Geomorphology of a thermo-erosion gully, Bylot Island, Nunavut, Canada[1,2]

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Abstract: A thermo-erosion gully has been monitored in the valley of glacier C-79 on Bylot Island since 1999. The main channel of the gully reached 390 m in length a few months after its initiation and grew between 38 and 50 m/year over the following decade, for an overall approximated average of 75 m/year. In 2009, the total gully length and area, including the main and relict channels, were 2500 m and 25 000 m^2, respectively. Gullies affect snow accumulation, and therefore ground temperature, local water flow, and drainage. Sinkholes, gully heads, pools, baydzherakhi, tunnels, and collapses were grouped as a function of time since gully formation in that area. Sinkholes and tunnels were formed every year after gully inception, and baydzherakhi were found in 3–10 year old sections of the gully. Stabilization of the gully floor and sides took about a decade.

Résumé : Le chenal principal du ravin a atteint une longueur de 390 m quelques mois après son amorçage et il s'allongeait de 38 à 50 m/a durant la décennie suivante, pour une moyenne approximative de 75 m/a. En 2009, la longueur totale du ravin atteignait 2500 m et sa superficie 25 000 m^2, en incluant le chenal principal et les chenaux reliques. Les ravins affectent la couverture de neige et donc la température du sol, l'écoulement local de l'eau et le drainage. Des trous de suffosion, le recul de tête de ravins, des bassins, des baydjarakhs, des tunnels et des effondrements étaient regroupés en fonction du temps écoulé depuis la formation du ravin dans la région. Des trous de suffosion et des tunnels se formaient chaque année après la formation du ravin et les baydjarakhs se trouvaient dans les sections du ravin formées de 3 à 10 ans auparavant. La stabilisation du fond du ravin et de ses pentes prenait environ une décennie.

[Traduit par la Rédaction]

Introduction

Ice-wedge polygons are common in continuous permafrost landscapes (Markov 1961; Popov 1961; Péwé 1966). During winter in areas with mean annual air temperatures of –2 °C or less (Hamilton et al. 1983), contraction stresses in frozen ground may lead to ground cracking and to ice-wedge development (Lachenbruch 1962). Permafrost terrain with ice wedges and massive ground ice is particularly vulnerable to climate changes, as thawing may result in surface disturbances and ground subsidence. Changes in snow accumulation in gully depressions enhance the thermal disturbance caused by the feature (Ishikawa 2003; Grosse et al. 2011).

During the summer of 1999, thermo-erosional processes initiated the development of a gully network in ice-wedge polygons on Bylot Island. A detailed description of gully formation and processes involved has been reported by Fortier et al. (2007). The main channel of the gully reached 390 m in length during this first summer. Infiltration of snowmelt runoff into open contraction cracks triggered the development of pipes and tunnels in the permafrost. Yearly repetition of this process and tunnel collapse led to the development of an extensive gully system. Heat transfer between the water flowing in the tunnel and the enclosing frozen ground was the main process of permafrost degradation (Fortier et al. 2007). The spatiotemporal evolution of the gully system has been closely monitored (1999–2010) since its inception (Godin and Fortier 2010).

A gully is "a steep sided trench or channel, often (several metres) deep, that is cut into poorly consolidated bedrock, weathered sediment or soil" (Thomas and Goudie 2000). This definition encompasses both permafrost and nonpermafrost environments. Mechanical erosion by water (Poesen et al. 2003) is the most common driver of gully development, but not in ice-rich permafrost. Heat transfer is usually the main process contributing to gully inception and development in permafrost. The term "thermo-erosion gully" is used here to distinguish the distinct genesis of gullies in permafrost terrain. This paper focuses on thermo-erosional gully development. Such gullying induces the formation of thermo-erosional and

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thermokarst landforms. These landforms change over time, with some stabilizing soon after their inception, some remaining active for several years and some developing after the stabilization of other landforms. Over the past 10 years, the gully head on Bylot Island migrated upstream, while the gully outlet evolved towards stabilization.

The objective of this paper is to describe the geomorphology of the recently developed thermo-erosion gully.

Study site

The study site (73°9′N–79°57′W) lies within the Qarlikturvik Valley (also known as glacier C-79 valley), which is located on the southwestern plain of Bylot Island (Fig. 1; Inland Waters Branch 1969). The valley is ~18 km long, oriented east-northeast–west-southwest, and bordered by plateaus up to 500 m above sea level. Two glaciers at the valley head feed a proglacial braided river that flows towards the sea. The tall valley walls connect the plateaus to a ~5 km wide terrace at the bottom of the valley. Syngenetic ice-wedge polygons have formed and aggraded in the peaty-silt terrace since the Late Holocene (Fortier and Allard 2004). The terrace has a gentle (~2%) slope, with a few ephemeral streams feeding the polygonal wetlands. The ice-wedge polygons have remained nearly intact for ~4000 years, but were recently altered drastically by the development of several thermo-erosion gullies (Fortier et al. 2006). In 2009–2010, 36 thermo-erosion gullies were mapped in the valley. There are several hundred gullies on 1972 aerial photographs of the unglaciated southwestern plain of Bylot Island.

