The Eureka River landslide and dam, Peace River Lowlands, Alberta

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Abstract: The Eureka River landslide of June 1990, at 50 Mm$^3$, is one of the largest historical landslides on the Interior Plains of Canada. It is one of seven large translational landslides to have occurred in the Peace River Lowlands within the last 65 years. Each landslide occurred in Quaternary sediments deposited within a preglacial valley. Each landslide formed a dam. The rupture surface of the Eureka River landslide in preglacial lacustrine sediment, 125 m below the Peace River Lowlands plains, extended beneath the river channel causing the channel to be elevated. The resulting landslide dam was over 20 m high, forming a lake exceeding 8 km in length. The river cut a new channel around the toe of the landslide, abandoning the prelandslide channel. As the new channel is free of armour, incision has been rapid. After 10 years, the dam now stands approximately 5 m high.

Key words: landslide, landslide dam, Peace River, Alberta, preglacial valley, geomorphology.

Introduction

In June 1990, the north bank of the Eureka River, from river km 10.3 to 11.9 (measured along the river from the confluence with the Clear River), slid southward and dammed the river (Fig. 1). The landslide remobilized approximately 50 Mm$^3$ of surficial material deposited within a preglacial valley. The landslide’s rupture surface, in preglacial lacustrine sediment, extended beneath the river channel, causing the channel to be thrust upward by approximately 25 m, creating a Type 6 landslide dam (Costa and Schuster 1988) and a lake exceeding 8 km in length.

Large translational landslides associated with preglacial valleys are common hazards within the Peace River Lowlands of western Canada. Other historic large translational landslides with rupture surfaces in preglacial lacustrine sediment in the Peace River Lowlands include: the 1939 Montagneuse River landslide (Cruden et al. 1997), the 1959 Dunvegan Creek landslide (Hardy et al. 1962; Pennell 1969), the 1973 Attachie landslide (Evans et al. 1996; Fletcher and Hungr 2000), and the 1990 Saddle River landslide (Cruden et al. 1993). Historic large translational landslides with rupture surfaces in till include the 1990 Hines Creek landslide (Lu et al. 1998) and the 1995 Spirit River landslide (Miller 2000).

In this paper, we document the largest historic landslide on the Interior Plains for which photographic records predating and postdating the landslide exist. We begin by describing the watershed’s physiography and stratigraphy and then discuss the material properties of the preglacial lacustrine sediment. We then describe the landslide’s kinematics, explore possible landslide triggers, provide a stability analysis, and classify the landslide and associated successive landslides. The landslide dam and its fluvial geomorphic effects are then described. Finally, hazards associated with such large translational landslides and dams are discussed. This landslide adds to a developing regional understanding of landslide hazards in an area of rapid oil and gas development. Landslide terminology throughout this paper follows Cruden and Varnes (1996).
Background

Watershed physiography

The Eureka River watershed lies mostly within the Peace River Lowlands, Alberta; a broad, gently undulating plain dissected by steep gorges (Fig. 2). The Eureka River flows westward within one of the gorges. The river’s headwater is the Clear Hills, at 1035 metres above sea level (m.a.s.l.); the mouth is at the Clear River confluence, at 490 m.a.s.l. The Clear River flows westward a further 15 km, and then southward for 31 km, to join the eastward-flowing Peace River at 380 m.a.s.l. The vertical drop of the Eureka River from the headwater to the mouth is 545 m. From the confluence of the Eureka and Clear rivers, at river km 0.0, to river km 53.2, the longitudinal slope of the Eureka River is 0.14°, flattening to 0.04° at river km 61.0 (Fig. 3).

Historical aerial photographs, from 1945 to 1998, show continuous landslide deposits from the Clear River confluence to river km 18.2 on the north bank of the Eureka River, and to river km 15.2 on the south bank. The only landslide active during this time interval, and in this reach, was the 1990 event. Further, the thick vegetation cover on the scarps of the inactive landslides in 1945 suggests that there had been no other large landslides for several years prior to 1945. Upstream of this lower reach, soil falls predominate where the Eureka River undercuts its banks.

