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Des défis du Nord au Sud

The Eureka Sound lowlands: an ice-rich permafrost landscape in transition

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ABSTRACT

The Eureka Sound area is characterized by cold polar desert conditions (MAAT of -19.7ºC), permafrost >500 m thick and a dynamic landscape where rapid change linked to melting ground ice is common. Massive ground ice and ice-rich sediments are widespread below the Holocene marine limit (150 m asl) where its distribution, content, and stratigraphic character are closely related to the nature of marine sediments. Since 1990 approximately 450 natural exposures of massive ice have been mapped, including ~100 that were studied in detail. Most are headwall exposures in retrogressive thaw slumps. These data are supplemented by 50+ core holes up to 15 m deep and GPR surveys. Over 25 years of observation the average retreat is 6.9 m/yr, but in the past 5 years the average is 7.5 m and the highest annual retreat is 23.9 m. Since 2005 many new slumps have formed and ice wedges have exhibited significant thaw degradation.

RÉSUMÉ

La région où se situe Eureka est caractérisée par des conditions polaires désertiques (avec une température annuelle moyenne de -19.7°C), par un pergélisol >500 m d'épaisseur riche en glace et d'un paysage dynamique où des changements rapides sont liés à la fonte de glace de sol massive sont communs. La glace de sol massive et des sédiments riches en glace sont très répandus en dessous de la limite marine holocène (150 m d'altitude), où sa distribution, le contenu, et le caractère stratigraphique sont fortement reliés à la nature des sédiments marins. Depuis 1990, environ 450 expositions naturelles de glace de sol massive ont été cartographiées, y compris ~100 qui ont été étudiées en détail. La plupart sont des retraits de mur frontal des glissements régressifs dû au dégel. Ces données sont complétées par plus de 50 trous de carottage jusqu'à 15 m de profondeur et des levés GPR. Une retraite moyenne de 6,9 m/an a été déterminée par plus de 25 années d'observations. Toutefois dans les cinq dernières années, la moyenne est 7,5 m et la retraite annuelle la plus élevée est 23,9 m. Depuis 2005, de nombreux nouveaux effondrements se sont formés et les coins de glace ont montré une dégradation de dégel importante.

1 INTRODUCTION

'With the Arctic warming roughly twice as fast as the rest of the globe, there is more need than ever to monitor the changing conditions there" (Nature 2011, p171). This is the underlying message of not only the 2011 special issue of Nature on the effects of climate change in the Arctic (called "After the Ice"), but also most integrated syntheses on climate change (e.g. ACIA 2005, IPCC 2007, 2013, Forbes et al 2011). The undisputed message is the Earth's cryosphere is experiencing unprecedented change, and one of the recommendations given very high priority by most reports is the need for more systematic research and long-term monitoring to reduce uncertainties related to cryospheric processes and feedbacks. Permafrost is a key component of the cryosphere through its influence on energy exchanges and hydrological processes, thaw subsidence and carbon fluxes, and hence the global climate system. The climate-permafrost relationship has gained added importance with increasing awareness and concern that rising temperatures will increase seasonal thaw and reduce permafrost extent. In response to these concerns two themes tend to dominate the Arctic climate change literature; the first is concerned with permafrost-carbon feedbacks (e.g. Koven et al 2011) and the second is the projection of permafrost loss based on simulation models (e.g. Slater and Lawrence 2013). In both cases the results are often forced with generalized data sometimes from the sub and low Arctic where permafrost is inherently unstable to quasi-stable. Projections for the high Arctic are often poorly constrained and based on limited data. A serious gap in many of these analyses is a lack in long-term baseline observations, information on active layer-permafrost dynamics and ground ice conditions for the Canadian high Arctic. The Canadian Arctic Archipelago covers approximately 1.42 x 10^6 km^2 and lies entirely within the continuous permafrost zone. Most global scale models perform poorly for this area, they also tend to use the IPA permafrost map as the basis for permafrost and ground ice conditions which due to scale is grossly oversimplified. Regional analyses of permafrost-ground ice conditions based on field studies remain an important source of information to help constrain modelled projections of climate change, particularly for regions like the Arctic Archipelago. This paper presents recent observations from a 25 year study on ground ice and related thermokarst driven landscape changes in the Eureka Sound lowlands of Ellesmere and Axel Heiberg Islands in the Canadian high Arctic. The primary aim of this paper is to highlight recent trends in summer climate and thermokarst activity. The latter will focus on two very dynamic thermokarst processes; the formation of retrogressive thaw slumps in areas of massive ground ice and thaw subsidence along ice wedge troughs.