The mean annual air temperature between 1971 and 2000 at Pond Inlet Airport, ~85 km southeast of the study site (Fig. 1), was –15.1 °C, and the mean annual precipitation was 190 mm, most of which (~75%) fell as snow (Environment Canada 2002). The active layer depth may vary from 50–60 cm in peaty-silts to 1 m in sands and gravels (Fortier et al. 2006). Regionally, the thickness of the permafrost has been estimated to be at least 400 m (Young and Judge 1986; Smith and Burgess 2000).

The study site is located on the terrace in the central part of the valley, south of the proglacial river (Fig. 2). The terrace surface near the study area comprises 2–4 m of ice-rich, fine to coarse aeolian sediments mixed with poorly decomposed peat (Fortier et al. 2006). The long axis of the gully runs up the gentle slope of the terrace. From late spring to early summer, snowmelt runoff flows into the gully. During the rest of the summer, the main hydrologic contribution comes from a small, captured stream, flowing from a valley located upstream, drainage from an adjacent wetland and rainfall. The gully outlet connects to a nearby (≈250 m) kettle lake discharge, which flows toward a proglacial river 1 km downstream.

Methods

During the 2009 field work, the gully was mapped using a differential global positioning system (DGPS), with a 0.5–1 m resolution in each dimension. Details of DGPS post-processing are given in Godin and Fortier (2010). The site was surveyed eight times between 1999 and 2009. Erosion markers were installed along the gully to monitor active thermo-erosion processes and gullying. While digitizing the

Fig. 1. Bylot Island is located in the eastern Canadian Arctic. Pond Inlet is located 85 km southeast of the study site on Baffin Island.

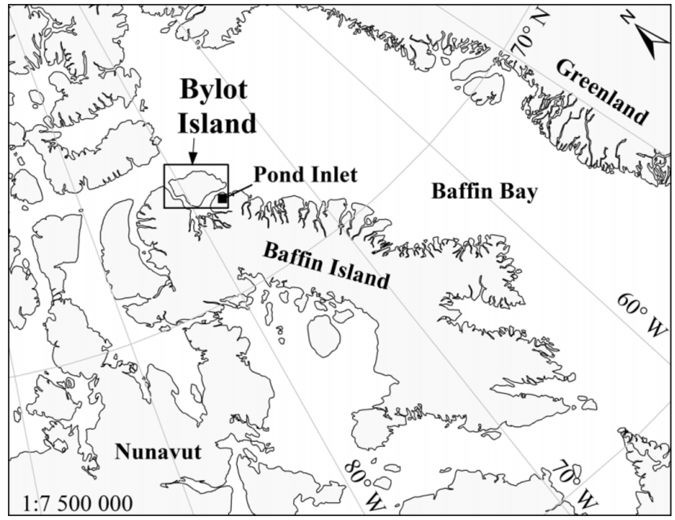
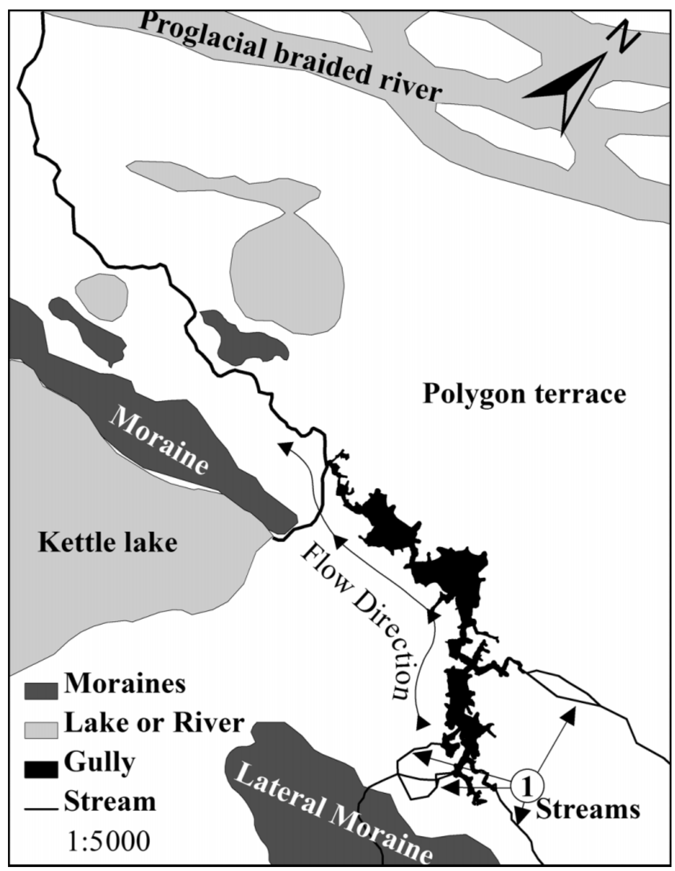


Fig. 2. Geomorphological context of the study site. Four streams (1) reached the network via sinkholes and gully head during the summer. Flow direction in the gully is oriented southeast or east to west. The gully outlet connects with a stream that flows in the proglacial river.



gully contours, permafrost degradation features were recorded in a geodatabase.