Valley slopes reflect the zone of instability; slopes are less steep where there have been large translational landslides (Fig. 3, Miller and Cruden 2001). Channel form also reflects the occurrence of landslides (Fig. 3); where large landslides have occurred, the channel is straight (terminology follows Selby 1985, p. 268). The channel gradient, however, is not obviously affected by the landslides (Fig. 3).

Stratigraphy

Within the Eureka River watershed, the thickness of the surficial sediment is generally between 40 and 75 m (Kerr 1971). Beyond the Peace River Lowlands, sediment thickness decreases (Ozoray 1982). In preglacial valleys, such as the Shaftesbury channel (Pawlowicz and Fenton 1995), sediment thickness can exceed 200 m (Kerr 1971). At the site of the Eureka River landslide, sediment thickness is at least 125 m. The Eureka River landslide is situated within a preglacial valley, as are other historic landslides at the Montagneuse (Cruden et al. 1997), Saddle (Cruden et al. 1993), and Spirit rivers (Miller 2000), and Hines (Lu et al. 1998) and Dunvegan creeks (Pennell 1969), indicated in Fig. 4. The stratigraphy within the preglacial valleys (Fig. 5) records the up-drainage advance of the Laurentide ice sheet and the subsequent down-drainage ice-front retreat. Bobrowsky et al. (1991), Bobrowsky and Smith (1992), and Catto et al. (1996) place the terminus of the Laurentide Glacier further west in British Columbia.

The main scarp of the Eureka River landslide, from 598 to 620 m.a.s.l., shows 15 m of normally graded rhythmite couplets (Fig. 6), overlain by 2–5 m of mottled grey and tan, clay and silt with no apparent structure. The uppermost mottled material likely formed following disturbance by
Fig. 2. The Eureka River watershed (Canada National Topographic System (NTS) map 84D, Energy Mines and Resources 1977) with climate stations and river gauges indicated. Kilometre markers are measured along the channel taken from 1:20,000 Alberta Provincial Base maps 84D/6NW and NE. The landslide is located at 56° 26' 00'' N, 119° 24' 15'' W on the north bank of the Eureka River, 11 km from the confluence of the Eureka and Clear Rivers. The crown of the landslide is at 620 m.a.s.l.; the toe is at 520 m.a.s.l. Contours are 100 feet. (© Produced under licence from Her Majesty the Queen in Right of Canada, with permission of Natural Resources Canada).

Fig. 3. Eureka River’s channel form, longitudinal slope (using channel length), and average valley slope (data from Alberta, Provincial Base maps 84D/6NW and NE).
Fig. 4. The Peace River Lowlands with historic large landslides and preglacial valleys indicated (dashed lines from Pawlowicz and Fenton (1995) and Kerr (1971)).

Fig. 5. Stratigraphy at the Eureka River, Montagneuse River (Cruden et al. 1997), Hines Creek (Lu et al. 1998), Spirit River (Miller 2000), and Saddle River (Cruden et al. 1993) landslides. Each landslide occurred within sediment deposited in a preglacial valley. Each of the sedimentary units within the preglacial valleys is prone to instability.
pedogenesis and frost action to the underlying rhythmite couplets. Each of the rhythmite couplets, 0.5–2 cm thick, consists of a tan silty layer overlain by a black clay layer. At least one turbidite, containing rip-up clasts, punctuates the exposed rhythmite couplets. The deposit also contains larger clasts, up to cobble size, presumed to be ice-rafted dropstones. We consider this uppermost unit to be a retreat phase postglacial lacustrine deposit. This classification is consistent with Ozoray (1982). Bobrowsky et al. (1991), Catto (1991), and Liverman (1991) describe similar assemblages in nearby areas within British Columbia and Alberta.

Beneath the postglacial lacustrine sediment, at the main scarp, is a diamicton (598–600 m.a.s.l.). The unit shows cemented silt layers up to 20 cm thick, punctuated by thin silty laminae. Where exposed, the contact of the diamicton and the overlying glaciolacustrine deposit is gradational, with thin beds of diamicton present in the basal 1 m of the overlying postglacial lacustrine assemblage (Fig. 6).

