MASS-MOVEMENT DISTURBANCE REGIME LANDSCAPES, HAZARDS, AND WATER IMPLICATIONS: GRAND TETON NATIONAL PARK, WYOMING

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✦ ABSTRACT

The Teton Range is the result of active crustal extension (normal faulting) and is the youngest range in the Rocky Mountains at approximately 2 million years old. This makes it a particularly attractive landscape to study, especially in terms of landform development and morphology because of its youth, state of seismic activity, and its recent deglaciation. These factors have combined to produce a unique fluvial landscape in that the fault-shattered metamorphic/igneous rocks of the range have been/are being eroded from their source cliffs at high rates which has covered the glacially scoured valley floors with colluvium such as talus slopes, rock slide, avalanche, and debris flow deposits. This project was focused on the characterization of all forms of mass movement, especially rock slides, multiple talus types (rockfall, alluvial, avalanche), protalus lobes, protalus ramparts, lobate and tongue-shaped rock glaciers, and their collective effects on water retention and its late-season delivery in the Grand Teton National Park, WY. A major goal of this project was to reclassify many of the mass movements in the park in an effort to streamline and simplify previous efforts by other scientists. Methods used during this study included field reconnaissance and measurements acquired during the summers of 2010 and 2013 and measurements taken from various datasets (NAIP imagery, shape files used within a GIS [ArcMap 10.0], and Google Earth™). Mass movement deposits, as well as ice glaciers and long-term snowbanks, were mapped and interpreted. Overall conclusions are that the major sources of mass movements from the Archean crystalline core of the range are the result of extensive jointing, fault-shattering, increased frost-wedging at higher altitudes, slopes steepened by prior glacial erosion, and extensive snow avalanches. Areas of Paleozoic sedimentary rocks marginal to the crystalline core produce rockslides as a result of steep dips and unstable shales beneath massive overlying carbonates. The presence of internal ground ice enables development of protalus lobes, thicker rock-fragment flows, and thinner boulder streams. Such ground ice is likely to enhance late-season water delivery downstream unless climate warming and recurrent droughts become too extreme.

✦ INTRODUCTION

The role of rockslides-rock avalanches in mountain landscapes has been well documented in Himalayan, Alpine, and some Rocky Mountain regions (Hewitt 2006, Shroder 1998a,b, Shroder and Bishop 1998, Shroder et al. 1999, 2010, 2011). Although the abrupt slope failures themselves are extremely short-lived events, the persistence of rock-wall detachment scars and the various deposits themselves can persist for long periods as influences on water diversion and impoundment, fluvial and lacustrine sediment retention, sediment H_20 storage as fluid (water) and solid (ice), and the release of occasional landslide lake outburst floods (LLOF). In addition the more incremental or slower accumulation of collateral mass-movement deposits, such as talus, alpine debris flow cones, snow avalanches, and some rock glaciers and boulder streams with internal ice permafrost also can have both similar and different effects.

The Teton Range is an area of high relief, containing high-gradient drainage and amplified rates of, often catastrophic, colluvial activity. High altitude snowfall (~2 km) is generated as moist Pacific air masses are uplifted over the mountain barrier (Foster et al. 2010). These streams are essentially derived from melt-waters of fresh annual snowpack, firn fields, and remnant glaciers of the late Pleistocene. Because the area is nearly deglaciated, excepting small vestiges of past sizable glaciers (the Teton glacier, etc.), Teton Range trunk-streams are "misfits" or "underfit", in that they are presently too small to have created the valleys they occupy (Huggett 2007). Mass movement processes such as debris flows, avalanches, and rockfalls / rockslides occur at extremely high rates in Teton Range. This high-frequency of mass movement is largely due to tectonic and climatic forcing, coupled with the effects of variable rock types and weathering, and post-glacial debuttressing of valley walls. These colluvial processes further complicate / perturb the "misfit" fluvial regime with high and frequent debris / rock inputs which overload or even dam streams, causing epicyclical responses (Hewitt, 2006) to the constant / chaotic inputs of mass movement.

This situation of continual or extended disruption of a fluvial regime has been discussed by Hewitt (2006) as a "disturbance regime", which refers to "the long-term or permanent consequences of relatively brief, but reoccurring episodes, usually of high magnitude or at critical sites", which produce landforms "whose location and history and are dependent on the disturbances" (Hewitt 2006, p367). According to Hewitt (2006, quoting Benda et al. 2003), the landforms of a disturbance regime "tend to be polygenetic and exhibit morphological heterogeneity", also called "patchy heterogeneous morphologies" (Burchstead et al. 2010). Though Hewitt was mostly concerned with mega-scale rock avalanches, or sturzstrom (Shroder and Weihs 2010), the same phases of the landslide interruption epicycle are observable in Teton canyons.

Teton Range canyons / streams fit under the classification of "a disturbance regime landscape", in that there is an abundance of frequently occurring mass movements that are clearly affecting the centerlines of trunk-streams. It is the primary goal of this paper to describe and explain selected examples of trunk-stream positions in terms of their incongruence / congruence with glacial trough centerlines, and the disturbances which have affected their locations in the Teton Range.

Problem statement

Studies concerning the role of various mass movements and their effects on water delivery are well documented in other ranges such as the Himalaya or other Rocky Mountain ranges; however, colluvial stream disturbances have not been explained or described for the Teton Range.

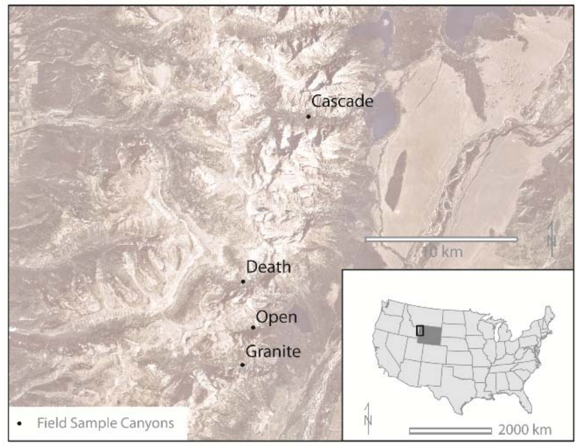

Figure 1. Study area canyons of Grand Teton National Park.

Purpose and objectives

The purposes and objectives of this study were to:

1. Begin with Marston et al.'s (2010) prior funded seminal work for UW-NPS Research Center on cross-valley profiles and mass-movement hazards in deglaciated canyons of Grand Teton National Park as a fundamental basis for further geomorphological mapping.

2. Apply Hewitt's (2006) methodology for recognizing landslide interruption epicycles in mountain valleys in the Grand Teton National Park, specifically Cascade, Death, Open and Granite canyons (Figure 1).