Aerial photography from 1958 (1:60 000) and 1972 (1:15 000) was used to validate whether the gully existed during these intervals and to localize thermokarst ponds in the gully area.

A satellite image (IKONOS, taken in 2007, 1 pixel = 1 m) was used to manually delineate gully layout for that year.

The thermo-erosional related features comprised: (*i*) sinkholes, tunnels, collapsed tunnels, and exposed ice wedges near the gully head; (*ii*) retrogressive thaw slumps, baydzherakhi, and stabilized retrogressive thaw slumps in older parts of the gully; (*iii*) alluvial levees and pools in the gully channel. Areas proximal to the gully were characterized by thermokarst terrain.

The geodatabase containing the summary of the gully features was integrated into a geographic information system (ArcGIS v9.3.1, ESRI, Redlands, California). A Bezier smoothing was applied over the digitized gully to minimize spatial error. Gully dimensions were established within ArcGIS by using the "calculate geometry" tool. The distribution of erosion features within the gully was represented using a standard deviation ellipse analysis (Lefever 1926; Gong 2002). This tool regroups several features in a single ellipse to show the most representative emplacement of a given feature. This representation is useful to assess the locus of gully activity and stability. In the current study, the area of the ellipse covers one standard deviation of the distribution of a given feature.

Results

Geomorphological context

Four distinct streams on the terrace entered the gully network during the summers of 2009 and 2010 (southeast of the gully, Fig. 2 (1)). At the gully head, the streams were within a radius of ~100 m. Two of the streams entered the gully via waterfalls; the others flowed into sinkholes and then through tunnels connected to the main channel of the gully.

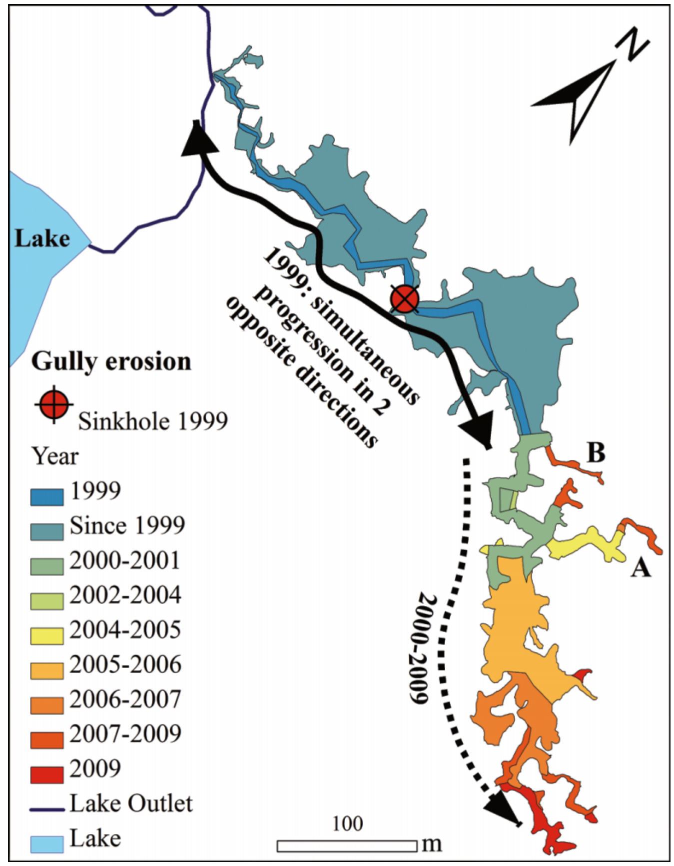
Gully geometry (summer 2009)

During the 10 years following its initiation, the gully system developed to ~750 m in length. If all the gully branches, relict channels and baydzherakhi are considered, the total erosion represents ~2500 linear metres and about 25 000 m^2 (Godin and Fortier 2010). Gully channels essentially follow the ice-wedge pattern (Figs. 2, 3). Recent collapse of tunnel roofs and polygon center remnants illustrate the link between the eroded ice-wedge network and the gully layout. As gully form stabilizes over time, channel angularity softened but maintained even after 10 years.

