ARCTIC ALASKA
ENVIRONMENTAL CHANGE
FIELD EXCURSION TO THE NORTH SLOPE
JUNE 1-15, 2016

Summer @ UAF

UNIVERSITY OF ALASKA FAIRBANKS

OFF THE BEATEN PATH

Alaska Geobotany Center

BIOL 495/695 Syllabus and Course Reader

BIOL 495/695 Arctic Alaska Environmental Change: 2016 Field excursion to the North Slope

Syllabus for 2016 UAF Summer Sessions Special Topic course, BIOL 495/695, Arctic Alaska Environmental Change: Field excursion to the North Slope, 1-15 Jun 2016

1. Course information

Title: Special Topic, Arctic Alaska Environmental Change: Field excursion to the North Slope
Number: BIOL 495 / 695
Credits: 3
Prerequisites: BIOL 115 & 116, or equivalent introductory physical science course intended for science majors in biology, geology or geography or instructor approval
Location: Murie Building, Room 230
Meeting time: 1 Jun, 9:00 am

2. Instructors and contact information

Prof. Amy Breen (instructor and course leader, has Wilderness First Responder Training), albreen@alaska.edu, International Arctic Research Center and Alaska Geobotany Center, Room 252 in Arctic Health Research Building; Prof. D.A. (Skip) Walker, (instructor) dawalker@alaska.edu; Dave Klein (instructor) dklein7@alaska.edu; Jason Clark (instructor and course manager) jaclark2@alaska.edu

3. Course readings/material:

Readings (see daily readings in the course schedule):
Daily readings: Each day 1-2 papers are required readings that we will discuss over breakfast and/or dinner. The required readings are in the "Syllabus and Course Reader". One student will be selected randomly each day to help instructors lead discussions.

Course library: The course also carries a book box with many other general references, relevant papers and books. Students can check these out for personal reading and as background for their course projects. The contents of the library are listed in the course "Syllabus and readings".

Good general references: These references provide a good overview of the Dalton Highway and research at the Toolik Field Station.

1. Brown, J. and Kreig, R. A. 1983. *Guidebook to permafrost and related features along the Elliot and Dalton highways, Fox to Prudhoe Bay, Alaska. Fairbanks,* AK: Division of Geological and Geophysical Surveys.
2. Huryn, A. and Hobbie, J. 2013. *Land of Extremes: a natural history of the Arctic North Slope of Alaska*. University of Alaska Press, Fairbanks.
3. Walker D. A., Hamilton, T. D., Ping, C.-L., Daanen, R. P. and Streever W. W. 2009. *Dalton Highway Field Trip Guide for the Ninth International Conference on Permafrost. Fairbanks, AK*: Division of Geological and Geophysical Surveys.
4. Hobbie, J. and Kling, G. 2014. *Alaska's Changing Arctic.* Oxford, New York.

Course equipment
The course will provide a large group meeting and eating tent, Coleman stoves, water purification, first aid kit, satellite phone, generator, and vehicles. Students will need to purchase food and have money for meals at Coldfoot and Prudhoe Bay. Students will need to enroll early and contact the organizers to get a list of required equipment including: tent, sleeping bag, sleeping pad, rain gear, footwear, sun protection, bug protection, personal gear and other camping equipment. For students traveling from abroad or that do not own extreme weather gear, tents, sleeping bags and sleeping pads are available from the course instructors or can be rented from UAF's Outdoor Adventures.

4. Course description:

Course catalog description:
BIOL F495_ Arctic Alaska Environmental Change: Field excursion to the North Slope. 4 Credits. Offered Summer 2016
15-day course, includes 12-day field excursion along the Dalton Highway, Brooks Range, Arctic Foothills Arctic Coastal Plain, Prudhoe Bay. Climate, geology, permafrost, soils, vegetation, wildlife, local people, infrastructure impacts. Special fees apply. Stacked with BIOL F695(3)

More detailed description: This course will consist of:

1. 2 days of preparation with lectures, local field trips in the Fairbanks area and logistics for the excursion.
2. 12 day field excursion
3. 1 day of student presentations when return to Fairbanks.

The trip will have a strong emphasis on Arctic environments, local people, and field sampling.

5. Course goals and student learning outcomes

The goals for the course are to: (1) Provide students with an in-depth field experience of Arctic environments, local people, and the oil industry's environmental research program and application to current Arctic issues. (2) Provide methods of field sampling of Arctic vegetation, soils, and permafrost in a variety of Arctic ecosystems. (3) Visit Arctic research sites, including Finger Mountain, Atigun Pass, Toolik Lake, Imnavait Creek, Happy Valley, Sagwon, and Prudhoe Bay.

6. Instructional method and grading criteria:

2-day preparation in Fairbanks:
Introductory lectures will give an overview of the course and Arctic ecosystems, permafrost and local people along the Dalton Highway. Students will develop a research topic to be examined during the excursion. On the third day students will visit local boreal forest ecosystems and the U.S. Army Cold Regions Research and Engineering Laboratory (CRREL) Permafrost Tunnel at Fox. Students should become familiar with the field guides (Walker et al. 2009, Brown & Krieg 1983, Huryn & Hobbie 2013) for the Dalton Highway route.

12-day field excursion:
The course will follow the route of the Dalton Highway. The course will examine Arctic environments, with in depth examination of the physical, biological, and human responses and adaptations to changing climate. We visit the old mining town of Wiseman to gain an understanding of village. We will establish camps in the Boreal Forest, Brooks Range, Arctic Foothills, and Arctic Coastal Plain — Coldfoot, Galbraith Lake, Happy Valley, and near Deadhorse — where we will camp and spend two days at each location exploring the local vegetation, soils, permafrost, geology, and land-use and climate-change issues. The course will have field lectures, conducted during hikes to different areas, using materials from past and existing research projects in the region. Students will learn the methods of vegetation, soil, and permafrost sampling and collect sample data from representative ecosystems. The course includes visits to the Arctic Research Station at Toolik Lake and the oilfield at Prudhoe with an overview of the environmental research of the oil companies at Prudhoe Bay. We will then return to UAF driving south from Prudhoe Bay to Fairbanks.

1-day presentation of student projects:
At the end of the course students will spend the morning working on their oral presentation that summarizes their observations during the excursion. Students will then present their findings in the afternoon with ample time for group discussions.

Research topics:
Students will develop a research topic that fits with the planned excursion. The topics should focus on descriptive aspects of Arctic environment along the climate gradient. Students should keep in mind that the analysis of the data will be limited by the short time available at the end of the course. At the end of the course, students will present 15-minute oral presentations summarizing aspects of their field observations, focusing on their research topic. Guidelines for these presentations will be handed out at the beginning of the course. Graduate students will also write a 10-15 page research paper focused on some aspect of observations during the course, which will be due 3 Jul 2014.

Academic integrity:
Plagiarism and cheating will not be tolerated. Plagiarism is presenting another's work as new or original without citing your source. For additional detail, see
http://www.uaf.edu/library/instruction/handouts/Plagiarism.html
Please speak with me if you have any questions about how to properly use other people's work.

Attendance policy:
Students are expected to actively participate in both the academic part and expedition part of camp, cooking, clean-up, waste management, emergencies, group decisions, and keeping a cheerful attitude in sometimes difficult field conditions such as rain, cold or snow.

7. Evaluation:

Summary of grading points:

Undergraduate student grading (BIOL 495 students):

Attendance and participation lectures, field trips, and discussions:	200 pts
Field notebooks and plant collections	200
Oral presentation of research topic	200
TOTAL	600 pts

Graduate student grading (BIOL 695 students):

Attendance and participation in discussions:	200 pts
Field notebooks and plant collections	200
Oral presentation of research topic	200
Final research paper	200
TOTAL	800 pts

These criteria may be modified somewhat as the course progresses.
Final grades will be as follows: greater than or equal to 90% = A; 80-89% = B; 70-79% = C; 60-69% = D; < 60% = F.

Graduate student grading:
Graduate students will be graded according to the same criteria as the undergraduate students except the graduate students are required to turn in 3-5 page research paper on a topic of their choice. Guidelines for this paper will be handed out on the first day of class. Due date is 3 Jul. Students should arrange for an incomplete grade if they cannot meet this deadline.

8. Support services:

Students are encouraged to contact the instructor with any questions, or to clarify the lecture or the assignments. We will be happy to review drafts of assignments and answer questions any time. Skip Walker's lab and office is in Arctic Health Research Building Room 254. While in Fairbanks, Amy will reside in Skip's lab. Lab phone 474-2459, Amy's cell phone: 907 750-1311, Skip's home phone: 451-0800.

9. Disabilities services:

The instructor will work with the Office of Disabilities Services (203 WHIT, 474 7043, to provide reasonable accommodation to students with disabilities.

10. Course schedule and reading assignments:

Date	Location	Activity	Reading to be done in preparation for each day
31 May	Fairbanks, Hess commons	Arrival, check into dorm	None
1-Jun	Fairbanks, Margaret Murie Bldg Room 230	**9:00 am:** Cold breakfast in break room **9:15 am:** Introductions, course schedule and expectations, readings & student projects, equipment list **(Amy Breen)** **9:45 am:** Overview of the Arctic system and Dalton Highway (**Skip Walker**) **10:45:** Overview of permafrost systems (**Yuri Shur & Misha Kanevskiy**) **Lunch:** In Murie break room **1:00 pm:** Risk assessment; health & safety (**Matt Irinaga**), equipment check **6:00 pm:** (College Pizzaria), dinner. **Night:** UAF Dorms or elsewhere in Fairbanks	Start on: **Marshall, R**. 1991 (reprint). *Arctic Village: A 1930s Portrait of Wiseman, Alaska,* University of Alaska Press, Fairbanks, pp. 3-44. Finish by breakfast 5 Jun.
2-Jun	Fairbanks, Meet at Arctic Health Research Building West Parking Lot	**Breakfast:** On own* **9:00 am:** Permafrost coring (**Misha Kanevskiy**), boreal forest plants (**Amy Breen**), animals (**David Klein**); meet at Arctic Health Bldg, West Parking Lot, travel together to coring sites **Lunch:** Fast food on way to Fox* **1:00 pm:** CRREL Permafrost Tunnel (Elliot Highway) (**Yuri Shur & Misha Kanevskiy).** Help pack trailer for trip. Get ready for next day's departure. **Dinner:** On own* **Night:** UAF Dorms or elsewhere in Fairbanks	**Kanevskiy, M.** et al. 2008. Late-Pleistocene syngenetic permafrost in the CRREL permafrost tunnel, Fox, Alaska. University of Alaska, Institute of Northern Engineering, Fairbanks, Alaska.
3-Jun	Fairbanks to Coldfoot, Meet at Arctic Health Research Building West Parking Lot	**Breakfast:** Sourdough Sam's **6:30 am:** Final packing, drive to breakfast. **8:00 am:** Drive to Yukon river, insect ecology (**Derek Sikes**) **Lunch:** Yukon River picnic **PM:** Drive to Coldfoot with stop at Finger Mountain **Dinner:** Coldfoot Truck Stop **Night:** Tent camp in Coldfoot vicinity	**Chapin, F. S.** et al. 2010. Resilience of Alaska's boreal forest to climatic change. Canadian Journal of Forest Research, 2010, 40(7): 1360-1370, 10.1139/X10-074

Date	Location	Activity	Reading to be done in preparation for each day
4-Jun	Coldfoot vicinity	**Breakfast:** In camp. **10:30 am:** Wiseman tour (**Jack Reakoff**) **Lunch:** Picnic at Koyukuk River bridge **PM:** Nolan Creek, ecology of willow carr **Dinner:** Cook camp dinner **8:00 pm:** Interagency Visitor Center presentation (**Heidi Schoppenhorst)**. Tent camp in Coldfoot vicinity.	**Marshall, R**. 1991 (reprint). *Arctic Village: A 1930s Portrait of Wiseman, Alaska,* University of Alaska Press, Fairbanks, p. 3-44
5-Jun	Coldfoot to Galbraith Lake	**Breakfast:** Coldfoot Truck Stop. **AM:** Drive to Galbraith Lake, with stops at frozen debris lobes, Sukakpak Mtn, Atigun Pass **Lunch:** Picnic at Atigun Pass summit **PM:** Set up camp. Catch up on readings, plant collections & projects **Dinner:** Cook camp dinner **Night:** Tent camp at Galbraith Lake.	**Daanen, R. P.**, G. Grosse, M. M. Darrow, T. D. Hamilton, and B. M. Jones. 2012. Rapid movement of frozen debris-lobes: implications for permafrost degradation and slope instability in the south-central Brooks Range, Alaska. *Natural Hazards and Earth System Sciences* 12: 1521-1537. **AND Walker, D. A.** et al. 2016. Circumpolar Arctic vegetation: a hierarchic review and roadmap toward an internationally consistent approach to survey, archive and classify tundra plot data. *Environmental Research Letters* doi:10.1088/1748-9326/11/5/055005
6-Jun	Galbraith Lake vicinity	**Breakfast**: In camp **AM:** Overview of releve sampling along Galbraith Creek **Lunch:** In camp **PM:** Atigun Gorge hike, plant collections **Dinner and night:** Tent camp at Galbraith Lake	**Breen, A. L.**, M. K. Raynolds, I. Timling, D. F. Murray & D. A. Walker. 2014. Ecology and Evolution of Plants in Arctic and Alpine Environments, pp. 149-178 *in* Rajakaruna, N., B. Boyd & T. Harris. (Eds.) Plant Ecology and Evolution in Harsh Environments. Nova Science Publishers, Hauppauge, New York.
7-Jun	Galbraith Lake-Toolik Lake	**Breakfast:** In camp **AM:** Brooks Range, Atigun Pass. Wildlife and alpine vegetation and landforms, north side of pass. **Lunch:** Sack lunch at solifluction lobes **PM:** South slope of Atigun Pass, Drive to Toolik **Dinner and night:** Toolik Field Station	**Huryn, A. and J. E. Hobbie.** 2012. **Chapters 1 to 5**, "Introduction", "Bedrock geology", "Glacial geology", "Permafrost and patterned ground", "Habitats and patterned ground", pp. 1-52 in *Land of Extremes*: *a Natural History of the Arctic North Slope of Alaska*. University of Alaska Press, Fairbanks, Alaska.

Date	Location	Activity	Reading to be done in preparation for each day
8-Jun	Toolik Lake vicinity	**Breakfast:** TFS. **AM:** Drive to Imnavait Creek. Imnavait Creek orientation, R4D research, point-frame sampling **Lunch:** Sack lunch from TFS at Imnavait Creek **PM:** Drive to Toolik Lake, Overview of research at TFS talk (**Donie Bret-Harte**), drive back to Galbraith Lake **Dinner and night:** Galbraith Lake	**Walker, D.A.** et al. 2014. **Chapter 3**, "Glacial history and long-term ecology in the Toolik Lake region." Pages 61-80 *in* Hobbie, J. E. and G. W. Kling (eds.) *Alaska's Changing Arctic: Ecological consequences for tundra, streams, and lakes*. Oxford University Press, New York, NY.
9-Jun	Galbraith Lake-Happy Valley	**Breakfast**: In camp. **AM:** Drive to Happy Valley. Visit poplar stand on Sag River. **Lunch:** Picnic along the drive **PM:** Catch up on readings, plant collections & projects **Dinner and night:** Happy Valley	**Hobbie, J. E. and G. W. Kling**. 2014. **Chapter 10,** "Ecological consequences of present and future changes." Pages 303-324 *in* Hobbie, J. E. and G. W. Kling (eds.) *Alaska's Changing Arctic: Ecological consequences for tundra, streams, and lakes.* Oxford University Press, New York, NY.
10-Jun	Happy Valley	**Breakfast:** In camp. **AM:** Orientation to Foothills landscapes and vegetation, Buckner sampling **Lunch:** In camp **PM:** Work on readings, plant collections and projects **Dinner and night:** Happy Valley	**Schurr, E. A. G.** et al. 2015. Climate change and the permafrost carbon feedback. *Nature* 520: 171-179.
11-Jun	Happy Valley-Sag River	**Breakfast**: In camp **AM:** Drive to Sag River camp, stops at Sagwon, gyrfalcon nest, orientation to Coastal Plain landscapes and vegetation **Lunch:** Sack lunch on Sag River by gyrfalcon nest **PM:** Work on class notes, plant collections, projects **Dinner and night:** Sag River camp	**Walker, D. A.** et al. 1998. Energy and trace-gas fluxes across a soil pH boundary in the Arctic. *Nature* 394:469-472.
12-Jun	Sag River	**Breakfast:** In camp **AM:** Hike to Percy Pingo **Lunch:** Sack lunch at pingo **PM:** Discussion of pingos and floristics of pingos **Dinner and night:** Sag River camp	**Walker, M.D.** 1987. **Chapters 1 to 2**, "Introduction" and "Background", pp. 1-34. *Vegetation and floristics of pingos, Central Arctic Coastal Plain, Alaska.* Doctoral thesis, University of Colorado, Boulder, Colorado.

Date	Location	Activity	Reading to be done in preparation for each day
13 Jun	Sag River-Prudhoe Bay-Sag River	**Breakfast**: Prudhoe Bay Hotel **All day** field trip in Prudhoe Bay oil field (**Kyla Choquette, and Tom Barrett**) **Lunch:** BP cafeteria **Dinner:** Prudhoe Bay Hotel or back in camp **Night:** Return to Sag River camp, discussion of BP tour	**Raynolds, M.** et al. 2014. Cumulative geoecological effects of 62 years of infrastructure and climate change in ice-rich permafrost landscapes, Prudhoe Bay Oilfield, Alaska. Global Change Biology doi: 10.1111/gcb.12500 **AND Streever, B.** et al. 2011. Environmental change and potential impacts: applied research priorities for Alaska's North Slope. *Arctic* 64: 390-397.
14 Jun	Sag River to Fairbanks	**Breakfast:** In Camp **AM:** Drive to Coldfoot **Lunch:** Coldfoot Truck Stop **PM:** Drive to Fairbanks. **Dinner:** Someplace in Fairbanks* **Night:** Dorms in Fairbanks	**Streever, B.** 2002. Science and emotion, on ice: the role of science on Alaska's North Slope. *Bioscience* 52: 179-184.
15-Jun	Fairbanks, Margaret Murie Bldg Room 230	**Breakfast:** On own* **8:00 am:** Unload vehicles **AM:** Prep of presentations **Lunch:** Take out pizza in classroom **PM:** Final presentations and evaluations **Dinner:** Celebration **Night:** Dorms in Fairbanks	None
16-Jun	Depart		

* Students will purchase these meals on their own

Course Library (2016)

Items not in manila folders:
Books, data reports, natural history guidebooks, guides to the Dalton Highway and floras

Argus, G. W. 2004. A guide to the identification of *Salix* (willows) in Alaska, the Yukon Territory and adjacent regions. 85 pp.

Armstrong, R. H. 1995. Guide to the Birds of Alaska. Alaska Northwest Books, Anchorage, AK. 322 pp.

Barreda, J. E., J. A. Knudson, D. A. Walker, M. K. Raynolds, A. N. Kade, and C. A. Munger. 2006. Biocomplexity of Patterned Ground Data Report. Alaska Geobotany Center, Fairbanks, AK. 224 pp.

Brodo, I. M., S. D. Sharnoff, S. Sharnoff, and S. Laurie-Bourque. 2001. Lichens of North America. Yale University Press, New Haven.

Brown J, Kreig RA. 1983. Guidebook to permafrost and related features along the Elliot and Dalton highways, Fox to Prudhoe Bay, Alaska. Fairbanks, AK: Division of Geological and Geophysical Surveys.

Cody, W. J. 2000. Flora of the Yukon Territory. NRC Research Press, Ottawa.

Collette, D. W. 2004. Willows of Interior Alaska. U.S. Fish & Wildlife Service. 111 pp.

Douglas, David C., Patricia E. Reynolds, and E. B. Rhode. 2002. Arctic Refuge coastal plain terrestrial wildlife research summaries. No. 2002-0001. US Fish and Wildlife Service, Reston VA.

Harris, J. G. and M. W. Harris. 1999. Plant Identification Terminology, an Illustrated Glossary. Spring Lake Publishing, Spring Lake UT. 197 pp.

Hasselbach, L. and P. Neitlich. 1988. A genus key to the lichens of Alaska. U.S. National Park Service, Gates of the Arctic NP&P, Fairbanks AK. 36 pp.

Hobbie, J. E. and G. W. Kling, editors. 2014. Alaska's Changing Arctic: Ecological consequences for tundra, streams, and lakes. Oxford University Press, New York, NY.

Hulten, E. 1968. Flora of Alaska and Neighboring Territories. Stanford University Press, Stanford, CA.

Huryn, A and Hobbie, J. 2013. Land of Extremes: a natural history of the Arctic North Slope of Alaska. University of Chicago Press.

Jorgenson, M.T. (ed.). 2011. Coastal Region of Northern Alaska. Guidebook to Permafrost and Related Features. Guidebook 10. State of Alaska, Department of Natural Resources, Division of Geological and Geophysical Surveys.

Marshall, R. 1991 (reprint). *Arctic Village: A 1930s Portrait of Wiseman, Alaska,* University of Alaska Press, Fairbanks,

Mull, C. G. and Adams, K. E.1985. Dalton Highway, Yukon River to Prudhoe Bay, Alaska: Bedrock geology of the eastern Koyukuk basin, central Brooks Range, and eastcentral Arctic Slope. 155 pp.

Munsell-Color. 1994. Munsell Soil Color Charts. Macbeth Div. of Kollmorgan Instr. Corp, NY.

National Geographic Society. 1987. Filed Guisde to the Birds of North America.

Sibley, D. A. 2000. The Sibley Guide to Birds. National Audubon Society. Alfred A. Knopf, New York. 154 pp.

Skinner, Q. D., S. J. Wright, R. J. Henszey, J. L. Henszey, and S. K. Wyman. 2012. A Field Guide to Alaska Grasses. Alaska Dept. of Natural Resources, Palmer, AK. 384 pp.

Streever, W. 2006. Long-term ecological monitoring in BP's North Slope oil fields. BP Exploration, Anchorage AK.

Streever, W. 2007. Long-term ecological monitoring in BP's North Slope oil fields. BP Exploration, Anchorage AK.

Streever, W. and S. Bishop. 2012. Long-term ecological monitoring in BP's North Slope oil fields. BP Exploration, Anchorage AK.

Viereck, L. A. and E. L. Little, Jr. 1994. Alaska Trees and Shrubs. University of Alaska Press, Fairbanks, Alaska.

Vitt, D. H., J. E. Marsh, and R. B. Bovey. 2007. Mosses, Lichens and Ferns of Northwest North America. Lone Pine Publisher. 296 pp.

Walker, D. A. 1985. Vegetation and environmental gradients of the Prudhoe Bay region, Alaska. US Army Cold Regions Research and Engineering Laboratory, CRREL85-14, Hanover, NH. 240 pp.

Walker DA, Auerbach NA, Nettleton TK, Gallant A, Murphy SM. 1997. Happy Valley Permanent Vegetation Plots. Boulder, CO: University of Colorado. Data Report.

Walker, D. A., H. E. Epstein, V. E. Romanovsky, C.-L. Ping, G. J. Michaelsen, R. P. Daanen, Y. Shur, R. A. Peterson, W. B. Krantz, M. K. Raynolds, W. A. Gould, G. Gonzalez, D. J. Nicolsky, C. M. Vonlanthen, A. N. Kade, H. P. Kuss, A. M. Kelley, C. A. Munger, C. T. Tarnocai, N. V. Matveeva, and F. J. A. Daniels. 2008. Arctic patterned-ground ecosystems: A synthesis of studies along a North American Arctic Transect. Journal of Geophysical Research - Biogeosciences 113:G03S01, doi:10.1029/2007JG000504.

Walker DA, Hamilton TD, Ping C-L, Daanen RP, Streever WW. 2009. Dalton Highway Field Trip Guide for the Ninth International Conference on Permafrost. Fairbanks, AK: Division of Geological and Geophysical Surveys.

Walker, D. A., M. K. Raynolds, M. Buchhorn, and J. L. Peirce. 2014. Landscape and permafrost change in the Prudhoe Bay Oilfield, Alaska. Alaska Geobotany Center, University of Alaska, Fairbanks, Alaska.

Walker MD. 1987. Vegetation and floristics of pingos, Central Arctic Coastal Plain, Alaska. Doctoral thesis, University of Colorado, Boulder, Colorado.

Items in manila folders (arranged alphabetically by author within subject folders): Journal articles and book chapters

ANIMALS

Amstrup, S. C., G. York, T. L. McDonald, R. Nielson, and K. Simac. 2004. Detecting denning polar bears with forward-looking infrared (FLIR) imagery. Bioscience 54: 337–344.

Barnes, B. 1989. Freeze avoidance in a mammal: body temperatures below 0 degree C in an Arctic hibernator. Science 244:1593–1595.

Buck, C. L., and B. M. Barnes. 1999. Annual cycle of body composition and hibernation in free-living Arctic ground squirrels. Journal of Mammalogy 80: 430–442.

Felchhelm, R. G., B. Streever, and B. J. Gallaway. 2007. The Arctic Cisco (*Coregonus autumnalis)* subsistence and commercial fisheries, Colville River, Alaska: A conceptual model. Arctic 60: 421–429.

Klein, D. R., Meldgaard, M., & Fancy, S. G. (1987). Factors determining leg length in Rangifer tarandus. Journal of Mammalogy, 68(3), 642-655.

Klein, D. R., & Bay, C. 1994. Resource partitioning by mammalian herbivores in the high Arctic. Oecologia, 97(4): 439-450.

Klein, D. R. 1995. Arctic ungulates at the northern edge of terrestrial life. Rangifer, 16(2): 51-56.

Klein, D. R. 1999. Comparative social learning among arctic herbivores: the caribou, muskox and arctic hare. Pp 126-140 In *Mammalian social learning: comparative and ecological perspectives* (HO Box and KL Gibson, eds.). Cambridge University Press, United Kingdom.

Klein, D. R. 2001. Similarity in habitat adaptations of Arctic and African ungulates: evolutionary convergence or ecological divergence? Alces, 37(2): 245-252.

Klein, D. R. and et al. 2005. Management and conservation of wildlife in a changing arctic environment. Pages 597-648 in S. J. Hassol, editor. Impacts of a Warming Arctic, Arctic Climate Impact Assessment Scientific Report. Cambridge University Press, UK.

Liebezeit, J. R., S. J. Kendall, S. Brown, C. B. Johnson, P. Martin, T. L. McDonald, D. C. Payer, C. L. Rea, B. Streever, A. M. Wildman, and S. Zack. 2009. Influence of human development and predators on nest survival of tundra birds, Arctic Coastal Plain, Alaska. Ecological Applications 19:1628–1644.

Long, R. A., T. J. Martin, and B. M. Barnes (Eds.). 2005. Body temperature and activity patterns in free-living Arctic ground squirrels. Journal of Mammalogy 86:314–322.

McDonald, T. L., W. J. Richardson, G. C. R, S. B. Blackwell, C. S. Nations, R. M. Nielson, and B. Streever. 2012. Detecting changes in the distribution of calling bowhead whales exposed to fluctuating anthropogenic sounds. Journal of Cetacean Research Management 12:91–106.

Noel, L. E., M. K. Butcher, M. A. Cronin, and B. Streever. 2006. Assessment of effects of an oil pipeline on caribou, *Rangifer tarandus*. The Canadian Field-Naturalist 120:325–330.

Owen, M., and J. Simerson. 2006. Can you hear me now? Polar bears and sensitivity to noise pollution. Conservation and Research for Endangered Species:1–2.

Sheriff, M. J., G. J. Kenagy, M. Richter, T. Lee, O. Toien, F. Kohl, C. L. Buck, and B. M. Barnes. 2011. Phenological variation in annual timing of hibernation and breeding in nearby populations of Arctic ground squirrels. Proceedings of the Royal Society B: Biological Sciences 278:2369–2375.

Sheriff, M. J., M. M. Richter, C. L. Buck, and B. M. Barnes. 2013. Changing seasonality and phenological responses of free-living male arctic ground squirrels: the importance of sex. Philosophical Transactions of the Royal Society B: Biological Sciences 368:20120480–20120480.

Stickney, A. A., T. Obritschkewitsch, R. M. Burgess. 2014. Shifts in fox den occupancy in the greater Prudhoe Bay Area, Alaska. Arctic 67: 196-202.

Williams, C. T., B. M. Barnes, and C. L. Buck. 2011. Daily body temperature rhythms persist under the midnight sun but are absent during hibernation in free-living arctic ground squirrels. Biology Letters 8:31–34.

Williams, C. T., M. J. Sheriff, J. A. Schmutz, F. Kohl, Ø. Tøien, C. L. Buck, and B. M. Barnes. 2011. Data logging of body temperatures provides precise information on phenology of reproductive events in a free-living Arctic hibernator. Journal of Comparative Physiology, B 181:1101–1109.

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Hinzman, L. D., C. J. Deal, A. D. McGuire, S. H. Mernild, I. V. Polyakov, and J. E. Walsh. 2014. Trajectory of the Arctic as an integrated system. Ecological Applications.

Raynolds, M. K., D. A. Walker, D. Verbyla, and C. A. Munger. 2013. Patterns of Change within a Tundra Landscape: 22-year Landsat NDVI Trends in an Area of the Northern Foothills of the Brooks Range, Alaska. Arctic, Antarctic, and Alpine Research 45:249–260.

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Ellis J. M., Calkin P. E. 1979. Nature and distribution of glaciers, neoglacial moraines, and rock glaciers, east-central Brooks Range, Alaska. Arctic and Alpine Research 11: 403-420.

Hamilton TD. 1986. Late Cenozoic glaciation of the Central Brooks Range. Pages 9-49 in Hamilton TD, Reed KM, Thorson RM, eds. Glaciation in Alaska: the Geologic Record. Anchorage: Alaska Geological Society.

HUMAN, INDUSTRIAL & SOCIAL SYSTEMS

Chapin, F. S., III, M. Hoel, S. R. Carpenter, J. Lubchenco, B. Walker, T. V. Callaghan, C. Folke, S. A. Levin, K.-G. Mäler, C. Nilsson, S. Barrett, F. Berkes, A.-S. Crépin, K. Danell, T. Rosswall, D. Starrett, A. Xepapadeas, and S. A. Zimov. 2006. Building resilience and adaptation to manage Arctic change. Ambio Special Report 35:198–202.

Klein, D. R. 1973. The ethics of hunting and the antihunting movement. Transactions of the Thirty-Eighth North American Wildlife and Natural Resources Conference. 18-21 March 1973. Wildlife Management Institute, Washington D.C.

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Key plant species for Dalton Highway plant communities
BIOL 495/695, Arctic Alaska Environmental Change: Field Excursion to the North Slope

Introduction:
Each day of the course will be in different ecosystems with different plant communities and dominant plant species. It is important to gain familiarity with as many of these key species as possible during the course. They provide strong information about ecosystem processes and function. The following list contains five species that are common in each of 14 ecosystems (70 species total). The instructors will pay particular attention to these species. Students should collect these & press these species for their notebooks, draw and note distinguishing characteristics, and be able to recognize these on sight. The list contains a Latin name (genus and species), the Flashcard growth-form set where details regarding distinguishing characteristics can be found, and the family name). Some common synonyms are also listed. We have organized the list by the ecosystems we will focus on from south to north along the latitudinal transect from Fairbanks to Prudhoe Bay. Distinguishing characteristics for most species are in the Flashcards, and we will be going over these during the field exercises and reviewing them periodically. For species that are not in the Flashcards (starred *), you will need to go to the plant floras for additional information. This is a basic list that everyone should strive to know by the end of the course. Interested students can, of course, learn more. Because of the later spring as we move northward, some species may not be sufficiently developed to see the distinguishing characteristics, but there may also be clues from the previous year's growth.

North Campus Lands boreal forest
White spruce forest
Picea glauca (tree, Pinaceae)
Alnus viridis (tall shrub, Betulaceae)
Linnaea borealis (prostrate dwarf shrub, Caprifoliaceae)
Cornus canadensis (prostrate dwarf shrub, Cornaceae)
Pleurozium schreberi (pleurocarpus moss, Entodontaceae)
Deciduous forest on disturbed site
Populus balsamifera (tree, Salicaceae)
Populus tremuloides (tree, Salicaceae)
Betula neoalaskana (tree, Betulaceae)
Rosa acicularis (low shrub, Rosaceae)
**Equisetum arvense/pratense* (horsetail, Equisetaceae)
Black spruce forest
Picea mariana (tree, Pinaceae)
Ledum palustre ssp. *groenlandicum* (= *Rhododendron* groenlandicum) (dwarf shrub, Ericaceae)
Vaccinium vitis-idaea (dwarf shrub, Ericaceae)
Cladina spp. (including *C. arbuscula, C. mitis, C. rangiferina*) (lichen, Cladoniaceae)
Hylocomium splendens (moss, Hylocomiaceae)
Tussock fen
Eriophorum vaginatum (sedge, Cyperaceae)
Larix laricina (tree, Pinaceae)
Betula nana (dwarf shrub, Betulaceae)
Rubus chamaemorus (dwarf shrub, Rosaceae)
Sphagnum spp. (incuding *S. warnstrofii, S. girgensohnii*)
Species around Smith Lake:
Salix alaxensis (tall shrub, Salicaceae)
Salix bebbiana (tall shrub, Salicaceae)
Comarum palustre (=Potentilla palustre) (forb, Rosaceae)
Calamagrostis canadensis (grass, Poaceae)
Shepherdia canadensis (dwarf shrub, Eleagnaceae)

Sukakpak Mountain, fen
Dasiphora fruticosa (=*Potentilla fruticosa*) (low shrub, Rosaceae)
**Andromeda polifolia* (dwarf shrub, Ericaceae)

Arctous rubra (prostrate dwarf shrub, Ericaceae)
**Trichophorum caespitosum* (sedge, Cyperaceae)
**Pinguicula vulgaris* (forb, Lentibulariaceae)

Atigun Pass, dry Alpine tundra

Dryas octopetala (prostrate dwarf shrub, Rosaceae)
**Geum glaciale (=Novosieversia glaciale)* (forb, Rosaceae)
Saxifraga tricuspidata (forb, Saxifragaceae)
Saxifraga bronchialis (forb, Saxifragaceae)
**Pedicularis lanata* (forb, Scrophulariaceae)

Atigun River, dunes and dry tundra

Elymus arenarius (grass, Poaceae)
Oxytropis maydelliana (forb, Fabaceae)
Rhododendron lapponicum (dwarf shrub, Ericaceae)
Saxifraga oppositifolia (forb, Saxifragaceae)
Thamnolia subuliformis/vermicularis (lichen, Thamnoliaceae)

Imnavait Creek and Toolik Lake, Dry acidic tundra/dry acidic snowbeds

Salix phlebophylla (prostrate dwarf shrub, Salicaceae)
Arctous alpina (prostrate dwarf shrub, Ericaceae)
Cassiope tetragona (dwarf shrub, Ericaceae)
Diapensia lapponica (prostrate dwarf shrub, Diapensiaceae)
**Hierochloë alpina* (grass, Poaceae)

Toolik Lake, moist nonacidic tundra

Carex bigelowii (sedge, Cyperaceae)
Salix arctica (prostrate dwarf shrub, Salicaceae)
Pyrola grandiflora (forb, Pyrolaceae)
Tomenthypnum nitens (moss, Brachytheciaceae)
Flavocetraria cucullata (lichen, Cetrariaceae)

Happy Valley, moist acidic tundra

Empetrum nigrum (dwarf shrub, Ericaceae)
Ledum palustre ssp. *decumbens* (*Rhododendron tomentosum* ssp. *subarcticum*) (dwarf shrub, Ericaceae)
Rubus chamaemorus (dwarf shrub, Rosaceae)
Petasites frigidus (forb, Asteraceae)
Peltigera aphthosa (lichen, Peltigeraceae)

Drive to Prudhoe Bay, Disturbed tundra along roads and pipelines, moist nonacidic tundra

Epilobium angustifolium (forb, Onograceae)
Salix lanata (=*Salix lanata* ssp. *richardsonii*) (low shrub, Salicaceae)
Lupinus arcticus (forb, Fabaceae)
Hedysarum alpinum (forb, Fabaceae)
Salix reticulata (prostrate dwarf shrub, Salicaceae)

Prudhoe Bay, Percy Pingo hike, wet nonacidic tundra

Carex aquatilis (sedge, Cyperaceae)
Arctophila fulva (aquatic grass, Poaceae)
Drepanocladus brevifolius (moss, Amblystegiaceae)
Eriophorum angustifolium (sedge, Cyperaceae)
Scorpidium scorpioides (moss, Amblystegiaceae)

Prudhoe Bay, Coastal tundra (may not see some of the saline habitat plants)

Salix rotundifolia (prostrate dwarf shrub, Salicaceae)
Dupontia fisheri (grass, Poaceae)
Carex subspathacea (sedge, Cyperaceae)
**Puccinellia phryganodes* (grass, Poaceae)
Cochlearia officinalis (forb, Brassicaceae)

Terms for Plant Identification (arranged somewhat by relevant growth forms):

Refer to Harris and Harris (1999) *Plant Identification Terminology* for complete definitions:

Trees and tall shrubs (>200 cm tall):

Alternate vs. Opposite venation or branching.
Ament or Catkin: a dense raceme of apetalous unisexual flowers (as in inflorescense of *Betulaceae* or *Salix*).
Bracts: leaf-like structures at the base of flowers, as in the woody bracts of cones of Pinaceae.
Capsule or Ovary: a dry dehiscent fruit containing numerous carpels and seeds (as in fruit of *Salix*).
Felty: dense hairiness (as in felt and the undersides of leaves of *Salix alaxensis*).
Dehiscent: fruits that split open at maturity along a suture (as in capsules of *Salix*).
Dentate: toothed leaf margin.
Dioecious (male and female flowers on different plants) vs. Monoecious: Know which families and/or species are dioecious or monoecious.
Elliptical: (as the cross section of the petiole of *Populus tremuloides*).
Fascicle: a tight bundle or cluster of needles as in some members of the Pinaceae family.
Pendulate: drooping downward (as in some catkins of *Salix* and *Betula*).
Petiole: stalk of leaf attachment to stem.
Precocious: early timing of flowering, before leaves form.
Raceme: an unbranched elongated inflorescence composed of numerous flowers with pedicels arranged along a central stalk, maturing from the bottom upwards.
Reticulate: net-like venation (as in *Salix bebbiana* or *S. reticulata*).
Samara: winged dry indehiscent fruit (as in *Betula* or *Acer*).
Serrate: saw-toothed leaf margin.
Stipules: leaf-like organs at base of true leaves (as in *Salix*).
Villous: long unmatted hairiness (as in hairs on branches of *Salix alaxensis*).

Low shrubs (40-200 cm):

Aggregate: densely clustered as in fruit of *Rubus idaeus.*
Apetalous flowers: without petals (as in Betulaceae and Salicaceae).
Comose: bearing a tuft of long white hairs as in the seeds of Salicaceae.
Connate fusion of like parts as in the petals in flowers of *Viburnum* which often form a cone-shaped structure.
Corymb: a flat-topped inflorescence as in *Spirea beauverdiana* (check out common inflorescence types in Harris and Harris, note difference in pedicel branching of corymbs, cymes and umbels).
Cuneate: wedge-shaped leaf base (as in *Betula glandulosa* and *Myrica gale*).
Glabrous: without hairs as in stems of *Salix pulchra.*
Inflorescence: the complete cluster of flowers including the axis and any bracts at base of the inflorescence.
Hip: berry-like fruit composed of ripened hypanthium that surrounds numerous achenes as in fruit of *Rosa acicularis.*

Hypanthium: cup-shaped base of flowers of Rosaceae consisting of fused corolla, calyx, and stamens.
Lanate: wooly hairiness as in leaves of *S. glauca.*
Oblanceolate: lance-shaped leaf that is broader at the tip than at the base as in *Myrica gale.*
Obovate: ovate-shaped leaf that is broader at the tip than at the base as in *Arctostaphylos uva-ursi.*
Palmate: palm shaped as in venation of Grossulariaceae and *Ribes.*
Pedicel: stalk-like attachment to the stem of flower, inflorescence, or fruit.
Perianth: in flowers, the group of corolla (petals) and calyx (sepals) considered together, or one or the other if one is absent.
Persistent: not droping in the fall as in leaves of *Salix pulchra.*
Pilose: long soft straight hairiness as in *Salix lanata.*
Pinnate: feather-like as in venation of leaves as in many species of Rosaceae (e.g., *Dasiphora fruticose, Rosa acicularis, Rubus idaeus*) or arrangement of leaflets in compound leaves (as in *Rosa*).
Scales: any thin flat structure as in the reddish-brown scales on underside of leaves of *Shepherdia canadensis.*
Scurfy pubescence: covered with scales as in Elaeagnaceae and *Shepherdia canadensis*
Simple: (not divided into leaflets) vs. Compound (leaves composed of numerous leaflets) leaves.
Stipules: leaf-like structure at the base of the petiole as in persistent stipules of *Salix lanata.*
Truncated leaf base: flat across the petiole (as in *Betula nana*).

Dwarf shrubs (<40 cm tall):
Achene a small dry indehiscent fruit with a single seed attached to the ovary at a single point.
Aggregate fruit as in *Rubus chamaemorus.*
Berry as in the fruits of *Vaccinium.*
Campanulate bell-shaped, as in flowers of numerous species of the Ericaceae family. (e.g. *Cassiope tetragona, Arctostaphylos, Vaccinium*).
Coriaceous leathery, as in the thick leathery leaves of many members of the Ericaceae family (e.g., *Arctostaphylos uva-ursi, Ledum, Rhododendron*).
Cyme flat or round-topped inflorescence as in *Cornus*. (Note: the 4 showy white petal-like bracts of *Cornus* are also part of the inflorescence.) Also note difference between a cyme and an umbel.
Crenulate wavy leaf margin as in *Dryas octopetala*.
Drupe: a fleshy indehiscent fruit with a stony indocarp that usually contains a single seed, as in fruits of *Cornus canadensis, Empetrum nigrum,* and *Arctous.*
Entire: Not toothed or divided as in the leaf margin of *Dryas integrifolia.*
Indehiscent fruits: those that do no split open at maturity to disperse seeds, as in Betulaceae family and *Dryas*.
Involute: decribes the inward rolled leaf margins as in many species of Ericaceae (e.g. *Ledum*, *Rhododendron*).
Orbicular: circular in outline as in the leaves of *Pyrola grandiflora* and *Salix rotundifolia.*

Pedicel: flower and fruit stalks that attach to the stem, Important in distinguishing *Ledum palustre* ssp. *decumbens* (pedicel strongly angled near the tip) from ssp. *groenlandicum* (smoothly arching pedicel).
Plumose: plume-like as in the feathery plumes attached to the achenes of *Dryas* that consist of persistent mature styles.
Prostrate: Lying flat on the ground, or creeping.
Pubescent: Covered in short soft hairs or bearing any kind of hairiness.
Rosette: a radiating cluster of leaves as in *Diapensia lapponica.*

Grasses:
Androecium: The male reproductive parts of a flower (includes the filaments & anthers); a collective term for all the stamens.
Awn: (dorsal vs. terminal) as in the awns of *Trisetum* (dorsal) and *Festuca* (terminal).
Caespitose: As in the growth forms of numerous tussock- or bunch-forming graminoids.
Caryopsis: The seed of a grass.
Culm: The hollow or pithy stalk or stem of a graminoid plant.
Floret: A small indifidual flower, as in the flower of a grass, within a larger group of flowers.
Glumes: (1st and 2nd): The pair of bracts at the base of a grass spikelet.
Graminoid: Grasslike
Gynoecium: Female reproductive parts of a flower (includes the ovaries, styles, & stigmas); a collective term for all the pistles.
Inflorescence: The flowering part of a plant; the arrangement of flowers on the axis of the plant. Typical inflorescences of grasses are spikes, racemes, and panicles.
Leaf sheath: The basal part of a leaf that surrounds the culm (as in the inflated leaf sheath of *Alopecurus alpinus*).
Lemma: The lower or outer of two bracts that subtend a grass floret.
Ligule: A tongue-shaped organ. In grasses and sedges, the membranous appendage where the leaf joins the leaf sheath.
Nerve: A prominent vein as in nerves on lemmas or leaves.
Palea: The uppermost bract of the pair of bracts that enclose the flower or seed of a grass floret.
Panicle: A branching racemose inflorescence, with the flowers at the base of the panicle maturing first.
Pedicel: The stalk of a single flower or fruit, or of a grass spikelet.
Pedicellate: Having a pedicel.
Raceme: An unbranched elongated inflorescence with pedicllated flowers, maturing from the bottom upward.
Racemose: Having flowers in racemes.
Rachis: The main axis of a compound structure as in a compound leaf or an inflorescence.
Rachilla: A small rachis, as in the rachilla within a grass spikelet.
Rhizome: A horizontal underground stem.
Sessile: Attached directly to the stem or culm without a petiole (as in a leaf) or pedicel as in a flower or fruit or spikelet of a grass.
Sheath: The extension of the leaf that wraps about the culm. In grasses the sheath is open.
Spike: An inflorescence of a grass or other plants with sessile spikelets of flowers.

Spikelet: A small spike or secondary spike as in the ultimate group of flowers in a grass that is subtended by a pair of glumes.
Stolon: A horizontal above-ground stem.

Sedges and rushes:
Achene: The seed of sedge, enclosed in the perigynium. Important to distinguish the cross-section shape of the achene (biconvex (two sided, derived from 2 carpels, hence flowers these have 2 stigmas per perigynium), vs. trigonous (3-sided, derived from 3 carpels, hence flowers of these have 3 stigmas per perigynium).
Beak of the perigynium: The extension of the perigynium that encloses the style of the pistle. The beak can have numerous forms including elongated, truncated, bidentate,
Bract: Refers to both the leaves at the base of a sedge spike, or the scale at the base of the perigynium.
Leaf sheath: (See Sheath under Grasses.) In *Carex* the leaf sheaths are generally completely closed. In grasses, the leaf sheaths are not completely closed; the margins overlap or are open.
Perigynium (pl. perigynia): The bottle-shaped organ (really a modified bract or leaf) that encloses the pistil and ultimately the achene of sedges.
Spike: The portion of the sedge that contains groups of male and/or female flowers. Sedge spikes can be staminate (male flowers only), pistillate (female flowers only), or contain both male and female flowers (androgynecandrous spikes have the staminate flowers above the pistillate flowers and gynecandrous spikes have the pistillate flowers above the staminate flowers).

Forbs:
Anthesis: When a flower is fully expanded and functioning.
Banner: The upper, usually largest, petal of in flowers of the Fabaceae.
Bulbils: A small bulb that is borne above the ground on the stem of viviparous (sprouting on the parent) plants, such as *Polygonum viviparum* and *Saxifraga cernua.*
Calyx: The portion of a flower consisting of the sepals.
Carpel: A seed bearing (female) reproductive organ of a flower formed from one modified leaf. The carpel is sometimes hollow forming a chamber that contains the seeds. Several carpels can be fused together to form a single pistil, and numerous seed containing chambers in a fruit. From Greek *karpos* 'fruit'. The Onagraceae have 4 carpels in each fruit, which dehisce longitudinally.
Cauline: Pertaining to the stem as in cauline leaves.
Ciliate: Having fine hairs in the margin, as in the margins of many leaves.
Compound flower: A flowering head in the Asteraceae consisting of both ray flowers and disk flowers.
Connate: Tubular or bell-shaped corolla, as in *Campanula*.
Corolla: The portion of the flower consisting of all the petals.
Corymbose: forming a corymb (see Harris and Harris, p. 158) as in the flowering head of *Achillea borealis.* A flat- or rounded-top inflorescence with the pedicels not arising from a common point – the lower pedicels being longer.

Cotyledon: The first leaf or leaves that form inside the seed and expand immediately as the seed sprouts. A monocotyledonous plant has one cotyledon; dicotyledonous plants have two.
Determinate and indeterminate inflorescences: Refers to how the group of flowers in an inflorescence develop. In determinate inflorescences the terminal flower develops first preventing further elongation of the flowering stem. In an indeterminate inflorescence, the lower flowers develop first allowing continued elongation of the main flowering axis, as in *Epilobium angustifolium* and many species of Brassicaceae.
Dehiscent: Opening at maturity to release contents, as in the capsules of Brassicaceae or other dehiscent fruits or anthers.
Disk florets: The central regular flowers (or tubular flowers) of a composite flower in the Asteraceae.
Filiform: Thread-like or finely linear.
Follicle: A dry dehiscent fruit composed of a single carpel that opens along a single side, as in many members of the Ranunculaceae (e.g., *Ranunculus, Aconitum, Caltha*).
Haustorium (pl. Haustoria): A specialized root-like organ used by parasitic plants to draw nourishment from the host plant. Examples include members of the Orobanchaceae (e.g. Boschniakia) and Santalaceae (e.g. *Geocaulon*).
Imbricate: Overlapping like shingles, as in the involucral bracts of *Solidago*.
Interpetiolar stipules: Small leaf-like appendages that occur between the bases of petioles of the opposite leaves of the family Rubiaceae.
Involucre: The whorl of bracts subtending a flower, as in the involucre of Asteraceae.
Involute: Rolled inward as in the leaf margins of *Saussurea anugustifolia*.
Keel: a prominent longitudinal ridge as in the keel of a boat, refers also the two united lower petals of flowers of the pea family (Fabaceae).
Legume: A dry dehiscent fruit of the Fabaceae. Also refers to members of the Fabaceae (formerly the Leguminosae).
3-merous flowers: Flowers of the Liliaceae that have three sepals (which may be petal-like), 3 petals, and 6 stamens.
Loment: A legume (pea pod) that has constrictions between the seeds, as in *Hedysarum*.
Mucronate: Tipped with a short sharp abrupt point, as in the leaves of *Saxigraga flagellaris.*
Pappus: bristles or hairs (which are reduced sepals) at the base of the both ray and disk flowers of Asteraceae.
Perianth: The collective term for corolla and calyx of a flower.
Phyllaries: Involucral bracts of the Asteraceae.
Ray florets: the strap-like flowers of a composite flower (or ligulate flowers).
Receptacle: The widened portion of the pedicel or peduncle on which the flower sits, as in the receptacle of Asteraceae.
Reflexed sepals: Sepals that are bent downward or backward as in those of *Saxifraga punctate* ssp. *nelsoniana.*
Reniform: Kidney-shaped, as in the leaves of *Oxyria digyna*.
Scape: A leafless elongated stem-like pedicel arising from the ground to support a terminal flower or group of flowers, usually from a basal rosette of leaves, as in *Parrya nudicaulis*, and *Papaver macounii*.

Schizocarp and mericarp: A dry indehiscent fruit that splits into separate one seeded segments (mericarps) as in some members of the Apiaceae.
Setose: Covered with bristles as in the in the leaf margins of *Saxifraga tricuspidata.*
Silique: An elongated (more than twice as long as wide) capsule of members of the Brassicaceae, which consist of two halves or valves separated by a persistent septum along which the capsule splits in half during dispersal of the seeds, as in *Descurainia*. Shorter capsules of the Brassicaceae are called silicles, as in *Cochlearia*.
Spatulate: rounded tip gradually tapering to the base, as in the leaves of *Lequerella arctica.*
Staminoidea: Infertile stamens, as in the numerous staminoidea that occur in *Parnassia* flowers.
Stipitate: Having a short stalk (or stipe) supporting a structure, as in the flowers of *Astragalus umbellatus.*
Stipules: Leaf-like appendage at base of leaf petiole. Note: in *Oxytropis* the stipules at the base of the basal leaves are membranous and brown, and the remains of which form masses of material that may not be recognized as stipules, as in the chestnut brown stipules of *Oxytropis maydelliana*.
Subulate: Sharp pointed or awn-shaped.
Succulent: Fleshy as in the leaves of *Sedum rosea* and other members of the Crassulaceae.
Twice pinnatifid: Twice-divided pinnate leaves, as in the leaves of *Artemisa frigida.*
Umbel and compound umbel**:** A flat- or rounded-top inflorescence with the pedicels arising from a common point, Compound umbels have two divisions of the pedicels as in *Heracleum lanatum*.
Wings: The two lower lateral petals on either side of the keel in flowers of Fabaceae.

Marshall, R. 1991 (reprint). *Arctic Village: A 1930s Portrait of Wiseman, Alaska.* University of Alaska Press, Fairbanks, Alaska, pp 3-44.

INTRODUCTION

BLANK spaces on maps had always fascinated me. So when I found in the spring of 1929 that I had a summer ahead in which to do whatever I desired, I took the atlas from my shelf and turned to the map of Alaska. Carefully examining it, I observed that only two really large sections were left uncharted, the one on the South Fork of the Kuskokwim River, southwest of Mount McKinley, the other at the headwaters of the Koyukuk River, north of the Arctic Circle. Mount McKinley was a great temptation, but on the whole the notion of a summer in the Arctic was even more alluring. I decided that there ought, however, to be some purpose back of this spree, so I rationalized a scientific investigation as a reason for my expedition. As a forester and plant physiologist, it seemed eminently appropriate that I should make a study of tree growth at northern timberline.

I cannot say that I learned very much either about tree growth or timberline. But I did come away with a vivid impression that the few white and Eskimo people who were scattered through this remote region were on the whole the happiest folk I had ever encountered. It is so easy, however, to found an erroneous impression on the superficial contacts of a couple of months that I decided to return for at least a year in order to make a detailed study of this civilization of the North.

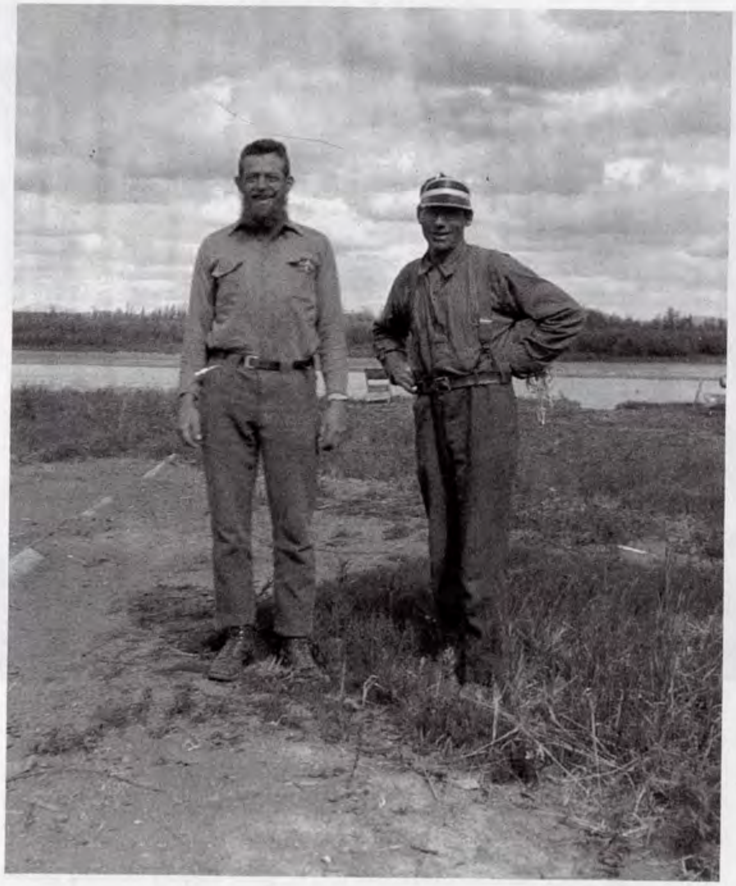
Consequently, in August, 1930, I set out for a second sojourn in Arctic Alaska. It took me about two weeks' travel by train, boat, and train again, to get from New York to Fairbanks, which, with its 2,000 people, is the metropolis of interior Alaska. At Fairbanks railroads and highways cease, so after a week of waiting for good weather in this city, I took the 200-mile airplane flight into Wiseman, the major settlement in the Koyukuk.

The welcome awaiting me when we landed would seem preposterous to any one with the conventional notions about the stolid frontiersman. Suddenly I realized to what good friends I was returning in Wiseman. The instant I stepped out of the plane, Martin Slisco, jovial roadhouse proprietor, ran up and threw both arms around my neck. Little Willie English, seven-year-old Eskimo boy, with whom I had hopping races the previous summer, was next, and he jumped all over me. Pete Dow, cynical old sourdough of thirty-two Arctic winters, nearly pumped my hand off, and his face was cracked with smiles. And following them came all the others, for every one in town, Eskimo and white, was out at the field.

The next two days were spent in Wiseman, conversing happily with friends whom I had known for a few weeks the year before, but who acted as if we were lifelong acquaintances. They were so eager to pour out the events of the past year to some one to whom they were not stale stories long ago. So I heard over and over, from a dozen different persons, each giving a slightly different version, the chief landmarks in the life of the community.

These two days were also spent preparing for a trip which Al Retzlaf, my previous year's partner, and I were taking to the sources of the North Fork River. Al was interested in gold prospecting and I in tree growth, but in spite of these mixed purposes, our partnership worked splendidly. We

THE AUTHOR'S CABIN—INSIDE.



THE AUTHOR AND ERNIE JOHNSON.

panned for gold and measured trees separately, but gloried together in nearly a month of travel up unvisited valleys, in scaling the summits of unknown mountains, in camping more than 100 miles from the closest neighbors, in discovering scenes of natural beauty fully as splendid as the world-famous Yosemite or Glacier Park.

When we got back from this trip at the end of September, Al returned to Fairbanks and I started to prepare a home for the ensuing year. I had rented from Martin Slisco a log cabin next door to the roadhouse. It was sixteen by eighteen feet, eight feet high, and had a board floor, canvas lined walls and ceiling, and a roof of split logs covered over with dirt and sods. On the south side, where it would admit the maximum sunlight, was one large window, five and one-half feet long and two feet high. Through it I could look out across the still unfrozen Koyukuk River to a range of steep, rugged mountains, all covered with snow.

The furnishings of my room included a spring bed, a double-decker wooden bunk, three chairs, a bureau with two deep shelves instead of drawers, a large, flat table for writing, located directly under the window, a cupboard for my kitchen utensils, a table for my wash basin and toilet articles, and a high, home-made cabinet for storing my medical, scientific, and photographic equipment. An iron heater in the middle of the room served the triple purpose of cooking, keeping the cabin warm, and drying damp clothing.

On top of the bureau was an analytical balance, my phonograph, and thirty precious records which I had shipped all the way from Baltimore. Near the window I had a crude bookcase made of old egg crates which contained some sixty-six volumes which I had brought with me, including works as varied as *The Magic Mountain*, *Mrs. Dalloway*, *Anna Karenina*, *Plays of Euripides*, *Erewhon*, *The Life of Sir William*

Osler, The Decline of the West, Ordeal of Civilization, Middletown, The Sexual Life of Savages, The Dance of Life, The Quest for Certainty, Social Psychology, Physics of the Air, The Universe Around Us, Minor Surgery, Medical Biometry and Statistics, and Thermodynamics. It was a splendid collection for a year of isolation, but the isolation proved so extremely social that I did not have time to read a quarter of the books.

This sociability commenced the first day I moved into my house, while I was still unpacking my belongings. Three veteran white inhabitants of the Koyukuk and one Eskimo dropped into my cabin, ostensibly to listen to my new records, but actually, I am sure, because they were afraid I might be a little homesick. Thereafter, there was scant opportunity for homesickness. Scarcely a day went by that at least five or six different people did not visit me. During the course of my year in the Koyukuk, every person around Wiseman except one three-months-old baby, and one Eskimo woman who could not speak English, spent at least several hours alone with me in my cabin.

Intimacy with these friendly people was also enhanced by many visits to their own homes. In addition to constant association with those who lived right in Wiseman, I paid about a dozen visits to each of the two nearby gold mining centers on Nolan Creek and Hammond River. Here the miners would invariably welcome me with glowing hospitality, banter with me in unfailing good humor, provide a comfortable bed for as many nights as they could induce me to stop with them, cook delicious meals for me, allow me to work with them in their mines, lend their magazines to me, and borrow my books in return.

Sometimes I took longer trips. Toward the end of October I indulged in my first dog mushing on a four-day journey

with Bobbie Jones, when the sky was sparklingly clear and the snow freshly white. The Stanich brothers and Tom Kovich, with whom we stopped along the way, were brimming with eager cordiality. A few weeks later Jesse Allen, Kenneth Harvey, and I went out on a ten-day expedition to haul sheep meat from the head of the Middle Fork. Later we were joined by Albert Ness, and the four of us spent vigorous days in the unique fellowship which comes from enduring mutual hardships, and snug nights in snow surrounded camps where we conversed on the widest variety of subjects. Then in March Ernie Johnson and I occupied a month in exploring the course of an unknown river, and returning to Wiseman by way of the most remote mining camps in the region, where we were welcomed as if we were long-lost friends. Finally, for fifty days during July and August Ernie and I were mapping the source streams of the Alatna and John River drainages, and meeting now and then, scores of miles from the closest human being, some solitary prospector.

But it is not the purpose of this introduction to go into the details of these different expeditions which I made by mushing, boating, and back packing. It is enough to state that a third of my 452 days in the Koyukuk were spent on such journeys. Sometimes they would lead me to isolated cabins, miles from the closest neighbors, where under the stimulation of unanticipated companionship, lonely men would speak to me of yearnings and strivings, of secret satisfactions, of long-buried tragedies, of splendid ideals, and of constant gropings through many years of contemplation for the meaning and significance of life.

However, the major share of my fifteen months in the Koyukuk was passed in the town of Wiseman. My typical day started around seven, when I arose, during much of the time before daylight, started my fire, crawled back into bed while the cabin warmed, and then breakfasted with Martin Slisco. After breakfast, while we waited for the dawn, he would regale me with the latest details of his contemporary love affair, or we would play the phonograph to drive away the last vestiges of drowsiness. I would spend most of the mornings in my cabin, reading, writing, fixing up my notes, and talking with visitors. After cooking my lunch I might spend the afternoon visiting around town, working at home, or going for a walk. I recall a number of evenings when I walked out along the trail, while the far below freezing weather made my nose tingle. The southern sky would be brilliant with sunset colors, the snow all around would change from a strange purple to a dark gray, and diminutive Wiseman when I returned would be twinkling with lights. Then I would repair to the roadhouse, which was at the same time a shelter from the trail and a social center for the community. Supper would always be a loquacious meal, with those who were eating there, and those who merely came to chat, talking back and forth without intermission. After supper I would sometimes remain at the roadhouse until bedtime, either listening to the conversation or dancing, sometimes receive visitors and play the phonograph in my own home, but most often drop around from one cabin or igloo to the other, talking with different friends until far into the night. Then I would walk home through the freezing air, while the northern lights rolled brightly across the heavens, and feel that life could not possibly be more splendid. While I was undressing I might play the *Hungarian Rhapsody*, the *Gymnopedie*, or perhaps Schubert's *Unfinished Symphony*. The last record for the evening I always put on just before turning out the gasoline lantern, and then I listened to it comfortably from bed. When the final note was over and the automatic stop had clicked, it generally took me about thirty seconds to fall asleep.

In the long days of spring and early summer, when there was no darkness at all, we often chatted or danced until two or three in the morning without noticing the time. I remember one night when Kaaruk and Oscar and I set out at midnight to tramp the hills for seven hours, picking cranberries, shooting ducks, stopping now and then to catch grayling at some deep hole in the river, watching the sun rise at one in the morning, and joking and laughing continually. I recall another night when Jennie Suckik and I climbed to see the midnight sun from Smith Creek Dome, and I shall probably never forget the strange light which saturated the vast expanse of wilderness. About this time Jesse Allen, Kenneth Harvey, and I set out on another trip, traveling by the cool daylight of night, and sleeping without blankets in the warmth of mid-day, and feeling that the whole world was a trifle upside down.

Those were days which made life seem a constant romance and adventure. To-day, when I am back in the more prosaic cities of the Atlantic seaboard, it sometimes seems difficult to realize that such events were ever real. But almost every month I receive letters from my Arctic friends, some from the Eskimo girls running to thirty pages, so I have sound evidence that my fifteen months in the Koyukuk was more than a glorious dream.

However, it is not my desire to write about my own adventures, but rather to describe in an objective manner the unusual civilization in which these adventures were set. For I think it would interest many readers to learn something about the independent, exciting, and friendly life of the Arctic frontier. Consequently, I am writing this book with the purpose of painting a complete picture of the civilization of whites and Eskimos which flourishes in the upper reaches of the Koyukuk, 200 miles beyond the edge of the Twentieth Century.

I

GEOGRAPHY

IN THE vast domain beyond the Yukon is a distinctive civilization which spreads across the entire drainage of the Koyukuk River lying north of the Arctic Circle, an area embracing approximately 15,000 square miles. While this territory seems rather small when blocked out on the map of Alaska, in comparison with more familiar places it takes on a greater significance. Belgium contains in all less than 12,000 square miles. New Jersey has but 7,500 and Massachusetts barely 8,000 square miles. It would take Massachusetts and New Jersey combined to equal the area of the upper Koyukuk.

But there is this striking difference between the places mentioned and the Arctic Koyukuk. Belgium has a population of 8,060,189 people. New Jersey has some 4,041,334 citizens. Massachusetts shelters 4,249,614 people, while the two states together have a population of 8,290,948. The upper Koyukuk has a total population of 127.

Stated in terms of density of population, it is interesting to compare this region with more familiar places:

Region	*Population per Square Mile*
Upper Koyukuk	0.0085
Alaska	0.101
Nevada (least densely populated state)	0.8

Region	*Population per Square Mile*
United States	41.3
New York	264.2
Rhode Island (most densely populated state)	644.3
Belgium	686.0
England	734.2
Manhattan Island	84,113.2

In other words, Alaska as a whole is some twelve times as densely populated as its remote segment, the upper Koyukuk. The least densely populated state in the Union has about one hundred times as many people to the square mile. The United States as a whole has five thousand times, England has nearly 100 thousand times, and Manhattan Island has ten million times as concentrated a population as the civilization of the Arctic Koyukuk.

The 127 people who make up this Arctic civilization include seventy-seven whites, forty-four Eskimos, and six Indians. The following table analyzes this distribution in a little greater detail:

Race	*Men*	*Women*	*Children*	*Total*
White	70	7	0	77
Eskimo	9	11	24	44
Indian	1	4	1	6
Total	80	22	25	127

It may seem that in taking an arbitrary fraction of the drainage basin of a single river as a unit of civilization, an artificial and illogical boundary has been set. Actually, however, the Arctic drainage of the Koyukuk represents a remarkably unified culture, and the break with an entirely different cul-

ture occurs just about at the Arctic Circle. North of the Circle is a civilization composed of a mixture of white men and Eskimos with their central occupation gold mining and their major language English. South of the Circle is a civilization composed almost entirely of Indians who have felt very little the influence of white culture, where the central occupations are still hunting, trapping, and fishing, and the major language is still Indian.

Practically on the Arctic Circle is the Episcopal Mission of St. John in the Wilderness. Around this focus on the north side of the Koyukuk is the Eskimo village of Alatna, and on the south side of the river the Indian village of Allakaket. Both villages show markedly the influence of the white man, much more so than do the Indians lower down the river, but they still remain closer to the down river than the up river culture.

The main Koyukuk River and the lower reaches of its five chief tributaries (the Alatna, South Fork, John, North Fork, and Middle Fork Rivers) flow through a flat, swampy country with no conspicuous relief. But in their upper reaches these streams cut through one of the most rugged terrains imaginable, with precipices rising sheer for hundreds and even thousands of feet, with deep, glacial canyons as sensational as Yosemite, and with great rock mountains jutting almost straight up from the valleys. This back country is virtually unknown even to the inhabitants of the Koyukuk. A few have been into it on hunting or prospecting expeditions, but there are still many areas which have never been visited by man.

The entire population of the Arctic Koyukuk focuses around two towns. Before going further it is necessary to explain that the word town as used in this region has a very different connotation than in the outside world. One can liken this difference to the diversity between the new and the old conception of the atom. The old conception was simply of one solid, substantial mass. But the new conception pictures the atom as consisting of a nucleus and everything in space which it affects. Similarly, the ordinary town consists of a solid, consecutive mass of houses occupying a limited position on the map. But the Koyukuk town consists of the nucleus of the store and everything in space which it affects; in other words, everybody throughout the entire region who comes regularly to the store to trade.

Actually only sixteen of the seventy-seven white people of the upper Koyukuk live permanently around the centers of trade. The majority, some forty-six people in all, live in mining communities of from two to eighteen people at Nolan Creek, Hammond River, Jim Pup, Emma Creek, the Porcupine, and Wild Lake. There are eight people who live close to some community (that is, from one to ten miles away), where they can easily get into contact with human beings, but where they are too remote for neighbors to hear their shouts or observe that no smoke rises from their chimneys in case of some disaster. One of these people, taken violently sick last summer, lay for three days alone in his cabin, too feeble to travel the necessary mile for help. But there are seven men who live in far more genuine isolation than this man. Their homes are from twenty to sixty miles from the closest humankind. The Eskimos almost always live in groups, residing around the towns chiefly in the winter and camping among the caribou hills or along the rivers in summer. It is only their hunters who live in isolation for periods of a few days or at most a few weeks.

Both of the upper Koyukuk towns are located on the main artery of the region, the Koyukuk River. Bettles is the first one reached on ascending this stream. It is thirty-eight miles by trail and ninety miles by the sweeping bends of the river

BETTLES

WISEMAN: The large building in the foreground is the roadhouse, and the second cabin beyond is the author's home.

