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Observation of Rapid Drainage System Development by Thermal Erosion of Ice Wedges on Bylot Island, Canadian Arctic Archipelago

Daniel Fortier,[1]* *Michel Allard*[2] *and Yuri Shur*[1]

[1] Institute of Northern Engineering and Department of Civil and Environmental Engineering, College of Engineering and Mines, University of Alaska Fairbanks, Fairbanks, AK, USA
[2] Geography Department and Centre d'études nordiques, Université Laval, Québec, Canada

ABSTRACT

Rapid development of a new drainage system was observed on Bylot Island. A 750-m long gully system was eroded in four years. The process was initiated by the formation of sinkholes eroded in ice wedges by runoff flowing into open frost cracks. The sinkholes evolved into underground tunnels cut in the ice-wedge network and the ice-rich permafrost. Widening of tunnels was followed by subsidence and collapse of their roofs and the development of open gullies. The drainage generally developed as the shortest line along the regional slope with some deviations caused by collapse of blocks of soil which temporarily obstructed the water flow. Retrogressive scarps exposed to flowing water retreated at maximum rates of up to 5 m/day for a total of 15 to 50 m during the summer. Scarps exposed to atmospheric heat and solar radiation retreated between 2.5 and 40 m over four summers with a mean of 15.5 m. Such slopes had nearly stabilised after four years with a retreat rate of only a few centimetres per year in the last year of observation. Copyright © 2007 John Wiley & Sons, Ltd.

KEY WORDS: ice wedges; snowmelt runoff; thermal erosion; drainage; tunnel; gully

INTRODUCTION

The formation of tunnels in permafrost was first mentioned by Leffingwell (1915). Mackay (1974, 1981, 1988, 1997), Hyatt (1992) and Seppälä (1997) stressed the important geomorphological modifications that can result from such tunnel formation. Shumskiy (1964) and Shur *et al*. (2004) described how tunnels in the permafrost become the location of water ponding leading to the formation of so-called thermokarst-cave ice or pool ice.

In this paper we show the significant impact that hydro-thermal underground erosion can have on the drainage system in a permafrost environment with well-developed ice wedges. On Bylot Island, during the summers of 1999, 2001 and 2002, we had a rare opportunity to observe tunnel inception and scouring, tunnel collapse, and the subsequent development of gullies several metres wide and hundreds of metres long closely associated with a network of ice wedges.

The objectives of this paper are: 1) to describe the formation of tunnels and the associated gullies; 2) to better understand the fundamentals of underground hydro-thermal erosion and the conditions that enable it to occur; and 3) to show the role of the underground

* Correspondence to: Daniel Fortier, Institute of Northern Engineering and Department of Civil and Environmental Engineering, College of Engineering and Mines, University of Alaska Fairbanks, PO Box 755900, Fairbanks, AK 99775-5900, USA. E-mail: ffdf@uaf.edu

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thermo-erosion process and subsequent gully development in the evolution of an ice-wedge polygon drainage system.

STUDY SITE

The study site (73° 10′ N, 80° 05′ W) is located on the southern plain of Bylot Island, in the valley of glacier C-79 (Inland Waters Branch, 1969) (Figure 1a). The surrounding plateaus and rolling hills have a relief of 300 to 500 m. A braided river flows from the glacier to the sea in a glacio-fluvial outwash plain that runs NE-SW in the centre of the valley. To the south, the outwash is bordered by a 3 to 4 m high terrace onto which aeolian accumulation has occurred since 3670 ± 110 BP ago (Fortier and Allard, 2004). The terrace slopes gently (0.5 to 0.6°) towards the glacio-fluvial outwash and is covered by flat to low-centred ice-wedge polygons, riddled by thaw lakes, punctuated by pingos and drained by small streams flowing to the main river (Figure 1b) (Zoltai *et al*., 1983; Allard, 1996). The study site is located at an elevation of about 20 m a.s.l., at the foot of north-facing gullies incised in peat-covered glacial deposits that form a thin mantle over bedrock. The study site locally concentrates the flow of a catchment area of about 2 km^2 consisting of a network of deeply eroded gullies formed in the 300-m high Lancaster formation bedrock (Figure 2) (Miall *et al*., 1980). Abundant snowmelt from the gullies feeds the fluvial system in early summer. Occasional rainfall events in midsummer also provoke high discharge events. Continuous sedimentation and a cold climate created conditions favouring the growth of syngenetic permafrost during the Holocene (Fortier and Allard, 2004; Fortier *et al*., 2006).

The centres of the polygons are generally very wet with plant communities primarily made up of hydrophilic graminoïds (Massé, 1998) and mosses. The ridges of low-centred polygons are better drained and support mesic shrubs and forbs species (Duclos, 2002). Small shallow ponds occasionally occupy the centre of the polygons, and some troughs are filled with standing water.

The mean annual air temperature (1971–2000) at Pond Inlet, the nearest community located about 85 km southeast of the study site is $-15.1\,^{\circ}C$ (Environment Canada, 2002). There is a strong correlation between air temperatures at Pond Inlet and Bylot Island (Fortier and Allard, 2005). The annual precipitation is 190 mm/year, of which 145 mm water equivalent falls as snow (Environment Canada, 2002). The mean summer (June to August) air temperature is about 4.5 °C (Cadieux *et al*., 2005).

Bylot Island is well within the continuous permafrost zone: the permafrost is probably more than 400 m thick (Heginbottom, 1995). The active layer is about 40 cm thick in the ice-rich peaty silt of the study site (Fortier *et al*., 2006).

METHODS

Snowmelt, Runoff and Frozen Ground Properties

Snow and rainfall data (1999–2002) were provided by Dr Gilles Gauthier (Laval University, Canada). Snowmelt was monitored from the last week of May (2 June 2002) until snow disappearance. Snow thickness was measured at a nearby long-term monitoring site comprising 50 equally spaced stations along two 250-m long transects. Snow cover at the study site was visually estimated every two to three days during our observation period. Daily precipitation was recorded during the summer (June, 1 to August, 20) using a rain gauge.

Between 22 June and 12 July 2001, snowmelt runoff discharge was measured at three stations (Figures 1b and 2) at 2–3 day intervals with a portable current meter (Global Water™). These discharge values are close to the daily peak flows because they were measured in the afternoon. They provide an estimate of the magnitude of flows before and after entering the tunnels and flowing into and out of the main gully. Station 1 was located on a small stream connecting the foot of the valley side and the terrace. Station 2 was located downstream of station 1 at the head of a waterfall created by headward erosion. Most of the water flowing into the tunnel network passed via this retrogressive head scarp. Station 3 was located downstream of station 2 at the gully outlet before its connection to a stream flowing to the glacio-fluvial outwash plain (Figures 1b and 2). Channel cross-sections were measured with a graduated rod and a measuring tape. Flow measurements were repeated three times close to the deepest part of the channel and their average value was calculated.