Gully spatial development, 1999–2009

The gully was initiated in 1999, and the location of the first sinkhole is indicated by a red circle on Fig. 3. The gully then progressed simultaneously in opposite directions (Fig. 3), with 390 m of gullying development in the first year. Subsequently, the network developed upstream and laterally on both sides of the initial channel. In the second and third years, the rate of gully extension was 51 m/year. From 2002 to 2009, the rate was 32 m/year during 2002–2005 and 38 m/year during 2006–2009. The overall rate of gully extension for the whole observation period (1999–2009) was ~75 m/year. These rates do not take into account erosion in directions normal and subnormal to the main axis. Gully geometry started to change with two branches bifurcating away from the main channel eastward in 2004 (Fig. 3, A) and 2007 (Fig. 3, B).

Fig. 3. Spatiotemporal evolution of the gully during the period from 1999 to 2009. The gully initiated in 1999, and the first observed sinkhole is shown by a red circle. The gully first evolved in two opposite directions: downstream and upstream (retrogressive erosion, double arrow line). A and B indicate the positions of two gully branches bifurcating away from the main channel.



Gully geomorphology

Thermo-erosional sinkholes, tunnels, and collapses

Thermo-erosional sinkholes were found exclusively in zones that had been eroded in the previous 2 years. A continuous water source was needed for the development of these sinkholes. Once initiated, ground subsidence enabled the drainage and capture of adjacent streams and groundwater into the sinkhole (Fig. 4A). Water then flowed within a short tunnel network connected to the gully. These tunnels developed preferentially in ice wedges. The tunnels usually bifurcated once or twice at ~90° or ~120° angles at the junctions of ice wedges. Later in the summer, the tunnel roof usually collapsed. This relict stream bed was composed of gravel and sand, without vegetation because of the stream that flowed toward the sinkhole at this location in previous years.

Distribution analysis of the gully morphology showed that the sinkholes were closely grouped on the south side of the gully <200 m from a ephemeral stream (Fig. 5).

The downward development of the tunnel was limited by its base level, which is the proglacial braided river level. Tunnel roof was thinning following two-sided permafrost thawing, which is the downward propagation of the thawing front from the surface and the upward propagation of the thawing front from the tunnel roof.

Fig. 4. Geomorphology of a thermo-erosion gully. (A) Tunnels and sinkholes are common in this area. The channel is no longer used because of the recent formation of a sinkhole upstream, indicated by S on the figure (July 2009). (B) A recently collapsed tunnel roof in a subcritical section of the gully (2009). (C) Baydzherakh in an advanced stage of degradation near the oldest zone of the gully (one decade). (D) Alluvial levees in the gully channel enabled the formation of small pools within sub-perpendicular gully branches. Note large amounts of deposited silt in the channel. (E) Thermokarst pond a few metres from an exposed ice wedge.