UPSTREAM FROM WISEMAN.

DOWNSTREAM FROM WISEMAN: The town is at the left margin of the photograph and the airplane field in the center.

from Alatna at the Arctic Circle. It is located at the confluence of the John River and the main Koyukuk, and merely consists of a dozen occupied houses on a willow flat about eight feet above the river. Some twenty-four of the Koyukuk inhabitants do their trading at the Bettles store and make its roadhouse the social center of their lives.

The term roadhouse too has a different connotation in Alaska than in the outside world. It is not primarily a place for wild parties and intoxication, though these are surely not barred, but is chiefly a haven of shelter along the dogteam road. Every old sourdough[1] in Alaska has recollections of plowing with his dogs through some blizzard, half frozen, half exhausted, wondering whether he would ever reach the end of his dark trail, when into the night would suddenly shine the cheering lights of habitation. He would drive his weary team up to a large log cabin, overjoyed in the assurance at last of a night of warmth and safety. Later, with his dogs tied up, he would sit by the glowing heater, removing his damp moccasins, rubbing his hands together, thawing out his whole chilled body, while the ever ready proprietor would be busy in his kitchen, no matter what the hour, preparing a sumptuous supper. Then while eating he would listen to the gossip of the country, give in exchange the news of the people from whom he had just departed, and feel the glow of friendship so distinctive of the northern hospitality.

Eighty-five miles up the Koyukuk from Bettles, upon a low flat where Wiseman Creek flows into the Middle Fork from the west, is the metropolis of Wiseman. Here some 103 people come to do their trading, of whom as many as eighty-one actually were in town at one time during the

[1] A sourdough is a veteran of the North country. The term is sometimes used to include any one who has witnessed the Yukon and its tributaries freeze up in the autumn and break up in the spring.

Christmas festivities. Wiseman boasts forty-eight occupied houses, located chiefly along three streets which run parallel to the river and two more avenues which extend back from the Middle Fork. However, when I speak of avenues please do not imagine paved boulevards. In summertime a Wiseman street is represented by a brown streak through the surrounding weeds and willows. In winter it is marked by a hard groove made by the passing sleds and walkers in the soft snow. In spring it is merely a part of town even a little muddier than the rest. At most places it is bounded by fences and cabins, except for the outside of the front street which is the bank of the river. This riverside road is the longest street in town, being about half a mile between its two extreme cabins.

It is not, however, the minor dimensions of town which are most impressive in Wiseman. It is instead the huge expanse which separates this lonely settlement from the outposts of Twentieth Century civilization. It is 200 miles airline to the closest pavement, the closest auto, the closest railroad, or the closest electric light at Fairbanks. The nearest hospital or doctor is 150 miles to the southwest at Tanana. Even steamboat navigation ends eighty-five miles down the river at Bettles, while the closest church is ninety miles further still at Allakaket. For many things which a person desires he must send 3,500 miles to Seattle. Such great distances give the Koyukuk an inaccessibility reminiscent of the Nineteenth Century frontier of the West, and an isolation which lies beyond the conception of most people in the closely populated regions of Twentieth Century mechanization.

II

CLIMATE

THERE are few civilizations set in such a varied yearly climatic cycle as is the Koyukuk. The difference between the short, sunless, snow-filled days of December, and the verdant, twenty-four-hour days of June and July, is almost the difference between two worlds. All the economic activities of the people, all the social habits, even the psychological reactions are revolutionized by the passage of the seasons.

Viewed from a scientific standpoint, the basic force behind all these differences is the fact that the earth's axis is tilted at an angle of 23° 27′ away from a perpendicular to its orbit. This tilt, as we all learned in fourth grade geography, means that in summer the northern regions of the earth are pointed toward the sun, while in winter they are faced so much away that even at mid-day the sun in many places cannot clear the horizon. The Arctic Circle is an imaginary line the same number of degrees from the North Pole as the earth's axis is tilted. As one travels north from the Circle there is a constant increase in the number of sunless days of winter as well as the days of midnight sun in summer, unless local topography happens to alter theoretical conditions.

Of course, in any mountainous country local topography does. Thus at Wiseman, which is only one degree north of the Circle, there are thirty-one consecutive days, from December 7 through January 6, when the sun can never be seen. This is because the 3,000 foot mountains which wall in the valley of the Koyukuk hide the sun even when it rises several degrees above the horizon. At the other extreme, although in theory one should be able to see the midnight sun for a couple of weeks in summer, the high mountains block it out entirely from Wiseman, and one must climb several hundred feet up some surrounding hillside before it becomes directly visible. But right in town one can always see some color in at least the northern sky for four and one-half consecutive months, from April 15 through August 28.

The amount of light in Wiseman on clear days at different periods of the year may best be indicated by the following chart:

Condition of Light	*Dec. 21*	*Feb. 21*	*Apr. 21*	*June 21*
First color in sky	8.00 A.M.	5.50 A.M.	....	....
First daylight	9.10 A.M.	6.40 A.M.	2.20 A.M.	....
Sunrise	....	10.40 A.M.	4.50 A.M.	1.25 A.M.
Sunset	....	3.15 P.M.	7.35 P.M.	9.30 P.M.
Last daylight	2.40 P.M.	5.20 P.M.	9.40 P.M.	....
Last color in sky	3.50 P.M.	6.10 P.M.	....	....

I have defined daylight as that amount of light necessary for trees to appear three dimensional and colored, in contrast to darkness when they are merely flat, black objects. In the shortest days of winter I could only read comfortably, without artificial light, for an hour and a half at midday, even when seated next to my south-facing window. On cloudy days at this time I had to burn my light all day long. At the schoolhouse the gasoline lantern is continually lighted during December and January. At the other end of the year it is unnecessary to use artificial light even at midnight from May 1 to about August 10.

THE CANYON OF THE KOYUKUK: Between Bettles and Wiseman.

The River Crossing at Wiseman—Winter.

The River Crossing at Wiseman—Summer.

Of course the winter is featured by cold as well as darkness. The coldest thoroughly authentic temperature was one of — 70° F. recorded at the Weather Bureau Station at Allakaket, just at the south edge of the region. In December, 1901, a standard Green thermometer, hung outside the old roadhouse at the mouth of Minnie Creek, registered — 72° F. There have been scores of days when the thermometer passed 60° below. In the winter of 1929-1930 this happened every day for two consecutive weeks. On the other hand, there have been relatively mild winters. One of these, according to all accounts, was the one which I spent in the Koyukuk, when the coldest temperature was merely minus fifty.

In summer the weather becomes as warm as in many mountainous parts of the United States. On July 29, 1923, the temperature in the shade rose as high as 90° F. at the Weather Bureau Station at Allakaket. It had been —69° F. on January 26 of the same year, making a six month span of 159 degrees Fahrenheit. In general any temperature over 70° F. is considered very hot weather.

The seasons in the Koyukuk are of considerably different definition and extension than those of the temperate zones. Winter, the dominant season, I would define as that time of the year when the ground is continuously covered with snow and when a person in traveling must be constantly prepared for below zero weather. Summer is the time when the rivers are open for navigation, the ground is free from snow except on the higher mountains, the leaves are out on the hardwood trees, and the mosquitos, and later the gnats, make life thoroughly miserable for any one not prepared to cope with them. Spring is a brief transition period between the two, when the snow is melting, the rivers breaking up, the ground exceptionally muddy, and travel an extremely difficult matter. Autumn is another brief transition period when the rivers are freezing and the ground is alternately bare and covered with snow. It would not be possible to set hard and fast limits to these seasons, but in a general way they have the following approximate duration at Wiseman:

Winter	October 10 to May 10	7 months
Spring	May 10 to June 10	1 month
Summer	June 10 to September 10	3 months
Autumn	September 10 to October 10	1 month

If you come to Wiseman for the first time during the summer months there is little to suggest the Arctic. You see green hills rising all around you to end in rocky summits three thousand feet above the valley, yet not a trace of snow on any of them. There is dark green timber in all of the valleys and well up on the south-facing mountain slopes. The muddy flats are covered with clumps of sedges, and many delicate flowers are in bloom from the valley floor to the highest summit. The turbulent Koyukuk boils along, now tumbling over shallow riffles, now racing squarely into cut banks, now flowing in orderly procession for a brief distance down some spruce-girded lane. The bright gravel bars which fill its valley are cut into innumerable abandoned channels where it has flowed along in former years, and where it may start to flow again at any moment. Almost everything at this season could find its counterpart in any wild temperate region. It is only when one walks about in the full daylight of midnight that there is a sense of the exotic, and the weird feeling of mystery that one anticipates at the ends of the earth.

There will come a night, sometime about the middle of August, when the thermometer will drop well below freezing, and in the morning when you rise to look outside your cabin you will notice the whole valley tinged with yellow.

From then on summer will keep building toward autumn. First the birches, then the cottonwoods, then the willows become bright golden. The hillside herbs turn red and purple and yellow. The muddy flats are frozen every morning, and every evening the winds down valley from the unexplored north bring promise, though summer has not yet departed, that winter is almost at hand. By the end of August there is a brief period each night when darkness settles down, and on clear nights one sees the first mild flashing of the aurora.

Early in September the birch leaves come down in great numbers, and soon after the cottonwood commences to lose its foliage. The rivers grow very shallow, and one knows that their source streams in the high Arctic Divide have frozen. On cold mornings the river is filled with slush ice which may keep flowing for several hours after the sun strikes the surface of the water. The ground becomes frozen even in midday, making walking delightfully easy after all the mud of summer. Some cloudy day the falling rain will commence to turn to snow, and in a few hours everything will be white. But it is still only autumn, and the sun is sufficiently high that a few hours of shining will melt all the snow in the valley. However, on the mountaintops the bare ground will not show up again for nine or ten months. Some time early in October there will be a little heavier snowstorm than before, and then one morning you will wake up to find that the ground is buried to remain so, and that winter is really at hand.

Shortly after the snow has come to stay the main river freezes solidly. The days grow rapidly shorter, so that where there were ten hours of sunlight at the autumnal equinox in September, two months later only an hour and a half is left. On the sixth of December half of the sun skids along the horizon for ten minutes, and then disappears, not to return until the seventh of January. During the sunless days, if it is clear weather, there is a continuous sunset effect in the sky to the south from the first break of dawn to the last twilight. I remember vividly, one clear, cold Sunday afternoon, walking back toward Wiseman from Hammond River. About three o'clock, just at dusk, I came to a place where the trail swung out on a high point overlooking the Middle Fork, and there below me was a barren plain of snow stretching half a mile through the twilight to the cold, black forest. Beyond that was the most gorgeous sunset you can imagine, a whole sky warm and glowing, and everything so quiet and beautiful that I wondered what commensurate value the outside world could possibly provide.

There are many crystal clear nights, when the whole sky is covered with surging waves of the aurora, bands of green and blue and purple and white which go shooting and rolling and twisting across the heavens in a display of brightness which is never the same for two consecutive seconds, which sometimes almost fades into oblivion, at others bursts in full vividness over the entire expanse of the sky, and even seems to dip down below the mountaintops and race along the surface of the earth. The full moon at this time of year is so bright that you can read fairly fine print by its light, and each star is a clear, sharp, twinkling beacon, almost a solid, substantial mass in its certain outlines. And as a frame for everything, above and below, are the cold, moonlit mountains rising whitely against the sky.

Early in January there is great excitement. The sun has returned again. The first time it shines in town old and young alike are exuberant. I walk down the street and an old sourdough, cutting wood in front of his cabin, shouts and points to the sun. At another cabin a man who has spent thirty-four

winters in the North is all beaming. He laughs and quotes these lines from "Paradise Lost":

> *"Oh thou that, with surpassing glory crowned,*
> *Look'st from thy sole dominion like the god*
> *Of this new World—at whose sight all the stars*
> *Hide their diminished heads—to thee I call."*

I drop in at an Eskimo cabin where three old women are seated on the floor sewing moose hide and gossiping. They point excitedly to the wall where the sunlight is shining, and one of them chants an Eskimo nonsense rhyme about the sun. Another says in broken English: "Me feel pretty young today. Old fellow come back again." Then they all laugh. I go out and pass by an igloo. A five-year-old Eskimo girl who can hardly speak English runs up to me.

"See that down there?" she says, pointing to sun, and then she jumps up and down in excitement.

At first the increase in sunlight is slow. It is almost two weeks before the sun shows up for even an hour. But around the beginning of February the days do noticeably lengthen, and by March they stretch so rapidly that one can readily observe that each day is longer than the one before. On the first of April there are six more hours of sunlight in town than on the first of March.

During these months of increasing daylight Wiseman shows its most typical appearance. The roofs of the houses, the fences, and the summer gardens which they surround are buried under a deep layer of powdery snow. Only the streets and the entries to the cabins are well packed down by the constant passage of sleds and walkers. By day a blue smoke rises from each occupied cabin, at night the light gleams out from the yellow windows in cheery warmth.

Viewed on clear days, the setting of Wiseman is inspiring until one becomes so accustomed to it as to forget the fact that men grow up and die without ever beholding even a small fraction of the beauty which constantly surrounds this village. As far as one can see, up and down the Koyukuk, the flat valley floor is flanked by pure white mountainsides, jutting into rocky pinnacles which catch the sunlight hours before the lowlands are out of shadow. Behind the town a series of terraces ascend to the base of sparkling summits which rise in opposition to the deep blue sky. Across the Koyukuk is a cut bank, projecting about fifty feet upwards. It is pleasant to climb to its top late in the afternoon and watch the evening and night come down on the valley. At first, backed by the bright, setting sun, Wiseman is an imposing-looking village, with its well-spaced group of houses arranged in orderly rows, up and down the river. But as the sun dips behind the mountains, and the chill evening winds arise, the town seems to shrivel and the all-abounding wilderness grows larger and more impressive. Dusk keeps on deepening, a single light shines from a cabin window, and before long all of Wiseman is aglow. Its brightness, however, seems very trivial in the infinite extension of the uninhabited mountains. As darkness descends, the wilderness keeps on expanding, until when night has fallen the sparkle of town seems only the tiniest oasis of warmth and comfort, almost lost in the all-pervading desolation and freezing and mystery of the Arctic.

But even an Arctic winter has an end. By the middle of April there is no more total darkness. The snow commences to melt on the roofs under the influence of the high sun. By the end of the month, here and there in the town, out on the river bar, along the lower south-facing hillsides, bare spots of ground appear. Early in May the niggerheads are already in blossom, though most of the ground is still buried under snow.

Before the middle of May there is so little snow left in the flats that sledding becomes impossible. Then some day, without any warning, you perceive that the ice is gone from the river and the water is running free. Day by day the snow continues to disappear until the flats are all bare, and the mountainsides show a spectacular speckling of white and green. Only the north slopes still remain a pure white. But not for long. Under the constant vigor of the sunlight, which is already shining more directly than it ever does in temperate zones, the snow is taken away so rapidly that you can almost see it disappearing, hour by hour. The side streams now are brim full of water, and little trickles which you can cross in summer without wetting the soles of your shoes are absolutely unfordable. A week or two after the streams around Wiseman are at the full, the ones up valley reach their peak. Then the flood comes down the river, covering gravel bars and little islands and everything from rim to rim of the valley floor, a wild, surging, absolutely uncontrollable torrent, sweeping along willows and large logs and even entire, uprooted trees, which go bobbing up and down as they ride the crest of the inundation.

Meanwhile the flowering plants are at the height of their glory. The delicate white blossoms of the Arctic anemone are sprayed along the warmer banks as early as the middle of May. They are shortly followed by the golden dandelions and buttercups, the purple violets, the goldthread, and the angowuk. Early in June the spruce forests are carpeted with the large, white blossoms of the Dryas, most widely distributed of all the Arctic plants. Throughout the month this eight-petaled flower makes the woods glorious. It's place of prominence is followed by the golden California poppy and the Arctic sage and later still by a myriad of blue forget-me-nots.

But long before the forget-me-nots are in bloom the insects come out so fiercely that most men must wear a head net and gloves constantly when they are working, to protect themselves from the mosquitoes. These are taken philosophically and jocularly. "Only about two more months of them," says Wes Ethrington, the day after they appeared, "and then a month of gnats, and then the snow again."

Sometime around the middle of June the first hot spell sets in. The temperature rises above 70° and feels more like 95° with the direct solar radiation. Then at last your mind, which has not quite been able to keep up with the rapid changes, drops the last thought of winter, and suddenly you realize that summer has come again.

III

HISTORY

FOR AT least a hundred years before the advent of the white race the Arctic Koyukuk was largely a no-man's land. The settlements of the Indians extended only up the main river for a short distance above the mouth of the South Fork. The Eskimos did not live in the region at all, but came across from the Kobuk and Noatak Rivers and over the low passes from the Arctic, to hunt and fish in the watersheds of the Alatna and John Rivers. But, in spite of these minor invasions, the greater portion of this 15,000-square-mile territory was as untraveled as it had been when it was buried under the last ice sheet.

The first white men to enter the Arctic Koyukuk were Lieutenant Henry T. Allen and Private Fred Fickett of the U. S. Army. Allen, later nationally famous as the commander of the Ninetieth Division of the United States Army in the World War, and internationally admired for his humane and tactful administration of the duties of commander of the American Army of Occupation in Germany after the War, had been commissioned to make an exploration of the almost unknown interior of Alaska. During the course of a 2,200 mile wilderness journey he reached in August 1885 a place about five miles above the mouth of the John River. Upon his return to the United States he made a remarkably accurate map of the Koyukuk River which the best previous sketches had made to head about 400 miles off course. This initial exploration of Allen's was augmented the following March by a journey of Lieutenant Stoney of the Navy across the drainage of the Alatna River. As a result of these two expeditions the outside world received its first knowledge of the upper Koyukuk.

During the subsequent years a few adventuresome prospectors commenced to search for gold in the remote sources of this river. John Bremner and Peter Johnson were the first to brave its isolation, coming there in 1887 and several years following. The dangers were genuine, but the isolation not quite sufficient, for Bremner encountered some Indians on the Hog River who murdered him in 1891. In that same year Johnnie Folger made his first journey to the Koyukuk, and for four years with a series of different partners he mined the placers on the South Fork, Chapman Creek, and Tramway Bar. The first gold in paying quantities was discovered at the latter place in 1893. In all there were probably eighteen or twenty different prospectors who visited the upper Koyukuk between 1887 and 1897.

Then came 1898.

From all over the world 80,000 people went stampeding to the North country, certain of a fortune in exchange for a few weeks, a few months, or a few years of adventure, depending on the degree of their optimism. Many of them knew nothing about the outdoors, most of them knew nothing about gold mining, and all of them knew nothing about the requirements for existence in the North. Possessing various degrees of stamina and adaptability to totally new conditions, some died, most gave up, a fraction remained and founded the present civilization of Alaska.

The great focus for the stampede of '98 was Dawson and

the Klondike. However, the good claims had already been staked in 1897. All one could do was to buy one of these for a huge sum of money, work for wages, prospect on worthless creeks, or move to some other part. A great many chose the latter course. About 1,000 people from this Dawson overflow boated to the Koyukuk in the late summer of 1898. Some 200 of these found their way north of the Arctic Circle, and staked out a great swath of country on the South Fork, Middle Fork, and Alatna drainages. Caught short by the early freezeup, they had to spend the winter on these bleak and frigid rivers, a pathetic band, isolated, incompetent, horribly homesick for the outside world from intercourse with which they were entirely severed. One can still see the crumbling remains of the cabins they built for themselves in clusters which they optimistically christened Arctic City, Beaver, Rapid City, Union City, Peavey, Seaforth, Soo City, Jimtown. One can still imagine the smoke rising from the chimneys of these long-forgotten settlements, while lonely men clustered around stoves, wondering if winter would ever end, wondering if they could outlast scurvy or starvation, wondering almost if there really were other people anywhere in the world. By a miracle only two out of the two hundred men marooned in the Arctic were frozen to death, one drowned, and one died of a fatty heart. The others, as soon as the river had broken in the spring, pulled stakes and dropped down to the Yukon, leaving as their only permanent record the names of their wives and sweethearts on most of the creeks flowing into the Koyukuk for a hundred miles north of the Arctic Circle.

Following their departure a hardier and more permanent group of miners entered the country. Four of them are still resident there to-day. The first real money was struck by Knute Ellingson on Myrtle Creek that summer. The first store was opened at the mouth of John River that autumn by

Gordon Bettles, after whom the settlement which grew up around it was named. The prospectors were still, however, leading a nomadic life, and they were still very much cut off from the normal comforts of even frontier existence. Probably the best picture of those nomadic days can be given by quoting the reminiscences of an old German named Carl Frank who still resides in the Koyukuk.

In those days we didn't have no time for prospecting anywhere. We was just running around wild, stampeding. It was in February, I remember, I was living just as comfortable as can be in a little cabin in Bettles, big stove, big woodpile, plenty of blankets, what more could I want? Only I was a little low in grub. Then some fool—we had awful liars in those days—we have them now—he whisper about something up North Fork, so we all wanted to get a claim of course there too. I got tangled up with Argo Bill as a partner. I had a hell of a time with him. He was a tough fellow. He always want to shoot the gun. He growl: "You do what I say or I blow out your brains." I don't understand much English in them days, but I understand that. Plummer, he come along with us. I didn't bother about getting snowshoes, I think I can just walk along the trail, it will be good and hard. So we set out along the trail, pulling and pushing on our sled. There wasn't hardly any dogs in the country in them days, and it cost you nearly a whole winter's outfit to buy one.

We met John Tobin, oh, what a liar he was. He whisper, something awfully rich on Alder Creek. There is no trail to Alder Creek broken out, just the deep snow. Plummer has snowshoes, Argo Bill has snowshoes. I want to get in on the rich creek, so I make snowshoe out of a young spruce tree and some board. I have to practice first. Of course you can slip along, slip along, but the thing gets so awfully sidling, and it's very slow. Bill and Plummer start out first, and so soon as I am ready I start along, tracking after them. I walk

and I walk, but I can't see anything of them. I came to a creek and I see where they blaze for me to write my name on, to stake out my own claim. But it get darker and darker and I still don't see anything of them. They climb right up over a mountain and I have awfully hard time with my boards. Pretty soon I come to a place where I can't see their trail no more, and then I really am lost. The snow gets hard and I take off my boards and walk on the snow. Then pretty soon it gets soft again and I have left my boards behind. I slip down under the trees and all over, and the snow gets down my neck and the snow gets up my sleeves and the snow gets inside my pants. Then I come to a creek and pretty soon I see a trail. Oh, my God, I feel good. I saw this must be Alder Creek. I don't know which way to follow trail, but I think maybe I go downstream. In a little while I see light through the trees, and then I see a fire, and pretty soon I find Plummer and Argo all set up comfortable there in the snow for the night. We all talk, but the worst was I couldn't make myself understood what I had done, I speak such awful English, and they just laugh at me. I say: "If I could only explain myself," and then they laugh some more.

After we get finished up with the Alder Creek business I go down to the mouth of the North Fork to rest up. I had hardly any food left, and I sit alone in my tent wondering what I should do. All at once I hear some one walking very quick up to the tent. I go outside and there is Tom Dowd. He looks as if he is pretty nearly crazy. He says: "Oh, an awful thing has happened. Andy Seaman shot himself." Andy was his partner, a nice young chap, about 27 years old, only he was so awful religious.

I say: "Where is he and how did it happen?"

Tom says: "He's on the North Fork, near Florence Creek. He was pulling his sleigh in front, and he had his shotgun lying on it all cocked, and the trigger must have caught on some willow, and now he's dying."

We went right out to Florence Creek, and find the poor fellow. He was suffering terribly. He was wounded all through the bowels, and the pain and the worry drive him crazy. His mind was in a terrible way. He kept shouting: "God forgive me, God forgive me, oh, my God, my God, forgive me, forgive me, oh, forgive me," all day long. What he wanted Him to forgive I don't know, but it made me feel bad. I think it was awful. If religion is any good for any one it should be to ease your dying moments. But his religion put him through the worst agony I ever saw a man go through.

"My God," I thought, "if religion is only that the last hour of the poor religious man is made so miserable, my God, I say, the hell with religion."

After three terrible days like this he died. We hauled him on a sled to Pope Creek, just across the Koyukuk from the mouth of the North Fork. There was a cabin there, and we put his body inside. Then we start to dig a hole, but the ground is so hard and frozen it take us two days just to dig a place big enough for his body. We finished it the evening of the second day, but we didn't bury him that night. We cooked our beans over a fire in the snow. The bacon was rotten. After a while I go in the cabin to get something. There was no window in it, no light at all. I open the door and all at once there was a rumpus there, a r-r-r-r. "I'm not afraid," I think, but I am though. You know how a man feels, you have a dead man there. I don't believe in a ghost but I think a little there was a ghost anyway. I stand perfectly still for a few seconds, and then all at once something rush by my leg, and out into the snow, and I nearly fall over. But it was no ghost, only a marten that had gotten inside some way.

The next day there was the funeral. We hauled him out of the cabin—it was 60° below zero that morning and the sun was blood red. Dowd was an Irishman, born in America, a nice, fine, slow spoken gentleman.

"Carl," he said, "I'm not a good talker, but I'll pray the Lord's prayer anyway."

Then we let him down. I was not religious, I never was,

but it comes so funny to a man when you remember back how we put the fellow down, there was no one around, only us two ragged stiffs kneeling there at the grave, and the sun so big, and the air so cold.

Then, when we had that fellow buried, that Tom Dowd, he want to go prospecting then. He say: "What we do now, Carl?" I say: "We go to Florence Creek, that is Get-Rich-Creek, but—" I say: "I have no grub." Dowd say: "That's all right, I have no grub either, we get some from Jack Hood."

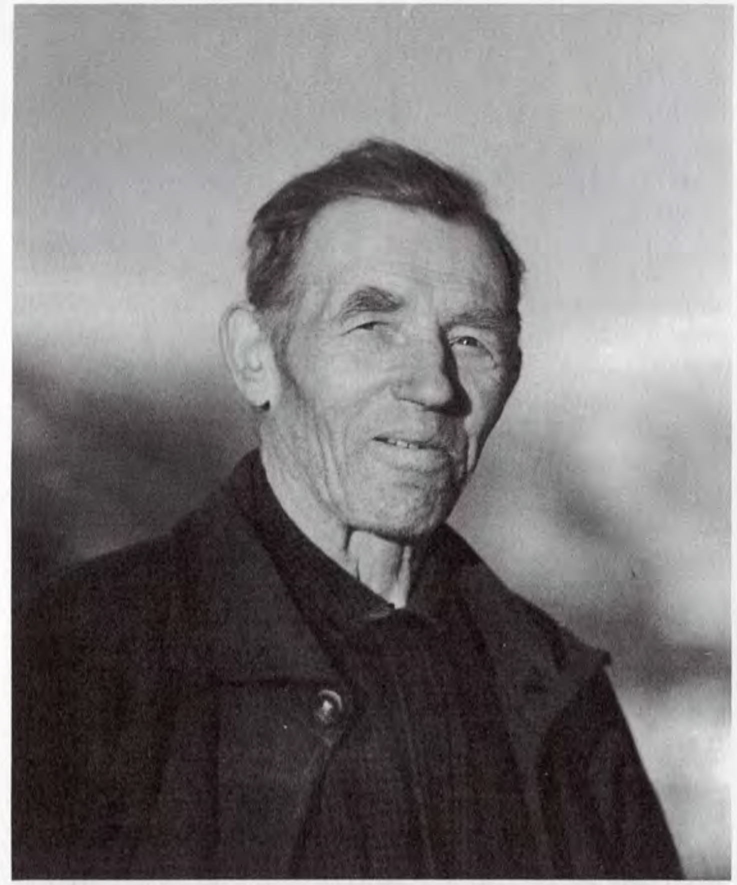
We go down to Bettles and talk to Hood, and he agree to grubstake us. But he was awful much in love with the natives down in Bettles in those days. He is too busy to haul grub for us, but we haul a little grub ourselves to Florence Creek, just enough to last a couple of weeks, till Hood gets up with the team. But Hood does not come. My partner goes out to get ptarmigan and he fall and sprain his ankle. I have a partner down at Bettles, and a partner in the hospital, and the grub thirty miles away. It was an awful time. Every day I go out to get ptarmigan, but I do not know how to do it very well, and we get pretty empty. A whole week more pass. Still Hood did not arrive. I don't know what we will do. I don't think people ever starve to death any more, but still I think, how can you live without any food? Finally Hood came just before the snow melt.

Then we start to prospect like the deuce. But we don't find any gold. We prospect four weeks, but still we don't find anything. Finally we think it's no use prospecting this way any more. So we come back to river to go up to Twelve-mile Creek. But meantime the river had broken up, and there was so much water everywhere we had to wait in the Canyon five or six days for the water to go down. We decide to cache our stuff here while we go up the river so we won't have so much junk to haul. I had a good feather bed and my pillows I brought from 'Frisco and lots of good clothes. Tom had his lots of clothes too. We made a big cache in a swampy place, and left everything there.

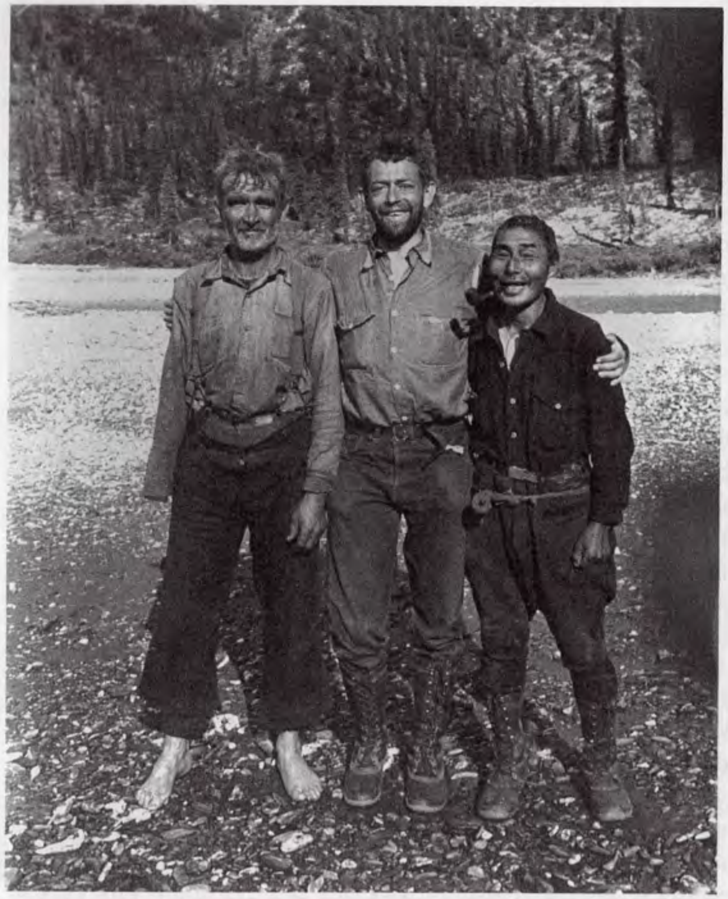
Then we set out for Twelvemile. Tom set out by river, but I said, no I will walk across country, it will be shorter and easier. They all draw us maps and tell us just where it is, back of big mountain range. But something go wrong. There was a mountain range all right, but it was the wrong mountains. I got lost for three days. It was the worst thing I ever had happen to me. I wander and wander and wander, and hardly sleep all that time, and I think now I really am a goner. Then I see some hills, way off in the distance, which look like they draw on the map. I went up to those hills and see far over there a creek. I went to the creek and it was running the wrong way. I say to myself: "My God, this creek is wrong." Then I don't know what to do. I wander all over some more, and finally I find myself on the river. Then I think I will not leave this any more. I keep follow it up all around big bends and points, but I'm scared to leave. Finally on the third day, I come to mouth of big creek, and there is Tom, all comfortable in camp.

We work all summer long. In the fall we come down from Twelvemile Creek, no gold after all. But we were going to get good clothes and go down to see town at Bettles. We got down to our cache and we couldn't find it. It had been a very dry summer, and the swamp had dried up, and the whole country was on fire, and our cache had all burned up. We couldn't find anything but a few burned blue dishes. That was all right, we couldn't help it. We go down to Bettles, as flat broke as could be.

Then the South Fork fellows come down. They have pretty good blowout. We drink that squareface gin. I didn't know that stuff. It look like water but it did not taste like water. They drag several fellows home on sleighs. By and by my partner fall on floor, off counter. The fellows say, "Let him lay there, he'll sleep good there." But just when



KNUTE ELLINGSON:
He struck the first real money in the Koyukuk in 1899.



JESSE ALLEN, AUTHOR, AND NUTIRWIK AT CANYON CREEK CAMP.

I leave, Pat Judge, he say, "Come, Carl, let's bring your partner home." So we took him home, he's nice and loose and limber, and I think he will be fine in the morning. Then he was sleeping to beat the deuce. After an hour I came away, he was still sleeping good.

Sunday morning I look out of my cabin. The sun was shining so nicely on the new snow, and I stretch my arms over my head, and think this is a great world. All at once I see Pat Judge run like the dickens. I run out, and shout, "What's the matter?" He say: "Tow Dowd is dead."

We bury him then. Israel make a speech. He mean it good, but it sound pretty funny. He says: "We're living here in a cold country, and if we take a drink we mean it well. It wasn't your business, God, to get sore at poor old Tom." Then we all start to sing a song, *Nearer, my God, to Thee*, or what was it they sing.

The history of the upper Koyukuk since 1898 may be told succinctly in the three parallel columns of the following table. The white population, the gold production, and the prostitutes all reached approximately simultaneous peaks in two different periods. The first occurred from 1900 to 1903 when successive stampedes were on to numerous creeks, and nearly $800,000 in "sunburned" gold was recovered from the shallow gravels of their valleys. The second was from 1908 through 1916 when the deeply buried bedrock on Nolan Creek and Hammond River was being mined.

Year	*Permanent White Population*	*Gold Production in Thousand Dollars*	*Prostitutes*
1898	200	...	0
1899	120	...	0
1900	270	107	2
1901	320	173	6

Year	*Permanent White Population*	*Gold Production in Thousand Dollars*	*Prostitutes*
1902	350	200	10
1903	300	301	7
1904	210	200	0
1905	220	165	1
1906	160	165	0
1907	120	100	0
1908	240	220	6
1909	230	420	5
1910	190	160	8
1911	160	130	5
1912	230	216	9
1913	250	368	8
1914	270	260	13
1915	300	290	14
1916	250	320	12
1917	200	250	7
1918	150	150	2
1919	130	110	2
1920	119	90	0
1921	107	78	0
1922	101	132	0
1923	97	37	0
1924	92	54	0
1925	88	50	0
1926	93	68	0
1927	98	78	0
1928	90	46	0
1929	83	32	0
1930	77	31	0
1931	71	27	0
Total	...	5,028	...

The first streams to yield gold in large quantities were Myrtle Creek, Emma Creek, and Gold Creek, all discovered in 1899 and 1900. They were located between sixty and eighty miles from the store at Bettles, and consequently it became necessary for a trading center to spring up a great deal closer to the hub of mining activity. As early as 1899 the town of Slate Creek was started at the mouth of the creek which bears that name. In the summer of 1900 one of the waves of green stampeders got as far up the Koyukuk as this point, then got cold feet, turned around, and departed. This incident was enough to change the first, unromantic appellation of the settlement to Coldfoot. The real boom in Coldfoot did not come until the next year when both the Northern Commercial Company, the great trading organization for all Alaska, and William Plummer opened stores. At the same time other essential concomitants of a frontier civilization had commenced to function, so that when the peak was reached in 1902 Coldfoot boasted one gambling hole, two road-houses, two stores, seven saloons, and ten prostitutes. But there were no churches, and indeed to this day no house of worship has ever appeared north of Allakaket.

The diggings on Myrtle Creek and Emma Creek were only a few miles from Coldfoot, so the men who were mining there paid frequent visits to town in quest of hilarity. Twice a year, in the spring and fall, there was a general reunion when the men came down from the more distant creeks to haul back their summer's or winter's supplies. In the normal celebrating which ensued some of the fellows became pretty wild. Fred Swift, one of the worst dare-devils, tried one night to enter the house of Lydia, fairest of all the prostitutes. Lydia was having a brief spasm of living steadily with another miner at the time, and she did not want to let Swift in. She locked the door, and Swift tried to break it down. He employed such energy that she became frightened and blazed away with her revolver, right through the door. The shot missed its mark, and Swift said in an ordinary, conversational voice: "A little more to the right, Lydia." Then he jumped to the left. Lydia fired according to his new direction and Swift said calmly: "A little more to the left, Lydia," and jumped to the right. After repeating this several times he tired of it, and built a bonfire on her front porch. Finally, in frantic desperation at the thought of losing her house, Lydia allowed him to enter.

In the spring of 1905 there was a large crowd of impatient miners waiting in Bettles for the first grub boats, that they might get provisions to outfit for the big stampede which was going up John River that season. The first boat brought no food but considerable dynamite. The exasperated miners broke into some of the boxes and took out the sticks. To these they attached cap and fuse, then ignited the latter, and chased each other around town with the blazing sticks of dynamite. Just before they exploded they would toss them over the bank into the river. Miraculously nobody was hurt, but there was a noise like a battle all through the night.

Nevertheless, in spite of such riotous antics, it was never a camp of desperados. "It was too hard for the really tough guys to come so far from comforts," one of the veterans of those early days once told me. Argo Bill had the reputation of being the bad man of the Koyukuk. He would bluster, and aim his gun at people, and growl that if they wouldn't do whatever he might desire he would blow out their brains. Finally a mild-mannered Austrian called his bluff, and subsequently Argo Bill became as docile as any one in camp. After the first few years nobody carried guns any more, except to go hunting.

It is truly remarkable that in this period there was almost

no trouble between the white people and the natives. It must have required an amazing amount of tolerance and consideration to bring about that happy state of affairs. The first contact had not been so propitious, for the whites had been very careless with their fires as well as their sexual activities, and the Indians had grounds to be incensed. The murder of John Bremner may be attributed to such early friction. The Eskimos, who commenced to migrate to the Koyukuk from the North at the same time that the big stampede of 1898 populated the country from the South, never had had any difficulty at all. There is not even a single instance, in the thirty-four year history of contact between these two races, of any serious physical injury being administered by a member of the one race to a member of the other.

Although there was still a great amount of gold being taken out of the ground after 1903, the number of people in the Koyukuk commenced to decline. This was largely because the cost of living was so high. In 1905 some exaggerated rumors of a bonanza on the John River brought probably 400 transient prospectors to the Koyukuk, of whom several dozen remained, but aside from this brief boom the population diminished steadily. In 1906 there was a big stampede to the adjacent Chandalar country, and this made such a marked depletion in the ranks of the Koyukuk that within two years its population was cut in half.

In the autumn of 1907 three Swedes, John and Louis Olson and John Anderson, were given a strip of ground 300 feet wide on Nolan Creek. They were supposed to prospect this ground and find if there was any gold deep down below the surface. If there was gold in paying quantities on this narrow sample then it would be worth the owners' while to mine the adjacent claims. Actually there was almost unbelievable fortune. The three Swedes took out $100,000 the first winter. In three years they recovered over a quarter of a million dollars from their narrow strip. No piece of ground in the whole country ever yielded so richly. The news of fortune, even with the slow communication of those days, speedily traveled all over Alaska. By the spring of 1908 there were over two hundred new men rushing into the Koyukuk, and half of them stayed. The lonesome valley of Nolan Creek throbbed with activity. Fifteen or twenty different outfits were sinking holes, and a dozen boilers chugged away, day and night. Responding promptly to the stimulus of gold, half a dozen prostitutes had arrived before the end of the summer. A new boom was on in earnest.

More gold was recovered from Nolan Creek and its tributaries in four years than had been taken from the entire Koyukuk in all the years before that. Then, just as the riches of Nolan Creek commenced to wane, Verne Watts finally located the deep channel of Hammond River in the spring of 1911. During the next five years over a million dollars came out of this valley. Food, clothing, machinery, and whisky were unloaded for both of these diggings at the site of Wright's old roadhouse at the mouth of Wiseman Creek (commemorating a transient prospector who stopped a few minutes to pan its gravels and perpetuate his own name). A new town first called Wrights, then Nolan, finally Wiseman sprang up at this point. Meanwhile Coldfoot lost ground steadily until twenty years later there were only mice and ptarmigan to hear the winds go howling down the valley of the Koyukuk.

The peak year in the Koyukuk's second boom period was reached in 1915 when over 400 tons of freight were brought in for the 300 whites and perhaps seventy-five natives living in the region. Sixty tons of booze alone came into the Koyukuk that year. Since this could not legally be sold to the natives, it made 400 pounds (including bottles and packing) for every white man, woman, and child. In some years there was as much whisky as food brought in. In fact, whisky had the priority over everything else, and "the trail never got so bad they couldn't haul whisky up here, no matter how scarce the food might be."

Some of the veterans of those days estimate that at least half of the money taken out of the ground went for booze and prostitutes. Two miners, flush with their winter's clean-up, spent \$1,500 in one night on champagne. John Bowman once squandered between \$10,000 and \$11,000 in less than two weeks around Wiseman, most of it on a single prostitute. Half a dozen different men must have lavished at least \$25,000 on the prostitutes. Of course if a person had only his sexual intercourse and a little to drink, he would not have to pay very much. There were standard prices. It cost \$20 to spend the night with a prostitute, sport, hooker, floozie, or chippy (the words were used interchangeably), or \$5 for a single act of copulation. But the men were easily conquered by feminine wiles. One fellow "gave a hooker \$2,500 to get an education, and she knew too much for him already." A favorite trick for a prostitute was to get angry at the man who was fond of her, and refuse to have anything to do with him. If the man was sufficiently enamored he might give her a present of a thousand dollars to buy back her affections. It was far from the ends of the world, and women were few.

After 1916 three things happened. The richest claims both on Nolan Creek and Hammond River were mined. The high wages of the World War period attracted many of the most energetic men to the Outside. Prohibition went into effect, and the freely flowing whisky which had been to many such an important feature of the life in the Koyukuk was over. Consequently, population and gold production both declined almost uninterruptedly. The last prostitute left in 1919, and none has ever returned for more than a few weeks. To-day there do not remain a quarter of the white people who made the gaudy civilization which brightened the Koyukuk in the booming years when gold was coming from the ground in hundreds of thousands of dollars and fortune seemed directly within everybody's grasp.

PART II

THE PEOPLE

CRREL Permafrost Tunnel
Fox, Alaska

A Guidebook

prepared by the:

Institute of Northern Engineering
University of Alaska Fairbanks
P.O. Box 755910
Fairbanks, Alaska 99775-5910

LATE-PLEISTOCENE SYNGENETIC PERMAFROST IN THE CRREL PERMAFROST TUNNEL, FOX, ALASKA

M. Z. Kanevskiy, H. M. French, and Y. L. Shur (eds.)

CONTRIBUTING AUTHORS:
Bjella, K.L.[1], Bray, M.T.[2], Collins, C.M.[1], Douglas, T.A.[1], Fortier, D.[2], French, H.M.[3], Kanevskiy, M.Z.[2], Shur, Y.L.[2]

[1]CRREL, [2]University of Alaska Fairbanks, [3]University of Ottawa, Canada

ABSTRACT

Late-Pleistocene syngenetic permafrost exposed in the walls and ceiling of the CRREL permafrost tunnel consists of ice- and organic-rich silty sediments penetrated by ice wedges. Evidence of long-continued syngenetic freezing under cold-climate conditions includes the dominance of lenticular and micro-lenticular cryostructures throughout the walls, ice veins and wedges at many levels, the presence of undecomposed rootlets, and organic-rich layers that reflect the former positions of the ground surface. Fluvio-thermal modifications are indicated by bodies of thermokarst-cave ('pool') ice, by soil and ice pseudomorphs, and by reticulate-chaotic cryostructures associated with freezing of saturated sediments trapped in underground channels.

INTRODUCTION

The CRREL permafrost tunnel is located at Fox, approximately 16 km north of Fairbanks, Alaska. Constructed 40 years ago, it is one of the few underground exposures of syngenetic Pleistocene-age permafrost. Naturally-occurring exposures of ice-rich permafrost quickly degrade and provide only opportunistic study. The permafrost tunnel allows hundreds of visitors the unhurried opportunity to become acquainted with ice-rich permafrost, and for professionals to study the peculiarities of syngenetic permafrost and its history.

This guide summarizes recent cryostratigraphic observations made from within the tunnel and re-evaluates earlier interpretations. Some observations have been described in previous publications (e.g. Shur et al. 2004, Bray et al. 2006) while others are presented in the NICOP proceedings volume (e.g. Bray 2008, Fortier et al. 2008, Kanevskiy et al. 2008).

THE CRREL TUNNEL

The CRREL permafrost tunnel was constructed in the early 1960's by the U.S. Army Cold Regions Research and Engineering Laboratory (CRREL) in order to test mining, tunneling, and construction techniques in permafrost. It continues to be maintained by CRREL for research opportunity. The plan of the tunnel is shown in figure 1. The tunnel entrance is located on the eastern margin of Goldstream Creek Valley where a steep 10 m high escarpment had been created by placer gold mining activities in the first part of the previous century. The surface of the valley that lies immediately above the long axis of the tunnel rises gently from the top of the escarpment in which the entrance is located towards the east side of Goldstream Creek Valley. The active layer of the terrain that overlies the tunnel is between 0.7 and 1.0 m thick. This is typical of the Fairbanks area.

The tunnel is composed of two portions (see fig. 1). The adit (a nearly horizontal passage from the surface into the hillside) was driven by the U.S. Army Corps of Engineers using continuous mining methods in the winters of 1963-64, 1964-65, and 1965-66 (Sellmann 1967). The winze (an inclined adit) was driven by the U.S. Bureau of Mines (USBM) from 1968 to 1969 using drill and blast, thermal relaxation, and hydraulic relaxation methods (Chester & Frank 1969). The adit extends approximately 110 m in length and is predominantly located in the frozen silt unit. The winze begins approximately 30 m into the adit and drops obliquely at an incline of 14% for 45 m, passing into the frozen gravel unit and ultimately into the weathered bedrock, where a Gravel Room was excavated (Pettibone & Waddell 1973). At the time of excavation, portions of the Gravel Room roof consisted of

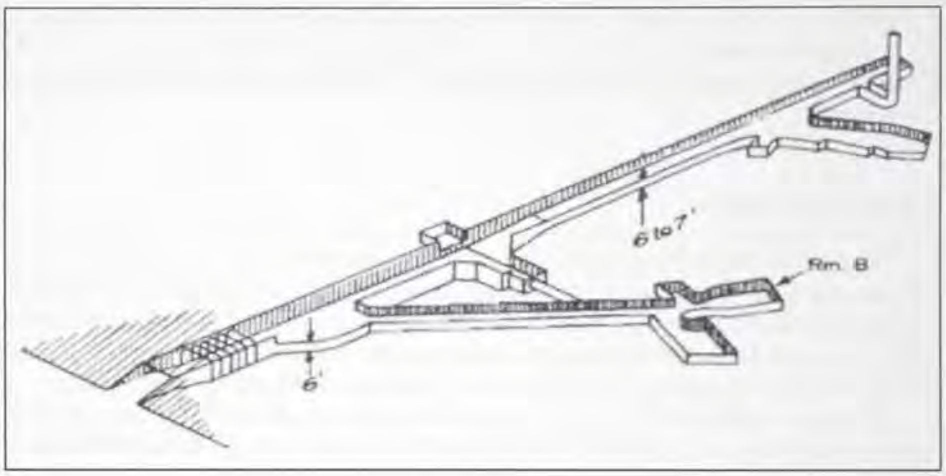
2.0 m of frozen gravel lying below the overlying silt unit. After the winze levels out adjacent to the Gravel Room, it continues for another 10 m into what is known as the CRREL Room. The tunnel is chilled by natural ventilation in winter and by artificial refrigeration in summer, supporting permafrost stability.

Many papers have been published on the geology, paleoecology, and cryostratigraphy of the sediments exposed in the tunnel (Sellmann 1967, 1972, Watanabe 1969, Hamilton et al. 1988, Long and Péwé 1996, Shur et al. 2004, Pikuta et al. 2005, Bray et al. 2006, Katayama et al. 2007, Wooler et al. 2007, Fortier et al. 2008, Kanevskiy et al. 2008) as well as their engineering properties (Chester & Frank 1969, Pettibone & Waddell 1971, 1973, Thompson & Sayles 1972, Johansen et al. 1981, Johansen & Ryer 1982, Garbeil 1983, Weerdenburg & Morgenstern 1983, Arcone & Delaney 1984, Delaney & Arcone 1984, Huang et al. 1986, Delaney 1987, Bray 2008). The problems of tunnel construction have also been described (Chester & Frank 1969, Dick 1970, Swinzow 1970, Linnell & Lobacz 1978).

Sediments exposed within the tunnel consist mainly of frozen silts that range in age from Wisconsin to Recent (fig. 2). They are eolian (i.e. wind-blown) in nature and are largely derived from the outwash gravels and braided stream deposits of the Tanana lowlands and surrounding hills. Ice-rich silts of eolian origin were also partly reworked and re-transported by slope and fluvial processes (Péwé 1975, Hamilton et al. 1988).The silts overlie fluvial gravels of Nebraskan age (Fox Gravel) that are derived from the sur¬rounding hills of the Yukon–Tanana Uplands. These gravels over¬lie Pre-Cambrian schist bedrock.

The silt overburden at the thickest point is approximately 14 m over the adit and 18 m over the Gravel Room.

Figure 1. Isometric view of the tunnel.



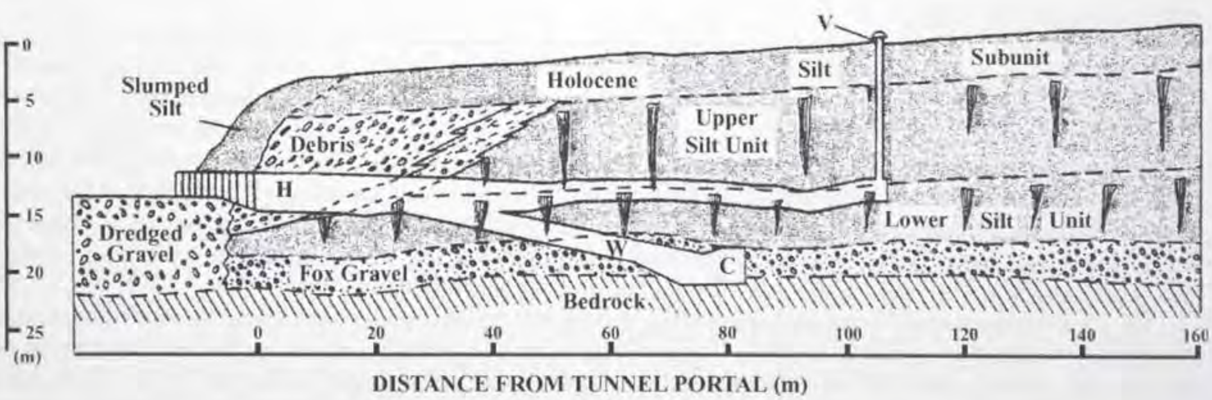


Figure 2. Cross section of the geology and permafrost features exposed in the CRREL tunnel near Fairbanks, Alaska. Two units of ice-rich Pleistocene-age silt are shown to be separated from two sets of ice wedges by an unconformity (dashed line). H—horizontal tunnel (adit); V—ventilation shaft (does not exist today); W—winze; C—chamber. (From Hamilton et al. 1988).

Near the tunnel portal, the fanlike deposits of poorly-sorted debris overlie the silts in an unconformable fashion; they formed between 12,500 and 11,000 years BP during deep erosion of the Goldstream Creek Valley slopes (Hamilton et al. 1988).

PREVIOUS OBSERVATIONS ON THE TUNNEL PERMAFROST

Sellmann (1967, 1972) was the first to provide information on the tunnel geology and permafrost. He described segregation ice, foliated wedge ice, and large clear masses of ice (buried 'aufeis'). He also identified two 'unconformities'. The upper was marked by the tops of small wedges and a change in soil color. The lower unconformity was identified by (i) a change in size and shape of wedges, (ii) a gap in radiocarbon ages obtained from organic material contained within the silty sediments in the tunnel walls of between 14,000 and 30,000 years BP, and (iii) a 20-fold increase in chemical concentration with depth. Sellmann suggested '…this unconformity was probably caused by some regional warming or local depositional or erosional event'.

Subsequently, Hamilton et al. (1988) obtained samples from the tunnel walls and reported upon 33 radiocarbon dates. The dates for silts are within the range 30,000 – 43,000 years BP. In addition, a diverse assemblage of animals' bones (bison, horse, mammoth (?), caribou (?) and arctic ground squirrel) and plant macrofossils (grasses and sedges) indicated a tundra or shrub-tundra environment. Hamilton et al. (1988) concluded that the tunnel '... provides continuous and undisturbed exposures of ice-rich silt that overlies gravel and bedrock' and that '…most of the pore and segregated ice formed during freezing of silt'. They identified pore ice, segregated ice, foliated wedge ice and buried surface ice, as previously documented by Péwé (1975) and concluded that 'most of the pore ice and segregated ice formed during freezing of silt and has been preserved since that time'. They also identified two independent systems of ice wedges and inferred a thaw unconformity between them (fig. 2). Other bodies of ice, described as '...horizontal, saucer-shaped bodies, 2–6 m wide and 0.5–2 m deep' (Hamilton et al. 1988) were interpreted as buried frozen thaw ponds formed in ice-wedge troughs. According to these authors, these ice bodies '…generally consist of 3 successive depth zones: (1) clear ice with vertical bubble trains, transitional downward into (2) ice containing reddish brown, suspended organic matter that overlies (3) a lenticular body of unusually ice-rich silt'.

CRYOSTRATIGRAPHY AND CRYOLITHOLOGY

Cryostratigraphy refers to the study of frozen layers in the Earth's crust. It is a branch of geocryology. It was developed first in Russia where the study of ground ice gained early attention (Shumskii 1959, Katasonov 1962, 1969) and subsequently led to highly detailed studies (e.g. Vtyurin 1964, Popov 1967, 1973, Gasanov 1969, Gravis 1969, Zhestkova 1982, Shur 1988, Romanovskii 1993, Dubikov 2002) that are unparalleled in North America. A summary of cryostratigraphic principles can be found in French (2007, 153-185).

Cryostratigraphy differs from traditional stratigraphy by specifically recognizing that perennially-frozen sediment and rock contain structures that are different to those found in unfrozen sediment and rock. Cryolithology is a related branch of geocryology and refers to the relationship between the lithological characteristics of rocks and their ground ice amounts and distribution. The structures, largely determined by the amount and distribution of ice within sediments are termed 'cryostructures'. Cryostructures are useful in determining the nature of the freezing process and the conditions under which frozen sediment accumulates.

A distinction must be made between epigenetic and syngenetic permafrost (fig. 3). Epigenetic permafrost refers to permafrost that forms subsequent to deposition of the host sediment and rock. By contrast, syngenetic permafrost refers to permafrost that forms at the same time as the host sediment is being laid down. These types of permafrost can be distinguished by analysis of cryostructures. The epigenetic-syngenetic distinction is extremely useful in the context of Quaternary paleo-environmental reconstruction.

Cryostratigraphy adopts many of the principles of modern sedimentology. For example, 'cryofacies' are defined according to volumetric ice content and ice-crystal size, and then subdivided according to cryostructure. Finally where a number of cryofacies form a distinctive cryostratigraphic unit, these are termed a 'cryofacies assemblage' (French 2007).

(i) Cryostructures

Russian permafrost scientists were the first to systematically identify cryotextures and cryostructures (Gasanov 1963, Katasonov 1969, Zhestkova 1982, Popov et al. 1985, Melnikov & Spesivtsev 2000). Unfortunately, these classifications are complex and unwieldy. For example, Katasonov's (1969) classification lists 18 different cryostructures and Popov et al's (1985) classification lists 14 categories.

A simplified North American cryostructural classification by Murton & French (1994) encompasses the range of cryostructures found in permafrost (fig. 4). Several Russian terms are transliterated. All cryostructures can be recognized by the naked eye.

The common cryostructures are:

(1) 'structureless' (Sl) - refers to frozen sediment in which ice is not visible and consequently lacks a cryostructure. (This category is termed 'massive' in the Russian transliterated literature).
(2) 'lenticular' (Le) - lens-like ice bodies that are described by inclination, thickness, length, shape and relationship to adjacent cryostructures.
(3) 'layered' (La) – continuous bands of ice , sediment or a combination of both.
(4) 'regular reticulate' (R) - a regular three-dimensional net-like structure of ice veins surrounding a mud-rich block
(5) 'irregular reticulate' (Ri) – an irregular three-dimensional net-like structure of ice veins surrounding a mud-rich block.
(6) 'crustal' (Cr) - refers to the ice crust or rim around a rock clast.
(7) 'suspended' (Su) - refers to grains, aggregates and rock clasts suspended in ice. (This category is termed 'ataxitic' in the Russian transliterated literature).

Figure 5 shows examples of cryostructures typical for the tunnel.

The micro-morphology of cryostructures can be observed using an environmental scanning electron microscope (ESEM). For example, Bray et al (2006) provide examples of structureless (i.e. 'massive') and lenticular cryostructures viewed conventionally and under ESEM (fig. 6).

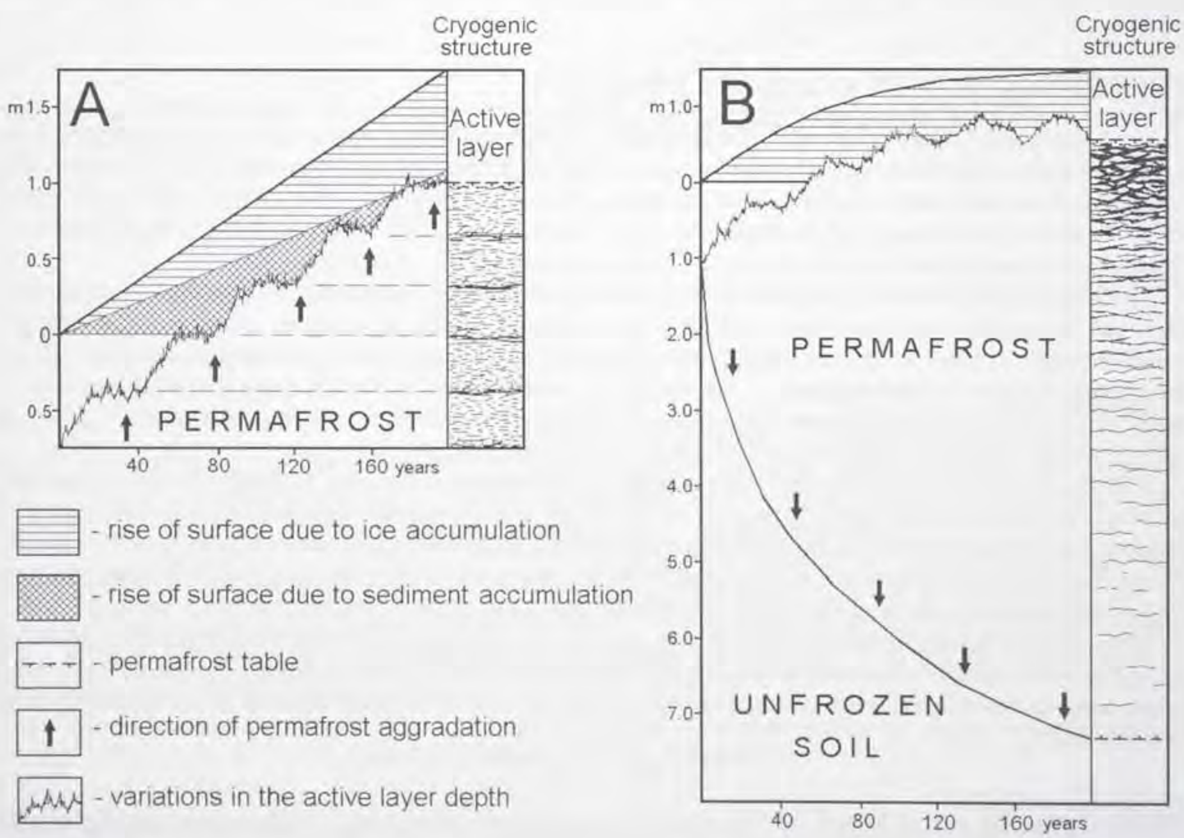


Figure 3. Mechanisms of (A) syngenetic (based upon Popov, 1967) and (B) epigenetic permafrost formation.

(ii) Thaw unconformities

Discontinuities in the nature and distribution of ground ice bodies related to permafrost thawing are termed 'thaw unconformities' (French 2007). They are the result of either thawing of frozen materials (primary thaw unconformity) or subsequent refreezing of previously-thawed material (secondary thaw unconformity). A primary thaw unconformity forms at depth below a residual thaw layer. In doing so, it truncates the top of an ice wedge. When permafrost subsequently aggrades, the original thaw unconformity at depth becomes a secondary (i.e. palaeo-) thaw unconformity and the new active layer-permafrost boundary becomes the primary thaw unconformity. The secondary thaw unconformity can be recognized by both the truncated ice wedge and by different cryostructures in the sediment above and below. Thaw unconformities can be further recognized by differences in stable isotope values, by heavy mineral assemblages above and below the unconformity, and by horizons of enhanced micro-organisms.

The manner in which permafrost degrades and subsequently forms again, and the cryostratigraphic evidence that it leaves, is illustrated in figure 7.

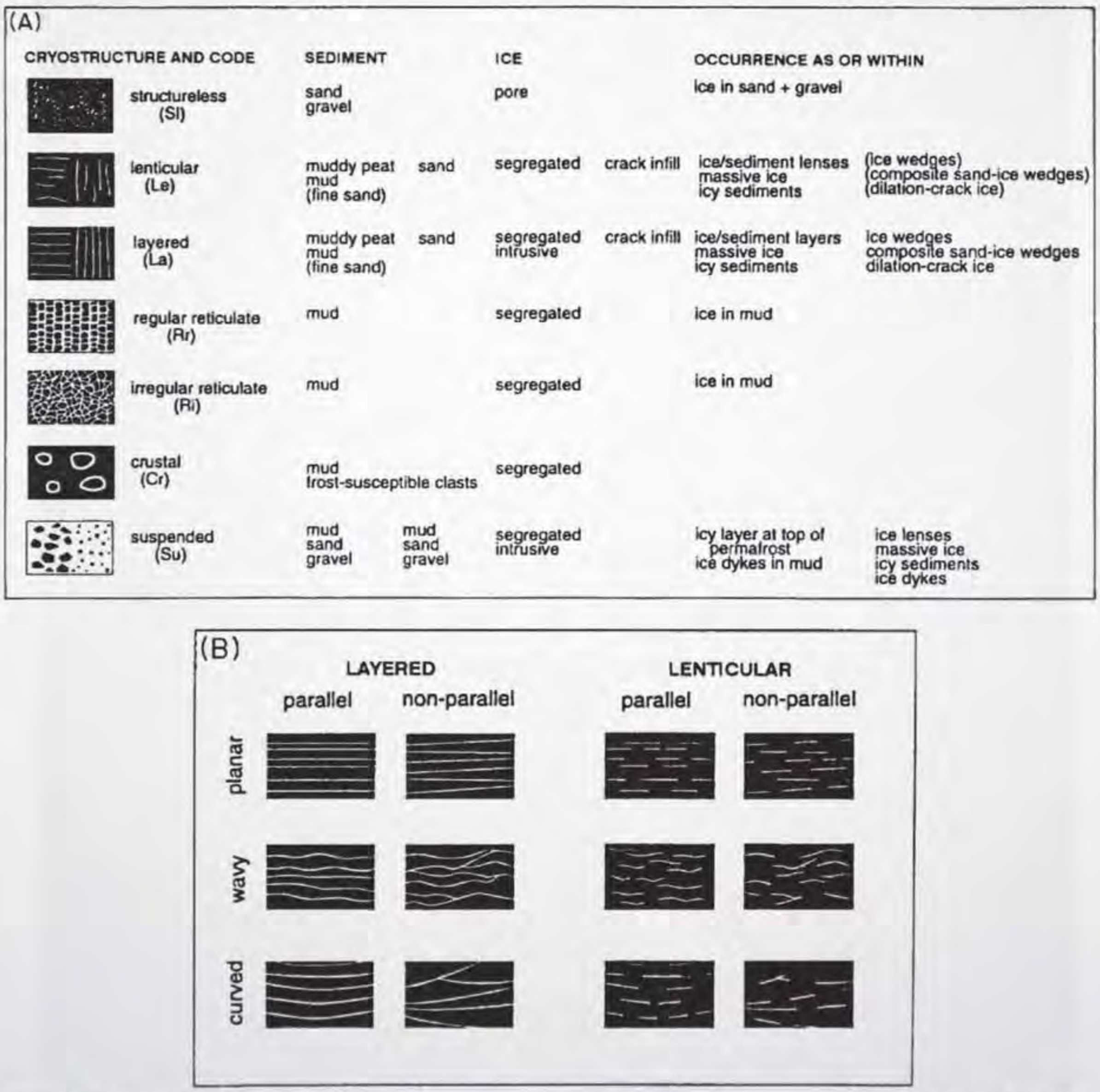


Figure 4. A North American classification of cryostructures. (A) Scheme proposed by Murton & French (1994). Ice is shown in white and sediment in black. (B) Terms and illustrations used to describe layered and lenticular cryostructures. (From Murton & French, 1994).

Figure 5. Photos showing (A) micro-lenticular cryostructure, location marker is 1.25 X 1.25 cm in size (photo by M. Kanevskiy), and (B) reticulate chaotic cryostructure, the handle of the knife is about 6 cm long (From Fortier et al. 2008).

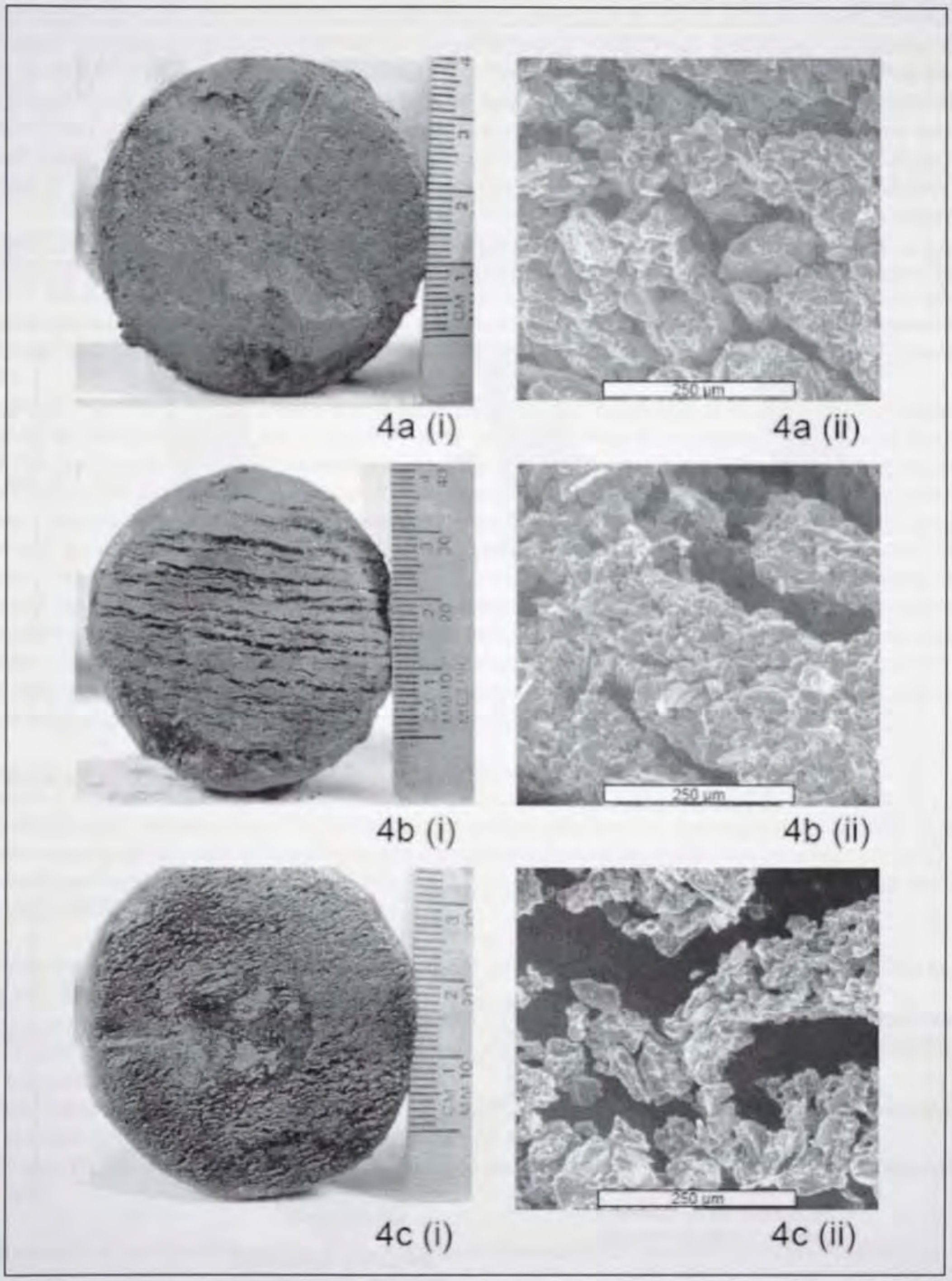


Figure 6. Example of cryostructures from the CRREL tunnel viewed conventionally and under ESEM. (a) Massive cryostructure. Image (i) is a macro-scale image typical of silt pseudomorphs. Note centimeter scale. Image (ii) is a micro-scale image using an ESEM. Bar scale indicates 250 mm. (b) Lenticular-layered cryostructure. Image (i) is a conventional macro-scale image. Note centimeter scale. Image (ii) is a micro-scale image using an ESEM that shows the soil and micro ice-lens morphology. Bar scale indicates 250 mm. (c) Micro-lenticular cryostructure. Image (i) is a macro-scale image. Note centimeter scale. Image (ii) is a micro-scale images under ESEM in which soil particles are generally suspended in an ice matrix. Bar scale indicates 250 mm. (From Bray et al 2006).