The stratigraphic context of the ice wedges and the properties of the surrounding soils were studied on borehole cores and from exposures in gully sides. The gravimetric water (ice) content (% dry weight) of the permafrost soils was measured on 48 samples taken from a core (0.4 to 1.8 m depth) retrieved in 2001 from the centre of a polygon located next to the gully system. The thermal regime of the undisturbed permafrost was monitored to a depth of 3 m with a

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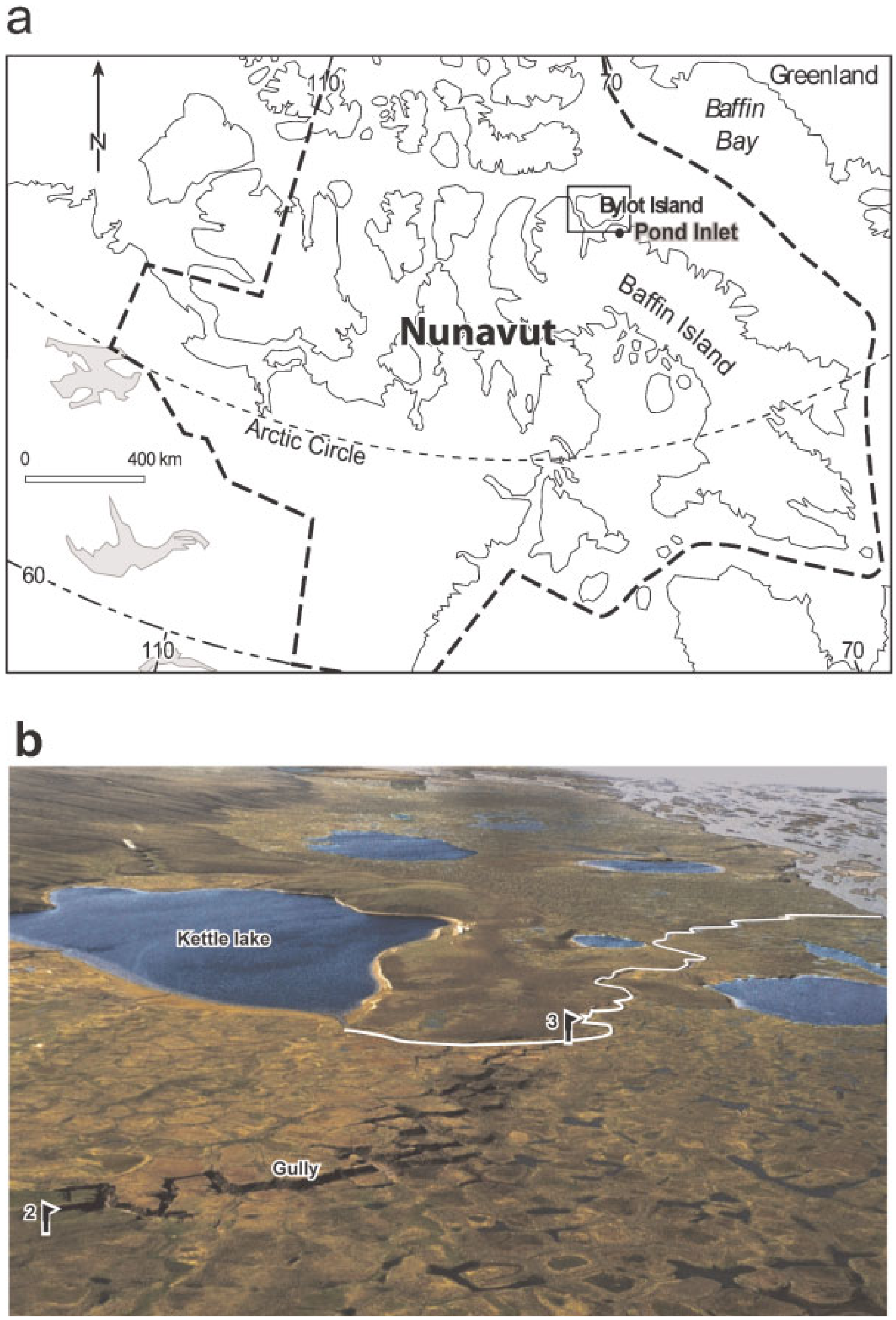

Figure 1 (a) The study site located on the southern plain of Bylot Island in the Canadian Arctic. (b) Oblique aerial view of the study site on the ice-wedge polygons terrace showing the gully in 2001. Flags: discharge measurement stations at the inlet (2) and the outlet (3) of the gully. The gully ends in a small stream (outlined in white) flowing into the glacio-fluvial outwash plain (upper right corner). This figure is available in colour online at www.interscience.wiley.com/journal/ppp.

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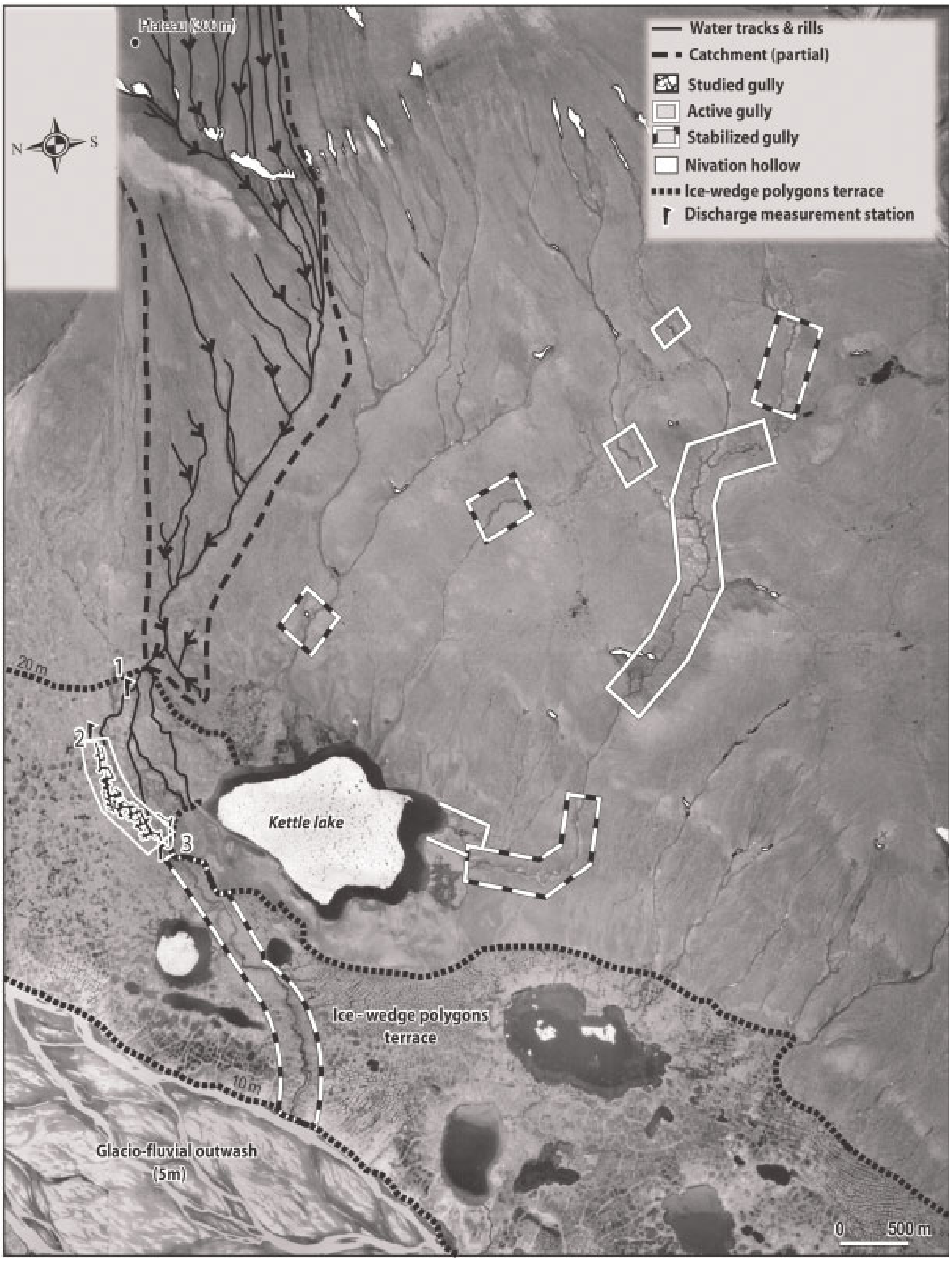

