide defined by Fleming *et al.* (1999), with the exception of the area of landslide deposits southeast of the landslide head, which is defined based on monitoring results from Coe *et al.* (2003). Topographic base map (including streams) from Messerich and Coe (2003).

Figure 3. Photograph of the active part of the landslide taken on 16 May 2000. This figure is available in colour online at www.interscience.wiley.com/journal/espl

area. The materials are heterogeneous, but typically consist of yellow, clay-to-sand sized grains with scattered patches of bouldery debris, reddish-brown and purple clay, and fan, pond and stream sediments. The majority of landslide material is from Tertiary andesitic flows that have been highly altered by acid sulfate hydrothermal alteration (Diehl and Schuster, 1996). The acid sulfate alteration causes surface water on the landslide to have a pH that ranges from 2 to 4 (Coe and Burke, 2003).

The active part of the landslide has probably been moving for about 300 years (Varnes and Savage, 1996). Movement since the 1950s is well documented (Crandell and Varnes, 1961; Smith, 1996; Savage and Fleming, 1996; Jackson *et al.*, 1996; Fleming *et al.*, 1999; Coe *et al.*, 2003; Schulz *et al.*, 2007). Annual rates of movement range from 0.1 m/yr at the head of the landslide to 1.5 m/yr at the toe to 7 m/yr in the neck of the landslide (Fleming *et al.*, 1999; Coe *et al.*, 2003). Movement is continuous throughout each year, but seasonally variable. Landslide velocity is generally positively correlated with groundwater pore pressure near the center of the landslide (Coe *et al.*, 2003; Schulz *et al.* 2007), and inversely correlated with pore-water pressure along the margin of the landslide (Schulz *et al.*, 2007). The inverse correlation along the margin may be due to a pore-pressure feedback mechanism (Moore and Iverson, 2002; Iverson, 2005; Schulz *et al.*, 2007) wherein landslide material dilates during acceleration, causing pore pressure to decrease and the landslide to decelerate.

The hydrology of the landslide has not been studied in detail. About one-third of the active part of the landslide occupies the former drainage of Slumgullion Creek (Figure 2). The upper reach of Slumgullion Creek now flows along the south flank of the landslide and merges with a creek coming off of the active part of the landslide just below the active landslide toe. Both creeks are perennial. Nearly all of the drainages on the active landslide merge to form the landslide creek (see Messerich and Coe, 2003). The landslide drainages are fed by perennial springs located near the head of the landslide (Figure 2). The body of the landslide contains numerous springs and sinks (Fleming *et al.*, 1999).

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Depressions on the landslide surface often contain ponds and/or fans. Whether or not a depression contains water (i.e. is a pond or fan) appears to be dependent on the location of surface drainages, which can alternate between supplying water to depressions, flowing into sinks or being diverted to new channels as landslide structures form or are destroyed by continuous movement. Throughout this paper, we use the term 'pond' to label and describe the depressions that we observed from 1998 to 2007 because all of the depressions contained water during at least part of the observation period, and most of the depressions contained water throughout the entire observation period. When discussing depressions that did not contain water, we use the term 'dry' or 'dry pond' when describing the depressions, and 'fan' or 'fan sediments' when describing sediments deposited in the depressions.

Methods

Field and GPS observations

We observed the eight ponds (Figure 4) during periodic trips to conduct campaign-style, rapid static GPS surveys of movement at 20 monitoring points on the landslide (see Figure 2 and Coe *et al.*, 2003). We conducted 30 surveys during the nine-year observation period (1998–2007), with each survey lasting one to three days. The shortest time between field campaigns was 17 days and the longest was 569 days. The mean time between campaigns was 112 days. Because GPS monitoring points were distributed across the entire landslide (Figure 2), we observed most of the surface of the landslide during each GPS survey. Therefore, during each survey, we observed ponds, streams and springs, and took notes on variability in locations, amounts of flow and water levels. We also took ground-based photographs of some ponds to document pond locations with respect to moving landslide debris and vegetation. At one pond (Pond 6, Figures 2 and 4) we measured the elevation of pond water level and the spatial location of the pond as part of our regular survey campaign. At Pond 6, we always attempted to set up our GPS station to measure the elevation of the pond water along the edge of a peninsula that extended into the pond. The horizontal position of this peninsula changed through time and this change was visible on our results (described later in the paper). The landslide moves downslope at Pond 6 at a rate of 4.5 m/year (1.2 cm/day). We conducted the GPS surveys to determine whether or not the pond water was also moving downslope.

Mapping of pond sediments

We mapped pond sediments in the field using 1:6000-scale aerial photographs from July 2000 as our base map. We identified pond sediments (Figure 5) in the field based on their sorted, generally silty-sand texture, internal stratigraphy and relatively consistent, light tan color. Pond sediments were easy to identify and map directly downslope from ponds, but became more difficult to confidently identify as distance from ponds increased and sediments became jumbled and mixed with landslide debris. We terminated pond sediment map units when the percentage of pond sediments exposed on the landslide surface dropped below about 60 percent. After field mapping, lines on the photographs were transferred to a 1:1000-scale July 2000 topographic base map (Messerich and Coe, 2003; see Figure 2) using a Kern DSR-11 photogrammetric stereo plotter (Chapius and van den Berg, 1988).

Analysis of aerial photographs

We used stereo aerial photographs taken in October 1939, October 1940 and July 2000 to measure and compare topographic profiles of pond locations. The scales of the photographs were 1:12 000, 1:12 000 and 1:6000 for the 1939, 1940 and 2000 photographs, respectively. The 1939 and 1940 photographs covered the upper and lower parts of the active landslide, respectively (see Figure 2 for coverage). The July 2000 photographs covered the entire active landslide. Figure 6 shows photographs from 1940 and 2000 in the region of the active landslide toe.

Each set of photographs was registered to a common ground coordinate system using ground control points and the Kern DSR-11 stereo plotter. The July 2000 photographs were registered using 40 ground control points that were surveyed and covered with clearly visible fabric targets several days prior to acquisition of photography (Messerich and Coe, 2003). Once the 2000 photographs were registered, the 2000 photographs served as a baseline source of ground-control points for photographs from 1939 and 1940. To register these older photographs, we followed guidelines outlined by Coe *et al.* (1997). We identified control points that were visible in the 2000, 1939 and 1940 photographs, and recorded their ground coordinates. These control points were located off of the active landslide and consisted of photo-identifiable objects such as road intersections, bushes, stream intersections, and stumps and trunks of trees. Eleven of these points were used to register the 1939 photographs and 17 were used to register the 1940 photographs. The overall root mean squared errors (RMSEs) from both registrations were 1.3 m in the Easting and Northing directions, and 0.9 m in elevation. Thus, all photographs were registered to a common ground coordinate system and subsequently measured topographic profiles could be directly compared. Walstra *et al.* (2007) describe a similar procedure for the registration of aerial photographs for landslide assessments.

Topographic profiles were measured from the 1939, 1940 and 2000 photographs in both transverse and longitudinal directions with respect to landslide geometry (see Figure 2 for profile locations). Profiles in the transverse direction were selected to document topographic changes (or lack thereof) at individual pond locations. The longitudinal profile was selected to document topographic changes along the approximate centerline of the active landslide. Each profile used starting and ending points off the active landslide surface. The spacing between individual measurement points along each profile was 2 m.