SYNGENETIC PERMAFROST

The permafrost exposed in the CRREL tunnel is typical of the Pleistocene-age permafrost referred to in the Russian literature as "Yedoma" or "Ice Complex" (e.g. Soloviev 1959, Katasonov 1978, Popov et al. 1985, Romanovskii 1993). Ice Complex sediments have been studied mainly in Central and Northern Yakutia; similar sediments were observed also in Chukotka, West Siberia, Taimyr, Alaska, and Canada (Vtyurin 1964, Péwé 1966, 1975, Popov 1967, Gasanov 1969, Katasonov 1978, Lawson 1983, Carter 1988, Hamilton et al. 1988, Shur et al. 2004, French 2007). This permafrost developed when long periods of uninterrupted cold-climate sedimentation allowed permafrost to form syngenetically.

Syngenetic permafrost forms in response to sedimentation (alluvial, slope, aeolian, lacustrine, etc.) that causes the base of the active layer to aggrade upwards. By definition, the permafrost is syngenetic because it is of the same

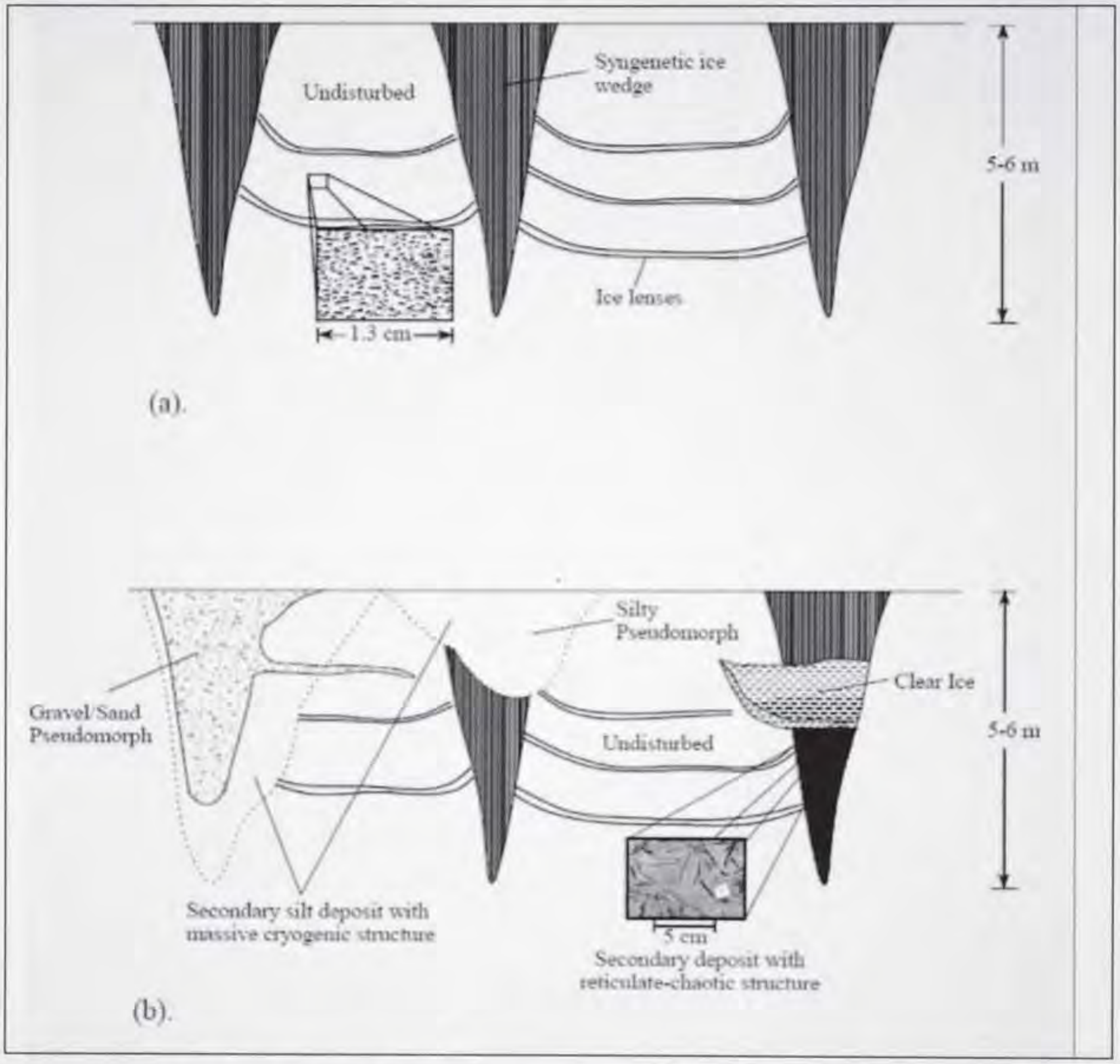


Figure 7. Schematic diagram of (a) undisturbed syngenetic permafrost and (b) typical modified permafrost exposure within the CRREL tunnel. In (a) the expanded image represents a micro-lenticular cryostructure, a reliable indicator of syngenetic permafrost. In (b) an idealized schematic shows typical secondary modification of original syngenetic permafrost. Expanded image represents reticulate-chaotic cryostructures. The reticulate-chaotic cryostructure is associated with 'clear ice', interpreted as thermokarst-cave ('pool') ice. (From Bray et al. 2006).

age (approximately) as the host sediment. It means that transformation of active-layer sediments into a perennially-frozen state occurs virtually simultaneously with sedimentation. Typically, syngenetically-frozen sediments are silty, or loess-like (up to 70-80% silt fraction), and ice-rich (the soil gravimetric moisture content may exceed 100-200%). They also contain rootlets, buried organic-rich horizons, and exhibit rhythmically-organized (i.e. layered) cryogenic structures.

The main locations where contemporary syngenetic permafrost is forming today are in the alluvial and deltaic environments (Shur & Jorgenson 1999) of Arctic North America (e.g. Colville River, Alaska; Mackenzie River, Canada), and in northern Siberia (e.g. Lena, Ob, Yenisey, Yana, Indigirka, Kolyma river valleys and deltas). Thickness of contemporary syngenetic permafrost usually does not exceed a few meters.

Pleistocene-age syngenetic permafrost occurs mainly in the continuous permafrost zone of Siberia, and its occurrence in the discontinuous permafrost zone of Alaska is a rare phenomenon. It is also found in the valleys and lowlands of adjacent unglaciated Yukon Territory, Canada. It should be mentioned that, under the current climatic conditions of the Fairbanks area, modern ice-wedge formation occurs very rarely and only in peat (Hamilton et al. 1983).

Syngenetic permafrost is often characterized by numerous ice veins and ice wedges. In contrast to epigenetic permafrost, in which ice wedges rarely exceed 4 m in depth, ice wedges in syngenetic permafrost may extend through the entire strata, either as huge wedges reaching 10-40 m in depth and 2-6 m in width or as small ice veins throughout the profile. Their varying width and depth reflect the varying rates of sedimentation and climate conditions. In syngenetic permafrost bodies, wedge ice can occupy 30-50%, and even more in some cases, of the total volume. In color, the wedges are grey because of the numerous inclusions of fine sediment. As such, they can easily be distinguished from smaller Holocene and modern ice wedges located in the top part of Yedoma sections, because the latter are usually white and opaque due to fewer soil inclusions and an abundance of air bubbles.

Radiocarbon dating and oxygen-isotope variations indicate that much of the syngenetic Pleistocene permafrost in northern Siberia formed between 40,000 and 12,000 years ago (Vasil'chuk & Vasil'chuk 1997, 2000). It was the dominant type of permafrost that formed in unglaciated lowlands during the Late Pleistocene. It is the main source of well-preserved Late-Pleistocene faunal remains (woolly mammoths etc).

CRYOSTRATIGRAPHIC MAPPING IN THE TUNNEL

Cryostratigraphic mapping uses the cryofacial method first proposed by Katasonov (1962, 1978). It is based upon two concepts, namely, that (1) the shape, size and spatial pattern of ice inclusions (i.e. cryostructures) depend on the conditions under which the sediment was deposited and then frozen, and (2) every cryofacies has its own specific cryostructure.

The results of cryostratigraphic mapping of the main adit of the tunnel (Bray et al. 2006) are shown in figures 8 and 9. Four categories of information are shown:

(a) Mineral sediment is grouped into the general categories of silt, sand and gravel.
(b) Cryostructures are identified as being either lenticular or micro-lenticular (i.e. syngenetic), massive (i.e. epigenetic) or reticulate-chaotic (i.e. epigenetic).
(c) Ice bodies are mapped as being either lenses (usually a layered cryogenic structure), wedge ice (formed in thermal-contraction cracks), or clear ice (occurring in association with wedge ice).
(d) Pseudomorphs are mapped where the sediment or ice is interpreted to be the result of ice-wedge modification.

Descriptions of the cryo-lithostratigraphic units, ice bodies and other features are shown in table 1.

In 2006, the 38 m long winze section was studied (Kanevskiy et al. 2008). Cryostratigraphic mapping of one wall and the ceiling of the winze was performed in the scale 1:20; several small sections were studied in detail (scale 1:4). Figure 10 shows the general view of the left wall and the ceiling of the winze. More detailed fragments are shown in figures 11 to 13.

CRYOSTRUCTURES

Recent publications (Shur et al. 2004, Bray et al. 2006, Fortier et al. 2008, Kanevskiy et al. 2008) show the variety of cryostructures that are present in the tunnel and describe typical features related to syngenetic permafrost formation.

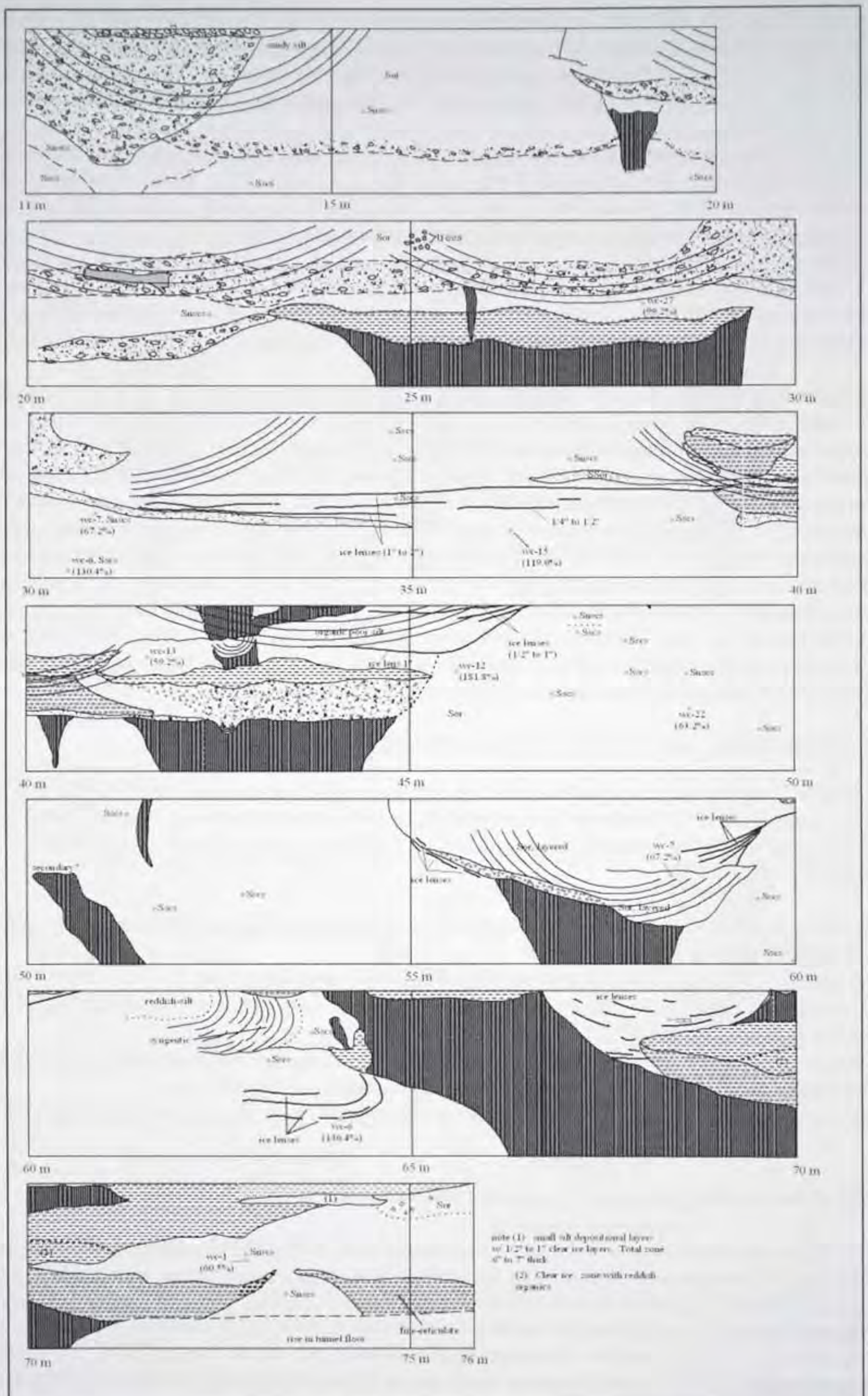


Figure 8. Cryostratigraphic map of part of the main shaft of the CRREL Permafrost Tunnel, left side (viewed from entrance) of the tunnel. For the legend, see Figure 9. (From Bray et al. 2006).

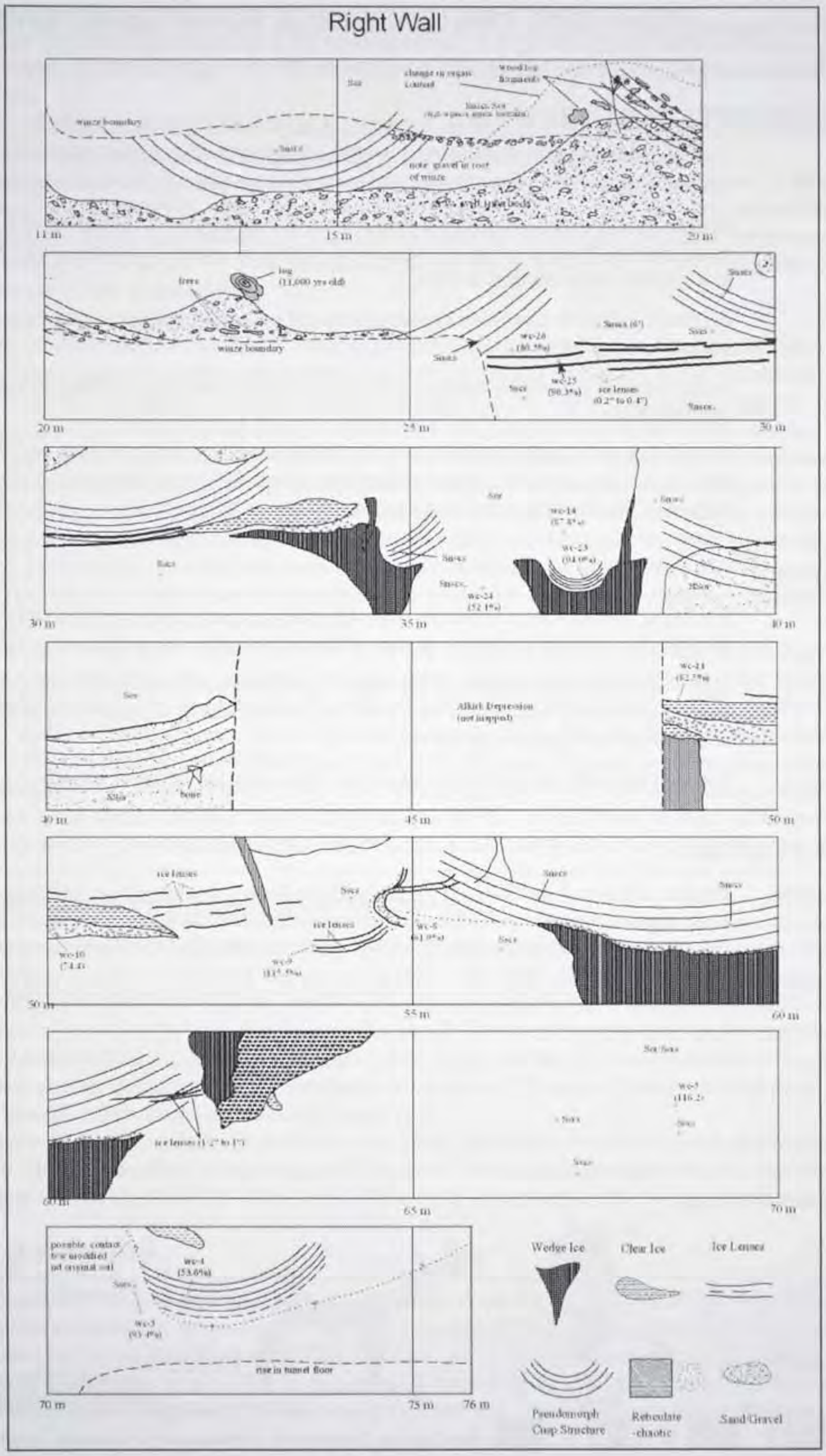


Figure 9. Cryostratigraphic map of part of the main shaft of the CRREL Permafrost Tunnel, right side (viewed from entrance) of the tunnel. (From Bray et al. 2006).

Table 1. Cryo-lithostratigraphic units, ice bodies and other features that were mapped in the CRREL tunnel (see figs. 8, 9).

1. CRYO-LITHOSTRATIGRAPHIC UNITS	
Sscs:	Fairbanks silt, representative of the original syngenetic permafrost, characterized by micro-lenticular and layered cryostructures. Average gravimetric water content 130%
Snscs:	Fairbanks silt, characterized by a massive cryostructure that is indicative of secondary modification. It contains no cryostructures typical of syngenetic permafrost. The average gravimetric water content is 69%
Sor:	Fairbanks silt with a massive cryostructure and possessing organics (rootlets, wood, animal bones, etc)
Ssor:	Sandy organic silt with a massive cryostructure and containing rootlets, wood and animal bones
Gr:	Gravel deposits; sandy, silt, imbricated. Where near the tunnel entrance, they may represent slope deposits. Deeper within the tunnel, the gravel deposits are directly related to the fluvial erosion and thaw-modification of ice wedges
2. ICE BODIES	
Ice lenses:	Lenses of ice that range in length from 10 cm to several meters and with thickness of between 0.5 to 10 cm. They form part of the micro-lenticular and layered cryostructures
Clear ice:	Lenticular-shaped ice bodies, often with aligned bubbles towards outer edges, and usually associated with reticulate-chaotic cryostructures in adjacent sediments. The ice is interpreted as thermokarst-cave ice
Wedge ice:	Foliated ice with vertical soil laminations, often grey to brownish in color
3. OTHER FEATURES	
Pseudomorphs:	Bodies of mineral soil ranging in composition from gravel to silt, commonly possessing high organic contents and often possessing reticulate-chaotic cryostructures. Interpreted as replacement deposits within previously thaw-eroded and truncated ice-wedge structures

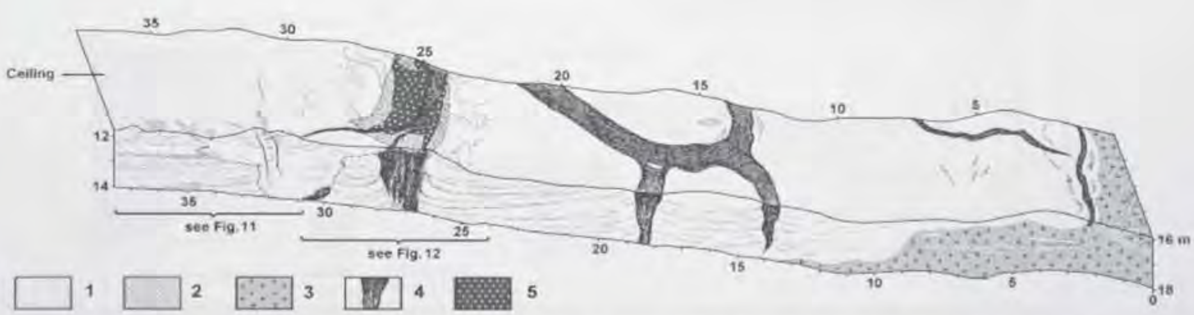


Figure 10. General view of the left wall and the ceiling of the winze. 1 – silt; 2 – sand; 3 – Fox Gravel; 4 – ice wedge; 5 – thermokarst-cave ice. See figures 11 to 13 for details. (From Kanevskiy et al. 2008).

Four types of cryogenic structure can be identified in the tunnel (Shur et al. 2004):

(1) A micro-lenticular cryostructure is the most common; it is formed by thin and short lenses of ice practically saturating the soil (fig. 5 A). The thickness of straight and wavy ice inclusions is generally less than 0.5 mm.

(2) A layered cryostructure is represented by repeated layers of ice with thickness of between 0.2 and 1 cm. The layers form series with the spacing between layers of between 2 and 5 cm.

(3) A lenticular-layered cryostructure is formed by ice lenses with a thickness from 0.5 to 1.5 mm and a length from a few millimeters to 1 cm. These lenses form continuous ice layers with soil inclusions.

(4) A reticulate-chaotic cryostructure is recognized by relatively thick multi-directional ice lenses (veins), often randomly oriented (fig. 5 B). This structure is interpreted as having formed following the closed-system freezing of a tálik or thaw layer.

The dominant cryostructure that can be observed in the CRREL tunnel is micro-lenticular (Shur et al. 2004). This is typical of syngenetic permafrost formation. The micro-lenticular term refers to the occurrence of very small, sub-horizontal (sometimes wavy), relatively short ice lenses. Usually, the thickness of uniformly-distributed ice lenses (and the spacing between them) does not exceed 0.5 mm (see fig. 5 A). In the winze (section 1, see fig. 13 A), several varieties of micro-lenticular cryostructure can be distinguished (e.g. latent micro-lenticular, micro-braided). Micro-lenticular cryostructures typically form more than 50-60% of the entire thickness of the syngenetic permafrost (Kanevskiy 2003). Usually the micro-lenticular cryostructure is combined with a layered cryostructure. In the tunnel, gravimetric moisture content of the sediments with micro-lenticular cryostructure varies from 80% to 240%. The great variability of gravimetric moisture content of silts can be associated with existence of the several varieties of micro-lenticular cryostructure mentioned above (see fig. 13 A). It is typical for syngenetic permafrost, and mostly linked to different rates of sedimentation.

Certain sections of the tunnel show bodies of clear ice. These are usually underlain by silt that exhibits a reticulate-chaotic cryostructure. This cryostructure is the most obvious type that is visible in the tunnel (Shur et al. 2004). However, it is not the most common cryostructure and is restricted to a few localities where it can be easily recognized by relatively thick ice veins, randomly oriented (see fig. 5 B). These multi-directional reticulate ice veins are thought to have formed by inward freezing of saturated sediments trapped in underground channels incised within the permafrost by thermal erosion. They form following cessation of flow as freezing occurs in sediments either laid down in the channel floor or which have slumped into the channel from the sides of the gully. Formation of the reticulate-chaotic cryostructure has been reproduced in laboratory experiments (Fortier et al. 2008).

AGGRADATION OF THE PERMAFROST TABLE

Seven thin organic-rich horizons can be observed in the upper part of the winze (see fig. 11). They occur at a depth of approximately 12-14 m below the ground surface. The AMS radiocarbon dates for organic-rich layers vary from 31,000 to 35,000 yr BP (fig. 11). Below each peat horizon, at a depth of approximately 0.4 to 0.6 m, are distinct icy layers (so-called 'belts' in the Russian literature). These are interpreted to be the temporary positions of the former permafrost table (i.e. base of the active layer) during the time of peat accumulation. In all probability, the peat reflects an environment of slower sediment accumulation. The approximate positions of the active layer during these periods are indicated by arrows in figure 11.

Numerous small cracks partially filled with ice (ice veins) extend downwards from the peat horizons to depths of up to 0.5 m. These cracks form polygons up to 0.5 m across. It is tempting to speculate that these were seasonal-frost cracks and that the seasonal-ice veins were subsequently incorporated into the syngenetic permafrost.

MASSIVE ICE BODIES

Bodies of massive ice are exposed in the wall and ceiling of the CRREL tunnel and impress the first-time visitor. These are the most visible expression of the ice-rich nature of Pleistocene-age syngenetic permafrost. Three types of massive ice can be identified: wedge ice, clear ice, and clear ice with wedge-ice inclusions.

Many of the ice wedges in the CRREL tunnel have been thaw-modified by fluvio-thermal erosion which promotes the formation of soil and ice pseudomorphs (see below). Figure 6 is a schematic diagram showing how thermal erosional processes may modify syngenetic permafrost.

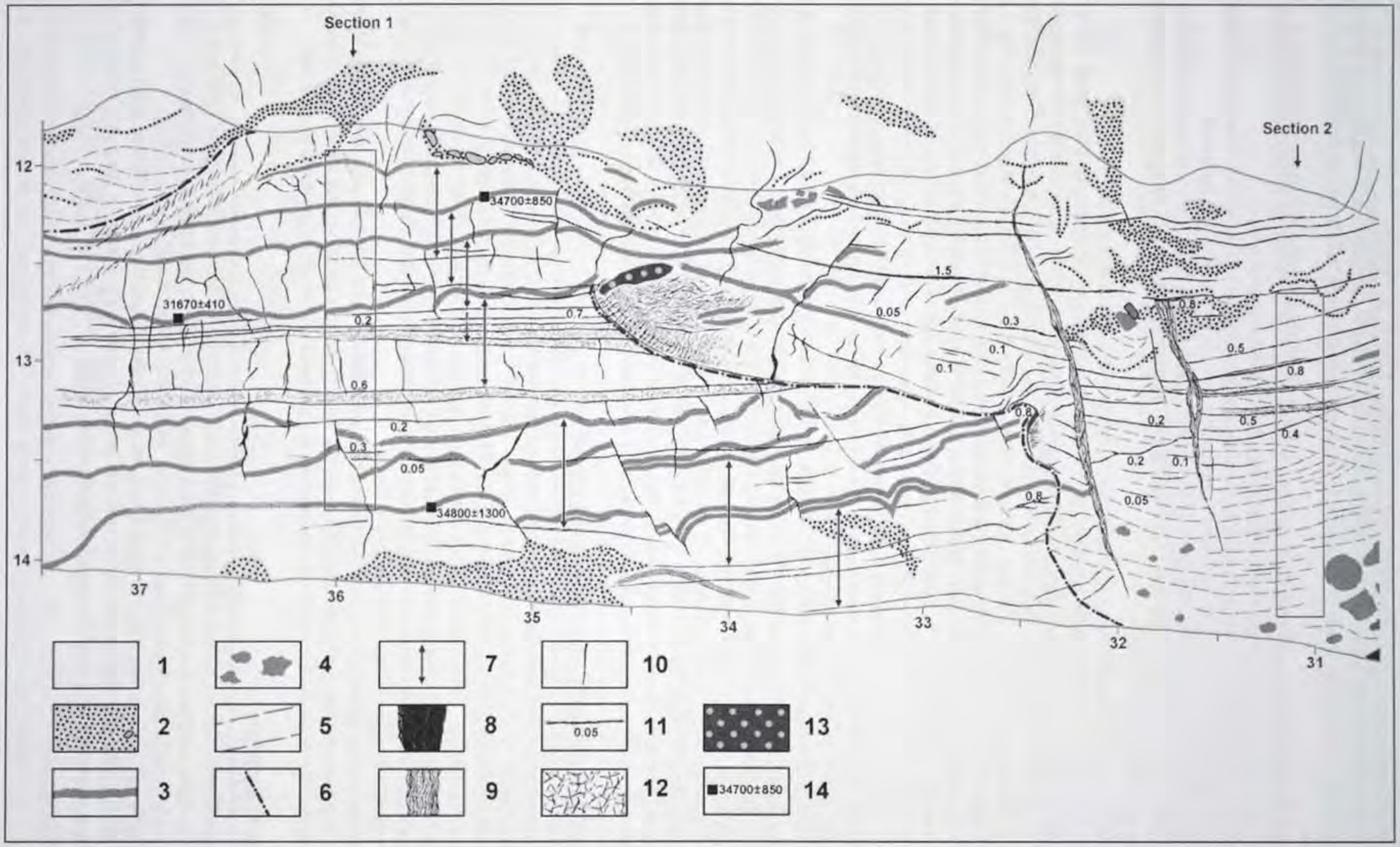


Figure 11. Cryostratigraphic map of the left wall of the winze, interval 31-37 m. 1 – silt; 2 – sand, gravel inclusion; 3 – in situ peat layers; 4 – inclusions of re-transported organic matter; 5 –lamination in silt; 6 – erosion boundary; 7 – approximate position of active layer at the periods of slower sedimentation; 8 – ice wedge; 9 – composite wedge (ice/silt); 10 – isolated ice vein; 11 – ice layer ('belt'), thickness in cm; 12 – reticulate-chaotic cryostructure; 13 – thermokarst-cave ice; 14 – radiocarbon date, yr BP. (From Kanevskiy et al. 2008).

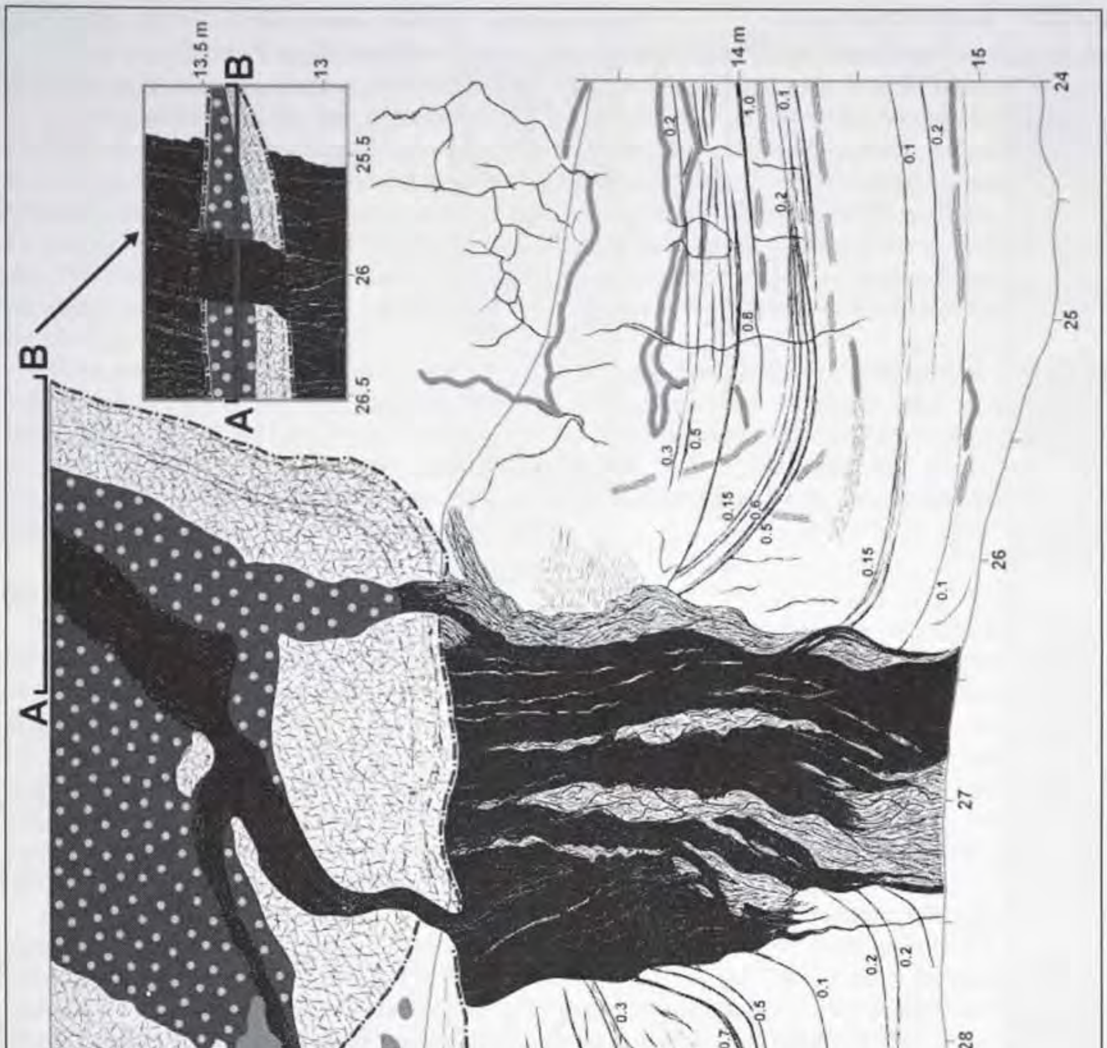


Figure 12. Cryostratigraphic map of the left wall of the winze, interval 24-31 m. For the legend, see Figure 11. A-B – schematic reconstruction of vertical section through the ice pseudomorph, located at the ceiling of the winze. (From Kanevskiy et al. 2008).

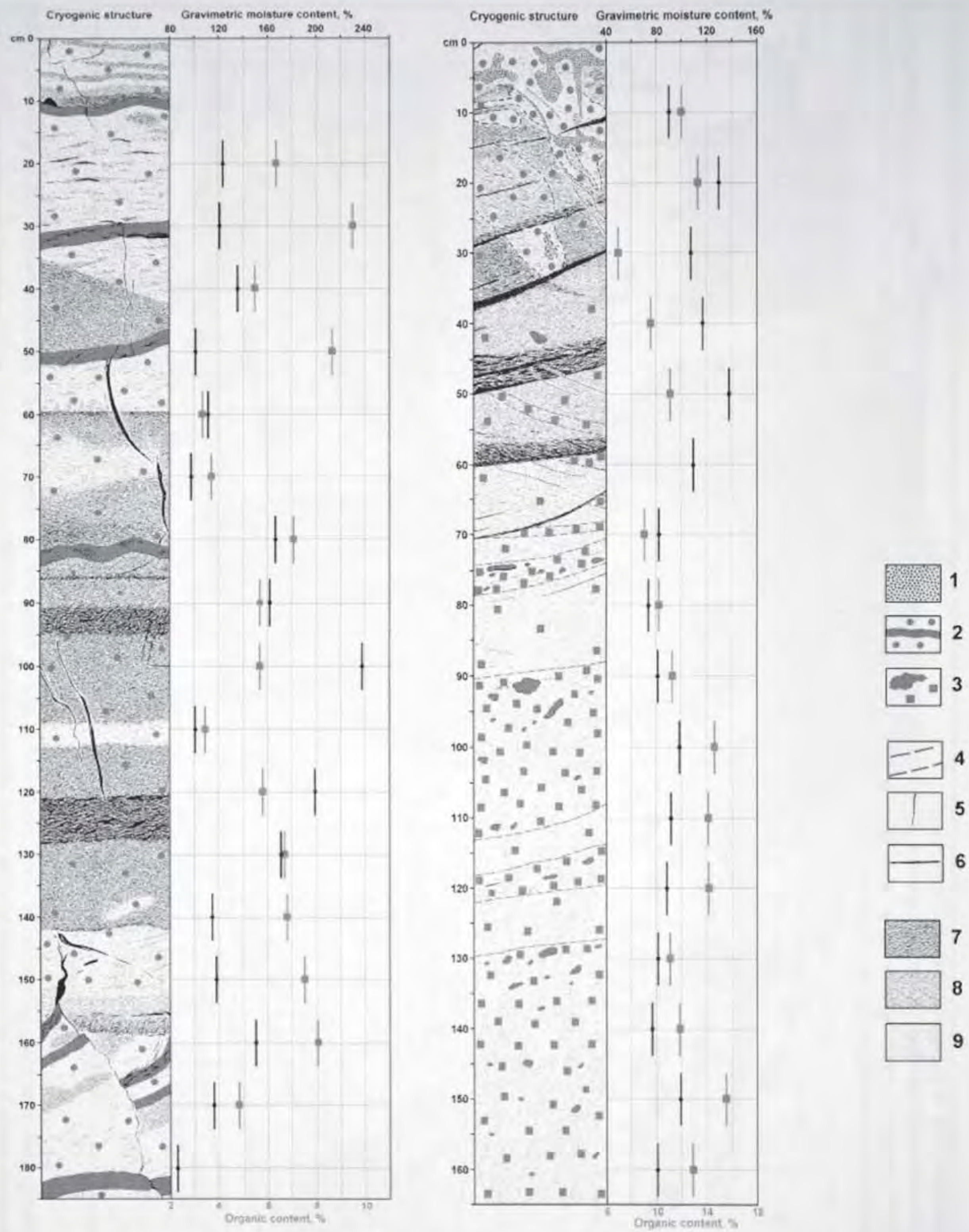


Figure 13. Details of cryogenic structure and properties of sections #1 (a) and #2 (b); location of sections is shown at Figure 11. 1 – sand; 2 – in situ peat layer and inclusions of organic matter; 3 – inclusions of retransported organic matter; 4 – lamination in silt; 5 – isolated ice vein; 6 – distinct ice layer ('belt'); 7 – micro-braided cryostructure; 8 – micro-lenticular cryostructure; 9 – latent micro-lenticular / porous cryostructure. (From Kanevskiy et al. 2008).

(a) Wedge ice

Wedge ice is the main type of massive ice that is present within the CRREL tunnel (fig. 14). It is easy to recognize by its wedge, or vertical, shape and by its foliated structure. Wedge ice is grey to dark brown in color; this reflects the presence of silt particles and organic staining within the ice. The size of the wedges is difficult to quantify. Although the wedges range in apparent width from 1 to 7 m, their true width probably varies between 0.5 and 3.0 m. It is also important to stress that only the middle and lower portions of the wedges are seen. Wedge ice is also present in the winze section where the wedges have an apparent width of up to 1.8 m. Here, the apex (nipple) of the wedge terminates at the stratigraphic contact between the overlying silts and the underlying alluvial gravels (fig. 10). The tunnel presents a great opportunity to see crossings of several ice wedges from inside: exposures of the wedge ice in the ceiling of the winze allow one to estimate the dimension of the ice-wedge polygons to reach 8-12 m (fig. 10).

When compared to the epigenetic ice wedges commonly described from northern and Central Alaska (e.g. Leffingwell 1919, Péwé, 1966), the wedges in the CRREL tunnel are average to large in size. However, when compared to some of the Late-Pleistocene syngenetic ice wedges described from Siberia along the Yana or Kolyma Rivers in northern Siberia (see Dostovalov & Popov 1966, Popov, 1973, Vasil'chuk & Vasil'chuk 1997), or the anti-syngenetic wedges inferred from the Pleistocene Mackenzie River Delta, Canada (see Mackay, 1995; French, 2007, 181), they appear to be average to small in size.

(b) Clear ice

We interpret the clear ice bodies in the CRREL tunnel to be thermokarst-cave ice (Shur et al. 2004, Bray et al. 2006). In North America, this is known colloquially as 'pool' ice (Mackay 1997). This is because ice-rich syngenetic permafrost is highly susceptible to thermal erosion that promotes the formation of subterra¬nean channels. When these channels are finally closed by sediment, water that is ponded behind the blockage begins to freeze. This process results in formation of thermokarst-¬cave ('pool') ice. The clear ice bodies are lenticular shaped. Their visible thickness in the tunnel ranges from a few centimeters to about 2 m and their extent beyond the ceiling is not known. The largest apparent horizontal extent of this type of ice that can be viewed in the tunnel is approximately 7 m. The alternative interpretation, that these clear ice bodies are buried surface, or pond, ice (Sellmann 1967, Hamilton et al. 1988), is not supported by the cryostructures present in the tunnel.

In the winze, a horizontal body of thermokarst-cave ice crosscutting the ice wedge is exposed on the ceiling (figs. 12, 14). Its thickness varies from 0.2 to 0.35 m and it is underlain by a silt layer (0.1-0.4 m thick) having a reticulate-chaotic cryostructure (Shur et al. 2004, Fortier et al. 2008). This massive ice body is aligned with the width of the ice wedge; however, it is wider than the wedge indicating that the initial subterra¬nean channel eroded laterally across the ice wedge into the enclosing sediments (see fig. 11, section A-B).

(c) Clear ice crossed by ice veins

There are many places in the tunnel where veins of wedge ice penetrate horizontal bodies of clear thermokarst-cave ice (fig. 15). This relationship demonstrates that the formation of wedge ice did not terminate when the cavity was filled with water and the water subsequently froze. Instead, it indicates that thermal-contraction cracking, ice-wedge formation, and permafrost growth continued after emplacement of the thermokarst-cave ice. The other example of where a horizontal body of thermokarst-cave ice is penetrated (i.e. crossed) by an ice wedge can be seen in the ceiling of the winze (see fig. 12, section A-B).

THERMAL EROSION, SOIL AND ICE PSEUDOMORPHS

Numerous sites of former gullies and underground channels can be observed in the silty sediments at various depths. They appear to have been cut by running water and afterwards filled with thermokarst-cave ice and soil whose structure and properties differ from the original syngenetic permafrost (Shur et al. 2004, Bray et al. 2006, Fortier et al. 2008).

The formation of thermokarst-cave ice is related to the gully erosion that must have occurred, especially during the spring snowmelt, during growth of the syngenetic permafrost on the relatively gently-sloping terrain of Goldstream Valley where the CRREL tunnel is located (Shur et al. 2004). It is necessary to stress that syngenetic permafrost, composed predominantly of ice-rich silty sediments, is especially susceptible to fluvio-thermal erosion (Shur et al. 2004, Bray et al. 2006, Fortier et al. 2007). Fluvio-thermal erosion occurs when surface runoff, from snowmelt, summer precipitation or thawing permafrost, becomes concentrated mainly along ice wedges causing preferential thaw. The gullies that result frequently assume an inverted 'T' cross-profile because water first erodes

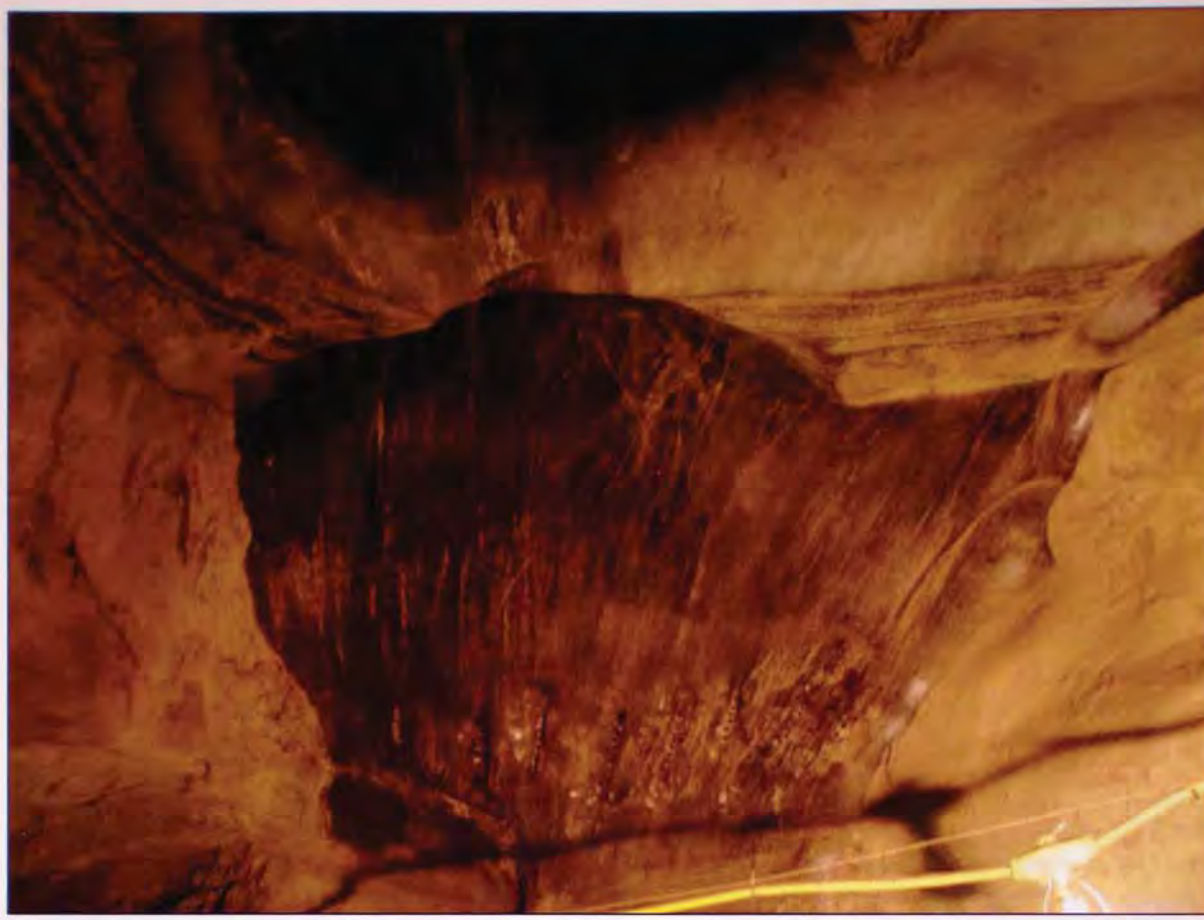


Figure 14. Photo showing ice wedge dissecting a horizontal lens of white thermokarst-cave ice (ice pseudomorph), right wall of the winze. The width of the wedge is 1.0 m. The thermokarst-cave ice body is underlain by several silt layers with reticulate-chaotic cryostructure. The same wedge exposed on the opposite wall of the winze is shown in Figure 12. (Photo by M. Kanevskiy).

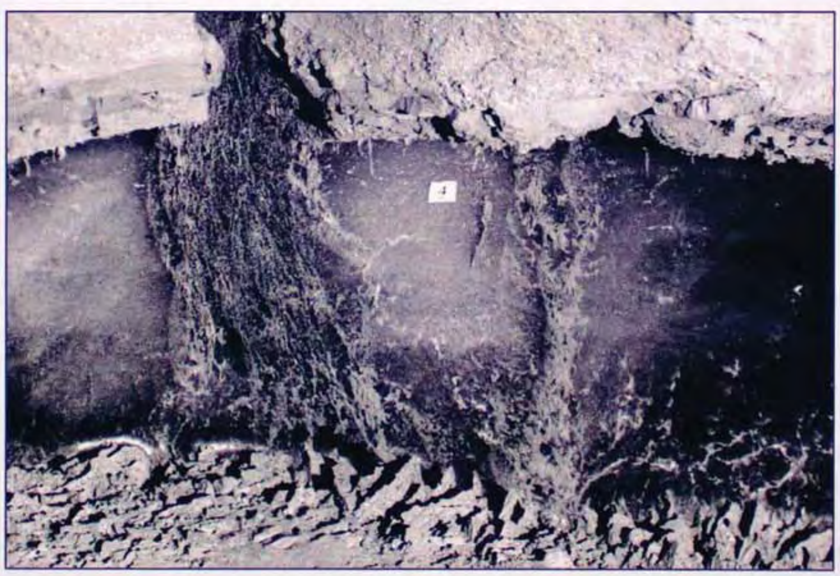


Figure 15. Photo showing veins of wedge ice penetrating a near-horizontal layer of thermokarst-cave ice. The location marker is 2.5 X 2.5 cm in size. (From Shur et al. 2004).

vertically and then, as the bed becomes armored with transported sediment from up-gully, laterally. This often leaves an organic-mat overhang. Slumping, piping and the creation of small tunnels above and adjacent to the partially-eroded ice wedge are also common. Fluvio-thermal erosion and the thaw-modification of ice wedges is a well known process in Arctic regions today (see French 2007, 191-2; Fortier et al. 2007). What is less well known is that, sometimes, standing water bodies accumulate in the channel floor behind slumped masses to form the ice bodies that Shumskii (1959) termed 'thermokarst-cave ice'.

The formation of pseudomorphs is related to the thaw-modification of ice wedges that would have occurred also during fluvio-thermal or underground-erosion episodes. Soil pseudomorphs are formed by silt or gravel filling the void left by the eroded ice wedge; ice pseudomorphs are formed by thermokarst-cave ice filling the void (fig. 6). These structures represent secondary infilling. Ice, sediment or an ice-sediment mix constitutes the infill. It is not surprising that the cryogenic properties of this infill material differ from the enclosing syngenetic permafrost. Formation of these structures is regarded as typical of syngenetic permafrost growth.

Both types of pseudomorphs (soil and ice) exist in the CRREL tunnel. However, they are difficult to recognize, especially ice pseudomorphs, because some have been subsequently modified by the penetration of ice veins. The incorporation of ice veins into ice pseudomorphs demonstrates that the formation of wedge ice is not terminated by the formation of thermokarst-cave ice (Shur et al. 2004). An ice pseudomorph modified by the penetration of ice veins is shown in figure 15.

Recent examination of the main adit showed that, of 20 ice wedges identified, 19 had been subject to thermal erosion. Approximately 60% of the channels cutting through the ice wedges and the enclosing syngenetic permafrost were partially or entirely filled by thermokarst-cave ice (Fortier et al. 2008).

In the winze, a gully filled with sediments can be observed at interval 29-35 m (see figs. 10 to 13). A truncated ice wedge affected by thermal erosion is located under this gully. The sediments filling the gully are mostly ice-poor stratified silts with lenses of sands. They contain numerous inclusions of organic material, which are interpreted as having been reworked by water. The organic content of the sediments in the gully varies from 7.0% to 22.8% by weight and is much higher in comparison with the original permafrost (Section 2, fig. 13 B).

Cryostructures in the lower part of section 2 (fig. 13 B, 60-165 cm) vary from latent micro-lenticular to porous (structureless). The gravimetric moisture content of this part of section 2 varies from 70% to 100% which is smaller than the water content of the original syngenetic permafrost. Such water content is unusual for sediments with very small amount of visible ice. It can be attributed to the higher organic content (fig. 13 B). Sediments with an organic content of 9-12% have a gravimetric moisture content of 70-80%, whereas sediments with organic content of 14-16% have a moisture content of 90-100%. The cryostructures and ice contents of the upper part of the section 2 (fig. 13 B, 0-60 cm) are similar to those of the original permafrost; here, the gravimetric moisture contents vary from 110% to 140%. This indicates change of sedimentation mode and decrease of sedimentation rate at the last stages of gully infilling.

ICE CONTENT

The ice content of sediments exposed in the tunnel varies widely. Although the weathered schist exposed in the lowest part of the Gravel Room contains a very small amount of visible ice, its gravimetric moisture content varies from 6.5% to 19.9%, averaging 11.7% (Hamilton et al. 1988). Gravimetric moisture content of alluvial gravel exposed in the lowest part of the winze generally is 8.9% to 10.3% (Hamilton et al. 1988). Typically, the gravel contains crustal cryostructures with thin ice crusts enclosing the gravel clasts. Close to the contact with overlying silt the thickness of ice crusts increases: sometimes it can reach 0.5-2.0 cm. Silt is generally ice-rich: gravimetric moisture content varies from 39% to 139% (Hamilton et al. 1988).

Recent studies show that the ice content of silt strongly depends on its cryostructure. For sediments with micro-lenticular cryostructure, gravimetric moisture content in the main adit varies from 80% to 180%, averaging 130% (Bray et al. 2006). A similar range (100-240%) is found in the winze (Section 1, fig. 13 A). For modified sediments with structureless (or massive) cryostructure, which fill gullies and soil pseudomorphs, gravimetric moisture content in the main adit varies from 50% to 95%, averaging 69% (Bray et al. 2006). For similar sediments in the winze, gravimetric moisture content is 70-100% (section 2, fig. 13 B, 60-165 cm). We associate the unusually high moisture content of ice-poor silt with the high content of reworked organic material in these sediments. The average gravimetric moisture content of the cross-stratified sands with structureless (or massive) cryostructure, filling underground channels, is 44.6%, whereas it is 107.7% in the surrounding permafrost with micro-lenticular cryostructure (Fortier et al. 2008). For sediments with reticulate-chaotic cryostructure, gravimetric moisture content in the main adit varies from 60% to 115%, averaging 85% (Bray et al. 2006).

TWO SILT UNITS?

The early studies in the tunnel (Sellmann 1967, 1972, Hamilton et al. 1988) did not adequately recognize the syngenetic nature of the permafrost. Two independent systems of ice wedges and an inferred thaw unconformity that separated the silts into an upper and a lower unit were recognized (see fig. 2). Some ice bodies, described as '...horizontal, saucer-shaped bodies' were interpreted as buried frozen thaw ponds formed in ice-wedge troughs, are better explained as bodies of thermokarst-cave ('pool') ice formed in underground channels.

There is now evidence that thermal-erosion processes were simultaneous with permafrost formation. First, the clear ice bodies are best explained as thermokarst-cave ice. Second, the occurrence of ice veins that penetrate thermokarst-cave ice means that ice wedges continued to grow after their partial thaw-modification and destruction by fluvio-thermal erosion and the subsequent pooling of water within the erosional void. Third, the dominant cryogenic structure is similar throughout the whole silt section. Fourth, the radiocarbon ages obtained from sediments within the tunnel do not reveal any sufficient break in sedimentation. In fact, no clear evidence of regional or widespread thermokarst can be found in the tunnel; instead, the thaw unconformities appear localized and connected with previous gullies and underground channels.

In summary, the cryostratigraphic data do not confirm the existence of two silt units divided by a continuous thaw unconformity, as described previously. Cryostructures, truncated ice bodies, and soil and ice pseudomorphs suggest a single sequence of continuous sedimentation and syngenetic permafrost aggradation in Late-Pleistocene time. This permafrost has been reworked by local thermal-erosional events.

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Resilience of Alaska's boreal forest to climatic change[1]

F.S. Chapin III, A.D. McGuire, R.W. Ruess, T.N. Hollingsworth, M.C. Mack, J.F. Johnstone, E.S. Kasischke, E.S. Euskirchen, J.B. Jones, M.T. Jorgenson, K. Kielland, G.P. Kofinas, M.R. Turetsky, J. Yarie, A.H. Lloyd, and D.L. Taylor

Abstract: This paper assesses the resilience of Alaska's boreal forest system to rapid climatic change. Recent warming is associated with reduced growth of dominant tree species, plant disease and insect outbreaks, warming and thawing of permafrost, drying of lakes, increased wildfire extent, increased postfire recruitment of deciduous trees, and reduced safety of hunters traveling on river ice. These changes have modified key structural features, feedbacks, and interactions in the boreal forest, including reduced effects of upland permafrost on regional hydrology, expansion of boreal forest into tundra, and amplification of climate warming because of reduced albedo (shorter winter season) and carbon release from wildfires. Other temperature-sensitive processes for which no trends have been detected include composition of plant and microbial communities, long-term landscape-scale change in carbon stocks, stream discharge, mammalian population dynamics, and river access and subsistence opportunities for rural indigenous communities. Projections of continued warming suggest that Alaska's boreal forest will undergo significant functional and structural changes within the next few decades that are unprecedented in the last 6000 years. The impact of these social–ecological changes will depend in part on the extent of landscape reorganization between uplands and lowlands and on policies regulating subsistence opportunities for rural communities.

Résumé : Cet article évalue la résilience du système forestier boréal de l'Alaska face aux changements rapides du climat. Le réchauffement récent est associé à la réduction de croissance des espèces d'arbre dominantes, aux maladies des plantes et aux épidémies d'insectes, au réchauffement et à la fonte du pergélisol, à l'assèchement des lacs, à l'augmentation de l'étendue des incendies de forêts, au recrutement accru d'espèces feuillues après un feu et à la diminution de la sécurité des chasseurs qui se déplacent sur le lit glacé des rivières. Ces changements ont modifié des caractéristiques structurales, des rétroactions et des interactions fondamentales dans la forêt boréale, incluant la réduction des effets du pergélisol des hautes terres sur l'hydrologie régionale, l'expansion de la forêt boréale vers la toundra et l'amplification du réchauffement climatique à cause de la diminution de l'albédo (saison hivernale plus courte) et des émissions de carbone provenant des incendies de forêt. D'autres processus sensibles à la température pour lesquels aucune tendance n'a été détectée incluent la composition des communautés végétales et microbiennes, les changements à long terme à l'échelle du paysage dans les stocks de carbone, le débit des cours d'eau, la dynamique de population des mammifères ainsi que l'accès aux rivières et les perspectives de subsistance des communautés indigènes rurales. Les projections concernant la poursuite du réchauffement indiquent que la forêt boréale de l'Alaska va subir au cours des prochaines décennies des changements fonctionnels

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F.S. Chapin III,[2] R.W. Ruess, E.S. Euskirchen, J.B. Jones, K. Kielland, and D.L. Taylor. Institute of Arctic Biology, Department of Biology and Wildlife, University of Alaska Fairbanks, Fairbanks, AK 99775, USA.
A.D. McGuire. Alaska Cooperative Fish and Wildlife Unit, US Geological Survey, University of Alaska Fairbanks, Fairbanks, AK 99775, USA.
T.N. Hollingsworth. USDA Forest Service, Pacific Northwest Research Station, Boreal Ecology Cooperative Research Unit, University of Alaska Fairbanks, Fairbanks, AK 99775, USA.
M.C. Mack. Department of Botany, University of Florida, P.O. Box 118526, Gainesville, FL 32611-8526, USA.
J.F. Johnstone. Department of Biology, University of Saskatchewan, 112 Science Place, Saskatoon, SK S7N 5E2, Canada.
E.S. Kasischke. Department of Geography, University of Maryland, 2181 LeFrak Hall, College Park, MD 20742, USA.
M.T. Jorgenson. Alaska Ecoscience, 2332 Cordes Way, Fairbanks, AK 99709, USA.
G.P. Kofinas. School of Natural Resources and Agricultural Sciences and Institute of Arctic Biology, University of Alaska Fairbanks, Fairbanks, AK 99775, USA.
M.R. Turetsky. Department of Integrative Biology, University of Guelph, Guelph, ON N1G 1G2, Canada.
J. Yarie. Forest Soils Laboratory, Department of Forest Sciences, University of Alaska Fairbanks, Fairbanks, AK 99775-7200, USA.
A.H. Lloyd. Department of Biology, Middlebury College, McCardell Bicentennial Hall 372, Middlebury, VT 05753, USA.

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[2]Corresponding author (e-mail: terry.chapin@alaska.edu).

et structuraux sans précédents depuis les 6000 dernières années. L'impact de ces changements socio-écologiques dépend en partie de l'ampleur de la réorganisation du paysage entre les zones sèches et humides et des politiques qui régissent les perspectives de subsistance des communautés rurales.

[Traduit par la Rédaction]

Introduction

The boreal forest is the northernmost and coldest forested biome. It is underlain by discontinuous permafrost (permanently frozen ground) that governs stand-level biogeochemistry and regional hydrology. The continental climate of interior Alaska results in cold winters (January mean –23 °C), warm summers (July mean 16 °C), and an annual precipitation (mean 287 mm) that is similar to that of midlatitude deserts (Hinzman et al. 2006). Despite low precipitation and warm summer air temperatures, permafrost, where present, results in cold, wet soils during summer that have in the past constrained rates of decomposition, nutrient turnover, and productivity (Van Cleve et al. 1991).

High-latitude amplification of 20th century global warming has caused Alaska's boreal forest to warm twice as rapidly as the global average (Arctic Climate Impact Assessment 2005; Hinzman et al. 2005; Intergovernmental Panel on Climate Change 2007). Mean annual air temperature in interior Alaska has increased by 1.3 °C during the past 50 years (Shulski and Wendler 2007) and is projected by downscaled climate models to increase by an additional 3–7 °C by the end of the 21st century (Walsh et al. 2008; Scenarios Network for Alaska Planning 2010). Precipitation has increased by only 7 mm in the last 50 years (Hinzman et al. 2006). Its projected continued increase will likely be insufficient to offset summer evapotranspiration (Scenarios Network for Alaska Planning 2010), leading to potentially drier soils and lower lake levels.

In response to a gradual Holocene cooling and moistening of climate, black spruce (*Picea mariana* (Mill.) Britton, Sterns & Poggenb.) assumed regional dominance 5000–7000 years ago, producing a landscape mosaic whose pollen and charcoal abundances were similar to those of today (Lloyd et al. 2006). People have occupied this region for at least the last 8000 years (Aigner 1986) and have therefore been part of the modern boreal forest since its inception. This boreal system has persisted relatively unchanged, within the detection capabilities of paleoecological indicators, for the past 6000 years, despite substantial climatic fluctuations such as the Medieval Warm Period and Little Ice Age (Mann et al. 2009). This suggests substantial historical resilience of the boreal forest to climatic change. However, warming since the 1950s appears to be unprecedented in at least the last 2000 years (Overpeck et al. 1997; Kaufman et al. 2009). In this paper, we synthesize and integrate the findings of the Bonanza Creek Long-Term Ecological Research (LTER) program, as detailed in other papers in this special issue. We conclude that the Alaskan boreal system remains quite resilient but is undergoing changes in ecosystem and landscape structure, feedbacks, and interactions that, with continued warming, will likely cause reorganization or potential transformation to a fundamentally different system.

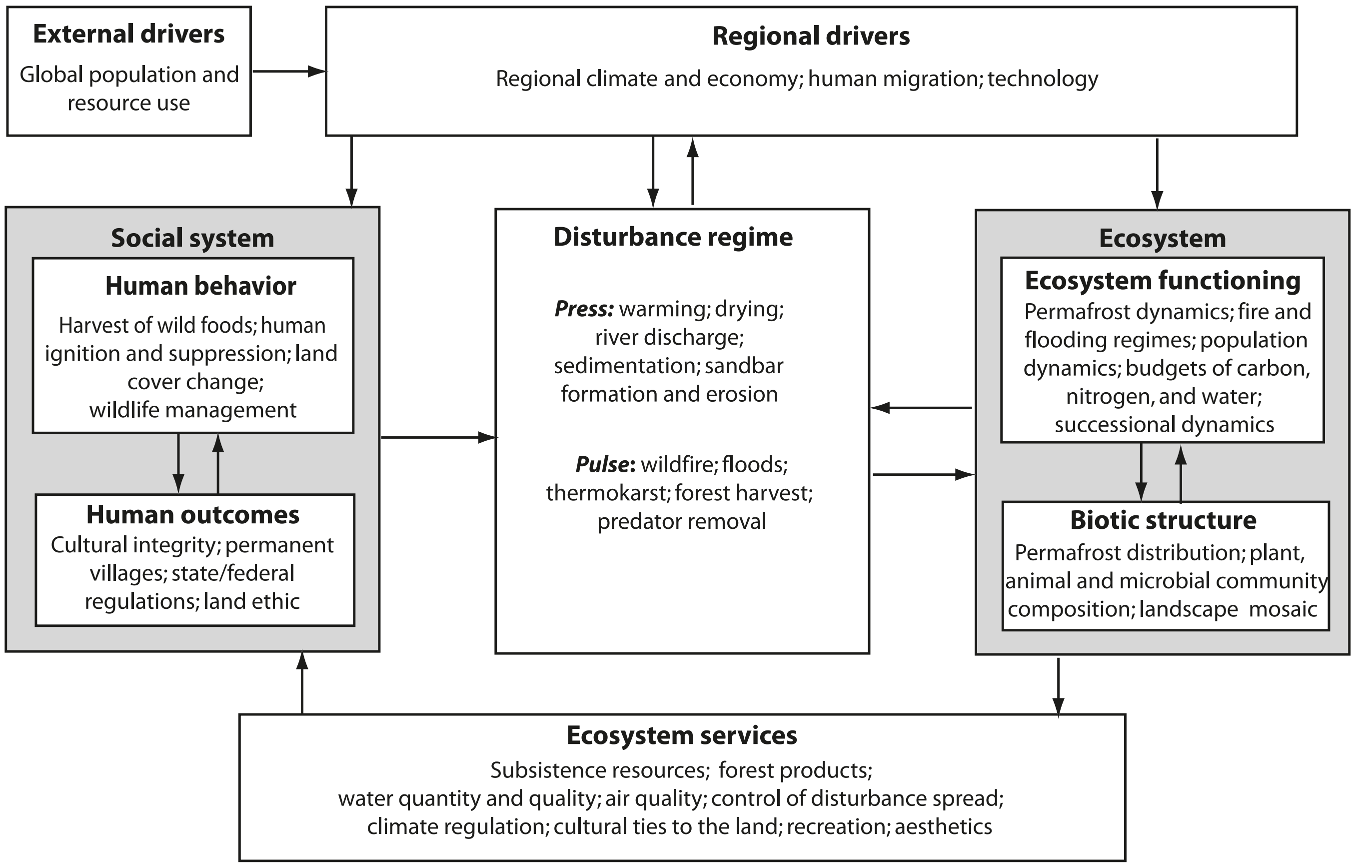
Conceptual background and framework

We view the boreal forest as a coupled social–ecological system in which social and ecological components interact in ways that govern the structure, functioning, and feedbacks of the system as a whole; a conceptual framework developed by the US LTER Network (Collins et al. 2007). Increases in the global human population and its use of resources and technology drive changes in Alaska that include climate warming, extraction of fossil fuels, and human immigration (Fig. 1). These changes in regional drivers, in turn, alter the natural disturbance regimes that govern the structure and functioning of ecosystems. Climate warming, for example, has modified ecosystem process rates through a variety of "press" perturbations (gradual changes) (Collins et al. 2007), including increased frequency of drought and lake drying (Fig. 1). In addition, warming modifies the frequency and severity of "pulse" perturbations (events), such as fires, floods, and thermokarst (ground depressions related to thawing permafrost). Climate warming and regional economic and demographic changes also influence social systems directly, thereby altering human effects on ecosystems, such as fire suppression, hunting, and land-use change.

The changes in press and pulse perturbations alter ecosystem structure and functioning in ways that affect "ecosystem services", the benefits that society derives from ecosystems (Fig. 1). Ecosystem services include food, fiber, and water (provisioning services); regulation of climate, disturbance regime, and hydrologic flows (regulatory services); and cultural, aesthetic, and recreational ties to the land (cultural services) (Millennium Ecosystem Assessment 2005). Ecosystem services, in turn, alter human behavior such as hunting patterns, food sharing, and demographic shifts, yielding human outcomes such as village structure, cultural integrity, land ethic, and hunting regulations (Fig. 1). The effect of ecosystem services on human behavior depends strongly on culture and human values. Although people have interacted with the boreal forest throughout its history, these interactions have changed substantially in the past century (Kofinas et al. 2010).

Vulnerability and resilience are conceptual frameworks that facilitate assessment of the long-term consequences of these changes. "Vulnerability" is the degree to which a system is likely to change due to exposure and sensitivity to a specified hazard or stress (Fig. 2, left side). Vulnerability also depends on a system's adaptive capacity to respond to the stress (Turner et al. 2003; Adger 2006). Adapting to climate change requires coping with current changes, learning the ecological and social consequences of change, innovating, and adapting to these changes. The boreal forest is ex-

Fig. 1. Diagram of the Alaskan boreal social-ecological system, based on the framework and concepts developed by the US Long-Term Ecological Research network (Collins et al. 2007). Global-scale drivers of change influence the regional drivers that directly perturb the boreal system through effects on social and ecological subsystems and the disturbance regime that links these subsystems. The resulting changes in dynamics alter ecosystem services, which modify human behavior and outcomes. Modified from Collins et al. (2007).