A 2 m high exposure of displaced diamicton, 100 m from the main scarp, shows a matrix-supported well-rounded pebbly sand unit, with predominantly metamorphic and igneous clasts of Laurentide provenance. Any pre-existing fabric in the pebbly sand unit was destroyed during the landslide event. The unit was traced laterally for about 150 m and is interpreted as an esker deposit. Alternatively, this unit might be interpreted as a fan, apron, or delta that developed along the margin of the retreating Laurentide glacier (as suggested in Catto (1991)).

The bottom of the till unit is exposed adjacent to the contemporary stream, from about 520 to 525 m.a.s.l. Here the till is massive, dark-grey, with occasional isolated angular sand blocks. The till is similar in composition to the underlying rhythmite deposit, though lacking structure. We classify this exposure as till, following Benn and Evans (1998).

Beneath the till are contorted, graded, clayey-silt rhythms, each about 1–3 cm thick (Fig. 7). The unit extends from about 520 m.a.s.l. to below the bottom of the valley floor (515 m.a.s.l.). The rupture surface daylighted within this unit, 1 m above the contemporary stream. The unit is classified as an advance phase preglacial lacustrine glaciotectonite deposit. This classification is consistent with sedimentological analyses performed at the Saddle River (Cruden et al. 1993), Montagneuse River (Cruden et al. 1997), and Hines Creek (Lu et al. 1998) landslides. The term glaciotectonite is used here because the original structure is apparent and discernable, though contorted (Benn and Evans 1998). Stratigraphic measurements made adjacent to the contemporary stream suggest the contortions are both glaciotectonic and landslide induced features (Miller 2000). Folds with axes perpendicular to the valley are likely glaciotectonic features, whereas folds with axes parallel to the valley are likely due to the landslide (Miller 2000).

Geotechnical properties of the preglacial sediment

The texture and Atterberg limits were determined for one sample of preglacial lacustrine sediment, collected adjacent to the Eureka River landslide’s rupture surface. Another sample from the same location was washed with hydrochloric acid to remove calcium carbonate, and analyzed for texture and Atterberg limits. Despite a 6.9% reduction in mass following the hydrochloric acid wash, there was no significant change in the texture and Atterberg limits of the sample.
The textures, Atterberg limits, and activity of preglaciated lacustrine sediments from the Eureka River landslide, the Montagneuse River landslide (Cruden et al. 1997), the Saddle River landslide (Cruden et al. 1993), the Dunvegan Creek landslide (Pennell 1969), and the Attachie landslide (Evans et al. 1996; Fletcher and Hungr 2000) are presented in Table 1. The texture and Atterberg limits of the sample of preglaciated lacustrine sediment collected at the Eureka River landslide are within the range of values found at the Attachie landslide, but well beyond the range of values reported for the Saddle River and Montagneuse River landslides. The activity of the sample from the Eureka River landslide is similar to that at the Attachie landslide, but dissimilar to the Dunvegan Creek, Saddle River, and Montagneuse River landslides. Variations in the activity and Atterberg limits suggest differences in clay mineralogy, possibly due to influences from different sediment sources.

The material properties of the sample of preglaciated lacustrine sediment from the Eureka River landslide were used to develop a preliminary estimate of the landslide’s residual friction angle, \( \Phi_r \), following Skempton (1985). Skempton’s (1985) empirical relationship between clay fraction and \( \Phi_r \) for soils with activities between 0.5 and 0.9, produces a \( \Phi_r \) of between 11 and 16° for the Eureka River landslide. Normalizing these values to an effective normal stress, \( \sigma'_{\text{ho}} \), of 100 kPa, following Skempton (1985), reduces the estimated \( \Phi_r \) to between 8.2 and 13.1°.