Results

Field observations

Periodic field observations made from 1998 to 2007 indicate that pond surface elevations and lateral extents are seasonally and annually variable according to precipitation conditions, and/or changes in the surface topography near pond sites that alter connections to the migrating stream network on the landslide. Mean annual precipitation at Slumgullion is about 650 mm (Figure 7). Roughly 60% of precipitation falls as snow from October through April. Snow melts in May and June, and rainfall, typically associated with thunderstorms,

Figure 4. Photographs of the eight ponds studied. See Figure 2 for locations. Photographs taken on dates shown.

occurs in July through September. Precipitation conditions during the observation period were highly variable. Total annual precipitation ranged from a 25-year high in Water Year 1999 (WY1999) to a 25-year low in WY2002 (Figure 7). An individual WY runs from 1 October to 30 September; thus, for example, WY 1999 refers to the period from 1 October 1998 to 30 September 1999. Below normal precipitation conditions began during WY2002 and continued

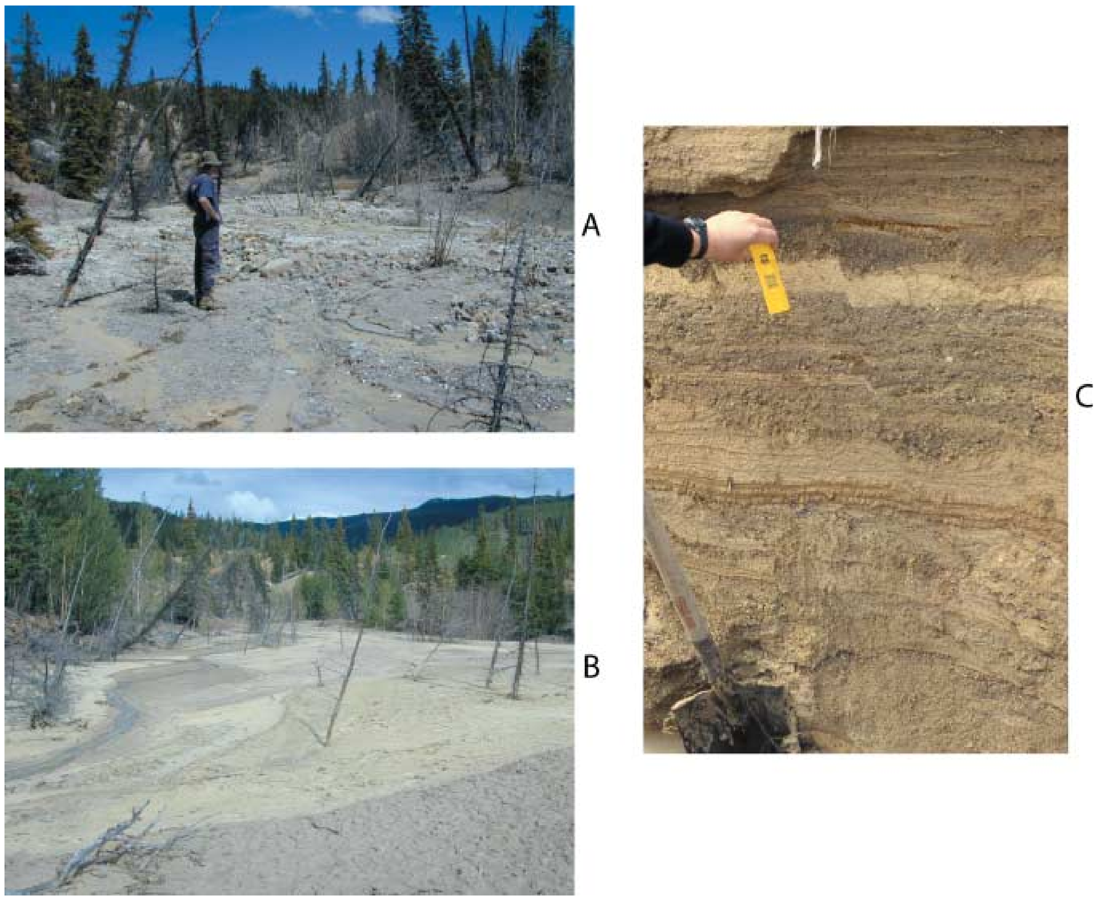

Figure 5. Examples of sediments deposited in ponds. (A) Coarse-grained sediments deposited in Pond 7. (B) Fine-grained sediments deposited in Pond 7. (C) Vertical section of pond sediments exposed by stream incision downslope from Pond 7. See geologist's hand and shovel for scale. The small normal fault to the right of the geologist's hand was probably created as the package of sediments moved over a convex upward bump adjacent to the pond. This figure is available in colour online at www.interscience.wiley.com/journal/espl

through WY2005 (Figure 7), with WY2002 classified as a severe drought (Pielke *et al.*, 2005). During this drought, we observed that two perennial springs located at the edge of the active landslide near the landslide head (Figure 2) flowed continuously and supplied water to the stream and spring network on the body of the landslide. Ponds reacted differently to precipitation conditions based on whether or not they were connected to this network. For example, the water level in Ponds 1, 2 and 8 (Figure 2 and Figure 4) decreased during the drought. Ponds 1 and 2 are located just off the active portion of the landslide adjacent to an active flank ridge, and Pond 8 is located on the advancing toe near the center of the landslide. There are no surface streams or visible springs that supply water to these ponds. From July 1998 to June 2001, these ponds appeared to be at average historical levels based on the level of the water surface with respect to long-term growth of algae. By June 2002, however, which was the peak of drought in the region (Figure 7), the water surface had dropped substantially and algae and evaporative salts were exposed at the surface and around the periphery of the ponds, respectively (see, e.g., Figure 8, 21 June 2002; Figure 4, Pond 8). By 2004, the pond surface levels had dropped still more, resulting in dead algae around the periphery of the ponds (see, e.g., Figure 8, 26 July 2004).

Conversely, Ponds 3–7 remained connected to the spring-fed stream network throughout the observation period and were mostly unaffected by the drought. Pond 7, for example, which is a seasonal pond located on the edge of a stream, continued to operate in a usual manner during the drought. Typically, Pond 7 only contained water during the summer thunderstorm season, when floods, hyperconcentrated flows, mud flows, or debris flows issued from the stream. When these events occurred, the southern corner of the pond filled with water, and the rest of the pond filled with coarse- (Figure 5(A)) to fine-grained sediment (Figure 5(B)). A longer-term record of deposition in Pond 7 (Figure 5(C)) is exposed by stream incision about 100 m downstream from the pond. The exposure reveals 2–30 cm thick layers of sediments of the same general variety as exposed on the surface of the pond, with a notable lack of debris-flow deposits. Our field observations indicate that one to three layers of sediments are typically deposited during each year.

Pond 6 (Figures 2 and 4) is interesting because the water surface elevation remained relatively static throughout the observation period, except for small-to-moderate decreases in January 2004 and August 2006, and a large decrease during the summer of 2002 (Figure 9). Field observations revealed that all of the decreases were caused when the pond was disconnected from the stream network by changes in landslide topography. When we first began observing this pond, it was fed by a stream from the south. However, between March and May 2000, this stream stopped flowing into Pond 6 because it was diverted to a different location by the changing topography of the landslide. During this time, the pond began to be fed by a stream from the north. Once this change occurred, the pond began to receive abundant sediment and a delta formed in the pond at the mouth of the stream. From May 2000 to June 2007, the delta continued to grow, and by June 2007 the entire pond had been filled by sediment (Figure 9).

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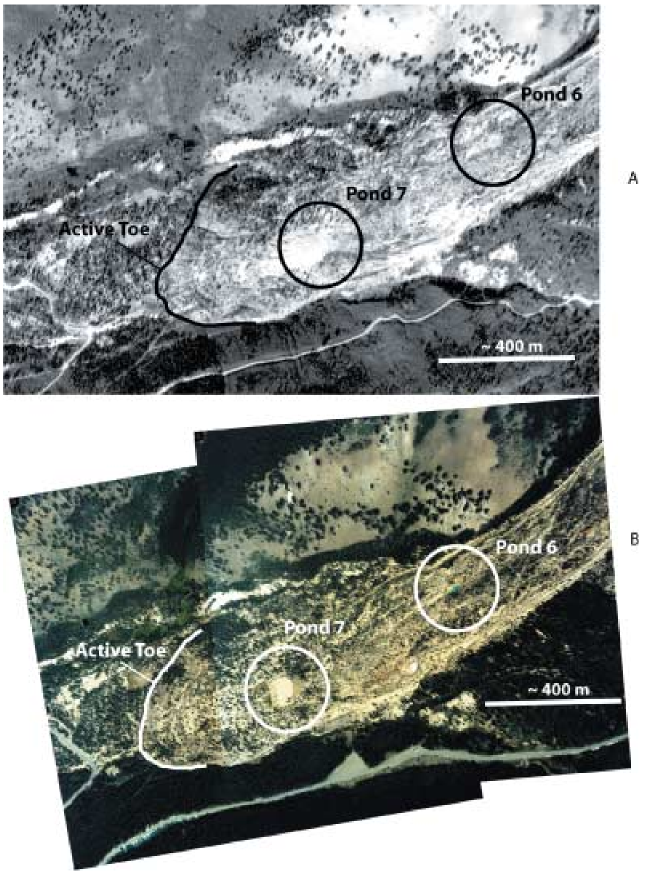

