Abrupt increase in permafrost degradation in Arctic Alaska

M. Torre Jorgenson,[1] Yuri L. Shur,[2] and Erik R. Pullman[1]

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[1] Even though the arctic zone of continuous permafrost has relatively cold mean annual air temperatures, we found an abrupt, large increase in the extent of permafrost degradation in northern Alaska since 1982, associated with record warm temperatures during 1989–1998. Our field studies revealed that the recent degradation has mainly occurred to massive wedges of ice that previously had been stable for 1000s of years. Analysis of airphotos from 1945, 1982, and 2001 revealed large increases in the area (0.5%, 0.6%, and 4.4% of area, respectively) and density (88, 128, and 1336 pits/km^2) of degrading ice wedges in two study areas on the arctic coastal plain. Spectral analysis across a broader landscape found that newly degraded, water-filled pits covered 3.8% of the land area. These results indicate that thermokarst potentially can affect 10–30% of arctic lowland landscapes and severely alter tundra ecosystems even under scenarios of modest climate warming. **Citation:** Jorgenson, M. T., Y. L. Shur, and E. R. Pullman (2006), Abrupt increase in permafrost degradation in Arctic Alaska, *Geophys. Res. Lett.*, *33*, L02503, doi:10.1029/2005GL024960.

1. Introduction

[2] Although thawing and settling of ice-rich terrain (thermokarst) is widespread in the subarctic zone where permanently frozen ground (permafrost) is discontinuous [*Jorgenson et al.*, 2001; *Halsey et al.*, 1995], the ground in the arctic zone of continuous permafrost has been considered stable because of much lower mean annual air temperatures (annual means −6 to −12°C). The risk for thaw subsidence in arctic lowlands is substantial, however, because of the high volume of ground ice at the top of the permafrost [*Nelson et al.*, 2001]. Under the cold climatic conditions in the Arctic, a polygonal network of wedge-shaped ice bodies form beneath a thin layer of seasonally thawed soil due to contraction cracking caused by large fluctuations in winter temperature [*Leffingwell*, 1919; *Lachenbruch*, 1962; *Mackay*, 1990]. Between the active layer and the ice wedges is a very thin transient layer [*Shur*, 1977] of usually frozen soil that adjusts in thickness to annual climatic variations and prevents the summer thawing front from encountering the ice wedges and initiating thaw settlement. In the area that we studied, large ice wedges typically are 2–4 m across the top and >3000 years old [*Jorgenson et al.*, 1998], indicating that the ice wedges have been stable over a long period under modest climatic fluctuations (Figure 1). The transient layer, however, provides only limited protection, and thermokarst due to human disturbance frequently has been observed in the Arctic, even at cold temperatures. Here we report an abrupt increase in natural degradation of ice-wedges during a period of an unprecedented 2–5°C increase in mean annual ground temperatures (MAGT) in northern Alaska since the 1980s [*Osterkamp*, 2003; *Clow*, 2003].

2. Methods

[3] We evaluated the degradation of ice wedges in northern Alaska at three spatial scales that included: (1) field surveys within two small, intensive sites (0.6-km^2); (2) photo-interpretation of a time-series of aerial photography within the two intensive sites; and (3) image processing of the spectral characteristics of aerial photography for two larger areas (14.5-km^2). During field surveys in Aug. 2003 and 2004, we sampled 43 plots in the two intensive areas (C1 and C2) 10–40 km west of the Colville River and 20 km south of the coast to assess changes in vegetation, microtopography, and soils associated with ice-wedge degradation evident on the ground. For vegetation, the cover of dominant live and dead species was visually estimated to evaluate patterns of plant mortality and recovery. For microtopography, relative elevations of the ground and water surfaces were surveyed with an auto-level. The stratigraphy of soil plugs from the active layer and shallow 1-m cores from the underlying permafrost was described using standard soil methods with special emphasis in differentiating fibrous peat comprised of varying plant remains. Thaw depths were determined with a metal tile probe.

[4] For the photo-interpretation of ice-wedge degradation over time, we delineated and classified thermokarst pits on photography from 4 July 1945 (1:45,000 scale, B&W), July 1982 (1:63,000 scale CIR), and 14–15 July 2001 (1:18,000 scale, color orthophoto mosaic by Aeromap, Inc., Anchorage, Alaska) to assess temporal changes in various stages of thermokarst within the two intensive sites. To compensate for differences in visibility of pits due to the varying resolution of aerial photographs, only large pits (>12 m^2) with water or aquatic sedge marshes were used for analysis.