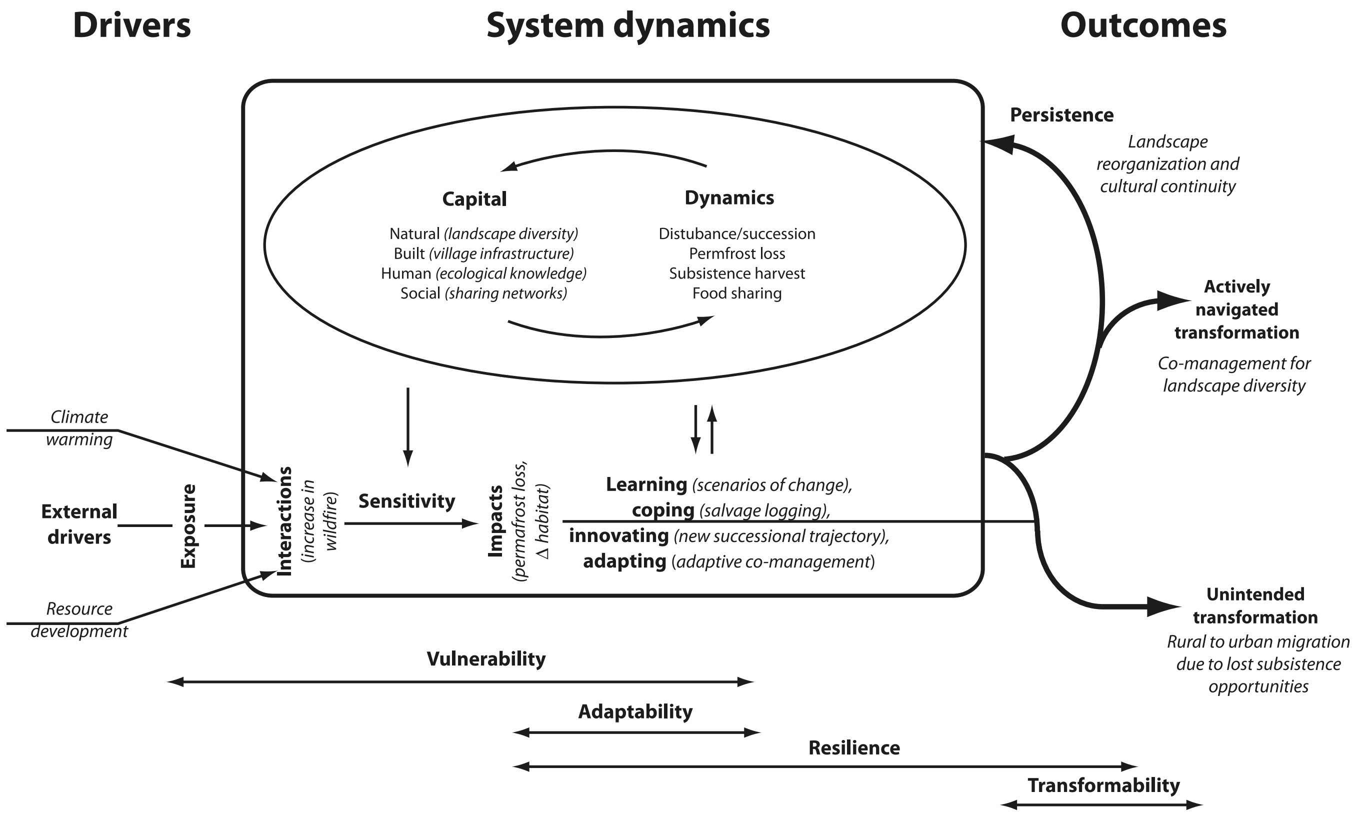
posed to a greater degree of climate warming than most places on Earth. This special issue documents the sensitivity and adaptive responses of the boreal system to climate warming with the aim of assessing the vulnerability of boreal forests to recent climate forcing.

"Resilience" is the capacity of the system to sustain its fundamental function, structure, and feedbacks when confronted with perturbations such as unprecedented warming (Walker et al. 2004; Folke 2006; Chapin et al. 2009*a*) (Fig. 2, right side). Resilience depends on the diversity of options available, the capacity to adapt to change, and the human capacity to adjust governance to implement new solutions. The boreal forest has properties that convey both *specific resilience* to particular perturbations (e.g., the semi-serotinous cones of black spruce that disperse seeds after fire) and *general resilience* (e.g., a diversity of species, successional trajectories, and patch dynamics that permit flexible response to a wide variety of expected and unforeseen perturbations) (Chapin et al. 2009*a*). Exceeding the resilience of the system may cause it to transform to some new state that has different system properties, that is sustained by different feedbacks, and that becomes resilient within this new domain of attraction (Fig. 2). The difference between persistence of a system (its resilience) or its transformation often depends on the balance between negative (stabilizing) feedbacks that tend to maintain the system in its current state and positive (amplifying) feedbacks that tend to push the system toward some new state (Chapin et al. 2009*a*). We assume that the boreal forest, like all ecosystems, has a suite of stabilizing feedbacks associated with competition, trophic dynamics, and successional cycles that sustain its characteristics over time. We hypothesize that certain climate-driven changes could initiate amplifying feedbacks that might transform the boreal forest to a new state (Fig. 2). Triggers for change might include the following:

- changes in soil moisture and hydrology associated with loss of permafrost, increased growing-season length, and seasonal timing of soil moisture recharge;
- changes in successional trajectory and biogeochemistry associated with changes in fire, flooding, or insect and (or) pathogen outbreaks;
- changes in abundance of keystone species (i.e., those that are disproportionately important relative to their biomass) or dominant species, including white spruce, alder, *Sphagnum* mosses, and snowshoe hares; and
- changes in human use of landscapes.

In this paper, we describe sources of boreal resilience. We also ask whether there have been changes that have the potential to trigger transformation and whether these transformations have begun to occur. A final consideration, about which we can only speculate, is the permanence or reversibility of ongoing or potential transformations.

Fig. 2. System perspective on the vulnerability and resilience of the Alaskan boreal forest, showing how changes in drivers (left side) modify the system to produce resilient or transformational outcomes (right side). The arrows at the bottom show the range of processes associated with vulnerability, adaptability, resilience, and transformability. Multiple external stresses (climate warming and resource development) interact to increase the extent and severity of wildfire. This, in turn, creates impacts (permafrost loss and habitat change). Social learning (learning, coping, innovating, and adapting) in response to these impacts has the potential to alter social–ecological interactions and various forms of capital of the system, which in turn, influence sensitivity to future events. Social learning also governs the relative likelihood of three potential outcomes: persistence of the existing system through resilience; actively navigated transformation to a new, potentially more beneficial trajectory through transformation; or unintended degradation to a new state due to vulnerability and failure to adapt or transform. Modified from Chapin et al. (2009*a*).



Climate sensitivity

Changes in uplands and well-drained sites

In contrast to the continuous permafrost zone of the Arctic, changes in discontinuous permafrost of the boreal forest are driven primarily by changes in ecosystems rather than by climatic change (Jorgenson et al. 2010). In interior Alaska, for example, changes in the insulative properties of snow, surface water, and vegetation and in the surface organic layer have altered permafrost integrity more than have changes in air temperature. Consequently, topographic and successional variations in these ecosystem properties lead to a spectrum of permafrost responses to climate warming (Jorgenson et al. 2010). In north-facing uplands, where there is no impoundment of surface water, gradual warming of permafrost has not greatly altered permafrost integrity except through occasional gully formation, whereas accumulation of surface water in thawing ice-rich lowlands or loss of organic insulation after severe fires leads to rapid permafrost loss.

Upland hydrology has also been relatively resilient to Alaska's long-term warming trend (Jones and Rinehart 2010). Although there is generally less discharge in warm years than in cool years, presumably due to greater evapotranspiration, there has been no detectable temporal trend in base flow or total summer discharge over the past forty years. Winter flow appears to be increasing in regions of the larger Yukon River Basin where permafrost is discontinuous (Walvoord and Striegl 2007). These results suggest that upland soils might continue to dry, and headwater streams could become ephemeral, if warming continues (Jones and Rinehart 2010).

Earlier snowmelt and later freezeup in interior Alaska (Euskirchen et al. 2010; Juday et al. 2005) have lengthened the growing season, causing trees to leaf out earlier in spring (Robin et al. 2008). Earlier spring leaf out enhances both photosynthesis and ecosystem respiration and, in moist systems, might increase carbon sequestration (Welp et al. 2007; Richardson et al. 2009). However, the functional consequences of longer and warmer growing seasons for water, carbon, and nutrient balance of Alaskan boreal ecosystems vary with soil moisture (McGuire et al. 2009), whose temporal trends depend on landscape position.

Fertilization and moisture exclusion studies show that moisture is now the primary factor limiting production of Alaskan boreal trees (Yarie and Van Cleve 2010). Nutrient responses occur primarily early in the season when soils are cold, in cool moist sites, or in wet years. These recent findings of strong drought effects contrast with earlier research, conducted when climate was cooler, which showed widespread nutrient limitation of forest production at our LTER sites (Van Cleve et al. 1986, 1991).

Because of the widespread occurrence of drought, most tree species in interior Alaska exhibit negative growth responses to warming (Juday et al. 2005; McGuire et al. 2010), a pattern that is consistent with declines since 1990 in greenness indices measured by satellites (Goetz et al. 2005; Lloyd and Bunn 2007; Verbyla 2008). Reductions in tree growth have been examined most thoroughly in white spruce (*Picea glauca* (Moench) Voss), the late-successional tree that dominates warm south-facing uplands and lowland floodplain forests. In this species dendrochronological, population, and experimental rainfall exclusion studies show that individual trees show a spectrum of growth responses to warming and rainfall, ranging from positive to negative. Negative responses of growth to temperature predominate over positive responses in this species (McGuire et al. 2010). White spruce is most negatively affected by warming in warm regions and in floodplain landscape positions where river levels are dropping. Even in cool environments such as tree line, formerly positive responses of tree growth to warming are now changing to growth reductions in many locations (Wilmking et al. 2004).

Recent extensive insect outbreaks on white spruce, aspen (*Populus tremuloides* Michx.), and larch (*Larix laricina* (Du Roi) K. Koch) might be both a response and contributor to reductions in tree growth. In the Kenai Peninsula of southern Alaska, extensive stand-level mortality of white–Sitka (*Picea sitchensis* (Bong.) Carrière) spruce hybrids, associated with warming and drying, led to dense growth of the grass *Calamagrostis canadensis* (Michx.) P. Beauv. and very poor spruce regeneration, suggesting a switch from forest to grassland (Berg et al. 2006). In interior Alaska, upland white spruce has dispersed into adjacent black spruce habitat after fire, producing seedling densities sufficient to generate fully stocked white spruce stands (Wirth et al. 2008). Together these studies suggest a low stand-level resilience of white spruce and perhaps other forest types, but the possibility of landscape-level resilience, if upland species shift into landscape positions that are currently dominated by black spruce. This landscape reorganization might occur extensively if wildland fire continues to increase in extent and severity in black spruce forests, as described later.

Changes in lowlands and poorly drained sites

In some poorly drained lowlands, permafrost shows low resilience to climate warming. Here thawing of high-ice-content permafrost, formed during the Little Ice Age, has in recent decades, caused surface subsidence and conversion of forests to ponds or wetlands, which absorb more radiation due to the low albedo (short-wave reflectance) of the wet surface, thereby accelerating the rates of thaw and landscape change (Jorgenson et al. 2010).

In lowlands underlain by gravel, high-ice permafrost is less common, and climate warming causes drying of lakes because of increased evapotranspiration and, in some situations, loss of permafrost and internal drainage (Jorgenson et al. 2010; Riordan et al. 2006). Willows, which colonize drained lake basins, have high rates of evapotranspiration, causing further drying. Lake drainage and wetland drying currently predominate over paludification in interior Alaska (Riordan et al. 2006).

In many peatlands and black spruce forests, the most widespread forest type in interior Alaska (Calef et al. 2005), permafrost has been relatively resilient. In these systems, water table depth and plant species composition strongly influence biogeochemical dynamics. *Sphagnum* and other mosses are keystone boreal taxa that account for 20% and 48% of production in uplands and wetlands, respectively (Turetsky et al. 2010). These mosses have tended to increase in abundance in late-successional stands, perhaps in response to insect-induced canopy reduction (Hollingsworth et al. 2010; Turetsky et al. 2010). The effective thermal insulation and low litter quality of mosses, especially *Sphagnum*, lead to cold, permafrost-dominated, nutrient-poor soils that constrain rates of decomposition, nutrient cycling rates, and therefore forest productivity (Turetsky et al. 2010). These conditions lead to carbon sequestration (Ping et al. 2005; Hollingsworth et al. 2008). Despite slow rates of overall carbon and nitrogen cycling, small pools of dissolved organic nitrogen (DON) cycle rapidly, with rates that are controlled by root and (or) mycorrhizal turnover (Ruess et al. 2006) and by competition for DON between plants (and their mycorrhizal fungi) and decomposers (Kielland et al. 2007; Chapin et al. 2009*b*). The stabilizing (negative) biogeochemical feedbacks associated with cold, wet soils contribute to the resilience of black spruce forests in landscape positions with stable permafrost (Johnstone et al. 2010*a*).

As in most forested ecosystems, Alaskan microbial communities are dominated by fungi, especially by ectomycorrhizal and ericoid mycorrhizal fungi (Taylor et al. 2010). Fungal diversity is at least 10-fold greater than plant diversity in the same sites. Fungal community composition is determined primarily by forest type, soil horizon, and time of year but does not differ among years. These observations suggest extreme fungal specialization to ecosystem structure and seasonality but low sensitivity to interannual variation in climate. We therefore hypothesize that changes in microbial communities and the biogeochemical processes mediated by them will be more sensitive to climatically driven changes in disturbance and vegetation than to the direct effects of climate warming.

Changes in floodplains

In river floodplains, successional pathways have changed in ways that might have important functional consequences. Thinleaf alder (*Alnus incana* (L.) Moench subsp. *tenuifolia* (Nutt.) Breitung) is a keystone nitrogen-fixing species that expanded in both old and young successional stands along the Tanana River during the 1990s (Hollingsworth et al. 2010; Nossov 2008). The alder expansion in older stands may reflect improved light availability associated with canopy reduction of white spruce by insect outbreaks (response to warmer summers) and ice storms (warmer winters) and of poplar by expanding beaver populations (unknown cause). Increased seed input from alder expansion in mature sites might explain recent increases in alder recruitment in early successional sites. In addition, moose, which have increased in abundance in the Tanana River lowlands due to predator control (see below), browse heavily on willows, releasing early successional alder recruits and invasive plants from competition. Modeling of the functional responses of these interactions between plant competition and herbivory (Feng et al. 2009) suggests that climate warming and predator control are facilitating invasion of exotic nitrogen fixers (Wurtz et al. 2008) and reducing carrying capacity for moose (reduced forage biomass and palatability) (Kielland et al. 2006). These changes indirectly reduce recruitment of white spruce (Angell and Kielland 2009). Although the data record is too short and the connections to climate too unclear to assess long-term persistence of these trends, the patterns suggest that forest resilience in floodplains depends on stand renewal through historically important successional pathways after river disturbance, but the long-term consequences of novel successional pathways characteristic of warmer drier conditions are less certain. The relative frequency of historical versus novel successional pathways will likely depend on changes in climate, herbivory by moose (a function of predator control and wildfire extent) (Kielland et al. 2006), and alder canker, an expanding forest pathogen that reduces alder abundance and its rates of nitrogen fixation and seed production (Ruess et al. 2009).

The reduction in alder growth and nitrogen fixation by alder canker is important because of alder's keystone role in nitrogen accumulation during floodplain succession (Ruess et al. 2009). Alder canker is most widespread in dry sites and dry years, suggesting a climate link to its spread. Green alder in the uplands also interacts with drought-mediated diseases (Mulder et al. 2008), with important implications for postfire nitrogen economy (Mitchell and Ruess 2009).

In summary, recent research suggests a high sensitivity of the Alaskan boreal forest to changes in moisture availability and to new species interactions (e.g., with insects and pathogens) that are emerging. This research suggests that controls over boreal forest dynamics are shifting from temperature and nutrient limitation, which were well documented in the 1970s and 1980s (Van Cleve et al. 1986, 1991), to more frequent limitation by drought.

Landscape consequences and societal implications

Fire and permafrost

Long-term studies by the Bonanza Creek LTER suggest that boreal forest stands show a mixture of sensitivity and resilience of different functional components in response to climatic change and that the net effect of these changes depends, in part, on reorganization of landscape patterns. We briefly summarize the landscape consequences of the changes described above and discuss their implications for society.

In the past decade, an increase in the number of years with extensive fires has doubled the annual area burned in interior Alaska compared with any decade of the previous 40 years and is 50% higher than any decade since 1940 (Kasischke et al. 2010). The area burned during late-growing season fires in the 2000s was three times higher than in any previous decade since 1950. Warmer and drier summers allow fires to continue burning in late summer, when soils are deeply thawed and have lower soil moisture, and therefore burn more deeply, creating a radically different soil environment for seedling establishment. These severe fires have disrupted conditions for black spruce regeneration that sustained black spruce dominance for thousands of years (Johnstone et al. 2010*b*; Turetsky et al. 2010). Stabilizing feedbacks that have sustained the resilience of black spruce forests include biogeochemical feedbacks (thermal insulation by mosses, presence of permafrost, moist soils, and low fire severity) and life-history feedbacks (high availability of black spruce seeds from on-site semiserotinous cones and effective establishment on organic seedbeds). The recent increase in mineral soil seedbeds and the reduction in fire interval have generated new successional trajectories dominated by deciduous tree seedlings (Johnstone et al. 2010*b*; Kasischke et al. 2010). In extremely dry sites, no tree recruitment may occur after fire (Kasischke et al. 2007; Johnstone et al. 2010*a*).

These observations suggest potential shifts in the relative abundance of forest types that currently dominate the Alaskan boreal forest: a decline in abundance of black spruce, which has dominated the lowland landscape and north-facing slopes for the last 6000 years; a potential increase of deciduous forests in former black spruce habitat; and a conversion to grasslands or shrublands on dry sites (Johnstone et al. 2010*a*). Deciduous forests, until now, have been largely restricted to south-facing uplands and floodplain corridors and have acted as a stabilizing feedback to fire probability and spread because of their high leaf moisture content and low flammability. As climate warms, however, vegetation effects on flammability decline, weakening this stabilizing feedback, so the areal extent of fire is projected to continue increasing with climate warming despite the shift to deciduous vegetation (Kasischke et al. 2010). Hardwoods accumulate less soil organic matter than spruce ecosystems (Van Cleve et al. 1983; Mack et al. 2008), so increased hardwood dominance might reduce carbon sequestration at landscape scales.

Historically, the high latent heat content of ice-rich permafrost enabled permafrost to persist after fire until moss-dominated black spruce communities fostered permafrost recovery (Jorgenson et al. 2010). With a switch to deciduous-dominated vegetation, the thick moss and organic layers are unlikely to rebuild, and permafrost will probably continue to degrade (Johnstone et al. 2010*b*; Jorgenson et al. 2010; Turetsky et al. 2010). We hypothesize that declines in stand-level resilience in community composition in both uplands (due to climatic sensitivity to drought) and lowlands (due to changes in successional feedbacks) convey substantial resilience of species composition at landscape-to-regional scales as a result of potential redistribution of stand types across the landscape. Long-term observations are required to test this hypothesis.

Landscape changes in the boreal forest alter its role in the global climate system. The most dramatic of these feedbacks is earlier snowmelt, which amplifies climate warming as a result of reduced albedo (Euskirchen et al. 2010). This is slightly offset by an increase in area burned, which increases albedo in winter (less forest cover to obscure the snow) (Liu et al. 2005; Randerson et al. 2006) and in summer (shift to a lower-albedo deciduous forest trajectory) (Euskirchen et al.

2010). Changes in trace gas (mainly CO_2 and CH_4) feedbacks from the boreal forest are less clear. Greater areal extent and depth of burning and insect outbreaks reduce carbon sequestration (McGuire et al. 2009), but the consequences of the hydrologic reorganization of landscapes are more complex. Sites that are drying generally show reduced rates of carbon, nitrogen, and water cycling, with greater declines in photosynthesis than in ecosystem respiration (and therefore a decline in carbon sequestration); sites that are getting wetter show the opposite trends (Euskirchen et al. 2010). Methane efflux also declines with drier conditions.

Mammals

Observed and projected changes in environment and vegetation will influence the responses of mammals to climate warming. In general, small mammals such as microtine rodents are more sensitive to interannual variations in weather than are large mammals such as moose and caribou that respond more strongly to variations in food supply and predation (Kielland et al. 2006). Snowshoe hares are quite sensitive to all of these factors (Kielland et al. 2010). Snow depth, rain-on-snow events, floods, and other climate-related events that are difficult to predict are projected to become more variable (Intergovernmental Panel on Climate Change 2007) and are likely to exert stronger effects on most mammals than will warming per se. Habitat changes resulting from warming-induced increases in floods, insect outbreaks, and wildfires could also strongly affect mammal distributions. In general, we hypothesize that most mammalian communities will show low resilience, with some species declining in abundance (e.g., lichen-dependent caribou), others increasing (e.g., fire-dependent moose and snowshoe hares), and others changing in species composition and distribution (e.g., microtine rodents that are sensitive to extreme events but some of which are favored by grasslands). These changes, if they occur, would substantially restructure the mammalian communities of interior Alaska.

Human communities

Changes in environment, ecosystems, and subsistence resources have important implications for Alaskan boreal communities, particularly those in rural areas where indigenous people have historically led a subsistence lifestyle as hunters and gatherers (Fig. 2). Warming directly affects communities as a result of thinner river ice and therefore reduced safety of winter travel and access to hunting grounds. Increased evapotranspiration and declining river discharge also reduce opportunities for barge delivery of fuel and increase the cost of living and therefore the dependence on subsistence harvesting (Kofinas et al. 2010). Now that communities are permanently situated rather than seminomadic, increased wildfire risk constitutes the major pathway by which warming influences rural communities. Wildfire constitutes a risk to life and property, reduces access to the land, threatens cultural resources, and reduces moose and caribou abundances for one to several decades (Maier et al. 2005; Chapin et al. 2008; Kofinas et al. 2010). Sources of resilience to address these changes include traditional sharing networks that maintain community identity while sustaining food supplies to the most vulnerable households and allowing hunters to borrow hunting equipment. As the abundance and distribution of subsistence resources change and access to hunting areas is modified, hunters will likely shift their hunting effort to those species that increase in availability and (or) accessibility. Development of community gardens or changes in hunting regulations to constrain competition from urban hunters could enhance resilience (Kofinas et al. 2010; Loring and Gerlach 2010). Changes in economic conditions, such as employment in rural and urban communities, will undoubtedly interact with the effects of climatic change, affecting human migration patterns and the overall resilience of villages. In summary, climate warming and socioeconomic changes challenge the resilience of rural indigenous communities, but indigenous culture has proven relatively resilient to even greater threats over the past century.

Many of the changes described above (e.g., wildfire risk and thawing permafrost) also affect larger communities and cities along the road network. However, urban areas are buffered by alternative income sources (jobs) and transportation options (roads) that reduce vulnerability. Rural-to-urban migration links villages with cities, putting pressure on public services (especially schools) in the cities but extends social networks of villages to tap urban employment opportunities (Kofinas et al. 2010).

Conclusions, uncertainties, and policy options

Although the Alaskan boreal forest and its people have been quite resilient to past warming, recent changes in structure and functioning suggest that the limits of this resilience have been approached and in some cases exceeded (Fig. 2). Disturbances (permafrost thaw, wildfire, insect outbreaks, disease, and drying of lakes and streams) are more extensive than at any time in the historical record, and the stabilizing (negative) feedbacks that previously constrained the magnitude of these changes or fostered recovery have been substantially weakened by warming. This suggests that the Alaskan boreal forest is on the cusp of potentially large nonlinear changes in structure and functioning. The major uncertainties concern the rates of change in disturbance and the extent to which these changes might be compensated at regional scales by redistribution of stand types across the landscape or by policies affecting rural–urban interactions. Current evidence from long-term studies at the Bonanza Creek LTER suggests the following.

- Permafrost will remain relatively resilient to continued warming except in high-ice-content lowlands and in areas burned by severe wildfires. The greatest sources of uncertainty are changes in snow cover, which will influence the rate at which these changes occur, and the extensiveness of severe wildfires and wetland formation that will likely trigger permafrost loss. In lowland areas, permafrost degradation causes a radical shift from black spruce ecosystems to *Sphagnum* bogs and herbaceous fens, whereas loss of permafrost and increased drainage in upland areas promotes replacement of black spruce by deciduous forests.
- New successional trajectories contribute to the variability of floodplain forests, with the long-term functional conse-

quences (e.g., nitrogen accumulation and suitability for wildlife) being uncertain.

- Increasingly extensive and severe wildfires are triggering a landscape transformation with potential expansion of deciduous forests from uplands to lowlands and north-facing slopes and perhaps the development of shrublands, grasslands, agriculture, or other novel ecosystem types in south-facing uplands and in lowlands. The future nature of upland ecosystems and the rates of transformation of lowland ecosystems are highly uncertain. These transitions and associated permafrost change will likely dominate changes in regional hydrology.
- The current mammalian fauna of interior Alaska will likely persist but reorganize into new patterns of distribution and abundance in response to both direct effects of climate and changes in habitat. The low degree of habitat fragmentation by human activities in Alaska provides greater opportunities for species migration and community reorganization in Alaska (and therefore regional resilience) than in most regions of the world. The net effect of these changes on subsistence opportunities for rural indigenous communities will depend on the rates and patterns of change, both of which are uncertain. Over the long term, resilience of both the animal and human communities may depend on policy decisions about the management of wildfire and human disturbances and the allocation of hunting opportunities between rural and urban hunters.

Global anthropogenic carbon emissions commit the planet to centuries of elevated CO_2 concentration in the atmosphere (Solomon et al. 2009). Actions taken globally to reduce emissions will influence the future magnitude and rate of change but are unlikely to prevent a continued warming trend (Fig. 1). Given the high certainty of continued warming in the Alaskan boreal forest, there are few policy options available at local-to-regional scales that could prevent the changes we have projected. However, local actions can influence the impacts of warming on ecosystems and society. For example,

- Agricultural and resource management policies will likely affect land-cover type, landscape connectivity, and species migration, and therefore, the trajectory of ecosystem change, particularly the rate of tree migration from south-facing uplands to lowlands and north-facing uplands.
- Fire management adjacent to human communities may influence the size, frequency, and configuration of future fires and therefore fire risk to these communities. In more remote areas, climate-driven changes in flammability will likely be more important than fire management in driving changes in fire regime and landscape heterogeneity.
- Development and resource management policies will influence the human exploitation of renewable and nonrenewable resources, the food and energy security of rural communities, and therefore, the integrity of the boreal forest as a coupled social–ecological system.

Continued research by the Bonanza Creek LTER has the opportunity to inform these policy choices and therefore the regional resilience of the Alaskan boreal forest.

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Natural Hazards and Earth System Sciences

Rapid movement of frozen debris-lobes: implications for permafrost degradation and slope instability in the south-central Brooks Range, Alaska

R. P. Daanen[1], **G. Grosse**[2], **M. M. Darrow**[3], **T. D. Hamilton**[4,*], **and B. M. Jones**[4]

[1]Water and Environmental Research Center, Institute of Northern Engineering, University of Alaska Fairbanks, Fairbanks, Alaska, USA
[2]Geophysical Institute, University of Alaska Fairbank, Fairbanks Alaska, USA
[3]Mining and Geological Engineering Department, Institute of Northern Engineering, University of Alaska Fairbanks, Fairbanks, Alaska, USA
[4]US Geological Survey, Alaska Science Center, Anchorage, Alaska, USA
[*]Emeritus

Correspondence to: R. P. Daanen (rdaanen@alaska.edu)

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Abstract. We present the results of a reconnaissance investigation of unusual debris mass-movement features on permafrost slopes that pose a potential infrastructure hazard in the south-central Brooks Range, Alaska. For the purpose of this paper, we describe these features as frozen debris-lobes. We focus on the characterisation of frozen debris-lobes as indicators of various movement processes using ground-based surveys, remote sensing, field and laboratory measurements, and time-lapse observations of frozen debris-lobe systems along the Dalton Highway. Currently, some frozen debris-lobes exceed 100 m in width, 20 m in height and 1000 m in length. Our results indicate that frozen debris-lobes have responded to climate change by becoming increasingly active during the last decades, resulting in rapid downslope movement. Movement indicators observed in the field include toppling trees, slumps and scarps, detachment slides, striation marks on frozen sediment slabs, recently buried trees and other vegetation, mudflows, and large cracks in the lobe surface. The type and diversity of observed indicators suggest that the lobes likely consist of a frozen debris core, are subject to creep, and seasonally unfrozen surface sediment is transported in warm seasons by creep, slumping, viscous flow, blockfall and leaching of fines, and in cold seasons by creep and sliding of frozen sediment slabs. Ground-based measurements on one frozen debris-lobe over three years (2008–2010) revealed average movement rates of approximately 1 cm day^{-1}, which is substantially larger than rates measured in historic aerial photography from the 1950s to 1980s. We discuss how climate change may further influence frozen debris-lobe dynamics, potentially accelerating their movement. We highlight the potential direct hazard that one of the studied frozen debris-lobes may pose in the coming years and decades to the nearby Trans Alaska Pipeline System and the Dalton Highway, the main artery for transportation between Interior Alaska and the North Slope.

1 Introduction

Climate change currently underway in the Arctic (ACIA, 2004; IPCC, 2007) is strongly affecting cryosphere dynamics and distribution, including warming and degradation of permafrost and reduction of its areal extent. Major impacts are observed in thermal conditions of near-surface permafrost and active layer (Romanovsky et al., 2002, 2007, 2010; Nolan et al., 2005; Kääb et al., 2007a; Jorgenson et al., 2006; Smith et al., 2010). Mountain permafrost is warming and retreating to higher elevations in response to climate change, and a similar response is recorded by diminishing glaciers (Nolan et al., 2005; Kääb et al., 2007a; Marchenko et al., 2007; Harris et al., 2009).

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To determine rates of permafrost change and resulting landscape dynamics, permafrost research over recent decades has increasingly focused on processes related to both the vertical and lateral movement of the ground as a result of transient temperature dynamics and phase change (Haeberli et al., 2006). Liquid water expansion upon freezing causes soil volume to increase and the soil surface to rise, whereas warming of the ground ice and phase change from a solid to liquid state causes loss of soil volume, structure and strength, resulting in greater susceptibility to erosion and mass wasting during thawing. The impact of soil strength reduction due to permafrost thawing is increased in mountainous settings where unconsolidated sediments exist, whereby slope destabilization may result in gravity-driven catastrophic downwasting (Kääb et al., 2005). Mass wasting features on permafrost-stabilized slopes have been researched extensively in many mountain regions (Wahrhaftig and Cox, 1959; Humlum, 1998b; Gorbunov and Seversky, 1999; Matsuoka et al., 2005; French, 2007; Gruber and Haeberli, 2007; Kääb et al., 2007b; Ikeda et al., 2008), allowing for the description and categorisation of many types of features with a unique set of characteristics. All have gravity in common as their driving force and many also share types of movement processes, such as sliding, flowing and creeping, typical for periglacial mountain regions (French, 2007).

In the European Alps, research has shown that degrading permafrost in bedrock regions causes hazards on steep slopes, in particular rock falls and landslides (Gruber and Haeberli, 2007). The number of rock falls and landslides greatly increases with increasing ground temperature, the loss of ice cement in bedrock cracks and pore space of rock debris, and the number of freeze-thaw transitions deeper in the rock face as mean annual ground temperatures approach the melting point of water. Higher substrate temperatures and liquid water released from ground ice results in decreased ground viscosity and slope stability. Slope stability is also a concern for permafrost-affected soils (Gude and Barsch, 2005; Lewkowicz and Harris, 2005; Harris et al., 2008b). Although ground ice slows the downslope movement of debris (Swanger and Marchant, 2007), warming atmospheric conditions can slowly degrade the ground ice, increasing the unfrozen water content in the ground, and releasing meltwater for increased pore water pressure and destabilizing slopes (Geertsema et al., 2006; Kääb et al., 2007b; Delaloye et al., 2008; Harris et al., 2008b; Ikeda et al., 2008; Roer et al., 2008; Lambiel et al., 2008).

Typical and dynamic permafrost-related mass wasting features on mountain slopes are rock glaciers, which are common in many cold climate regions (Barsch, 1977; Kääb et al., 1997; Humlum, 1998a, b; Isaksen et al., 2000; Berthling et al., 2003; Haeberli et al., 2006; Farbrot et al., 2007; Ballantyne et al., 2009; Brenning and Azocar, 2010). Rock glaciers generally consist of blocky debris (Wahrhaftig and Cox, 1959) with interstitial ice, although abundant fines have also been observed in these features (Ikeda and Matsuoka, 2006). A model for grain size distribution in rock glaciers is provided by Haeberli et al. (1998). The supply of debris to these features seems to control most of the movement on the order of centimetres to metres per year (Degenhardt, 2009), but other mechanisms such as warming result in rock glacier thinning are also described (Krainer and He, 2006; Roer et al., 2008; Ikeda et al., 2002).

Rock glaciers were recognised in Alaskan mountain ranges as early as the 1950s (Wahrhaftig and Cox, 1959), yet they remain understudied with respect to other mountain permafrost regions, likely due to limited access to the remote mountain areas and limited hazards to infrastructure. Wahrhaftig and Cox (1959) studied approximately 200 rock glaciers in the Alaska Range. Studies of the Fireweed rock glacier in the Wrangell Mountains have shown a movement rate of 3.5 m per year (Bucki et al., 2004; Bucki and Echelmeyer, 2004). In the Kigluaik Mountains on the Seward Peninsula, Calkin et al. (1998) report the presence of ten active tongue-shaped rock glaciers and, based on lichenometry, suggest that some of them formed about 3000–4000 yr ago. In the Brooks Range, Calkin et al. (1987) found similar ages using lichen indicators, but they suggest that the initiation time coincided with the retreat of the Pleistocene glaciers in the region.

Although our frozen debris lobes resemble rock glaciers, in some respects, they differ in source area, composition and mechanism and rate of movement, as discussed below. We focus our study on permafrost-regulated, rapidly moving, elongate, partially frozen debris-lobes found on mountain slopes in the Brooks Range of Alaska. Although initially identified as active (Hamilton, 1978a, 1979b, 1981), others described these features as mostly inactive rock glaciers (Kreig and Reger, 1982; Brown and Kreig, 1983) and they have never been investigated in great detail. In this paper, we show for the first time that these features exhibit rapid movement, possibly linked to climate change and permafrost degradation, and may pose a hazard to nearby transportation infrastructure. The objectives of this paper are to characterise various processes that are typical for these landforms and assess the frozen debris-lobe dynamics over the last 60 yr, using (a) ground-based topographic surveys; (b) remote sensing-based analysis of lateral movements; (c) permafrost and active layer temperature measurements; (d) sedimentological and geotechnical analysis; and (e) tree ring analysis. We also describe the placement of frozen debris-lobes within the continuum of mass movement features. One particular frozen debris-lobe front is now very close (<70 m) to the Dalton Highway corridor; this individual feature poses a substantial hazard to the transportation infrastructure between Interior Alaska and hydrocarbon exploration and production centres of northern Alaska within the next few decades at current movement rates.

2 Study area

The frozen debris-lobes we studied are located in the south-central Brooks Range of Alaska, USA. Based on historical repeat photography and lichenometry, Ellis and Calkin (1979) suggest that the most rapid retreat of alpine glaciers in the area occurred after ca. 1870, followed by a deceleration after the mid-1900s (Ellis and Calkin, 1984). Lake varves, in a glacially fed lake in the central Brooks Range, indicates that this portion of the Brooks Range has experienced atmospheric warming of 3.7 to 5.0 °C during the summer months since the Little Ice Age as suggested by Bird et al. (2009) and by Evison et al. (1996), who studied glacial retreat dynamics. In a review paper, Molnia (2007) confirmed this overall pattern of glacial retreat since the late 19th century for the Brooks Range, which is in line with that of other Alaskan mountain ranges. Nolan et al. (2005) provide an overview of glacier thinning over the period from 1956 to 2003 for the McCall glacier northeast of our research area in the Brooks Range. They found an increase in thinning from 0.35 m yr^{-1} before 1993 to 0.47 m yr^{-1} thereafter.

Ellis and Calkin (1984) also suggest that rock glaciers, now present in cirques of the central Brooks Range, were probably initiated by increased mass wasting from over-steepened valleys and cirque walls after late glacial deglaciation. Ellis and Calkin (1979) report that most of the active rock glaciers are found north of the continental divide and at altitudes higher than 1350 m. a.s.l., with lower limits for tongue-shaped rock glacier snouts at 1200 m. a.s.l. for inactive forms and at 1300 m. a.s.l. for active forms. Hamilton (1978a, b; 1979a, b; 1980, 1981) mapped the surficial geology of the central Brooks Range. He recognised and mapped a variety of mass wasting features in the region including rock glaciers, talus cones, rock slides, slush flow deposits as well as open system pingos, other frost mounds and aufeis, indicating groundwater flow. Using aerial photography and helicopter traverses, Hamilton was the first to map the location and widespread occurrence of lobate mass wasting features in the central Brooks Range that are distinct from the previously described rock glaciers, due to their material content, location, shape and vegetation cover. In his maps he named them flow slides.

Frozen debris-lobes FDL-A, -B and -C examined in this paper are located in the south-trending valley of the Dietrich River (Fig. 1), a short distance south of the Continental Divide. They are located near the Dalton Highway transportation corridor (see Fig. 1, lower inset), about 65 km north of Coldfoot and 170 km south of Deadhorse. Most of our detailed observations are from a frozen debris-lobe (site FDL-A) located at 67°48.669′N/149°49.185′W, which is also the frozen debris-lobe closest to the Dalton Highway. Based on a 0.5-m resolution WorldView-1 satellite image from 24 August 2008 the lobe front was approximately 75 m away from the Dalton Highway and 310 m from the Trans Alaska Pipeline System (TAPS), which is buried in alluvial sediment in this section. Paralleling the highway is the Dietrich River, which provides the main drainage for the region.

The study area is mapped as continuous permafrost (Jorgenson et al., 2008), however, permafrost in the valleys of the southern Brooks Range is relatively warm (−1.0 °C at 24 m depth; measured in 2010) (CADIS, 2010) and shallow. Winter snow pack insulation causes warmer subsurface conditions relative to the air temperature. Combined, these characteristics result in permafrost that is highly susceptible to degradation and active layer deepening. Slope and aspect also contribute to the difference between warmer and cooler sites, with north facing slopes being predominantly cooler; this is evident from the altitudinal tree line, which varies strongly between slope aspects in this region. In addition, vegetation and presence or absence of soil organic layers is highly variable in the study area and strongly impacts the soil thermal regime. Deep borehole measurements indicate that the permafrost is generally warming in northern Alaska (Smith et al., 2010). Chandalar Shelf borehole temperatures, 30 km north of our research site, warmed about 0.04 °C yr^{-1} over the last two decades, which is similar to the Coldfoot deep borehole at 24 m depth.

3 Methods

3.1 Field observations

For this reconnaissance study, we collected data from frozen debris-lobes (FDL-A, -B, and -C) (see Fig. 1), with the majority of measurements taken on FDL-A. Field observations included an assessment of morphometric parameters such as overall shape, elevation, and slope and aspect configuration; identification and characterisation of surface features such as scarps, crevasses and vegetation cover; and identification of clear indicators for active movement such as toppling trees, split tree trunks, trees overrun and partially buried by the frozen debris-lobe, over-steepening of slopes, block fall of frozen, but otherwise unconsolidated fine-grained sediment, scratch marks on frozen blocks from sliding, and rippling and uplifting of frozen topsoil in front of the feature due to compression from the pressure of advancing debris.

Our field observations span multiple seasons (late winter: April; late summer: August) and the years 2008–2010. In the spring of 2008, we installed five metal marker pins near the terminus of FDL-A to measure movement rates with a Differential Global Positioning System (DGPS). During the summers 2008 and 2009, we repeated these location measurements to assess local movement rates. We also collected DGPS data in transects across FDL-A and along its boundaries, including the terminus and the headwall (Fig. 2b).

A time-lapse camera was installed to take photographs of the terminus of FDL-A at one-hour intervals throughout the 2009 summer. Ground temperature data loggers were

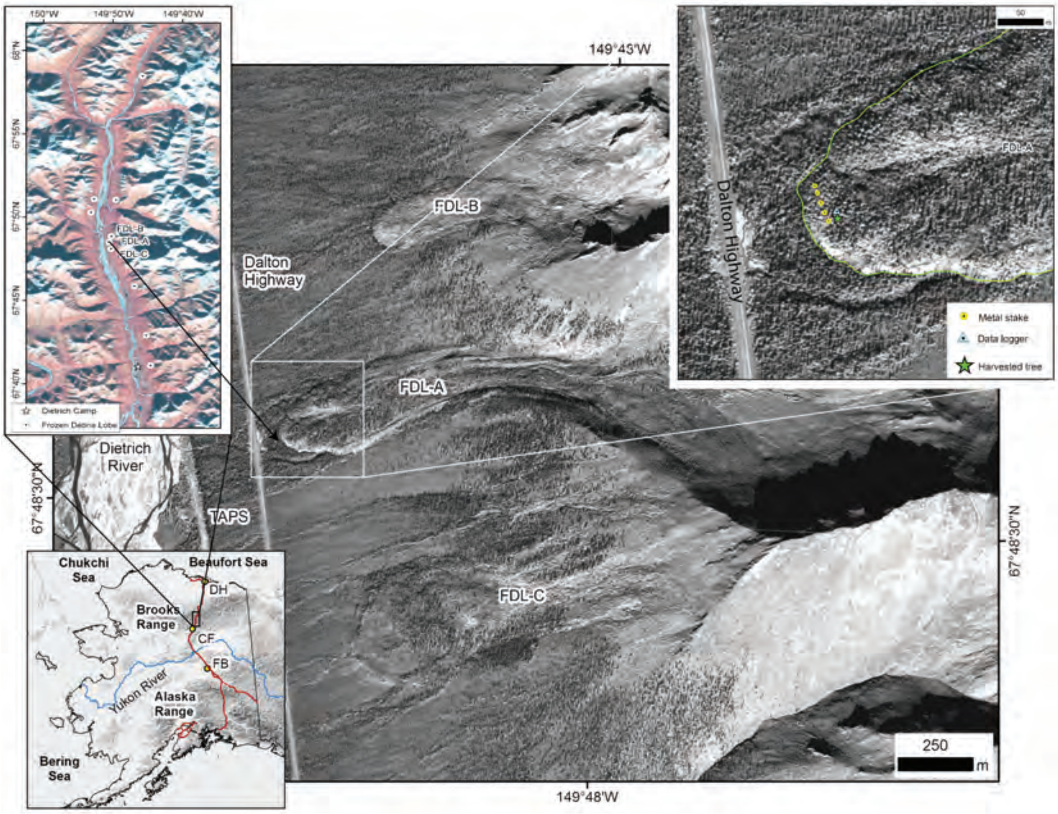

Fig. 1. Location of the study area and frozen debris-lobes (FDL) A, B and C (Image from 2008 WorldView-1, © Digital Globe). The Dietrich River, the TransAlaska Pipeline System (TAPS) and the Dalton Highway are shown to the west of the studied FDLs. Lower left inset indicates the study area (black rectangle) within the Brooks Range; FB is Fairbanks, CF is Coldfoot and DH is Deadhorse at the northern end of the Alaska Highway system (maroon line). Upper left inset shows the distribution of frozen debris-lobes along the Dalton Highway and this map is based on a Landsat-5 TM image (bands 4-3-2) from 19 September 2010 (USGS/NASA). Upper right inset shows the terminus of FDL-A with measurement locations.

installed to measure the air and surface temperatures, as well as the ground temperatures at depths of 0.50 m and 1.80 m on top of the feature near the terminus (near marker pin 1 (MP-1); Fig. 1) and additionally in front of the terminus. We also collected two soil samples from the upper 1 m of FDL-A for grain size analysis and determination of soil plasticity, and harvested a living tree that grew on the frozen debris-lobe to analyse tree ring growth.

3.2 Remote-sensing and terrain analysis

We used high-resolution aerial and satellite data from 1955, 1979 and 2008 to delineate the frontal terminus and sides of the three frozen debris-lobes in the study area (Table 1). All images were co-registered to the 2008 WorldView image and mapping was done in a desktop GIS environment. Additionally, we used a 5-m horizontal and 0.1-m vertical resolution, airborne interferometric synthetic aperture radar (IfSAR) derived digital elevation model (DEM) to assess the three-dimensional morphometry of the three features in our study area. The DEM was post-processed by removing all sinks and smoothing it with a 3×3 low pass filter. Once complete, we derived secondary parameters, including slope, aspect, planimetric and profile curvature, watershed size, and flow accumulation of the study area (Figs. 2 and 3).

4 Results

4.1 Field observations

The frozen debris lobes flow from mountain slopes (see Fig. 4), and consists of a mixture of coarse- to fine-grained debris (including woody debris from shrubs and trees where the frozen debris-lobes extend below forest limits). We also observed massive and interstitial ice within the frozen debris-lobes. Most of these features are covered with stands of drunken spruce trees (see Fig. 4c), which alerted us to their movement and prompted our investigation into these features. Trees growing on the slopes of the frozen debris-lobes experience great horizontal stress in the root zone due to ground movements, which can result in splitting of live

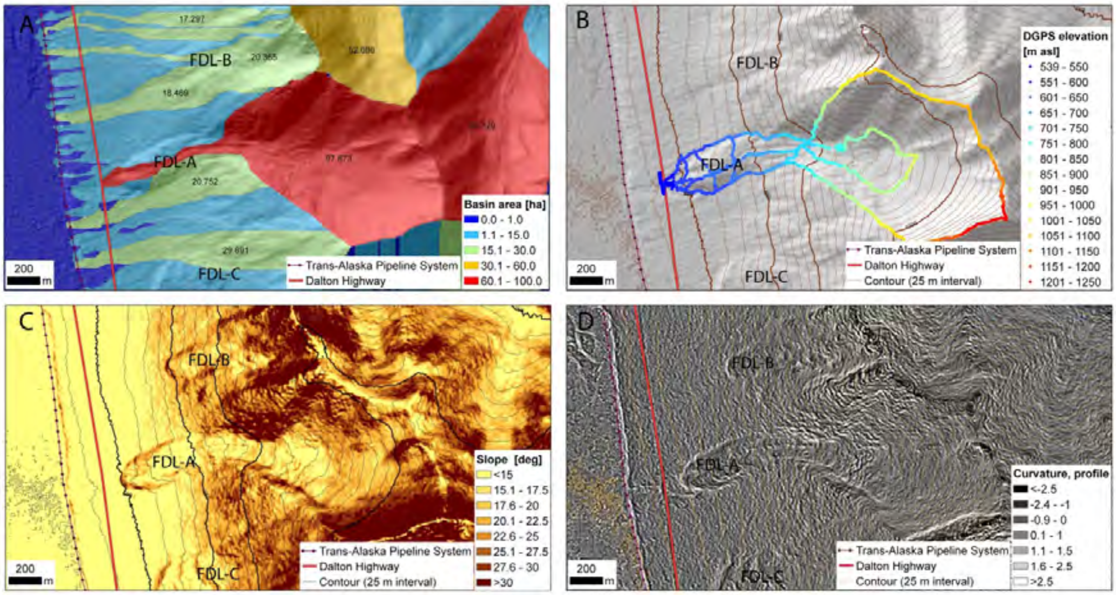

Fig. 2. IfSAR DEM-derived morphometric characteristics of frozen debris-lobe A (FDL-A). **(a)** Watershed area (ha) contributing to FDL-A (ca. 98 hectare); **(b)** shaded relief map with contour lines and an overlay of DGPS data, observed points are forming a line; **(c)** slope map; **(d)** curvature map.

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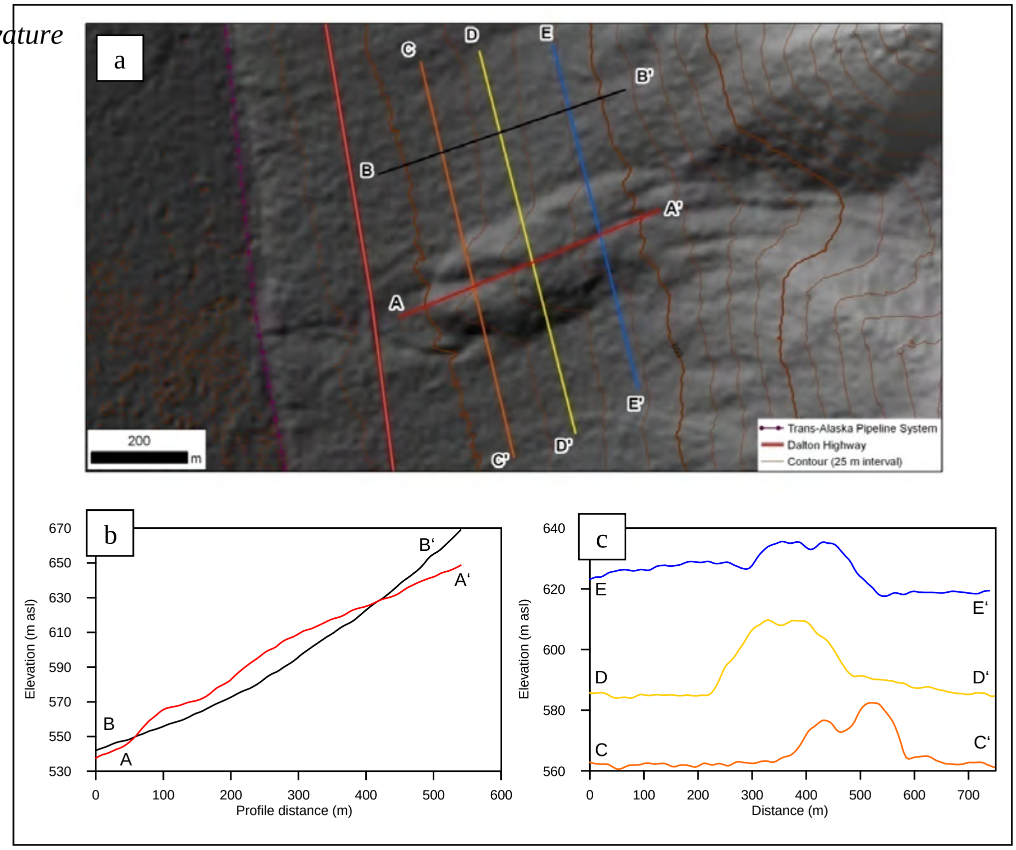

Fig. 3. Surface morphometry of frozen debris-lobe A (FDL-A). **(a)** Location of cross sections and long profiles; **(b)** long profiles, over FDL-A (A–A′) and next to FDL-A (B–B′); **(c)** three cross sections over FDL-A (C–C′, D–D′, and E–E′).

Figure 3. Surface morphometry of frozen debris-lobe A (FDL-A). a) Location of cross sections and long profiles; b) long profiles, over FDL-A (A-A'), and next to FDL-A (B-B'); c) three cross sectionsover FDL-A (C-C', D-D', and E-E').

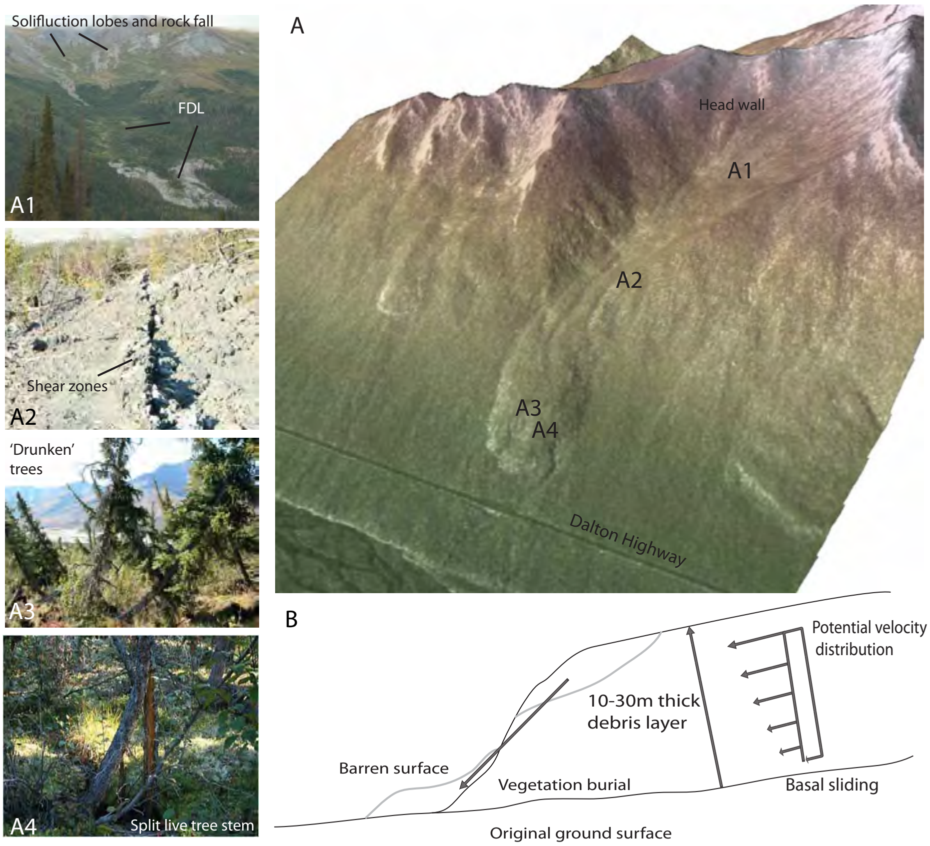

Fig. 4. Schematic overview of some observed processes on a frozen debris-lobe. (A1) Photograph of a frozen debris-lobe with upper zone of solifluction lobes, image shows FDL across the valley from FDL-A; (A2) shear zones in the middle section of FDL-A; (A3) "drunken" trees visible on FDL-A; (A4) split tree trunk on FDL-A; (B) profile of frozen debris-lobe front and associated movement; the flow vectors are assumed and parallel to the flow direction; background elevation model of FDL-A with upper contributing zone based on multiple DGPS surveys.

Table 1. Remote-sensing data used for mapping frozen debris-lobes.

Image type (ID)	Acquisition date	Ground resolution	Distance of FDL-A lobe front to Dalton Highway(m)*	Period (yr)	Front advance rate (m yr^{-1})
Aerial (AB1TAL00000009R_15_a)	22 July 1955	1:60 000	220 m	–	–
Aerial (AR5790027841791)	1 July 1979	1:64 000	170 m	24	2.1
WorldView-1 (AUG08WV010000008 AUG24220130_P1BS_052031425010_15_P003)	24 August 2008	0.5 m	70 m	29	3.4

* Distance was measured between the toe of the front lobe and the highway embankment toe; for the 1955 image, distance was measured to the location where the highway is currently located.

tree trunks (see Fig. 4d). Frozen debris-lobes are generally 500- to 2000-m long and 50- to 500-m wide, with a riser height (elevation above surrounding soil) between 5 and 30 m. Frozen debris-lobes move down slope from small drainage basins, typically less than 1 km^2 (Fig. 2a). Frozen debris-lobes are often located in small valleys; these valleys may or may not have a headwall providing a source of debris due to intense frost weathering. Often, solifluction lobes can be found at higher elevations just above these features, providing an additional material source. In larger basins frozen debris-lobes are absent and debris accumulates as alluvial fans.

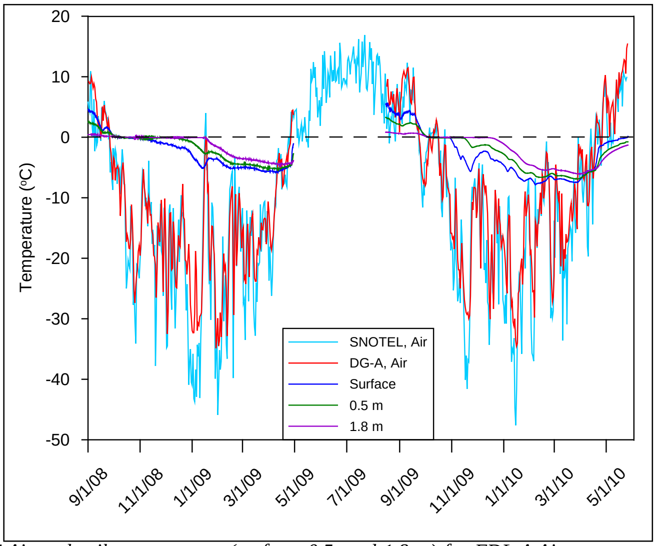

Figure 5.Air and soil temperatures (surface, 0.5, and 1.8 m) for FDL-A.Air temperatures from the SNOTEL weather station in Coldfoot are included for comparison.

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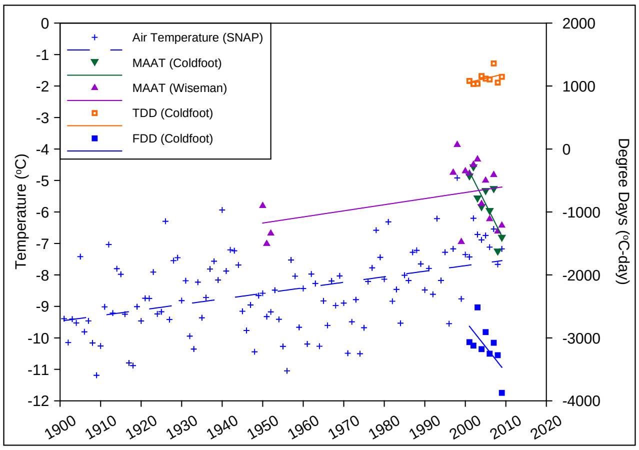

Figure 6.Mean annual air temperature near FDL-A.Reanalysis data is from SNAP; measured data was taken from the SNOTEL site near Coldfoot (60 km south of FDL-A) and near Wiseman (50 km south of FDL-A) (National Water and Climate Center, part of the Natural Resources Conservation service); thawing and freezing degree days are for Coldfoot.

Fig. 6. Mean annual air temperature near FDL-A. Reanalysis data is from SNAP; measured data was taken from the SNOTEL site near Coldfoot (60 km south of FDL-A) and near Wiseman (50 km south of FDL-A) (National Water and Climate Center, part of the Natural Resources Conservation service); thawing and freezing degree days are for Coldfoot.

Movement of frozen debris-lobes can be observed in the form of cracks in the surface of these features. These cracks are up to 25 cm wide, several metres deep and stretch over tens of metres laterally, and are often responsible for split tree trunks. Along the terminus of these features, trees are commonly overrun as the frozen debris-lobe advances. We observed live trees being pushed over and partially covered by flowing mud (summer) and by frozen debris slabs (winter). Frozen debris slabs slide over weak zones, and are observed during spring protruding out from the surface near steep slopes. In addition, frozen soil and debris slabs buckle in front of the frozen debris-lobes, forming cavities that collapse upon spring thaw. This formation seems to be the result of the feature pushing on the frozen soils in front of the lobe.

Active layer detachment slides and retrogressive thaw slumps are also observed on these features. Typically these features mobilize the sediments and accelerate the alluviation processes. Mobilization of the rocks was also observed from a steep section of the lobe terminus.

4.2 Ground thermal regime

To understand changes in ground temperature, we collected both ground and air temperatures on FDL-A (Fig. 5). The 2008–2009 mean annual air temperature is −5.1 °C and the mean annual ground temperature at a depth of 0.5 m in mineral soil is −0.3 °C. We compared these data to those from the Coldfoot Snow Telemetry (SNOTEL) site, which correlates strongly with a 2008–2009 mean annual air temperature of −7.7 °C. The difference of 2.6 °C is most likely due to atmospheric inversion with FDL-A 200 metres elevated over the valley floor where the SNOTEL site is located. Long-term changes in ground temperature are mostly a function of changes in mean annual air temperature. We compared the short record for mean annual air temperature in the region to climate reanalysis developed by the Scenarios Network for Alaska Planning (SNAP) (Fig. 6). This dataset was downscaled from multiple global circulation models (http://www.snap.uaf.edu). The short-term trend of decreasing mean annual air temperature measured near Coldfoot, Alaska appears contradictory to the regional warming trend in the long-term record and permafrost temperatures (Fig. 6). However, the trend in thawing degree days since 2001 indicates a warming trend for the region in summer, in spite of relatively cold winters.

Although the soil temperature dataset from FDL-A only spans two years, it provides some insights into the ground thermal regime. Soil temperatures collected on FDL-A indicate an active layer at least 1.8 m deep (see Fig. 5). Movement of FDL's causes debris exposure to the atmosphere, whereas nearby stable areas typically bear continuous cover of moss and tussocks, resulting in colder soil conditions. Additionally, the surface on these features is also much better drained, reducing the thermal offset.

4.3 DGPS surveys

The results of the marker pin survey on FDL-A with a DGPS demonstrated 1.3 m of movement during four months between April and August 2008, which indicates an average rate slightly greater than 1 cm day^{-1}. Repeat measurements in 2009 also showed a similar daily movement rate.

A survey of the entire feature during 2009 is presented in Fig. 2b. The DGPS was carried in a backpack while climbing over the feature. These data provide a snapshot of the geometry that can be compared with future datasets.

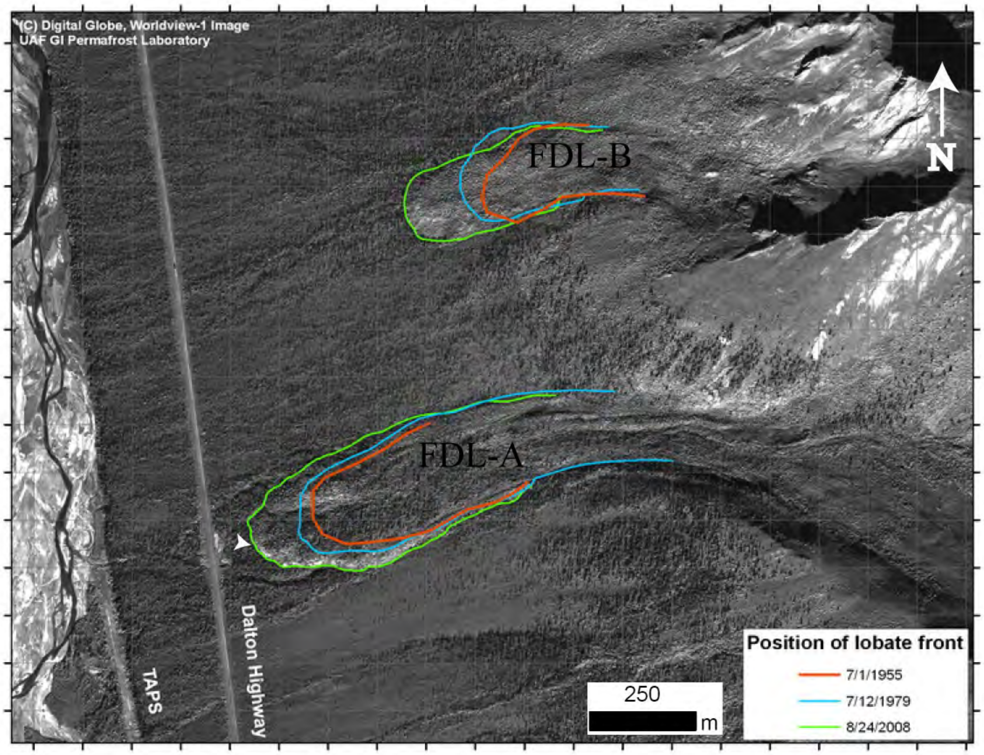

Fig. 7. Comparison of the termini positions of FDL-A and FDL-B between 1955 and 2008 from high-resolution aerial and satellite based remotely sensed imagery. Two drainage channels in front of the lobe drain FDL-A and are not part of the moving body. The white arrow near the terminus of FDL-A indicates the repeat photo camera location illustrated in Fig. 10.

Figure 7. Comparison of the termini positions of FDL-A and FDL-B between 1955 and 2008 from high-resolution aerial and satellite based remotely sensed imagery. Two drainage channels in front of the lobe drain FDL-A and are not part of the moving body. The white arrow near the terminus of FDL-A indicates the repeat photo camera location illustrated in figure 10.

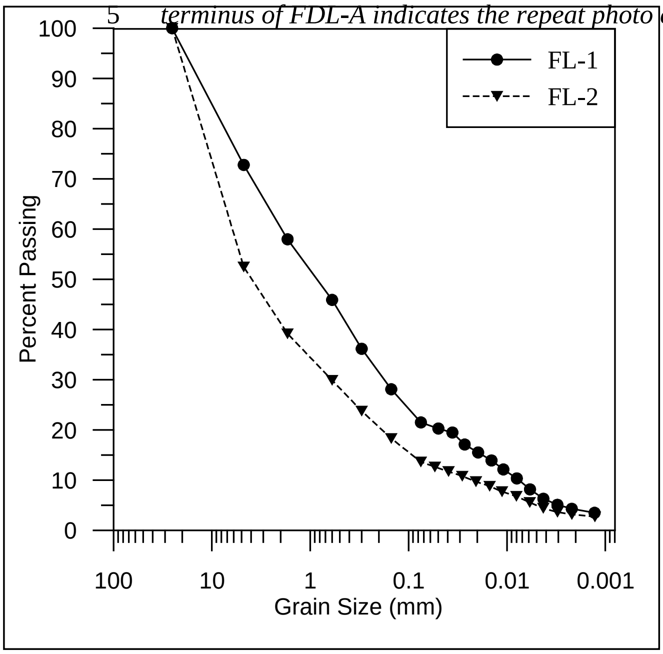

Fig. 8. Grain size distribution of two sediment samples taken from the top portion of the front lobe of FDL-A.

Grain size distribution of two sediment samples taken from the top portion of the front ... DL-A.

4.4 Remote-sensing and terrain analysis

FDL-A is elongate, varies in width from 140 to 260 m and has a length of 1200 m beyond its source area (Fig. 3). Based on the IfSAR DEM, the height of the front of this frozen debris-lobe above the surrounding area is 20 m, rising to 25 m approximately 275 m behind (upslope) from the lobe front, and decreasing again to 10 to 18 m approximately 500 m behind the lobe front. This indicates that debris movement is episodic or pulse like, with periods of faster movement alternating with periods of stagnation and buildup. Such movement pulses likely are controlled by slope, ground temperature, liquid water content and debris supply. The slope of the surrounding area north and south of the frozen debris-lobe front ranges from 5° to 17.5° (Fig. 3). While the surface of the frozen debris-lobe has roughly the same slope along its long axis, multiple areas along the lobe front as well as its sides have slopes exceeding 30°, which agrees with field observations of steep and several metre high scarps. The entire watershed contributing to FDL-A is about 98 ha, one of the largest watersheds on this mountain slope (Fig. 2a). The lobe front, with its approximately 20-m height and 170-m width, advances a debris volume of about 34 m^3 day^{-1} (12 410 $m^3 yr^{-1}$) at the current movement rate of 1 cm day^{-1} observed with ground measurements between

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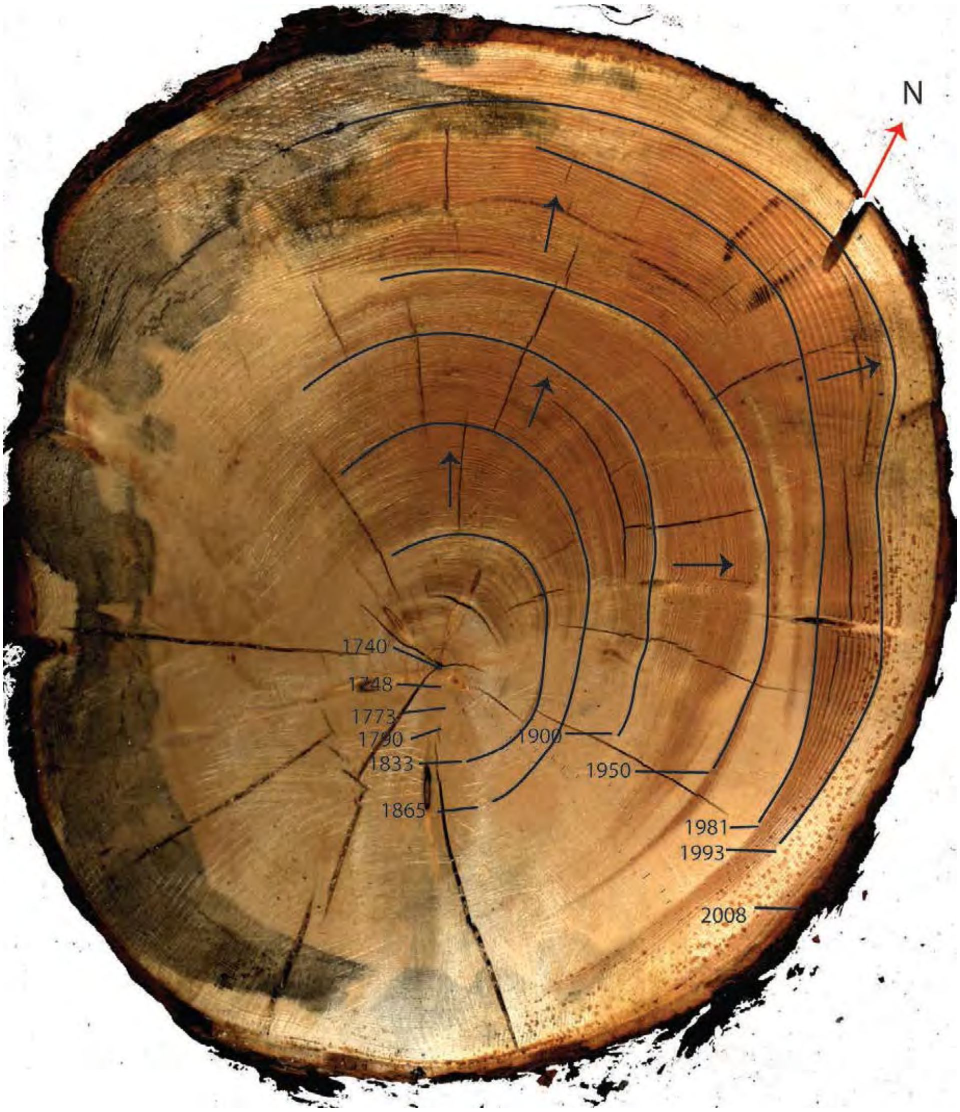

Figure 9. *Tree rings with compression wood direction shifts. The tree was harvested in 2008. Changes in the direction of compression wood growth (indicated by black arrows) occurred in the years indicated in the figure.*

Fig. 9. Tree rings with compression wood direction shifts. The tree was harvested in 2008. Changes in the direction of compression wood growth (indicated by black arrows) occurred in the years indicated in the figure.