Figure 6. Aerial photographs of the active toe and vicinity. Pond locations are circled. (A) Photograph taken in 1940. (B) Photograph taken in 2000. This figure is available in colour online at www.interscience.wiley.com/journal/espl

Ground-based, time-lapse photographs of Pond 6

Ground-based photographs of Pond 6 taken between 1998 and 2007 show landslide material and vegetation moving into and through the pond as the pond water remains stationary (Figure 10). For example, note the position of the labeled tree (Figure 10), then observe how its position changes through time. In 1998, the tree was well upslope from the pond. By 2002, the tree had entered the water, and by 2007 the tree was more than halfway across the pond. The same progression was seen with the peninsula and the channel (Figure 10). GPS monitoring of point MP13 (located about 130 m upslope from the pond, Figure 2) during the observation period indicated that landslide material moved through the pond at an average horizontal rate of 1.2 cm/day (Coe *et al.*, 2003) for a total horizontal displacement of about 40 m.

GPS observations of Pond 6

GPS monitoring of Pond 6 showed that the elevation of the surface of the water (Figure 11(A)) stayed relatively constant throughout the nine-year observation period, except during periods when the stream that supplies water to the pond was diverted by changing landslide surface topography. With the exception of the September 2002, January 2004 and August 2006 observations, when the pond was disconnected from the stream network, the surface of the water fluctuated by only about 30 cm (between elevations of 3113.8 and 3114.1 m, Figure 11(A)). The elevation changes would be much larger than 30 cm if the pond were moving downslope with the landslide material. By using an average landslide slope of 8º, and a horizontal movement rate of 4.5 m/yr (based on movement of point MP13), we estimate that a representative block of landslide material near Pond 6 would have decreased in elevation by 5.7 m during the nine-year observation period.

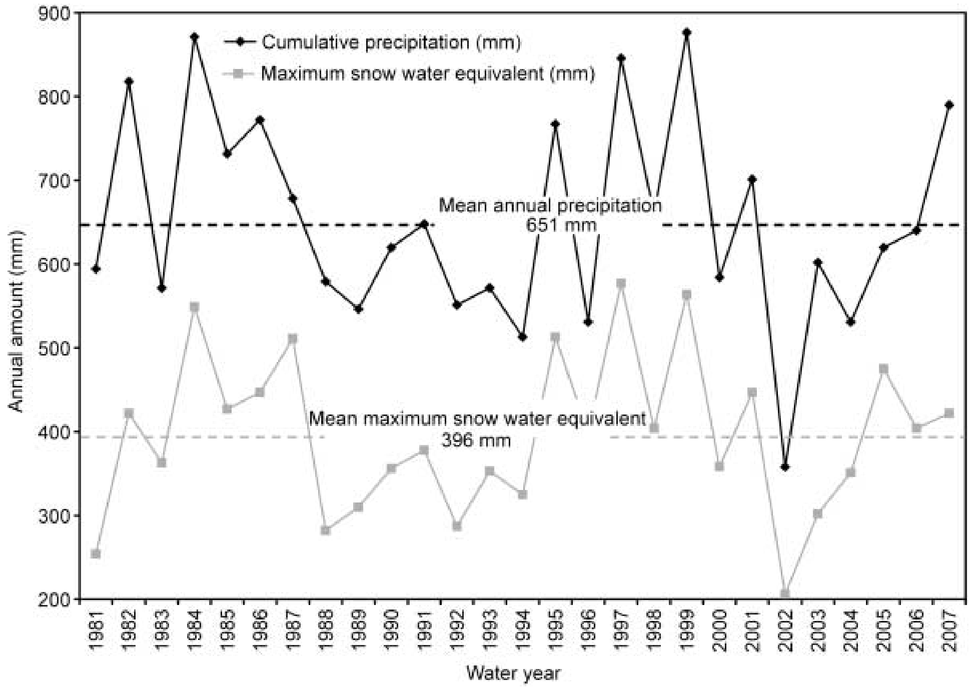

Figure 7. Precipitation records from the Slumgullion SNOTEL station (unpublished data, Natural Resources Conservation Service, US Department of Agriculture). See Figure 1 for approximate location of the station. Maximum snow water equivalent refers to the amount of water derived if snow on the ground were melted. Mean values from the 27-year period of record are shown by dashed lines.

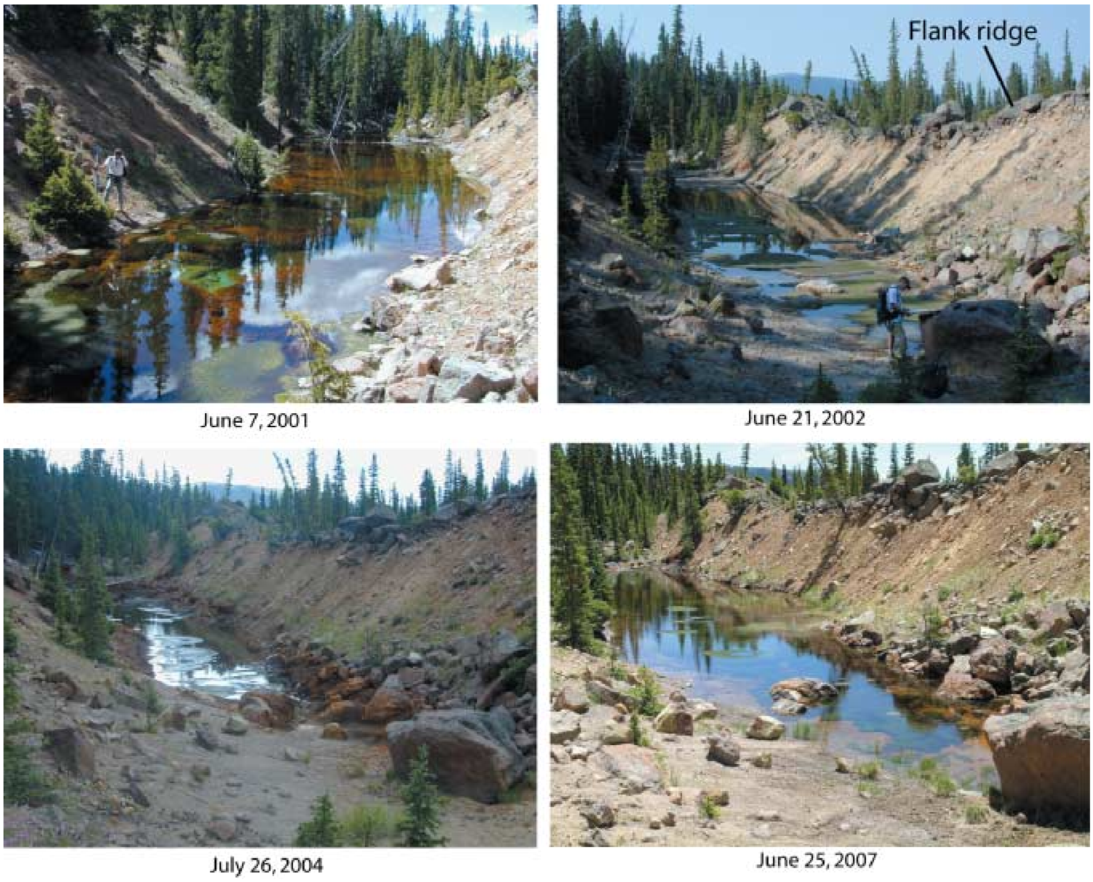

Figure 8. Time-lapse photographs of Pond 1. Dates on which photographs were taken are shown. This figure is available in colour online at www.interscience.wiley.com/journal/espl