Figure 2 Aerial photograph showing the studied thermo-erosion gully. The flags show the location of the discharge measurement stations on the terrace (1), at the inlet (2) and the outlet (3) of the studied gully. Other active and inactive gullies on the lower portion of the slope and on the ice-wedge polygon terrace are shown (see legend).

thermistor cable (YSI™ sensor 44033 at depths of 2, 10, 20, 30, 40, 80, 120, 180, 200, 300 cm) and a Campbell Scientific CR-10X data-logger installed in the centre of a polygon located in the area of gully formation but not affected by erosion.

Mapping

Polygon dimensions were measured in the field with a measuring tape. Tunnel and gully mapping is based on repeated vertical and oblique aerial photographs and ground surveys using a handheld Global Positioning System (GPS) with a precision of 6 to 8 m. In 1999, reference benchmarks were installed to track headward erosion and the formation of the gully system and the widths of 56 gully cross-sections were measured with a measuring tape. The retreat rates of 15 retrogressive scarps exposed to flowing water were measured. During the study period, about 50 tunnels were observed after collapse of tunnel sections. Three intact tunnels were entered through abandoned sinkholes and their depths, lengths, widths and heights were measured.

RESULTS

Dimensions and Cryostratigraphy of the Polygons

The polygons generally have four or five sides that are 10 to 25 m long, perimeters of 60 to 80 m and surface areas of around a few hundred square metres. The ridges of the low-centred polygons are a few decimetres high and a few metres wide. The ice-wedge tops are located at depths ranging from 0.4 to 0.7 m. The stratigraphy of the upper 3 to 4 m of the terrace's ice-rich syngenetic permafrost is composed of fine to medium, poorly sorted loess and poorly decomposed, low-density, fibrous organic matter. The organic-rich loess of the upper 2.5 m of the terrace has a mean grain size ranging from 5 to 6 Φ (15 to 31 μm) (Fortier and Allard, 2004). The average gravimetric water content of the loess in the upper part of the permafrost (0.4 to 1.8 m) was 110%. At the base of the valley side, the soils also comprise colluvial silts and fine sands interstratified with loess and organic matter.

Three generations of ice wedges were observed. First-generation (primary) wedges are oriented preferentially along the slope. Their exposures varied from 6 to 8 m wide (which is much greater than their true widths) at the top and 3 to 4 m at the bottom of 4 m-deep gullies. The depth of their bases is unknown. Generally, the second-generation wedges are oriented at an angle of 90 to 120° to the primary ones and their exposures were of similar size. Third-generation ice wedges are noticeably smaller than the others with exposures not exceeding 2 m. They subdivide the polygons formed by ice wedges of first- and second-generation wedges and show no preferential orientation.

Snowmelt Runoff

Snowmelt usually began at the end of May or the beginning of June. By mid-June, around half the terrace was snow-free, and snowmelt usually lasted until the end of June (Table 1). Residual snow patches and nivation hollows persisted in shaded areas of the watershed well into August. During the 1999 to 2002 period, the annual average snow thickness varied between 26 and 55 cm (Gagnon *et al*., 2004).

During the observation years, snowmelt proceeded in two main phases. On north-facing valley slopes where thermo-erosion occurred, snowmelt water percolated through the snow cover and mainly refroze deeper in the snow-pack, forming basal ice without giving rise to surface runoff for a period of 25–30 days. During the second period, which lasted approximately 10 to 15 days, residual snow-pack melt and basal ice melt generated substantial amounts of surface runoff. In 2001, for example, meltwater remained within the snow-pack until 16 June while subsequent runoff was high ($\approx$0.2 to 0.4 m^3/s) until the beginning of July and ended by mid-July (Figure 3). At the end of June, the flow was about four times higher than that measured two weeks later. During the study period, runoff associated with a heavy rain event was observed only once, on 12 July 2001.

Underground Thermo-erosion and Tunnelling through Permafrost

The first formation of tunnels in the permafrost by underground thermo-erosion was noticed in mid-June

Table 1 Snowmelt dynamics on the terrace of the study site.

Year	Thickness (cm)	100% cover	75% cover	50% cover	25% cover	0% cover
1999	35	<27 May	05 June	13 June	15 June	>24 June
2000	44	<31 May	11 June	16 June	19 June	>24 June
2001	55	<28 May	05 June	11 June	24 June	27 June
2002	26	NA	NA	04 June	08 June	16 June

Data by courtesy of Dr Gilles Gauthier, Laval University. Thickness: mean snow cover thickness before the onset of snowmelt. Cover: mean arial snow coverage. NA: not available.

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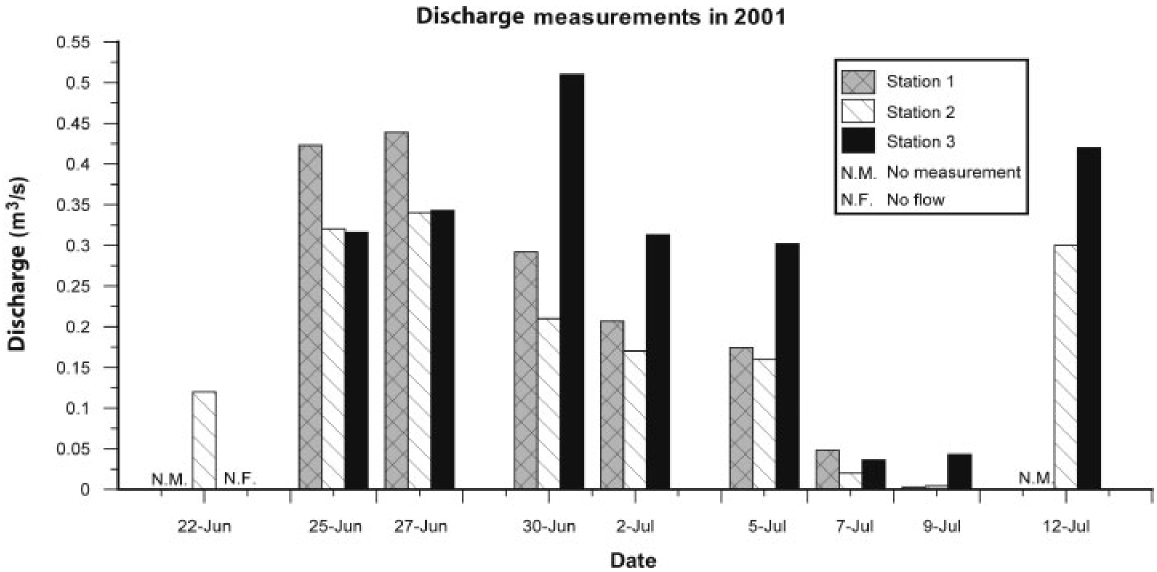

Figure 3 Discharge values measured during the summer of 2001. See Figures 1 and 2 for measurement station locations.