3. Adapt the landslide interruption epicycles of Hewitt (2006) to small discharge canyon streams in Grand Teton National Park as paradigmic examples of mass-movement control of small-basin alpine drainages (Shroder et al. 1999).

4. Develop an overview of ground-water and ground-ice retention in Teton alpine valleys relative to mass-movement deposits as porous and permeable aquifers, as well as partial aquitards to surface flow.

Significance

Mass movements, glaciers, and rock glaciers in GTNP affect water delivery throughout the park, can pose hazards to park users, and are known to create long and short term changes in the park landscape. Because mass movements, glaciers, and rock glaciers largely dictate water releases downstream, it is imperative that data be created about their character and the manner in which they affect the fluvial regime. Additionally, because these features are often hazardous to humans, studies that investigate their genesis and continued evolution are warranted in the interest of park safety.

✦ BACKGROUND

The Teton Range is the result of active crustal extension or normal faulting, and is one of the youngest ranges of the Rocky Mountains. Although movement on the Teton Fault began ~9 mya, the Teton Range itself is essentially largely Quaternary-aged (i.e. perhaps only ~2 million years old) (Smith et al. 1993, Byrd et al. 1994). Fault scarps and associated stratigraphy reveal Holocene slip rates of 0.45 – 1.6 mm/yr for the Teton Fault that are consistent with a range of recurrence intervals for large, scarp-forming earthquakes of 1600 to 6000 yr. Hampel et al. (2007) have even hypothesized that the post-glacial slip rates on the Teton Fault increased as a result of melting of the Yellowstone ice cap and deglaciation of the Teton range. The influence of high ground accelerations associated with large earthquakes was judged as particularly important in the Teton region by Smith et al. (1993).

Mass movements in the Teton Range are profuse, and much of the valley floors of the range are blanketed by the deposits of various forms of mass movements that occur above (rockfalls, rockslides, debris flows, avalanche, etc.). Case (1989), in his original aerial photographic mapping of landslides of Wyoming, was able to delineate the greater proportion of the state's slope failures, but given the massive size of the undertaking without enough (or any) field checking for accuracy in mapping, inevitable omissions or errors accumulated. Mass movements come in a great many variable types, which over time, weathering, soil formation, and revegetation, can become obscured as to genesis. The prior work of Marston et al. (2010), which used Case (1989) as an initial base map in five canyons (Paintbrush, Cascade, Garnet, Death, Granite), provides a foundation for mapping of the existing deposits which were omitted/not recognized, misclassified, and/or delineated inaccurately. According to Marston et al (2010), because a propensity of mass movement deposits in GTNP are found below Pinedale glacier trimline positions, glacial debuttressing has played a large role in mass movement deposition since the most recent glaciation. Already in the European Alps, the western Himalaya and elsewhere, degradation of rockwall permafrost is thought to have caused some rock failure such that formerly frozen surficial fracture systems no longer retain bearing strength, thus spontaneously failing (Arsenault and Meigs 2005, Cossart et al. 2008, Stoffel and Huggel 2012).

The Pleistocene glacial history of the Teton region has been worked out in fair detail by Pierce and Good (1992), Licciardi and Pierce (2008) and a number of others. Old glaciations, perhaps from the early and middle Pleistocene, are known to have occurred in the Jackson Hole region but their record is poor and whether or not they occurred in the rising but lower altitude Tetons of the time is unknown. The oldest, fairly well understood Munger (Bull Lake?) filled all of the Jackson Hole valley with ice from about 157 to 151 ka, until about 136 ka (Pierce and Good 1992, Licciardi and Pierce 2008); Precambrian crystalline rocks from the Tetons in the tills and moraines in Jackson Hole show it is likely that cirques and valleys from the Tetons were also filled with ice at this time. During the succeeding Pinedale glacial, ice advanced from highlands into Jackson Hole from the east, northeast, and north, perhaps because of storms moving up through the Snake River Plain in the Idaho region and building up the Yellowstone ice cap. Glaciers of the Teton were smaller, although they did join the south-flowing Snake River Lobe of that ice north of Jenny Lake. To the south of there, however, end moraines of valley glaciers from the Tetons extended no more than 2.5 km beyond the mountain front. The precise timing of the Pinedale glaciers from the Tetons has recently been constructed with cosmogenic dating of boulders and glacially striated bedrock at about 14 ka (Licciardi and Pierce 2008), but some radiocarbon dates indicate Pinedale ice as late as 9000 yr, perhaps persisting to warm altithermal time around 6000 years ago (Love et al. 2003) when it would have all melted away in most of the lower Teton Valleys. This 14,000 to 9000 year estimation of the melting away of Pinedale ice in Late Pleistocene to Early Holocene time is important because it would be the approximate beginning time for emplacement of the post-glacial, mass-movement debris in the valleys of the Teton landscape. Prior to that time, of course, all mass-movement debris from the valley sides would have fallen on the glacier ice of the time and been transported to the mountain front where it would have contributed to the end moraines encircling the lakes of Jenny, Bradley, Taggart, and Phelps, as well as the mouths of Glacier Gulch and Open Canyon (Love et al. 2003).

The relative youth of the Tetons makes them a particularly useful place to study in terms of landform development and morphology, recent deglaciation, and state of seismic activity, which can be expected to produce significant mass movement through valley wall deglacial debuttressing as the last of the Pinedale ice melted away, as well as horizontal seismic accelerations to the glacially oversteepened valley walls. Foster et al. (2010) noted that the tall, steep hillslopes of the Teton canyons provided shading, as well as plentiful snow influx from avalanching, and insulating debris cover from rockfalls to the valley floor glaciers. Tranel et al. (2011) found that the large talus and mass-movement aprons in some places indicate that the interglacial alpine fluvial system is locally transport-limited and unable to keep pace with the mass movement. The thick mass-movement fans may also enhance chemical weathering of the bedrock beneath them by trapped ground water which would have thereby facilitated periodic glacial incision or fluvial transport during peak flow conditions.

These factors have combined to produce a unique post-glacial, fluvial landscape in that the fault-shattered, metamorphic and igneous, crystalline rocks of the range have been/are being eroded from their source cliffs at high rates that have covered the glacially scoured valley floors with colluvium such as talus slopes, rockslides, avalanche debris, and debris flow deposits. These colluvial deposits have apparently caused the centers of "trunk" stream channels to be variably incongruent with the glacially eroded valley centers.