2 STUDY AREA

This study focuses on ground ice and thermokarst in the Eureka Sound Lowlands (ESL); a flat to gently rolling area on central Ellesmere and Axel Heiberg Islands in the Canadian high Arctic (Figure 1).

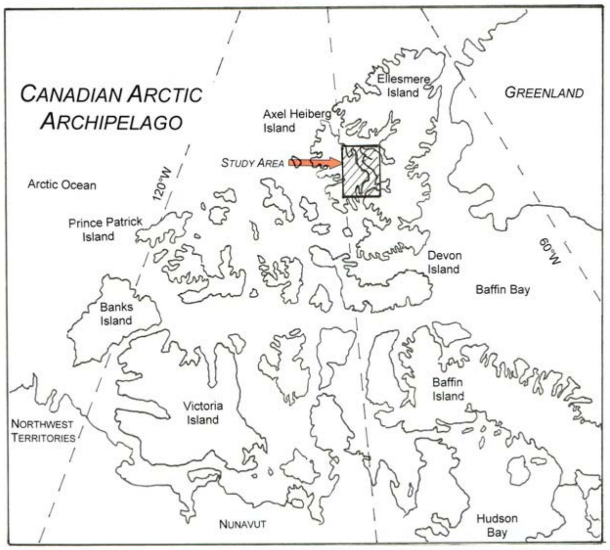

Figure 1. Location of the Eureka Sound Lowlands (ESL) in the Canadian High Arctic

The subdued landscape of the Eureka Sound Lowlands (ESL) lies below 200m asl and covers roughly 750 km^2 including; the Fosheim Peninsula, May Point, Depot Point and the emergent surfaces adjacent to Eureka Sound and Mokka Fiord. The ESL is part of an inter-montane basin that has an area of roughly 15,000 km^2 that is bordered by the Sawtooth Mountains to the east and the Muller ice caps to the west. The Environment Canada Eureka Weather Station is located on the north side of Slidre Fiord on the Fosheim Peninsula at 80°00′N, 85°55′W; roughly in the center of the ESL. Eureka is also a logistics hub that includes Fort Eureka a Canadian Military base, Polar Continental Shelf Program fuel storage and a year round gravel runway that is one of the more important transportation nodes in the high Arctic.

2.1 Climate

The ESL is characterized as a polar desert, with mean annual precipitation of only 68 mm, making it one of the driest regions of Canada. The thaw season, defined as the period when mean air temperatures remain above 0°C, usually varies between 3-6 weeks in length and is used to calculate the thaw degree days (TDD). A 60+ year climate record (1947-present) shows a mean annual air temperature (MAAT) of −19.7°C (s.d =1.3°C), and mean February and July temperatures of−38.2°C and 5.6°C, respectively. However, since 1980 the MAAT is -18.8°C (s.d.=1.5°C) and the mean February and July air temperatures have risen to -37.4°C and 6.2°C (Figure 2). The mean July air temperature for the last 10 years is 7.7°C and of particular interest were the mean July temperatures for 2011 and 2012 which were 9.8 and 8.7, respectively. The long-term average TDD is 359, since 1980 it is 401 and for the past 10 years it is 485. In 2011 and 2012 there was 676 and 575 TDDs but in 2013 the TDD totaled only 225, one of the lowest on record. There has been an increase in both the length and the mean temperature of the thaw season since 1980 with the greatest increase in the last 10 years including two of the warmest summers on record in 2011 and 2012. The significance of this is reflected in increased thermokarst.

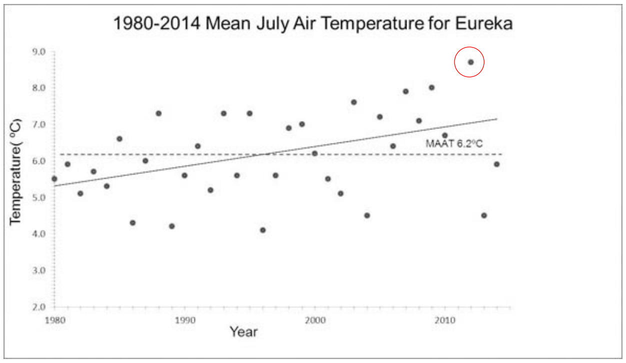

Figure 2. Plot of July temperatures since 1980 showing a general warming trend (solid line) with recent temperatures consistently above the mean (dashed line)

The climate of the Fosheim Peninsula is influenced by the surrounding mountains which limit the influx of cold air from the Arctic Ocean; July temperatures tend to be warm for this latitude (Edlund and Alt 1989) and have ranged as high as 19.4°C. It has been suggested that the weather station's location on the coast and the presence of a semi-permanent temperature inversion cause the weather station temperatures to be cooler than those further inland (Alt et al. 2000; Atkinson 2000). Data from Hot Weather Creek, a Geological Survey of Canada climate change observatory that was active for 6 years in the 1990s, suggest that that mean July temperatures inland may be as much as 3°C warmer (Alt et al. 2000).