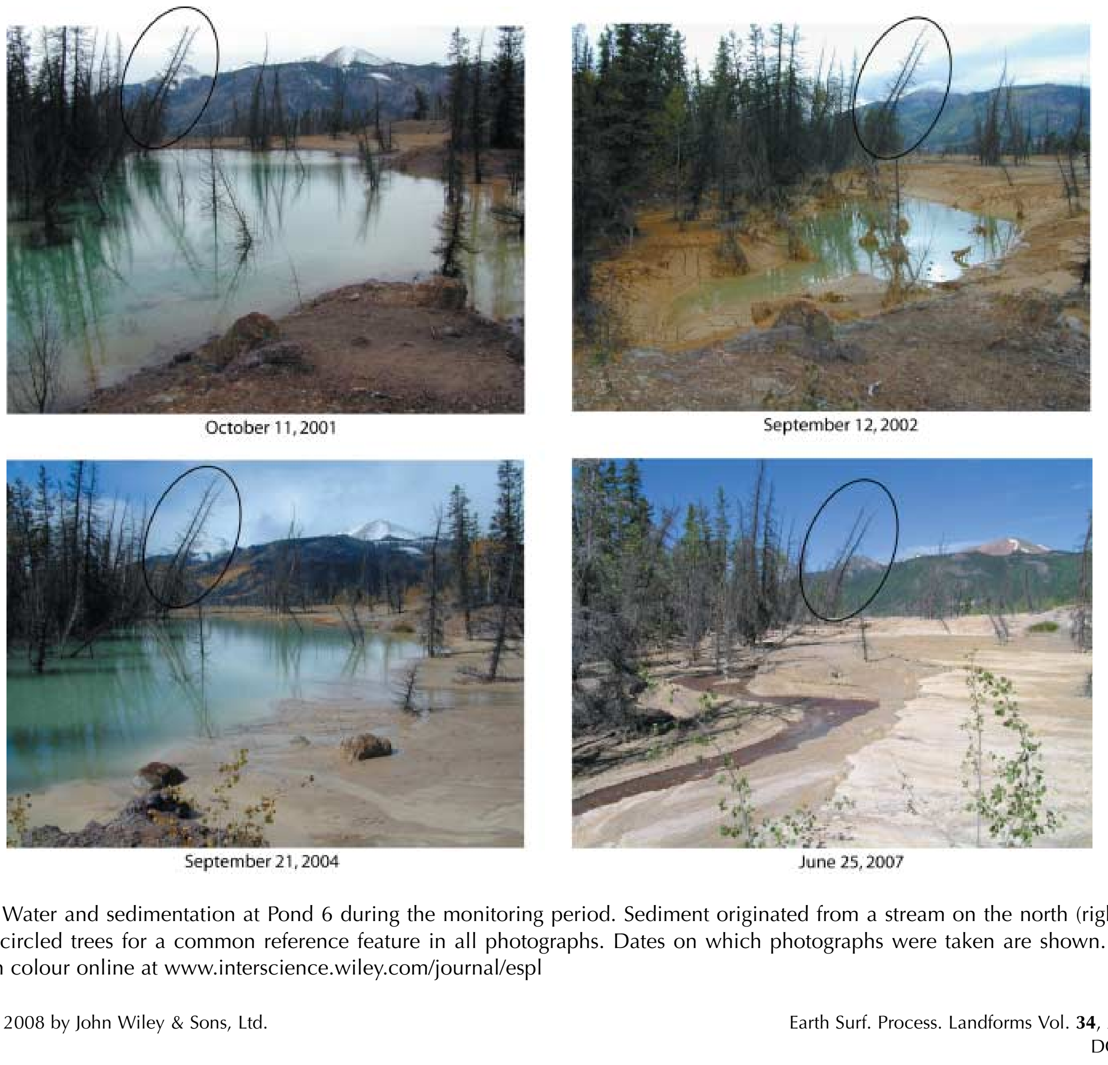

Figure 9. Water and sedimentation at Pond 6 during the monitoring period. Sediment originated from a stream on the north (right) side of the pond. See circled trees for a common reference feature in all photographs. Dates on which photographs were taken are shown. This figure is available in colour online at www.interscience.wiley.com/journal/espl

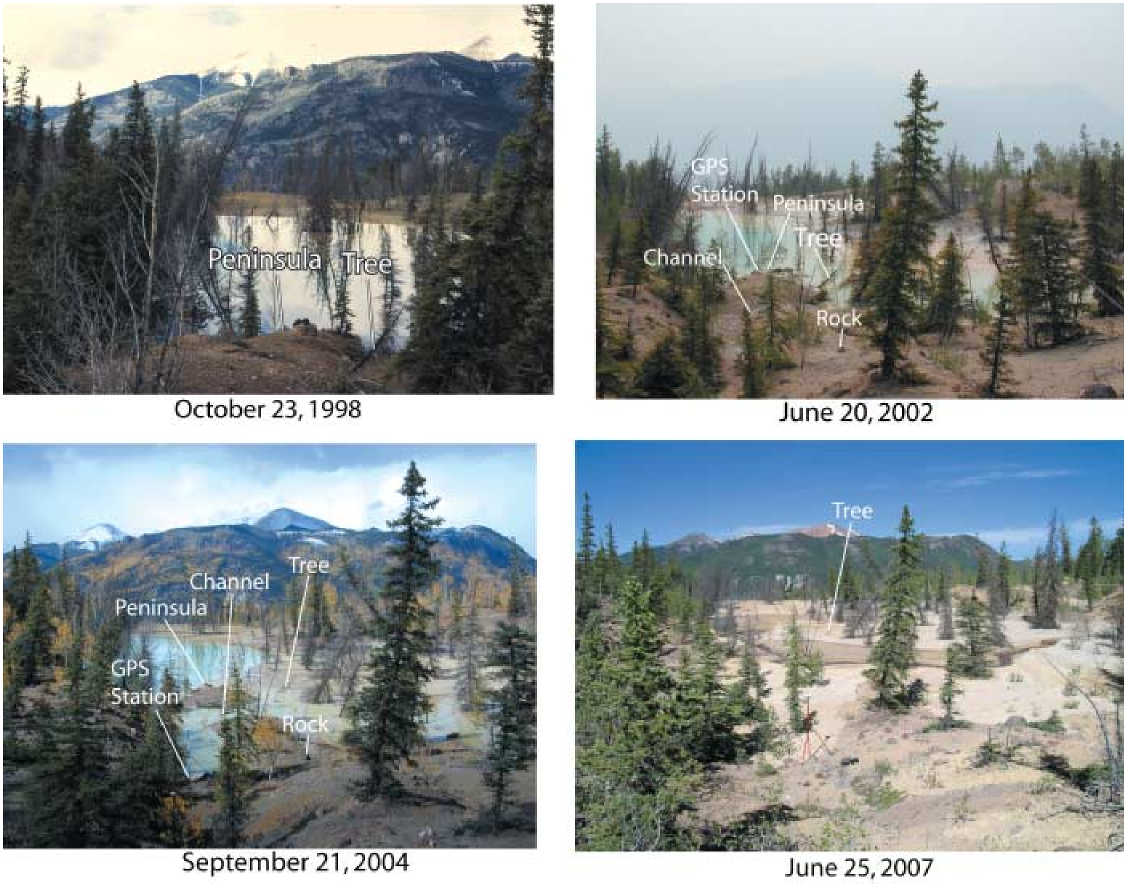

Figure 10. Time-lapse photographs of Pond 6 showing geographic items of interest, location of GPS station, and a tree and peninsula moving through the pond as the water remained stationary. Dates on which photographs were taken are shown.

The horizontal position of each GPS measurement location at Pond 6 is shown in Figure 11(B). Recall from the methods section that we always attempted to set up our GPS station to measure the elevation of the water at Pond 6 along the edges of the peninsula visible in Figure 10. We could not establish a fixed GPS benchmark at pond 6 because the primary goal of the work was to monitor changes in water elevation, which fluctuated. The sequential series of ground-based photographs (Figures 9 and 10) shows the fluctuation in water level, as well as the peninsula as it gradually moved into the pond as time progressed. This pattern is also seen in the GPS measurement locations at Pond 6 (Figure 11(B)). The measurement locations gradually migrate to the southwest, in the same direction of landslide movement. This pattern persisted until the summer of 2004, when the sequential photographs reveal that the peninsula had moved so far into the pond that it was breached by water (Figure 10, September 2004 photo). Instead of the usual measurement position along an edge of the peninsula, the measurement location in September 2004 was at the most upslope end of the pond, which was similar to the initial measurement position at the start of the observation period (Figure 11(B)). These GPS data indicate that the horizontal position of the pond remained spatially stationary during the observation period.