[5] We classified the spectral characteristics of 1945 and 2001 photography with image-processing software (Imagine 8.2, ERDAS, Inc.) to map deep (>30 cm) waterbodies across two larger areas (B and C, 14.5 km^2 each). Spectral analysis was not conducted with the 1982 photography because the scale was too small to consistently differentiate thermokarst pits and because the spectral classification for shallow waterbodies was poor. A ecological land survey (ELS) map for the area [*Jorgenson et al.*, 2003] was used to exclude lakes and streams from the analysis because wave reflection in 2001 photography and ice in 1945 interfered

[1]ABR, Inc., Fairbanks, Alaska, USA.
[2]Department of Civil and Environmental Engineering, University of Alaska, Fairbanks, Fairbanks, Alaska, USA.

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Figure 1. View of massive ice wedges near the ground surface in arctic Alaska (above). Thawing of the massive ice causes the surface to collapses and impound water (below, different site).

with spectral classification of larger waterbodies. For 1945, orthophoto mosaics were created for the areas and were balanced to account for systematic variation of illumination across the photos. Water was differentiated by choosing the grayscale threshold value that best depicted the transition from wet sedge meadows (water < 10 cm deep) to deeper water areas consistent with the waterbodies photo-interpreted for the two intensive areas. For 2001, waterbodies were identified by a two-step classification procedure using three bands in the true-color image. First, an unsupervised classification was done and each pixel was assigned up to five ranked classes. Second, a fuzzy classification was run to determine the pixel's most likely class based on surrounding pixel values to better group pixel associated with tiny waterbodies. The resulting set of classes then was evaluated manually to determine which classes best matched the waterbodies photo-interpreted in the two intensive areas. The classified images from both years were overlain to create a four-class image of landscape change: (1) unflooded in both years; (2) flooded in 1945, unflooded in 2001; (3) flooded in both years; or (4) unflooded in 1945, flooded in 2001. The ELS map was used to differentiated waterbodies in upland areas (e.g., alluvial-marine and eolian deposits) and lowland areas (e.g., drained-lake basins, swales) for analysis.

3. Results

[6] We classified ice-wedge degradation into six stages based on differences in trough depth, water depth, soil stratigraphy, and vegetation evident in the field: (1) undegraded ice wedges with no evident surface changes; (2) initial degradation with barely evident settlement associated with thawing of the transient layer and greening of tussocks; (3) intermediate degradation with obvious settlement, shallow standing water, and robust green tussocks; (4) advanced degradation with deep, water-filled pits and dead submerged tussocks; (5) initial stabilization with robust aquatic sedges in shallow water; and (6) advanced stabilization associated with thick peat accumulation, reestablishment of a permafrost layer above the ice wedges, reduction of surface water, and establishment of mosses (stages 4 at bottom of Figure 1; stages 1, 2, 4, 6 shown in Figure 2). This evolution is accompanied by large changes in mean trough depth, total soil thickness above the ice wedge, frozen soil thickness above the ice wedge, and thickness of new peat layers (Figure 2). Changes in the active-layer depth, however, were small because of the limited amount of soil over the ice wedges.

[7] Photointerpretation of ice-wedge degradation within two 0.6 km^2 intensive areas (C1 and C2) revealed that the area of thermokarst pits increased slowly from 0.5% in 1945 to 0.6% in 1982, then increased abruptly to 4.4% by 2001 (Figure 3). Similarly, pit density (i.e., the density of pits $>12\ m^2$) increased slowly from 88/km^2 in 1945 to 128/km^2 in 1982, then rapidly to 1336/km^2 by 2001. This 74-fold increase in the rates of change in areal extent between the 1945–1982 and 1982–2001 periods was due mostly to the

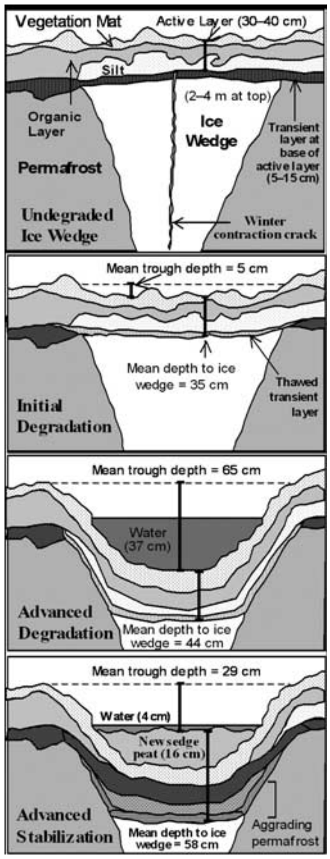

Figure 2. Stages of ice-wedge degradation, including an undegraded ice wedge with a thin protective soil layer; initial degradation with robust tussock growth; advanced degradation with a water-filled pit; and advanced stabilization with new peat. Intermediate degradation and initial stabilization stages are not shown. Data represent means from field surveys at 43 plots.