2.2 Permafrost

Permafrost in the ESL is continuous, deep and relatively cold. The permafrost thickness measured at an abandoned Panarctic Oil well site in the middle of the Fosheim Peninsula (Gemini E-10) is 502 m thick (Judge et al. 1981). Temperatures sensors installed in two 17m core holes near Eureka indicate that the depth of zero annual amplitude is ~15.4 m with a stable ground temperature of -16.5°C measured over a 10 year period. Throughout the region, the active layer is relatively thin with a mean thickness of ~60 cm (based on 10 shallow temperature cables) depending on local microclimatic conditions, however, the active layer can be as shallow as 30 cm and as deep as 90 cm. Ground ice is widespread throughout the study area; pore ice and thin lenses of segregated ice are found in abundance and often grade into massive ice. Ice wedges are ubiquitous over much of the ESL and occur in all surficial materials, although they vary in density, depending on soil type. Widespread massive ice is responsible for extensive retrogressive thaw slump activity. The ice-rich nature of the silty-clay surficial materials is conducive to the occurrence of active layer detachments (Lewkowics 2007). Patterned ground is relatively absent except for areas of tussock tundra.

3 GROUND ICE

The term ground ice refers to all types of ice contained in freezing and frozen ground (van Everdingen 1998) and even though it is an important geological component of permafrost not all permafrost contains ground ice. In areas of continuous permafrost ground ice takes many forms and varies greatly in its distribution and volume. Processes associated with permafrost formation can lead to ice accumulating in volumes far in excess of normal soil moisture conditions; in some cases far in excess of saturation. In areas where ground ice is present it plays a major role in landscape evolution.

3.1 Previous Research

The systematic investigation of ground ice conditions in the Eureka Sound Lowlands began in 1989 (Pollard 1991). Ground ice mapping and stratigraphic research identified a close association between massive ground ice occurrence, the Holocene sea level history and the distribution of marine sediments (Pollard and Bell 1998, Pollard 2000a). According to Bell (1996) the Holocene marine limit in the ESL is 145-150 m. Initially 7 areas were identified where clusters of retrogressive thaw slumps occurred and formed the basis for stratigraphic analysis (Couture and Pollard 1998, Pollard 2000a). A preliminary model linking late Pleistocene/early Holocene glaciation and runoff, marine deposition, Holocene emergence and permafrost formation was developed by Pollard and Bell (1998) and Pollard (2000b). Other studies involved detailed estimates of ground ice content (Couture and Pollard 1998), analysis of retrogressive thaw slump headwall retreat (Grom and Pollard 1998) and modelled projections of increased active layer depths (Couture and Pollard 2007). More recently our research has focused on different aspects of geophysical analysis (Thompson et al. 2012), microbiology (Juck et al 2005, Stevens et al 2007, 2008) and planetary analogues (Haltigin et al 2012). All of these studies involved detailed analysis of ground ice conditions and thermokarst and formed the basis for a program of ongoing monitoring and the establishment of long-term field sites.

3.2 The Nature and Distribution of ESL Ground Ice

Permafrost systems in general are complicated by the uneven and unpredictable distribution of ground ice. Two forms of ground ice that are particularly important in the ESL in terms of the response of permafrost landscapes to climate change (warming) are ice wedges (polygons) and thick, nearly pure bodies of horizontally extensive massive ice. Ice wedges occur throughout the study while massive ice deposits tend to be linked more closely to the surficial geology. Melting massive ice and ice wedges lead to widespread ground subsidence (thermokarst) and as such ground ice play a major role in the evolution of ESL landscapes. The volume of excess ice is the main variable that determines the potential magnitude of thermokarst. Volumetric ice contents in marine sediments in the ESL frequently exceed 60-70% (Pollard 2000a & b). Detailed information about the active layer is also needed before a realistic prediction about thaw subsidence can be made. Areas where the active layer is relatively thin, like the ESL, are vulnerable to small increases in summer temperature and/or the thaw season duration because its buffering capacity is limited. Small changes in the active layer can normally occur without serious impact.