Pond sediments

Results from pond sediment mapping indicate that the largest ponds (Ponds 3, 6 and 7, Figures 2 and 4), which are also tied into the stream network and therefore receive a regular supply of sediment, have the most extensive deposits downslope from their positions on the landslide (Figure 2). Pond 3 has deposits that extend a horizontal distance of about 500 m downslope. The closest GPS monitoring point to Pond 3 is MP6 (Figure 2), which moves at a horizontal rate of 1.6 m/yr (Coe *et al.*, 2003). If we use this rate as a guide for movement of sediment at Pond 3, then the downslope pond deposits suggest that Pond 3 has been stationary for about 312 years of movement. Pond 6 has deposits that extend about 500 m downslope and then merge with deposits from Pond 7 (Figure 2). The closest GPS monitoring point to Pond 6 is MP13 (Figure 2), which moves at a rate of 4.5 m/yr. This rate of movement and the length of downslope sediments suggest that Pond 6 has been stationary for about 111 years of landslide movement. Pond 7 has deposits that extend 670 m to the active toe (Figure 2). The closest GPS monitoring point to Pond 7 is MP16 (Figure 2), which moves at a rate of 2.6 m/yr. If we use this rate as a guide for movement of sediment at Pond 7, then Pond 7 has been stationary for about 258 years of movement. None of these estimates of stationary times are exact because the rate of movement is variable, not constant, as sediments progress downslope from each pond. However, in general, the results indicate that the ponds have been stationary for a long period of time, in the range of 100–300 years.

One additional pond (Pond 4, Figure 2) is connected to the stream network and has a mappable downslope band of pond sediments that is about 100 m long. We attribute this limited band of sediments to the small size of the pond (Figure 4) and to the fact that the downslope sediments remain in a fairly narrow stream drainage, where they were reworked, eroded away, became diffuse in extent and thickness and were therefore difficult to identify and map.

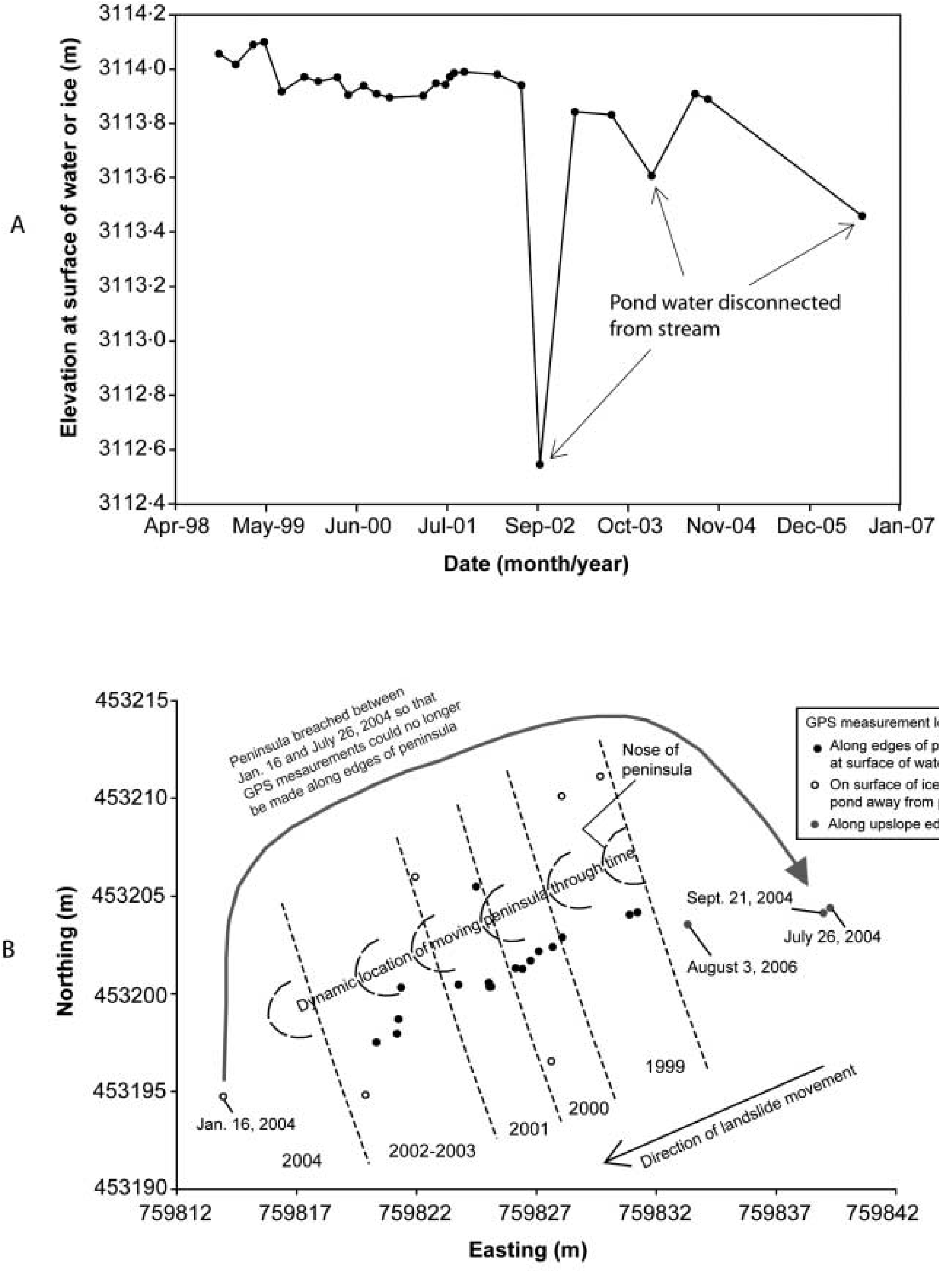

Figure 11. Diagram showing results from GPS monitoring of Pond 6. No measurements were made in 2007 because the pond was filled with sediment and did not contain standing water. (A) Diagram showing variation in the elevation of the surface of the water during the study period. Elevations are given in the North American Vertical Datum of 1988. (B) Diagram showing variation in the horizontal position of the GPS measurement location during the study period. Each dot is the position of the GPS measurement location on a date when data were collected for determining the elevation of the surface of water in the pond. Coordinates are given in the Colorado State Plane coordinate system.

At the other four ponds (Ponds 1, 2, 5 and 8), we could not identify any downslope pond sediments. At Pond 5, we attribute the lack of sediments to the fact that the pond was fed by a very short drainage, which did not produce sediment during the observation period. Ponds 1 and 2 are on the outboard side of the active landslide, were not connected to the stream network, and did not receive sediments during the observation period. Similarly, although Pond 8 is on the advancing toe of the active landslide, it was not connected to the stream network and did not receive sediments during the observation period.