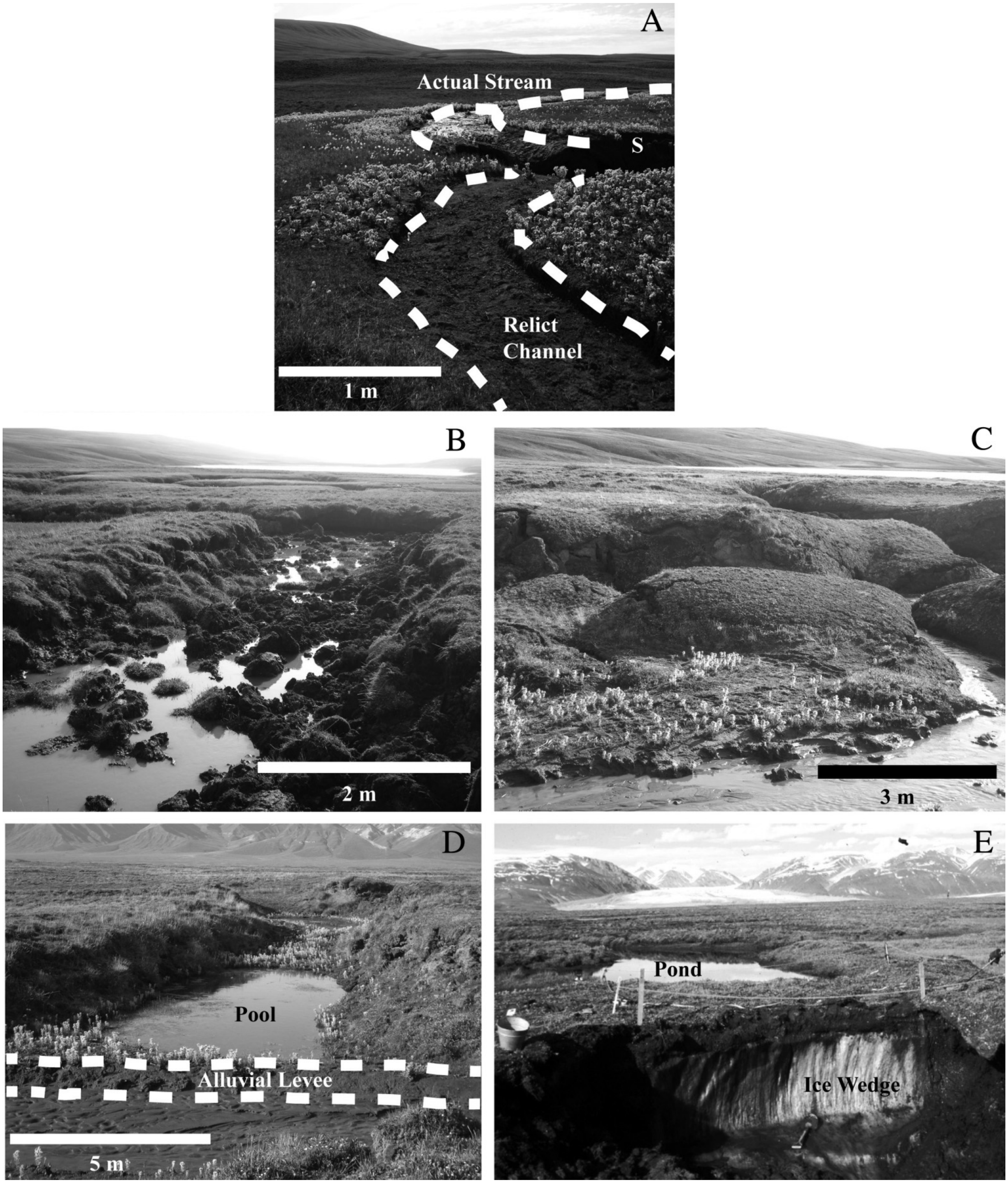
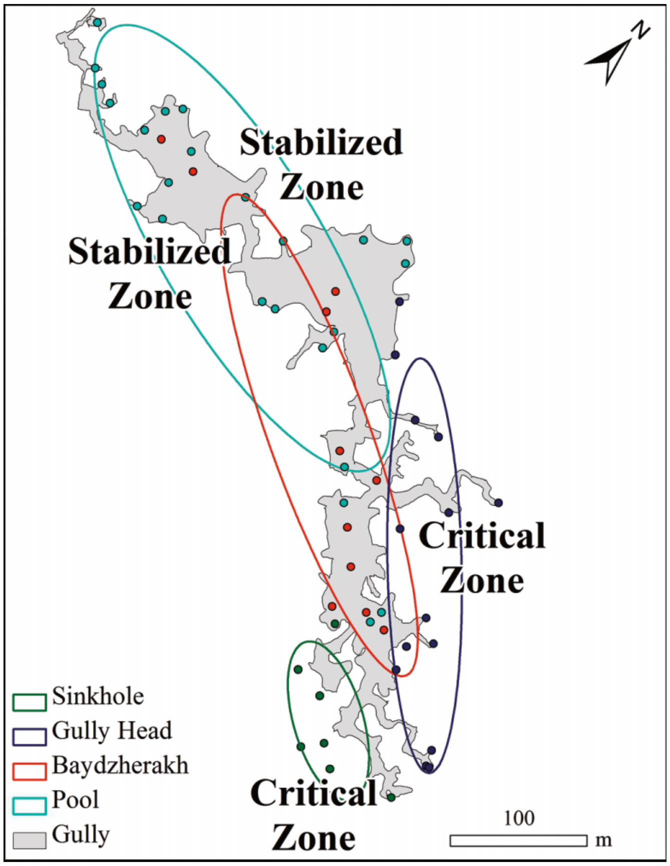


Fig. 5. Standard distance of gully landforms from the ellipse centroid. A smaller ellipse indicates that the form is concentrated in a smaller area, and a larger ellipse indicates a wider distribution. Colored dots in the figure indicate the emplacement of landforms. Zones that were in very active development in 2009 are indicated by the "critical zone" marker, and zones that are stabilized are indicated by the "stablized zone" marker.



Collapses (Fig. 4B) are characteristic of an advanced stage of thermo-erosion. A collapse could also occur when tunnel tops are sufficiently rich in ice or when the base of an exposed ice wedge melts when in contact with stream water. Water flow in the gully created thermo-erosional niches at the base of the exposed ice-wedges. Ice wedges and adjacent frozen ground collapse by brittle failure followed soon after the niche formation, a process similar to what was described by Hoque and Pollard (2005).

Gully head

Gully heads are located in highly active thermo-erosion zones. During the snowmelt period and early summer, most gully heads were connected to the network by waterfalls, some active for the whole summer. Polygons adjacent to gully heads were often saturated and fed for the entire summer by a small stream draining into the gully. Most gully heads were found on the north side of the gully main channel, opposite of the sinkholes (Fig. 5).