Massive ground ice and ice-rich sediments are widespread below the Holocene marine limit (150 m a.s.l.) where its distribution, content, and stratigraphic character are closely related to the nature of the fine-grained marine sediments that blanket much of the area (Pollard 2000a&b). The thickness of marine deposits is extremely variable ranging from a thin veneer roughly 100-150 cm thick to extensive gently sloping plateau features with 10-15 m of horizontally laminated silty clay and fine sandy silt underlain by weather Tertiary bedrock. The most significant ground ice exposures occur in association with the latter (Figure 3).

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Figure 3. A thick (3.4 m) body of layered massive ice overlain by 7 m of horizontally laminated silty clay. This is one of 24 exposures that formed near Eureka in 2011-2012.

Bell (1996) mapped the distribution of marine sediments and explained their occurrence as a function of late Pleistocene and early Holocene glaciation and Holocene sea level change. Paleoclimate records for the Canadian high Arctic indicate a continuously warm period from roughly 10ka to 5 ka BP followed by a cooling trend for 4-5 ka and then the current period of warming beginning in the early 1900s (Bradley 1990, Koerner and Fisher 1990, Taylor 1991). By combining the relative sea level record with ground ice elevation data (Figure 4) we show that much of the massive ice formation corresponds with a period of rapid emergence during the early Holocene when temperatures were not unlike the present. This period also corresponds to the deglaciation of highland ice caps that occurred on many of the structural ridges higher than 500 m that surround the ESL (Bell 1992). We propose that runoff from melting ice caps were probably the main source of water for ice formation. As emergence progressed permafrost quickly began to aggrade into newly exposed marine sediments forming mainly pore ice and thin horizontal ice lenses. These lenses have the similar salinities as the enclosing sediments (~6-7‰). As freezing encroached on the underlying fractured and weathered Tertiary bedrock the ice lenses are much thicker and tend to form a reticulated cryostructure.

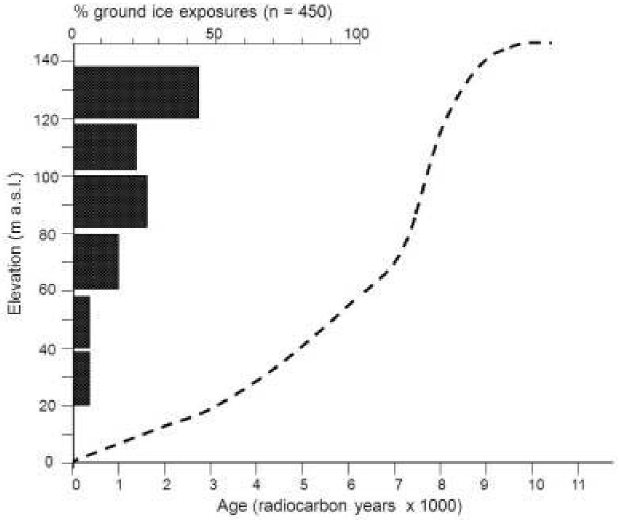

Figure 4. Elevational distribution of ground ice exposures superimposed on a provisional relative sea level curve for northern Fosheim Peninsula (Bell 1996). This diagram is modified from Pollard and Bell 1998 to include our entire ground ice data base.

The fractured and unconsolidated nature of the Tertiary bedrock forms an excellent aquifer and would have been saturated with meltwater from melting ice caps. As the freezing front encroached on the underlying bedrock the continuous supply of meltwater probably mixed with brackish seawater supported the formation of massive ice. The ice body would continue to form as long as temperatures were sufficiently cold and ground water supplied. At a critical depth the combined thickness of sediment and ice would generate sufficiently high overburden pressures to limit ice formation. At several sites the base of the massive ice includes fragments of the underlying bedrock. Massive ice bodies tend to have very low salinities (2-3‰) and isotopic signatures∂(^{18} from -36 to -28‰ SMOW) that suggest that glacial meltwater is the main source for the massive ice.

Since 1990 approximately 450 different natural exposures of massive ice have been mapped including more than 100 that were studied in detail. Most of these exposures are in the headwall of retrogressive thaw slumps. Additional information of ground ice stratigraphy was obtained from GPR surveys and data from 50 boreholes ranging in depth from 2-17 m. Figure 5 shows the location of the Holocene marine limit, together with the marine sediments and recent ground ice exposures.