Aerial photographs and topographic profiles

The longitudinal profile (P11–P19, Figure 2) measured along the length of the landslide from 1939/1940 and 2000 photographs (Figure 12) show that the shape of the landslide surface has not substantially changed during the 60- to 61-year period, although the upper part of the landslide has decreased in volume (by thinning, Figure 12(A)) and the area near the landslide toe has advanced and bulged upwards slightly (Figure 12(C)). The average elevation difference between 1939/1940 and 2000 elevations

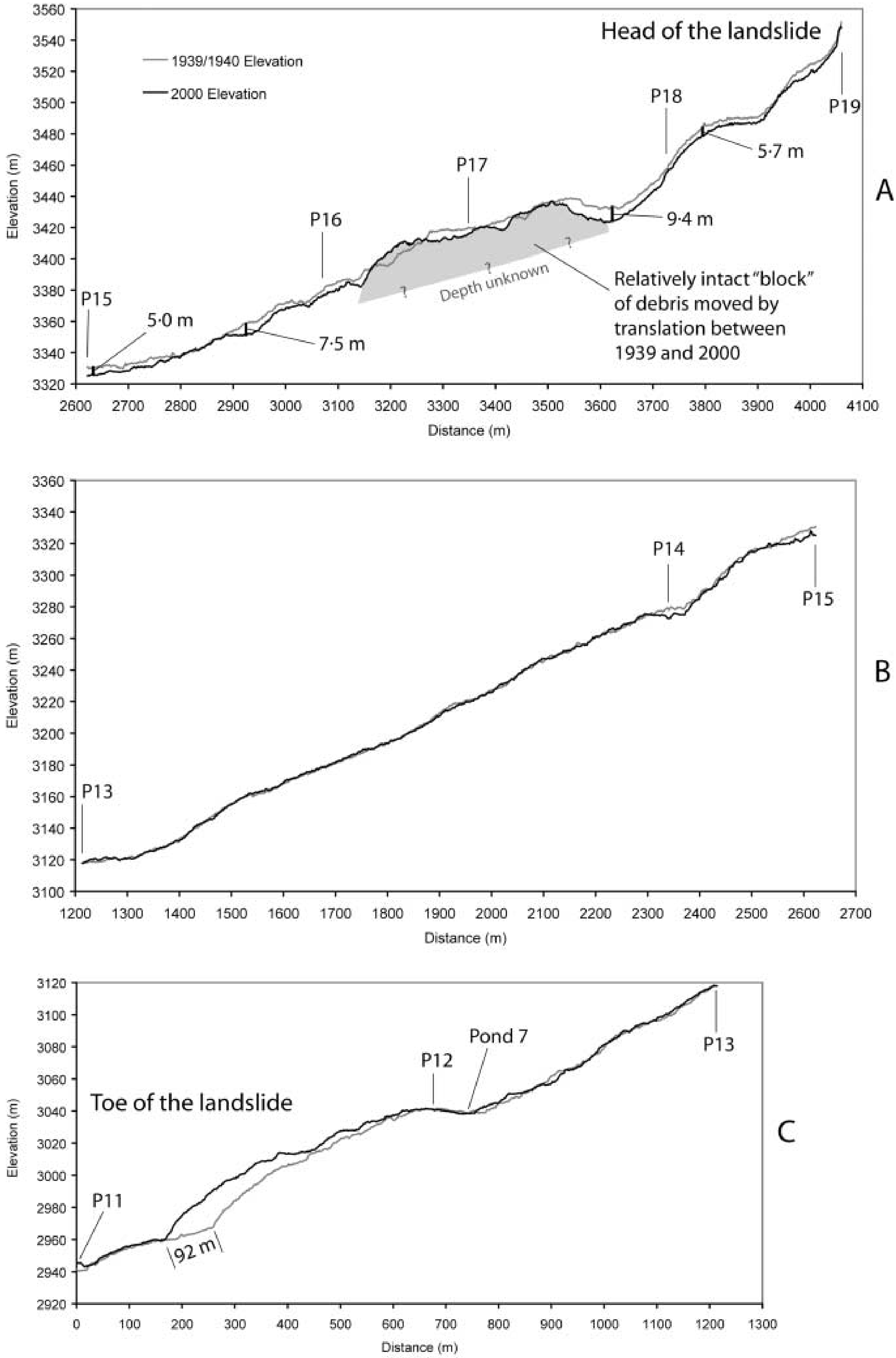

Figure 12. Longitudinal topographic profiles measured from 1939/40 and 2000 aerial photographs. Profile position is along the approximate centerline of the landslide (see Figure 2 for location). Vertical exaggeration is 3×. Points where the profile changes direction are shown (i.e. P12, P13, P14 etc.). Cumulative distances beginning at Point P11 are shown on the *x*-axis. (A) Profiles measured between P19 and P15. (B) Profiles measured between P15 and P13. (C) Profiles measured between P13 and P11.

along the entire length of the longitudinal profile (P11 to P19, Figure 2) was −0.16 m. Assuming that minimal debris was eroded (removed) from the active landslide between 1939 and 2000, one might expect this number to be closer to 0. Several probable explanations exist for why the average elevation change was less than 0. First, our vertical measurement RMSE of 0.9 m has undoubtedly played a role in causing the average value to be different from 0, and second, our single longitudinal profile location fails to take into account any lateral spreading of the landslide toe.

The shape of the middle and fastest moving part of the landslide is remarkably consistent, with the profile between points P12 and P14 being nearly identical in 1939/1940 and 2000 (Figure 12(B)). The average elevation difference along the length of the profile between these points was 0.15 m, with the 2000 elevations, on average, being slightly above the 1939/1940 elevations, but also generally within our vertical measurement RMSE of 0.9 m.

The shape of the landslide surface at the upper part of the landslide (upslope from profile point P14) is also very similar between 1939 and 2000, with the exception of the area located between profile points P16 and P18 (Figure 12(A)). In this area, it appears that a relatively intact "block" of landslide debris slid downslope between 1939 and 2000. The elevation change between 1939 and 2000 for the rest of the upper part of the landslide has been almost entirely negative (i.e., the landslide surface was substantially lower in 2000 than in 1939). If the area containing the intact block is excluded, then the average vertical lowering of the landslide surface between profile points P14 and P19 was 3.4 m. Using 61 years as the lowering period results in a vertical lowering rate (i.e. a material depletion rate) of 5.6 cm/yr. During the same time period in which the upper part of the landslide was being depleted of material, the landslide toe advanced 92 m (Figure 12(C)), at an average rate of 1.5 m/yr. Fleming *et al.* (1999) estimated that the toe was advancing at rates ranging between 0.8 and 2.0 m/yr in the mid-1990s, suggesting that the rate of 1.5 m/yr has been fairly constant. Therefore, between 1939 and 2000, neither the depletion of material at the head of the landslide, nor the continued growth of the landslide toe, has affected the overall movement rate of the landslide.

Transverse profiles (Figure 2 and Figure 13) indicate that pond locations (depressions on the landslide surface) have remained stationary for the 60- to 61-year period. Five of the seven ponds that contained water in 2000 did not contain water in 1939 and 1940. Ponds 1 and 2 contained water in both the 1939 and 2000 photographs; however, the elevation of the base of the ponds dropped by 3.5 m over the 61-year period (Figure 13(A) and (B)). The position of the ponds remained static in the landslide movement direction, but migrated by 8 to 10 m toward the centerline of the landslide (Figure 13(A) and (B)). This lateral migration is consistent with the decreasing landslide volume of the upper part of the landslide and the slow, east-to-west movement (about 20 cm/yr) of landslide deposits along the east flank of the well defined active landslide (see Figure 2 and monitoring point MP1 in Coe *et al.*, 2003).

Pond 3 contained water in the 2000 photographs, but did not in the 1939 photographs. The landslide surface morphology at the Pond 3 location was similar in 1939 and 2000, but the pond depression was more pronounced in 2000 (Figure 13(C)). Pond 3 was about 3 m higher in 1939 than in 2000. Visual inspection of the pond location in the 1939 photographs indicated that the location was an active fan, rather than a water-filled pond. Figure 13(C) also suggests that steep slopes adjacent to the active landslide near Pond 3, which were bare of vegetation in both sets of photographs, have degraded by as much as 3.5 m in 61 years.

Pond 4 occupies the downslope end of a major depression that is manifested as a pull-apart structure along the inboard side of the north flank of the landslide (Figures 2 and 13(D)). This pull-apart structure was one of the focus areas of a previous study of landslide structures at Slumgullion (see, e.g., Fleming *et al.*, 1999). The pond in the structure contained water in the 2000 photographs, but did not in the 1940 photographs. The shape and elevation of this depression, as well as the landslide surface along the transverse profile, were essentially the same in 1940 and 2000 (Figure 13(D)). In 1940, the pond location contained fine-grained fan or pond deposits.

Pond 5 occupies a relatively minor, but persistent, depression in the landslide surface (Figure 13(E)). The pond contained water in the 2000 photographs, but did not in the 1940 photographs. Although the depression at Pond 5 was better defined in 2000 than in 1940, the major topographic shapes and elevations of the landslide surface along the transverse profile were essentially the same in 1940 and 2000 (Figure 13(E)). In the 1940 photographs, the pond location contained fine-grained fan or pond sediments.