✦ STUDY AREA

Location

The Teton Range is located in northwestern Wyoming at about latitude 43 º 46' 46" N longitude 110º 50' 10" W. The range is approximately 72 km long and 20 km wide and is flanked by the south-flowing Snake River ~10 km to the east and by the Wyoming / Idaho state border ~5km to the west (Figure 1). The Tetons are the first topographic barrier to moist Pacific air masses moving through the western Snake River Plain (Foster et al. 2010). These conditions favor orographic uplift of moist air masses that in turn drop large (modeled at >200) amounts of precipitation high in the range (Foster et al. 2010). This situation has created "elevated peaks high above the surrounding topography" (Foster et al. 2010), or "topographic lightning rods" (Brozović et al. 1997, Foster et al. 2010).

Quaternary chronology

The chronology of glacial and other landforming events in the Tetons and Jackson Hole area has been evaluated by a succession of scientists who were impressed with the abundance of glacial, fluvial, mass-movement, and faulted landforms that occur in the area. Plentiful cirques high in the Tetons, glacierized and glaciated valleys with striking parabolic cross sections, prominent terminal and recessional moraines, strong river terraces, extensive and pervasive, polygenetic talus slopes and rockslides, and fresh fault scarps at the front of the Teton Range that truncate lateral moraines; all of these landforms have generated the impressive and iconic landscape of the Grand Teton National Park in western Wyoming. When the first government-sponsored surveys first came to the region from the 1870s to 1900, early researchers noted that glaciers had produced piedmont lakes impounded by morainal material (Bradley, 1973), and the fluvial terraces along the Snake River were related to glacial outwash (St. John 1877). Blackwelder's classic work in 1915 established the Buffalo, Bull Lake and Pinedale glacial stages from recognition of moraines near Togwotee Pass into Jackson Hole and glaciated valleys on the north and south sides of the nearby Wind River Range. Fryxell (1930, 1935) then successfully applied Blackwelder's glacier-stage nomenclature to the Tetons and Jackson Hole area and his application of the terminology has stood the test of time.

Most recently Pierce and Good (1992) and Licciardi and Pierce (2008) carried the Quaternary chronologies forward and updated them with fairly precise cosmogenic exposure-ages of the Bull Lake and Pinedale glaciations in the Teton Range and Jackson Hole areas. One of the primary changes between the older work and the newer is the attribution of a lesser role to glaciers from the Teton Range and an enhanced role to the southern part of the Yellowstone ice mass. Boulders that occur along the southern limit of the penultimate ice advance in Jackson Hole give erosion corrected ages of 136±13 ka, with oldest ages of 151-157 ka that correlate with Marine Isotope Stage (MIS 6), as well as with Bull Lake glacial deposits of the Wind River range and West Yellowstone. The timing of the next major glacial advance in the region is the Pinedale (MIS 2) with maxima varying from ~18.8 to ~16.5 ka, to possibly 14.6ka around the southern margins at Jenny Lake. Late Pinedale events in the Teton – Jackson Hole region suggest that major advances or stillstands of a large valley glacier from the east flank of the Teton Range occurred > 6kyr after the global Last Glacial Maximum (LGM) just prior to the younger

Dryas, followed by full and rapid deglaciation of the whole Yellowstone Plateau.

An important observation is the strong climatic gradients observed between the floor of Jackson Hole to the east and deeper into the mountains to the west as a reflection of the westerlies and the resulting rain-shadow gradients. For example (Pierce and Good 1992), the mean annual precipitation at Jackson is only about 4.3 cm, whereas only 10 km west at Wilson it is about 10 cm.

✦ METHODS

Field reconnaissance

Each mass movement deposit visited in the field was visually observed and photo documented. For some deposits, such as the Rendezvous rockslide, observing both the headscarp and toe was possible (because it was accessible), providing more comprehensive field data for those deposits. Slope angles of deposits were measured using an Abney hand level where vegetation was permitting. Dipping strata were measured with a Brunton pocket transit. Water output from mass movements was measured qualitatively where possible, such as return flow from a landslide's toe.

Geomorphological mapping

Combined with data collected in the field, geomorphological mapping was concluded in the laboratory using ArMap 10.0. Several data were used to map/delineate and describe landforms under study. These data included (but were not limited to):

- National Areal Imagery Program (NAIP) imagery (2012, 2009) ~1m resolution
- National Elevation Dataset (NED) ~10m resolution
- Geologic Map by Love et al. (1992)
- Landslides of Wyoming shapefile by Case (1989)
- Google Earth™ imagery

In general, field and lab-discovered mass movements, protalus lobes, glaciers, rock glaciers, and slow rock fragment flow deposits were digitized using NAIP imagery and aided by Google Earth™. The digitized outputs (a shapefile) were then accoutremented with attribute data such as mean elevation and mean deposit aspect. This attribute data was subsequently used in Excel 2010 to produce summary statistics of the various deposits.

Landslide interruption cycle identification

The Hewitt (2006) landslide interruption cycle of five phases, each with variable sediment assemblages, constitutes the *landslide-interrupted valley landsystem,* which will be utilized as the theoretical background for the primary mapping tool. As Hewitt (2006) has noted, this type of landsystem creates naturally fragmented fluvial systems in which a *disturbance regime geomorphology* can be identified. The five phases to be identified are: (1) *Landslide complex*, with associated mass movement emplacement forms (rock slide, rock fall, debris flow, etc.) (Shroder et al. 2005); (2) *Impoundment complex*, with associated aggradation and constructional landforms upstream of the barrier, and possible downstream erosion and/or sedimentation; (3) *Degrading interruption complex*, with trenching and removal of the impoundment complex and downstream sedimentation; (4) *Superimposed interruption complex*, with exhumation of buried valley fill and incision into pre-landslide valley floor; and (5) *"Shadow" interruption complex*, with minor but persistent legacies of interruption, mainly bedrock forms.

The application of Hewitt's (2006) landslide interruption cycles to landforms in the Teton Range involved:

- Modifying of the landslide interrupted valley landsystem to reflect the different nature of the Teton mountain valleys from those in the western Himalaya with large-discharge rivers in them where the landsystem categorization was first developed.

- Differentiating landslide impoundments from paleo-beaver dam impoundments, or even paleo-beaver dam modification, or augmentation of a prior mass-movement impoundment.

- Mapping the locations of streams running over bedrock, as opposed to other kinds of sediment covering bedrock where possible in the field.

- Differentiating between the kinds and sources of the sediment covering the bedrock.

- Insofar as possible, estimate grain size, thickness, and porosity/permeability by visual inspection as well as to enable a first approximation of hydraulic retention and conductivity for ground water estimations.

- Estimate ground ice on the basis of surface morphology (rock-glacier and boulder lobe characters and profiles).