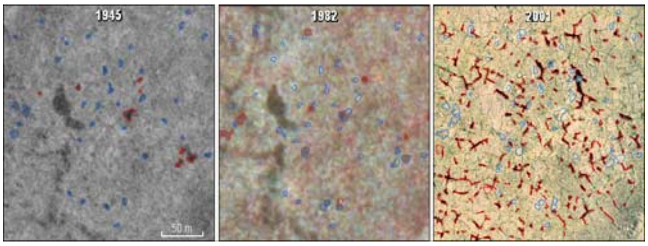

Figure 3. Abundance of degrading (advanced stage only in red) and stabilizing (initial and advanced stages in blue) thermokarst pits within the 0.6 km^2 intensive study area C1 (also see Figure 4) mapped on airphotos taken in 1945, 1982, and 2001. Note that: (1) degrading pits become stabilized by the time of the subsequent photography (19–27 yrs); (2) few degrading pits in 1982 continued to degrade in 2001; (3) all the other degrading pits in 2001 were not present in 1982; and (4) the drainage of the large wet meadow (dark patch left of center in 1982) into adjacent thermokarst pits by 2001.

initiation of new pits and not the expansion of old pits. In contrast, the percent area of stabilizing pits (initial and advanced combined) showed only gradual changes over the three dates (1.6%, 2.2%, and 3.0%, respectively), as earlier degrading pits became stabilized. The analysis reveals that initial and advanced degradation (stages 1 and 2, Figure 3) can happen within 20 yrs (possibly within 10 yrs), and that advanced stabilization is achieved within another 20–30 yrs.

[8] Spectral analysis of waterbody characteristics over two larger areas (29 km^2 total) provided similar results, indicating that ponding in upland areas where thermokarst pits are more distinct and easily differentiated, increased from 1.7% in 1945 (a wet year) to 4.3% in 2001(a dry year). When both uplands and lowlands are combined, new pits covered 3.8% of the overall area (area C only in Figure 4). These analyses also reveal that there has been a substantial redistribution of water from round, flooded polygon centers and low-lying basins into linear degrading troughs. The mapping of waterbodies associated with advanced degradation provides only a minimal estimate of the extent of recent degradation, however, because ground-based observations indicated that many collapsing polygonal troughs over ice-wedges were in an initial stage of degradation that could not be mapped. Because ice wedges occupy 10–30% of the volume of the top 2 m of terrestrial deposits [*Leffingwell*, 1919; *Brown*, 1967; *Jorgenson et al.*, 1998], the analysis indicates that at least 15–40% of the ice wedges have undergone advanced degradation.

4. Discussion

[9] These changes highlight several important processes. First, there has been an abrupt increase in ice-wedge degradation since 1982 that appears to be beyond normal rates of change in landscape evolution. The prevalence of dead tussocks, which originally develop slowly over centuries [*Fetcher and Shaver*, 1982], indicates that the surface was stable for at least centuries. The presence of large ice wedges, which require millennia to develop [*Jorgenson et al.*, 1998; *Fortier and Allard*, 2004], indicate that the wedges were stable for a long period prior to the recent degradation. Second, sites that reach the advanced-degradation stage become stabilized quickly by the accumulation of vegetation and organic matter. Although there is a positive feedback of accumulating water that initially enhances further heat gain and degradation, biological processes quickly develop to provide a strong negative feedback to limit the vertical extent of degradation. Third, once stabilized, the tops of ice wedges are much deeper and better protected from subsequent degradation because of the added peat. This stabilization process indicates the processes are not cyclical, reinforcing the conclusion that the recent degradation is a highly unusual phenomenon. The thick sedge peat and complex ice structures that form over wedges that have previously degraded and stabilized was lacking on the recently degraded wedges we studied.

[10] Although this study only quantified recent thermokarst within a relatively small area west of the Colville River where new high-resolution photography was available, we have also photographed similar changes near Barrow, Cape Simpson, Prudhoe Bay, Kaktovik, and the Noatak River. Our observations suggest, however, that the extent of degradation was more severe in our study area than in these other areas. We speculate that this may be due to larger increases in permafrost temperatures around our study area than in other areas in northern Alaska [*Clow*, 2003], or to variations in ice-wedge volume.