April and August 2008. Debris advancing every year at the lobe front equals about 22×10^6 kg, or about 440 truckloads per year for a 50-ton dump truck.

Analysis of an optical image time series covering the period from 1955 to 2008 shows an increase in the rate of advance of the frontal lobe of FDL-A and FDL-B (Fig. 7). During the 1955–1979 period, FDL-A moved approximately 50 m, which indicates an average rate of 2.1 m yr^{-1}. During the 1979–2008 period, the terminus advanced 100 m, which indicates a higher average rate of 3.4 m yr^{-1}. The average movement rate for the entire observation record is 2.8 m yr^{-1}. The current (2009–2010) movement rate we measured in the field is just over 1 cm day^{-1} ($\sim$3.6 m yr^{-1}), which may indicate an ongoing increase in the average movement rate. In summary, the front of FDL-A advanced 150 m since 1955, is currently less than 70 m away from the Dalton Highway, has a base cross-sectional width of 170 m, an average height of 20 m, and an accelerating terminus movement rate of 1 cm day^{-1}.

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4.5 Debris composition

Based on the geological map of the Chandalar Quadrangle by Brosgé and Reiser (1964), the bedrock immediately adjacent to the frozen debris-lobes in this reconnaissance study consists of slate, phyllite, schist, phyllitic siltstone, schistose sandstone and limestone. The frozen debris-lobes contained clasts of these rock types embedded in a complex fine grained matrix with massive and interstitial ice. We took two

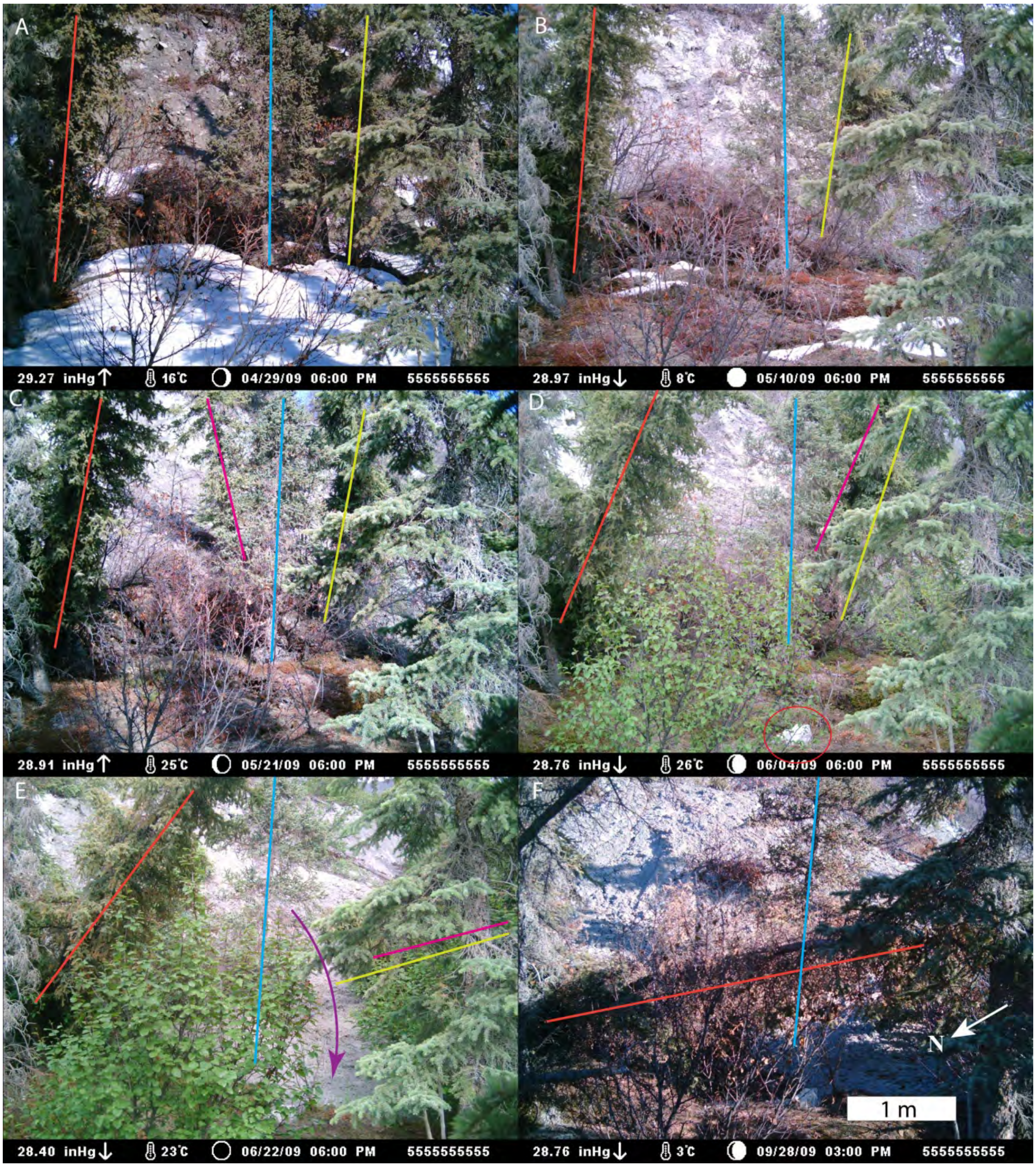

Fig. 10. Repeat photography of a portion of the FDL-A terminus. The steep face slumps down as it saturates during rain events. Straight lines follow positions of individual trees in the images (magenta tree fell from the top of the terminus); the red circle indicates the appearance of a rock from upslope; the purple arrow indicates mud flow from the steep slope. Photographs were selected from 29 April to 28 September 2009.

Figure 10. Repeat photography of a portion of the FDL-A terminus. The steep face slumps down as it saturates during rain events. Straight lines follow positions of individual trees in the images (magenta tree fell from the top of the terminus); the red circle indicates the appearance of a rock from upslope; the purple arrow indicates mud flow from the steep slope. Photographs were selected from April 29 to September 28, 2009.

surface sediment samples from FDL-A, which included fist-sized, tabular clasts of biotite garnet schist and quartz. Grain size analysis was performed for the sediment fraction smaller than 75 mm. The two samples were classified as silty sand with gravel and silty gravel with sand (GM), according to the Unified Soil Classification System (Fig. 8). The sediment from these samples had a specific gravity of 2.79, a liquid limit of 32 and no plastic limit.

4.6 Tree ring analysis

One unique aspect of the frozen debris-lobes in our study area is the presence of dense spruce and alder vegetation, due to their occurrence below the treeline. The tilted, or "drunken" trees (see Fig. 4c), may be a result of rapid movement and soil disturbance after a previous phase of stability that allowed tree growth, or a result of degrading permafrost and melting ground ice that also leads to surface disturbance

Figure 11. Striations on the underside of a block of frozen silty debris indicate sliding motion in the frontal lobe of FDL-A during and after soil freezing. Below is the original ground surface, littered with recent organic matter, and above is the seasonally frozen ground that slid over the original ground. Ice can be seen in cracks as well as streaks on the underside from the sliding motion.

Fig. 11. Striations on the underside of a block of frozen silty debris indicate sliding motion in the frontal lobe of FDL-A during and after soil freezing. Below is the original ground surface, littered with recent organic matter, and above is the seasonally frozen ground that slid over the original ground. Ice can be seen in cracks as well as streaks on the underside from the sliding motion.

and irregular subsidence. The trees respond by growing additional wood on the leaning side of the tree in an attempt to maintain a vertical growth position. Tree ring analysis of one drunken spruce tree growing on the surface of FDL-A showed evidence of distinct shifts in direction rather than a continuous change in direction (Fig. 9). The tree rings indicated that this tree formed compression wood (i.e., larger tree rings on one side of the tree trunk) over periods ranging from one to four decades.

4.7 Characterisation of movement processes in frozen debris-lobes

The major types of movements observed at our studied frozen debris-lobes include: (1) permafrost creep during winter and summer, (2) basal sliding, and (3) several forms of active layer movement such as mudflows, detachment slides, gelifluction and sliding of frozen soil slabs during fall and early winter freezing. All forms of movement are driven by gravity and largely oriented in the same downslope direction. As a result of movement, deep crevasses or cracks occur within the frozen debris-lobe. These forms of movement also cause trees to lean and become buried under debris at the terminus, likely causing inclusion of considerable amounts of organic matter under and within the frozen debris-lobe.

Permafrost creep occurs in areas with steep gradients and is visualized in surface cracks and deformed trees (Fig. 2c). With the currently available data basal sliding was not directly observed, but indirect evidence suggests that it occurs. Organic soils, overrun by the frozen debris-lobe, could provide a weak layer that allows sliding as the weight of the debris pushes down the slope. Bucki and Echelmeyer (2004) identified such a sliding layer inside the Fireweed rock glacier. Other forms of sliding of the frozen debris in the active layer have been observed, such as in Fig. 11.

Downslope motion of seasonally frozen soil is common on steep terrain underlain by permafrost. A number of different movement types were observed in the field. During the summer of 2009, a steep portion of the terminus collapsed and caused a mudflow in front of the time-lapse camera. Beginning with the spring snowmelt, rapid movement of coarse debris and fine sediment occurs as mudflows from the terminus and the sides of the frozen debris-lobe. As the summer progresses and the active layer deepens, mudflows increase in size. Hourly time-lapse photographs revealed mudflows on FDL-A during or immediately after rain events, with rain water possibly contributing to the saturation of fine-grained sediments and leaching of fines from coarser debris (Fig. 10). From the images it is difficult to quantify rates of motion; however, the mudflow we measured exceeded 10 m in length in a single summer. The soil flows from the top of the steep terminus, likely driven by pore water pressure, to the base of the terminus where it accumulates. Pore water pressure builds as a result of snowmelt, rainfall and/or ground ice melt. Exposed frozen ground below the flowing soil contains both massive and interstitial ice, which may contribute to the mudflow as it melts during the summer. These mudflows cause erosion of frozen debris-lobes, and they add to alluviation along small creeks that flow into the Dietrich River and clog culverts underneath the Dalton Highway. At one location, 24 km south of the study area, an alluvial fan has recently formed directly downstream of a frozen debris-lobe. The alluvial fan is still mostly unvegetated, and may represent a case of frozen debris-lobe destabilization, further indicating the delicate balance between debris accumulation and debris erosion.

Detachment slides and retrogressive thaw slumps occur on steep parts of the sloping surfaces of frozen debris-lobes. Vegetation holds the upper weak layers together, where high liquid water content causes the upper layer to slide. At the upper end of the detachment slide a head wall remains that is sensitive to thaw slumping, because the ground ice is exposed to air temperatures. The overall flow of the frozen debris-lobe causes the slope to change locally due to variations in velocity. Dynamic accumulation and depletion of debris occurs along the slope as a result of overall frozen debris-lobe movement.

Gelifluction in the active layer occurs mostly at a higher elevation where solifluction lobes deliver debris to the top of the frozen debris-lobe. These solifluction lobes can be observed on high resolution satellite images and at the ground surface.

Finally, during the fall and winter, we observed scour marks resulting from sliding motion that likely occurred during seasonal freezing. Several metres of the steepest sections of FDL-A's terminus moved by sliding during the fall, with entire frozen soil slabs overtopping the original ground surface and the movement causing the formation of scour marks along the base of the frozen soil slabs (Fig. 11).

The rates of movement of other frozen debris-lobes in the region vary. Some features appear to be stable, based on straight trees and older shrubs growing on the surface. Other features are becoming progressively more barren, as trees and shrubs are toppled and buried below a thick layer of debris. Thus, our preliminary findings reveal that frozen debris-lobes are dynamic periglacial surface features that may exhibit local controls as well as regional responses to changes in climate and other types of disturbance.

5 Discussion

While frozen debris-lobes have been identified in the south-central Brooks Range of Alaska, they may occur elsewhere. A feature of similar size and altitude which is also covered with trees, some with split trunks, was described by Blumstengel and Harris (1988) in the St. Elias Range. Blumstengel and Harris (1988) observed that the lower lobe terminus of the St. Elias feature is in the riverbed of the Slims River. Thus, damming of rivers and subsequent flooding are additional potential hazards associated with these features when they terminate in a river or alluviate the river bed.

5.1 Movement of frozen debris-lobes

Frozen debris-lobes exhibit a variety of soil motion processes in a continuum from gelifluction to permafrost creep (Haeberli et al., 2006). Large cracks in the surface of FDL-A suggest internal stresses due to heterogeneous movement. These cracks may be the result of internal movement of the frozen core rather than surface movement within the active layer (Matsuoka and Humlum, 2003; French, 2007). Based on the measured flow rate of frozen debris-lobes, basal sliding also should be considered as one of the processes causing motion. If basal sliding occurs, internal stresses can be expected as the centre of the feature would move more rapidly compared to the longitudinal edges of the debris-lobe due to differences in friction. Observations of soil surface deformation and buckling of vegetation in front of the terminus may be evidence of sliding. Some sliding at the surface occurs during freezing when the frozen soil slides as a rigid body over the unfrozen soil of the active layer (Fig. 11). This shearing may result from increased pore water pressure within the unfrozen portion of the active layer. This pressure may build during the fall as the freezing front penetrates the surface and confines ground water flow. Should increasing temperatures result in the development of a talik within a frozen debris-lobe, sliding or deformation in the talik may become the dominant and potentially catastrophic type of movement, when large parts of the frozen debris-lobe become unstable and slide down the slope.

Permafrost creep in these features is the suggested cause for surface deformation. This creep is also referred to as interstitial ice/debris flow (Haeberli et al., 2006), indicating its dependence on the presence and amount of interstitial ground ice. If creep occurs throughout the thick layer of frozen debris, it results in faster movement in the upper surface of these features. This type of motion leads to steepening slopes near the terminus of the feature that will increase the potential for mudflows when the frozen debris on the steep slopes thaws in spring.

Mudflows occur on the steepest slopes (10–50°) of the frozen debris-lobe surfaces and take place when the soils become oversaturated (Harris et al., 2008a), mainly during rain events in summer. Rapid warming of the soil surface can also cause oversaturation through melting of the pore ice. Mudflows were recorded with time-lapse photography and are best characterised by flowing of saturated debris. Trees were observed to slide down the terminus slope within an hour. Shrubs and trees were subsequently buried as they accumulated at the foot of the terminus. The overall slope (5–10°) of the surrounding ground, including the slope of the ground in front of the frozen debris-lobe, does not sustain the movement of mud and debris, causing it to accumulate near the toe of the terminus. A creek draining the feature, however, continues to carry the sediment to lower elevations towards the Dalton Highway and the Dietrich River.

Table 2 contains a summary of the postulated controls on the formation and behaviour of frozen debris-lobes, which are similar to those presented by Matsuoka et al. (2005). Ground ice plays a key part in the motion of creeping permafrost. We postulate that ice formation due to liquid water infiltration through surface cracks may be a major driving force contributing to the dynamic nature of frozen debris-lobes. Underground ice forms or degrades with fluctuations in available energy sinks and sources. Climate is, therefore, a strong factor in the stability of frozen debris-lobes and will determine their shape and occurrence. During summer, a portion of this ground ice melts and, together with rain events, provides lubrication for soil movement through gelifluction (Lewkowicz and Harris, 2005; French, 2007; Harris et al., 2008b) and mudflows.

5.2 Climate change effects

Glacier dynamics have been directly linked to climate variation in the Brooks Range. Although very different in appearance and material, we mention them because of their sensitivity to regional changes in climate. Almost all glaciers in Alaska are losing mass due to climate warming (Calkin et al., 1998; Arendt et al., 2002, 2009; Nolan et al., 2005; Berthier et al., 2010). The movement rates of frozen debris-lobes are close to the movement rates reported for common glaciers in the Brooks Range (Nolan et al., 2005). Rock glaciers in contrast generally move at rates of millimetres to metres per year (Humlum, 1997; Roer et al., 2008; Haeberli et al., 2006; Krainer and He, 2006; Hausmann et al., 2007; Ikeda et al., 2008). Ikeda et al. (2008) suggested that climate change causes some rock glaciers to move at faster rates. Saturation of fine-grained sediments was identified as causing acceleration in rock glaciers (Roer et al., 2008; Kääb et al., 2007b; Ikeda et al., 2008; Riff et al., 2008). Warming summers with more potential for liquid water infiltration may have contributed to the acceleration observed in the European Alps. More precipitation throughout the year or an enhanced hydrological cycle is also projected for Interior Alaska, as a result of climate change (Huntington, 2006). We postulate that increased precipitation and, thus, runoff over and through a frozen debris-lobe and its watershed, may promote erosion and the formation of a talik within the frozen debris-lobe. This allows for increased instability of frozen debris-lobes during the entire year, including the winter, and could result in movement rates far exceeding the rate of 1 cm day^{-1} currently observed. This increased erosion rate may result in the release of massive amounts of sediment to areas in the valley below the frozen debris-lobe (i.e., the Dalton Highway and Dietrich River).

Tree-ring observations suggest that the movement rate of frozen debris-lobes is episodic, depending on long-term soil thermal behaviour, debris accumulation and short-term soil climate shifts. In particular inter-annual snow cover variation can play a major role in controlling those short-term variations due to the impact of snow cover on the ground thermal regime as an insulating layer.

The rate of movement of frozen debris-lobes in the southern Brooks Range has increased over the last 50 yr. Even though short-term records (decade), indicate a local cooling trend, the long-term records (century) indicate warming. Estimated climate effects on frozen debris-lobes are, at this point, mainly based on processes that we observed over a short time period. Ground temperature is expected to affect processes such as creep, gelifluction, sliding layers, talik formation and wetness. Warmer wetter ground is less viscous and greater pore pressure (enhanced by taliks) can reduce resistance between soil particles leading to potential sliding planes in the debris and deeper wetter active layers enhance frost action and gelifluction. Thus, these features may serve as an “early warning indicator” of general slope instability on hill slopes underlain by permafrost. However, longer time series and ground observations are necessary to better understand the relative importance of individual processes in FDL movement and vulnerability to acceleration.

5.3 Frozen debris-lobe hazard

Frozen debris-lobes constitute a potential hazard to the transportation corridor running through the Dietrich River valley. Large quantities of slope debris have accumulated over millennia in small contributing valleys. This debris has very likely remained frozen during the last few thousand years, because the region is underlain by continuous permafrost. We have found evidence that these features exhibit episodic movement and there is some evidence this movement is controlled by their internal thermal state.

Currently, the frozen debris-lobe closest to the Dalton Highway (FDL-A) poses the largest hazard to local infrastructure. Even though its terminus is currently located outside the highway right-of-way, the frozen debris-lobe is producing large amounts of sediment, which has already led to the burial of culvert inlets. At the very least, recent sediment alluviation associated with frozen debris-lobes is increasing maintenance costs for culverts along parts of the Dalton Highway system in the Brooks Range. In a worst case scenario, rapidly moving frozen debris-lobes could become a direct threat to the highway. Based on the observed

Table 2. Controls on frozen debris-lobe(FDL) dynamics.

Controls	Processes	Significance
Bedrock	Debris source; controls size, shape, and abundance of rock debris	Rock abundance controls FDL size; platy debris promotes internal sliding
Weathering	Produces fine-grained rock material (regolith) for incorporation into FDL	Fine-grained debris holds moisture, promoting ice-rich permafrost and internal FDL creep
Position on valley side	Catchments above limits of last glaciation are strongly weathered, with incorporation of eolian silt and organic soil material	Promotes nourishment of FDL and robust FDL movement
Catchment geometry	Shape, size, and slope of catchment controls amount and rate of debris fed into FDL	Affects size, shape, and flow rate of FDL
Climate and climate change	Affects permafrost stability, depth of active layer, and water influx from precipitation and snowmelt runoff	Influences growth of massive ground ice and interstitial ice; deeper active layer with warming climate promotes surface movements; warming permafrost promotes internal movements
Vegetation	Roots stabilize FDL surfaces; vegetation shades surfaces in summer	Inhibits surface erosion by preventing rain impact and runoff; vegetation canopy and ground cover maintain cooler surfaces and near-surface permafrost

rate of motion of FDL-A, the timing of this potential hazard will likely occur within the next 20 yr if the current rate is maintained. However, the question remains whether FDL-A's movement rate can further increase undercurrent or future climatic, permafrost and hydrological conditions to the very fast rates of comparable slope features observed in other regions of the world (Kääb et al., 2005). In this case, direct impacts on the Dalton Highway may be only years away.

Currently there are more than ten FDL's identified along the transportation corridor and hundreds in the south central Brooks Range (Hamilton and Labay, 2012). The pipeline is in many cases further removed from the direct impact of FDLs, because it is located closer to the centre of the valley and often buried in this region. More detailed monitoring of FDL's in the Brooks Range transportation corridor would provide very useful data and information to land managers, transportation planners and infrastructure engineers, allowing a better assessment of existing and potential hazards to a safe and continuous operation of the Dalton Highway, TAPS and any planned future infrastructure in this region.

6 Conclusions

Frozen debris-lobes are wide-spread on mountain slopes in the south-central Brooks Range. These features are active and can move at rates greater than 1 cm day^{-1}. Their movement is a combination of creep, sliding and flow. They consist of frozen and unfrozen fine- to coarse-grained debris, as well as organic fine and coarse woody matter.

The location of a frozen debris-lobe (FDL-A) within 70 m of the Dalton Highway and close to the Trans Alaska Pipeline System is a concern because of the potential amount of sediment and debris that may impede travel along the Dalton Highway as well as access to TAPS. If current movement rates of this frozen debris-lobe are maintained, the feature will reach and seriously disturb the only road to the Alaska North Slope oil and gas fields in about 20 yr.

An observed acceleration of movement over the last 30 yr is likely related to active layer deepening and permafrost warming as a result of changes in climate. Climate projections for Interior Alaska point to the increased warming, slope destabilization and, thus, hazard potential by these features over the coming years.

While providing us with insight into the overall characteristics and movement of frozen debris-lobes, the results of this reconnaissance-level investigation remain preliminary. Further research is necessary to better understand the internal stratigraphy and seasonal behaviour of frozen debris-lobes, in particular, ice content, liquid water content and seasonal movement rates. In addition to continuing field observations and remote-sensing analysis, we suggest utilising geophysical methods and employing a drilling programme in future investigations.

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LETTER

Circumpolar Arctic vegetation: a hierarchic review and roadmap toward an internationally consistent approach to survey, archive and classify tundra plot data

D A Walker[1], F J A Daniëls[2], I Alsos[3], U S Bhatt[4], A L Breen[5], M Buchhorn[1], H Bültmann[2], L A Druckenmiller[1], M E Edwards[6], D Ehrich[7], H E Epstein[8], W A Gould[9], R A Ims[7], H Meltofte[10], M K Raynolds[1], J Sibik[11], S S Talbot[12] and P J Webber[13]

1 Institute of Arctic Biology, University of Alaska Fairbanks, USA
2 Institute of Biology and Biotechnology of Plants, Münster University, Germany
3 Tromsø Museum, UiT, The Arctic University of Norway, Tromsø, Norway
4 Department of Amospheric Sciences, University of Alaska Fairbanks, USA
5 International Arctic Research Center, University of Alaska Fairbanks, USA
6 Department of Geography and Environment, University of Southampton, UK
7 Department of Arctic and Marine Biology, UiT—The Arctic University, Norway
8 Department of Environmental Sciences, University of Virginia, USA
9 US Forest Service International Institute of Tropical Forestry, Rio Piedras, Puerto Rico, USA
10 Department of Bioscience, Aarhus University, Denmark
11 Slovak Academy of Sciences, Bratislava, Slovak Republic
12 US Fish and Wildlife Service, Anchorage, AK, USA
13 Department of Plant Biology (Emeritus), Michigan State University, Ann Arbor, MI, USA

E-mail: dawalker@alaska.edu

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Abstract

Satellite-derived remote-sensing products are providing a modern circumpolar perspective of Arctic vegetation and its changes, but this new view is dependent on a long heritage of ground-based observations in the Arctic. Several products of the Conservation of Arctic Flora and Fauna are key to our current understanding. We review aspects of the PanArctic Flora, the Circumpolar Arctic Vegetation Map, the Arctic Biodiversity Assessment, and the Arctic Vegetation Archive (AVA) as they relate to efforts to describe and map the vegetation, plant biomass, and biodiversity of the Arctic at circumpolar, regional, landscape and plot scales. Cornerstones for all these tools are ground-based plant-species and plant-community surveys. The AVA is in progress and will store plot-based vegetation observations in a public-accessible database for vegetation classification, modeling, diversity studies, and other applications. We present the current status of the Alaska Arctic Vegetation Archive (AVA-AK), as a regional example for the panarctic archive, and with a roadmap for a coordinated international approach to survey, archive and classify Arctic vegetation. We note the need for more consistent standards of plot-based observations, and make several recommendations to improve the linkage between plot-based observations biodiversity studies and satellite-based observations of Arctic vegetation.

1. Introduction

Accurate and consistent approaches for documenting the composition and structure of Arctic vegetation and its relationships to the environment are essential to ground-based and remote-sensing studies that attempt to understand Arctic biodiversity and the causes of circumpolar vegetation change (Bunn and Goetz 2006, Bhatt *et al* 2010, Elmendorf *et al* 2012, 2015, Meltofte *et al* 2013, Myers-Smith *et al* 2015b). The International Biological Program (IBP) Tundra Biome stimulated Arctic vegetation research between 1967 and 1974 (Brown *et al* 1980, Bliss 1981, Bliss *et al* 1981), which led to numerous

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syntheses in the 1990s (Chapin *et al* 1992, Oechel *et al* 1997, Wielgolaski 1997). More recently the Flora Group within the Conservation of Arctic Flora and Fauna (CAFF) made major progress toward an integrated circumpolar view of Arctic vegetation. CAFF is the biodiversity working-group of the Arctic Council, which is an intergovernmental forum promoting international cooperation, coordination and interaction among the eight Arctic Nations.

The *Annotated PanArctic Flora* (*PAF*) *Checklist* (Elven *et al* 2011) was first proposed at the 1975 International Botanical Congress in Leningrad as a means to assess panarctic plant diversity (Murray and Yurtsev 1999). The PAF was completed under the leadership of Reidar Elven and colleagues at the University of Oslo, and is now a living updatable online annotated checklist that provides a consensus of the names for all Arctic vascular plants. A new Arctic Vegetation Archive (AVA) initiative, described later in this paper, relies heavily on the PAF for standardized plant names. The Circumpolar Arctic Vegetation Map (CAVM), which was first proposed at the 1992 International Arctic Workshop on Classification of Arctic Vegetation in Boulder, CO (Walker *et al* 1994), and the map was completed in 2003 (CAVM Team 2003, Walker *et al* 2005). The CAVM provided a framework for the Arctic Biodiversity Assessment (ABA) (Meltofte *et al* 2013), which included three circumpolar vegetation-related syntheses devoted to plants (Daniëls *et al* 2013), fungi (Dahlberg *et al* 2013), and terrestrial ecosystems (Ims *et al* 2013). In sections 2, 3 and 4 of this review, we use several products from the ABA, along with other sources, to describe our current hierarchical understanding of Arctic vegetation at circumpolar, regional, and land-scape levels. In section 5 we focus at the plot level. We describe an example plot archive from Arctic Alaska, and make several recommendations that provide the beginning of a roadmap for more consistent international approaches to surveying, archiving, and classifying Arctic plot data.

2. Circumpolar patterns: the north–south influence of zonal climate and sea ice

The Arctic bioclimate zone occupies the land area beyond the northern climatic limit of forests. The zone has cold winters (mean January temperatures well below freezing) and cool summers (mean July temperatures below about 10 °C–12 °C). The Arctic zone covers 7.1×10^6 km^2, or about 4.8% of the land area of the Earth. Of this, glaciers cover about 29%; the remaining area constitutes the Arctic Tundra Biome, which has an Arctic flora, and tundra vegetation composed mostly of various combinations of herbaceous plants, small shrubs, mosses, and lichens (Walker *et al* 2005).

The Arctic Tundra Biome is essentially a long narrow ecological transition zone between the boreal forest and the Arctic Ocean. Eighty percent of the entire lowland portion of the Arctic zone lies within 100 km of the cooling influence of seasonally ice-covered seas with roughly 177 000 km of highly dissected coastline. This narrow circumpolar ribbon of tundra is divided into five Arctic bioclimate subzones (figure 1, inset map). The subzone boundaries are based primarily on the Arctic phytogeographic zones of Boris Yurtsev (Yurtsev 1994) and are defined according to summer temperatures and dominant growth forms of plants in the zonal vegetation types. The subzones as delineated by geobotanists are generally closely aligned with land-surface summer-warmth index classes (figure 1, main map) that were derived from the Advanced Very High Resolution Radiometer satellite data (Raynolds *et al* 2008a). The map also shows areas where some adjustments in the subzone boundaries are needed, particularly along steep coastal temperature gradients, on islands, and in mountainous areas.

The growth forms and diversity of plant species that comprise tundra plant canopies are related to the available summer warmth along latitudinal and altitudinal gradients. For example, the vertical structure of zonal vegetation varies from very small plants (<2 cm tall) in a single discontinuous layer in subzone A to complex plant canopies with two to three layers in subzone E, which can include shrubs that exceed 80 cm tall (Walker *et al* 2005). Species richness in the five Arctic subzones increases twenty-fold from north to south, but the number of endemics increases only about a three-fold (Daniëls *et al* 2013). Within Arctic mountain ranges, floristic richness in altitudinal bioclimatic belts is similar to the richness in latitudinal bioclimate subzones with similar summer temperature regimes, but strongly modified by the effects of slope and duration of snow cover (Sieg *et al* 2006).

Subzone A is the coldest (mean July temperatures less than 3 °C), smallest (approximately 2% of the area of the Arctic) and most unique subzone, with tundra unlike that elsewhere in the Arctic. The subzone lacks dwarf shrubs, all woody plants, sedges, bog mosses (*Sphagnum*), and peat in wetlands, all of which are among the dominant characteristics of tundra vegetation in subzones further south. A new class of vegetation, the *Drabo corymbosae-Papaveretea dahliani* (Daniëls *et al* 2016), has been described recently to characterize the zonal vegetation of subzone A. Subzone A is also the most threatened subzone. It is restricted to parts of the Arctic that, until recently, were generally surrounded by summer coastal sea ice all summer. Melting of the summer ice will result in higher summer temperatures on the adjacent land areas. Only a 1 °C to 2 °C increase in July mean temperatures in subzone A would permit the establishment of woody dwarf shrubs, sedges, and a large group of species that are generally currently missing in subzone A (Walker *et al* 2008).

A circumpolar map of Arctic aboveground phytomass on zonal sites (figure 2(a)) is based on the strong

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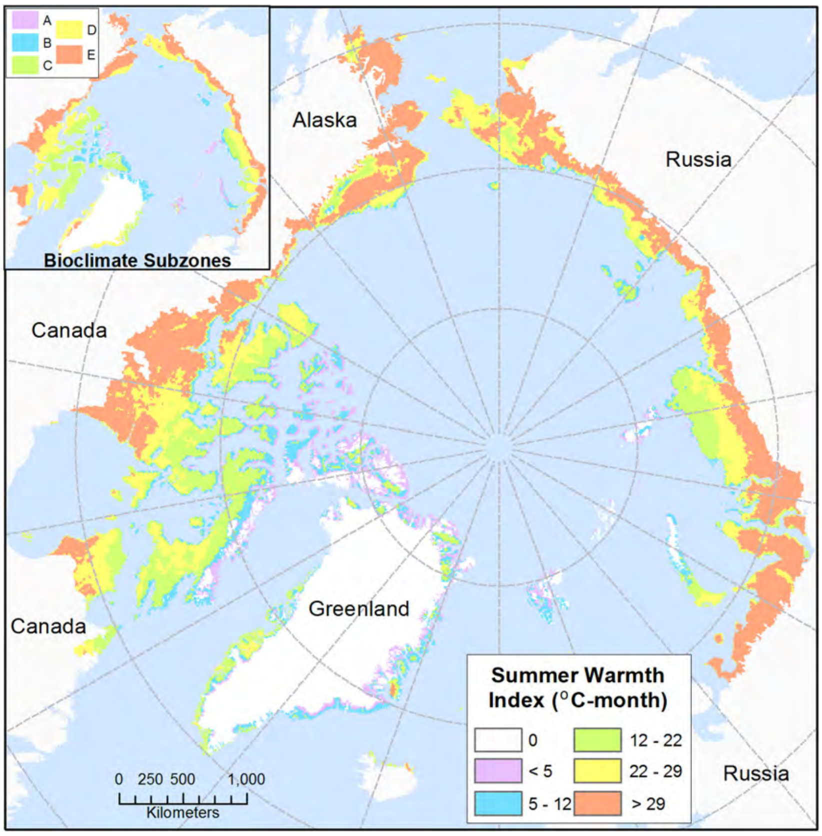

Figure 1. Map of the 22 year (1982–2003) mean of the summer warmth index (SWI = sum of monthly mean temperature above freezing) of arctic tundra, based on AVHRR land surface temperature data. Compare to inset map showing Arctic bioclimate subzones according to the CAVM Team (2003). The AVHRR-derived temperature is the land-surface radiant temperature, which characterizes the environment of low growing tundra plants within the surface boundary layer better than climate station temperature data, which are measured 2 m above the ground. On a monthly basis, Arctic mid-summer land-surface temperatures are warmer than air temperatures by about 2 °C, but vary considerably under different summer climate regimes. Adapted from Raynolds *et al* (2008a).

correlation between phytomass and the Normalized Difference Vegetation Index (NDVI) (figure 2(a), inset regression curve). The NDVI is a 'greenness index' derived from spectral-reflectance data. NDVI values are calculated from a variety of optical sensors aboard Earth-orbiting satellites, and are used for monitoring vegetation biomass, productivity, and related properties (Tucker and Sellers 1986) (see legend of figure 2 for how the index is calculated). In the Arctic, NDVI is often well correlated with ground measurements of phytomass, the leaf-area index (LAI), carbon dioxide flux and other measures of tundra photosynthetic activity (Stow *et al* 2004). The phytomass values reported in figure 2(b) were obtained from plots of zonal vegetation along two latitudinal transects in North America and Eurasia that cross all five Arctic bioclimate subzones (Raynolds *et al* 2012).

Temporal changes in tundra greenness are monitored annually using the NDVI (Bhatt *et al* 2010, Epstein *et al* 2014). The maximum NDVI (MaxNDVI) is an index of the peak greenness and the peak phytomass reached in a given summer. A general increase in MaxNDVI occurred from 1982 to 2013 in most of the Arctic (figure 3) (Bhatt *et al* 2013). This is generally attributed to increased growth of warmth-adapted plants, particularly deciduous shrubs (Myers-Smith *et al* 2015a), but there is considerable spatial and temporal variation. Some areas, particularly much of Arctic Russia and southwest Alaska, show recent (1999–2011) declines in midsummer temperatures and MaxNDVI, which suggests decreased productivity is linked to documented increased midsummer cloudiness and cooler midsummer temperatures (Bhatt *et al* 2013).

3. Regional patterns

3.1. The east-west influences of geography, geology, and history

Much of the regional variation in Arctic productivity (figure 2) and biodiversity (figure 4) can be attributed to historical patterns of glaciation, changes to the positions of the Arctic coastlines, and differences in parent material. For example, the amount of time since deglaciation accounts for about 34% of the variation in circumpolar aboveground phytomass and NDVI patterns (Raynolds and Walker 2009).

Global cooling over the past ~50 million years (MY) led to particularly dramatic changes in the environment of the Arctic. The cooling was linked to a

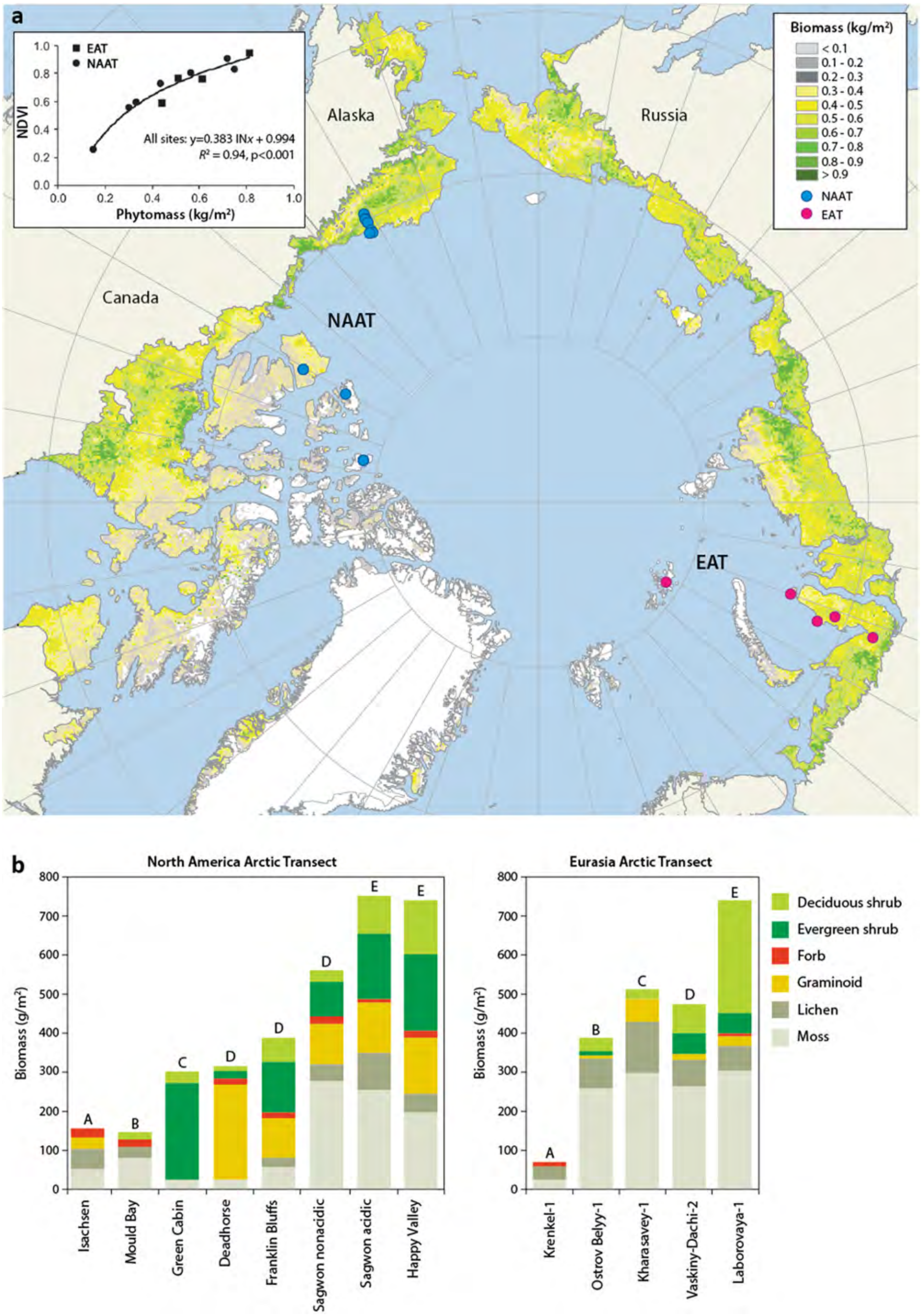

Figure 2. Aboveground zonal phytomass in the Arctic. (a) Zonal phytomass map based on NDVI-phytomass regression (inset graph, upper left). NDVI (normalized difference vegetation index) is interpreted as the photosynthetic capacity of the vegetation and is calculated by the formula NDVI $=$ (NIR $-$ R)/(NIR $+$ R), where NIR is the near-infrared band of the spectrum and R is the red band of the spectrum. The relation was calculated using GIMMS3g AVHRR maximum NDVI 8 km data for years during which the phytomass was collected (2003–2010). The bioclimate subzone of each location is indicated by the letter above each bar. (b) Clip-harvest samples of zonal vegetation were made along pan-Arctic transects in North America (NAAT, blue dots) and Eurasia (EAT, red dots) summarized for each location along the NAAT and EAT by plant functional type. Adapted from Raynolds *et al* (2012) for the Arctic Biodiversity Assessment (Meltofte *et al* 2013) and reprinted by permission of CAFF.

drop in levels of atmospheric greenhouse gases and to continental drift, which altered ocean currents and patterns of global heat transport. The fossil record indicates that over much of this period climates were temperate, and lower-elevation terrain within the present-day Arctic was forested (Miller *et al* 2010). Between 2 and 3 MY ago, a major climatic transition featuring growth of sea ice and cooling of the Arctic Ocean led to forest retreat, the development of tundra vegetation, and permafrost expansion. The past ~2 MY have seen repeated advance and retreat of ice sheets (the Quaternary glaciations), but these have been geographically asymmetric. Ice repeatedly spread across large areas of Canada, Greenland, northern Europe and northwestern Russia, whereas Beringia, which extends from northeast Siberia to far northwest

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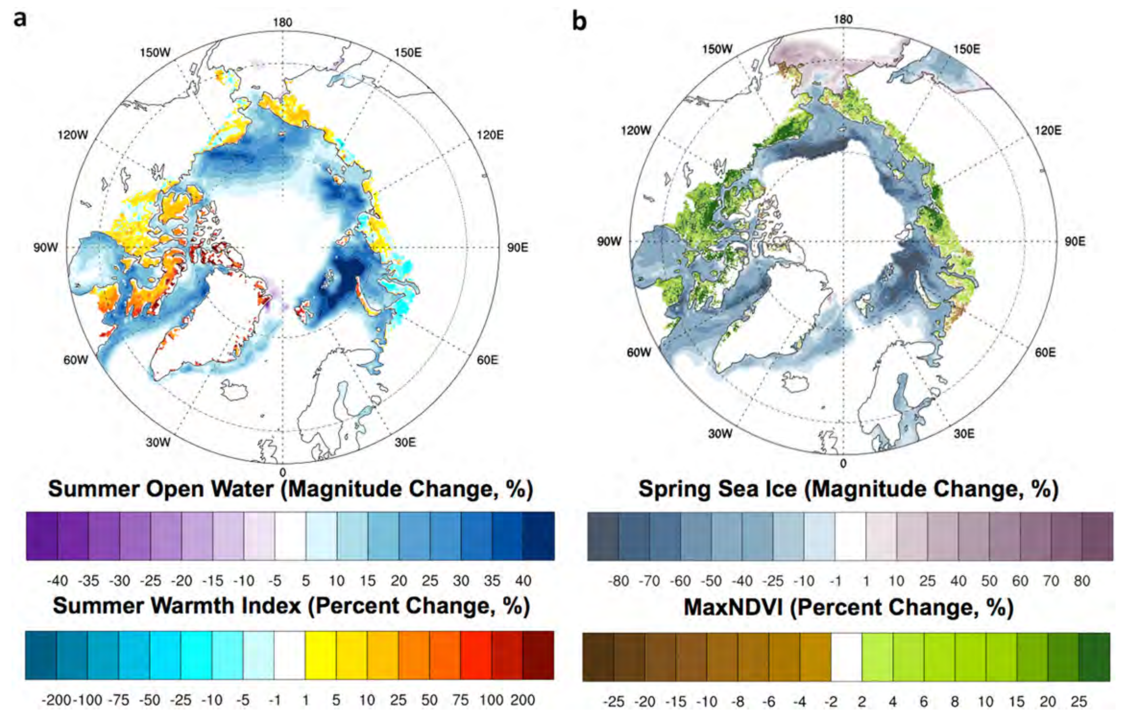

Figure 3. (a) Circumpolar changes in summer open water and the summer warmth index (SWI); and (b) the extent of spring sea ice and maximum NDVI (MaxNDVI). Changes in summer open water were determined during May–August SWI is the annual sum of the mean monthly temperatures exceeding freezing. The changes of sea-ice breakup are represented by 50% sea ice concentration. The annual maximum NDVI is usually reached in early August. The sea ice concentration and open water data were derived from SMMR and SSM/I passive microwave records. NDVI and land surface temperatures (SWI) information were derived from AVHRR data and the NDVI is from the Global Inventory, Modeling and Mapping Studies (GIMMS) dataset. (Adapted from Bhatt *et al* 2010, updated to 2013).

Canada, experienced only local mountain glaciations. During periods of lowered sea level, Beringia included the large land bridge that became exposed in the area of the present-day Bering Strait. The glaciated regions were subject to large-scale processes of erosion and deposition that eliminated the vegetation, though the extent of the ice varied spatially and temporally during the Quaternary period (Edwards *et al* 2000). During glacial periods, the climate over most of Beringia was cold and dry, which limited woody vegetation. The fossil record indicates the vegetation was dominated by graminoid species and forbs that have tundra and steppe affinities today (Anderson *et al* 2004). Nevertheless, the heterogeneity of Beringian landscapes almost certainly afforded local refugia for a range of woody plants (Brubaker *et al* 2005). In relatively warm, interglacial periods, such as the current Holocene (the past ~11 000 years), the dry herbaceous vegetation switched to mesic communities featuring a greater dominance of shrubs (Anderson *et al* 2004).

The Arctic is presently divided into floristic provinces and subprovinces that reflect the geographic history described above (Yurtsev 1994). The most recent iteration of these divisions has five phytogeographic provinces and 21 subprovinces (figure 4, legend upper left). There are 2218 recognized vascular plant species in the Arctic, distributed in 430 genera and 91 families (Elven *et al* 2011). Floristic diversity is low compared to other biomes and is less than 1% of the world flora. Thirty-six percent of the species belong to only four families: Asteraceae (254), Poaceae (224), Brassicaceae (133) and Cyperaceae (190) (Daniëls *et al* 2013). Floristic diversity varies widely across the phytogeographic provinces, largely a consequence of the varied glacial histories. The Beringian group of provinces has relatively high floristic diversity (315–825 species; average 621 species), which reflects its vast unglaciated areas, whereas the heavily glaciated North Atlantic group has relatively low diversity (215–649; average 449) (figure 4). Of the 106 Arctic endemics, the Beringian provinces have 39; whereas, European Russia-West Siberia provinces have only three (Daniëls *et al* 2013).

3.2. Genetic diversity

Genetic diversity within species is essential to long-term persistence of floristic diversity because it provides the opportunity for species to adaptively respond to changing climate. Similar to the patterns of floristic diversity, the highest levels of genetic diversity and most local genetic markers are found in Beringia with lower numbers in the North Atlantic region (Eidesen *et al* 2013). While Beringia has generally been inferred as a long-term refugium for Arctic plants (see above), there has been intense debate about the history of the plants in the repeatedly and heavily glaciated amphi-Atlantic region (Brochmann *et al* 2003). Genetic evidence indicates that a few species may have been

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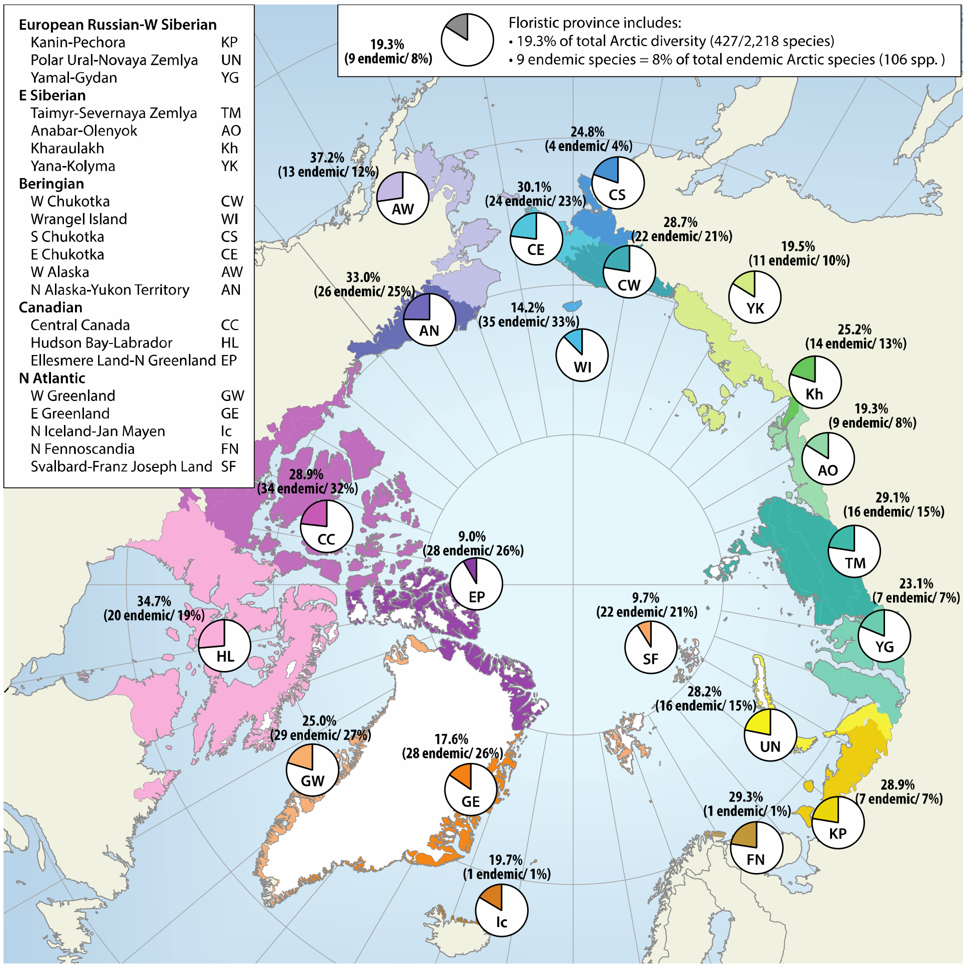

Figure 4. Vascular-plant species richness within each phytogeographic province (colors and codes on the background map) as a percentage of the total Arctic species richness (2218 species). The number of endemic species is shown in parentheses with percentage of the total arctic endemic species (106). From Daniëls *et al* (2013). Floristic provinces are according to Elven *et al* (2011) (reprinted by permission of the CAFF).

able to survive *in situ* during the last glacial maximum (Westergaard *et al* 2011), whereas the majority of species colonized post-glacially (Alsos *et al* 2015). This is reflected in the low number of Arctic endemic species (figure 4), the very few species endemic to any of the floristic provinces and the overall low levels of genetic diversity (Eidesen *et al* 2013). Genetic studies of 1200 populations of 27 northern vascular plant species combined with distribution modeling predict that most northern plant species will lose ranges at a higher rate than temperate species. The predicted loss of genetic diversity is overall less than range loss, but varies with species traits, such as adaptation to dispersal and growth form (Alsos *et al* 2012).

3.3. Productivity and diversity hotspots

No Arctic region is considered a global-scale hotspot of biodiversity (Vane-Wright *et al* 1991, Myers *et al* 2000, Meltofte *et al* 2013), but unglaciated regions, particularly in Beringia, have relatively high floristic diversity compared to the rest of the Arctic. Relatively large areas (100–1000 km^2) with locally high productivity and diversity also occur in association with unique physiographic features that influence local climate. These include the Arctic 'oasis' along the 70 km long Lake Hazen, near the northern limit of land (81.8° N) on Ellesmere Island (Svoboda and Freedman 1994), and the coastal plain of the Arctic National Wildlife Refuge in northeastern Alaska, where the eastern Brooks Range makes a turn toward the Arctic coast and compresses three Arctic bioclimate subzones to within 50 km of the Arctic Ocean.

The concept of hotspots needs to distinguish areas containing many endemic Arctic species with high conservation priority from local thermal hotspots with high biological productivity. The presence of anomalously tall shrubs or trees is an indicator of thermal hot spots in the Low Arctic (Forbes *et al* 2010, Lantz *et al* 2010, Tape *et al* 2012), but not necessarily hot spots of diversity. An area of particularly lush shrub and poplar growth in northern Alaska is the north-flowing Chandler River in the central part of the Arctic Foothills (Tape *et al* 2011). The presence of balsam poplar (*Populus balsamifera*) is another good indicator of local thermal hot spots because these trees often form small boreal enclaves that occur on thermally warm valleys and south-facing slopes of the Brooks Range, often near springs associated with limestone bedrock areas. Summer-warmth-index maps derived from satellite data indicate that about 40% of the balsam poplar stands in northern Alaska occur in sites with relatively high summer ground-surface temperatures (Breen 2014).

Remote sensing can be a useful tool to help identify potential hot spots of diversity and high productivity. In the Bathurst Inlet area of northern Canada, areas of relatively high species diversity correspond to areas with high diversity of spectral-signatures on Landsat images (Gould and Walker 1997, 1999). In Svalbard, a combination of remote sensing tools, digital elevation models, and detailed ground-based surveys were used to verify the presence of locally rare thermophiles in this High Arctic environment (Karlsen and Elvebakk 2003), and have recently been used to develop habitat suitability and species distribution models (Nilsen *et al* 2013). However, as shown in the discussion of subzone A, it is the *lack* of species from the south that give the extreme High Arctic areas their special character and conservation value.

4. Landscape-scale patterns

Major landscape-scale differences in productivity and species diversity can be attributed to underlying geology and topography, and resulting differences in soil, snow and wetland distribution. Successional patterns related to streams, lakes, fire, coastal flooding and humans are additional landscape-level factors. The effect of soil pH on Arctic vegetation is a particularly important factor that has been described in numerous studies (Edlund 1982, Elvebakk 1997, Walker *et al* 1998). For example, a striking substrate pH boundary stretches 800 km across the northern front of the Arctic Foothills in northern Alaska (figure 5). The boundary is thought to be caused by different ages of loess deposits on either side of the boundary, possibly enhanced by a regional climate boundary that coincides with the northern front of the Arctic Foothills (Zhang *et al* 1996). Differences in soil pH across the boundary affect the composition and structure of plant communities, and a wide variety of ecosystem properties and processes, including soil temperature, active-layer thickness, photosynthesis, respiration, decomposition, and fluxes of trace gases energy and water (Walker *et al* 1998). Similar patterns are seen in mountain ranges and other terrain with adjacent areas of carbonate-rich and acidic bedrock (Edlund 1982, Cooper 1986, Elvebakk 1994). Older landscapes generally have more leached soils with lower soil pH than younger surfaces. For example, the area near Toolik Lake, Alaska, has been subjected to repeated glaciations during the Pleistocene, leaving several glaciated landscapes of different age that span over a MY of glacial history within about 100 km north of the Brooks Range. Each different-aged glacial surface can be recognized by characteristic suites of landforms, periglacial features, soils and vegetation that are legacies of its geomorphic history (Hamilton 1986). Difference in productivity on the different-age surfaces can be inferred from NDVI patterns and corresponding biomass data (Walker *et al* 1995).

Landscape-scale maps at fine scales (approximately 1:5000 scale and finer) can display transitions in plant communities along mesoscale hill slopes (toposequences), riparian areas, snowbeds, and wetlands. Variation related to patterned-ground features is especially common in the Arctic (Washburn 1980). A study of non-sorted circles along the Arctic climate gradient found that major differences in soil moisture, soil temperature, and site stability occur within spatial distances of a few centimeters, and that the vegetation biomass and thickness of the plant layer on the patterned-ground features affect the soil thermal, hydrological, and nutrient properties (Kade *et al* 2005, Walker *et al* 2011, Frost *et al* 2013). Maps of patterned-ground landscapes ranging in size from about 4 m^2 to 1 ha are sometimes made at very fine scales (1:500 scale or finer) (Chernov and Matveyeva 1997, Raynolds *et al* 2008b).

Animals are also a major factor affecting landscape-level vegetation and productivity patterns. Rich habitats are often associated with areas of high animal use such as the south-facing gravelly slopes of pingos (Walker 1990), bird cliffs (Williams and Dowdeswell 1998), and archeological sites near polynyas in the central and High Arctic (Schledermann 1980, McCartney and Helmer 1989, Murray 2005). Animals can have both negative and positive effects on productivity. Resampling vegetation within herbivore exclosures at Barrow, Alaska, in the 1950s and 1970s found that lemmings and other herbivores outside the exclosures had reduced the relative cover of lichens and graminoids while the relative cover of deciduous shrubs increased; consequently, a wide variety of ecosystem properties, including thaw depth, soil moisture, albedo, NDVI, net ecosystem CO_2 exchange, and methane efflux were affected (Johnson *et al* 2011). Outbreaks of insect defoliators have also been shown to dramatically impact deciduous shrubs in low-arctic Greenland (Post and Pedersen 2008) and at the forest-tundra interface in Northern Fennoscandia (Jepsen *et al* 2013). These pulses of defoliation lead to changed nutrient cycling, and increased understory vegetation and indirectly affect herbivore community composition. Abundant semi-domestic reindeer populations, in combination with cyclic vole populations, appear to be able to counteract the climate-driven increase in shrub growth in some areas of the Low Arctic (Ravolainen *et al* 2014). One of the most dramatic examples of herbivore overabundance is the case of snow geese, which permanently transformed and partially destroyed large areas of salt-Marsh vegetation along the Hudson Bay in Canada (Jefferies *et al* 2006).

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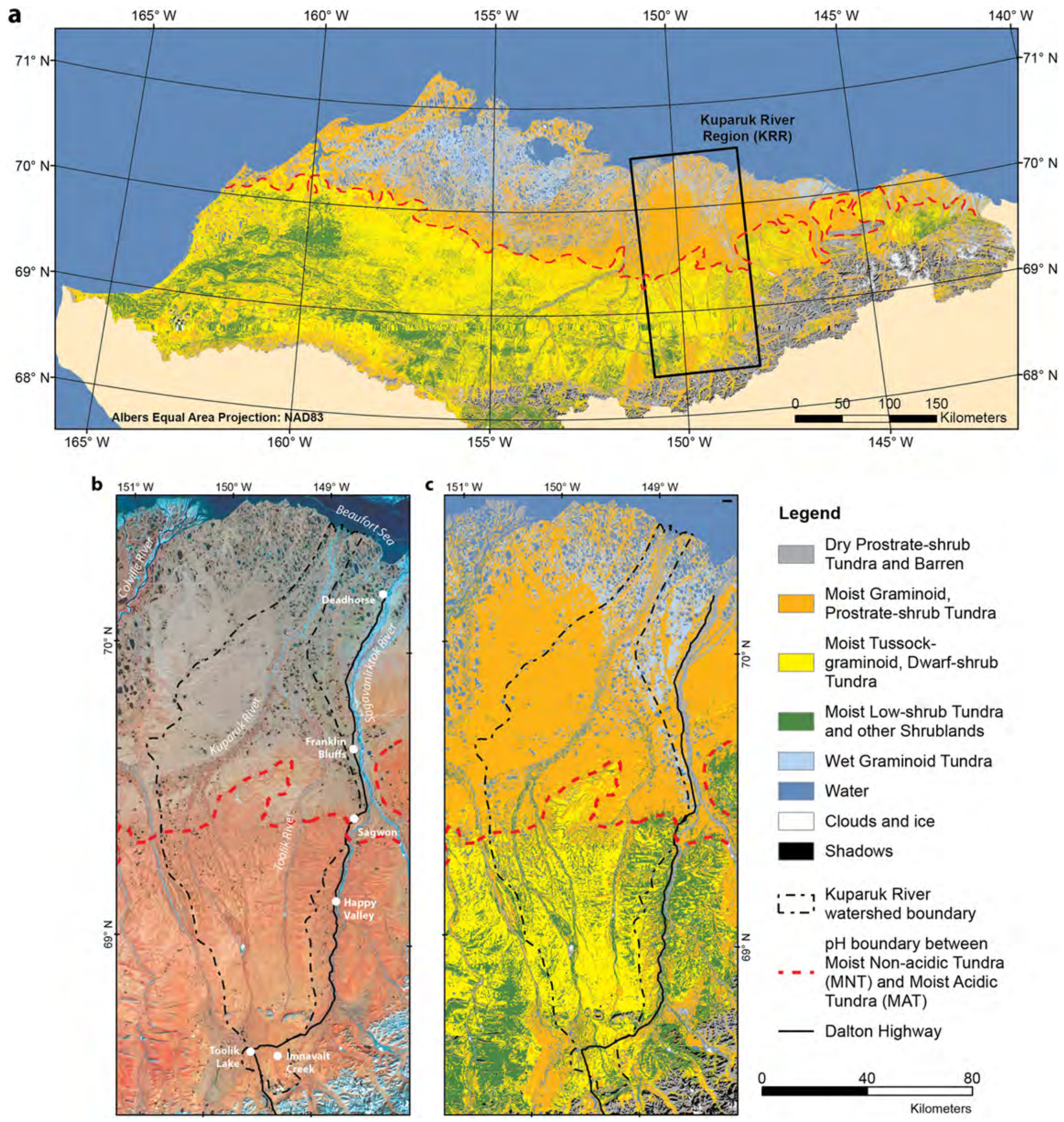

Figure 5. (a) Land-cover map of northern Alaska (adapted from Muller *et al* 1999). The map shows the distribution of major physiognomic groups of tundra types. The red dashed line separates mainly graminoid and prostrate-shrub-dominated tundras (orange) on in the northern part of the map from shrubbier tussock tundra (yellow) and low-shrub tundra (green) in the southern part. Wet tundra (light blue) also occurs on flat landscapes of northern coastal plain. The black rectangle contains the Kuparuk River region, an intensively studied Arctic watershed. (b) Landsat MSS false-color infrared mosaic of the Kuparuk River watershed (dashed black line). In this region, the gray area north of the red dashed line has predominantly moist nonacidic tundra (MNT). Redder areas south of the boundary have mainly moist acidic tundra (MAT). The redder tones of MAT are due mostly to more dwarf and low shrubs (e.g., *Betula nana, Ledum palustre* ssp. *decumbens*, and *Salix pulchra*). MNT vegetation has fewer shrubs, more erect dead sedge leaves, and more exposed soil patches due to a greater abundance of non-sorted circles. (c) Land-cover map of the Kuparuk River Region derived from the Landsat data (Muller *et al* 1998). Landsat data are courtesy of the US Geological Survey Alaska Data Center.

5. Plot-level observations: a panarctic vegetation plot archive

A conceptual diagram summarizes the four levels of observation of circumpolar Arctic vegetation and typical research topics described above, along with, monitoring, integration and modeling tools that can be applied across scales (figure 6).

Our knowledge of Arctic floristic (plant-species) and vegetation (plant-community) response to environmental gradients at all these scales relies on rather sparse ground-based plot data collected during expeditions and at Arctic observatories since the late 1800s. Vegetation data are usually collected from small plots that describe the structure, composition, and site factors of the plant canopy in common vegetation habitat types (figure 7).

5.1. Arctic vegetation plot databases

Plot based survey data are increasingly gathered and stored in large vegetation databases (Schaminée *et al* 2011). The Arctic Vegetation Archive (AVA) is an effort to assemble historic Arctic vegetation plot data into a single publically accessible database and to apply it to northern issues, including a much needed circumpolar Arctic vegetation classification (Walker

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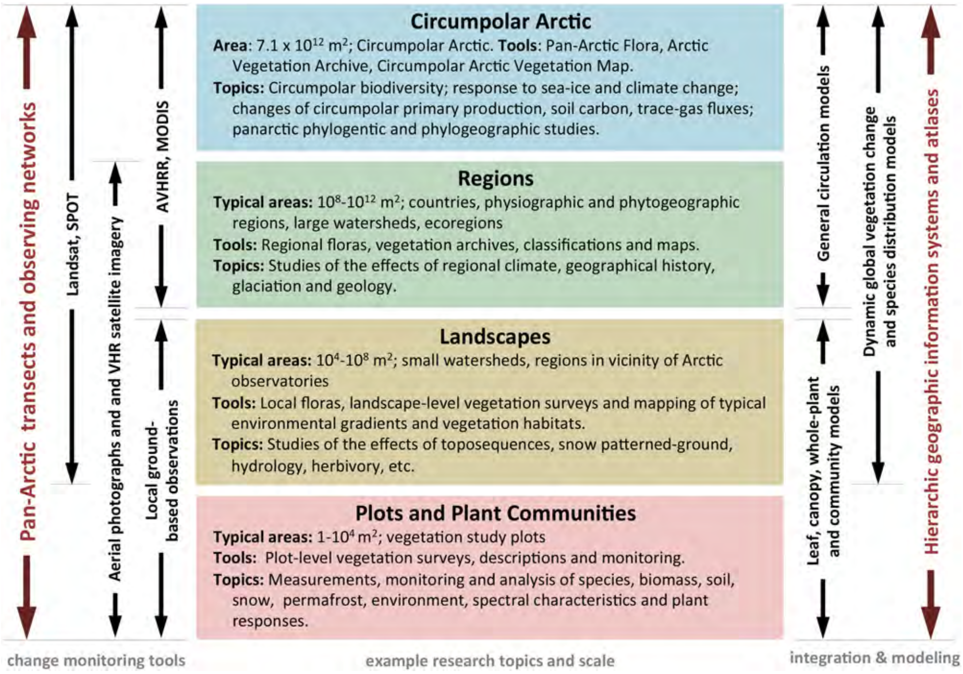

Figure 6. Hierarchy of levels of observation of Arctic vegetation. Left-hand vertical arrows show examples of monitoring tools that are effective across levels; right-hand bar shows corresponding examples of integration and modeling tools. Red highlighted monitoring and integration tools indicate methods used to examine vegetation at the full range of scales.

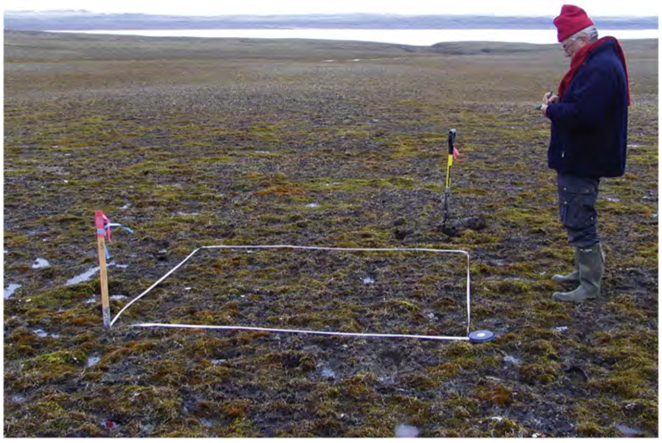

Figure 7. A vegetation survey being conducted in a wet vegetation plot located near Isachsen, Ellef Ringnes Island, Nunuvut, Canada, 78° 47′N, 103° 35′W, part of the North America Arctic Transect (blue dots on figure 3), using the Braun-Blanquet approach (Westhoff and van der Maarel 1978). This simple survey method is used widely across the Arctic.

and Raynolds 2011, Walker 2014). Prototype databases for the AVA are under development for Greenland (AVA-GL) (Bültmann and Daniëls 2013) and Arctic Alaska (AVA-AK) (Walker *et al* 2013). The AVA-AK is nearest to completion and currently contains species and environmental data from approximately 3000 vegetation plots in 24 datasets in northern Alaska (Walker *et al* 2016). The archive is accessible through the Alaska Arctic Geoecological Atlas (figure 8), a web-based portal at the University of Alaska. Each dataset has a 'Catalog' record with a detailed description of the dataset. Downloads or links to plot photographs, maps of plot locations, soil and environmental data, biomass and spectral data information and key data reports and publications are also provided wherever available.

The raw and standardized plot data are stored in .csv files, and a Turboveg database contains the species data from all AVA-AK datasets with consistent plant nomenclature and header data (a standardized set of key environmental variables). Turboveg is the most widely used software program specifically designed for

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Figure 8. Home page for the plot archive within the Alaska Arctic Geoecological Atlas, showing locations of 38 currently known Arctic tundra plot datasets. Twenty four of these (dark and light green points) are in the AVA-AK Turboveg database; 17 (dark green) have complete catalog data records; the gray datasets are still being evaluated for inclusion. Clicking on a point or dataset name leads to a large scale image that shows individual plot locations and a Catalog data record which explains the data and provides links to the species data, plot photos, and other ancillary information if available.

the storage, selection, and export of vegetation plot data (Hennekens and Schaminée 2001). Plot data stored in Turboveg can be exported for further analysis by other spreadsheet and database tools (e.g., Microsoft Excel and Access, Twinspan, Canoco, PC-ORD, and JUICE). A key aspect of the AVA is a PanArctic Species List (PASL), which standardizes species names across datasets in the Turboveg database (Raynolds *et al* 2013). The AVA-AK Turboveg database follows as closely as possible the database protocols being developed for the European Vegetation Archive (Chytrý *et al* 2016). The data are also being exported to the VegBank plot database, which is used for the US National Vegetation Classification (USNVC) (Peet *et al* 2012). The AVA-AK is registered in the Global Index of Vegetation-plot Databases (Dengler *et al* 2011).