1999. Runoff channelled along the trough of a polygon disappeared into an open thermal contraction crack. Running water was heard to flow underground, but the flow was not visible from the surface. Within a few days, flowing water had enlarged the frost crack, which evolved into a fully developed sinkhole with a 3 to 4 m vertical waterfall ending in plunge pools (Figure 4).

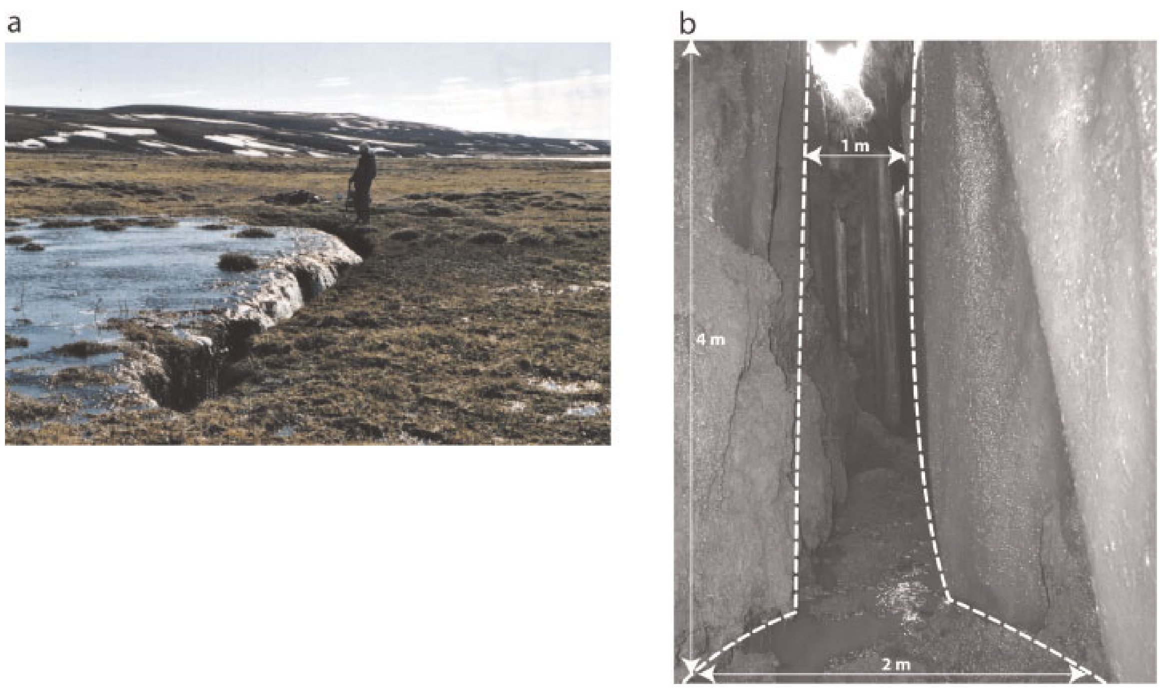

Figure 4 (a) Photograph (person for scale) of a waterfall forming a plunge pool in an ice wedge. (b) Cross-sectional view of (a) after partial collapse of the tunnel. A 2-m wide tunnel formed at the bottom of the waterfall. This figure is available in colour online at www.interscience.wiley.com/journal/ppp.

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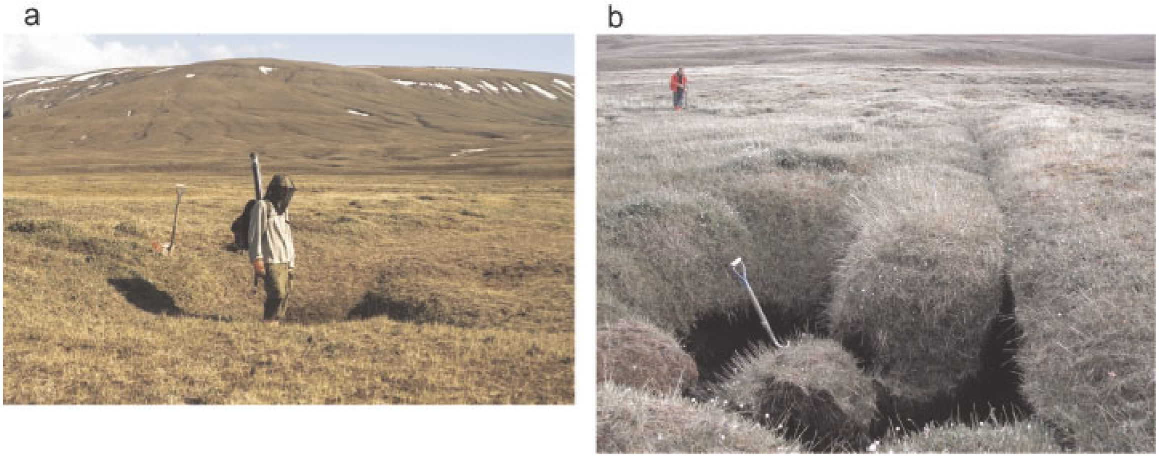

Figure 5 (a) Typical saucer-shape depression at the junction of ice wedges after cessation of runoff. (b) Sinkholes along an ice-wedge trough. The hole was about 2 m deep. This figure is available in colour online at www.interscience.wiley.com/journal/ppp.

Water flowing over the ice wedges melted their upper parts and created numerous saucer-like surface depressions (Figure 5a) that were semi-circular to elliptical, several decimetres deep, and had diameters ranging from 1 to 2 m. During the summers of 1999 to 2002, surface thermo-erosion created funnel-shaped sinkholes along ice-wedge troughs (Figure 5b) in late-June to early July (snowmelt runoff peak). Some sinkholes and tunnels were preserved for more than one year and served as preferential flow paths during the next snowmelt runoff period.

Around mid-June 2001, when surface water flowed into the frost cracks, the temperature of the upper permafrost and the frozen active layer varied between around $-2\,^{\circ}$C at the surface to below $-15\,^{\circ}$C at a depth of 3 m (Figure 6). Table 2 shows thawing and freezing dates for the active layer at different depths along with thawing degree-days. The width of tunnels eroded during this period varied from 0.1 to 10 m in a few days. Tunnel scouring ceased with the end of runoff. It was reactivated on 12 July 2001 during and after a rain event of 15 mm. Measurement of flow at station 2 and out of the main tunnel showed that the discharge values were similar in magnitude to snowmelt runoff values ($>0.3\,m^3/s$).