✦ PRELIMINARY RESULTS AND DISCUSSION

Geomorphological mapping

Efforts to expand and correct/reclassify previous mapping done by Case (1989) and Marston et al. (2010) resulted in many new, or corrected, delineated deposits. This study identified 57 features (glaciers [10], rock glaciers [20], protalus lobes [22], slow rock fragment flows [3], rockfall talus [1], and a single slump block). Some of these features were not previously mapped by Case (1989) or Marston et al. (2010) (2 protalus lobes and 1 rock glacier). There were also some features from Case (1989) that required reclassification based on field and laboratory findings. These reclassifications were performed for protalus lobes, rock glaciers, and slow rock fragment flows that were purported by Case (1989) to be something different. In total, there were 12 deposits in which there were discrepancies between Case's (1989) and this study's findings. Because of the use of DEMs, high resolution imagery of various dates and digital globes such as Google Earth™, our findings are more resolute than Case's (1989) work, which was limited by the use of aerial photography and sporadic fieldwork, if any.

Landslide interruption epicycles

Our preliminary assessments of the canyons selected for this study indicate a number of examples of phenomena which will be useful in applying the Hewitt (2006) designations. For example, Phase One (emplacement) and Phase Two (impoundment) can be observed in Cascade Canyon where a debris flow has

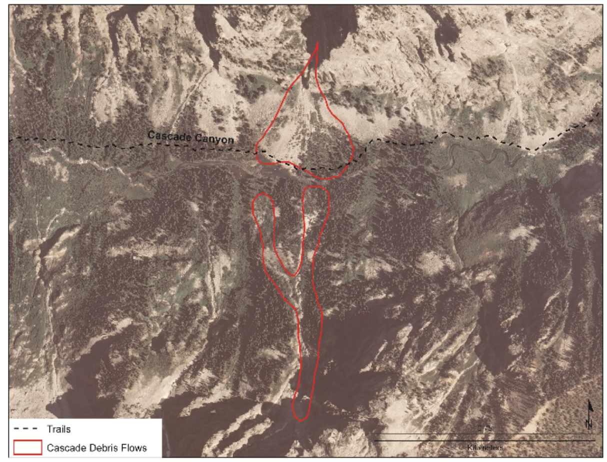

Figure 2. Planimetric view of opposing debris flow deposits and subsequent impoundment in Cascade Canyon using NAIP aerial photo (2009).

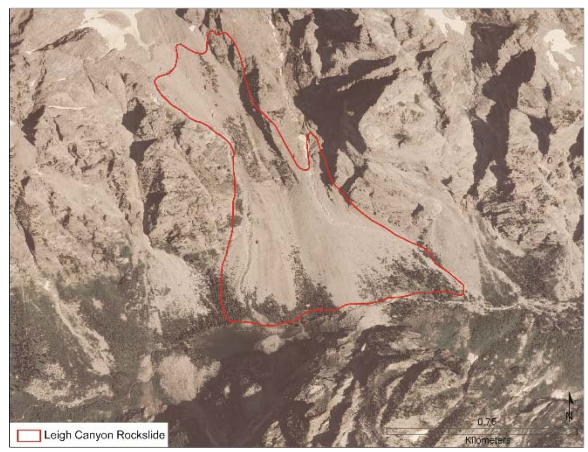

Figure 3. Planimetric view of a rockslide deposit and subsequent impoundment in Leigh Canyon using NAIP aerial photo (2009).

emplacement of the landslide, or limited discharges may have precluded significant downcutting and lateral erosion to take parts of the valleys into the more advanced stages of Hewitt's classification scheme. The Rendezvous Rockslide in Granite Canyon (Figure 4) may fit some of the criteria for Phase Four in that the impoundment area has been drained and slightly incised; however, the incision is limited to the lake sediments and not the bedrock beneath. No evidence of Phase Five (epigenetic gorges) were observed during this study. It is likely that the scales of Teton processes (such as mass movements and water discharge) and time since the most recent deglaciation have not been of sufficient magnitude to produce the landforms that Hewitt (2006) found in the much older and larger-scale Himalayas.

Rendezvous rockslide

The Rendezvous Rockslide is a large slope failure on the northwest-facing side of Rendezvous Mountain above Teton Village near Jackson Hole, WY. It can be accessed from the top by an arduous and dangerous descent from the tramway ride stop at the top of the ski area, or from the trail up Granite Canyon to the toe of the failure in Granite Creek. The slope failure descends from an altitude of about 9600 feet (2925 m) on a generally northwestern trajectory, down to an altitude of about 7880 feet (2400 m) in Granite Canyon, which is a descent of about 1720 feet (525 m) vertically downward in about 1.2 mi (2 km) downslope (Figures 4 and 5).

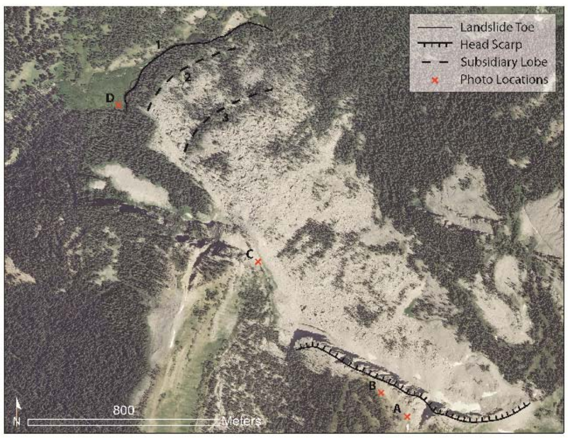

Figure 4. Planimetric view of Rendezvous Rockslide using NAIP aerial photo (2009). Adjacent photos are located on map with red X's. Several lobes (1, 2, 3) are thought to represent at least three separate failure events contributing to this deposit.

The bedrock geology map of the region (Love et al. 1992) shows that the direct or proximate cause of the landslide was the presence of the massive carbonates of the Gallatin Limestone, Bighorn Dolomite, and the Darby Formation that occur above and overload the unstable Park Shale Member of the Gros Ventre formation (Love et al. 2007). The combination of the weight of these overlying formations totaling some 980 feet (300 m) of thickness, as well as the ~20° angle of dip to the northwest down into the Granite Canyon valley, and the abundant precipitation and high freeze and thaw of the mountain environment have combined to produce the slope failure.

The uppermost main scarp of the slope failure at the top of the Rendezvous Mountain ridge has been eroded down to a low rounded ridge at about 9560 feet (2914 m) that is only about 160 feet (48 m) high, but on the nearby left (west) flank of the landslide the huge cliffs of Gallatin Limestone, and Bighorn Dolomite rise some 600 feet (180 m) above the surface of the landslide, where major rock blocks have been provided to the failure. Furthermore, about halfway down the lateral west margin of the landslide, a second major lateral scarp cliff of dominantly carbonate rocks of the Darby Formation also rises a similar amount in huge cliffs.