A preliminary cluster analysis of the first 16 datasets (1568 plots) produced a dendrogram with 17 clusters with sensible ecological organization, mainly along a complex soil-moisture/ soil-pH gradient. The diagnostic, constant, and dominant taxa in these clusters appear to show strong correspondence to previously described Br.-Bl. classes and alliances described elsewhere in the Arctic (Walker *et al* 2016).

5.2. Toward a coordinated international approach to survey and archive plot data

Although the AVA-AK database is a significant step toward developing a classification for Arctic Alaska and the circumpolar region, the datasets in the archive show considerable variability in quality. The data were collected during a period of over 65 years using a wide variety of survey methods. *Incompatible methods* included: (1) project-specific sampling protocols that made it difficult to compare datasets from different locations; (2) data that were collected from plots with obviously heterogeneous vegetation; (3) doubtful or incomplete taxonomic determinations. *Missing information* included: (4) data that were published only in summary form for vegetation types but not for the individual plot samples; (5) missing important ancillary information, such as plot coordinates, photographs of the vegetation, nature of the soils, or positions along slope, soil moisture, or snow gradients; (6) loss of the original data and/or critical metadata due to the death

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of the author(s); and (7) datasets that were unavailable because they were obtained for private industry and considered proprietary information.

Considerable progress toward a roadmap for international vegetation surveys has been made and summarized in a recent review (De Cáceres *et al* 2015). This framework is not reviewed here, but is an essential starting point for new vegetation surveys. Below, we provide some specific suggestions for future surveys in the Arctic. In most respects, these suggestions follow the 'analytic research phase' of the Braun-Blanquet (Br.-Bl.) approach described by Westhoff and Van der Maarel (1978) with rather minor adjustments specific for Arctic situations. We add some additional suggestions, such as collection of biomass and soil data, which greatly increase the value of plot data for remote-sensing and other applications.

5.2.1. Choice of area for a vegetation survey

The Arctic is remote and under-sampled. New surveys should focus in areas that have good logistical support, such as the existing network of Arctic Observatories, where researchers can spend the time necessary to produce high-quality datasets and where there is a likelihood that the plots will be revisited in the future for comparative monitoring studies. Special efforts should also be made to identify 'hotspots' of productivity, diversity, and endemism that are not represented at the main Arctic observatories. Remote sensing, local knowledge, and gaps in the existing plot network can aid in identifying these areas. Field camps should be considered to examine vegetation variation in ecological situations that are not adequately represented at the Arctic observatories or in the existing AVA.

5.2.2. Local floras

It is best to conduct vegetation surveys in conjunction with taxonomists who can devote the time necessary to make professional herbarium voucher collections and produce floristic surveys that include complete vascular-plant, bryophyte, and lichen species lists from a full suite of habitat types at each station. A standardized method of making local floras has been applied to approximately 500 locations in Russia (Tolmachev 1931, Yurtsev *et al* 2004, Balandin 2008, Khitun *et al* 2016). The Russian approach to making local floras should be considered and modified if necessary for other Arctic countries. The Pan-Arctic Flora and Pan Arctic Species List will need to be regularly updated as new floristic information is gained. There is also a critical need for a new generation of Arctic vegetation scientists with strong taxonomic training to make these floristic surveys.

5.2.3. Selection of plant communities in representative habitat types

Considerable debate surrounds the topic of plot selection, particularly whether to select sample sites preferentially based on expert knowledge, often in relation to typical habitats, as in the Br.-Bl. approach (Mueller-Dombois and Ellenberg 1974), or to use random approaches, including stratified random sampling, which better meet statistical assumptions required for ecological studies, but which under-sample rare habitat types. In practice, a compromise is often necessary to meet the realities imposed by budgets, available time, and other logistic constraints, while at the same time avoiding the circular reasoning of only documenting preconceived vegetation types (De Cáceres *et al* 2015). An in-depth field reconnaissance guided by fine-scale aerial imagery of the study area should precede the formal survey to assess the habitat variation within the local region. Most of the Arctic is still in a natural state, so a good approach is to focus on the natural habitats and prioritize the sampling according the most- to least-common habitat types within a local landscape. First target the most abundant stable zonal sites, where the vegetation is mainly a product of long-term adaptation to the local climate. Then sample other common plant communities that are apparent at landscape scales including vegetation along toposequences, snow gradients, chronosequences associated with stream terraces and lake succession, different bedrock and soil types, and finally in small-scale special habitats associated with such features as rocky talus slopes and blockfields, frost boils, perennial springs, dunes, and zoogenic communities. Another approach that yields high-quality data is to sample a given habitat type across a broad regional gradient. Examples include sampling zonal sites along climate (Matveyeva 1998) or elevation (Sieg *et al* 2006) gradients. Other examples have focused on snowbeds (de Molenaar 1976), pingos (Walker 1990), riparian habitats (Schickhoff *et al* 2002), poplar groves associated with springs and warm habitats (Breen 2014) and anthropogenically disturbed areas (Sumina 2012).

5.2.4. Centralized-replicate sampling approach

Within a given a representative habitat type, a relatively small sample plot should be placed within a larger visually homogenous area of vegetation with relatively homogeneous plant-species composition, canopy structure, and local environmental factors, so as to avoid obvious transitions or boundaries between plant communities (Mueller-Dombois and Ellenberg 1974). The specific sites for plots generally should be at least partially subjectively chosen (rather than randomly located) to avoid obvious transitions between plant communities. This is a particularly important consideration in Arctic patterned-ground landscapes, where considerable habitat variation may be unnoticed on aerial photographs and can occur within a few centimeters. Make replicate samples (5–10) in areas of the same habitat type. Sampling along disturbance gradients or chronosequences can be done in a similar way by choosing sample sites in plant communities that occur in multiple areas of the landscape. This sampling approach is good for classification but may not be compatible with experimental studies that require a purely random sampling design for making statistical inferences. In these cases, a statistician should be consulted to help design a sampling approach (De Cáceres *et al* 2015).

5.2.5. 'Minimum-area' plots
Ideally, the plots should be of the same size to compare the species diversity within them, and should contain a high percentage (90%–95%) of the total number of species in the plant community, but also be as small as possible so as to avoid sampling several plant communities in the same plot. Methods of determining the minimum area are described in the literature (Westhoff and van der Maarel 1978) but are sometimes difficult to apply to surveys that include many vegetation types with widely divergent vertical structure, or that are in areas of complex microtopography, such as areas of permafrost-related patterned-ground. A rough rule of thumb is that the plot size in m^2 should roughly equal the height of the vegetation in decimeters (Barkman 1989). Chytrý and Otýpková (2003) recommend 16 m^2 for most grassland, heathland and other herbaceous vegetation, 50 m^2 or low-shrub vegetation types and 200 m^2 for woodlands.

5.2.6. Permanent plot markers and photographs
The corners of the plot should be permanently marked and labeled in a manner that will be still be visible or at least locatable (e.g. with metal detectors) many years in the future. Plot documentation should include high-resolution GPS coordinates of the plot corner markers, and photographs of the vegetation landscape and soil with the plot number clearly visible. Visits to the plots in winter to collect snow data will require marking the plots with long vertical poles to aid in locating the plots in snow-covered landscapes.

5.2.7. Description of the sample site
Include habitat type, geographic coordinates, elevation, photos, slope, aspect, soil moisture regime, snow regime, pH, landform, parent material, geological setting, surface geomorphology, active-layer thickness, disturbance types and degree, animal sign, and stability of the soil. A standardized data form with codes or standard names for the various factors should be used so that this is part of the record for the plot. A list of required and recommended fields used for the AVA-AK are in Walker *et al* (2016).

5.2.8. Cover estimates for all vascular plants, lichens, and bryophytes
It is highly advisable to collect small samples of all species encountered in a plot to avoid misidentification. Expert taxonomists in various plant groups will probably be needed, especially for the mosses, liverworts, lichens, grasses, sedges, and willows. Cover estimates can use direct percentage cover estimates or classes, such as Br.-Bl. cover-abundance scores (Westhoff and van der Maarel 1978).

5.2.9. Characterize the soil
At a minimum photograph the soil profile, make a brief description, and collect soil samples from the plant rooting zone and the top mineral horizon for later physical and chemical analysis. Preferably, work with a soil scientist experienced in Arctic soils.

5.2.10. Biomass and spectral data
Biomass data and ground-based spectral data are necessary for linking remote-sensing spectral information to actual plant production. The methods for harvesting, sorting, and categorizing biomass samples can strongly impact the reported biomass values and need to be standardized to make the data comparable between datasets. This was attempted during the IBP in the late 1960s and 1970s (Wielgolaski *et al* 1981) with some success, but the methods need to be revisited and a manual developed that incorporates new knowledge and better serves the remote-sensing community. Standardized procedures are also required for collecting LAI and spectral-radiometric data for use in calculating vegetation indices, such as the NDVI. The use of spectral data in phytosociological studies is relatively new and sampling should be developed with the advice of a remote-sensing specialist.

5.2.11. Other data
Every attempt should be made to make the data as widely useful as possible. Vegetation scientists should return to their plots in other seasons, other years, and with experts in a variety of disciplines, for example, soils, remote sensing, snow ecology, and animal ecology, to help interpret the causes of the spatial and temporal patterns. The information is also essential to interpret changes to such things as active layer depths and trace-gas fluxes. However, care must be taken to protect the plots and surrounding vegetation from trampling during the revisits because these sites are extremely valuable and should be protected.

5.2.12. Publication of plot data
In the past, many journals would only publish synoptic or summary tables for vegetation types because of limited space, but recent wide acceptance of supplemental data files for on-line publications now make publishing the complete plot data a standard practice. We also highly recommend formal data reports for each survey that provide full methods, photographs, and all the ancillary data collected from the plots.

5.3. Toward an Arctic-wide vegetation classification
In polar regions of Canada, Greenland, Iceland, Svalbard, Russia, and the United States, the Br.-Bl. approach (Braun-Blanquet 1932, Westhoff and van

der Maarel 1978, Dengler *et al* 2008) has historically been the most commonly used vegetation-survey method. This has resulted in compatible preliminary structured syntaxonomical and nomenclature surveys that can serve as a foundation for future sampling and a coherent consistent classification system across the Arctic (Bültmann and Daniëls 2013, Daniëls and Thannheiser 2013, Nilsen and Thannheiser 2013). Of 16 datasets in a preliminary analysis of the AVA-AK, thirteen followed the Br.-Bl. approach for sampling and five of these followed the International Code of Phytosociological Nomenclature (ICPN) for naming plant communities (Walker *et al* 2016).

The Br.-Bl. approach is primarily a floristic-based approach at all levels of its hierarchical framework, which consists of four primary vertical levels of organization (class, order, alliance, and association). At the lowest level, an association is a floristically defined plant-community type with a set of diagnostic species. The methods of naming new units is strictly defined by the ICPN (Weber *et al* 2000), and acceptance of new units requires formal publication according to the code. The approach is described in several textbooks although none incorporates the latest computer-based approaches for using the method. Arctic countries outside of North America will likely continue to use the Br.-Bl. approach for vegetation surveys and classification in the near future.

In North America, a relatively new EcoVeg vegetation classification approach has developed in the last 40+ years (Jennings *et al* 2009, Faber-Langendoen *et al* 2014). The method is an eight-level physiognomic-floristic-ecological classification approach (Class, Subclass, Formation, Division, Macrogroup, Grpoup, Alliance, and Association). The highest level in the EcoVeg approach is the formation class, which is a broad combination of dominant plant growth forms adapted to certain environmental conditions. The methods of field surveys, classification, and naming communities are described in several publications (FGDC Vegetation Subcommittee 2008, Jennings *et al* 2009, Faber-Langendoen *et al* 2014). The approach was adopted by North American land-management agencies as the vegetation standard for the US National Vegetation Classification (USNVC) (Faber-Langendoen *et al* 2014) and the Canadian National Classification (CNVC) (MacKenzie and Klassen 2004). It will likely continue to gain favor in North and South America.

We do not advocate one approach over the other because each approach has its advantages and will likely be continued where it is now practiced. However, one major advantage of the Br.-Bl. method for Arctic vegetation classification is that it has been applied in most regions of the Arctic and new data and analyses can build on the existing data and typologies. There is currently a lack of such an Arctic tradition with the EcoVeg approach. We recommend that future Arctic vegetation surveys adopt sampling methods that are compatible with the Br.-Bl. approach. These survey methods are generally compatible with the USNVC methods, and the data should be useable in classifications using either approach. With the advent of massive vegetation databases in the Arctic, both systems could be used to develop independent classifications from the same database, and evaluated regarding the efficacy of each.

6. Conclusion

Satellite-based remote-sensing data provide the means to characterize and monitor changes to Arctic tundra vegetation at circumpolar, regional, and landscape scales, but we will continue to need information collected from vegetation plots at the ground level to make sense of the spatial and temporal patterns observed from space. Although vegetation plot data are expensive to obtain, particularly in remote areas, the data and resulting classifications provide a set of operational units that are useful for description, understanding and management of vegetation and vegetation change at all scales in a rapidly changing Arctic. Moving forward with future vegetation surveys and analyses in the Arctic should build on the information collected by previous vegetation scientists, but also learn from the these previous surveys to create datasets that can be used for a wide variety of applications. For now we recommend continued collection of plot data following the Br.-Bl. protocols, mainly because the method has been used in most areas of the Arctic. We also recommend a series of international workshops to standardize plot-based observations and to begin a more focused effort to develop a truly circumpolar characterization and classification of Arctic vegetation.

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Chapter 7

ECOLOGY AND EVOLUTION OF PLANTS IN ARCTIC AND ALPINE ENVIRONMENTS

Amy L. Breen*[1], *David F. Murray*[2], *Martha K. Raynolds*[3], *Ina Timling*[3] *and Donald A. Walker*[3]

[1]International Arctic Research Center, University of Alaska, Fairbanks, Alaska USA
[2]University of Alaska Museum of the North, University of Alaska, Fairbanks, Alaska USA
[3]Institute of Arctic Biology and Alaska Geobotany Center, University of Alaska, Fairbanks, Alaska USA

ABSTRACT

The main structuring element of a terrestrial biome lies in its vegetation. Hierarchical patterns, from the level of the plant community to the global biome, are at their core a reflection of the evolutionary response of plants to their environment. These processes provide the framework for our chapter on ecology and evolution of plants in arctic and alpine environments. Arctic and alpine plants grow above latitudinal and altitudinal treelines around the world. Short-statured shrubs, herbaceous plants, lichens, and mosses comprise the low vegetation of these regions that is collectively referred to as tundra. Arctic and alpine tundras are viewed as growing in uniformly and predictably harsh environments with low temperatures, even during the growing season. The harshness attributed to the tundra, however, vastly oversimplifies the limitations plants face in these environments. The Arctic is not spatially uniform at any scale; neither is the Alpine. The arctic flora in particular, with a history that exceeds two million years, developed through multiple glacial periods. There is ample evidence of major climatic changes over millennia through which tundra vegetation has persisted despite the perceived harshness. Components of the arctic flora may be ancient, but the modern flora is an amalgam of Tertiary, Quaternary, and Holocene contributions. Herein, we focus on recent insights into the ecology and evolution of arctic and alpine plants gained from molecular ecology, modeling, and remote-sensing studies. We review the history and evolution of arctic and alpine floras and discuss the current status of arctic and alpine plant biodiversity. We then

· Corresponding author: albreen@alaska.edu

discuss the potential for arctic and alpine plants to adapt to a changing climate. We conclude with an overview of plant cross-kingdom interactions, with a focus on the plant-ectomycorrhizal fungi symbiosis in arctic and alpine environments.

INTRODUCTION

Arctic and alpine plants grow above latitudinal and altitudinal treelines around the world (Figure 1). Treeline is the limit of forest; beyond it conditions limit the growth, survival, and reproduction of trees. Short-stemmed shrubs, herbaceous plants, lichens, and mosses comprise the low vegetation of these regions. The vegetation of this treeless landscape is collectively referred to as tundra. Compact life forms are common, such as plants with dense basal rosettes or forming cushions, which protect vulnerable growing tissues from drying winds in summer and from blowing snow in winter.

Arctic and alpine tundras are viewed by some as uniformly and predictably harsh environments. Growth and productivity are constrained by the physical environment: timing of snowmelt, topography, moisture availability, exposure, and aspect. The vegetation is formed by species sufficiently tolerant of cold summer temperatures at any given location to survive freezing temperatures during the growing season, although frost-hardiness and frost-avoidance are not unique to arctic and alpine plants. The harshness attributed to the tundra, however, vastly oversimplifies the limitations plants face in these environments. To characterize tundra as harsh clearly represents our temperate zone bias (Murray, 1987). This bias makes it difficult to not view tundra plants as perilously close to the limits of life—which is simply not so. As Raup (1969) wrote:

> "…what we need is a first class Eskimo(sic) botanist—one who thinks of the tundra as a home, and a very good place to live. I think he would see the plants as they are, members of an ancient flora remarkably well adjusted to the habitat."

The Arctic is not spatially uniform at any scale; neither is the Alpine. Arctic and alpine environments are climatically variable from day to day, month to month, and year to year, yet they are predictable within limits. The Arctic flora in particular, with a history that exceeds two million years, developed through multiple glacial periods with contrasting demands imposed by the changing biological and physical environments over millennia through which tundra vegetation persisted, although the floristic composition of tundra varied over time—despite the perceived harshness.

Physiognomic similarities among the tundra regions can lead us to equate arctic and alpine environments. Dissimilarity among tundra types exists, however, notably in geographic distribution. Arctic tundra is beyond the latitudinal limit of trees in the northern hemisphere and comprises nearly 5% of the terrestrial surface of the Earth, or over 7 million km^2 (Walker et al., 2005). Approximately 5 million km^2 of the Arctic is covered by vegetation, and the remainder is covered by ice. In contrast, the Alpine is beyond the altitudinal limit of trees and comprises 3% of the terrestrial surface of the earth (Körner, 2003). Approximately 4 million km^2 of alpine tundra is scattered globally, with 82% occurring in the northern hemisphere. Plant species that occur in both the Arctic and Alpine, are designated as arctic-alpine taxa.

Our discussion here of alpine tundra is limited to the northern hemisphere as this is where most high altitude tundra occurs and where it is the most similar to the Arctic.

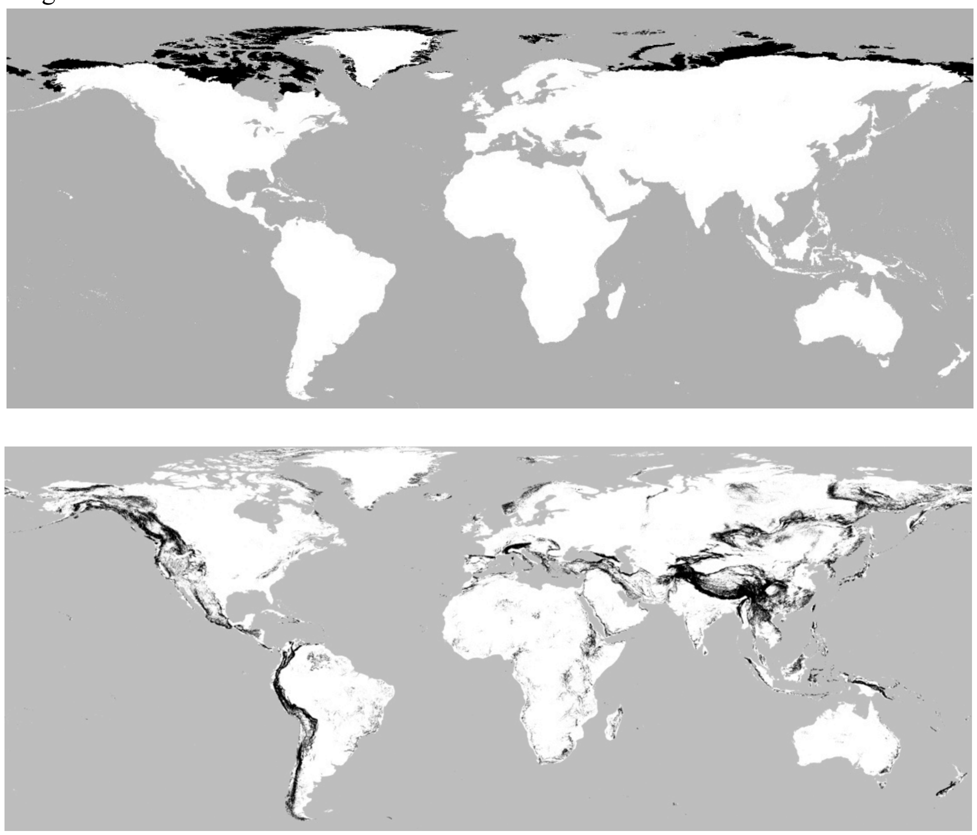


Figure 1. Global geographic distribution of arctic (top) and mountainous (bottom) regions. Alpine regions are fragmented and confined to above-treeline elevations within the mountainous regions, and thus difficult to depict at the global scale. The top panel is adapted from CAVM Team (2003), and the bottom panel is adapted from Körner et al. (2011).

Climate

The critical climatic attribute shared by arctic and alpine environments is low temperatures during the growing season. The arctic photoperiod is continuous during the growing season north of the Arctic Circle (latitude 66º 33' 44" N). However, during the continuous daylight of an arctic summer, the sun's angle remains low and the solar radiation is less intense than at lower latitudes. Mean July air temperature in the High Arctic is <6º C and can reach 10-12º C at its southern limit in the Low Arctic (Walker et al., 2005). There is a four-fold difference in length of the growing season across this gradient ranging from a few weeks to over three months.

A comparison of mean air temperature of the warmest month across alpine sites in the northern hemisphere shows a range from 5º C in the Austrian Central Alps to 8.5º C in the Rocky Mountains at Niwot Ridge in Colorado (Körner, 2003).

Plant Adaptations

The most profound limitation of the short growing season is its effect on plant reproduction. We discuss plant reproductive adaptations here and then specifically address the process of adaptation through natural selection in the *Adaptation and the Response of Arctic and Alpine Plants to Climate Change* section of this chapter. Plants must progress through anthesis, pollination, and seed set during a relatively short span of summer warmth. It is no surprise, therefore, that annuals are rare; the best example of an arctic annual being the arctic-alpine/bi-polar *Koenigia islandica* (Polygonaceae) (Jónsdóttir, 2011). The primary means to avoid this limitation is through vegetative reproduction; that is, by rhizomes (*e.g.*, graminoids), runners (*e.g.*, *Potentilla anserine* [Rosaceae], *Saxifraga flagellaris, Saxifraga platysepala* [Saxifragaceae]), bulbils (*e.g.*, *Bistorta vivipara* [Polygonaceae], *Saxifraga cernua, Saxifraga foliolosa,* viviparous grasses in *Festuca* and *Poa* [Poaceae]), or by producing seeds apomictically (*e.g.*, *Potentilla* spp.).

The majority of arctic and alpine plants can reproduce sexually, despite the prevalence of vegetative reproduction (cf. Murray, 1987), or if apomicts, may nevertheless require pollination. Self-incompatibility is rare and autogamy provides more assured seed set, but this can lead to genetic homogeneity and inbreeding depression. Mixed mating overcomes the many limitations imposed by arctic and alpine environments. Polyploidy buffers plants against the effects of inbreeding and genetic drift (Brochmann et al., 2004). An adaptation that is totally unexpected is heliotropism, the remarkable tracking of the sun by plants such as *Dryas* (Rosaceae) and *Papaver* (Papaveraceae). This occurs in conjunction with parabolic corollas (Kevin, 1972a, 1975; Wada, 1998) such that reflection of solar radiation from the inner surface of the corollas is focused on the reproductive structures thus warming them above ambient temperature and hastening development, as well as providing basking sites for insects (Hocking & Sharplin, 1965).

Advantages accrue to plants capable of producing pre-formed flower buds that overwinter surrounded by scales and leaves. These buds are developmentally advanced, in some cases up to and including meiosis, thus important steps of morphogenesis have already been completed when flowers open the following spring. Sørensen (1941) provided an excellent discussion in which he documented the wintering floral stages with photographs of meticulous dissections and cleared tissue.

Plants that are self-compatible and autogamous are more assured of seed set, although with some genetic cost through reduced recombination that accompanies inbreeding. Most outcrossers are self-compatible and through mixed-mating gain reproductive advantages. The outcrossing species are primarily wind- and insect-pollinated. Some are self-incompatible obligate outcrossers that require the mediation of insects (Kevin, 1972b). These plants offer both attraction and reward to potential pollinators. Attraction lies in flower shape and color, but “colors” not entirely within the spectra visible to humans. Among the white- and yellow-flowered taxa, so numerous in the flora, are ones with spectacular color elaborations in the ultra violet range, invisible to us but sensible to insects (Kevin, 1972c). Floral reward is typically in the form of pollen and nectar.

The relationship between plant and pollinator, attraction and reward, is so well established, co-evolved, that the loss of pollinators can limit the range of plants. Savile (1959) noted that the northern limit of Fabaceae in the islands of the eastern Canadian Arctic correlates well with the disappearance of the bumblebee. To someone accustomed to Low and

Middle Arctic floras (*sensu* Polunin, 1951), when on the ground, the absence of legumes in the High Arctic is noticed.

Vegetation Patterns

The response to summer temperatures is so consistent that the Arctic has been divided into five bioclimate subzones based largely on characteristic vegetation (CAVM Team, 2003). The bioclimate subzones are separated by approximately 2ºC in mean July temperature. Similar changes in temperatures occur with elevation in alpine areas in the Arctic, with elevation belts corresponding to the Arctic bioclimate subzones separated by approximately 333-m based on the adiabatic lapse rate of -6º C/1000 m (CAVM Team, 2003), although these subzones have been shown to shift upwards in Greenland due to a more continental climate with earlier snowmelt (Sieg & Daniëls, 2005).

In the Alpine, generally three biogeographical zones or alpine belts are recognized (Wielgolaski, 1997). The lower belt, with no trees and often with tall shrubs, is called the Low Alpine. The next belt, without shrub thickets and with a dominance of graminoids, is the Mid Alpine, although sometimes it is divided into two belts with the upper belt being referred to as Subnival. The belt of limited vegetation beyond the Mid Alpine that occurs on the highest peaks may be called either the High Alpine or Nival Belt. For a comparison of the biogeographical zones and belts of the Arctic and Alpine that pre-dates the Circumpolar Arctic Vegetation Map (CAVM Team, 2003) see Figure 1.1 in Wielgolaski (1997).

Each bioclimate subzone in the Arctic has characteristic plant growth forms. Bioclimate subzone A is the coldest part of the Arctic and includes mountain elevations closest to permanent snow cover. Most of the ground surface is barren, with only sparse vascular plant cover. What little vegetation is present grows mostly in soil cracks related to patterned ground or in sheltered areas provided by topography, where plants are protected from the wind and have a warmer microclimate. Nonvascular plants and biological soil crusts—consisting of a mixture of fungi, algae, and crustose lichens—are dominant, with a few scattered herbs (Vonlanthen et al., 2008). In bioclimate subzone B, there are a few more species of vascular plants and greater plant cover. Bare ground and biological soil crusts are still common, especially on ridges, dry hill slopes, and on the tops of hummocks. In bioclimate subzone C, the vegetation is still patchy, but covers most of the ground in flat, moist areas. Shrubs start to become an important component of the vegetation in sheltered sites. Bioclimate subzone D is mostly vegetated, with a mix of sedges, erect dwarf shrubs, forbs, lichens, and a thick layer of mosses (Kade et al., 2005). Bioclimate subzone E is adjacent to treeline and has the tallest shrubs and the most continuous vegetation cover.

Variation in plant communities at the sub-meter scale also occurs in relationship to patterned ground in most arctic and alpine areas (cf. Murray, 1997). Soil-frost processes create a range of patterned-ground features from 10- to 30-m diameter polygons with centers, rims, and troughs, to 1- to 5-m diameter frost circles and hummocks (Raynolds et al., 2008). Microhabitats associated with small differences in elevation above the water table, or differences in frost activity, are populated by different species. For example, in tussock tundra, shrubs grow on the warmer, well-drained areas, while mosses grow in the cooler, moister depressions.

Winter conditions affect plants mainly through snow and wind. Plants beneath the snow cover are protected from extreme temperatures, desiccation, abrasion, and herbivory, but can experience shorter growing seasons (Walker et al., 2001a). Taller vegetation is sometimes found in areas protected by moderately deep snow cover, but in deeper accumulations, snowbed plant productivity is strongly limited by the short snow-free period. Some evergreen species have developed the ability to photosynthesize beneath thin snow cover, giving them a head start in spring (Starr & Oberbauer, 2003). Plant communities specifically adapted to very short growing seasons are found in these snowbeds (Billings & Mooney, 1968).

Soil pH has a strong effect on arctic and alpine vegetation. Non-acidic areas in Arctic Subzones D and E are characterized by deeper thaw, non-tussock forming sedges and forbs, and frost circles with bare ground in their geomorphically active centers (Walker et al., 1998). In contrast, acidic areas have a deep moss layer, commonly including *Sphagnum* species (Sphagnaceae), which insulates the soil from summer warming. Plants growing in these moist, acidic soils include tussock sedges and ericaceous shrubs. This tussock tundra is found on old soils throughout Beringia, the vast region spanning from northeast Russia east across the Bering Land Bridge to northwest North America, that remained ice-free during Quaternary glaciations (see below). The effects of soil chemistry are especially obvious in areas of thin soil that are common in the Alpine, where plants are growing close to the source bedrock. Limestone bedrock weathers quickly and does not form soil as well as acidic bedrock, resulting in dry, calcium-rich soils, supporting vegetation that is often sparse, but forb-rich (Walker et al., 2001b).

Since the 1960s, many reviews on the topic of the ecology and evolution of plants of arctic and alpine environments have been published. In this chapter we provide a list of recommended readings by topic (Table 1) and summarize the insights gained from molecular ecology, modeling, and remote-sensing studies. We first provide an overview of the history and evolution of arctic and alpine floras and then discuss the biodiversity of arctic and alpine plants and their potential for adaptation to climate change. We conclude with an overview of plant cross-kingdom interactions, with a focus on the plant-ectomycorrhizal fungi symbiosis in arctic and alpine environments.

Table 1. Prominent reviews recommended by the authors on the topic of the ecology and evolution of arctic and alpine plants

Regional focus	Topic focus	Reference
Arctic	Adaptation	Savile (1972)
Arctic	Ecology	Chernov (1985)
Arctic and alpine	Biodiversity	Chapin & Körner (1995)
Arctic	Vegetation ecology	Bliss (2000)
Arctic	Phytogeography	Abbott & Brochmann (2003)
Alpine	Ecology	Körner (2003)
Arctic and alpine	Evolution	Abbott (2008)
Alpine	Vegetation ecology	Ellenberg (2009)
Arctic	Fungal ecology	Timling & Taylor (2012)
Arctic	Ecology & evolution	Brochmann et al. (2013)
Arctic	Biodiversity	Meltofte (2013)

HISTORY OF ARCTIC AND ALPINE FLORAS

There is a distinct arctic flora, one restricted to regions north of the latitudinal treeline, consisting of taxa that do not have ranges south of the Arctic, but including taxa with minor excursions into the northern boreal alpine zone. There are notable disjunctions south from the Arctic, as in North America, into the southern Rocky Mountains, and these comprise the arctic-alpine flora. There is also a distinct alpine flora that does not reach the Arctic, but is restricted to the southern Rocky Mountains and mountain ranges such as the Alps, Carpathians, Altai, and Caucasus.

The classic late 19th century model proposed a once widespread Tertiary arctic flora driven by advancing Pleistocene ice sheets south into high mountains, leaving nothing in their wake, the *tabula rasa* (clean slate) hypothesis (Nathorst, 1892). These southern migrants remained in the mountains and ascended to their summits when the post-glacial climate ameliorated and plants from south of the maximum extent of glaciation could migrate northward to repopulate the Arctic. Thus, the alpine flora was, by this reckoning, a Quaternary derivative of an early Tertiary arctic flora (cf. Darwin, 1859).

Weber (1965, 2003) has sought an explanation for the disjunctions of alpine plants in the Altai of south-central Siberia and in the southern Rocky Mountains of western North America. He presented abundant examples of taxa shared by both mountain systems and absent from the area between. To reconcile the huge geographic separation today, he envisioned (as did Darwin) a once more-or-less continuous blanket of these taxa at some time during late Tertiary and the subsequent destruction of these plants in the intervening area during the Quaternary. His proposal is logical and derives from inferences from long and detailed studies of floras; however, it must be said that this explanation is without empirical evidence.

Tolmachev (1960) proposed that the arctic flora had been derived from the alpine floras from the mountain ranges of Eurasia and North America. Although Hultén (1958) had earlier supposed a circumpolar arctic tundra at the onset of Quaternary glaciations, he accepted Tolmachev's hypothesis and put forward his own argument in favor of this account of history. He was aware of a common floristic core in mountain ranges surrounding the Arctic (W. A. Weber, pers. comm.). Which flora is the antecedent, arctic or alpine, is a question that remains unanswered.

Late Tertiary floras as reconstructed from plant remains at Lava Creek on the Seward Peninsula in Alaska USA (Hopkins et al., 1971) and at Kap Køpenhavn, 82° N latitude in Greenland (Bennike & Bøcher, 1990) do not provide evidence for a continuous late Tertiary arctic tundra. However, from that flora of Tertiary forests and forest-tundra, plants of bogs and similar cold sites, pre-adapted to conditions that would become widespread in the Quaternary, survived the shift from forest to tundra. Plants of pond margins and waterways faced little change in habitat as the cooling progressed. We can presume these plants persisted wherever riparian habitats remained extant (cf. Johnson & Packer, 1965). Macrofossils from the Tertiary Beaufort Formation of arctic Canada (Matthews & Ovenden, 1990) generally support this view. The occurrence of *Saxifraga oppositifolia* and *Dryas integrifolia* (Figure 2) in Canada and Greenland raise an important question: does the presence of quintessential tundra plants in today's world signify tundra in Late Tertiary?

Formation of the circumpolar arctic tundra we see today progressed throughout the nearly 2.5 to 3 million years of the Late Tertiary (Pliocene) and Quaternary. Arctic tundra reached its geographic extent, floristic richness, and current zonation (see Daniëls et al., 2013 for details) in post-glacial time. Those arctic areas wholly covered by ice sheets during the last glacial maximum, of course, were colonized as recently as 6,000 to 10,000 years ago. Components of the arctic flora may be ancient, but the modern flora is an amalgam of Tertiary, Quaternary, and Holocene contributions. As the vast continental ice sheets withdrew and eventually disappeared, plants moved onto the deglaciated terrain, a great many of them from south of the former ice margin, but patterns of plant distribution suggest also other sources, ones from within the area thought to be *tabula rasa*.

Fernald (1925) and Hultén (1937) drew our attention to areas of persistence, where plants adapted to harsh conditions avoided Quaternary glaciations in ice-free periglacial refugia when most of northern Eurasian and North America were otherwise ice covered. This meant *tabula rasa* but with special cases of plant survival. From centers of persistence, plants emerged and became geographically and ecologically sorted according to their dispersing ability and thresholds of tolerance to various abiotic and biotic limiting factors. Some plants moved faster and farther than others and established a circumpolar existence; some developed southern extensions along the Cordillera and formed the arctic-alpine flora. Others have continued to occupy restricted areas despite the millennia since their release from glacial conditions.

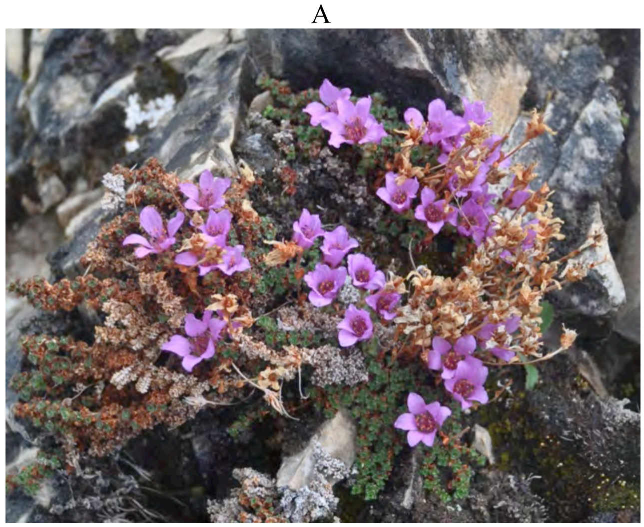
Whereas the boundaries of the huge Beringian refugium, as proposed by Hultén (1937), are now well documented by both geological and biological data, the extent and even the existence of smaller arctic and alpine refugia are still debated. Beringia is vast, but nunataks, used here in its broad sense as any non-glaciated area surrounded by glacier ice, are smaller in area, and discrete. Periglacial refugia have been used to explain numerous disjunct distributions, especially in alpine systems. Even so, questions remain: where did nunataks occur, when were they ice-free, when and how did the plants arrive at these locations, how did they survive there, and are they necessary to explain floristic novelties?

With the advent of molecular analysis of the genome in both plants and animals, and the rise of the field of phylogeography (Avise, 1994) there came an additional line of evidence by which to identify refugia, centers of phylogenetic and geographic origin, routes of migration, and instances of long distance dispersal from known sources.

More recently, information on plant cover has been gleaned from the bulk DNA extracted from frozen soil cores gathered at several sites in the Arctic: Russia, United States (Alaska), and Canada. Techniques have been developed that provide, for the most part, greater resolving power (*i.e.,* the ability to identify more taxa to species, than could be achieved through palynology alone). Importantly, these cores have been taken from exposures that date back to the last glacial maximum (cf. Willerslev et al., 2014).

Prior to molecular genetics, the thinking was that plants isolated for long periods of time in nunatak refugia would exist where, due to isolation, an influx of new genotypes was nil. Random fixation of genes by genetic drift and removal of less fit gene combinations through intense stabilizing selection acting upon these small populations would, theory predicts, result in a gene pool of low diversity but consisting of genotypes admirably adapted to the narrow constraints of a harsh, full glacial nunatak existence. This presumably left the survivors poorly equipped for post-glacial dispersal—except, perhaps, for the polyploid taxa.

A



B

Figure 2. A) *Dryas integrifolia* and B) *Saxifraga oppositifolia* were both a component of the Late Tertiary arctic flora as reconstructed from plant remains at Lava Creek on the Seward Peninsula in Alaska (Hopkins et al., 1971) and at Kap København in Greenland (Bennike & Bøcher, 1990) (photo credits: Martha Raynolds).

A tenet of phylogeography is that plant genomes undergo steady mutation in the neutral, or non-coding, regions of the genome. The longer populations are isolated, the longer the time for the fixation of unique gene combinations and rare alleles; hence, genetic identities form as long as interbreeding with other populations does not occur, as that would swamp any unique haplotypes. In phylogeography, the expectation is for greater genetic diversity as the signal of refugia persistence.

Disjunct occurrences in mountains were taken by some to be *prima facie* refugial survivors; the bicentric distribution pattern in the Scandinavian mountains is an example (Dahl, 1955). Although a thorough reconsideration by Brochmann et al. (2003) concluded that refugia were not necessary to account for both the disjunctions and endemics, a more recent study (Westergaard et al., 2011) has found examples explained by nunatak survival. Thus, these publications are a perfect illustration of the wisdom of Berg (1963):

> "…most … biogeographers explain the arctic-alpine disjunction in terms of glacial survival…It is my opinion that no single explanation can account for all the arctic-alpine disjunctions…a great deal of argumentation has resulted from a futile search for one universal cause."

The aggregate of disjunct occurrences of Rocky Mountain plants in eastern North America are what first led Fernald to propose his persistence theory (Fernald, 1925). What made some of his examples controversial was the absence of geological evidence for ice-free areas. Ives (1974), in his splendid review of biological refugia and the nunatak hypothesis, chastised those making claims for periglacial refugia without supporting evidence for full glacial, ice-free conditions, even in the face of strong geological evidence against such claims.

A counter-argument to refugial survival was that disjuncts were ecological specialists that arrived at their current position in post-glacial time. Why they remain today as small isolated populations was thought to be the result of drastically reduced ecotypes, the conservative species of Fernald (Fernald, 1925), the rigid species of Hultén (Hultén, 1937), and thus a genetically determined inability to disperse and compete elsewhere. An excellent review of Fernald and Hultén and the debate over refugial existence or post-glacial arrival is provided in Raup (1941, pt. 1).

Long distance dispersal has always been offered as a mechanism to explain disjunct species, but one which we are unlikely to confirm by direct evidence. Savile (1956, 1972), a great field biologist, believed in the efficacy of winter transport by strong winds over a landscape of ice and snow. However, for some geographic problems, greater distances must be traversed. Plant propagules are believed to have been carried across the Atlantic Ocean by migratory waterbirds such as those moving from western Europe to northeastern North America, contributing to the Amphiatlantic flora. The discussion has long gone back and forth, with reasons supporting both why long distance dispersal is probable and why it is not (Dahl, 1963; Löve, 1963).

Abbott & Brochmann (2003) have provided an excellent review of the molecular evidence for transatlantic dispersal. Since then, more examples have appeared: *Carex bigelowii* (Cyperaceae; Schönswetter et al., 2008) and *Saxifraga rivularis* (Westergaard et al., 2010). Moreover, in a remarkable study Alsos et al. (2007) demonstrated how Svalbard could be supplied with plants from elsewhere in post-glacial time, even from distant sources, without involving refugial populations—a suggestion that would have been in conflict with glacial geologists who have said that periglacial refugia did not exist there.

Mountains high enough to support alpine vegetation today were for the most part ice-covered during glacial maxima, certainly during the last glacial maximum, but alpine plants could have persisted in peripheral nunataks at the margins of an ice shield as Schönswetter et al. (2004) postulated for *Ranunculus glacialis* (Ranunculaceae) in the Alps. In the case of *Eritrichium* (Boraginaceae; Stehlik et al., 2002) at high elevations in the Alps, snow and ice would make refugia problematic. Similarly, Marr et al. (2008), having examined the genetics of *Oxyria digyna* (Polygonaceae) over much of the North America Cordillera and elsewhere, reported genetic diversity among disjunct occurrences that they interpreted as the consequence of periglacial refugia, albeit where geological evidence for ice thickness would appear to rule out ice-free areas. The implication is that genetic evidence trumps geological projections, yet the genetic diversity could be the result of post-glacial secondary contact and the rare alleles, at least in small populations, by fixed random processes.

How did the arctic species disperse southward down the Rocky Mounting chain, getting as far south as Montana, where there are about 100 arctic taxa found in the alpine zone (P. Lesica, pers. comm.)? There are even some arctic-alpine plants on the summits of the San Francisco Peaks of Arizona (Deaver Herbarium; www.nau.edu/deaver). We assume this is due to migrations southward from the Arctic, but it remains unclear when this would have occurred. During the glacial maxima, ice cover was nearly complete and thus, we presume, a barrier to dispersal. Prior to the final glacial advances and/or as glaciers receded in early post-glacial time, there would have been both the arctic environment and open corridors through which plants could have dispersed southward from Beringia (and some southern alpine plants northward). Thus a post-glacial process cannot be ruled out; in fact it seems likely. Despite numerous studies and discussions on the history and evolution of arctic and alpine floras for more than a century, there is still much to be learned.

BIODIVERSITY OF ARCTIC AND ALPINE PLANTS

Species richness of arctic and alpine plants tends to decline with increasing latitude and elevation. Low temperatures and a short growing season are environmental filters that are hypothesized to exclude species from increasingly more severe climates (Chapin & Körner 1995; Walker, 1995). There is no consensus, however, on a single explanation for the decline in biodiversity. Hypotheses fall into two groups, those based on ecological mechanisms of species co-occurrence and those based on evolutionary mechanisms governing rates of diversification and Earth history (Payer et al., 2013). These hypotheses are not necessarily mutually exclusive, as observed patterns may be due to interactions between both abiotic and biotic factors.

On a more regional scale, species richness of arctic and alpine plants is best explained by the ancestral stock of species, long-distance migration following deglaciation, evolution of new taxa, and proximity to a rich species pool as within and near Beringia (Chapin & Körner, 1995; Murray, 1995). Migration is essential for the assemblage of arctic and alpine floras, especially following glacial periods and associated extinctions. In the Arctic, the flora tends to intergrade continuously from a few centers of persistence. In contrast, alpine floras are more discrete due to their restricted habitat and geographic isolation, thereby leading to higher levels of endemism. Thus, mountain ranges in different regions tend to have disparate assemblages of alpine dominants, while the dominant plant species across the Arctic tend to have a circumpolar distribution.

Whereas it is often remarked that the flora of arctic and alpine regions is species-poor, even depauperate, the question arises: species-poor in relation to what? Summer climate is sufficiently cool and winds strong enough to preclude trees and tall shrubs, thus a major component of boreal and temperate vegetation is missing from tundra. But, are there niches unfilled? Are there families, genera, or species missing that we should expect? These questions have not been addressed, but are of interest as we discuss the flora of arctic and alpine regions in this section of the chapter.

Our knowledge of the arctic flora differs for each of the three main taxonomic groups of plants—vascular plants, bryophytes, and algae. Vascular plants are the best-known group. This is due in part to the recent publication of the *Checklist of Panarctic Flora (PAF) Vascular Plants* (Elven, 2011). The Panarctic flora includes 2,218 taxa, 91 families and 430 genera; which is less than 1% of the world flora (Daniëls et al., 2013). There are few gymnosperm taxa: 96% of the flora are angiosperms. Eight species-rich families account for more than 50% of the flora, of which the top three families are Asteraceae (254 taxa), Poaceae (224 taxa), and Cyperaceae (190 taxa). About 5%, or 106 taxa, are endemic. Most endemics are Beringian, occur arctic-wide, and are forbs. There are no endemic woody species.

As a whole, the arctic flora is viewed as taxonomically, ecologically, biologically, and genetically coherent with the many species having a circumpolar distribution. Biodiversity is low in comparison to temperate or tropical ecosystems. Trends in species richness are largely attributed to history, including glaciations, land-bridges, and north-south trending mountain ranges (Yurstev, 1994). Bryophytes are ubiquitous in the Arctic and contribute significantly to species richness, particularly in moist to wet habitats (Daniëls et al., 2013). There are an estimated 900 arctic bryophyte species and approximately 4,000 freshwater and marine algal species. The biodiversity of microalgae is still largely unknown. At present, there are few introduced species (101 taxa; Elven, 2011). The most widespread non-native stabilized introduced species are *Lepidotheca suaveolens* (Asteraceae, pineapple weed), *Plantago major* subsp. *major* (Plantaginaceae, common plantain), and *Trifolium pratense* (Fabaceae, red clover). Most of the introduced species are not invasive and are restricted to disturbed habitats. For example, hay brought in to protect disturbed slopes from erosion where the trans-Alaska oil pipeline passes through the Arctic created an influx of invasive species, but most were gone after the first winter. Although not currently a threat in the Arctic, invasive species are likely to increase due to increasing human activity coupled with climate change. For example, a recent study showed that visitors to Svalbard transport a minimum of four seeds on their shoes. Most of these seeds are from species known to be invasive elsewhere and over a quarter of these seeds were found to be capable of germination under current climatic conditions (Ware et al., 2012).

Plant species diversity of the world-wide alpine flora is much greater than in the Arctic. Körner (2003) estimates 8,000-10,000 vascular plants, comprising 100 families and about 2,000 genera, or nearly 4% of the world flora. The most common families in the Alpine are similar to those also common in the Arctic: Asteraceae, Poaceae, Brassicaceae, Caryophyllaceae, Cyperaceae, Rosaceae, and Ranunculaceae. Regional alpine floras, from the Teton Range in Wyoming to the Hokkaido alpine zone in Japan, typically include 200-280 species, with a mean diversity of 241 species from nine distinct mountain ranges (Körner, 2003). In contrast, in the Arctic, mean species richness of vascular plants from the 21 Panarctic floristic provinces (Elven, 2007) is estimated at 544 species (Daniëls et al., 2013). The most species rich floristic province is Western Alaska (825 species), and the least species rich region is Ellesmere Land-North Greenland (199 species). These data are not directly comparable to estimates of diversity for alpine floras as floristic provinces are not analogous to more regional mountain ranges. Within the alpine zone, total plant species richness within a given region declines by about 40 species of vascular plants per 100 m of elevation (Körner, 2002). Mosses (also see Chapter 12) and lichens (also see Chapter 3) deviate from this pattern as they often increase in abundance with increasing altitude, although their richness eventually decreases at the highest altitudes. Most alpine species occur at 1,000 m or lower, although a few species have been found as high as 5,900 m in the Tibetan Himalaya (Rongfu & Miehe, 1988) or 6,300 m on Mount Everest (Grabherr et al., 1995). Given the geographic isolation of mountains that often are functionally islands, endemism is high with the highest degree of endemism found at moderate, rather than at extreme altitudes.

There are several stressors to arctic biodiversity (Meltofte, 2013). These fall into two categories: anthropogenic and climatic stressors. Anthropogenic stressors include increased development, such as infrastructure associated with oil, gas, and other resource extraction. Further development will be made possible by increased opportunities for transportation including shipping lanes, road building, and regular air service to remote localities. There are also stressors from contaminants, such as persistent organic pollutants, and increased potential for oil spills.

Climatic stressors are the most serious threat to plant biodiversity in the Arctic and equally, or more so, to alpine environments. Climate warming is predicted to lead to migration of plants northward, altering the structure of vegetation through additions or even replacement from the sub-arctic to the low Arctic to the high Arctic. Terrestrial habitats in the Arctic are bounded to the north by a coastline so there is the potential that high arctic ecosystems may only survive in isolated refugia or in mountain habitats. A similar scenario is predicted for the Alpine, with expansion of treeline vegetation to higher elevations. Snowbed specialists, adapted to late snow melt and low soil temperatures are among the most threatened as both conditions are likely to be altered by climate change (Björk & Molau, 2007).

Many studies document changes in arctic and alpine plant distributions consistent with climate warming predictions. Re-sampling studies from over 100 mountains in Scandinavia and Europe, as well as on the arctic islands of Spitsbergen and Greenland, show that species richness on mountain summits has increased (Birks, 2013). This increase is predominantly an altitudinal ascent of grasses, dwarf shrubs, and low shrubs. In central Norway, Klanderud & Birks (2003) showed that changes in species richness from 1930 to 1998 varied by elevation belt. Total plant species richness in the lowest elevation belt (1,600-1,800 m) increased by 8-14 species, while in the mid-elevation belt (1,800-2,000 m) total plant species richness increased by 5-8 species. Above 2,000 m, little or no change in species richness was observed. No high-alpine species had gone extinct, although a few species had decreased in frequency since 1930. In Montana's Glacier National Park, arctic-alpine plant cover declined over two decades of study (1988-2011) with a concurrent increase in mean summer temperature (Lesica, 2014). Plants restricted to high elevations declined more so than those with a broader elevational distribution. In alpine areas of Europe, Gottfried et al. (2012) found increases in warm-adapted species and declines in cold-adapted species over a relatively short time period from 2001-2008. Warming experiments have shown an increase in shrubs in the Low Alpine in Europe (Cannone et al., 2007) and Asia (Klein et al., 2007) and from multiple sites across the Circumpolar Arctic (Elmendorf et al., 2011; Walker et al., 2006). Increases in satellite measures of greenness (related to aboveground plant biomass) have been observed (Epstein et al., 2012), as well as increases in shrub cover based on repeat photography in the warmest parts of the Arctic (*e.g.*, Tape et al., 2006), although grazing by reindeer, lemmings, and voles may limit shrub expansion (Olofsson et al., 2009). Studies in colder subzones of the Arctic have found increased vegetation cover and height, but little change in community composition (*e.g.,* Hudson & Henry, 2009), except in recently deglaciated areas where succession is occurring.

Equating biodiversity with species richness is one measure, but there is another level to be considered. From molecular studies, we now know that genetic diversity within Linnean, or biological, species can be high. The problem comes in assessing Linnean diversity, for there is often no parallel morphological differentiation to provide visible markers to genotypic boundaries. There is great genetic variation within the species (cf. Brochmann & Brysting, 2008). Reticulate evolution among arctic plants involves multiple genomes, secondary contact, hybridization, and polyploidization, all of which provide raw material for infraspecific variation and differentiation.

Some of the best information on biological species diversity comes from studies of *Draba* (Brassicaceae), initiated by Brochmann and continued by him with students and colleagues in Oslo. Grundt et al. (2006) conducted intraspecific crossing studies of three circumpolar diploid species in *Draba* and found, despite observations of limited morphological and genetic diversity, evidence for cryptic biological species, ones reproductively isolated from one another and thus evolutionarily independent. Hybrids from within populations were mostly fertile (63%), while those from within and among geographic regions (Alaska, Greenland, Svalbard, and Norway) were mostly infertile (8%). These results suggest that infraspecific diversity may be higher in the Arctic than previously realized.

Genetic diversity is essential to long-term persistence of arctic and alpine biodiversity as it provides opportunities for species to respond to changing environmental conditions. As abundance and geographic distributions of species shrink, genetic variability for selection to act upon is also often reduced. For most arctic and alpine plants, we lack information on how genetic variation, and therefore evolutionary potential, is generated and maintained. Species richness is often used as a surrogate for genetic diversity in conservation planning, although we are still learning how these two levels of biodiversity are related. To date, a few studies have addressed whether species and genetic levels of biodiversity co-vary in arctic and alpine plant communities.

Taberlet et al. (2012) showed that for the flora of the Alps and Carpathians, species richness and genetic diversity of high mountain vascular plants are not correlated. Their results showed that genetic diversity is instead associated with glacial history of a species, which in turn was linked with environmental and ecological characteristics of glacial refugia, range shifts, and associated demographic processes. In contrast, Eidesen et al. (2013) showed that patterns of genetic diversity across 17 vascular plant species are analogous to large-scale patterns of species diversity in the Arctic. Diversity was highest in Beringia and decreased gradually into more recently deglaciated regions. It should be noted that both of these studies assessed neutral genetic diversity, which is not under selection.

An aspect of genetic diversity in arctic plants was noted many decades ago as chromosome counts of northern plants were becoming known and diploids and polyploids were identified. It was further noted that there are more polyploids at higher latitudes than at low latitudes (Hagarup, 1928). The relationship between the frequency of polyploids and the northernmost regions became the preoccupation of many, for whom the underlying belief was that polyploidy *per se* gave the plants advantages in cold climates. The advantages of genetic diversity from multiple sets of chromosomes was presumed to endow polyploids with the ability to persist in the rigorous conditions, such as in glacial refugia (see above) and also to have the capacity to spread aggressively during deglaciation (Löve, 1959).

Johnson & Packer (1965, 1967) and Johnson et al. (1965) demonstrated, at Ogotoruk Creek in northwest arctic Alaska, a relationship between the frequency of polyploid taxa along gradients of soil texture, moisture and temperature, depth to permafrost, and degree of geomorphic disturbance. The diploids and low polyploids were found on more stable Tertiary surfaces, and the higher polyploids were found in habitats of the sort that became common and widespread during cold intervals of the Pleistocene, suggesting their more recent divergence.

Brochmann et al. (2004) examined the observations and explanations for polyploidy in arctic plants, particularly what can be concluded from recent molecular studies. Essentially, polyploidy is the means by which reticulate evolution proceeds and by which hybrids can gain fertility, stability, and independence. Research with hybrids showed there is *interspecific* gene flow across ploidy levels (Brochmann et al., 1992a), which demolishes the simplistic but long held belief in strong reproductive barriers between diploids and tetraploids and so-called abrupt speciation. Surprisingly, there can be two or three different parental species, all polyploids sharing parts of their genomes, which form polyphyletic hybrids. These hybrids attain fertility through polyploidization. Hence, taxa of different parental combinations, formed at different times and places, can exist within the same Linnean species (Brochmann et al., 1992b). Recent studies have shown that polyploidy has occurred at different times and places within *Vaccinium uliginosum* (Ericaceae; Eidesen et al., 2007) and that different ploidy levels overlap across the circumpolar distribution of *Saxifraga oppositifolia* (Müller et al., 2012).

Changes in biodiversity, driven by climate and other anthropogenic stressors, will provide new opportunities for recruitment and require adaptation and adjustment of arctic and alpine floras. Crawford (2008) argues that many widespread arctic and alpine plants occupy a range of different habitats, in terms of temperature and soil-moisture content for example, and are ecotypically diverse. If so, this should help buffer these species against extinction with increases in global temperatures. For other plants that are of recent origin or which are narrowly distributed, such ecotypic diversity does not exist. For species that may be outcompeted by more thermophilous species invading from the south, their survival depends on their ability to colonize newly deglaciated land at higher latitude or altitude where temperatures remain low. For alpine species that are already restricted in high altitude mountain ranges, there may be no new suitable habitat to exploit. If so, these species are likely to be among the most endangered in the future (Birks, 2008). In the next section of this chapter, we discuss adaptation and the response of arctic and alpine plants to climate change.

Adaptation and the Response of Arctic and Alpine Plants to Climate Change

Climate change in recent decades has led to changes in the composition and distribution of vegetation in arctic and alpine environments. These regions are changing, and as a consequence their biodiversity is also changing (Callaghan et al., 2004). Predicted increases in temperatures globally are 0.1°C per decade, which is amplified in the polar region compared to lower latitudes (ACIA, 2005).

In response to temperature increases, shrubs and trees are extending their limits both northwards and upwards. How will arctic and alpine plants be affected by climate change? Birks (2008) stated this question well:

> "Will arctic plants be pinched between advancing shrub tundra and forest and the rising sea-level in the low-land Arctic? Will alpine plants be squeezed off the tops of mountains?"

It is likely that some arctic and alpine plants will become extinct, particularly those with small endemic populations at the limit of plant life in the High Arctic or at high altitude. If we look to the past, however, to when the climate warmed in the early Holocene, temperatures were about 2°C warmer in the Arctic. Arctic and alpine plants persisted, and no arctic-alpine species with a fossil record is known to have gone extinct in the Quaternary (Birks, 2008). It therefore is likely that more ecotypically diverse species are resilient to climate change and will survive and adapt as long as some suitable habitat remains.

Ecotypes, variants within species, have long been recognized among temperate plants, where ecotypes show various morphological features adaptive to particular environmental conditions. A selective advantage may also accrue to ecotypes in their native site without a change in morphology as to be recognized as taxonomically distinct. There are many examples of ecotypes along latitudinal and altitudinal gradients, even along local gradients of microtopography where adaptations are less morphological and mainly physiological (Chapin & Chapin 1981; Shaver et al., 1979). Ecological amplitude in geographically wide-ranging species derives from the formation of entities with genetically fixed, adaptive properties. The effectiveness of this process is not fully appreciated. For ecotypes to undergo speciation there would first need to be sufficient genetic variation within them, and second, selection pressure to drive the process of differentiation. Absent one or both, further divergence does not occur; moreover, the infraspecific ecotypes allow for persistence across a range of environmental conditions. Raup (1969) evaluated the breadth of tolerance by species to gradients of soil moisture, plant cover, and geomorphic disturbance and found that some species exhibit great tolerances. This capacity of some tundra plants is a function either of phenotypic plasticity or of genetically fixed ecotypic differentiation, or a bit of both. It is likely that more ecotypically diverse species will have large ecological amplitudes, will be resilient to climate change, and will survive and adapt as long as the thresholds of tolerance to limiting factors are not exceeded.

Temperature, photoperiod, concentration of CO_2, and light intensity all affect photosynthesis and photosynthetic efficiency of plants. Species occurring in both arctic and alpine tundra provide examples of ecotypic differentiation for those environmental parameters. Ecotypes of these species are differentially adapted to the low light intensity and long photoperiod of the Arctic and to the high light intensity and short photoperiod of the Alpine. Even differences in the production of leaves, leaf width and thickness, and concentration of chlorophyll have been identified as part of ecotypic differentiation of physiological traits (cf. Mooney & Billings, 1961; Tieszen & Bonde, 1967).

Much of what we know about adaptation in arctic and alpine plants is based on common-garden studies as a means of identifying genetically controlled responses among plants grown in different adaptive norms. Work has ranged from the early reciprocal transplant studies of Clausen & Hiesey (1958) with *Potentilla glandulosa* and Clausen et al. (1948) with *Achillea* *lanulosa* (Asteraceae) across an elevational gradient in California, to work by Mooney & Billings (1961) with *Oxyria digyna* from a broad latitudinal range of arctic and alpine populations, to work by Tieszen & Bonde (1967) with *Deschampsia caespitosa* (Poaceae) and *Trisetum spicatum* (Poaceae) from arctic and alpine sites. The work of Clausen and colleagues revealed a sequence of climatic races. Mooney & Billings (1961) showed a clear differentiation of physiological traits in *Oxyria digyna* over a latitudinal gradient from northern Alaska south through the Rocky Mountains to Colorado. A more recent study returned to two separate reciprocal transplant experiments in Alaska 30 years later, *Dryas octopetala* subspecies along a snowbank gradient in the Alpine and *Eriophorum vaginatum* (Cyperaceae) along a latitudinal gradient in the Arctic (Bennington et al., 2012; McGraw & Antonovics, 1983; Shaver et al., 1986). For both species, differential survival in the ecotypes' native site provided strong evidence for local adaptation in these long-lived species. These findings show a broad range of ecotypes that would likely respond differently to climate change. Ultimately, the ecotypic differentiation revealed by these and other studies of arctic and alpine plants suggests extinction of wide-ranging species would be unlikely.

Just how the genes underlying genetic variation control ecotypic differentiation in arctic and alpine plants is unknown. Molecular evidence based on non-coding regions of the genome, so usefully applied in phylogeography is, however, neutral to the effects of selection. A focus on adaptive rather than neutral genetic variation will be needed for predicting responses to climate warming (Crawford, 2008). If we assume ecotypic diversity is a surrogate for adaptive genetic variation, it would seem, as discussed above, that species with high ecotypic diversity are likely to survive climate warming. To date, the genetics of adaptation have largely been studied in model organisms with short generation times and not for long-lived arctic and alpine plants.

We must note that an important distinction between arctic and alpine environments is both day length and light intensity. Phenology is often related to day length in plants. For example, arctic and alpine populations of *Oxyria digyna* show ecotypic differences in flower and rhizome production, and in growth responses, to temperature and day length (Mooney & Billings, 1961). Consequently, southern ecotypes cannot simply migrate northward to cooler temperatures in a warming climate, as day length varies from about 15 hours of solar radiation on the summer solstice at Niwot Ridge in Colorado (40º N) to continuous low intensity 24-hour solar radiation north of the Arctic Circle (>66º 33' 44" N). There are clearly limits to arctic and alpine plants escaping climate change by extending their ranges northwards and upwards.

Several recent global modeling studies have shed light on potential future states of vegetation in arctic and alpine environments. Alsos et al. (2012) analyzed range-wide genetic diversity of 27 northern plant species and used species distribution modeling to predict their future distributions and levels of genetic diversity through 2080. Their work predicts range reduction and loss of genetic diversity in nearly all species in their study, according to at least one scenario. Species that were more vulnerable to losses in genetic diversity lacked traits for long distance dispersal and had high levels of genetic differentiation among populations. In another study, Pearson et al. (2013) used ecological niche models, based on statistical associations between vegetation and climate, to predict the future distribution of arctic vegetation. Their study predicts that at least half of vegetated areas will shift to a different vegetation class, for example from graminoid tundra to shrub tundra, by 2050. Moreover, their model predicts woody plant cover, or shrub tundra and forest, will increase by as much as 52%. In contrast, Breen et al.'s (2014) regional modeling study for Alaska tundra predicts more modest shifts in woody plant cover. Their study used a state and transition model that is driven by both climate and fire dynamics. Treeline advance varies by the climate model used to drive the simulations. With greater tundra fire activity, 12% of tundra transitions to forest, and 11% of graminoid tundra transitions to shrub tundra, by 2100. In contrast, with more modest tundra fire activity, the amount of tundra that transitions to forest nearly doubles to 20%, but there is little change in the relative amounts of graminoid and shrub tundra.

Arctic and Alpine Plant Interactions with Other Organisms: The Ectomycorrhizal Symbiosis

Virtually every plant is full of endophytes (fungi, bacteria, viruses) that occur in all organs of plants. As in other ecosystems, plants in the Arctic and Alpine interact with organisms across kingdoms, including plants, animals (mammals, birds, insects), fungi, bacteria, archaea, and viruses. Many of these complex interactions, both direct and indirect, occur simultaneously. These interactions happen with different degrees of specificity and range from antagonistic to mutually beneficial. The outcome of such interactions depends in large part on the environment (Partida-Martinez & Heil, 2011), which in the Arctic and Alpine are dominated by low temperatures and a short growing season.

The use of molecular methods has not only revealed a great biodiversity of organisms in arctic and alpine environments, but also highlights the complex interactions of plants with other organisms, including fungi (Dahlberg et al., 2013; Gao & Yang, 2010; Timling & Taylor, 2012). Fungi are ubiquitous and benefit plants as mutualistic mycorrhizas (also see Chapters 2, 5) and saprotrophs by providing nutrients and water; they can harm plants as pathogens. We will illustrate how molecular methods have shed light on plant interactions with other organisms through the example of ectomycorrhizal fungi (EMF).

The ectomycorrhizal symbiosis is abundant throughout the Arctic and Alpine, where the fungi associate with shrubs, as well as a few sedges and forbs. Although EMF associate with only about 6% of the vascular plants in the Arctic, these plants are important components of plant communities that cover up to 69% of the ice-free Arctic (Walker et al., 2005). The symbiosis seems especially important in biomes with low nutrient availability, where the fungus provides nutrients and water to the plant and the plant provides carbohydrates to the fungus. In the Arctic, 61-86% of nitrogen in ectomycorrhizal plants is provided by their fungal symbionts while the plant provides 8-17% of photosynthetic carbon to the fungi (Hobbie & Hobbie, 2006).

In contrast to vascular plants in the Arctic, EMF associated with shrubs do not follow the classic pattern of species richness decline with latitude, which suggests that fungal species richness is not governed by temperature (Bjorbaekmo et al., 2010; Timling et al., 2012). The species-rich EMF communities that have been observed on host plants in the Arctic and Alpine are dominated by a few families that are especially species-rich (Thelephoraceae, Cortinariaceae, Inocybaceae) (Blaalid et al., 2014; Gao & Yang, 2010). Similarly, many plant communities are dominated by a few species-rich families (Asteraceae, Brassicaceae, Caryophyllaceae, Cyperaceae, Fabaceae, Poaceae, Ranunculaceae, and Rosaceae; Daniëls et al., 2013). This suggests that some plant and fungal families are especially adapted to arctic and alpine environments.

Furthermore, EMF communities in the Arctic appear to be dominated by generalist fungi with wide ecological amplitudes and which are excellent dispersers (Geml et al., 2012; Timling et al., 2012). In contrast to boreal (Taylor et al., 2010) and temperate forests (Ishida et al., 2007) and Mediterranean woodlands (Morris et al., 2008), the ectomycorrhizal symbiosis seems to have lower specificity in the Arctic and Alpine (Botnen et al., 2014; Gao & Yang 2010; Timling et al., 2012). While EMF communities in boreal, temperate and tropical climates show distinctive phylogeographic distribution patterns, with restrictions to continents and sub-continental regions (Geml et al., 2008; Talbot et al., 2014), the majority (73%) of EMF species observed in studies from Svalbard and across the entire bioclimatic gradient of North American Arctic have occurred also in other regions within and beyond the Arctic (Geml et al., 2012; Timling et al., 2012). Such wide distributions within the Arctic have been also observed for lichens (Geml et al., 2010) and vascular plants (Alsos et al., 2007). The wide distribution of fungi and lichens might be aided by the smaller size of their propagules. Finally the wide distribution suggests that terrestrial and trans-ocean long distance dispersal must be a common phenomenon in the wide open landscapes of the Arctic, aided by wind, snow, driftwood, sea ice, birds, and mammals (reviewed in Alsos et al., 2007).

Nevertheless, within the Arctic and Alpine, EMF communities show distribution patterns at the regional and local scale that often parallel those of plant communities found there. Ectomycorrhizal fungal communities associated with *Dryas integrifolia* and *Salix arctica* (Salicaceae) change gradually across the five bioclimatic subzones of the North American Arctic, corresponding with climate, plant productivity, glaciation history, geology, and soil factors (Timling et al., 2012). At a local scale, EMF communities often correlate with habitat, successional stage, plant community, and bedrock and edaphic factors such as pH, carbon, and nitrogen (Blaalid et al., 2014; Fujimura & Egger, 2012; Yao et al., 2013; Zinger et al., 2011).

Climatic changes in the Arctic have led to pan-arctic shrub expansion (Tape et al., 2006) and increases in plant productivity (Bhatt et al., 2010) and nutrient cycling (Rustad et al., 2001). Long-term warming experiments show not only changes in plant communities (Walker et al., 2006) but also changes in EMF community structure associated with *Betula nana* (Betulaceae), one of the shrubs most responsive to climate warming (Deslippe et al., 2011). The mutualistic nature of the ectomycorrhizal symbiosis, the low host specificity and the wide distribution support the idea that EMF may play critical roles in the expansion of shrubs in the tundra. Evidence from past climate changes comes from paleobotanical studies which show that plant and fungal communities changed with past glacial and interglacial cycles, with an increase in shrubs and trees and their ectomycorrhizal symbionts since the last glaciation (de Vernal & Hillaire-Marcel 2008; Lydolph et al., 2005). Soil analyses along a bioclimatic gradient in the North American Arctic show that subzone A, which is devoid of woody species, harbors EMF species, probably as spores, and that soil fungal communities in subzone E greatly overlap (74%) with communities of the boreal forests (Timling et al., 2014). Furthermore, a bioassay with soils from above treeline showed that these soils provide sufficient inoculum for a significant growth of conifers (Reithmeier & Kernaghan, 2013). The authors concluded that spores in the soils and shrubs above treeline could facilitate treeline expansion. With a warming climate one might expect changes of EMF community composition with a northward shift of some EMF taxa. Finally EMF may be critical in facilitating an establishment of woody species in subzone A and a treeline expansion into subzone E.