3.2.1 Cryostratigraphy

Most exposures include horizontally layered massive ice several meters thick conformably overlain by 1-7 m of massive to weakly laminated marine sediment. A typical section includes a thin surface layer 5-20 cm thick of massive fine sand with a few large sub-rounded clasts (gravel to cobble sized) that grades into a massive fine silty sand. The fine sand unit varies in thickness between 1-3 m. This unit grades into a finely laminated silt and silty-sand. Layers (laminae) 1-1.5 cm thick of well sorted silt and fine sand are interpreted as rythmites associated with episodic glacial melt water deposition into a shallow marine (brackish) system. Soil salinities range between 2-5‰. The active layer is typically 30-90 cm deep. In the Eureka Sound lowlands the pattern in ground ice is closely linked to the thickness of marine sediments, and the presence of terraced marine deposits has proven to be an extremely good indicator of massive ice.

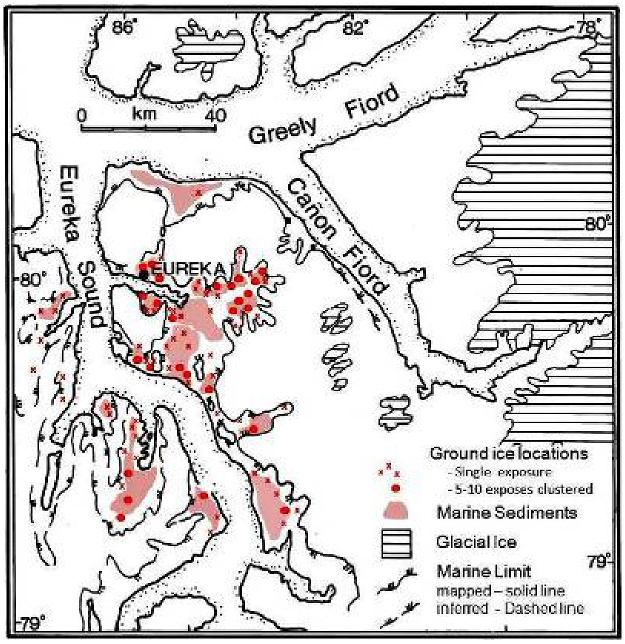

Figure 5. Study site map showing the Holocene marine limit at~145m asl in relation to the distribution of marine sediments and ground ice exposures. A veneer of marine sediments blanket all surfaces below 145M asl but in places forms thick laminated terraces (shown in pink)

In the ice-rich areas below marine limit three patterns in ice content profile occur. The first is marked by moderately high ice contents (20-30% by volume), occurring as pore ice and thin discontinuous ice lenses and layers near the base of the active layer. Beneath the active layer, ice contents increase sharply so that at depths of roughly 1.0-1.3m they reach 60-99% where layers of pure ice and muddy ice predominate; the sediment in the ice is mainly silt or clay sized. Pore and lens ice occur where the lithology changes to coarse sandy material or weathered bedrock. The second pattern includes a layer of weakly laminated sandy silt 5-7 m thick with ice contents ranging from 8-15% near the base of the active layer, to 12-17% at depths of 2-3 m. Near the base of this unit, ice contents increase to 30-40% as the clay and silt contents increase. Pore and lens ice grade into more regular vertical and reticulate patterns of ice veins. At depths ranging from 7-10 m, faintly laminated silty clay materials abruptly change to massive ice with fragments or layers of silty clay (60-70% ice) to pure horizontally foliated massive ice (90-100% ice). The third pattern is intermediate between the first and second patterns. In this case, ice contents increase gradually (15-30%) with depth usually reflecting a rise in silt content. At a depth of roughly 4-5 m, interbedded layers of silty ice and icy silt with ice contents of 65-78% (sometimes higher) predominate for up to 5 m. At a depth of 9-10 m ice contents decrease to 50-60% and below this depth, they drop off dramatically (5-10%) as materials grade into a weathered bedrock substrate.

4 THERMOKARST

The nature and magnitude of thermokarst is directly related to two important variables, (1) the thermal stability of the upper part of permafrost, including the depth of the active layer, and (2) ground ice contents. Retrogressive thaw slumps are one of the more spectacular forms of back-wearing thermokarst. Most of the 450 natural massive ice exposures were in retrogressive thaw slumps (Figure 6).

Figure 6. A large retrogressive slump with a 20 m headwall. Note the person for scale.