Retrogressive thaw slump

Retrogressive thaw slumps were formed active thermo-erosion zones that included exposed ice wedges. Recent thaw slumps were characterized by steep, arcuate, and often well-defined head-walls with poor drainage. Thaw slumping was commonly active for 3–5 years, and several slumps remained active after cessation of thermo-erosion processes in an area. Thaw slumps were distributed throughout the gully, but with higher concentration and larger size in the sub-stabilized to stabilized zone. Recently collapsed sections of the gully featured numerous exposures of ice wedges and ice-rich permafrost.

Thermokarst terrain

Thermokarst subsidence occurred adjacent to the gully margins. Little ground subsidence was observed in active thermo-erosion zones, but it was frequent in the stabilized section of the gully, between 5 and 10 years old.

Baydzherakh

A baydzherakh (Fig. 4C) represents an advanced stage of ice-wedge polygon degradation. Collapse and melting ice wedges create isolated ice-rich permafrost mounds within the gully network. Exposure of these polygon center remnants to solar radiation, sensible heat, precipitation, and mechanical erosion (e.g., following snow dam breaching) resulted in further ground subsidence and, in some cases, complete degradation of the mounds. Baydzherakhi were observed in parts of the gully 3–10 years old, especially where the gully initiated (Fig. 5).

Pools

Sediment transport and sedimentation in the gully channel was significant due to the large input of silt originating from the gully head area. During peak flow, silt deposited in the gully was remobilized and redeposited downstream in low-gradient areas of the gully channel. Sediment deposited on the side of the channel formed alluvial levees (Fig. 4D). These levees dammed water drainage from retrogressive thaw slump and gully branches sub-perpendicular to the main channel and enabled the development of small pools, a few hundred metres squared in area and ~30–50 cm deep. When snowmelt runoff filled the pools and during precipitation events, water spilled over the levees and into the main channel. Pools occurred mostly in older stabilized sections of the gully (Fig. 5).

Thermokarst pond

Near the gully, low-center polygons and degraded ice-wedge junctions often contained thermokast ponds (Fig. 4E). Comparison of a 1972 aerial photograph and an IKONOS 2007 satellite image of the area near the current gully head showed that thermokarst ponds contained by ice-wedge polygon rims were drained following retrogressive erosion of the gully walls. Since thermokarst ponds are widely distributed along the gully, their drainage can occur in active and stabilized sectors of the gully.

Discussion

Development of thermo-erosion gullies

Rates of thermo-erosion processes varied from year to year, depending on local snow depth, the speed at which the snow cover melted, and runoff. Erosion rates during the first year of gully development were extremely rapid. Fortier et al. (2007) reported maximum erosion rates up to 5 m/day for short periods, with a total of 390 m during the first year of gully inception. These high rates during the first year illustrate the instability of gully morphometry and the high sensitivity of the permafrost landscape to thermo-erosion related disturbances.

In a continuous permafrost environment, the expansion and spatial evolution of thermo-erosion gullies is faster than in a nonpermafrost environment by one to two orders of magnitude. The long-term (10 years) erosion rate average for the main channel of the gully was 75 m/year. When all gully branches are taken into account, rates were 250 m/year or 2500 m^2/year. If the first year is excluded from the calculation, then average erosion rates varied between 30 and 50 m/year for the gully main channel, and ~1140 m^2/year area. This is two orders of magnitude higher than rates of ~1 m^2/year for silty-clayed soils in a semiarid region in Spain (Ries and Marzolff 2003) and for a loess plateau in China (Wu and Cheng 2005). In humid areas, especially during extreme precipitation events, higher rates were reported (~80 m/year in clay for a coastal hillslope in California; Swanson et al. 1989). Tunnel collapse, ice-wedge exposures, and retrogressive thaw slumping trigger positive feedback effects that increase gully development.

Active permafrost degradation processes at a specific location in the gully change over time as the gully network enlarge. Figure 6 shows the typical distribution of landforms in the gully during the study period, in relation to time. Figure 7 is a conceptual model of gully zonation which comprises typical landforms within critical (youngest), subcritical, sub-stabilized (older), and stabilized (oldest) zones of the gully. It helps to understand the distribution of landforms and the spatiotemporal dynamics of thermo-erosion gullies. For instance, during the interval from 1999 to 2009, the location of sinkholes moved retrogressively ~300 m upstream (Fig. 8). Sinkhole inception occurred yearly, which suggests that runoff water infiltrated previously formed cavities in the active layer and the permafrost or open frost-cracks (Fortier et al. 2007; Douglas et al. 2011). Sinkhole deactivation usually resulted from stream capture by another sinkhole upstream. Although sinkholes are ephemeral features, they were central to gully dynamics and were the main driver of retrogressive gully development. Sinkholes were always connected to the gully by tunnels. Tunnels are also ephemeral features and usually collapsed after one to a few years by thawing followed by brittle failure (Hyatt 1992). These collapses constantly changed the location and number of gully heads. Streams and runoff flowed into the gully either indirectly via sinkholes and tunnels or directly via waterfalls at gully heads. Unlike sinkholes, some gully head waterfalls were reactivated after 4–5 years of inactivity (no flow) due to gully-induced hydrological changes upstream on the terrace. Collapse of tunnel roofs exposed numerous ice-wedges which rapidly ablated. This permafrost degradation process lasted for several years (5–10 years) and eventually stabilized or slowed down. Due to the high number, large size, and differential orientation of the ice wedges, thaw slumping significantly contributed to gully enlargement. Baydzherakhi and themokarst features were typical of the sub-stabilized zone where Godin and Fortier (2010) demonstrated a clear decrease in gully channel cross-section angularity with time. Pools exclusively formed in the older, low-gradient, stabilized portion of the gully channel. Also, a stabilized gully 1.5 km in length located 2 km east of the current study gully has seen no significant enlargement since 1958, as an aerial photography (1958), an IKONOS 2007 image interpretation, and field observations (2009) confirm. Furthermore, sedimentation rates of ~1 mm/year measured in polygons near the gully area (Fortier et al. 2006) would imply that, in theory, several decades at least would be required to fill in the gully.