CONCLUSION

Despite considerable progress made in recent years, there remains much to learn about the ecology and evolution plants in arctic and alpine environments. Molecular ecology, modeling, and remote sensing studies, along with future fossil discoveries, will continue to build upon our knowledge of these biomes and improve our understanding of their potential response to future climate change. Brochmann et al. (2013) write that the species-poor arctic flora is likely to be adapted to environmental change, through selection for high mobility and buffering against inbreeding- and bottleneck-induced gene loss via polyploidy. However, today's flora of arctic and alpine environments will be challenged by a climate warmer than the Holocene and over a shorter period. There is a need to begin focusing on adaptive, rather than neutral genetic variation, to predict how arctic and alpine plants will respond to climate warming over the next century.

There is also a need to intensify biodiversity research on arctic and alpine floras, with an emphasis on vegetation classification, monitoring, and modeling (Daniëls et al., 2013). Efforts such as the Arctic Vegetation Archive (Walker et al., 2013) are underway to improve coordination and cooperation among arctic nations and to produce a pan-arctic vegetation classification. Furthermore, the archive will provide vegetation data from across the Circumpolar Arctic for use in biodiversity and ecosystem models. Jónsdóttir (2013) is also leading an initiative to develop a research framework on biodiversity-shaping forces that considers different spatial and temporal scales and identifies commonalities across biological hierarchies and organisms. This framework will provide for testing hypotheses about biodiversity trends in the face of climate change in the Arctic and Alpine.

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1

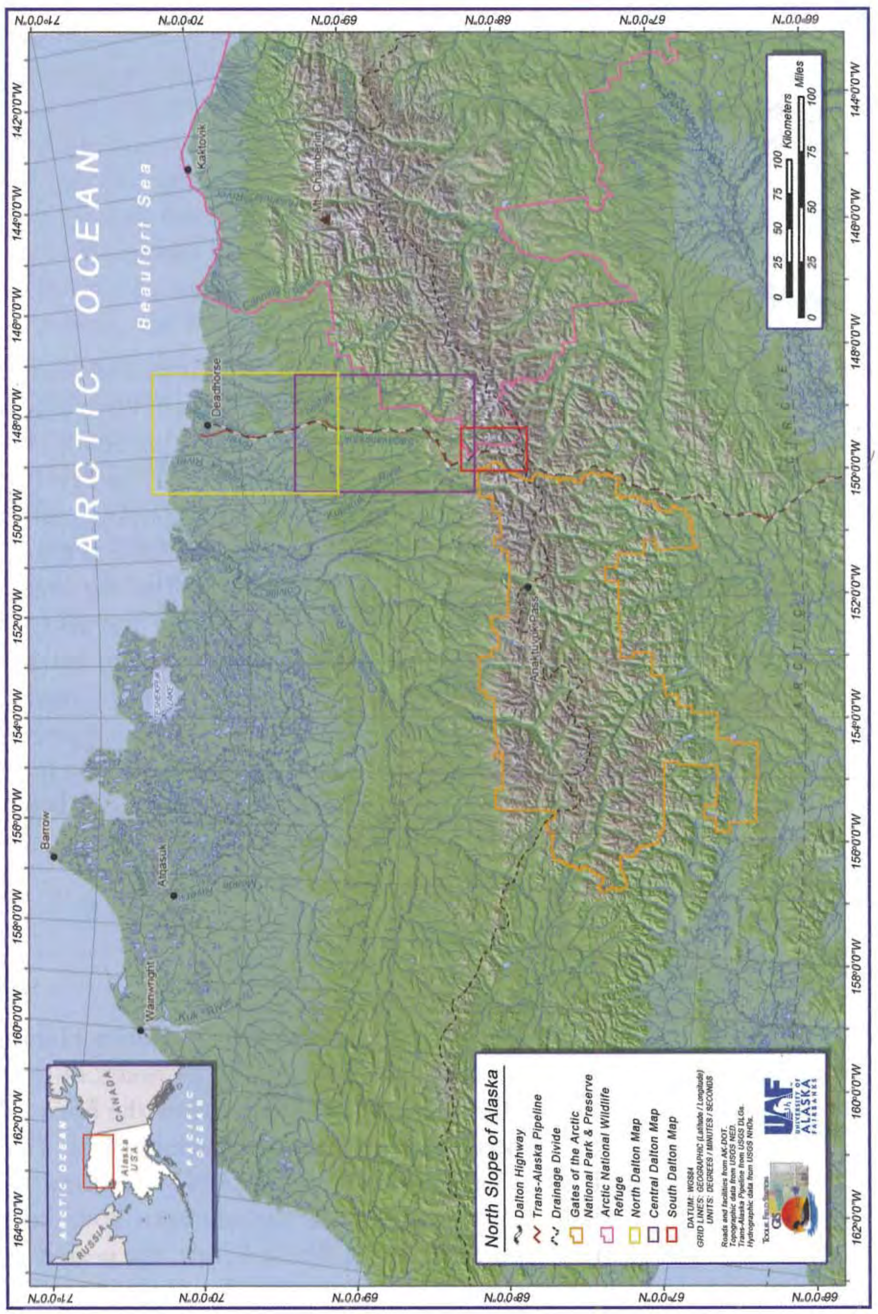
Introduction

The North Slope

The North Slope, also known as the Arctic Slope, is that part of northern Alaska where rivers drain into the Arctic Ocean north of Point Hope (Map 1). It is enormous, with an area equal to Nebraska or South Dakota (about 200,000 km^2) and extending from about 68°N to 71°N at its greatest width. The North Slope contains three major physiographic regions: the Brooks Range, Arctic Foothills, and Arctic Coastal Plain. These are arranged as east–west trending bands parallel to the arctic coast, with the Arctic Coastal Plain most northerly and the Brooks Range most southerly. Each region has characteristic plant and animal communities due to differences in geology, topography, and climate. Nevertheless, they are all considered tundra. There are two types of tundra. Alpine tundra refers to mountain habitats *above* the tree line. Arctic tundra refers to habitats *beyond* the northern tree line. This book is an introduction to the natural history of the North Slope, the only arctic tundra in the United States.

Why the "Arctic"?

The Arctic is precisely defined as that part of the Northern Hemisphere where the sun is visible above the horizon for 24 hours during the summer solstice (around June 21) and is hidden below the horizon for 24 hours during the winter solstice (around December 22). The lowest latitude at which this occurs is about 66°33′N, which marks the position of the Arctic Circle and delimits the southern boundary of the Arctic. As one moves north from the Arctic Circle, the period of continuous daylight ("polar day") or darkness ("polar night") lengthens. At the southern North Slope village of Anaktuvuk Pass (68°8′N) in 2010, for example, the sun rose on May 25 and did not set again



Map 1. North Slope of Alaska.

until July 18 (a 54-day polar day), while the polar day for the northern North Slope town of Barrow (71°17′N) lasted from May 11 to August 1 (82 days). Conversely, the polar night lasted from December 7, 2010, to January 4, 2011 (28 days), at Anaktuvuk Pass and from November 19, 2010, to January 21, 2011, at Barrow (63 days). Those unfamiliar with the relatively low arctic latitudes of the North Slope may have the impression that polar day and night are relatively constant periods of light and darkness. This is not so. During the polar day there are noticeable declines in light and temperature even on the summer solstice because the sun strikes the ground at a low angle during the early morning hours. During the polar night, even on the winter solstice, there is sufficient midday twilight to perform outdoor chores without a headlamp or lantern.

Having provided this pleasingly precise definition of the Arctic, it is important to point out its shortcomings. First, its precision is illusory. The Arctic Circle is not fixed but varies over about 2° during 40,000-year cycles caused by wobbles in the angle of the Earth's axis. Consequently, the location of the famous monument marking where the Dalton Highway crosses the Arctic Circle is only an approximation. Second, it does not adequately delimit the distribution of ecosystems containing communities of organisms adapted to arctic conditions. For example, no biologist would argue that the western shore of the Hudson Bay near Churchill, Manitoba, is not a typical arctic ecosystem, complete with polar bears, arctic foxes, and collared lemmings. Yet Churchill (58°45′N) is clearly south of the Arctic Circle and thus experiences no midnight sun. To accommodate such discrepancies, it has been suggested that the Arctic be defined to include areas of the Northern Hemisphere that have mean July temperatures of 10°C or less. The "10°C July mean-temperature rule" roughly determines the northern limit of tree growth[1] and the southern limit of continuous permafrost.[2] This definition is ecologically sound because it is based on a single key attribute that underlies the structure and function of all arctic communities: a long annual period of deep, dark cold.

Low Arctic versus High Arctic

The Arctic is often subdivided into the "high Arctic" and "low Arctic." The high Arctic includes habitats relatively close to the North Pole (e.g., greater than 75°N). With the exception of limited sedge meadows near streams and below lasting snowbanks, these are best characterized as rocky barrens populated by relatively few plant species (e.g., fewer than

150 vascular plant species). The low Arctic includes habitats closer to the Arctic Circle. These generally have comparatively lush vegetation and high plant diversity (more than 250 vascular plant species). Although all this may seem to be an exercise in hair splitting, it is important to be aware of the large differences between these subdivisions because much of what has been written about arctic ecosystems is based on the high Arctic and may not directly apply to the low Arctic. The North Slope, the focus of this book, provides an excellent example of a low-arctic ecosystem.

Climate

Temperature

The average annual temperature of the North Slope is about –12°C. The warmest month is July (mean near arctic coast = 5 to 8°C, mean in foothills = 12 to 13°C). The coldest is February (mean near arctic coast = –29 to –27°C; mean in foothills = –30°C) because the Arctic Ocean becomes completely covered by ice at this time, dramatically decreasing the transfer of heat from the relatively warm ocean. The major variation in temperatures across the North Slope is related to the distance south of the arctic coast. Coastal regions have warmer winters and colder summers, and interior locations have colder winters and warmer summers. For example, long-term climate records show average July high temperatures of 8°C for Barrow (coastal) to 19°C in Umiat (interior) and average February low temperatures ranging from –30°C at Barrow and –35°C at Umiat. To put all this in perspective, the summer freeze-free period at inland locations can be as much as 30 days; the freeze-free period at Barrow, however, is only 10 days. Needless to say, summers are short everywhere on the North Slope, a place where snowfall can be expected any day of the year!

Precipitation

Precipitation on the North Slope is highly influenced by seasonal patterns of freezing of the Beaufort, Chuchki, and Bering Seas. As continuous sea ice develops, atmospheric moisture is reduced and, as a consequence, precipitation on the North Slope is low from November through April. On the other hand, the lack of near-shore sea ice results in relatively high amounts of precipitation during July and August. Annual precipitation measured on the North Slope ranges from a minimum of 150 mm/yr near Barrow to 550 mm/yr in portions of the Brooks Range (North Slope annual average = 250 mm, or about 10 inches). There are problems with these estimates due to the difficulty of sampling blowing snow, but they nevertheless indicate low precipitation by any measure. To provide perspective, a common definition of a desert is a region receiving less than 250 mm of precipitation annually. By this definition, the North Slope is surely close to a desert—but a very strange one. During summer, the soils of much of the foothills and coastal plain are water saturated and often covered by thick, spongy layers of sphagnum moss. This is because the shallow layer of soil that thaws every summer—the "active layer" (0.3–1.0 m deep, or about 12–39 inches deep)—is sealed by the continuous layer of watertight permafrost below. In low-lying, poorly drained habitats, the water released as the active layer thaws simply has no place to go! The occurrence of water-saturated soils is further facilitated by low rates of evaporation and transpiration (water loss through plant tissues) due to cool summer temperatures. As a consequence of this conspiracy between a shallow active layer, a continuous layer of underlying permafrost, and cool summer temperatures, pools of standing water, extensive wetlands, ponds, lakes, rivers, and streams are all common features of the North Slope landscape. These features defy traditional definitions of a desert, to say the least.

Seasonality

It has been suggested that the traditional concept of the four seasons—winter, spring, summer, autumn—is not useful for the Arctic. There is good reason for this because the calendar dates that define the seasons of temperate regions are of little relevance here. The arctic year can be divided more usefully into a short warm period and a long cold period. The warm period is delimited by the consecutive days centered on mid-July when the *mean* daily temperature is above freezing (usually 10 to 30 days, although frost may occur on any day). The concepts of spring and autumn are reduced to brief seasons ("spring" from mid-May through mid-June, and "autumn" from mid-August through mid-September). The cold period begins in September when soils and surface water freeze ("freezeup") and lasts until they thaw ("breakup") in May. This concept of a "warm season" and a "cold season" makes sense ecologically and is similar to the concept of tropical seasonality, where the year is divided into monsoon and dry seasons, rather than the traditional four seasons.

Snow

The North Slope is blanketed with snow for almost nine months each year. Consequently, the ecology of plants and nonmigratory animals living here cannot be fully appreciated without some knowledge as to how they are affected by this. Snow performs four major ecological roles. First and foremost it provides *insulation* that maintains moderate soil temperatures even in the depth of winter. Second, high *humidity* within the snowpack reduces desiccating effects of dry winter air on buried plants and animals. Third, it provides critical winter *habitat* for small mammals (e.g., lemmings and voles) while simultaneously hampering foraging by the burial of food sources used by some larger animals (e.g., Dall sheep, caribou) and birds. Finally—although this role occurs only during the spring thaw—the melting snowpack provides a spatially variable water supply that is critical in determining the distribution of tundra plant species.

Insulation

Snow provides excellent insulation. Once a snow layer reaches a depth of 20–80 cm (about 8–31 inches), an uncoupling of air and ground temperatures occurs.[3] The depth at which this occurs is called the "hiemal threshold." The wide range in snow depths at which the hiemal threshold occurs is due to differences in the amount of insulation provided by different types of snow. Low-density "new snow" (about 0.1 g/cm^3) provides the highest insulation because it contains many dead-air spaces; high-density "old snow" (about 0.4 g/cm^3) provides less insulation due to few dead-air spaces. As new snow becomes old snow a predictable process of structural metamorphism occurs, resulting in the difference in dead-air volume. As snow falls, it forms a layer of the familiar lacy snowflakes (new snow). Within a short time (hours to days depending upon factors such as snow depth, temperature, and wind packing), the snowflakes are compacted into a layer of tightly packed grains (old snow) and the volume of trapped dead air declines while the snow's strength and ability to support weight increase. Regardless of age or state of metamorphism, the insulation value of a snow layer approaching 40–50 cm (about 16–20 inches) is sufficient to maintain relatively constant ground temperatures. When insulated by 50 cm or more of winter snow, ground surface temperatures in the foothills, for example, rarely fall below –4 to –10°C (about 25–14°F) even though air temperatures of –30°C (about –22°F) or less are frequent and long lasting.[4] The insulating effect of snow across the vast distances of the North Slope varies because the winter snowpack is relatively deep inland but relatively shallow near the coast due to wind. Consequently, mean winter ground surface temperatures range from about −6°C in the southern foothills to −20°C on the northern coastal plain.

Humidity

A thick snow layer maintains relatively high levels of internal humidity. Following the relatively rapid conversion of new snow to old snow, further structural changes occur that are related to the temperature gradient that forms within the snow layer. The top of the snow layer, which is in contact with the atmosphere, is usually colder than the bottom, which is in contact with the ground. The relatively warm temperature of the ground causes a high rate of sublimation[5] of the ice grains in the bottom of the snow layer. The water vapor produced contributes to high levels of relative humidity (e.g., 100 percent) within the air spaces of the snowpack. Such high levels of humidity greatly enhance the ability of buried plants to avoid desiccation during winter.

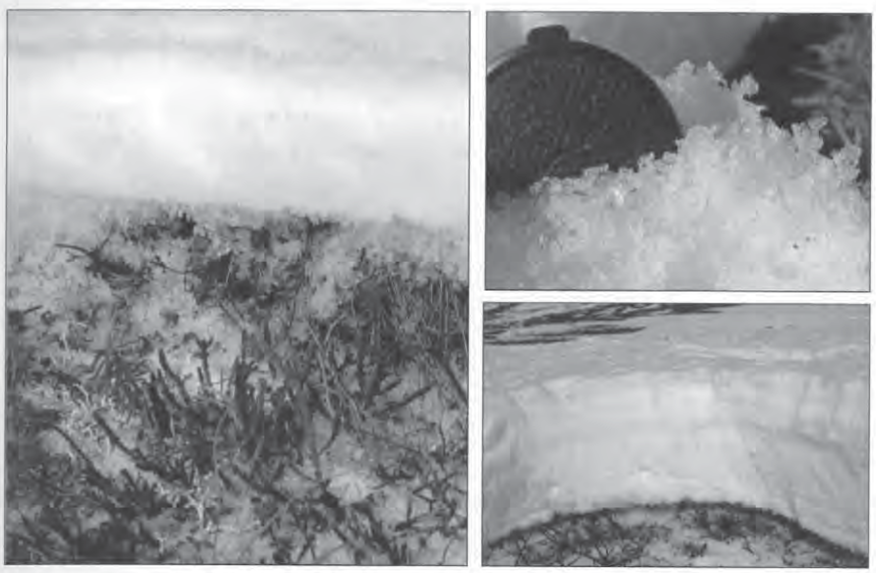


Fig. 1.1. The easily navigated space provided by the depth hoar layer at the base of the winter snowpack is used for habitat by lemmings, voles, shrews, and weasels. ***Left:*** Detail of depth hoar layer (about 5 cm deep) showing the loose, nonadhesive crystals formed by vapor produced by sublimation of the snow layer adjacent to the ground (Dalton Highway, MP 355, ADH). ***Upper right:*** Detail of ice crystals forming depth hoar (Atigun Pass, ADH). ***Lower right:*** A 4 cm deep layer of depth hoar formed at the base of a 33 cm snowpack (Atigun Pass, ADH). The depth hoar layer was lightly brushed to remove the loosely packed crystal layer to better show its extent and location.

Habitat

A thick snow layer provides space near the ground that allows construction of runways and nests by voles, lemmings, shrews, and weasels. This space results from the redistribution of water from the bottom of the snow layer to the top. This process is driven by the diffusion of vapor produced by sublimation at the bottom of the snow layer to the colder top of the snow layer. Here it refreezes and becomes incorporated into a "snow-ice matrix." The snow grains near the top of the snowpack thus increase in size and density at the expense of snow grains near the bottom. As a consequence the bottom of the snow layer is converted to a relatively open matrix of large, brittle, nonadhesive crystals known as "depth hoar" (Fig. 1.1). It is the open space provided by depth hoar that is used as habitat by lemmings, voles, shrews, and weasels, all of which remain active beneath the snow during winter.

2

Bedrock Geology

Suspect Terrane?

Most large landmasses are formed by the fusion of distinct blocks of continental crust ("terranes") with different geographical origins. Consequently, each terrane has a unique combination of rocks and fossils.[1] Alaska consists of at least 13 terranes, with one of the largest being the North Slope. The North Slope is considered a "suspect terrane" because it was joined to the northern Alaskan landmass after migrating from an uncertain ("suspect") location. Information based on the fossil record, however, indicates that the North Slope terrane was always part of North America—it just changed location. The most credible explanation of how the present-day North Slope came to be involves the rifting (splitting) of a large chunk of land from northernmost Canada, followed by a jackknife rotation centered near the Mackenzie River delta, and then a final collision with the north coast of ancestral Alaska. The rifting and subsequent rotation was caused by seafloor spreading during the Early Cretaceous (145–112 million years ago, or mya). Consequently, much of the very old geologic features of the North Slope, including the deposition of the oil-producing strata of the Prudhoe Bay oil fields, occurred while it was a part of present-day northern Canada. Following the joining of the North Slope terrane with ancestral Alaska, a long period of north–south contraction resulted in the uplift of the modern Brooks Range and the shedding of enormous volumes of eroded materials toward the north that eventually covered much of the foothills and coastal plain.

Bedrock of the North Slope

The bedrock of the North Slope terrane is primarily sedimentary in origin. In many places it has been distorted by overthrusting[2] and uplifting. This began with the earliest stirrings of the Brooks Range during the Middle Jurassic (202–146 mya, i.e., prior to the migration of the North Slope terrane away from northern Canada) and continued through the Early Cretaceous (146–100 mya) and the Quaternary (less than 1.8 mya). The bedrock of the present-day Brooks Range is often exposed and readily observed. In the foothills and coastal plain, however, it is usually concealed by sediment layers, or more technically "strata," thousands of meters thick. Although the local bedrock geology of the North Slope can be exceedingly complex due to extreme displacement and distortion, the regional pattern is simple. As one ventures from the Continental Divide toward the Arctic Ocean, the successive exposures of bedrock become younger due to northward dipping of bedrock.[3] To simplify this overview of the geology of the North Slope, we will restrict ourselves to bedrock types observed along the convenient cross section provided by the Dalton Highway. These major bedrock types are presented in order of age, from oldest to youngest, beginning at Atigun Pass.

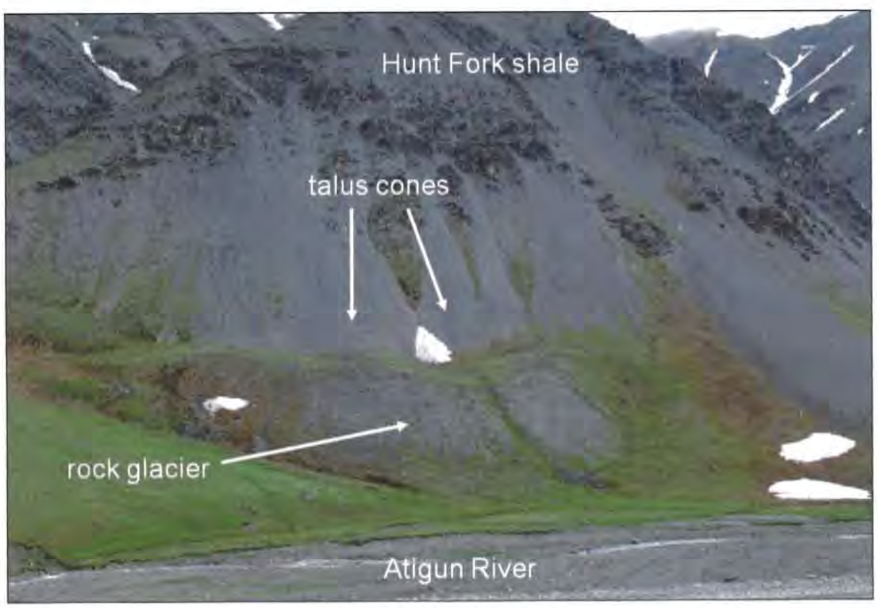


Fig. 2.1. Hunt Fork Shale (Dalton Highway, MP 248). Note the ice-cored rock glacier and the talus cones. The symmetry of the steep talus cones is maintained by ice, which functions as cement. Both rock glaciers and steep, symmetrical talus cones are typical permafrost features of the Brooks Range (ADH).

Hunt Fork Shale[4] and Kanayut Conglomerate

Two major types of bedrock dominate the landscape from Atigun Pass to Atigun Gorge. The Hunt Fork Shale (Upper Devonian, 385 mya, Figs. 2.1, 2.2) is a thinly foliated, dark brown to black shale averaging more than 1,000 m thick. It was deposited in a shallow marine habitat and now forms many of the interior peaks of the Brooks Range. Its presence is indicated by distinctive steep slopes with symmetrical talus[5] cones that occasionally terminate in ice-cored rock glaciers[6] (Fig. 2.1). Slopes formed by the Hunt Fork Shale have a distinctive shining appearance when viewed from a distance on a sunny day.

Like the Hunt Fork Shale, the Kanayut Conglomerate (Upper Devonian and Lower Mississippian, 385–345 mya, Fig. 2.3) forms many of the interior peaks of the Brooks Range. Conglomerates, or "pudding stones"—so named because they resemble fruitcake—are fine-grained sandstones containing "clasts," or rock particles, of various types and origins. The Kanayut Conglomerate averages 2,600 m in depth and contains clasts of chert (a fine-grained type of rock consisting primarily of tiny quartz crystals) and quartz that may be as large as 23 cm in diameter, indicating that the depositional environment

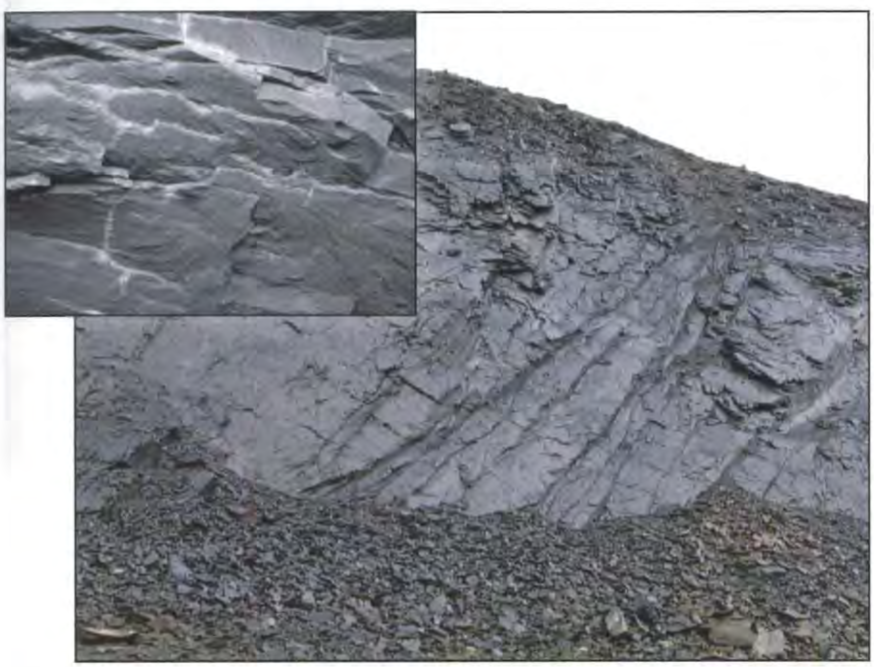


Fig. 2.2. Road cut showing laminated Hunt Fork Shale. Inset shows detail of shale (Dalton Highway, MP 248.4, ADH).

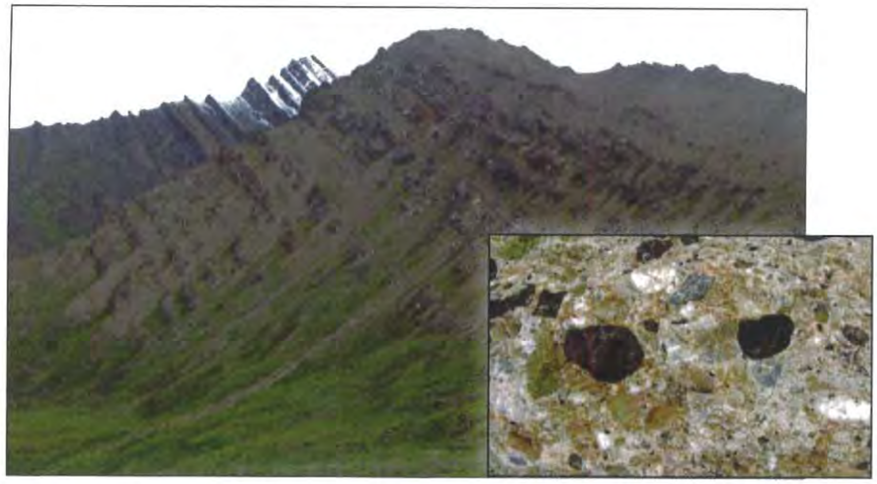


Fig. 2.3. Southward-dipping bed of Kanayut Conglomerate (Dalton Highway, MP 259, ADH). Dark layers of the erosion-resistant conglomerate are shown in relief against the lighter, readily eroded shale. Inset shows detail of conglomerate with distinctive chert clasts.

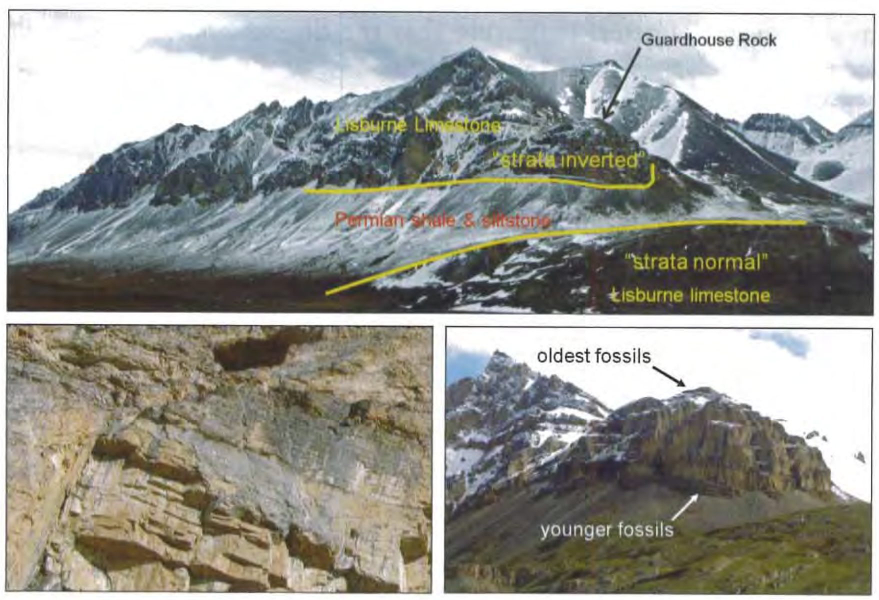


Fig. 2.4. ***Top:*** Cliffs of Lisburne Limestone near Atigun Gorge (Dalton Highway, MP 270, ADH). Two major layers of Lisburne Limestone can be seen. Both layers have identical stratigraphy. The lower layer ("strata normal") is in its original position. The upper layer ("strata inverted") has been overturned due to folding. ***Lower left:*** Detail of Lisburne Limestone cliff showing typical yellowish color when weathered (Guardhouse Rock, ADH). ***Lower right:*** Guardhouse Rock overlooks the Dalton Highway near Atigun Gorge (ADH). This feature is formed from an overturned layer of Lisburne Limestone. Consequently, the fossils of its summit are older than those of its base.

was a broad alluvial plain with numerous braided rivers that transported rock debris from an ancient mountain range. Compared with other regional rock types, it is both readily identifiable and extremely erosion resistant.

Lisburne Limestone

The Lisburne Limestone (Upper Mississippian through Upper Permian, 326–254 mya, Fig. 2.4) is a 600 m or more thick layer of limestone (calcium carbonate–based bedrock), dolomite (magnesium carbonate–based bedrock), and chert that underlies much of the North Slope. It contains abundant fossils of bryozoans, crinoids, corals, brachiopods, foraminiferans, and algae, indicating that it was deposited in a shallow marine environment (Fig. 2.5). In the cold and dry climate of the Brooks Range it is also resistant to erosion and is an important cliff-forming rock. The southern walls of the Atigun Gorge, for example, are formed by towering gray to yellowish-gray cliffs of Lisburne Limestone that reach as high as 700 m (Fig. 2.4). During the overthrusting of the bedrock strata that now comprise the northern front of the Brooks Range, its great tensile strength allowed it to withstand extreme distortion and

Fig. 2.5. Invertebrate fossils from Lisburne Limestone. ***Upper left:*** Mollusk shells (base of Guardhouse Rock, H.M. Rantala). ***Upper right:*** Brachiopod (base of Guardhouse Rock, ADH). ***Lower left:*** Coral (Kuparuk River, H.M. Rantala). ***Lower right:*** Crinoid stem (summit of Guardhouse Rock, ADH).

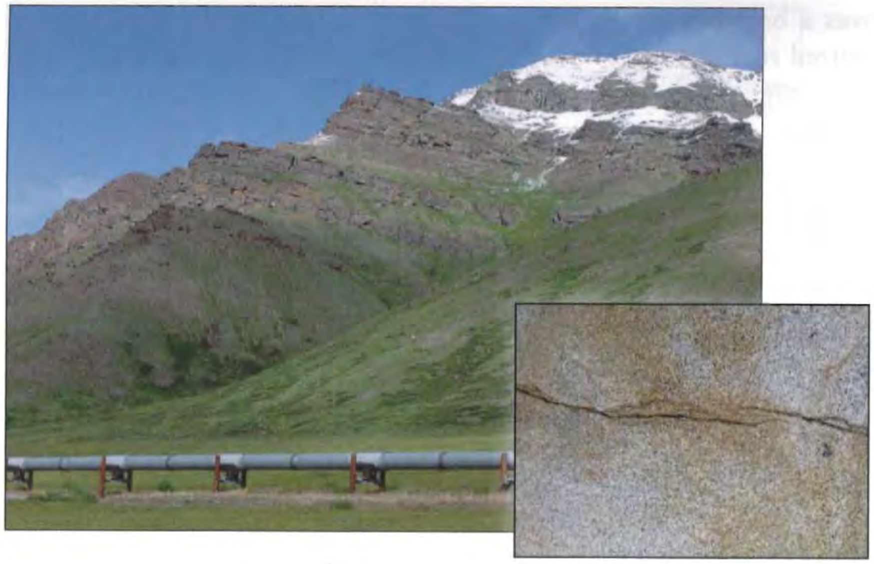


Fig. 2.6. Sandstone and shale of Fortress Mountain Formation near Atigun Gorge (Dalton Highway, MP 273, ADH). Inset shows detail of sandstone.

Fig. 2.7. ***Left:*** Aerial view of Slope Mountain and Imnavait Mountain showing well-defined bedrock strata (ADH). Slope Mountain consists primarily of sandstones of the Nanushuk Group. The upper layers were deposited in a freshwater river environment. The lower layers were deposited in a marine environment. ***Upper right:*** View of Slope Mountain from Dalton Highway (ADH). ***Lower right:*** Outcrop of Prince Creek bedrock (Dalton Highway, MP 352, ADH). Exposures of the Prince Creek Formation along the Colville River contain abundant dinosaur fossils.

Fig. 2.8. Franklin Bluffs (Dalton Highway, MP 385, ADH). The Franklin Bluffs consist of soft siltstone that weathers into distinctly rounded, pastel-colored landforms. The sediments forming this siltstone were deposited about 43–56 mya in a protected delta or lake environment.

folding while maintaining structural integrity. In some cases it was folded to such an extent that adjacent layers are, in reality, stratigraphical mirror images[7] (Fig. 2.4).

Fortress Mountain Formation[8]

North of Atigun Gorge, the bedrock geology changes abruptly from Lisburne Limestone to the Fortress Mountain Formation and the related Nanushuk Group (Early Cretaceous, 122–100 mya, Figs. 2.6, 2.7) that form the upper strata of large portions of the major mountains to the north (e.g., Imnavait and Slope Mountains, MP 301). The Fortress Mountain Formation consists of about 3,000 m of dark shale, sandstone, and conglomerate that was deposited in coastal deltas. Unlike the preceding bedrock types, the Fortress Mountain Formation and Nanushuk Group were deposited after the North Slope terrane joined present-day Alaska.

Prince Creek and Sagavanirktok Formations

As one proceeds north of Slope Mountain in the central foothills, outcrops of bedrock are few because they are usually buried by deep Quaternary (less than 1.8 mya) sediments. The porous sandstone and fractured limestone strata that are "reservoir rocks" for the oil reserves of the northern coastal plain, for example, are buried at average depths exceeding 2,500 m. There are a few exceptions, however. The Prince Creek Formation (Late Cretaceous, 86–66 mya, Fig. 2.7) is a 600 m thick layer of nonmarine sandstone, conglomerate, shale, and coal that underlies portions of the central North Slope. Outcrops occur along the Dalton Highway in the Sagwon Uplands (MP 353). Exposures of the Prince Creek Formation along the lower Colville River have revealed dinosaur, fish, turtle, mammal, and plant fossils ranging in age from 66 to 76 mya. The Franklin Bluffs (Fig. 2.8), a spectacular outcrop of the Sagavanirktok Formation (Late Cretaceous to Late Miocene, 84–3 mya), is the youngest significant exposure of bedrock along the Dalton Highway. The Sagavanirktok Formation is as much as 2,600 m thick and consists of sandstone, soft siltstone, conglomerate, and coal deposited in coastal marine and freshwater habitats. The Franklin Bluffs were formed by erosion of outcrops of the Sagavanirktok Formation along the eastern bank of the Sagavanirktok River. They are composed of a soft siltstone that weathers into distinctly rounded pastel-colored landforms. The sediments comprising this siltstone were deposited in a protected delta or lake environment during the Eocene (56–34 mya).

3

Glacial Geology

The North Slope has been subject to at least five to seven episodes of glaciation during the past three million years. The glaciers involved were relatively small because of snow starvation due to the complete freezing of the Arctic Ocean. As a consequence, they were more similar to extra-large alpine glaciers than to the massive ice sheets that affected more southerly regions. In no case did the North Slope glaciers produce ice sheets with their own watershed divides, as required for the formation of continental ice sheets. Present-day glaciers occur in the Brooks Range, where they are limited to mountain cirques (Fig. 3.1).

Pliocene and Early Pleistocene Glaciers

Evidence of the earliest glaciation (Late Pliocene, more than 2.6 mya) on the North Slope consists of outwash[1] and erratics.[2] These ancient glaciers, which were as wide as 330 km and extended north to within 30 km of the modern-day arctic coast,[3] deposited the "Kuparuk gravels" and the "Gunsight Mountain erratics." Erratic boulders of Kanayut Conglomerate (Fig. 2.3) as large as 1.5 m in diameter, deposited 100 km or more into the foothills, provide evidence that the interior cirques[4] of the Brooks Range were the sources of these glaciers. After these earliest episodes, the next major glaciations were the Anaktuvuk (Early Pleistocene, less than 2.5 mya) and Sagavanirktok (Middle Pleistocene, 0.8 mya, Fig. 3.2). The Sagavanirktok- and Anaktuvuk-age glaciers extended as far as 70 km north of the Brooks Range.

Late Pleistocene Glaciers

Evidence of major glacial activity on the North Slope during the Late Pleistocene includes moraines[5] (Figs. 3.3, 3.4), kames[6] (Fig. 3.4), kettle lakes[7] (Fig. 3.5), and U-shaped glacier-scoured valleys (Fig. 3.6), in

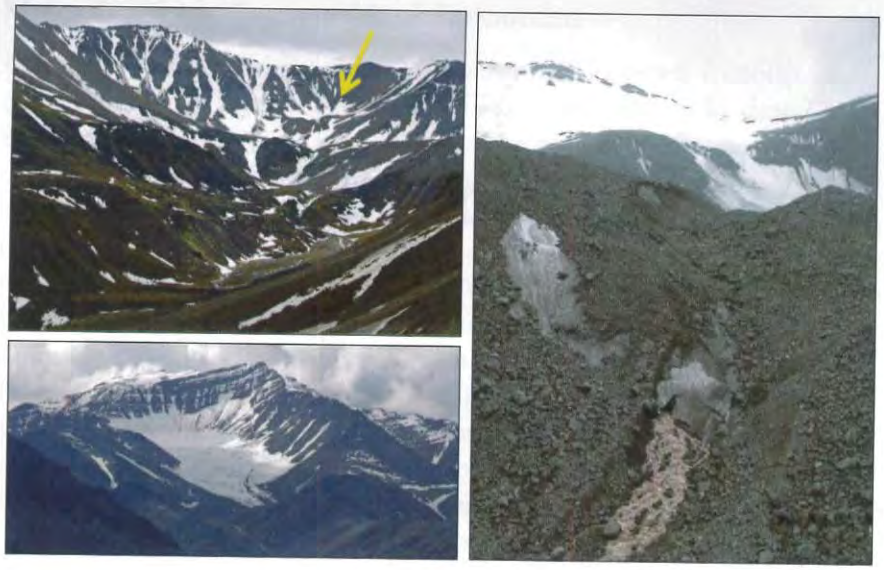


Fig. 3.1. ***Upper left:*** Grizzly Glacier at Atigun Pass. The glacier (arrow) is buried beneath rock debris (ADH). ***Lower left:*** Cirque glacier in Brooks Range. This glacier, informally known as the "Gates glacier," is visible from the Galbraith Lake airport road (MP 275, ADH). ***Right:*** Surface of Gates glacier showing cover of rock debris and a meltwater channel. The reddish color of the meltwater is due to rock flour produced by abrasion between the glacier and underlying bedrock (S.M. Parker).

Fig. 3.2. ***Upper:*** Southwest view from the divide between the Toolik Lake (top center) drainage and the Kuparuk River drainage showing boulder field deposited by the Itkillik I glacier (about 30,000 years ago, ADH). ***Lower:*** View northeast showing deposits from the Sagavanirktok River–age glacier (more than 125,000 years ago, ADH).

Fig. 3.3. ***Upper:*** End moraine of a cirque glacier in the headwaters of the Ribdon River marking the position of the glacier's most recent advance (ADH). The stream has breached the moraine to produce the V-shaped outlet behind the helicopter. ***Lower:*** Itkillik II–age end moraine of a glacier that advanced through Atigun Gorge (about 11,500 years ago, ADH).

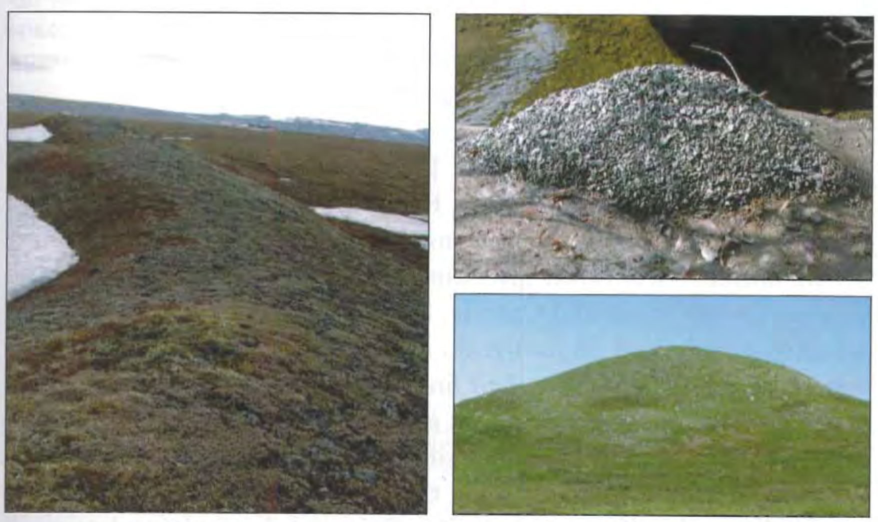


Fig. 3.4. ***Left:*** Itkillik II moraine showing heath vegetation typical of well-drained tundra soils (Toolik Field Station, ADH). ***Upper right:*** Heap of gravel deposited after melting of ice below a talus slope (Atigun Gorge, ADH). This is a small-scale version of the process that forms kames, or hill-size heaps of gravel deposited by a melting glacier. ***Lower right:*** An Itkillik II–age kame near the Toolik Field Station (ADH).

Fig. 3.5. ***Upper left:*** Kettle lake with slumping banks indicating subsidence, possibly due to melting of buried glacier ice (Toolik Field Station, ADH). ***Upper right:*** Galbraith Lake from summit of Guardhouse Rock (ADH). The ancestral Galbraith Lake was formed upstream of an end moraine during the final retreat of the Itkillik II glaciation. At this time the lake extended about 18 km south up the Atigun River valley. The draining of the ancestral lake exposed deep accumulations of fine sediment now evident as the dunes along the Atigun River as it enters Atigun Gorge. ***Lower left:*** Atigun River upstream of the south shore of the ancestral Galbraith Lake (MP 253, ADH) showing coarse sediments. ***Lower right:*** Atigun River channel downstream of the shoreline of ancestral lake (MP 260) showing abrupt change from coarse to fine sediments (ADH).

addition to outwash and erratics. The Late Pleistocene Itkillik glaciation (less than 125,000 years ago, Figs. 3.2, 3.3, 3.4, 3.5) consisted of three major advances and recessions that were synchronous with those of the familiar Wisconsin glaciation that affected much of the Midwestern United States and Canada. Itkillik II refers to the last advance, which began around 20,000 years ago and ended about 11,500 years ago when the glaciers made their final retreat to mountain cirques. The moraines and kames of the Itkillik glaciation are well preserved as permafrost features and can be identified today as sharp moraine crests, a notably irregular rolling terrain, the occurrence of compound kames kettle lakes with collapsing banks (Fig. 3.5), and clearly defined outwash channels. The Itkillik glaciers left a striking geomorphological footprint over much of the North Slope (Figs. 3.2, 3.3, 3.4, 3.5) that is responsible for the abundant heaths and numerous kettle lakes of the foothills.

Fig. 3.6. Northward view of Atigun River from Atigun Pass showing a classic U-shaped, glacier-carved valley (ADH).

4

Permafrost and Patterned Ground

Permafrost

Because of low annual temperatures, the entire North Slope is underlain by permafrost (Fig. 4.1). Permafrost shapes landscapes, creates dams for ponds and lakes, blocks the downward movement of groundwater, and restricts plant roots to the surface of the soil. It is clearly a major factor defining and controlling the arctic environment. Despite its ubiquity, however, permafrost is difficult to directly demonstrate.

The permafrost layer beneath the North Slope is 90–600 m thick. In the foothills at Toolik Lake it is 200 m thick while on the coastal plain at Prudhoe Bay it is 600 m thick. This corresponds to the distribution of the annual mean air temperatures (lowest in the north). Rising air temperatures and increasing snow cover on the North Slope have in historical times warmed the permafrost. However, here the permafrost is in no danger of thawing for some time. For example, temperatures in a series of boreholes along the Dalton Highway have been measured every year for more than 30 years. At a depth of 20 m, the depth at which seasonal cycles in temperature are dampened, temperatures have risen from –25 to –8°C, from –9 to –7°C, and from –5 to –4°C at three locations from the shore of the Beaufort Sea at Prudhoe Bay to the mountains.

One of the major variables affecting the environmental role of permafrost is the amount of ice it contains (Fig. 4.1). Permafrost may or may not contain inclusions of ice. Alaskans pay attention to the presence of ice in permafrost when building a house, a road, or an airport runway, for if large quantities of permafrost ice melts the soil collapses and structures are destroyed. To avoid melting this type of ice, the 1,289 km Trans-Alaska Pipeline is raised above ground for 50 percent of its length due to the presence of permafrost containing dangerous amounts

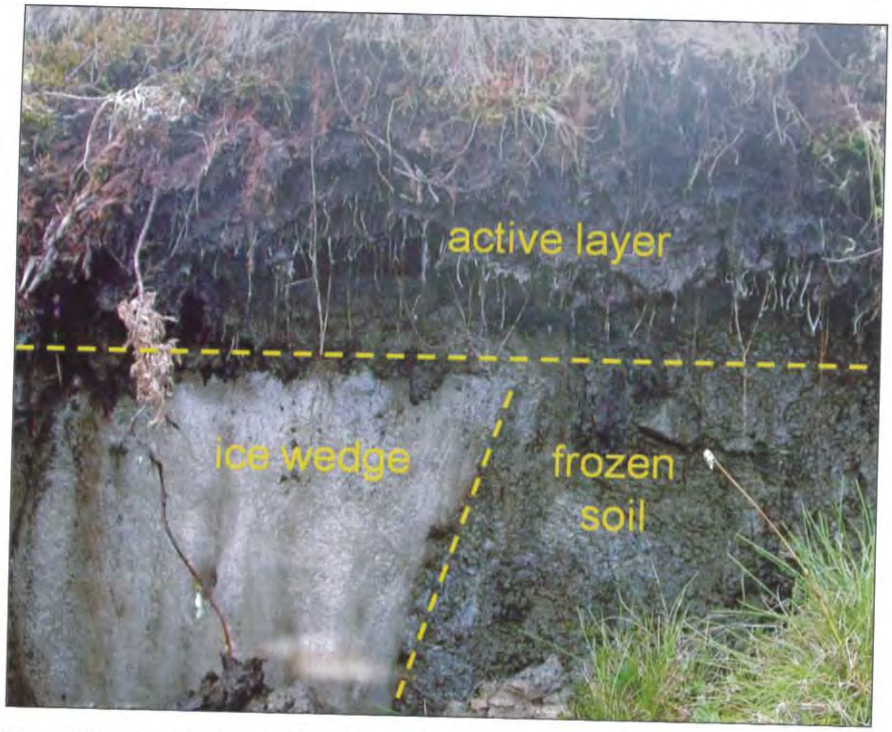


Fig. 4.1. Tundra soil exposed by a thermokarst in a tributary of the Toolik River (ADH). The top ~50 cm of soil is the active layer, or the layer of soil that thaws and freezes each year. Below the active layer is permafrost, or permanently frozen soil, rock, and ice. At left is an ice wedge forming the perimeter of a frost polygon. To the right is frozen soil.

Fig. 4.2. ***Left:*** "Percy pingo" is one of many pingos in the Toolik River pingo field on the coastal plain (Dalton Highway MP 398, JWS). Pingos from this field range from about 1 to 20 m in height. Several can be seen to the west of the Dalton Highway (MPs 375–400). White-fronted geese are in the foreground. ***Right:*** Percy pingo during summer (ADH).

of ice. This aboveground construction was much more expensive than burial of the pipeline.

Pingos are excellent natural examples of the power of ice within the permafrost layer to affect landscapes (Fig. 4.2). These are conical hills up to several hundred feet high caused by the accumulation of a core of ice. One type of pingo (closed system pingo) begins with the draining of a lake that formerly contained unfrozen water all winter (i.e., more than 2 m deep) with unfrozen, water-saturated sediments. The loss of the insulating lake water allows the top of the sediments to freeze and soon a new permafrost layer forms above the old one. The two layers are separated from each other by a thick layer of water-saturated sediments. A freezing front forms beneath the new permafrost layer, drawing water to it. The expansion of water when it freezes increases the volume of water and sediment beneath the new permafrost layer; the layer is pushed up when the volume of the freezing water increases as ice is formed. Little by little the volume of ice increases over the centuries and the pingo pushes upward. Because of the requirement for water-saturated sediments for their formation, almost all pingos are found on drained lake beds or on the floodplains of major rivers. Conical pingo-like mounds in the foothills are likely kames.

Although the permafrost of the North Slope is not yet beginning to thaw, there is good evidence that a warming climate is increasing the occurrence of thermokarsts (Fig. 4.3). Thermokarsts occur when the

Fig. 4.3. A massive thermokarst in a tributary of the Toolik River caused by thawing of ice-rich permafrost (July 2004, ADH).

thawing of permafrost results in the softening and eventual slumping of soils to produce features such as sinkholes and mudslides. Thermokarsts are so named because of their similarity to features of karst topography, a landscape shaped by dissolution of soluble bedrock, usually limestone or dolomite,[1] resulting in abundant sinkholes. In the case of thermokarsts, sinkholes and related features are formed by warming of the permafrost and the thawing of ground ice (thus the *thermo* in thermokarst) rather than the dissolution of bedrock by water. One example of a recent thermokarst is in a small tributary of the upper Toolik River near the Dalton Highway (Fig. 4.3). A stream flowing in the active layer suddenly began to erode the permafrost and melt extensive accumulations of ice. An erosion channel 3 m deep soon formed and large amounts of nutrients and sediment were transported into the Toolik River (Fig. 4.4).

Patterned Ground

Patterned ground refers to the occurrence of repeated, often geometrical patterns on the ground's surface. These are best appreciated when viewed from the air. The processes causing patterned ground are varied but are usually related to cycles of freezing and thawing of the active layer. Perhaps the best-known type of patterned ground is the frost polygon network (Figs. 4.5, 4.6). When saturated tundra soils freeze in autumn, cracks form that result in a pattern of interlocking polygons like those formed as a mud puddle dries. The frost polygons on the Arctic

Fig. 4.4. Sediment entering the Toolik River from a tributary draining an active thermokarst (August 2003, W.B. Bowden). Sediment-laden water is to the left; clear water is to the right.

Coastal Plain, however, may be tens of meters across. During snowmelt the next spring, water fills the cracks. When this water freezes the following autumn, it forms an ice wedge (Fig. 4.1) that pushes the soil up and away to form a low ridge. Over hundreds of years the ice wedge can become several meters deep as it expands. Shallow ponds often form in frost polygons due to the raised ridges forming their perimeters. The eventual development of drainage channels that breach the polygon perimeters, however, causes the ice wedges to melt and the ground above them to settle, changing low-center polygons to high-center polygons. Stream channels in fields of frost polygons often follow a zigzag pattern with deep, circular pools occurring where the ice wedges of intersecting polygons have melted and collapsed. These are called *beaded streams* due to the occurrence of a series of relatively evenly spaced pools (the beads) connected by shallow channels (Fig. 4.7).

Another type of patterned ground formed by the freeze-thaw cycle is the frost boil and the related sorted circle and stone stripe. Frost boils (Fig. 4.8) are patches of bare soil that occur within the otherwise continuous tundra vegetation. Frost boils form when a pocket of wet clay or silt oozes to the surface due to pressure exerted when the surrounding soil freezes and expands during winter—similar to the squeezing of a tube of toothpaste. Before freezing, the oozing clay or silt forms a

Fig. 4.5. View of Galbraith Lake toward the southwest showing numerous low-center frost polygons with uniform rectangular shapes. The Galbraith Lake airstrip is to the upper right.

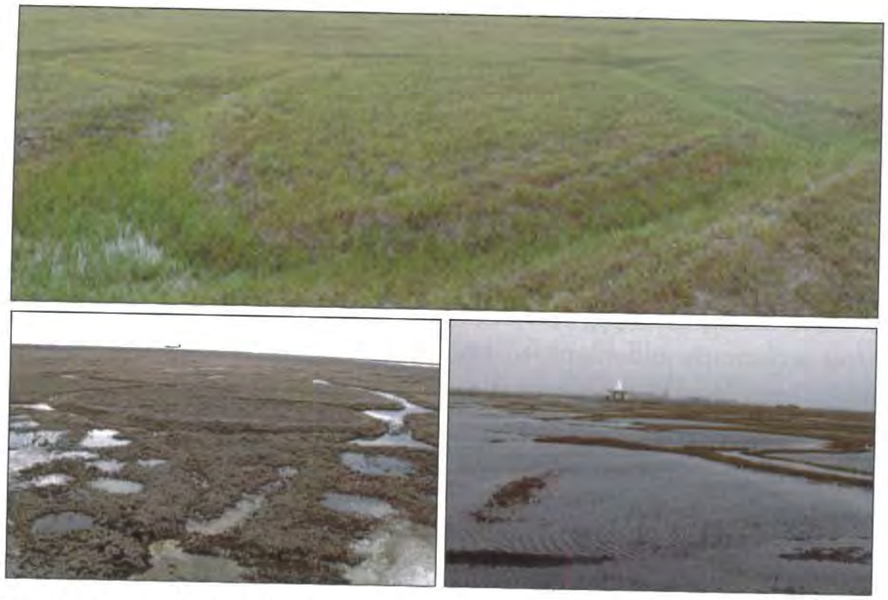


Fig. 4.6. ***Upper:*** High-center frost polygon (Deadhorse, ADH). ***Lower left:*** Low-center frost polygons (Deadhorse, ADH). ***Lower right:*** Flooded low-center frost polygons. Only their elevated rims are exposed (Deadhorse, ADH).

Fig. 4.7. ***Left:*** Horsetail watertracks (Slope Mountain, ADH). ***Right:*** Beaded stream (Toolik Field Station, J.P. Benstead). The channel flows along the perimeters of frost polygons.

mound as much as 10 cm or more above the surrounding tundra. During spring, the mound thaws and collapses to form the bare soil patches that indicate the presence of an active frost boil. Frost boils are important to the dynamics of tundra plant communities because they provide rare patches of bare soil, allowing germination and growth of seedlings unable to compete with mature plants.

Sorted circles (Fig. 4.8) are essentially frost boils that form in ground containing stones of various sizes rather than a dense mat of vegetation. When the center of the frost boil rises during winter, the stones roll to its perimeter to eventually form a circle. If a number of such sorted

Fig. 4.8. ***Upper left:*** Frost boil in ancestral Galbraith Lake sediments (Dalton Highway, MP 266). ***Lower left:*** An array of frost boils (Dalton Highway, MP 266). ***Upper right:*** Sorted circle (Ivishak River, ADH). ***Lower right:*** Stone stripes (Toolik Lake Field Station, ADH).

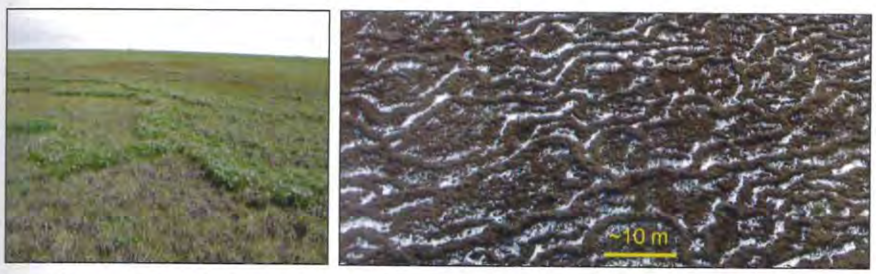


Fig. 4.9. ***Left:*** Detail of string bog showing one plant community on the elevated "strings," which is dominated by *Dryas*, and another in the surrounding sedge and grass wetland. ***Right:*** Aerial view of a string bog (Galbraith Lake, ADH).

circles occur in close proximity, which is often the case, the ground will appear to be covered with interconnected rings of large rocks surrounding discs of gravel and pebbles. Stone stripes (Fig. 4.8) occur when what would otherwise be a sorted circle forms on a slope. Under the influence of gravity, stones are slowly (e.g., a few centimeters a year) displaced downslope, where they eventually produce stripes rather than circles.

String bogs (Fig. 4.9) are relatively flat wetlands with raised string-like features in repeated, roughly parallel lines. The processes producing them are poorly understood. The simplest explanation begins with the accumulation of masses of dead sedge and grass leaves that are rolled into linear "strings" by sheet flow of water during spring thaw. These are formed perpendicular to the direction of flow. Because the strings of string bogs are raised above their immediate surroundings they become colonized by plants requiring microhabitats that are relatively well drained. Once colonized and stabilized by roots, strings become permanent features that support plants such as *Dryas* and dwarf willows in habitats otherwise dominated by sedges. String bogs are common in low-gradient wetlands throughout the North Slope, particularly on the coastal plain.

A final example of patterned ground is the horsetail watertrack (Fig. 4.7). Watertracks are small channels (10 to 20 cm wide) that carry surface water downslope during snowmelt. In summer they are evident as verdant stripes that contrast strongly with the subdued colors of the surrounding tundra. Horsetail watertracks are groups of parallel watertracks that show elegant, synchronized curves, like a horsetail billowing in the wind. These curves are formed by solifluction, which is the downhill slipping of soils as the active layer thaws during summer. This downhill movement is slow (a few centimeters a year), but over long periods of time the once relatively straight watertracks become curved.

5

Habitats and Ecology

Terrestrial Habitats

Reflecting the varied geology, topography, and climate, the Arctic Coastal Plain, the Arctic Foothills, and the Brooks Range contain distinctly different habitats with characteristic plant and animal communities. At the broadest scale, differences among these habitats are driven by water availability and elevation and proximity to the Arctic Ocean, which control patterns of temperature (**Fig. 5.1**). As one moves south from the coastal plain and across the foothills, there is an increase in mean July temperature accompanied by an increase in the height and diversity of plants.[1] For example, only dwarf shrubs (less than 15 cm) are found on the coastal plain adjacent to the arctic coast, where mean July temperatures are only 5–7°C. In the southern foothills, however, woody shrubs as high as 40–200 cm are common. Here the mean July temperatures are 12–13°C. The influence of the Arctic Ocean is also shown by patterns of plant biomass.[2] Although biomass is not particularly high anywhere on the North Slope due to severe nitrogen limitation, plant biomass ranges from 160–370 g/m^2 near the arctic coast to 850–1300 g/m^2 in the foothills, a difference of four to five times. Of course, plant biomass again decreases as one proceeds south into the Brooks Range, where mean temperatures decline approximately 6°C for every 1,000 m rise in elevation (**Fig. 5.1**).

At a finer scale, aspect and relief together determine other characteristics of terrestrial habitats on the North Slope. Aspect affects the amount of available heat, the degree of exposure to prevailing winds, and the duration and amount of snow cover. Relief affects drainage and soil type. Knowing this, some generalizations about habitat distribution can be made depending upon where one happens to be while on the North Slope (**Fig. 5.2**). Windward slopes and hilltops are both exposed to prevailing winds and well drained. Consequently, their dry soils support heath communities, or more simply heaths (**Fig. 5.3**). Under the extreme conditions

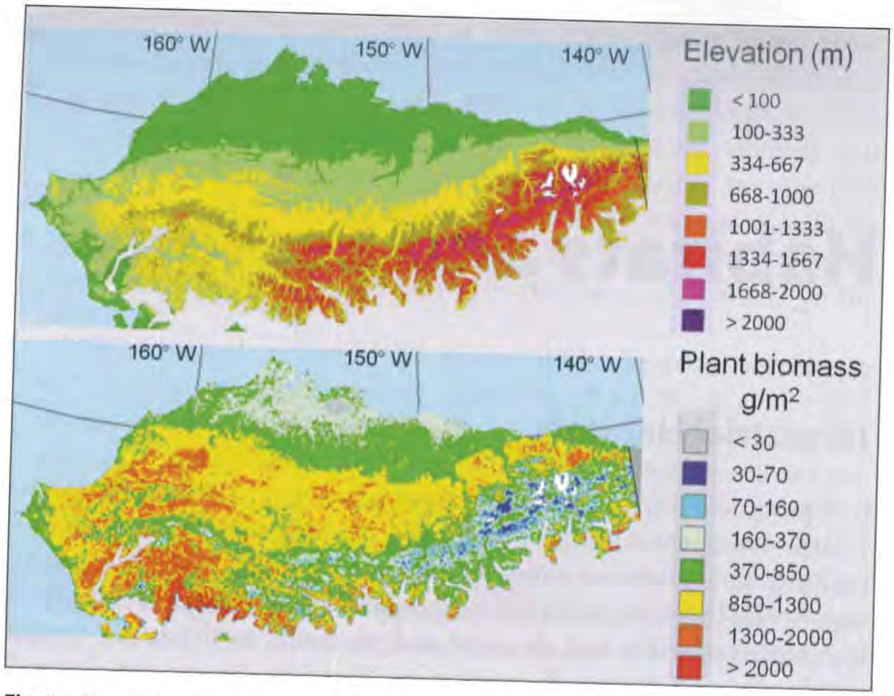


Fig. 5.1. ***Upper:*** Approximate boundaries of the major physiographic provinces of the North Slope as delineated by elevation: Arctic Coastal Plain (less than 100 m above sea level), Arctic Foothills (101–1,000 m above sea level), Brooks Range (1001–2,000 m or more above sea level). ***Lower:*** Pattern of plant biomass (grams of dried aboveground plant material per m^2) of the North Slope. Adapted with permission from Raynolds, Walker & Maier (2006).

of exposed mountain ridges, soils may form only beneath isolated cushion plants such as purple mountain saxifrage (*Saxifraga oppositifolia*). Lee slopes, protected from wind, accumulate large drifts of snow during winter, which may provide a source of water during spring and early summer (**Fig. 5.4**). Here, snowbed habitats support plant communities that depend on the water from melting snow, such as arctic white heather (*Cassiope tetragona*) and mountain avens (*Dryas integrifolia*). Habitats with less extreme relief, such as the gently rolling tussock tundra of the foothills and the flat terrain of the coastal plain, support both relatively well-drained habitats populated by cotton grass (*Eriophorum vaginatum*) and shrubs (*Salix* spp., *Betula* spp.) and poorly drained habitats populated by sedges (*Carex* spp., **Fig. 5.3**). Extensive tracts of brown soils—with high organic matter content and a loamy texture allowing deep oxygen penetration—are found in moist, moderately well-drained tussock tundra while the soils of poorly drained, wet sedge meadows are usually saturated with water (water-logged), resulting in anoxia (absence of oxygen), low rates of decomposition, and the accumulation of peat. Within this

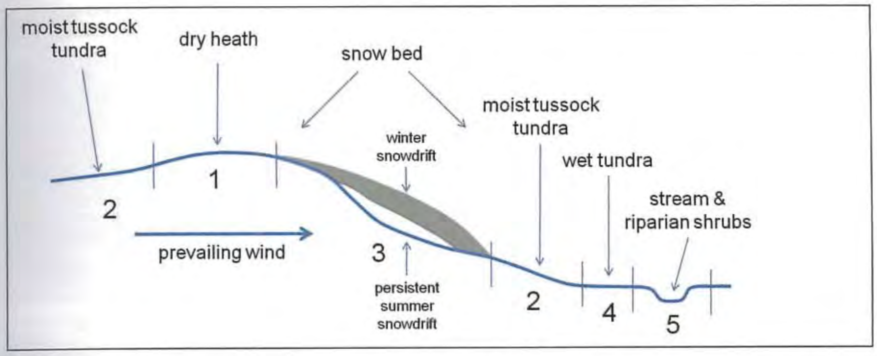


Fig. 5.2. Relationships among major terrestrial habitat types, topography, and direction of prevailing wind. Adapted and redrawn from Raynolds, Walker & Maier (2006).

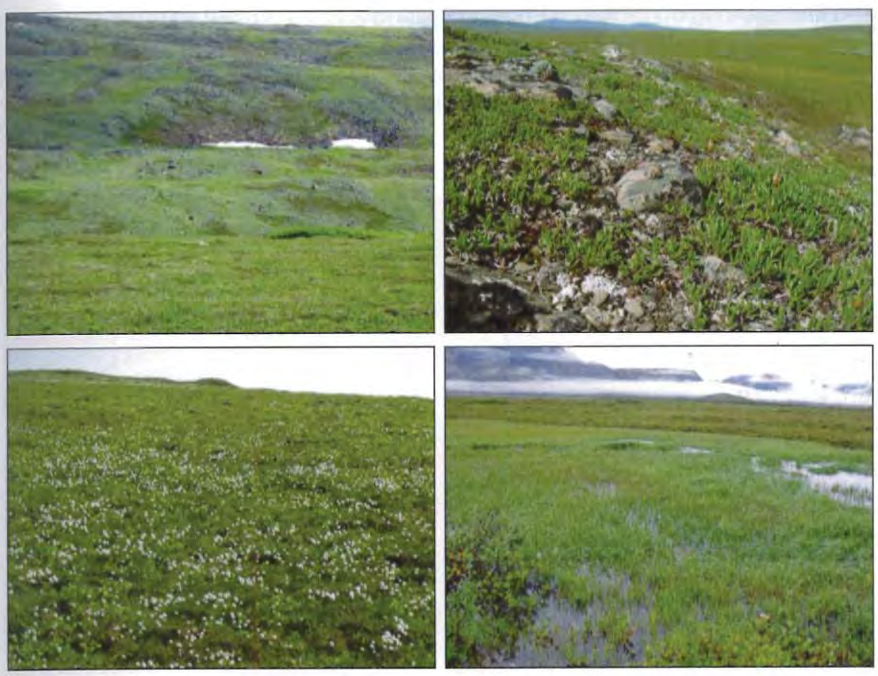


Fig. 5.3. ***Upper left:*** Mosaic of major foothills habitats. The gray and blue-green patches in the top half of the photograph are dry heaths; the darker green-brown areas above the snowbanks are snow beds. Moist tussock and shrub tundra is in the foreground (Galbraith Lake, ADH). ***Upper right:*** Detail of a *Dryas*-dominated heath (Toolik Field Station, ADH). ***Lower left:*** Moist cottongrass tundra dominated by *Eriophorum vaginatum* (Slope Mountain, ADH). ***Lower right:*** Wet sedge and grass tundra (Pump Station 4, ADH).

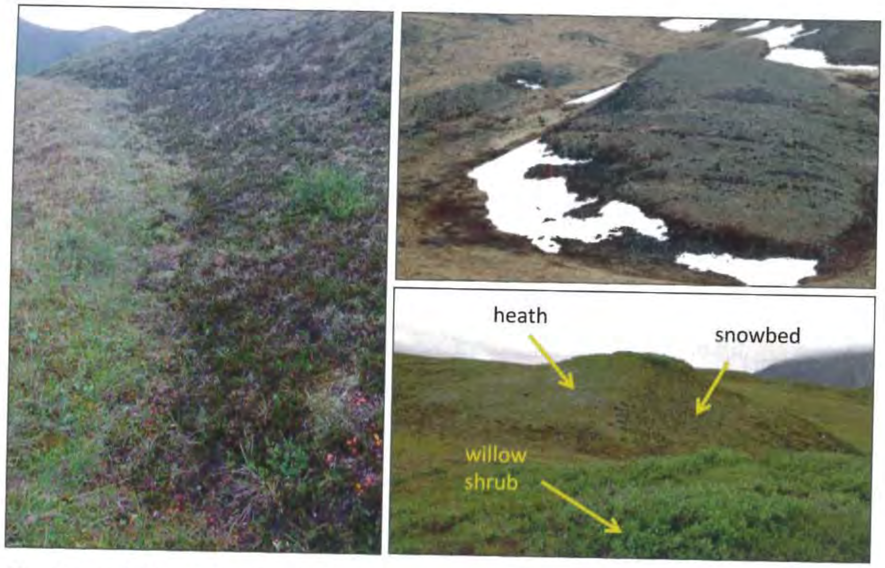


Fig. 5.4. ***Left:*** The left side of this photograph shows moist tussock tundra. The right side (dark green) at the base of the slope is a snowbed community dominated by arctic white heather (*Cassiope tetragona*, Atigun Gorge, ADH). ***Upper right:*** High-relief glacial drift showing dry heath on ridge tops (grayish) and snowbed communities and a persistent snow drift near base (dark brown, Toolik Field Station, ADH). ***Lower right:*** A kame showing a dry heath community on its upwind flank and a snowbed community dominated by *C. tetragona* on its lee flank. Willow shrub habitat is in the foreground.