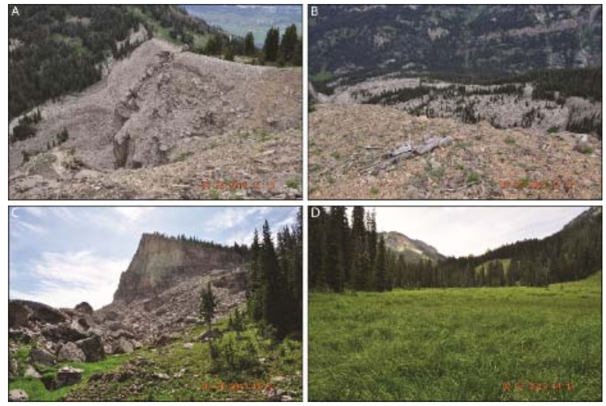

Figure 5. Photos taken at Rendezvous Rockslide. Photos A and B taken at head scarp. Photo C was taken at midslope and margin of landslide. Photo D was taken at the landslide toe, which caused impoundment of Granite Creek and the deposition of sediments in this now-drained lake.

The overall surface of the landslide mass is a hummocky rolling terrain, with several undrained depressions, one of which holds an intermittent small lake and grassy area in the upper third of the mass, about 580 m down from the top of the slide. The depression appears to hold water for at least part of the year as a result of the presence of impermeable shale remnant masses that outcrop around the depression. The landslide measures about 2 km in length and is about half a km wide in the middle and lower sections, although it is only about 0.2 km wide at the narrower top. The uppermost part has a considerable amount of remnant shale exposed in heaps and mounds where it has been forced by the presumably repetitive slope movements. The lower two thirds of the landslide mass has plentiful large boulders, ten of which were measured on Google Earth™ and found to average ~21 m in diagonal long axis. Midway down the slope failure on the left (southwest) flank where a change in movement direction occurs from WNW to NW, two offsetting shear planes daylight at the margins to form two stepped terraces, each about 10 m high, the risers of which are the surface manifestation of the shear planes. It is apparent that the main mass of the slope failure has subsided here to show the shear planes.

The lowermost toe of the Rendezvous Rockslide is estimated to be at least 60-80 m thick (perhaps thicker), and appeared to be comprised of at least two main episodes of movement, one above and overriding the other. The lowermost part of the toe was composed of large boulders, the interstices of which appeared filled with weathered debris and soil, whereas the uppermost part of the toe appeared to have a dominance of open matrices, with far less weathering degradation. This lowermost and apparently older toe is conspicuous in having a plentiful forest cover of large, old-growth conifers that are well-rooted in the weathered soils that have developed throughout the lower portion of the mass. Although no date estimates could be obtained from soil thickness measurements in the limited time available for investigation, it is possible that these two movement lobes may represent dominant movement during times of greater precipitation and/or permafrost degradation at the end of the Pleistocene or early Holocene. Presumably, the slope failure would not have been active because it was frozen during Pinedale glaciation and any prior mass movement would have been removed by glacier ice in Granite Canyon, but in the subsequent melting away of Pinedale ice, and the unstable climate associated with the Little Ice Age in late Holocene, other movement could have occurred.

The Rendezvous Rockslide was low velocity as it moved slowly and most probably intermittently down into Granite Canyon but did not climb the north side of the canyon at all. In all probability the mass was comprised of many small and independent motions that collectively brought successive masses of rock fragments of the Paleozoic rock fragments to lower positions in the valley over multiple years of rock motion, more in wetter years, and less in drier.
The hydraulic associations of the Rendezvous Rockslide are unknown, although it is likely that almost all of the precipitation that falls upon the feature infiltrates downward through the profuse open porosity between the plentiful boulders and becomes groundwater. Much of this water would have been absorbed by the shales lower down and contributed to their plasticity. The basal slip surface(s?) of the landslide appear to be largely on the uppermost Death Canyon Limestone Member of the Gros Ventre Formation.

The overall slope of the outer landslide surface is about 13°, which compares with the generalized dip of the bedrock at this location of ~20°. The ~7° difference in amounts can be explained by the landslide accumulating mass in the valley below, which would reduce the overall surficial gradient. It is thus apparent that the cause of the slope failure was the incompetence of the Park Shale, which permitted the motion of the landslide to run largely along the bedding planes at the top of the Death Canyon Limestone. This latter rock unit forms a prominent planar shelf of resistant bedrock that is characteristic of much of the high topography elsewhere on Rendezvous Mountain (shown in Figure 30 of Love et al. 2007), as well as the head of Death Canyon, where it forms a prominent landform of conspicuous size that is known as the Death Canyon Shelf. This feature extends southwest-northeast laterally within the Grand Teton National Park for over 5 km, averaging some 300-400 m in width with conspicuous rockfall talus and protalus lobes along it with conspicuous rockfall talus cones and protalus lobes and ramparts that have developed from the rock fragments emanating from extensive freeze and thaw of the rock cliffs of Big Horn Dolomite and Darby Formation rocks outcropping along it.

The toe of the Rendezvous Rockslide partially blocked the course of Granite Creek that runs in the bottom of Granite Canyon, which caused upstream aggradation and the deposition of a flat plain of finer clastics (gravels, sands, silts, and clays). The aggradational plain extends upstream for ~450m and ~200 m in width and is covered with sedges, grasses, and willows in between the multiple channels of Granite Creek that have developed over the aggradational plain. In the bottom of Granite Creek stream channel, swirls of ascending sand-grain plumes attest to water under pressure from below that is rising into the channel from shallow groundwater base flow that comes in from uphill water sources, probably from both sides of the canyon; landslide source on the southwest side, as well as the opposite northwest side where water can also infiltrate into the low gradient slopes there.

The Rendezvous Rockslide is a reasonably typical type of slope failure that is fairly characteristic all over various parts of Wyoming in other uplifted mountain terrains where the carbonates and shales of the Lower Paleozoic produce similar mass-movement landform features. For example, at the head of Shell Canyon in the western Bighorn Range, a similar suite of rocks has produced even larger and more pronounced failures of the bedrock, with extensive glide blocks, rockslides, and slow-flow features under similar structural conditions of dipping carbonates and shales at moderate to higher altitudes.