Empirical \( \Phi_r \), peak friction angle (\( \Phi_p \)), and cohesion (\( c \)) values from the Montagneuse River (Cruden et al. 1997), Saddle River (Cruden et al. 1993), Dunvegan Creek (Pennell 1969), and Attachie (Evans et al. 1996) landslides are presented in Table 2. Table 2 shows reasonable consistency among the Saddle River, Montagneuse River, and Dunvegan Creek landslides; however, the \( \Phi_r \) from the Attachie landslide is inconsistent with the other landslides. The \( \Phi_r \) value generated for the Eureka River landslide, using Skempton (1985), falls close to the \( \Phi_r \) values of the Saddle River, Montagneuse River, and Dunvegan Creek landslides.

The 1990 Eureka River landslide

Landslide kinematics

The 1990 landslide is located between river km 10.3 and 11.9 on the north (left) bank of the Eureka River. The surface morphology indicates the landslide is composed primarily of four main blocks (Figs. 8, 9, and 10).

Block 1 is 1370 m wide (extending from river km 10.3 to 11.9) and 300–375 m long. The block was translated southwards by a maximum of 30 m (determined by comparing pre- and post-landslide aerial photographs, Figs. 8 and 11) between river km 11.3 and 11.9. The extent of the displacement of this block decreased westward, with less than a 2 m southward translation, between river km 10.3 and 10.8. There was little disturbance to the forest cover atop this block, indicating a deep, planar rupture surface and translational motion.

Block 1 consists of the colluvial deposits of three neighboring and older landslides (Figs. 11 and 12). The earlier slides likely occurred several years prior to 1945, as the main scarp had been extensively modified by 1945 (Canada aerial photographs A8303, 54–56). The margins of block 1 follow the main scarps of the eastern and central older landslides, as well as a major fracture of the western older landslide (Fig. 12), suggesting the reactivation and amalgamation of the rupture surfaces of the older slides. Discontinuous north–south fractures, approximately coinciding with the lateral margins of the central earlier landslide, accommodate the differential displacement across the block (Figs. 9 and 12).

The toe of block 1, which includes the pre-landslide riverbed, was thrust upward by up to 25 m (based on the extent of flooding upstream of the landslide dam) at river km 11.4. Upstream (east) and downstream (west) of river km 11.4, uplift decreases. Bedding orientations measured at the toe of the landslide suggest folding of the toe occurred initially, followed by rupturing, then thrusting of the hanging wall (Miller 2000). Subsequent stream incision around the toe of the landslide has left lengths of the original channel elevated above the contemporary channel. The rupture surface was located some 20 m beneath the pre-landslide stream channel, 1 m above the 1999 river level, at river km 11.4. Further incision is likely to occur.

Block 2, upslope of block 1, was in place prior to the landslide, so the movement of block 2 represents landslide retrogression. The upper surface of the block was the modified landslide scarps of the eastern and central older landslides (Figs. 8 and 11). The block extends from river km 11.0 to 11.7, although only the portion between river km 11.3 and 11.7 is intact. Between river km 11.0 and 11.3, the block is highly fractured. Between river km 11.3 and 11.7, the block is about 350 m wide and between 55 and 110 m long. During the landslide, this block dropped 20–45 m, translated southward 20–35 m, and rotated backwards about 20° (based on four tree tilt measurements and a pre- to post-landslide topographic comparison). The rotation is indicative of a curved rupture surface beneath the block, with a radius of curvature estimated at 190 m.

Between blocks 1 and 2 are a minor scarp, 3 tension cracks, and 4 uphill-facing scarps. These create a graben with a width of approximately 60 m. The cracks and uphill-facing scarps suggest that the graben formed in response
to near-surface extension, likely due to the backwards rotation of block 2.

Block 3 extends from river km 11.2 to 11.6. The block is wedge shaped and almost devoid of vegetation, suggesting that most of this block was not previously exposed at the surface. Vegetation was found in one small area (less than 2 m²) near the centre of the block. The block was translated southward about 30–50 m and downward about 20 m.

Upslope of block 3 are blocks 3E (east) and 3W (west) (about 0.75 ha each). The translation of these blocks was similar to that of block 3, although with some rotation. Their surfaces now strike at 206° and dip 13°N (3E), and 136° and 10°N (3W). The upper surfaces of the blocks were formerly at Lowland plain level and are treed.