As with Pond 4, Pond 6 occupies a major and persistent depression in the landslide surface. The shape and elevation of this depression, as well as the landslide surface along the transverse profile, were essentially the same in 1940 and 2000 (Figure 13(F)). The pond contained water in 2000, but did not in 1940 (Figure 6). The landslide surface at the pond location in 1940 was covered with fined-grained fan or pond sediments.

Pond 7, as described earlier, was typically dry most of the year. This pond occupies a major and persistent topographic depression on the landslide surface (Figures 12(C) and 13(G)), that was previously described by Parise and Guzzi (1992) and Fleming *et al.* (1999). The shape and elevation of this depression, as well as the landslide surface along the transverse profile, were very similar in 1940 and 2000 (Figures 12(C) and 13(G)), but the pond did not contain water in either set of photographs.

Pond 8 occupies a minor, but persistent depression near the toe of the landslide (Figure 13(H)). The shape of this depression and the overall landslide surface in the vicinity of the pond are similar with regard to major elements, but also reflect major changes in elevation due to the advancing toe and expansion of the toe towards the south. The elevation of the base of the pond was about 5 m higher in 2000 than in 1940. The pond contained water in the 2000 photographs, but did not in the 1940 photographs.

Discussion

Our results indicate that ponds on the Slumgullion landslide (1) are sensitive to changes in the local, spring-fed stream network, or to climatic conditions if they are not linked with the stream network (e.g. Ponds 1 and 2), and (2) remain spatially stationary while landslide debris moves downslope and through the pond locations. Pond sediments indicate that three pond locations have remained stationary (in the direction of landslide movement) for between 100 and 300 years. Topographic profiles from 1939/1940 and 2000 indicate that all eight pond locations have remained stationary for at least 60 years. Movement rates documented by Fleming *et al.* (1999) and Coe *et al.* (2003) suggest that total landslide movement in the vicinity of the ponds during this 60-year period ranged from 90 to 270 m. In order for the ponds to remain stationary while the landslide moves continuously, the location of the ponds must be mimicking depressions on the basal surface of the landslide. The presence of depressions indicates that the topography of the basal slip surface is irregular.

Movement of landslides is usually concentrated on a basal slip surface (Hutchinson, 1970; Prior and Stephens, 1972;

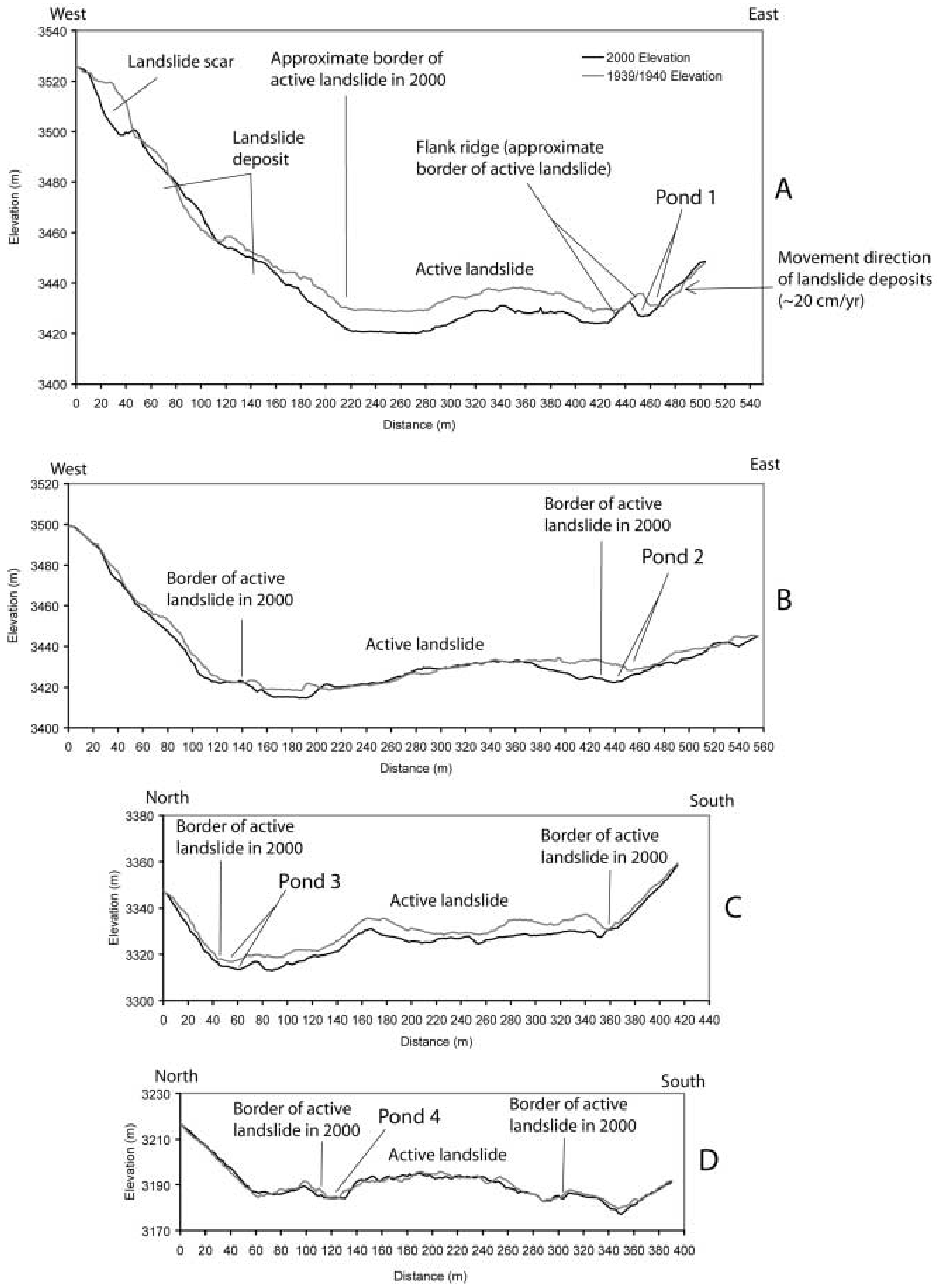

Figure 13. Transverse topographic profiles at each pond location (labeled) measured from 1939/40 and 2000 aerial photographs. See Figure 2 for locations of the profiles. Vertical exaggeration is 2× . Cumulative distances begin on the west or north side of the landslide and increase to the east or south, respectively. Active landslide borders were mapped from fieldwork and 2000 aerial photographs. The landslide border in 1939/40 may have been at a different location. Profiles are designated as follows: (A) profile at Pond 1, (B) profile at Pond 2, (C) profile at Pond 3, (D) profile at Pond 4, (E) profile at Pond 5, (F) profile at Pond 6, (G) profile at Pond 7 and (H) profile at Pond 8.

Keefer and Johnson, 1983; Baum and Johnson, 1993). Stout (1971) suggested that irregular basal slip surfaces were common in many landslides and that such surfaces had profound effects on slope stability calculations and mitigation designs. Hutchinson (1983) argued that accurate knowledge of the location and shape of basal slip surfaces was also critical for properly placing instrumentation, conducting sampling operations and designing mitigation schemes. Flageollet *et al.* (2000) stressed the importance of defining buried topography for understanding and modeling landslide dynamics. Multiple studies have described how variations in the geometry of slip surfaces can affect pore-water pressures, movement and internal deformation of landslides (see, e.g., Iverson, 1986; Zhang *et al.*, 1991; Baum and Johnson, 1993; Baum *et al.*, 1998; van Asch *et al.*, 2006, van Asch *et al.*, 2007). At most landslides, however, limited subsurface explorations