[11] We infer that the degradation was due to warm summer temperatures that occurred during 1989–1998, and probably was initiated by extreme summer weather in 1989. At the Kuparuk Oilfield 60 km to the east, the summer thawing degree-day sum was 43% higher, and precipitation was 60% higher, in 1989 than the 1983–2002 averages (667 TDD and 72 mm, respectively; Figure 5). Conditions of both elevated summer temperatures

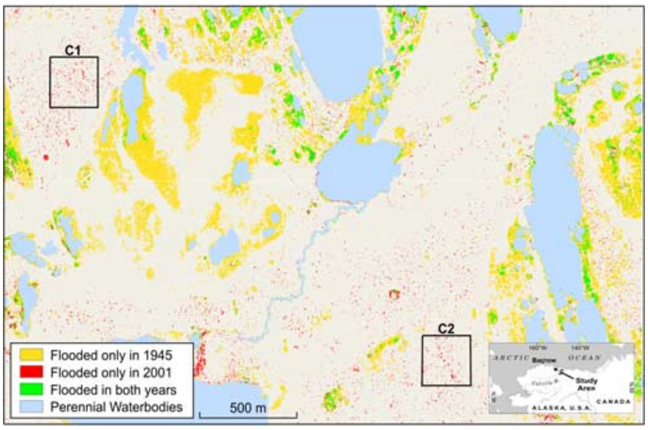

Figure 4. Abundance of new waterbodies attributed to ice-wedge degradation (red) within a 29 km^2 west of the Colville Delta on the central Beaufort Coastal Plain in northern Alaska. The prevalence of waterbodies in drained-lake basins (yellow) indicates 1945 was much wetter than 2001, therefore, the new waterbodies in 2001 were due to thermokarst, not precipitation or seasonal differences in photography. Deep lakes were excluded from the analysis. More detailed analysis of changes in area C1 are present in Figure 3.

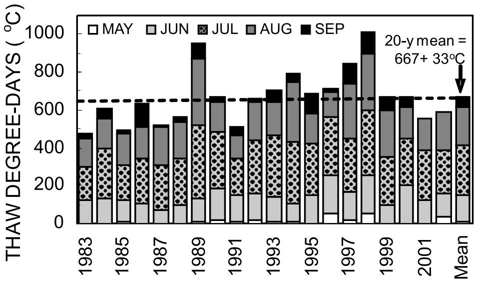

Figure 5. Summer thawing-degree day sums (based 0°) for the Kuparuk Airport, 70 km east of the study area.

and precipitation are particularly conducive to increasing thaw depths, because wetter soils conduct more heat down from the surface than do dry soils. Active-layer measurements near Prudhoe Bay indicate that thaw depths in 1989 were 13–25 cm deeper than those during the previous two years [*Romanovsky et al.*, 2003], sufficient to thaw the transient layer and some underlying ice. Regional analyses indicate that MAGT near the study area increased by up to 5°C, whereas, the central and eastern portion of the Beaufort Coastal Plain had more modest increases of 2–3°C since the mid-1980s [*Clow*, 2003; *Romanovsky et al.*, 2003]. Geothermal analysis of data from deep boreholes indicates that MAGT during the late 1990s and early 2000s were the warmest in last two centuries [*Lachenbruch and Marshall*, 1986; *Zhang and Osterkamp*, 1993; *Osterkamp*, 2003], and paleoecological reconstructions indicate that recent air temperatures (particularly 1989 and 1998) were the warmest in the last millennium [*Intergovernmental Panel on Climate Change*, 2001]. The recent thermokarst of large ice wedges developed over thousands of years provides additional evidence that recent temperatures exceed natural climatic variability.

[12] The presence of some ice wedges with thick accumulations of peat over them, indicates that there may have been some earlier warming that also caused degradation. Although, minor levels of degradation may happen continually due to microtopographic and surface-water readjustment as wedges grow, we speculate that the degradation evident in 1945 might have been caused by rapid warming from ~1850 to 1950. The recent degradation, however, is one to two orders of magnitude greater. What also is evident is that the degradation is caused by the relative temperature change in thaw depths, not the crossing of some absolute threshold. Because ice wedges form just beneath the active layer in equilibrium to a long-term regime of summer temperatures (see top of Figure 1), a change in MAGT from −11 to −8°C can have a similar effect on changing the active-layer regime as a change from −6 to −3°C. Thus, ice-wedge development leads to permafrost instability even in very cold climates.