Block 4 is located between river km 10.3 and 10.8, 200 m north of the Eureka River. The block is 150 m long and 450 m wide. The block was displaced southward and downward less then 2 m, in response to a loss of toe support following the movement of block 1. There was no disturbance to the forest on this block. Because block 4 consists of colluvium of the western earlier landslide, its movement represents a reactivation.

The volume of the 1990 landslide was estimated using Cruden and Varnes (1996, Fig. 3.6), which characterized the displaced material as a half ellipsoid. With a landslide length of 600 m, a width of 1370 m, and a depth of 115 m, the displaced volume is 50 Mm³. In Fig. 10 there appears to be a 15% loss of volume when pre-landslide topography is compared to post-landslide topography. A post-landslide loss of volume at the landslide toe has occurred due to stream erosion. Error in the topographic maps due to forest cover may also contribute to this apparent loss of volume.

**Landslide trigger**

The conditions for the Eureka River landslide were created by streambed incision and accompanying slope softening, due to lateral and vertical unloading (Imrie 1991; Matheson and Thomson 1973). While there were preparatory movements (Fig. 12), the greatest displacement occurred in June 1990. Here we explore possible triggers for this event.

A review of the Canadian National Seismological Data-Base (Geological Survey of Canada 2000) found no earthquakes in the spring of 1990 that might have triggered the Eureka River landslide. Thus, the possibility of a seismic trigger was eliminated.

Possible hydrologic triggers include a decrease in slope toe support due to streambed incision and an increase in the slope pore-water pressure due to an elevation of the water table. Cruden et al. (1993) and Lu et al. (1998) attribute the Saddle River and Hines Creek landslides, respectively, to streambed incision during a major storm in mid-June 1990. Cruden et al. (1993) calculated a return period of 50 years for the flow in the Saddle River during this storm. A storm trigger for the Eureka River landslide is not obvious. Maximum instantaneous and maximum daily stream discharge data from the Eureka River gauge (Environment Canada, Water Survey of Canada 1998) have 15 April as 1990’s extreme values. Further, the 15 April maximum instantaneous discharge, 16.7 m³/s, ranks as the 13th largest

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flow out of 21 records from 1976 to 1998; this discharge is well below the mean of 27.2 m$^3$/s. At the Clear River gauge (Environment Canada, Water Survey of Canada 1998a, Fig. 2), the maximum instantaneous discharge for 1990 occurred on 12 June (98.2 m$^3$/s, rank 8/23 (1972–1998), mean 114.7 m$^3$/s), indicating that the June storm may have affected...
the lower Eureka River watershed, although the station’s maximum daily discharge occurred on 26 April.

Precipitation data from the town of Eureka River (Environment Canada, Climate Services 1998) shows above-average rainfalls for June 1990 (94.4 mm, mean 63.9 mm, rank 7/36 (1962–1997)). The Clear Hills Lookout (Environment Canada, Climate Services 1993) shows above-average rainfalls for May (rank 5/29 (1961–1991)) and June 1990 (rank 4/29 (1961–1991)). Although neither of these stations recorded a storm as intense as the one that was attributed to triggering the Saddle River landslide (Cruden et al. 1993), the heavy 1990 precipitation points to elevated pore pressure as a possible trigger for the landslide.

Longer-term hydrologic trends were examined (Miller 2000) because Laycock (1990) had suggested that agricultural practices in the Peace River Lowlands have increased

**Stability analysis**

In the absence of subsurface information, we assume that the displaced material is a single wedge of rock with a vertical main scarp at the crest of the valley, the water table is at the ground surface, and the surface of rupture is horizontal, at residual, and with no cohesion. We use these assumptions in the model developed by Norrish and Wyllie (1996, eq. 15:3). The assumption of the water table being at the ground surface was supported by abundant pre-landslide surface ponding near the landslide (Fig. 11) and above-average precipitation in June 1990 (Environment Canada, Climate Services 1998b).
The assumptions of no cohesion and the friction angle being at residual were made because pre-1990 landslide movement was apparent (Figs. 11 and 12). The assumption that the main scarp was at the crest of the valley was made to simplify calculations, and is true for those areas in which there was no retrogression (see Figs. 8 and 11). The assumption of rupture surface horizontality is based on the assumption of a stratigraphic control on the rupture surface. These assumptions were also made in the analyses of the Saddle River and Montagneuse River landslides (Cruden et al. 1993, 1997). With these assumptions and a factor of safety of 1, Norrish and Wyllie’s (1996) equation becomes