Alpine debris cones and fans

The alpine debris cones and debris fans emplaced in the many canyons of the Grand Teton National Park have been formed by a variety of processes that require considerable elucidation to delineate process and form in such a fashion as to serve as explanatory templates for similar assessments in other mountain ranges. Certainly such work has been undertaken in many other environments such as in the European Alps, the Canadian Rockies, the Colorado Rockies, the Himalaya, and elsewhere, but still a refresher look at the features can be instructive, particularly where common polygenetic overprinting of process and form can cause misinterpretation. All of the alpine mass movement forms can constitute a variety of talus accumulations that originate solely, or by some combination of falling rock fragments, running water from rainstorms or snow melt on mountain slopes, or by snow avalanches (White 1981). These accumulations of alpine debris form in cones or fans at various angles of repose, from the steep ones at >35° - 45° for **rockfall talus**, to 35° - 38° for the upper slopes <28° for the talus toes of **alluvial talus**, and <25° for **avalanche talus**, which also has gently concave-up toes where the snow avalanches sweep out. In addition, talus subtypes occur of the more strongly developed **avalanche boulder tongues** or avalanche roadbank tongues, and the **protalus ramparts** with rock fragments that move out over snow banks to pile up as ridges at the base of talus slopes (White 1981). All of these constitute what can be quite polygenetic forms of alpine talus that is emplaced in what has been considered a reasonably non-catastrophic fashion, wherein most of the rock clasts come down as isolated unit rockfalls and slides, or are emplaced by snow avalanches, or by isolated rivulets in rainstorms (Figures 6 and 7).

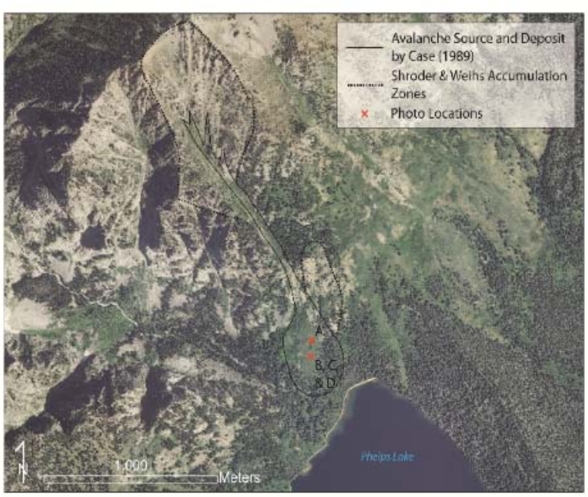

Figure 6. Planimetric view of Death Canyon avalanche source and deposit using NAIP aerial photo (2009). Avalanche polygon delineated by Case (1989) (solid line) was modified in this study (dashed line). Adjacent photos are located on map with red X's.

Figure 7. Photos taken in and near Death Canyon of an avalanche source and deposit. Figure A shows source area. Figure B shows large boulder transported during 2010-2011 avalanche season. Figures C and D show still growing downed trees beneath boulder and on deposit area.

This reasonably non-catastrophic emplacement is contrasted with the truly catastrophic, very rapid, and large forms such as the larger rockfall debris streams, rock avalanches, or sturzstroms, rock slides, debris slides, or the large slump-earthflows and large slow debris flows that are so characteristic of areas with extensive and weak sedimentary rocks.

The Teton Range has a sedimentary rock cover of Lower Paleozoic shales and carbonates to the west and south that lap onto the Precambrian crystalline rocks at the core of the upfaulted block that constitutes the highest portions of the range (Figure 8). These sedimentary rocks, with their unstable shales beneath massive carbonates, are the main source of rock rubble that makes up the widest variety of mass movement in the outer fringes of Grant Teton National Park. In the high core of the range, however, the crystalline rocks of gneisses, granites, amphibolites, metagabbros and the like are not as conducive to the formation of large mass movements as are the less stable sedimentary rocks, but still the extensive jointing, fault shattering, steep slopes, and higher altitudes that promote considerable freeze and thaw can also contribute large quantities of rock rubble that are commonly remobilized into extensive talus cones and fans, as well as the glacier moraines, rock glaciers, protalus lobes, and the like that characterize much of the range. Unlike the Himalaya and Hindu Kush where the relief and steep slopes are so much greater and high seismicity and massive rock avalanches from crystalline rocks abound (Shroder and Weihs 2010, Shroder et al. 2011), large rock avalanches have not been mapped in the Tetons, although a variety of large failures of sedimentary rocks are known. The nearest known seismically induced rock avalanche or rapid rockslide was the one across the Madison River in 1959 125 km NNW of the Teton Range that was