The tunnels were mainly opened within ice wedges and followed their polygonal layout. Generally, the tunnels had oval cross-sections and flat floors, and were located at depths ranging from several decimetres to 4 m beneath the surface. Larger tunnels were always located between 3 to 4 m in depth, which is presumably the depth reached by winter frost cracks.

Tunnel roofs were composed of the active layer, ice-rich fine-grained soil of the upper permafrost, and remnants of ice wedges (Figure 7). During the summer, the stress exerted on the tunnel roofs by the overburden resulted in creep subsidence visually estimated to be in the order of several decimetres. The roofs of most of the observed tunnels eventually collapsed (Figure 7b) during the first summer of a tunnel's formation, although some did not collapse until the following year.

The observed network of tunnels comprised a main trunk and tributary tunnels created at the junction of ice wedges (Figure 8). Flow deviation around collapsed blocks created a zig-zag pattern of tunnels that followed the pattern of the ice wedges forming the polygons. *In-situ* measured dimensions of the tunnels are given in Table 3. Tunnel 1, a part of the main trunk of the tunnel network, was formed in 1999 by melting a downslope-oriented ice wedge and the surrounding sediment in the polygons. Tunnel 2 was a tributary of Tunnel 1 located at an ice-wedge junction. Tunnel 3 was first observed before snowmelt began in the summer of 2001 which indicates that it formed during the previous summer. In 2001, the underground water flow therefore took place partly through an inherited tunnel system. This tunnel was eroded in a downslope-oriented ice wedge and the surrounding sediments. It was enlarged during the summers of 2001 and 2002 (Figure 7a) and collapsed in August 2002.

Tunnel Collapse and Gully Formation

Tunnel collapses resulted in sub-vertical to oblique scarps with waterfalls and plunge pools undercutting the base of the exposed ice wedges (Figure 9a). For the study period, the maximum retreat rate of the scarps (2

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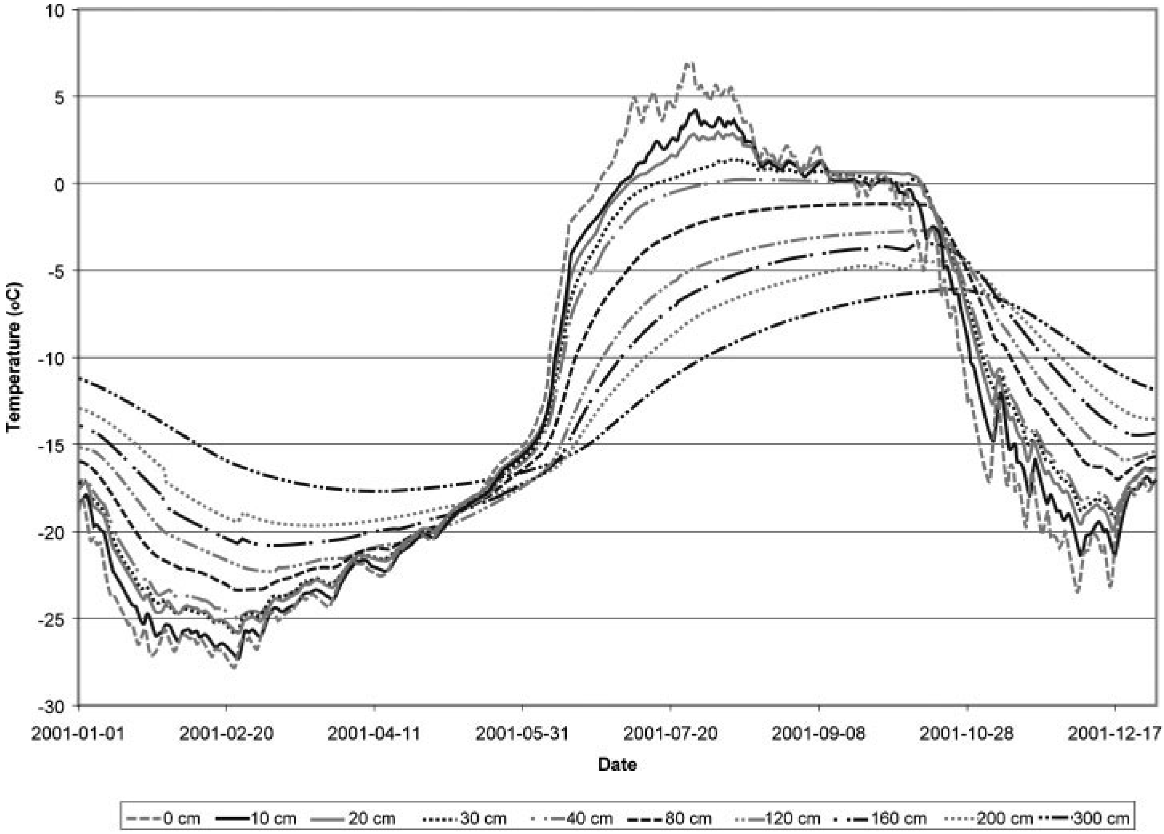

Figure 6 Ground thermal regime of a low-centred polygon in undisturbed permafrost at Bylot Island in 2001.

to 6 m wide and 2 to 4 m high) exposed to runoff (0.21 to 0.34 m^3/s) occurred at the end of June and early July and varied between 1 and 5 m/day with total retreats between 15 to 50 m during the summer. Blocks of ice wedge and the overlying active layer collapsed episodically and contributed to the retreat. On certain occasions, tunnels were formed under low hydraulic gradients (0.02 to 0.1) due to diversion of water along collapsed blocks or thaw slumps. A few tunnels were cut within the frozen peaty-silt sediments rather than within ice wedges (Figure 9b).

Runoff diminished considerably in July, falling below 0.05 m^3/s (Figure 3). When ground ice exposed in a scarp was not affected by running water but exposed to radiation and atmospheric heat, retreat continued by thaw slumping, collapse of active-layer overhangs, slope failure and mudflows. The sediments filled the bottom of the gully channels and had a strong impact on their geometry. The retreat after four summers as measured along 43 scarps in 2002, varied between 2.5 and 40 m with a mean of 15.5 m (standard deviation of 7.5 m). These values are similar to

Table 2 Thermal regime of the active layer.

Depth (cm)	Thawed	Frozen	Days unfrozen	TDD
0	27 June	21 September	85	256
10	03 July	24 September	83	147
20	07 July	14 October	99	117
30	17 July	04 October	89	49
40	04 August	24 September	52	6

Depth: thermistor depth; Thawed/frozen: date when the ground thawed or froze in 2001; Days: number of days in the unfrozen state; TDD: thawing degree-days.