$$\tan \Psi_f = \frac{\tan \Phi_r (\gamma_r - \gamma_w)}{\gamma_w} = 0.99 \tan \Phi_r$$

where $\Psi_f$ is the slope of the valley (12.8°), $\gamma_r$ is the unit weight of the soil (19.5 kN/m³; an average from Table 1), and $\gamma_w$ is the unit weight of water.

Equation [1] estimates $\Phi_r$ at 12.9°. This value is reasonably consistent with the preliminary $\Phi_r$ estimated using Skempton (1985) of 8.2–13.1°. Laboratory soil shear testing, to better define the $\Phi_r$, was not performed in this study because undisturbed samples could not be collected without drilling.

Landslide classification

The Eureka River landslide showed activity prior to 1990 (Fig. 12). The landslide’s peak velocity, in June 1990, can be judged only from the observation that several trees behind the main scarp were snapped (L. Foster, Worsley, Alberta, personal communication, August 1999), suggesting very rapid movement (3 m/min) or faster. Between 6 October 1992, when the first post-landslide aerial photograph (Alberta, AS4333, 175) was taken, and 1 May 1998, when the most recent aerial photograph (Alberta, AS4917, 45) was taken, blocks 2 and 3 moved southward approximately 25–30 m, indicating the landslide remained active.

Successive smaller landslides also suggest continuing activity. Instability is common at the crown and toe of the landslide. At the toe, the stream has cut a new channel to the south of the original channel. Approximately 20 m of incision has taken place in 9 years (2.2 m per year). Due to this rapid incision, slides, falls, and flows from both banks are common (Fig. 13).

At the crown of the landslide near the landslide access (Fig. 9), about 30 m of scarp retrogression occurred between 6 October 1992 and 1 May 1998 (estimated by comparing Alberta aerial photographs AS4333, 177 (1992) and AS4917, 45 (1998)). This retrogression includes soil falls and slides (Fig. 14). Flows, generated from within the apron of displaced material immediately below the main scarp, are also common.

Four slide-flows to the west of block 3 and one to the east of block 3 were mapped (Fig. 9). The largest of these, located directly west of block 3, is 65 m long, 35 m wide, and up to 2.5 m deep, with a volume of 3000 m³. Many smaller slides, flows, and falls were observed from the south flank of block 3 (Fig. 9).

West of block 4, on the west bank of a small tributary stream that enters the Eureka River at river km 10.35, a sizeable slide is present (Figs. 8 and 9). The slide may be a
The landslide dam

Landslide dam type and status

The 1990 Eureka River landslide dammed the Eureka River between river km 10.7 and 11.7. The landslide’s rupture surface extended beneath the former riverbed; thus, the landslide dam is a Type 6 dam, according to the landslide dam classification system of Costa and Schuster (1988). The Eureka River is cutting a new channel around the toe of the landslide, abandoning a length of the pre-landslide channel (Figs. 1 and 9). By August 1999, the original channel was 20 m above the new channel at river km 11.4.

At its maximum, the dam stood over 20 m, creating a lake exceeding 8 km in length (river km 11.5 to 19.8). The lake’s high stand is recorded by flood deposits in the branches and on the bark of trees at river km 17.1, 6 m above the contemporaneous stream. By 6 October 1992 (Fig. 15), the lakeshore had regressed almost 2 km to river km 17.8, which corresponds to a decrease in lake depth of about 5 m. By 1 September 1997, the lakeshore had regressed a further 3.1 km to river km 14.7 (Alberta aerial photograph AS4892, 148), which corresponds to a further decrease in depth of about 8 m. By that time, the lake had also divided at river km 11.7, creating a small lake between river km 11.5 and 11.6, and a large lake between river km 11.7 and 14.7. By 9 August 1999, our aerial survey found that flooding extended to river km 13.5 (Fig. 16), equivalent to a lakeshore regression of 1.2 km and a shallowing of about 3 m since September 1997. The Eureka River needs another 5 m of incision into the landslide dam to completely drain the lakes. With the decreasing depth of flooding, the downstream hazard of dam breach is diminished.