Fig. 6. Typical distribution of landforms within the gully system as observed in 2009 in relation with their age. Gully geomorphology age range from 1999 to 2009, from left to right: minimum, first quartile (Q1); median, third quartile (Q3); and maximum. Dots indicate outliers. Critical zone, where the thermo-erosion is the most active, and stabilized zone, where thermo-erosion is no longer active, are indicated on the figure.

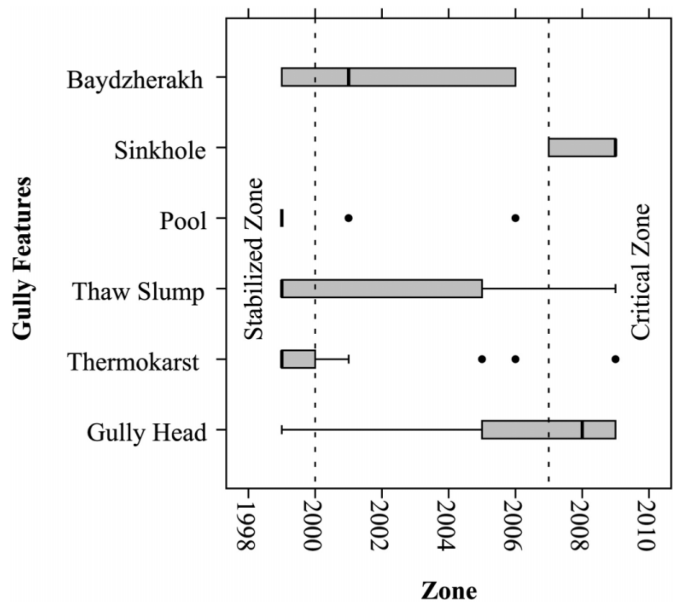


Fig. 7. Conceptual model of gully geomorphology development. Gully inception triggers the activity of various modes of permafrost degradation processes and the development of landforms, which evolve spatially over time in zones characterized by different levels of thermo-erosion activity. Gullies remain in the environment for tens to hundreds of years and thus represent a new state of the periglacial landscape.

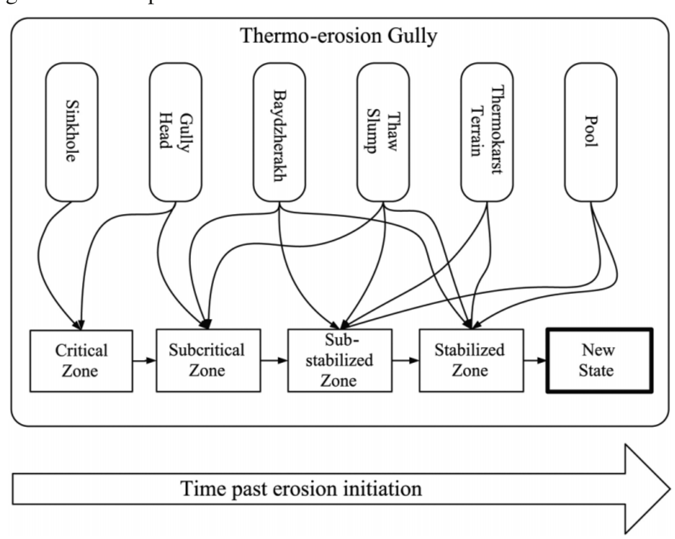
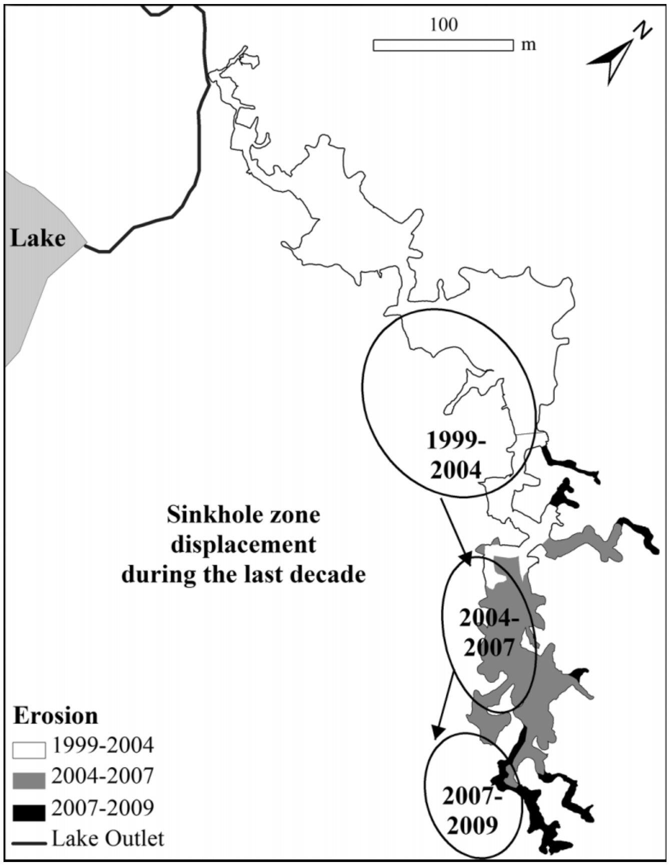