The average length of exposed headwall is ~87 m (the longest was >350 m) and average headwall height is 5-7 m. The highest head walls are over 20m (see Fig 6) but heights around 5-10 m are common. The headwall position of 30-40 slumps is being monitored by a combination of dGPS, total station surveys and tape surveys. In addition late July thaw depths (active layer) are also being monitored in the vicinity of the headwall. In some cases sites are measured annually but most are surveyed every second or third year. With the exception of only a few slumps that have remained active through the entire study most (70%) have stabilized after 7-8 years, (20%) after 10 years and (4%) stabilized only 2-3 years after their initiation. Slumps stabilized very quickly following 2 or 3 years of rapid retreat. Stabilization occurs when the sediment from the active layer buries the head wall. Retreat rate varies across the headwall in each slump; the maximum retreat at a point along the headwall can be misleading so we have been taking the average of the most active 10m section. Over 25 years of observation the average retreat for the sites monitored is 6.9 m/yr, the highest annual retreat was 23.9 m. A comparison between average annual retreat and mean annual air temperature does not show any significant relationship however there is a strong relationship between the years with highest annual retreat and the warmest mean July air temperatures or summers with above average thaw degree days. The number of (active) slumps based of aerial surveys tends to correspond with years with the warmest mean July temperatures. Between 1991-1995 there were 85-90 active slumps on the Fosheim Peninsula, from 1995-2000 the number of active slumps dropped to less than 60, from 2000-2005 the number of active slumps went up to 75-77, since 2005 the number has risen annually by 10% peaking in 2012 when 240 slumps were documented. The 2011 and 2012 summers represent a significant jump in thermokarst activity (Section 2.1 notes that mean July temperatures in 2011 and 2012 were 9.8 and 8.7, respectively). Not only did the number of new slumps nearly double but headwall retreat rates were the highest measured ranging between 6-20 m/yr. and averaging >10m. Another interesting patter was the tendency for some areas to develop multiple slumps (Figure 7).

Figure 7. Our South Slidre River study site alone had 43 active retrogressive thaw slumps including Figure 6 (red arrow) and several others of similar magnitude. The previous maximum slump number for this site was 13 in 1995).

The increased availability of reasonably priced high resolution satellite imagery for the high Arctic (IKONOS, QuickBird, WorldView 2 & 3) is changing the way we are monitoring the more visible types of thermokarst. Beginning in 2012, a new component of our monitoring program involves mapping onto geo-referenced satellite images dGPS data from our long-term data base and ongoing surveys. For example Figure 8a & b are WorldView2 satellite images from 2009 and 2012 from the Eureka area with dGPS (Trimble 5800) headwall collected in 2013 and 2014 mapped onto the 2012 image. Figure 8b is an enlargement of the upper right corner of Figure 8a. Many of the retrogressive thaw slumps mapped in 2012 do not appear on the 2009 image. The arrows in Figure 8a represent slumps in our data base which have at least 2 years of survey data. For example Slump A on Figure 8b does not appear on the 2009 image, but survey data from 2012-14 show that the length of its headwall has increased by 24 m (35%) and the headwall position has retreated more than 20 m over 40% of its length.