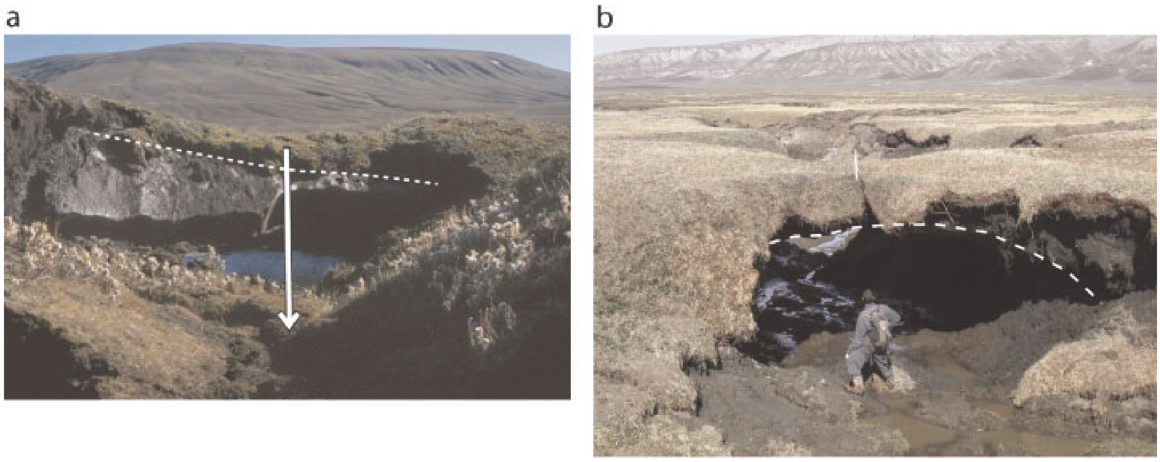

Figure 7 (a) Collapse of a tunnel roof. The top of an eroded ice wedge (broken line) is visible a few decimetres under the surface of the intact part of the roof. (b) Tunnel 3, exposed by roof collapse in 2000 and enlarged during the summers of 2001 and 2002. The broken line shows the ceiling of the tunnel. The length of the tunnel can be estimated by the collapsed part visible in the background. This figure is available in colour online at www.interscience.wiley.com/journal/ppp.

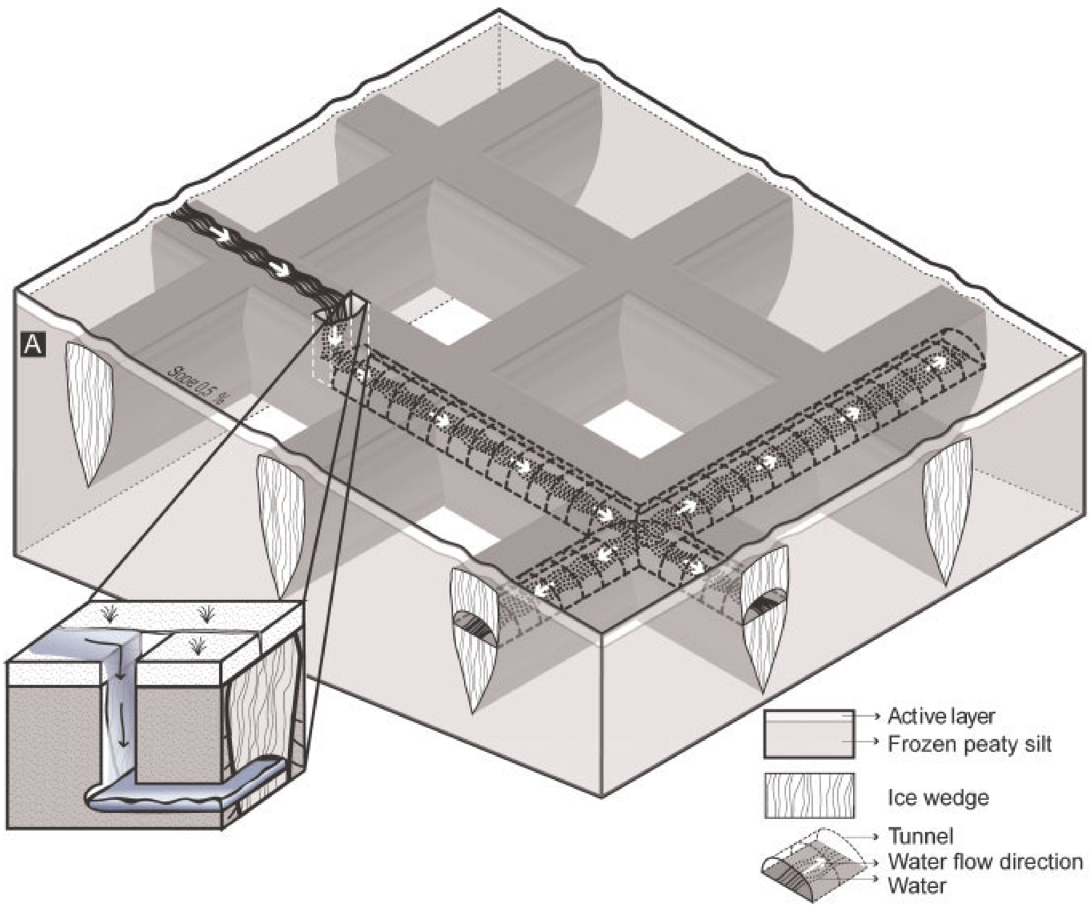

Figure 8 Sketch of idealised tunnel network through permafrost with main and tributary tunnels.

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Table 3 Dimensions of tunnels.

Tunnel	Depth (m)	Max. height (m)	Max. width (m)	Length (m)
1	3.5 to 4	1	9.7	25
2	3.5 to 4	1.75	3.75	16
3	4	0.5	5.5	16

Depth: tunnel's floor depth under the surface of the polygon; Max. height: maximum height of the tunnel; Max. width: maximum width of the tunnel; Length: tunnel's length.

measured retrogressive retreat of slopes for the Canadian and Russian Arctic (Mackay, 1966; French and Eggington, 1973; French, 1974; Are, 1985; Lewkowicz, 1987, 1988; Robinson, 2000; Shur *et al*., 2002).

Thaw slumped material, especially collapsed mats of poorly decomposed surface peat with rooted vegetation, accumulated on gully slopes and protected permafrost from further thawing. The retreat rate of such slopes ($n = 40$) initiated in 1999 was only a few centimetres per year in 2002, about an order of magnitude slower than the retrogressive erosion rate observed during the first two years.

Measurements of gully cross-sectional widths ($n = 56$) in 1999 revealed that they varied from 1 to 16.5 m with a mean of 6.4 m. After four summers, the main gully system was over 750 m long. Figure 10 illustrates change that took place along a gully section between August 1999 and August 2002 after the erosion of the ice wedges. Based on Figure 11a, the area covered by the interconnected gullies in 2002 was about 20 000 m^2 and the estimated volume of eroded frozen ground and ice wedges was around 60 000 m^3, assuming a general box shape for the gullies and a mean depth of 3 m. A total of ten active and inactive gully systems with a zig-zag pattern were identified on the terrace and the lower portion of the valley slopes (Figure 2).