Geomorphic effects of the landslide dam

River km 9.8 to 12.4, and km 14.6 to 17.8, were traversed to assess the geomorphic implications of the landslide and dam. The area upstream of the dam was traversed to assess the thickness of lacustrine sediments, and to search for recent slope failures. Upstream of the landslide lake (river km

westward extension of the 1990 landslide, although it was not considered as such in the volume calculation. To the north of the slide are two slide-flows.

The 1990 landslide was a translational block slide since blocks 1, 3, and 4 exhibit no rotation. Because block 1 contains the colluvial deposits of three older landslides, the 1990 landslide can be classified as enlarged. The retrogression of blocks 2 and 3 is also accommodated by the enlarged descriptor. The style of activity would be multiple because the blocks are in contact with each other and likely share a rupture surface.

Groundwater conditions prior to the 1990 landslide are not known. However, it is likely that the water table was near the surface, since June 1990 had above-average rainfall (Environment Canada, Climate Services 1998; Environment Canada, Climate Services 1993). The landslide is therefore described as wet. The landslide is composed of earth, since more than 80% of the material is finer than coarse sand. Therefore, the 1990 Eureka River landslide was an enlarged, multiple, very rapid, wet, earth slide; it continues as a slow earth flow. A similar transition to a slow earth flow has occurred at the Edgerton landslide, Alberta (Cruden et al. 1995).

Fig. 13. Complex earth slide – earth flow from the south bank of the Eureka River (near river km 11.4). Instability at the toe of the landslide is common, due to the rapid post-landslide incision of the Eureka River into the landslide dam.

Fig. 14. A rotational earth slide at the landslide’s main scarp (near the start of the surveyed transect, looking east). This movement has occurred after the movement of the main body of the landslide (June 1990) and represents continuing retrogression of the landslide’s main scarp.

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Fig. 15. Interpretation of Alberta aerial photographs AS4333, 179 and 181 (6 October 1992), showing the extent of flooding due to the landslide dam (box delineates Figs. 8 and 9).
Fig. 16. View of the lake created by the landslide dam. The photo was taken in August 1999, at stream km 11.7, 100 m north of the Eureka River, looking upstream. The lake extends for 1.8 km and is approximately 5 m deep.

Fig. 17. Point bar aggradation of about 1.5 m at river km 10.6 (looking downstream). This sediment deposit consists mostly of rounded cobble-sized mud ball alluvium. Mud ball alluvium was seen downstream of other landslides in the region (Miller 2000).
14.6 to 17.8), thick vegetation obscured the lacustrine sediments in formerly flooded areas adjacent to the Eureka River channel. Within the Eureka River channel, upstream of the lake, only channel sediments (cobbles and boulders) were observed. One translational landslide with a rupture surface in till was seen at river km 17.8. In formerly flooded areas adjacent to the remaining lake (river km 11.5 to 12.4), lacustrine deposits of generally less than 0.1 m were observed. Thicker lacustrine deposits are expected in areas that remain flooded. Deltaic sediments were observed at the mouth of a tributary that enters the lake at km 11.8, the thickness of which could not be assessed.

Below the lake, along the length of the new stream (river km 10.8 to 11.5), streambed armour is mostly absent; thus, stream incision into the dam has been rapid. This rapid incision has promoted bank instability. Between river km 11.0 and 11.1, and between river km 11.4 and 11.5, instability at the toe of the landslide has consumed much of the pre-landslide channel, causing the reintroduction of the old channel’s alluvium (cobbles and boulders) into the new channel. This coarse alluvium was subsequently washed downstream, leaving the channel unprotected. Until armour is re-established, the streamed remains susceptible to erosion.