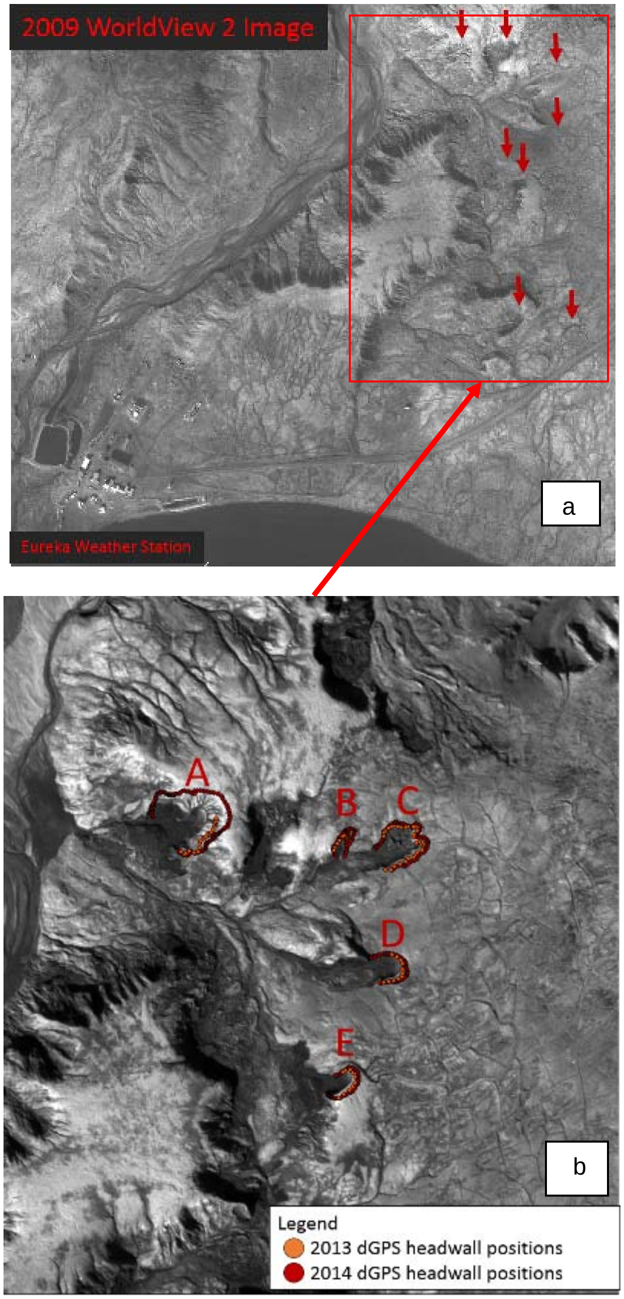

Figure 8. WorldView 2 images showing retrogressive thaw slumps in the Eureka area. The Eureka weather station is located in the lower right corner of Figure 8a. Also note the distinct marine sediment plateau in Figure 8a.

4.1 Ice Wedge Subsidence

The upper 7 m of frozen sediment of the Fosheim Peninsula is estimated to be approximately 49% ground ice by volume including 2-4% as wedge ice (Couture and Pollard 1998). The average ice wedge width and depth are 1.46 m and 3.23 m based on a survey of 150 ice wedges (coring and natural exposures). In some higher level marine sediments ice wedges are up to 2.5m wide and up to 5.5 m deep. Well-developed ice wedge polygons are ubiquitous in areas covered by marine sediments (Figure 9) and poorly developed on higher weathered bedrock surfaces.

Since 2000 many of the stable ice wedge polygons at our monitoring sites have exhibited significant subsidence along the ice wedge troughs (Figure 9a). In 2005 we began measuring the relative change in the trough morphology by differential leveling (Figure 9b).

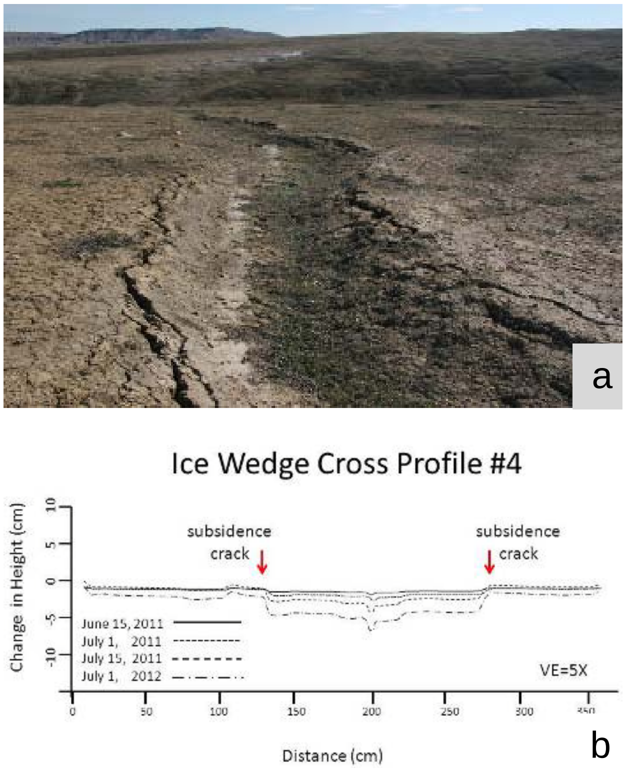

Figure 9. Gradual subsidence of ice wedge troughs is occurring through the ESL; 9a shows typical down faulting of sediments directly above the ice wedge. 9b shows the results of a series of surveys during 2011.

For much of the Arctic enlarged troughs are not unusual and thaw degradation along the top of ice wedges leads to the formation high-centered polygons and tundra ponds. In polar desert settings like Eureka, ice wedges often lack a surface expression (no trough structure). In the ESL where active layers are generally thin, the top of the ice wedge is usually in equilibrium with current active layer depths. As a result, any increase in the active layer depth will result in a subsidence equal to the increased thaw depth. Figure 10 illustrates how an increase in active layer depth of 5 cm can produce a 4 cm subsidence over the ice wedge in a single season, since it may take several years for the new active layer to stabilize the total collapse may be much higher. Figure 9a shows the formation of normal faults and illustrates the initial phase of this process. The proposed model is based on the differential thaw subsidence due to differences in ice content between the top of permafrost and the ice wedge.