DISCUSSION

Snowmelt Runoff and the Process of Underground Thermo-erosion

In the High Arctic, the snowmelt period is usually the most important hydrologic event of the year. At the beginning of the snowmelt runoff period, the ground is still frozen (Figure 6). The soils of the active layer that were saturated during the previous fall have a very low hydraulic conductivity and surface runoff over these soils is important (Kane and Stein, 1983; Granger *et al*., 1984; Roulet and Woo, 1986; Rovansek *et al*., 1996; Quinton and Marsh, 1998; Zhao *et al*., 2002). Our observations show that during this period, runoff was concentrated in the troughs of the polygons and that the water flowing down into the frost cracks enlarged the conduits and formed tunnels at the bottom of waterfalls by thermal and mechanical erosion. The tunnels were essentially eroded into ice wedges which are the most susceptible to thermal erosion because only thermal and not mechanical energy is required to remove the products of erosion. This concurs with empirical data from Russia showing that susceptibility to erosion increases according to soil type from peat to ice-poor silty clay, ice-rich silty-clay, sand and ice-rich silt, with the most susceptible being ice (Konstantinova, 1982).

We believe that thermo-erosion of ice wedges was facilitated by proceeding along open frost cracks. Underground thermo-erosion might therefore be more likely to occur in areas of active ice-wedge cracking with site-specific conditions to concentrate runoff.

Most of the larger tunnels were formed at the depth where the ground temperature was around $-15\,^{\circ}C$ (Figure 6). Thousands of cubic metres of ice wedges and surrounding frozen ground were eroded each summer. This became possible when running water created a very effective convective heat exchange at the water-ice (or frozen soil) interface. Ice melt due to convective heat exchange with running water can be estimated using the following equation (Tomirdiaro, 1972; Shur, 1977; Are, 1980, 1985):

$$h = \alpha(\Delta T_1)t/(\rho + c\Delta T_2) \quad (1)$$

where h is the thickness of melted ice (m), α is the heat-exchange coefficient between water and ice, (W/m^2 K), ΔT_1 (K) is the temperature difference between water and the melting ice, t is time (s), ρ (J/m^3) is the volumetric latent heat associated with ice melting, c (J/kg K)is the volumetric specific heat of ice, and ΔT_2 (K) is the difference between initial ice temperature and ice melting point. The second term in the denominator ($c\Delta T_2$) is about 10% of the first one (ρ) for the Bylot Island case study and the thickness of melted ice can therefore be approximated from the reduced equation:

$$h = \alpha(\Delta T_1)t/\rho \quad (2)$$

Equation 2 describes the heat-exchange process as ablation with the removal of thawed material and the

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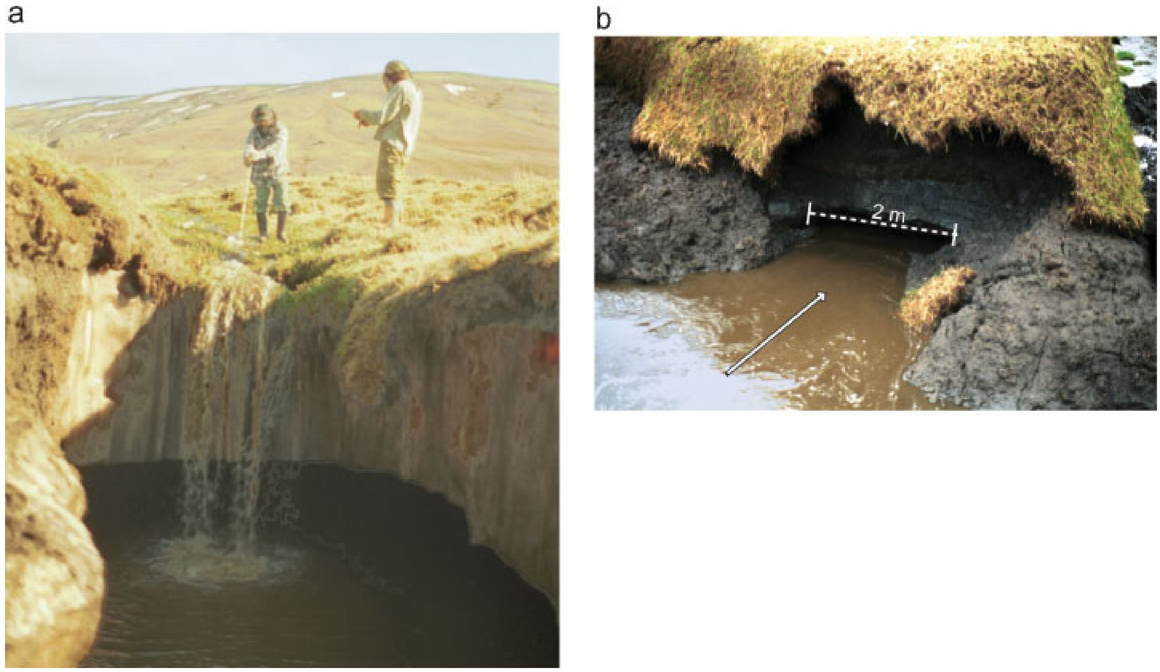

Figure 9 (a) Retrogressive erosion scarp exposed to flowing water. Note the thermo-erosional plunge pool at the bottom of the waterfall and the thermo-erosional niche under the eroded ice-wedge wall. (b) Tunnel cut in the ice-rich loess due to flow diversion in the gully. The arrow shows the direction of the water flow.

constant impact of running water on the melting surface. Shur (1977) evaluated α for a vertical wall of ice or frozen ground in contact with still water (free convection) and found that it varies, depending on water temperature, from about 100 to 300 W/(m^2 K). Are (1980, 1985) and Zotikov (1982) used well-known empirical heat-transfer equations to determine α for heat exchange between running water and ice (forced convection) and found it to be about 1000 W/(m^2 K). Laboratory experiments with water running over small ice plates (Pekhovich and Shatalina, 1970) and field experiments involving lake-ice floes up to

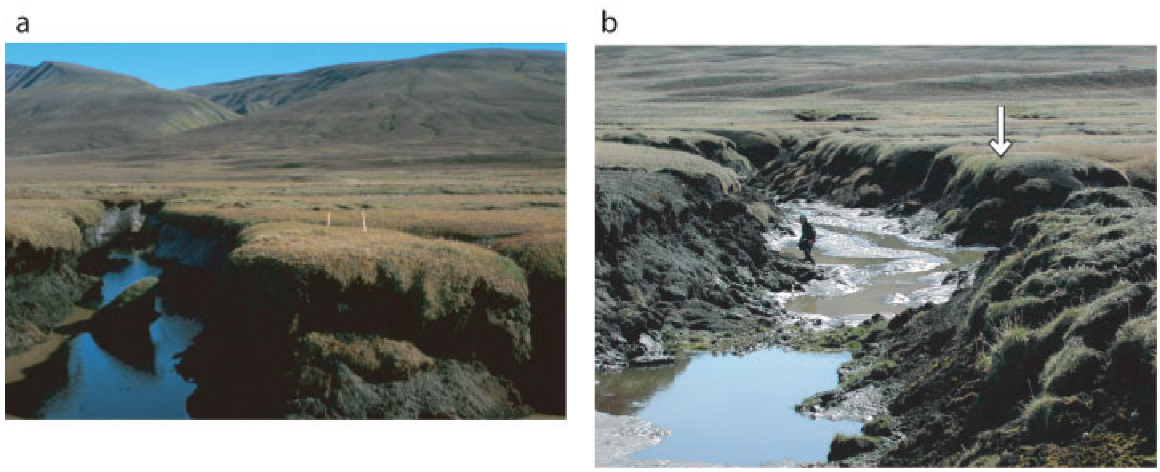

Figure 10 Evolution of a gully section. (a) August 1999 (pegs are 50 cm long): note the collapsed tunnel roof standing in the channel partly filled with water. (b) August 2002 (person for scale): the arrow points at the location of the pegs shown in (a).