Downstream of the landslide dam, at river km 10.8, landslide-induced streambed aggradation exceeds 2 m. The thickness of this deposit decreases rapidly in the downstream direction (Fig. 17). Below river km 9.8, sedimentation was observed only on the lower portions of point bars (Alberta aerial photographs, AS4917, 23–26, 34–36, and AS3867, 64–67, 93–94).

Regional hazards

The Eureka River landslide contributes to our understanding of hazardous landslides associated with preglacial valleys. Each of the sedimentary deposits within the region (Fig. 5) is prone to instability (Miller and Cruden 2001; Miller 2000). However, landslides in till or preglacial lacustrine sediment, within preglacial valleys, are potentially greater hazards due to their size and tendency to form landslide dams. As the Eureka River follows the preglacial Shaftsbury Channel (Fig. 4), landslides within both till and preglacial lacustrine sediment can be expected along the river’s length, as each unit is destabilized by the degrading Eureka River (Miller and Cruden 2001).

Slope instability in the preglacial lacustrine sediment requires the greatest setback of structures from the valley’s edge. Landslides in this unit leave valley slope angles of about 10° (Figs. 3 and 10); therefore, permanent structures should be set back from the valley crest by an angle of 10° from the base of the preglacial lacustrine deposit. Where necessary, temporary structures might be constructed within this setback distance, depending on the expected life of the structure, and the rate of the upstream progress of large translational landslides in preglacial lacustrine sediment. The front of the upstream progression of the large landslides can be inferred from valley slope angles (Fig. 3) as being just downstream of the valley slope maxima.

Valley slope angles (Fig. 3) might guide the location of infrastructure that traverses this valley. As the landslides are prone to reactivate (Cruden et al. 1993; Hardy et al. 1962; Lu et al. 1998; Miller 2000), the construction of infrastructure on the gentle slopes left by landslides in preglacial lacustrine sediment is not advised. Infrastructure should be situated some distance upstream of the valley slope maxima.

Catastrophic landslide dam breach is also a hazard. The width of this dam, nearly 1 km, suggests that a dam breach would involve rapid incision into the dam during a flood. The lack of bed armour and the erodible nature of the preglacial lacustrine sediments, at the Eureka River landslide, makes this dam particularly susceptible to such a phenomenon.

Conclusions

Three surficial units were recognized at the Eureka River landslide, a postglacial lacustrine unit, till, and a preglacial lacustrine unit. The rupture surface of the 1990 landslide was within the preglacial lacustrine unit. The Eureka River is incising into sediment within a preglacial valley at the site of the 1990 landslide.

Aerial photographs show evidence of large prehistoric landslides in the lower Eureka River watershed to river km 15.2 on the south bank, and to river km 18.2 on the north bank. Here, valley slopes are about 10°, and the channel is mostly straight. Upstream of this reach, valley slopes steepen, and the river meanders. In the watershed the interval between large landslides in preglacial lacustrine sediment is at least 55 years, and likely substantially longer.

The June 1990 Eureka River landslide was an enlarged, multiple, wet earth slide, with an estimated volume of 50 Mm$^3$ and a friction angle of near 13°. Between 6 October 1992, and 1 May 1998, the main body of the landslide remained active, with movement of 25–30 m. Instability at the crown and toe of the landslide is ongoing. The likely trigger for the landslide was above-average precipitation in the spring of 1990.

The 1990 Eureka River landslide dammed the Eureka River between river km 10.7 and 11.7. The Type 6 landslide dam was, at its maximum, over 20 m high, creating a lake exceeding 8 km in length. By August 1999, the lake was reduced to a length of 1.8 km. The Eureka River has incised into the landslide dam another 5 m to completely drain the lake. The landslide removed the streamed armour from the channel, leaving the streamed susceptible to erosion.

Permanent structures should be set back from the valley crest by an angle of 10° from the base of the preglacial lacustrine deposit. Infrastructure might cross the valley upstream to avoid landslides that are prone to reactivate.

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References
Alberta, Forestry, Lands and Wildlife. 1991. Forest Management Unit Map Area, 1 : 100 000, Edmonton, AB.
Environment Canada, Climate Services. 1998. Canadian monthly climate data CD, Eureka River, Environment Canada, Ottawa, ON.