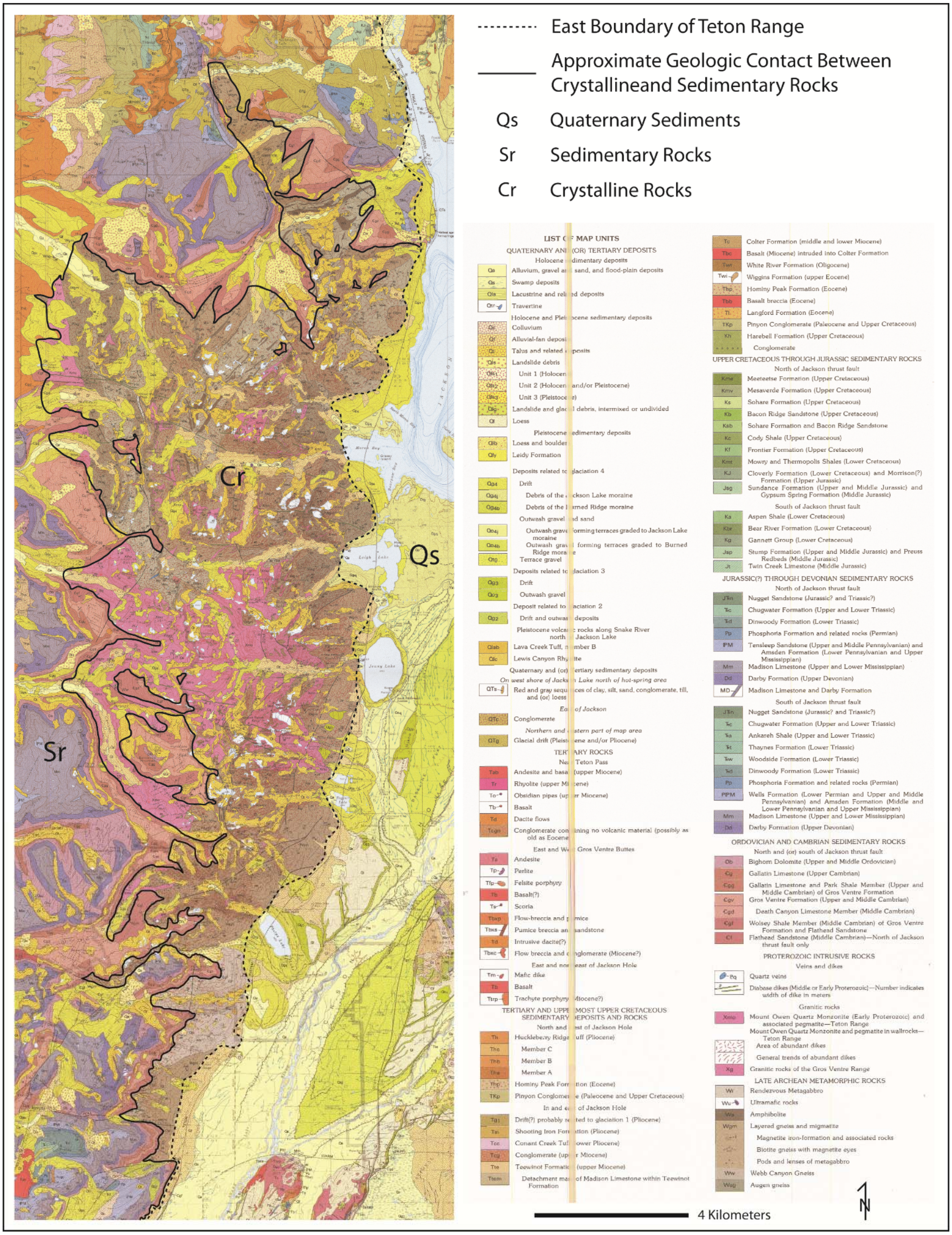

Figure 8. Geologic map of the Teton Range. Modified from Love et al. (1992).

produced by a magnitude 7.5 earthquake. The well-known Gros Ventre Rockslide of 1925 that occurred just 20 km east of the Teton Range was caused mainly by increased shear stress exerted from excess snowmelt and rainwater loading, coupled with the Tensleep Sandstone overlying Amsden shale dipping into the Gros Ventre River valley where the river had previously undercut the slope and thereby reduced shear resistance. Numerous other slump, slow debris flows and earthflows are known in the general region as well, mainly from combinations of weak and dipping sedimentary rocks and plentiful precipitation that loads slopes with heavy surcharge weights, dissolves cements, causes extensive freeze and thaw, brings in seepage pressures, and causes friable shale to lose strength, which collectively results in many slope failures.

Glaciers and rock glaciers

Glaciers are masses of permanent snow and ice that move slowly downhill under the influence of gravity, and rock glaciers are similar masses of ice that are completely or largely covered with rock fragments, which obscure the ice cores or interstitial ice cements beneath. Rock glaciers are distinguished from other forms of alpine debris by having steep fronts at the angle of repose that are caused by the top moving faster than the base, with the result that the rock fragments pile up steeply at the front (Giardino et al. 1987). The formation of glaciers occurs as winter snows accumulate to such an extent that all of the winter's accumulation does not melt entirely away each summer, and instead some snow converts to crystalline firn or pellets of ice that last through the melt season into the next winter. After the passage of some 3 – 5 years, the mass of stored ice can become sufficiently large so as to begin flow downhill as an ice glacier. Rock glaciers can begin formation as either masses of rock fragments into which meltwaters drain and refreeze to reconstitute as interstitial ice cements. They can also begin as masses of snow and ice in snow banks, or even ice glaciers that become covered with so much rock debris that the original snowbank or glacial origin can be lost and the masses come to move downslope through various causes of freeze and thaw or permafrost creep, but always with steep fronts of rock fragments at the angle of repose. Talus slopes can fill with such expanding interstitial ice to bulge out at the bottom as protalus lobes, which if developed to an extent that the projection of the talus slope down to the horizontal is greater than the projecton of the lobate front down to that same plane, then the feature is considered to be a protalus lobe, but if the lobate

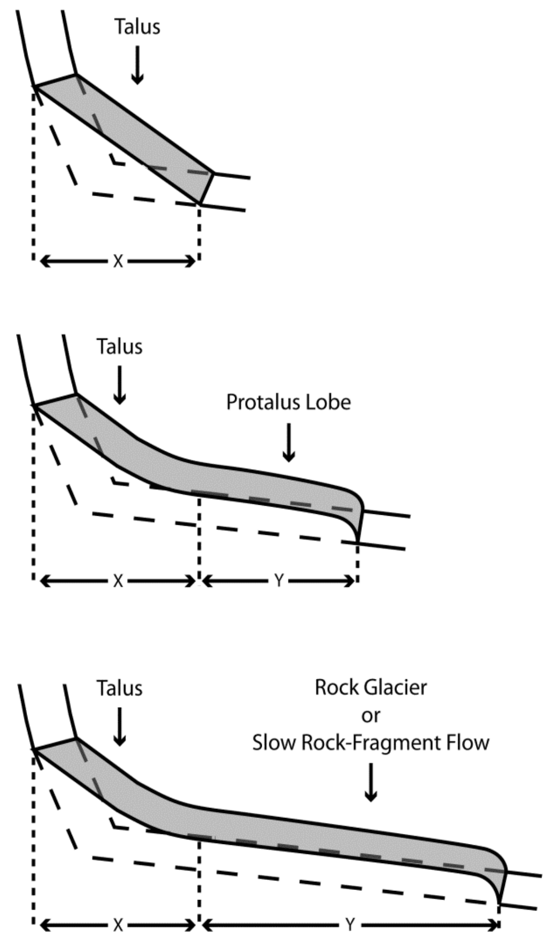

Figure 9. Generalized graphic of talus slopes, protalus lobes, and rock glaciers or slow rock-fragment flows.

front bulges out and travels further than the projected talus extent, then the feature is a lobate rock glacier (Figure 9). On the other hand, the ice-cored rock glaciers can commonly form from the overall downwasting of the ice glaciers until all, or almost all, of the original ice is covered up and the true glacial nature of the feature is lost. A renewal of cold temperatures or development of greater refreezing of the down-trickling meltwaters from snow melt or the like can then cause the ice core to push the terminus of the feature forward again so that a steep front of rock fragments at the angle of repose develops, at which point a tongue-shaped rock glacier with an ice core has formed.

Because both glaciers and rock glaciers represent ice resources stored away for different amounts of time until finally melted, they constitute important sources of meltwater when finally converted from solid into liquid at some point in their history.

Thus in Grand Teton National Park, it is important to better understand the hydrologic cycle as it is constituted in the region, especially if this hydrologic cycle changes greatly in coming years due to climate warming and drying.