[13] The abrupt increase in the extent and rate of permafrost degradation has significant ecological implications for tundra ecosystems. The degradation of ice wedges has caused a substantial redistribution of surface water from the adjacent tundra to the degraded trough network and broadened the ecological impacts. If trends continue, 10–30% of the terrestrial landscape may be directly affected, greatly altering local biodiversity, plant communities, and wildlife use; modes of soil respiration and organic-matter accumulation; and sinks and sources of trace gases.

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References

Brown, J. (1967), An estimation of volume of ground ice, Coastal Plain, northern Alaska, technical report, Cold Reg. Res. and Eng. Lab., Hanover, N. H.

Clow, G. (2003), GTN-P monitoring network: Detection of a 3 K permafrost warming in Northern Alaska during the 1990s, paper presented at NSF SEARCH Open Science Meeting, Natl. Sci. Found., Seattle, Wash., 27–30 Oct.

Fetcher, N., and G. R. Shaver (1982), Growth and tillering patterns within tussocks of *Eriophorum vaginatum*, *Holarctic Ecol.*, *5*, 180–186.

Fortier, D., and M. Allard (2004), Late Holocene syngenetic ice-wedge polygons development, Bylot Island, Canadian Arctic Archipelago, *Can. J. Earth Sci.*, *41*, 997–1012.

Halsey, L. A., D. H. Vitt, and S. C. Zoltai (1995), Disequilibrium response of permafrost in boreal continental western Canada to climate change, *Clim. Change*, *30*, 57–73.

Intergovernmental Panel on Climate Change (2001), *Climate Change 2001: The Scientific Basis*, Cambridge Univ. Press, New York.

Jorgenson, M. T., Y. Shur, and H. J. Walker (1998), Factors affecting evolution of a permafrost dominated landscape on the Colville River Delta, northern Alaska, in *Proceedings of Seventh International Permafrost Conference*, *Collect. Nord.*, *57*, 523–529.

Jorgenson, M. T., C. H. Racine, J. C. Walters, and T. E. Osterkamp (2001), Permafrost degradation and ecological changes associated with a warming climate in central Alaska, *Clim. Change*, *48*, 551–579.

Jorgenson, M. T., J. E. Roth, M. Emers, S. Schlentner, D. K. Swanson, E. Pullman, J. Mitchell, and A. A. Stickney (2003), An ecological land survey for the Northeast Planning Area of the National Petroleum Reserve–Alaska, report, ConocoPhillips, Anchorage, Alaska.

Lachenbruch, A. H. (1962), Mechanics of thermal contraction cracks and ice wedge polygons in permafrost, *Spec. Pap. Geol. Soc. Am.*, *70*.

Lachenbruch, A. H., and B. V. Marshall (1986), Changing climate: Geothermal evidence from permafrost in the Alaska Arctic, *Science*, *234*, 689–696.

Leffingwell, E. de K. (1919), The Canning River region of northern Alaska, *U.S. Geol. Surv. Prof. Pap. 109*.

Mackay, J. R. (1990), Some observation on the growth and deformation of epigenetic, syngenetic and anti-syngenetic ice wedges, *Permafrost Periglacial Processes*, *1*, 15–29.

Nelson, F. E., O. E. Anisimov, and O. I. Shiklomonov (2001), Subsidence risk from thawing permafrost, *Nature*, *410*, 889–890.

Osterkamp, T. E. (2003), A thermal history of permafrost in Alaska, in *Proceedings of 8th International Conference on Permafrost*, edited by M. Phillips, S. M. Springman, and L. U. Arenson, pp. 863–869, A. A. Balkema, Brookfield, Vt.

Romanovsky, V. E., D. O. Sergueev, and T. E. Osterkamp (2003), Temporal variations in the active layer and near-surface permafrost temperatures at the long-term observatories in northern Alaska, in *Proceedings of 8th International Conference on Permafrost*, edited by M. Phillips, S. M. Springman, and L. U. Arenson, pp. 989–994, A. A. Balkema, Brookfield, Vt.

Shur, Y. L. (1977), *Thermokarst*, Nedra, Moscow.

Zhang, T., and T. E. Osterkamp (1993), Changing climate and permafrost temperatures in the Alaska Arctic, in *Permafrost: Sixth International Conference, Proceedings*, pp. 783–789, S. China Univ. of Technol. Press, Guangzhou, China.

M. T. Jorgenson and E. R. Pullman, ABR, Inc., PO Box 80410, Fairbanks, AK, 99708, USA. (tjorgenson@abrinc.com; epullman@abrinc.com)

Y. L. Shur Department of Civil and Environmental Engineering, University of Alaska, P.O. Box 755900, Fairbanks, AK 99775-5900, USA. (ffys@uaf.edu)