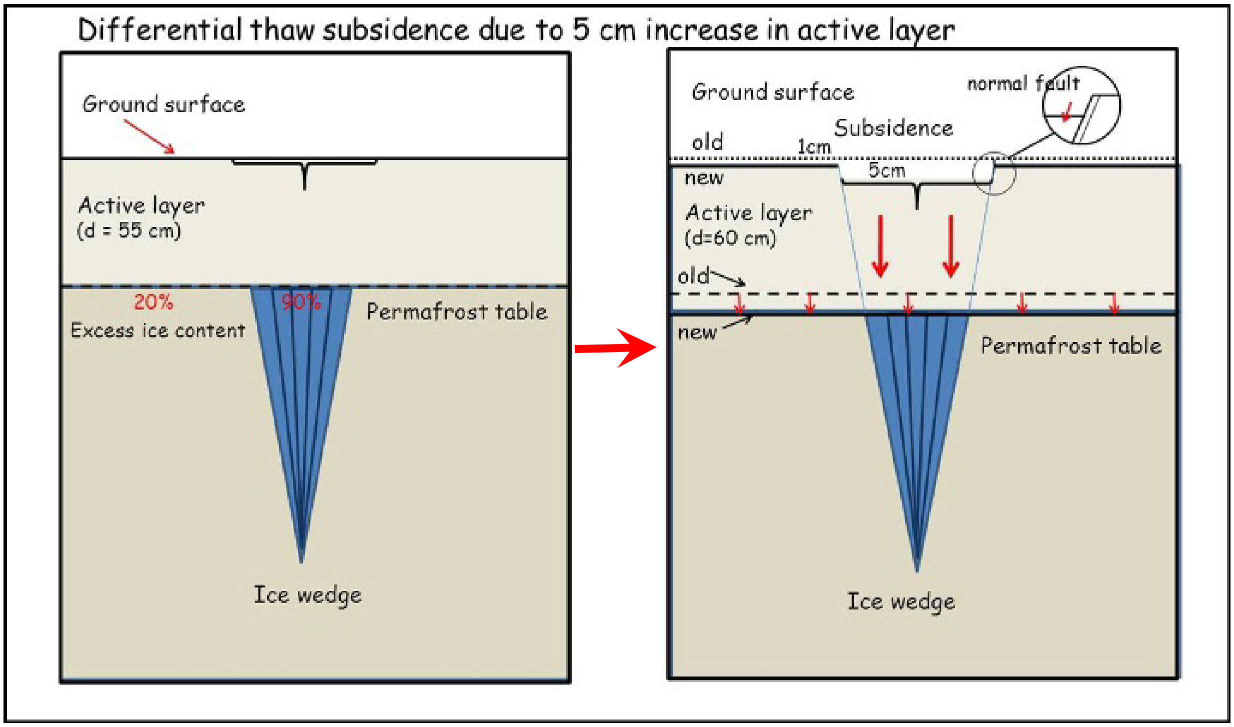

Figure 10. This figure presents a model to explain the differential thaw subsidence occurring in well-drained areas of the ESL. Given the extremely dry polar desert conditions and lack of surface water and vegetation differential subsidence is driven entirely by differences in ice content between the ice wedge and the upper few decimeters of permafrost.

4.2 Ice Wedge Melt Out

Several studies have characterized thermal erosion of ice wedges by running or standing water, the development of beaded drainage and the formation of tundra ponds. In 2012, several completely thawed ice wedge networks were documented (Figure 11). However, two interesting aspects of their occurrence are; 1) the very short time period in which they formed (a single thaw season-2012), and 2) the absence of running water. In each case the wedge was rapidly back wasting aided by sloughing of the sediments overlying the wedge (i.e. small scale active layer slides). At several sites the main body of the wedge melted so rapidly that polygon centers remained undisturbed. At several sites a single ice wedge oriented parallel to the main slope collapsed. In all cases the process was similar to a retrogressive thaw slump only back wasting was limited to the ice wedge.

5 CONCLUSIONS

The main conclusions to be drawn from this ongoing research program are: (1) despite low MAATs, cold winter conditions and short summer (thaw) seasons ice-rich permafrost, even in high Arctic polar deserts, is extremely vulnerable to increased summer temperatures, (2) polar desert systems characterized by relatively thin active layers and lacking surface vegetation will respond very quickly to warming with multiple, large retrogressive thaw slumps forming in only 1 or 2 seasons, (3) given their ubiquitous distribution in continuous permafrost ice wedges are a potentially significant form of ground ice. In areas characterized by a thin active layer ice wedges will be one of the first ground ice features to respond to increased summer warming and entire polygon networks are capable of collapsing in only 1 or 2 seasons. (4) Long-term monitoring and observation programs are essential components of climate change

Figure 11. Catastrophic melt out of ice wedges in 2011 and 2012

research and are necessary to properly interpret and parameterize climate change models.

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