Most of the glaciers recognized in the Teton Range have a north or east aspect, with the exception of Falling Ice Glacier on Mount Moran, which faces to the southeast. All these recognized ice glaciers occur on the east side of the range (Fryxell 1935), which is no doubt due primarily to the westerly winds that blow snow over the range to form east-facing cornices that avalanche down to form the glaciers on that side, as well as to decreased solar radiation on the east sides of the range as the sun is more intensive late in the day in the west. Many of the tongue-shaped rock glaciers on the high peaks of the Teton Range, on the other hand, occur on the west and north sides where snow does not accumulate as much as to the east, although the rock fragments from high freeze and thaw are still abundant, and the altitude is great enough and the topographic and rock-fragment shielding is enough so that the ice does not melt away (Figure 10). These tongue-shaped rock glaciers show by their occurrence in deep cirques that they were indeed once ice glaciers, and their lobate nature, with many longitudinal ridges and furrows and steep fronts at the angle of repose show that they remain active.

Glaciers in the Teton Range were first missed by the early explorers in the Hayden surveys of the 19th century, but later in the 20th were noted and described by a variety of people (Fryxell 1930, 1935, Reed 1964, 1967, Edmunds et al. 2011), but because most are small and not the lengthy and impressive ice masses of the polar regions or the high alpine ranges of the Himalaya and other famous ranges, the Teton glaciers have been considerably discounted by many. Devisser (2008) has noted that in the Tetons, roughly 276 permanent snow and ice bodies occur with a minimum elevation of 2694 m, a maximum of 4096 m and an average of 3127 m, and a total combined area of 6.9 km ^{2}. Many of these are immobile snowfields, not true, moving glaciers. The Teton Range does contain 10 of Wyoming's named glaciers, the largest of which is Teton Glacier, which Devisser (2008) calculated at 0.30 km^2, but Edmunds et al. (2011) noted its diminution successively from 1967 at 0.258 km^2 to 2006 at 0.215 km^2. Edmunds et al. (2011) also assessed Middle Teton Glacier and Tepee Glacier for the same time period and noted collectively that the greatest loss of glacier mass occurred from 1983 to 1994, which coincided with a rise in temperatures and a reduction in snowpack.

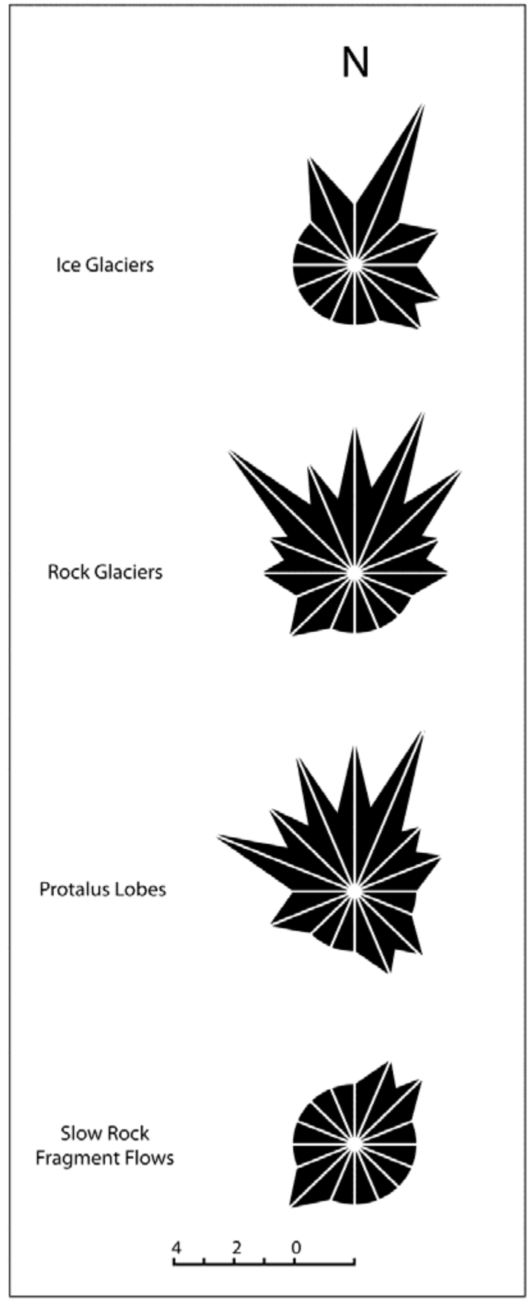

Figure 10. Radar diagrams of aspect directions of various deposits in GTNP. Note that these data represent features mapped for this project, and not for the entirety of the range.

The rock glaciers of the Teton Range are largely unnamed and unstudied, which is unfortunate because they do represent a fair amount of unaccounted for permafrost ice that most likely provides considerable late summer – early fall meltwater overland flow and groundwater base flow into the mountain streams.

✦ MANAGEMENT IMPLICATIONS

Mass movements, rock glaciers, and glaciers as water retention features

Results of this study confirm and reiterate that mass movements, rock glaciers, and glaciers in GTNP attenuate water at up-elevation locations, and arguably for years to centuries depending on the feature. These landforms, then, affect the local hydrologic cycle, and perhaps dramatically, through the lag and pulse effects of stored or mobilized water downstream, respectively. This situation has the potential to exacerbate the ever-present, cascading, trophic effect on most biota in the absence or abundance of water. Monitoring of precipitation and stream discharge at plentiful locations throughout the park would help future scientists better untangle the influences of landform classes or individual features on water distribution in GTNP.

Lack of gauges in GTNP streams

At present, there are no water discharge recording stations in GTNP streams. This situation impedes long term monitoring of the quantity and timing of up-canyon water attenuation/release. Adding these gauges would benefit further investigations of the water resources in that having baseline data of water availability would help insure correct management treatments be used to secure floral and faunal assemblages, as well as any ecosystem services supported by water. Additionally, streams in Grand Teton National Park are undisturbed (anthropogenically) and so these "pristine" waterways could serve as benchmark streams for unmodified/modified water delivery system studies, a practice of the U.S. Geological Survey.

Future hazards to park visitors

While it is fairly impossible to "predict" individual mass movements with any certainty, (sans the precursors to an imminent failure like ground cracks) the ability to recognize areas more prone to failure is likely a more fruitful pursuit. Because the geologic contact (unconformity) between the Archean crystalline rocks and Paleozoic sedimentary rocks of the range creates a situation of comparatively more labile, or "weaker" rocks, at the margins of, and inside the canyons of GTNP, prudence should be exercised in not placing campgrounds and hiking trails on or near such areas where, for example, daylighting sedimentary rocks are overlying crystalline basement rocks.

Nomenclature of Teton landforms

While a vast majority of the landforms in GTNP are properly classified, this study has uncovered several instances in past studies where a feature has been incorrectly classified. In doing so, we have reclassified any landforms whose character does not match the textbook description of the feature. This may have implications to park managers in terms of signage, or park information provided to the general public.

✦ ACKNOWLEDGEMENTS

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