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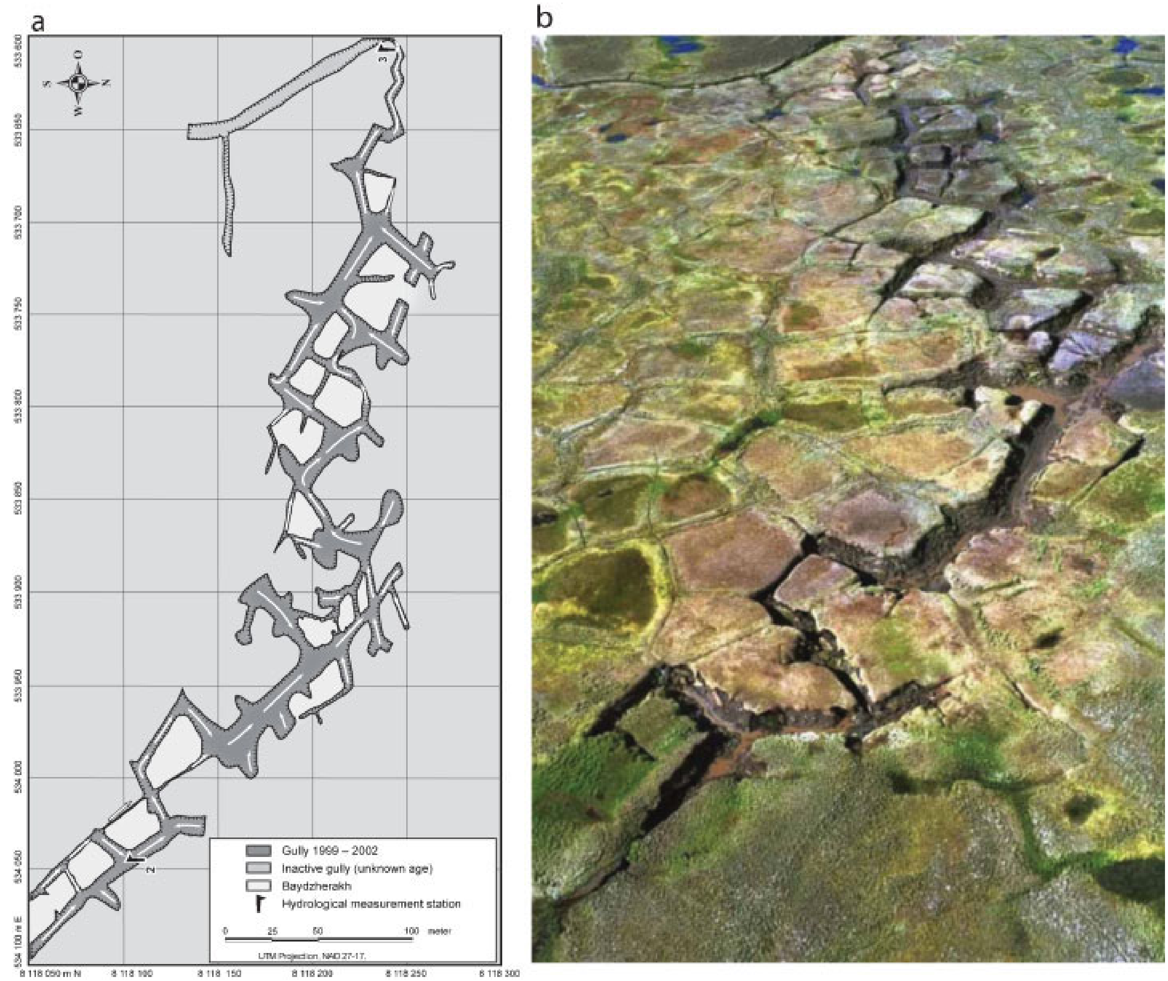

Figure 11 (a) Map of the gully system in 2002 (UTM) units. (b) Oblique aerial view of the upper part of the gully in July 2001. The highest point is in the lower left corner (same orientation as (a)).

100 m wide towed by a boat (Khudiakov *et al*., 1978) showed that α can be as high as 3000 W/(m^2 K). With such high α values, the convective heat exchange between running water and ice is very efficient even at low water temperatures. Based on Equation (1), the ice thickness melted would be 0.7 m/day or about 20 m/month if the water temperature was 1 °C and α was 2000 W/(m^2 K). This is in a good agreement with our observations of 1 to 5 m/day (this rate includes block collapse) and shows that large volumes of permafrost can be removed by hydro-thermal erosion under cold water and soil temperature conditions as long as runoff discharge is high and sustained ($>$0.15 to 0.45 m^3/s for 10 to 15 days in our case).

Gully Development and Rearrangement of the Hydrography of Polygonal Terrain

The gully which initially formed in 1999 accommodated runoff from the watershed and created a new hydrologic connection between the valley slope and the proglacial river. Hydrologic measurements from the beginning of the summer 2001 (e.g. Figure 3, 25 to 27 June) showed that the discharge at the main gully inlet (station 2) and at the gully outlet (station 3) were similar, indicating that the slope runoff was the main source of water running down the gully. However, later in the summer, the discharge at station 3 was nearly twice that at station 2, indicating that drainage from

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the active layer encompassing those eroded polygons located along the gully margins and water from melting ice wedges and thawing permafrost exposed in gullies were then major sources of water flowing in the gully. The presence of other active and inactive gullies on the terrace and the lower portion of the slopes indicates that the process of thermo-erosion is an effective driver of the hydrographic development of the local drainage network.

Gully Stabilisation

Along the gully margins, thaw settlement followed surface destabilisation when excess water from the melting ground ice drained out of the soil. Ground surface subsidence in turn caused the thawing front to gradually penetrate into the ice-rich upper part of the permafrost and led to further subsidence. However, where the soils were well drained and not affected by mass movement, many retrogressive thaw slumps initiated in 1999 were nearly stabilised in 2002 suggesting that the active layer was close to a new equilibrium. We believe that this can be explained by the development of a low-thermal conductivity, dry porous layer of organic-rich silt protecting the frozen soil beneath it from further thawing.

CONCLUSION

Our observations on Bylot Island demonstrate that thermo-erosion by snowmelt runoff can initiate internal tunnelling and gullying of ice-rich permafrost with a well-developed system of ice wedges. This can lead to rapid and far-reaching hydrologic and geomorphologic modifications to Arctic terrain at a local scale.

On Bylot Island this process gave rise to an underground drainage network of quickly progressing tunnels along a main central axis. In four summers, this underground network evolved into a continuous system of gullies over 750 m long and covering an area of about 20 000 m^2.

Further studies on a regional scale are needed to answer whether these underground thermo-erosion processes occur widely or are the result of a rare combination of site-specific factors.

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