Fig. 8. Sketch showing the displacement of the sinkhole zone upstream during the study period.



These observations illustrate well how thermo-erosion gullies impacted the permafrost over the long term. Thermo-erosion gullies remain in the environment for tens to presumably hundreds of years and thus represent a new state of the periglacial landscape (Fig. 7).

Long-term dynamics of thermo-erosion gullies: the importance of feedback mechanisms

Undisturbed ice-wedge polygons wetlands were highly sensitive to thermo-erosion related disturbances. Over the short term, gully dynamics were strongly affected by numerous positive feedback mechanisms, operating preferentially in the critical and subcritical zones (Fig. 7). During initiation and the first stage of erosion, the gully's morphometrics and thermal regime were in disequilibrium, and channel development rates were thus very high (Sidorchuk 1999). Positive feedback mechanisms were instrumental in accelerating gully spatial development by thermo-erosional processes. Positive feedback effects were essentially related to (*i*) heat advection by running water, which enabled the development of sinkholes and tunnels in the permafrost, and (*ii*) tunnel collapse and exposure of undisturbed permafrost to surface conditions. Other positive feedbacks effects were related to snowpack variability, snowmelt dynamics, changes in local hydrology (e.g., stream capture), and thermokarst pond drainage which reactivated thermo-erosion related landforms, such as gully heads and retrogressive thaw slumps.

In the long term, negative feedback mechanisms contributed to stabilize older sections of the gully. After a decade, about half the gully was characterized by chaotic thermokarst terrains, drained polygons, baydzherakhi, and wide channel cross sections. Long-term terrain settlement and drainage of the gully walls, and of polygons and baydzherakhi along the gully channel, created, in several areas, an insulative dried peat cover colonized by mesic vegetation. This would lead to changes toward a new thermal regime, slow down permafrost degradation processes, and stabilize the terrain. However, in some old areas such as the gully outlet, snow accumulation in depressions, snowmelt water damming in the channel, and summer exposure resulted in further permafrost degradation.

Collapsed materials in actively eroding zones, such as tunnel roofs and slumps, were important contributors to the gully stream sediment charge. During peak flow (late spring, early summer), meanders and alluvial levees and were formed in low-gradient sectors of the gully channel, especially near the gully outlet. Alluvial levees and meanders acted as a buffer to mechanical erosion by preventing the stream from eroding the gully walls. This negative feedback effect further contributed to stabilize older sections of the gully.

Conclusions

Over the past decade, spatial monitoring analysis of a gully within continuous permafrost in the valley of glacier C-79 revealed a nonlinear rate of development which was extremely rapid at first. The area proximal to the gullied area attained a threshold during the year of its formation and was rapidly developing toward a new state of equilibrium. For subsequent years, until a decade past its initiation, the gully developed at a much slower but quasi-steady rate. Gully inception triggered various permafrost degradation processes and the development of landforms, which evolved spatially over time in critical to stabilized zones, and in all cases toward an equilibrium state. Gully development was closely related to the configuration of the ice wedge network and driven by convective heat transfer between flowing water and the permafrost. Thermo-erosional forms were found in the gully in areas where water was entering into the network. Various positive and negative feedback mechanisms contributed to gully dynamics by enhancing permafrost degradation or gully stability, respectively. Gullies remain in the environment for tens to hundreds of years and thus represent a new state of the periglacial landscape.

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