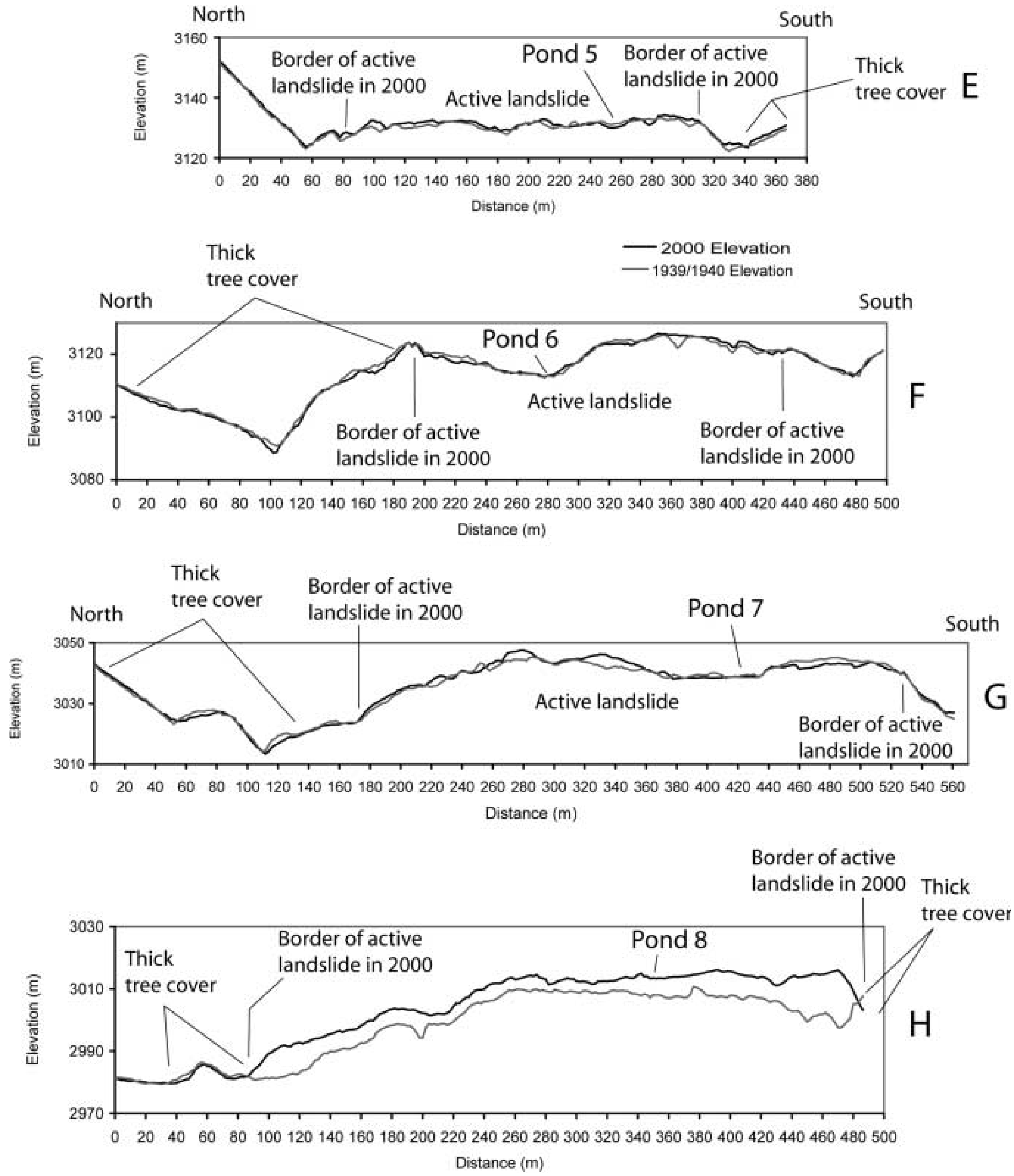

Figure 13. (*Continued*)

typically result in poorly defined basal slip surfaces that are usually idealized as smooth, arcuate surfaces for the purposes of modeling and stability analyses.

Hutchinson (1995) observed that surface morphology of existing slope failures can reveal the type of landslide, which in turn may help predict future behavior such as velocity, pattern of response to rainfall and likely travel distance. Our results suggest that for translational landslides that have moved distances greater than the dimensions of the largest basal-topographic irregularities (convex upward bumps and concave upward depressions as viewed from a position normal to the direction of landslide movement, e.g. Figure 12), landslide surface morphology can be used as a guide to the morphology of the basal slip surface. At Slumgullion, we do not have data to constrain the physical properties of the slip surface, but we suspect that the character of the surface is variable. That is, in some locations it is at the interface between altered bedrock and active landslide debris, and in other locations it is at the interface between older, inactive landslide debris and younger, active landslide debris. The maximum dimension of the largest basal topographic irregularities at Slumgullion is about 200 m (Figure 12, see numerous bumps and depressions between distances 2300 and 4100).

There are at least two areas at Slumgullion where it may not be possible to use surface morphology as a guide to the morphology of the basal slip surface. These include the face of the advancing toe and the flank ridges (see Figure 8, and Parise, 2003, for examples of flank ridges). The flank ridges were probably formed by a variety of processes related to lateral displacement of debris along the boundaries of the landslide, injection or tilting of material, diminishing material through time, creep movement of material normal to the active landslide and erosion and incorporation of basal material (Keefer and Johnson, 1983; Fleming and Johnson, 1989; Corominas, 1995; Fleming *et al.*, 1999). The existence of the relatively intact "block" of debris near profile point P17 suggests that the basal slip surface beneath the block is relatively smooth, which enabled the block to move intact without being disrupted by basal irregularities.

Examples of other landslides where it might be possible to use surface morphology as a guide to the shape of the basal-slip surface include the Acquara-Vadoncello (Wasowski and Mazzeo, 1998) and Alverà (Angeli and Silvano, 2004) landslides in Italy, the Super-Sauze landslide in France (Malet and Maquaire, 2004) and the Minor Creek landslide in California (Iverson and Major, 1987). All of these landslides appear to have moved by translation, probably for distances greater than the dimensions of their largest basal-topographic irregularities.

Additional implications of basal topographic control of landslide surface morphology are that (1) dateable sediments or organic material (see, e.g., Lang *et al.*, 1999) from basal layers of stationary ponds on a moving landslide will yield ages that are younger than the date of landslide initiation, and (2) landslide surface features such as faults, streams, springs and sinks might also be controlled by basal topography. The first implication could potentially be surmounted by dating material that was formerly in a pond, but is now located at a maximum distance downslope from the pond. For example, material from the furthest downslope extent of pond sediments shown in Figure 2 would yield ages closer to the date of initiation of recurrent movement at Slumgullion than would material from the present day pond locations. The second implication could be verified at Slumgullion by repeating the detailed mapping of landslide surface structures that was completed in the 1990s by Fleming *et al.* (1999).

Our topographic profiles indicate that the upper part of the Slumgullion landslide is becoming thinner by downslope movement of material and the toe is accumulating material and advancing downslope. The head of the landslide was depleted at a mean vertical lowering rate of 5.6 cm/yr between 1939 and 2000, while the toe has advanced at an average rate of 1.5 m/yr. If the upper part of Slumgullion continues to become depleted, without any resupply of material from the headscarp, then at some point in the future the landslide should stabilize and stop moving. Fleming *et al.* (1999) suggest a similar scenario based on mapping of surface structures and changes in landslide morphology.

Conclusions

We have presented multiple pieces of evidence that show that ponds on the Slumgullion landslide remain spatially stationary, while landslide debris moves through the pond locations. Whether or not a depression contains water is a function of the local, spring-fed stream network, or climatic conditions if the depressions are not linked with the stream network. The stationary nature of the ponds indicates that the irregular landslide surface morphology is a reflection of the topography of the basal surface of the landslide. Topographic profiles from aerial photographs taken in 1939/40 and 2000, and a map of pond sediments, suggest that the major features of the basal slip-surface have not changed significantly for 60 to 300 years. These results indicate that for translational landslides that have moved distances larger than the dimensions of the largest basal topographic irregularities (about 200 m at Slumgullion), it should be possible to use the surface topography as a guide to the shape of the basal slip surface. Areas that are possible exceptions to this statement include faces of landslide toes and flank ridges. Additional implications of this work are that dateable sediments or organic material from basal layers of stationary ponds on a moving landslide will yield ages that are younger than the date of landslide initiation, and that landslide surface features such as faults, streams, springs and sinks might also be controlled by basal topography.

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