Fig. 5.5. Shrub habitat with felt-leaf willow (*Salix alaxensis*, tall gray-green riparian shrubs), Richardson's willow (*S. richardsonii*) and arctic dwarf birch (*Betula glandulosa*, short green shrubs), and Siberian alder (*Alnus fruticosa*, dark shrubs at upper right) (Ivishak River, ADH). The inset shows the contrast between Richardson's willows (foreground) and the taller and more sparsely leaved felt-leaf willows along a streambank (background).

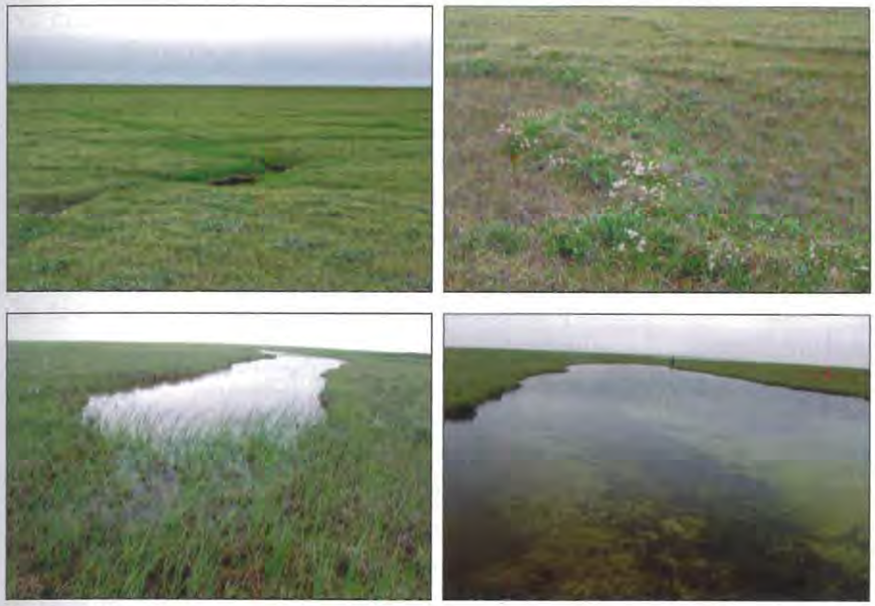


Fig. 5.6. ***Upper left:*** Coastal plain landscape where frost polygons provide the only significant form of topographic relief (Dalton Highway, MP 385, ADH). ***Upper right:*** "Strings" of string bog showing the importance of subtle differences in topographical relief to habitat structure on the coastal plain. The strings support a plant community dominated by *Dryas*. Between strings, the plant community is dominated by wet sedge and grasses (Dalton Highway, MP 395, ADH). ***Lower left:*** Shallow thaw pond showing associated sedge-grass wetland habitat (Dalton Highway, MP 384). ***Lower right:*** Oriented thaw pond with benthic (bottom) habitat visible (Dalton Highway, MP 395, ADH).

general habitat matrix the banks of rivers and streams (riparian habitats) usually support willow thickets dominated by *Salix alaxensis* (**Fig. 5.5**). These generalizations provide a rough guide to the terrestrial habitats of the North Slope. The remainder of this section is devoted to the specifics of its three major ecological regions.

Arctic Coastal Plain

The Arctic Coastal Plain is formed from sediments deposited during the mid- to late Quaternary (less than 1.8 mya). It is roughly defined by the area of the North Slope with elevations of 60 m above sea level or less (**Figs. 5.1, 5.6**). The lowest-lying habitats with water-saturated soils (thaw pond margins, flooded depressions of low-centered polygons) are colonized by wetland plants such as the sedges *Eriophorum angustifolium* and *Carex aquatilis* and grasses such as *Dupontia fisheri* and *Arctophila fulva*. Plant assemblages consisting of dwarf willows (*Salix rotundifolia*), saxifrages (*Saxifraga hieracifolia*, *S. hirculus*), coltsfoot

(*Petasides frigidus*), buttercups (*Ranunculus nivalis*), and mountain avens (*Dryas ajanensis*) occur in drier habitats such as the perimeters of low-centered polygons (**Figs. 4.5, 4.6**). Finally, in exceptionally well-drained habitats, such as former beach ridges (which may occur as far as 50 km south of the present-day coast) and pingos (**Fig. 4.2**), plant assemblages similar to those of foothills habitats are found (cottongrass *Eriophorum vaginatum*, Labrador tea *Rhododendron tomentosum*, lingonberry *Vaccinium vitis-idaea*).

Because the coastal plain is relatively flat, subtle differences in elevation between the raised ridges and the centers of frost polygons are important in determining the distribution and activity of lemmings and birds. The activity of brown lemmings (*Lemmus trimucronatus*) is concentrated in high-centered polygons that provide a mix of relatively dry (centers) and wet (perimeter troughs) habitats (**Figs. 4.5, 4.6**). Both winter and summer activity is greatest in the troughs that provide cover and food in the form of sedges and grasses. Snowy owls (*Bubo scandiacus*), important predators of lemmings, arrive on the coastal plain in May to nest on high-centered polygons kept snow-free by wind.

Arctic Foothills

Repeated glaciations have shaped much of the rolling terrain of the Arctic Foothills (**Figs. 3.2, 3.4, 5.4**). Here important high-relief habitats include kames and kame-like features, moraines, and mounds of drift. Extensive heath (mountain avens *Dryas ajanensis,* crowberry *Empetrum nigrum,* bearberry *Arctous rubra,* alpine azalea *Kalmia procumbens,* Labrador tea *Rhododendron tomentosum,* net-leaf willow *Salix reticulata*) and fellfield[3] communities are found on upwind slopes. Snowbed communities dominated by arctic white heather (*Cassiope tetragona*) develop on lee slopes (**Figs. 5.3, 5.4**). Vast expanses of low-relief, moist tundra are covered by cottongrass (*Eriophorum vaginatum*) and a rich assemblage of shrubs (diamond-leaf willow *Salix pulchra*) and dwarf shrubs (arctic dwarf birch *Betula nana* and *B. glandulosa*, lingonberry *Vaccinium vitis-idaea*, alpine blueberry *V. uliginosum*, cloudberry *Rubus chamaemorus*, Labrador tea *Rhododendron tomentosum decumbens*). Sedge meadows and willow thickets are found in poorly drained lowland habitats and along watertracks (**Fig. 5.3**). Thickets of felt-leaf willow (*Salix alaxensis*) and Richardson's willow (*S. richardsonii*) are prominent along streams (**Fig. 5.5**).

Brooks Range

The Brooks Range contains peaks as high as 2,700 m. Such high elevations at arctic latitudes result in habitats more similar to the polar deserts of the high Arctic than those found elsewhere on the North Slope. These habitats (more than 1,800 m in elevation) are barren with the exception of sparse patches of crust-forming lichens and isolated vascular plants such as purple mountain saxifrage (*Saxifraga oppositifolia*) and alpine draba (*Draba alpina*) in protected locations (**Fig. 5.7**). At lower elevations, dry meadow and heath communities dominated by mountain avens (*Dryas ajanensis*) are common, and meadows in protected valleys may be surprisingly lush with carpets of moss dotted with buttercups (*Ranunculus nivalis*), coltsfoot (*Petasides frigidus*), bear flower (*Boykinia richardsonii*), and a variety of composites. On dry slopes and fellfields a sparse but diverse assemblage of showy flowers may be found, including blackish oxytrope (*Oxytropis nigrescens*), prickly saxifrage (*Saxifraga tricuspidata*), arctic cinquefoil (*Potentilla hyparctica*), arctic forget-me-not (*Eritrichium aretioides*), and Pallas' wallflower (*Erysimum pallasii*). The floodplains of the larger rivers,

Fig. 5.7. ***Upper left:*** Steep, well-drained, boulder-strewn mountain slope (Atigun Pass, ADH). ***Upper right:*** Sheltered, moist mountain meadow (Atigun Pass, ADH). ***Lower left:*** Fellfield with sorted circles (Ivishak River, ADH). ***Lower right:*** Exposed mountain ridgetop (Guardhouse Rock, Atigun Gorge, ADH).

such as the Atigun and Canning Rivers, and their headwater valleys contain sedge meadows and willow thickets (Richardson's willow *Salix richardsonii*, diamond-leaf willow *S. pulchra*, felt-leaf willow *S. alaxensis*), dwarf birches (*Betula nana* and *B. glandulosa*), and soapberry (*Shepherdia canadensis*). One important characteristic of mountain habitats in the eastern Brooks Range is the abundance of limestone that provides the raw material for calcium-rich and relatively alkaline soils that support a characteristic community of vascular plants, lichens, and mosses. This effect can be observed as far north as the arctic coast, where the calcium-rich soils of the floodplains and deltas of the large rivers draining the eastern Brooks Range may support twice the plant species found in habitats containing acidic tundra soils.

Freshwater Habitats

Probably the two most critical factors controlling the structure of freshwater communities on the North Slope are winter freezing and low nutrient availability. With few exceptions (see "Spring Streams and Aufeis" below), freshwater habitats on the North Slope less than about 2 m deep freeze solid during winter. By "freeze solid" we mean that the entire water column freezes from top to bottom, which is significantly different from the covering of surface ice on "frozen" lakes and rivers in north-temperate climates. As a rule of thumb, only North Slope lakes and rivers 3 m or more in depth contain sufficient unfrozen water to support significant populations of fish during winter.[4] Lakes with depths exceeding 3 m are relatively numerous here, but only the larger rivers, such as the lower reaches of the Sagavanirktok and Colville Rivers, are deep enough to contain habitats that remain unfrozen during winter.

Freshwater animal species able to tolerate freezing are few. Consequently, total freezing results in freshwater communities with low diversity compared with those of lower latitudes. In addition to having low diversity, freshwater habitats of the North Slope are also relatively unproductive. This is due to low temperatures and low nutrient availability (particularly phosphorous, which is required in relatively large quantities by all organisms to produce ATP—an energy-carrying molecule—nucleic acids, and cell membranes). In fact, most freshwater habitats of the North Slope are ultra-oligotrophic, which means that nutrient concentrations are exceedingly low. This is attributed to two main factors. First, peaty soils with little mineral content and low soil temperatures result in slow rates of weathering and soil microbial activity, and thus the amounts of phosphorous released from terrestrial to aquatic habitats are very small. Second, abundant iron-rich sediments chemically bind phosphorus, which further reduces its availability to freshwater organisms.

Lakes and Ponds

Lakes and ponds are very abundant on the North Slope (**Fig. 5.8**). Despite the desert-like level of annual precipitation, the permafrost forms a barrier to drainage; small ponds and shallow lakes form anywhere there is slight elevation above a delta or coastal plain. In fact, in some areas of the coastal plain, the entire landscape is made up of lakes and shallow ponds lying on top of old drained lakes (**Fig. 5.8**). Ponds are always small and shallow (less than 0.5 m deep) while lakes are large in area and usually deeper than a meter. While there are a handful of relatively deep lakes (more than 20 m) in the foothills and mountain regions of the North Slope, the vast majority of lakes are less than 3 m deep and found on the coastal plain. The uniformity of the depth of the coastal plain lakes appears to be controlled by the volume of ice in the original frozen sediment. When the frozen ground thaws, the ice melts and the soil consolidates to form a lake basin. A calculation shows that thawing a soil with 10 percent of its volume made up of wedge ice (**Fig. 4.1**) will result in a lake with a depth of 1.4 m.

Because the maximum ice thickness that develops on lakes and ponds during the winter is 1.7–2.0 m, all ponds and shallow lakes less than 1.7–2.0 m freeze solid. Fish are usually, but not always, excluded from the shallow lakes. Ikroavik Lake near Barrow, for example, is 2.5 m deep and contains stickleback fish while another Barrow lake, Imikpuk, is approximately 3 m deep and has no fish. By determining the presence of fish, water-column freezing has dramatic effects on the diversity of communities found in lakes (higher diversity, complex food webs, fish are top predators) or ponds (lower diversity, simple food webs, copepod zooplankton are top predators).

There are literally tens of thousands of thaw ponds on the coastal plain of the North Slope (**Fig. 5.8**). These arise when ice wedges form in permanently frozen soil. Over hundreds and thousands of years the soil above the ice wedges is pushed upward by a few centimeters to create a network of polygons, each 10–40 m across and each containing a pond (**Fig. 4.6**). The pond water and dark sediment capture sunlight and the warm water deepens the pond through thawing of the ice-rich soil. These shallow thaw ponds may completely cover large regions or over time the ice wedges may thaw, the pond boundaries break down, and

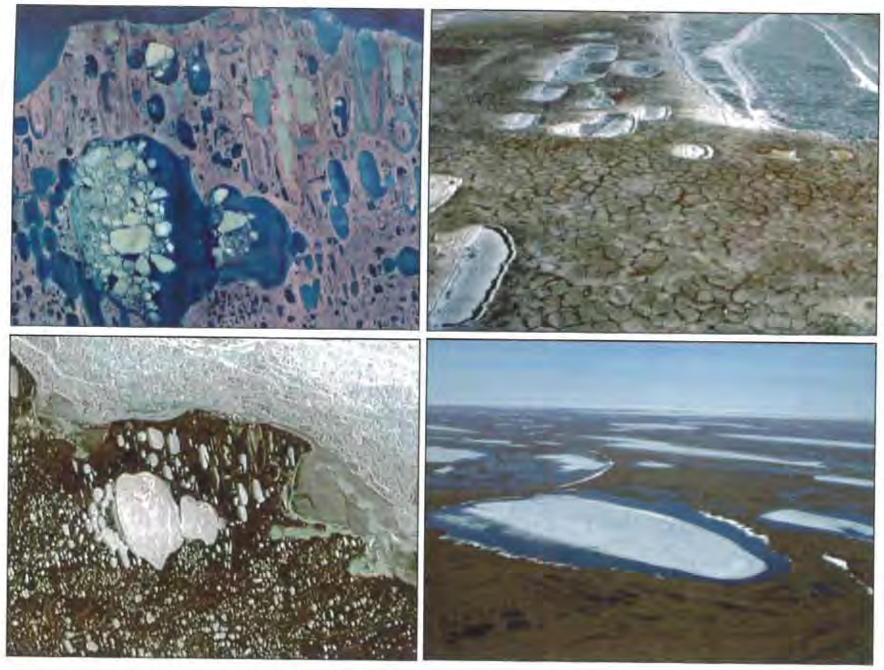


Fig. 5.8. ***Upper left:*** Satellite image of Teshekpuk Lake and abundant oriented thaw ponds during late summer. Thaw ponds in various states of the thaw pond cycle are apparent. Some contain water; others have drained and are dry. Teshekpuk Lake is formed by the coalescence of several adjacent thaw ponds. Note the late-season ice remaining on the surface of Teshekpuk Lake. In some years ice is present year-round (L. Moulton, MJM Research, with permission). ***Lower left:*** Satellite view of the Teshekpuk Lake region during winter. Note that some oriented thaw ponds contain ice while others are dry. The land-fast ice of the Beaufort Sea is visible (L. Moulton, MJM Research, with permission). ***Upper right:*** Frozen oriented thaw ponds and frost polygons (Deadhorse, October 2007, ADH). Note drifts on western shores of ponds indicating direction of prevailing winds from the east-northeast. ***Lower right:*** Oriented thaw ponds and lakes with moats of open water surrounding ice in early summer (National Petroleum Reserve, L. Moulton, MJM Research, with permission).

larger thaw lakes form. Thaw lakes are most numerous on the northern coastal plain where they contribute 15–40 percent of the ground cover. They are usually elliptical or cigar-shaped with long axes oriented about 10–20° west of north (**Fig. 5.8**). This pattern has stimulated much study of how groups of essentially round ponds evolve into groups of elliptical and uniformly oriented lakes. The secret is found in the occurrence of relatively continuous, unidirectional winds caused by the combination of a semipermanent cap of high air pressure over the high Arctic ("polar high") and the rotation of the Earth. As these winds sweep across a lake's surface they drive water toward the downwind shore, which tilts the water's surface (higher on the downwind shore), resulting in opposing circulation cells—one that travels northward and then eastward along the shore and another that travels southward and then eastward. These opposing circulation cells erode and elongate the opposite poles of the basin via both thermal processes that thaw sediments and mechanical processes that sweep them away. Over time these basins become elongated in a north–south direction to eventually produce a cigar-shaped, oriented lake. Thaw ponds undergo long development cycles, from patterned ground to lakes and back again. These cycles have been ongoing since the Pleistocene (an age of 12,600 years was reported for the basin of a thaw lake near Barrow). As a consequence, about 50–75 percent of the Arctic Coastal Plain is covered by thaw lakes and their drained basins in various stages of succession (**Fig. 5.8**).

The thaw ponds near Barrow have been the focus of intense ecological study since the 1970s. These ponds are typically less than 1 m in depth. Consequently they freeze solid and do not contain fish. Their relatively simple food webs are actually composed of two semi-independent webs—a water-column web and a benthic (bottom sediment) web. The water-column web is based upon phytoplankton (suspended algae) grazed by zooplankton (up to eight species of water-column-inhabiting crustaceans such as *Daphnia* and fairy shrimp). These in turn are prey for predatory zooplankton such as *Heterocope*, the top predator in fishless ponds (although shorebirds such as phalarope may also be predators of zooplankton). Because phytoplankton production is low due to phosphorus limitation, however, most animal production occurs via the second web, which is based on decaying sedge and grass leaves (detritus), the food of protozoa and larvae of midges[5] living on the bottom.

Although thaw ponds and shallow thaw lakes are most numerous, significant freshwater habitat is also provided by deeper thaw lakes, such as Teshekpuk Lake. Teshekpuk Lake (70°36.613′N, 153°37.660′W)[6] was formed by the coalescence of several adjacent basins[7] and is the largest lake on the North Slope, the third largest lake in Alaska, and an internationally significant molting habitat for waterfowl (**Fig. 5.8**). Because of its large size and high latitude, it is ice-free for only about six weeks each year and winter ice often persists until August. Consequently, Teshekpuk Lake provides poor summer habitat for fish because its relatively low temperature results in low growth rates. Nevertheless, its relatively great depth (maximum about 10 m) provides critical winter habitat for broad whitefish (*Coregonus nasus*), least cisco (*C. sardinella*), and arctic grayling (*Thymallus arcticus*). Teshekpuk Lake is most usefully considered a "lake system" rather than a lake because the low relief of its drainage (a few meters above sea level) results in a maze of interconnected waterways covering 32,600 km^2 rather than only the 830 km^2 area of its main basins. Although Teshekpuk Lake receives water from numerous inlets, the Miguakiak River is its only outlet. The flow of this river reverses at times due to its low channel gradient.

Lakes and ponds are much less abundant in foothills habitats than on the coastal plain; nevertheless, they provide significant habitat. The kettle lakes[8] of the glaciated regions of the Arctic Foothills can be relatively large and deep and often occur in clusters or lake districts (**Fig. 5.9**). Toolik Lake (68°37.930′N, 149°36.419′W), for example, is a compound kettle lake with a surface area of 1.5 km^2 and a maximum depth of 25 m (**Fig. 5.9**). The food web of Toolik Lake has been studied since the 1970s. As with tundra thaw ponds, two largely separate food webs have been identified: water column and benthic. The energy source for the water-column food web are tiny algae that swim using flagellae (photosynthetic flagellates) and bacteria that feed on dissolved organic molecules that enter the lake with water draining from the tundra. Other nonphotosynthetic flagellates feed on bacteria. The flagellates and bacteria are prey for zooplankton, including seven species of crustaceans and eight species of rotifers.[9] The low productivity of the flagellates and the bacteria, however, supports such a sparse population of zooplankton that they are unable to effectively fuel the growth of fish (lake trout *Salvelinus namaycush*, arctic grayling *Thymallus arcticus*, round whitefish *Prosopium cylindraceum*, slimy sculpin *Cottus cognatus*, burbot *Lota lota*). Consequently, the fish of Toolik Lake feed primarily on prey from the lake's bottom rather than zooplankton. The benthic (bottom) food web is supported by the primary production of diatoms[10] in the sediment and by organic particles that settle from the water column. These provide food for snails, such as the great pond snail (*Lymnaea stagnalis*), and midge larvae (Chironomidae), which are in turn prey for slimy sculpin, grayling, and young lake trout. Adult lake trout, the top predators in many foothills lakes, feed on smaller fish and snails. Compared with the coastal plain and foothills, lakes are sparse in the Brooks Range, although there are several relatively large and deep lakes formed by the damming action of glacial end moraines. Lake Peters (69°19.334′N, 145°02.717′W) and Lake Schrader (69°22.864′N, 144°59.611′W) are examples of glacier-formed mountain lakes exceeding 50 m in depth.

On the coastal plain, water depth is the main factor determining whether fish are present in a lake. In the foothills and Brooks Range, however, additional factors related to topography are important. Here, deep lakes with gentle outlet gradients may contain up to five fish species (lake trout, arctic grayling, round whitefish, slimy sculpin, and burbot). As the gradient of a lake's outlet becomes steeper, however, the pool of potential colonizers becomes smaller due to differences in swimming ability among species. For example, lakes with moderate outlet

Fig. 5.9. ***Upper left:*** Toolik Lake is a large, compound kettle lake in the Arctic Foothills (Toolik Field Station, JWS). ***Upper right:*** A "lake district" with numerous kettle lakes (Itkillik River drainage, J.P. Benstead). ***Lower left:*** Not all lakes formed by glacial processes are kettle lakes. Galbraith Lake has formed upstream of an end moraine that serves as a dam (ADH). ***Lower right:*** A lake district formed by kettle lakes in the Oksrukuyik Creek drainage in early winter (ADH).

gradients may be colonized by grayling and sculpin but not lake trout, and lakes with steep outlet gradients may be colonized only by grayling or may be fishless.

The presence or absence of fish and the type of fish species can have important consequences for aquatic community structure. Lakes with lake trout, for example, tend to have large populations of nonbiting midges because predation by lake trout reduces the abundance of the small fish (e.g., sculpin) that in turn prey on insect larvae. On the other hand, the great pond snail (*Lymnaea stagnalis*), important prey for lake trout, is rare on open sediments in such lakes. Finally, lake trout indirectly affect the abundance of zooplankton by feeding on smaller, zooplankton-feeding fish or by modifying their behavior, causing them to seek refuge in shallow water and thus reducing their effect on zooplankton abundance in the deep-water habitats.

Rivers and Streams

The running-water habitats of the North Slope range from headwater seeps to large rivers such as the Meade, Colville, and Sagavanirktok

(Fig. 5.10). Almost all the major rivers of this region arise on the northern flanks of the Brooks Range and have channels that travel across the foothills and coastal plain before emptying into the sea. Only one major river, the Inaru—which flows only during spring snowmelt—has a drainage that is entirely within the coastal plain.

The Colville River drains an area of 53,000 km^2 and is both the largest and longest river of the North Slope. Its headwaters arise in the Brooks Range and eventually coalesce to form a 600 km long channel that traverses eastward across the foothills before abruptly turning north. As it enters the sea the Colville River divides into as many as 30 distributaries[11] that flow across a 650 km^2 delta (70°25.388′N, 150°36.069′W). Because flow from frozen upstream tributaries effectively ceases during winter, seawater enters the main channel to form a saltwater wedge extending as far as 60 km upstream. Because of its relatively great depth, the lower channel of the Colville River (maximum depth = 12 m) remains unfrozen during winter and provides critical winter habitat for populations of whitefish (*Coregonus*) species.

Fig. 5.10. ***Upper left:*** View of upper Sagavanirktok River toward north. The Dalton Highway is to the left (west). Pump Station 3 is near the horizon (September 2007, S.M. Parker). ***Upper right:*** The Sagavanirktok River in winter (January 2008, ADH). The view is northward with the Dalton Highway and Pump Station 3 to the west. The blue ice is overflow ice, or aufeis. ***Lower left:*** Sagavanirktok River during breakup (May 2010, ADH). ***Lower right:*** Sagavanirktok River shortly after breakup (May 2010).

The major rivers between the Colville and the Mackenzie include (from east to west) the Aichilik, Hulahula, Sadlerochit, Canning, Sagavanirktok (**Fig. 5.10**), and Kuparuk (**Fig. 5.11**). These are relatively small compared with the Colville River, however. The Sagavanirktok River, the largest of these, drains only about 5,700 km^2, or only about 11 percent of the area drained by the Colville River. Also unlike the Colville River, the rivers of the eastern North Slope tend to run directly from the Brooks Range to the Arctic Ocean. Because of their relatively short channels and their locations in a geologically active portion of the eastern Brooks Range, these rivers contain immense amounts of eroding gravels, cobbles, and boulders. The volume of sediment, in fact, is so large that these eastern rivers lack the power to move it effectively and have become braided[12] (**Fig. 5.12**). Braided channels are relatively shallow and provide poor overwintering habitat for fish.

Headwater Streams

Although less conspicuous than large rivers, headwater streams are more numerous. The topographical position of a headwater stream has important effects on the gradient and size of the rocks forming its channel, on the temperature of its water, and on the amount and variability

Fig. 5.11. ***Left:*** View of Kuparuk River west of the Dalton Highway during summer (July 2007, J.P. Benstead). The Kuparuk River, a large tundra stream, has been the focus of research by scientists from the Toolik Field Station since the mid-1970s. ***Upper center:*** Kuparuk River during summer (July 2003, J.P. Benstead). ***Lower center:*** Kuparuk River during winter when the completely frozen river channel is often obscured by drifting snow (March 2007, Dalton Highway, ADH). ***Right:*** View of Kuparuk River west of the Dalton Highway during winter (April 2009, ADH).

of its discharge.[13] Factors such as these control the type of habitat the stream provides. For example, is a stream's channel steep and bouldery or relatively flat and peaty? Is its flow relatively constant or will there be periods of very high and very low flow? Will it freeze solid during winter or flow year-round? All these factors have important effects on the diversity and types of organisms one might find in a given stream.

Four types of headwater streams occur on the North Slope—mountain, glacier, tundra, and spring streams. Mountain and glacier streams are found in the Brooks Range, where their steep channels arise in high-elevation valleys (**Fig. 5.13**). Mountain streams gain their water as run-off from rain and snowmelt. Glacier streams receive most of their water from melting cirque glaciers. The bouldery channels of mountain and glacier streams consist of alternating cascades and pools. The channels of glacier streams differ from mountain streams, however, by having angular stones rather than the rounded stones commonly found in stream channels (**Figs. 3.1, 5.13**). This difference in shape is due to the fact that the stones in the channels of glacier streams are relatively close to their bedrock sources and so have not been significantly tumbled and eroded

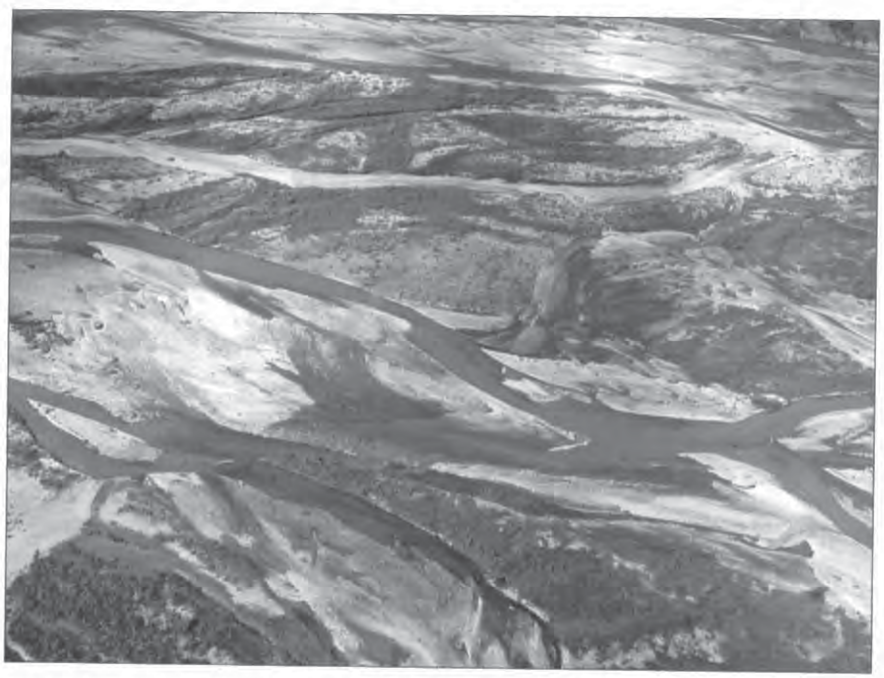


Fig. 5.12. Braided channel of the Ivishak River (ADH). Numerous parallel channels, or distributaries, are visible. Roots of willow thickets provide patches of stability to the otherwise-shifting floodplain gravels. These gravels may contain large volumes of sub-surface flow, which contribute to downstream aufeis formation during winter.

by flowing water. Glacier streams also carry suspended rock flour,[14] causing their water to be tinged with colors ranging from blue-white to rusty red (**Figs. 3.1, 5.13**). The tundra streams of the foothills and coastal plain are the most abundant stream type; they contribute more than 30,000 km or 82 percent of the total stream length on the North Slope (**Figs. 5.11, 5.13**). Like mountain streams, their flow originates primarily as rain and snowmelt. Unlike mountain streams, however, their basins are covered by deep layers of peat underlain by permafrost. Tundra streams often have beaded channels consisting of a series of pools connected by short channels (**Figs. 4.7, 5.13**).

Food webs of mountain, glacier, and tundra streams are well known. Those of glacier streams are the simplest and are based on biofilms[15] on stone surfaces, biofilms that serve as food for midge larvae (Chironomidae). Mountain streams and foothills tundra streams have more complex food webs containing numerous species of midge, black fly, stonefly, mayfly, and caddisfly larvae that feed on biofilms, organic particles suspended in the current, or one another. These insects, in turn, are food for arctic grayling (*Thymallus arcticus*), slimy sculpin (*Cottus cognatus*), and

Fig. 5.13. ***Upper left:*** Mountain stream (Holden Creek, Dalton Highway, MP 267.6, ADH). ***Upper right:*** Frozen channel of Trevor Creek, a mountain stream (Dalton Highway, MP 258.5, October 2007, ADH). ***Lower left:*** Tundra stream (Toolik River, Dalton Highway, MP 292, ADH). ***Lower right:*** Glacier stream draining the informally named "Gates glacier" west of the Dalton Highway, MP 275.

round whitefish (*Prosopium cylindraceum*). Food webs of tundra streams of the Arctic Coastal Plain are less well known. The peat bottoms of their low-gradient channels provide habitat for larvae of midges and caddisflies and amphipods, snails, and fingernail clams (Sphaeriidae), some of which are prey for arctic grayling, least cisco (*Coregonus sardinella*), and nine spine sticklebacks (*Pungitius pungitius*).

Spring Streams and Aufeis

Although spring streams are correctly considered headwater streams, they are different enough to warrant a separate discussion. This is based on the fact that many spring streams of the North Slope flow year-round, whereas most mountain, glacier, and tundra streams freeze solid during winter (**Figs. 5.14, 5.15**). With the exception of a few truly hot springs, such as Red Hill Spring (69°37′35.98″N, 146°01′37.57″W, water temperature = 33°C) and Okpilak Spring (69°19′49.45″N, 144°02′41.06″W, water temperature = 49°C), there are two types of springs on the North Slope: mountain and tundra. Mountain springs have winter temperatures ranging from 4–11°C (remember that air temperatures range below –30°C during winter, which makes water temperatures of 10°C or even 4°C seem pretty warm by comparison). Mountain springs are found where the Lisburne Limestone contacts layers of sandstone along the lower slopes of the northern Brooks Range (**Fig. 5.16**). The source of the water discharged by these springs is poorly understood. Tundra springs have winter water temperatures of 1°C or so and are in the foothills. Tundra springs are fed by water from upstream lakes and the deep gravel beds of large rivers (**Fig. 5.16**). Although many tundra springs flow year-round, some will freeze when their groundwater sources become depleted during winter.

Although spring streams provide a negligible amount of habitat (less than 1 percent of total stream length) on the North Slope, they have important consequences for biodiversity because they provide 100 percent of flowing stream habitat during winter. This is critical for organisms unable to tolerate freezing, such as some stoneflies (e.g., *Isoperla petersoni*), caddisflies (e.g., *Glossosoma nigrior*), and Dolly Varden char (*Salvelinus malma*). The American dipper (*Cinclus mexicanus*) and the northern river otter (*Lontra canadensis*) also require warm springs as overwintering habitat on the North Slope. Unlike most other North Slope streams, the food webs of warm springs include abundant large predacious stoneflies that are important prey for Dolly Varden char (*Salvelinus malma*) and the American dipper (*Cinclus mexicanus*), a semiaquatic songbird.

Fig. 5.14. Kuparuk River during snowmelt (May 2007, J. P. Benstead). During spring, completely frozen river channels gradually thaw from the surface down.

Fig. 5.15. ***Left:*** Main channel of the Atigun River (Atigun River 2 Bridge, ADH) during late winter showing exposed bottom sediments across the entire channel beneath the pipeline crossing. The complete freezing of upstream sources of water results in the cessation of flow during winter. ***Right:*** First flow of water in the main channel of the Atigun River during 2011 (May 15, 2011, Atigun Gorge, ADH). At the time this picture was taken surface water had been flowing on the surface of the river channel for less than 24 hours. The water is flowing over the deep layer of ice remaining frozen to the bottom.

Fig. 5.16. ***Upper left:*** A warm spring near the Ivishak River during July 2009 (ADH). ***Upper right:*** Ivishak Hot Spring during January 2008. When this picture was taken the air temperature was −40°C and the water temperature was about 5°C (ADH). ***Lower left:*** Aerial view of Ivishak River during February 2009 (ADH). Extensive warm spring activity in the eastern tributaries of the Sagavanirktok River (e.g., Echooka River, Ivishak River, Ribdon River) results in long river reaches that remain unfrozen during winter. ***Lower right:*** Tundra-spring tributary of the Kuparuk River (ADH). The aufeis formed by this spring is visible through the willow scrub at the center of the photograph. This spring's source is the deep gravels of the braided Kuparuk River channel upstream.

A good question concerning the fate of the water flowing from springs during winter is "How far from the spring's source will water flow before freezing?" The answer can be found in the form of river icings or aufeis.[16] An aufeis is a mass of ice formed by the successive freezing of overlying sheets of water that flow from a spring (**Fig. 5.17**). In the case of warm springs, the water in a spring stream's channel will flow downstream, cooling along the way until it reaches a point in a river channel that has become frozen solid or is otherwise dammed. Here water must flow over the river ice, where it eventually freezes and forms an aufeis. Since this process continues throughout the winter, the volume of an aufeis can be enormous, ranging up to 5 m or more thick and many square kilometers in area. Depending on its volume and the conditions of a given summer, an aufeis may persist well into August or even later. Aufeis are abundant in the eastern Brooks Range, where they

Fig. 5.17. ***Upper left:*** Melting aufeis formed by a tundra spring west of Galbraith Lake (June 27, 2009, ADH). This aufeis is visible from the Dalton Highway during much of the summer. ***Upper right:*** New overflow ice forming on the surface of the Galbraith Lake aufeis (March 2008, ADH). ***Lower left:*** Aerial view of Ribdon River aufeis (July 2004, S.M. Parker). ***Lower right:*** Calcium carbonate slush ($CaCO_3$) released from aufeis. Freezing results in the precipitation of calcium carbonate that was dissolved by stream water passing over limestone. This white, chalky powder is deposited on the floodplain as the aufeis melts (Galbraith Lake aufeis, June 27, 2009, ADH).

are usually associated with warm springs. The aufeis on the Echooka River (about 21 km^2 in early summer, 69°18.097′N, 147°36.383′W) is the largest and most spectacular and can be seen from outer space (take a moment to search for this aufeis—and others—using Google Earth). In some years residual ice from the Echooka aufeis may persist through the summer to freezeup. The presence of an aufeis on a river can have a large effect on seasonal flow patterns since the water that accumulates as ice during winter is slowly released during summer. Aufeis are often used by caribou seeking relief from mosquitoes.

So, back to the original question—how far from a spring's source will water flow before freezing? The answer depends on the water's temperature and volume. Sadlerochit Spring, a large and relatively warm mountain spring (about 11°C, 69°39.367′N, 144°23.706′W), flows about 5 km before forming a persistent aufeis (i.e., large enough to last through August). Echooka Spring, another large but relatively cool mountain spring (about 4°C, 69°16.069′N, 147°21.341′W), flows for about 6 km before forming an aufeis, as does Ivishak Hot Spring, a small but moderately warm mountain spring (about 7°C, 69°01.852′N, 147°43.003′W).

Although the presence of open streams flowing over relatively long distances during the North Slope winter may be hard to believe, this is indeed the case. One of us (ADH) actually measured the flow dynamics of a small, spring-fed tributary of the Ivishak River during January 2008 (**Fig. 5.16**). On a day when air temperatures hovered around −40°C (about −40°F), the temperature of the stream's water decreased from 5.4°C to 4.5°C over 200 m of channel length. Given this observed decrease in temperature, the water of this shallow stream (average depth = 13 cm, or about 5 inches) would have to travel 800 to 1,200 m before its temperature decreased to 0°C. Since a relatively large amount of heat must be lost before water at 0°C is converted to ice, the water in this stream would travel several more kilometers before actually freezing.[17] One must be mindful that these statistics are based on a very shallow stream; the channels of larger rivers such as the Ivishak, Echooka, and Sadlerochit are much deeper and will thus be less directly affected by air temperature. Given these facts, the occurrence of rivers on the North Slope that flow distances of 5–6 km before freezing during winter loses much of its mystery.

3

Hobbie, J. E. and G. W. Kling (eds). 2014. Alaska's Changing Arctic: Ecological consequences for tundra, streams, and lakes. Oxford University Press, New York, NY, pp. 61-80.

Glacial History and Long-Term Ecology in the Toolik Lake Region

Donald A. Walker
Thomas D. Hamilton
Hilmar A. Maier
Corinne A. Munger
Martha K. Raynolds

Introduction

An understanding of the glacial history of the Toolik Lake region is key to understanding the present-day variety of landforms, soils, and vegetation that occupy the extraordinarily beautiful landscapes that surround the lake. Landscape evolution following deglaciation proceeds much slower in the Arctic than in more temperate regions. Glacial surfaces with ages that span hundreds of thousands of years are present within a few kilometers of the Toolik Field Station. These surfaces have not been altered by agriculture or other anthropogenic influences, so the region is also an excellent laboratory to study the effects of glacial age on arctic ecosystem function. Such studies can help us understand how arctic systems change over long periods of time and provide insights about how they might change in the future. Here we synthesize information from maps and analysis of glacial geology, landforms, and vegetation of the 823-km^2 area of Toolik Lake and the upper Kuparuk River region (Figure 3.1).

Glacial Geology

The complicated topography of the upper Kuparuk River region (Figure 3.1) is the result of deposits laid down by valley glaciers that flowed into the region from the south during three major glacial advances that span the period of time between the middle Pleistocene (610,000–132,000 yr BP) and the late Pleistocene (115,000–10,000 yr BP) (Table 3.1; Figure 3.2). The glacial deposits are assigned to the Sagavanirktok (middle Pleistocene), Itkillik I, and Itkillik II (late Pleistocene)

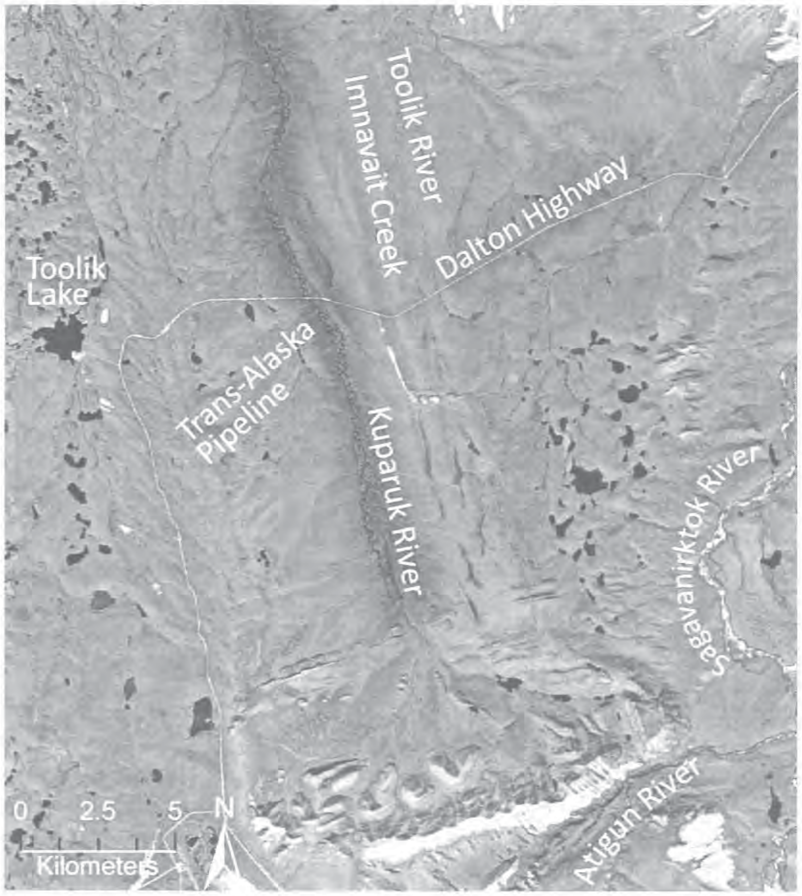


Figure 3.1 False color-infrared SPOT satellite image of the mapped area in the Toolik Lake region. Surfaces glaciated during the Itkillik glaciations have a browner tone and numerous lakes, whereas areas glaciated during the older Sagavanirktok glaciation have a brighter red tone and few lakes (see Figure 3.2 for glacial boundaries). The brightest red areas, e.g., along streams and water tracks, are densely vegetated with shrubs. More intermediate red tones are generally tussock tundra with varying amounts of shrubs. The more brownish tones are less densely vegetated and include dry tundra, nonacidic tundra, and wet tundra. Rocky areas are either light colored, such as the limestone in the lower right corner or greenish gray as in the sandstones and conglomerates along the north side of the Atigun River. Image taken July 28, 1989. [REFER COLOR IMAGE]

glaciations of the central Brooks Range glacial succession (see Hamilton 1986, 2003a, and 2003b for more details on the glacial chronology of the region).

Most of the upper Kuparuk River watershed, including Imnavait Creek, developed on Sagavanirktok-age glacial deposits (light and dark purple, Figure 3.2). The Sagavanirktok River glaciation consisted of several glacial events dating broadly from about 610,000–132,000 yr BP. During the oldest recognized

Table 3.1 Glacial sequence in the central Brooks Range based on Hamilton (1994)

Glacial periods (k yr BP)	Glaciation	Phase
Holocene (12–0)	Neoglaciation	
Late Pleistocene (115–12)	Itkillik II	Latest Itkillik II readvance (12.8–11.4)
		Main Itkillik II advance (25–17)
	Itkillik I	Phase B (>55)
		Phase A
Middle Pleistocene (610–132)	Sagavanirktok	Late Phase
		Main Phase
Early Pleistocene (2,580–610)	Anaktuvuk	
Late Tertiary (>2,580)	Gunsight Mountain	

Timing of ages of glacial periods in thousands of years before present (BP) are based on Richmond and Fullerton (1986) for main chronology and on Gibbard et al. (2009) for the end of the Tertiary. Radiocarbon dates for the Itkillik advances are from Hamilton (2003a).

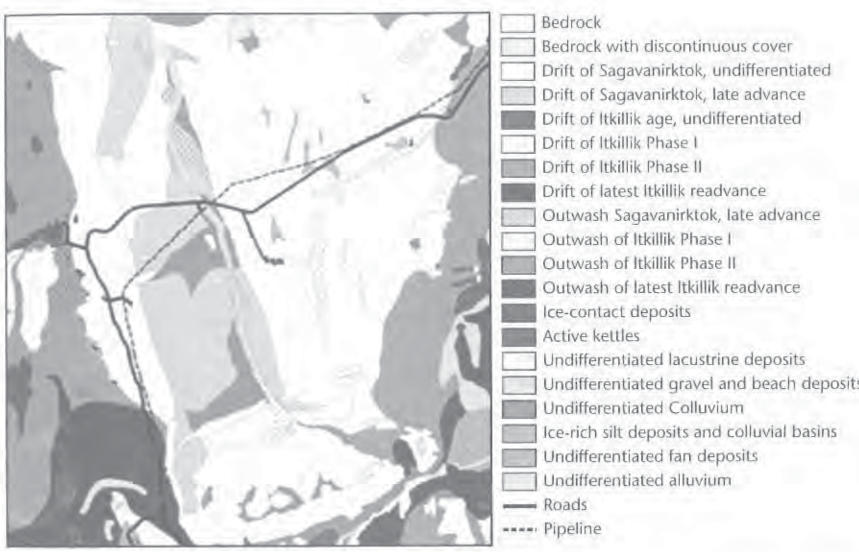


Figure 3.2 Simplified glacial geology map of the upper Kuparuk River region (based on Hamilton 2003a).

(maximum) Sagavanirktok-age advance, large valley glaciers flowed north down the Itkillik, Sagavanirktok, and Kuparuk river drainages and reached their northern limit 50–60 km beyond the northern flank of the Brooks Range. End moraines of the Sagavanirktok River glaciation are about 25 km north of northern edge of the map along the Kuparuk River and about 9 km north along the Toolik River and about 55 km north along the Sagavanirktok River. Several massive lateral moraines of Sagavanirktok age trend SSE to NNW across the central part of the upper Kuparuk River watershed (Figure 3.3, top). These broad, gently sloping moraines are rounded by slope-erosion processes and rise about 100 m from the valley bottoms to their crests (Hamilton 1986, 2003a, 2003b). The glacial till of these moraines is well covered by varying thicknesses of loess and colluvial sediments. Only a few widely dispersed glacial erratics protrude above the tundra surface. The hillcrests and slopes are generally topped by 20–40 cm of peat and tundra vegetation of various types that have been described in studies of a west-facing hill toposequence at Imnavait Creek (Walker and Walker 1996).

A less extensive late Sagvanirktok-age advance (darker purple on Figure 3.2) overflowed the upper Kuparuk drainage from the west and south to form moraines and outwash remnants that are intermediate in appearance between those of the maximum advance and the subsequent Itkillik I moraine succession.

Glaciation of Itkillik I age occurred more than 55,000 calibrated ^{14}C years BP (referred to here as "yr BP"; light green on Figure 3.2) and probably postdates the warmest part of the last interglacial ~120,000 yr BP. These glaciers abutted Sagavanirktok-age moraines on the west side of the upper Kuparuk River drainage, and bedrock ridges on the south side, but overflowed divides on the east and southeast sides of the upper Kuparuk River watershed. Itkillik I moraines are modified by weathering and erosion, but on a much smaller scale than deposits of the Sagavanirktok River glaciations; stony surfaces, small patterned-ground features, and lakes are more common on the Itkillik I surfaces than on the Sagavanirktok surfaces.

The subsequent but much less extensive Itkillik II advances, which date between about 25,000 and 11,400 yr BP (medium green on Figure 3.2 and Figure 3.3, bottom), are contemporaneous with the major late Wisconsin advances of the standard North American glacial succession. Two major advances of Itkillik II age took place between about 25,000 and 17,000 yr BP, forming extensive ice-stagnation features around Toolik Lake. Glacial flow patterns during the Itkillik II advance were generally similar to those of the present-day river drainages. A subsequent Itkillik II re-advance occurred between 12,000 and 11,400 yr BP (dark green on Figure 3.2). Itkillik II age surfaces are generally rocky and gently undulating with small blockfields (areas with >50% cover of moderate to large-sized angular rocks). Crests are slightly flattened, and loess and vegetation cover are locally absent. Kettle lakes with irregular shorelines are common.

Bedrock exposures also occur within the mapped area. These include Cretaceous-age fluvial and shallow marine conglomerates, sandstones, shales, and siltstones (Chandler, Toruk, and Fortress Mountain formations) that outcrop along several WSW–ENE trending ridges (Brosgé et al. 1979). A mountain, composed of Pennsylvanian and Mississippian limestone (Lisburne formation), occurs in the southeast corner of the map (Figure 3.1).

Figure 3.3 Glaciated terrain of Sagavanirktok and Itkillik II age. Top—Sagavanirktok-age moraine, Imnavait Creek vicinity. Note long smooth-flanking slopes and continuous vegetation cover broken only by dispersed large, weathered erratic boulders of resistant lithology. The vegetation is a rather homogeneous cover of tussock tundra. The cottongrass, *Eriophorum vaginatum*, is flowering profusely in this photo. Bottom—Drift and outwash surfaces of Itkillik II age, looking northeast from Jade Mountain, west of Toolik Lake. Note narrow moraine crests, steep-flanking slopes, and discontinuous vegetation cover on the moraine and kame features. Minor as well as major depositional features are well preserved. Photos by D. A. Walker. [REFER COLOR IMAGE]

Surface Geomorphology

The surficial deposits of Pleistocene age within the study area have been modified by a variety of postglacial (Holocene) geomorphologic processes (Figure 3.4). Common periglacial surface geomorphic features within the mapped area include sorted and nonsorted circles (frost boils) and stripes, turf hummocks, gelifluction lobes and terraces, water tracks, ice-wedge polygons, wetland features (strangmoor, aligned hummocks, palsas), and thermokarst features. More detailed descriptions of the various surface geomorphic units are contained in the maps of Walker and Walker (1996) for Imnavait Creek.

The broad hillslope deposits of the Sagavanirktok-age surfaces are more modified by weathering and erosion than are the younger Itkillik-age surfaces. Area analysis of the three dominant glacial surfaces (Figure 3.5) shows that the Sagavanirktok-age surfaces have more well-developed and indistinct hillslope water tracks than the Itkillik I and Itkillik II age surfaces (55%, 13%, and 9%, respectively), fewer lakes (1%, 2%, and 5%), more wetlands (3%, 2%, and 1%), fewer stony surfaces (1%, 8%, and 3%), fewer nonsorted stripes (12%, 13%, and 16%), and fewer areas with greater than 20% cover of nonsorted circles (1%, 1%, and 3%).

Vegetation

The vegetation map of this part of the upper Kuparuk River basin (823 km^2, Figure 3.6) shows 14 map units originally mapped at 1:25,000 scale (Walker and Maier 2008). Within these map units, 33 distinct plant-community types and 19

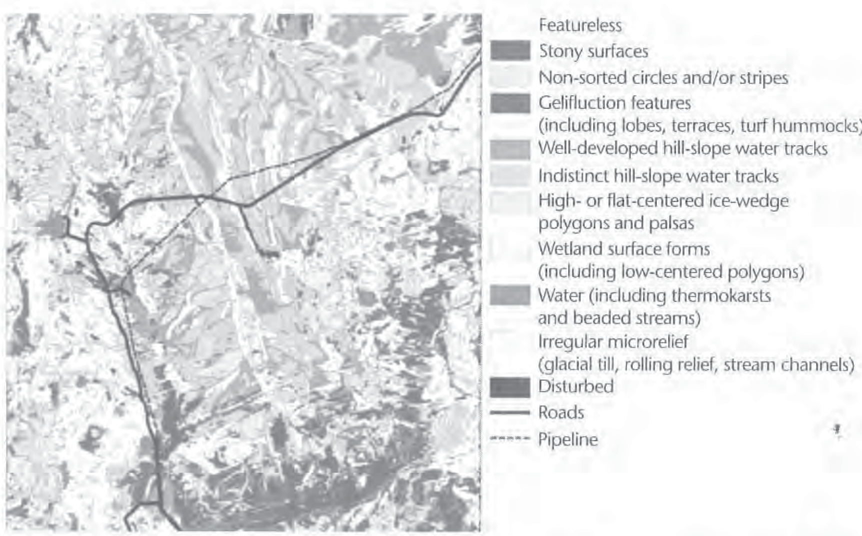
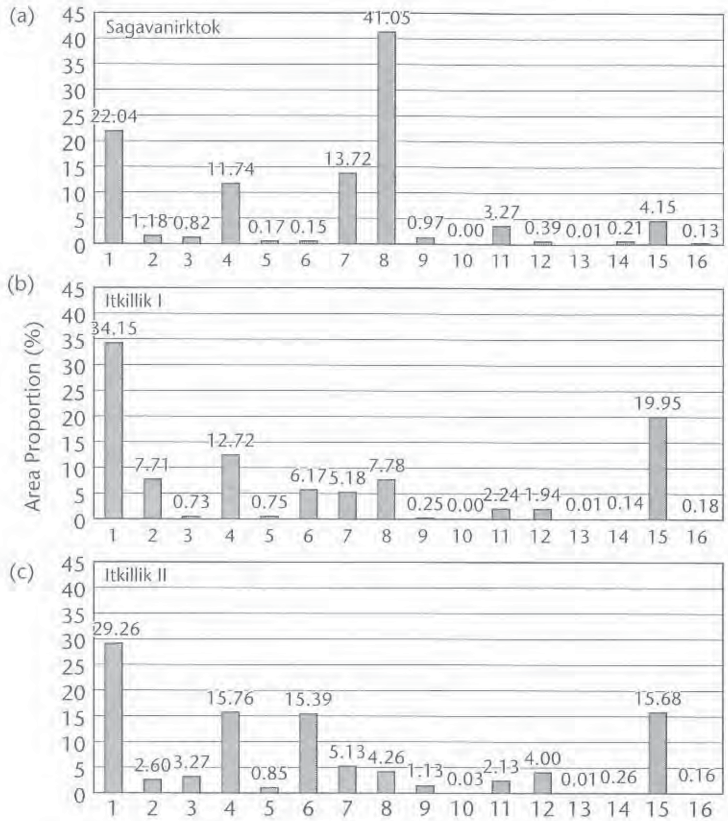


Figure 3.4 Surface geomorphology of the upper Kuparuk River region (based on Walker and Maier 2008). [REFER COLOR MAP]



1. Featureless
2. Stony surfaces
3. Non-sorted circles
4. Sorted and non-sorted stripes
5.Turf hummocks
6. Gelifluction features
7. Well-developed hill-slope water tracks
8. Indistinct hill-slope watertracks
9. High-or flat-centered ice-wedge polygons and palsas
10. Low-centered ice-wedge polygons
11. Wetland surface forms
12. Water
13. Pond Complex
14. Thermokarst and beaded streams
15. Irregular microrelief
16. Disturbed

Figure 3.5 Area analysis of surface geomorphology units (Figure 3.4) on each glacial surface (based on Munger et al. 2008).

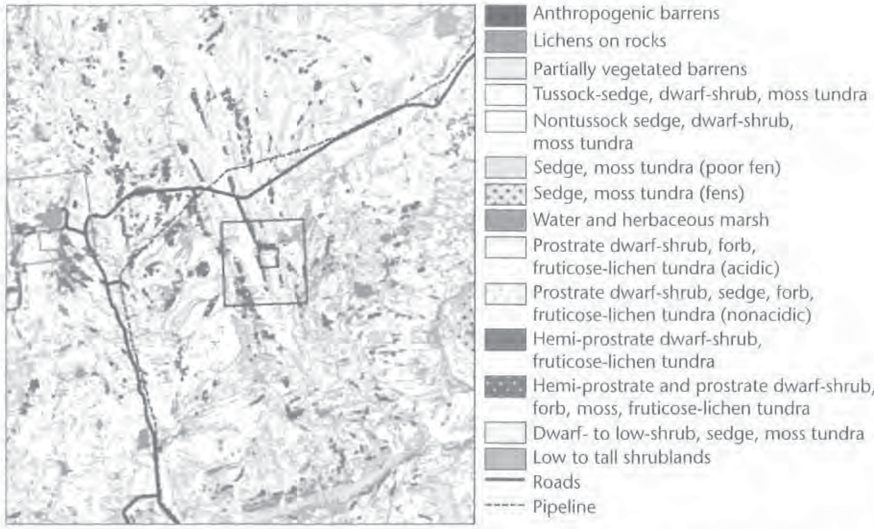


Figure 3.6 Vegetation of the upper Kuparuk River region. Red rectangles are areas of vegetation maps at Toolik Lake LTER research area (Walker and Maier 2008). Black boxes are areas of maps of research areas at Imnavait Creek (described in Walker and Walker 1996). [REFER COLOR MAP]

subtypes are recognized. Seven of the most common plant-community types are shown in Figure 3.3 (top) and Figure 3.7. Most of the published plant-community information in the Toolik Lake region comes from 81 permanent vegetation study plots at Toolik Lake and 73 plots at Imnavait Creek (Walker et al. 1994). The locations of the study plots and details of plant-species cover, soil properties, site factors, and photographs of each study plot are contained in two data reports (Walker et al. 1987; Walker and Barry 1991), which are available at the Arctic Data Coordination Center (ADCC), Boulder, Colorado, at http://adcc.colorado.edu and the Arctic Geobotanical Atlas at http://www.arcticatlas.org/support/.

Description and Environmental Controls

Toolik is within arctic bioclimate subzone E, the southernmost subzone of the circumpolar Arctic (Walker et al. 2005). Mean July temperatures in subzone E are typically 9°C –12°C. The zonal vegetation is dominated by either low-shrub tundra (in areas with warmer soils and thick active layers) or tussock tundra (in areas with ice-rich permafrost, cold wet soils, and shallow active layers).

Jorgenson (1984) first described the contrasting nature of the vegetation on different-age glacial surfaces near Toolik Lake. Several more recent studies have expanded on his observations (e.g., Walker et al. 1995, 1996; Munger et al. 2008).

Sphagno-Eriophoretum vaginati (Walker et al. 1994) (moist acidic tussock tundra; Figure 3.3, top) is the zonal plant association on ice-rich sediments with shallow active layers and low soil pH (3.8–5.5 in the Toolik Lake area); the nomenclature

Figure 3.7 Common plant communities in the Toolik Lake region: Top left—barrens, *Cetraria nigricans-Rhizocarpon geographicum* comm. on block field of Itkillik I–age surface; Top right—Dry acidic tundra, *Selaginello sibericae-Dryadetum octopetalae* on south-facing Itkillik II–age kame west side of Toolik Lake. Middle left—Snowbed, *Carici microchaetae-Cassiopetum tetragonae* on steep north-facing slope of Itkillik II–age till deposit. Middle right—Moist nonacidic tundra, *Dryado integrifoliae-Caricetum bigelowii* on Itkillik II–age hill slope west of Toolik Lake. Lower left—Shrub tundra, *Sphagno-Eriophoretum vaginati betuletosum nanae* in water track margin. Lower right—Wetland, *Carex aquatilis-Carex chordorrhiza* comm. in rich fen complex on Itkillik II surface. Also see Figure 3.3 for photo of tussock tundra (*Sphagno-Eriophoretum vaginati* typicum), the most common and zonal vegetation type in the region. Photos by D. A. Walker. [REFER COLOR IMAGE]

for plant associations follows that of the Braun-Blanquet approach (Westhoff and van der Maarel 1978). This plant association is common throughout much of subzone E of Beringia on older surfaces that have been unglaciated for long periods of time and where ice-rich permafrost and cold wet soils have developed on stable hillslopes. It is found extensively in northern Alaska, northwestern Canada (Lambert

1968), and Chukotka (Alexandrova 1980). Typical taxa in this association include tussock cottongrass (*Eriophorum vaginatum*), a mixture of dwarf shrubs (including *Betula nana, Ledum decumbens, Salix pulchra, Vaccinium* spp.), mosses (including *Sphagnum* spp., *Dicranum* spp., *Aulacomnium* spp., *Polytrichum strictum, Hylocomium splendens*), and lichens (*Peltigera aphthosa, Cladonia* spp., *Dactylina arctica*); the species nomenclature here for the most part follows that used by Walker et al. (1994). Either graminoids or shrubs can be dominant in response to local variations of temperature, moisture, and nutrients. These physiognomic variations within the association can be important from an ecosystem-function perspective. For example, the relative cover and size of deciduous shrubs can strongly affect microclimate, net primary productivity, energy, water and trace-gas fluxes, and animal habitat characteristics (Myers-Smith et al. 2011).

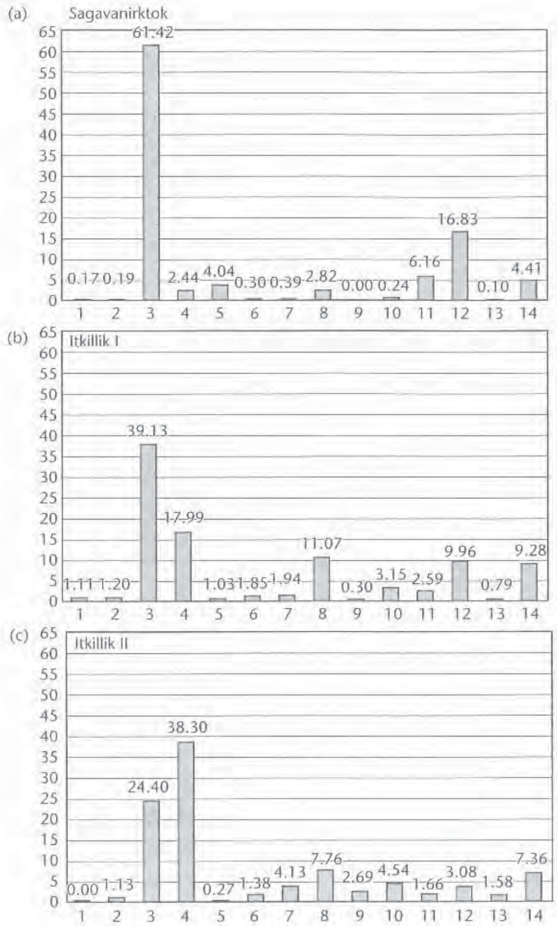
The plant association *Dryado integrifoliae–Caricetum bigelowii* (Walker et al. 1994) (moist nonacidic tundra; Figure 3.7, middle right) occurs on surfaces with somewhat higher soil pH (5.5–7.5) and warmer (in summer) soils, which are typically found on mesic sites of loess deposits, solifluction features, frost-boil complexes, alluvial terraces, and younger glacial surfaces (Walker and Everett 1991; Kade et al. 2005). Characteristic plant species in this association include graminoids (*Carex bigelowii, C. membranacea, C. scirpoidea, Eriophorum triste, Arctagrostis latifolia*), prostrate dwarf shrubs (*Dryas integrifolia, Salix arctica, S. reticulata, Arctous rubra*), forbs (*Bistorta vivipara, Senecio atropurpureus, Chrysanthemum integrifolium, Pedicularis lanata, P. capitata, Tofieldia coccinea, Astragalus* spp., *Oxytropis* spp., *Hedysarum* spp., *Saxifraga oppositifolia*), bryophytes (*Tomentypnum nitens, Hylocomium splendens, Aulacomnium turgidum, Rhytidium rugosum, Ptilidium ciliare*), and lichens (*Thamnolia* spp., *Cetraria* spp., *Peltigera* spp.).

Variations of plant-species composition within moist acidic and moist nonacidic tundra occur along the north-south climate gradient in northern Alaska (Kade et al. 2005) and Russia (Matveyeva 1998). For example, plant communities in bioclimate subzone E and the southern part of subzone D are more species rich and have more erect shrubs than similar communities farther north.

Relationship of Vegetation to Glacial History

Compared to the younger Itkillik-age surfaces, Sagavanirktok-age glacial surfaces have much greater cover of tussock-sedge, dwarf-shrub, and moss tundra (unit 3 in Figure 3.7); much less nontussock-sedge, dwarf-shrub, and moss tundra (unit 4); relatively high cover of erect dwarf-shrub tundra types (sum of units 12, 13, and 14); more poor-fen wetlands (unit 5); fewer rich fens (unit 6); fewer snowbeds (units 10 and 11); and less dry nonacidic tundra (unit 9) (Figure 3.8).

Each glacial surface also has distinctive patterns of biomass (Figure 3.9 and Figure 3.10). The normalized difference vegetation index (NDVI) is a commonly used measure of photosynthetic capacity that is derived from the red and infrared bands of multi-spectral data (Tucker and Sellers 1986). NDVI is strongly correlated with tundra aboveground biomass at the circumpolar scale (Raynolds et al. 2012) and with a variety of tundra biophysical properties including aboveground biomass, leaf area index, and trace-gas flux at the local scale (Shippert et al. 1995;



1. Lichens on rocks
2. Partially vegetated barrens and revegetated disturbed areas.
3. Tussock sedge, dwarf-shrub, moss tundra
4. Non-tussock sedge, dwarf-shrub, moss tundra
5. Sedge, moss tundra (poor fens)
6. Sedge, moss tundra (fens)
7. Water and deep water herbaceous marsh
8. Prostrate dwarf-shrub, forb, fruticose-lichen tundra (acidic)
9. Prostrate dwarf-shrub, sedge, forb, fruticose lichen tundra (nonacidic)
10. Hemi-prostrate dwarf-shrub, fruticose-lichen tundra
11. Hemi-prostrate and prostrate dwarf-shrub, forb, moss fruticose-lichen tundra
12. Dwarf-shrub, sedge, moss tundra
13. Dwarf- to low-shrub moss tundra
14. Low to tall shrublands

Figure 3.8 Area analysis of vegetation units on the major glacial units (from Munger et al. 2008.)

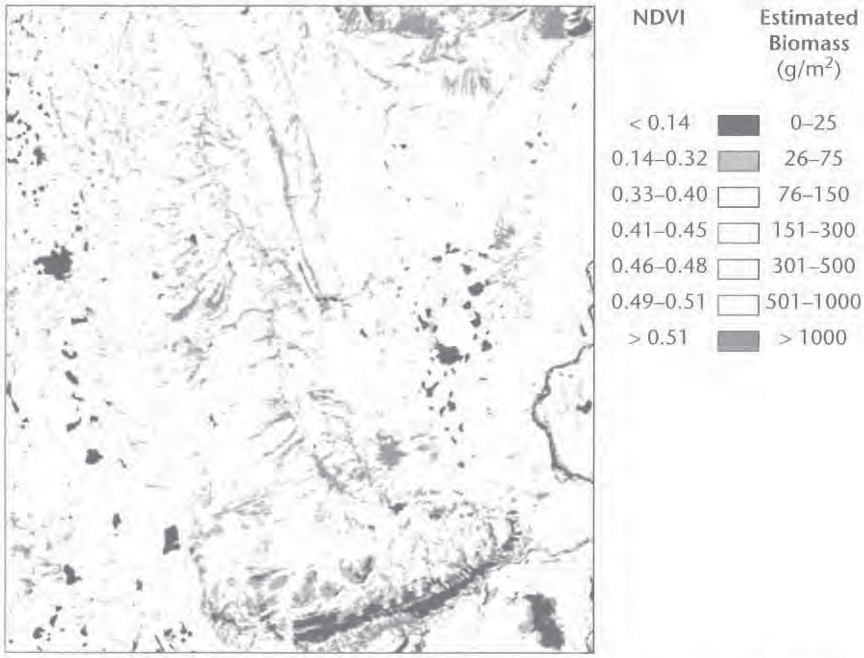


Figure 3.9 NDVI of the upper Kuparuk River region. The image is from a SPOT-1 multispectral image acquired July 29, 1989. NDVI = (NIR – R)/(NIR + R), where R is the reflectance in the red band (630–690 nm) and the NIR is the reflectance in the near infrared band (760–900 nm). Compare the pattern on the NDVI image with the glacial geology map (Figure 3.2) and note greater biomass per unit area on the older Sagavanirktok-age surfaces. Based on Shippert et al. (1995). [REFER COLOR MAP]

Stow et al. 2004). Older landscapes in the Toolik Lake region have higher NDVI (Figure 3.10). The higher NDVI values of the older landscapes are due in part to more-shrubby zonal vegetation. Biomass of the *Sphagno–Eriophoretum vaginati* tussock tundra, which grows on the older acidic surfaces, is about 25% greater than its nonacidic counterpart *Dryado integrifoliae–Caricetum bigelowii*, which is dominant on the younger glacial surfaces (512 g m^{-2} vs. 403 g m^{-2}; see also chapter 5). Older landscapes have more water tracks filled with high-biomass shrub tundra (average of 735 g m^{-2} in the shrubby plots of Walker et al. 1995). The older surfaces also have less area with low NDVI, such as lakes and ponds, nonsorted circles, bare soil, and cobbles.

Several studies have shown broad-scale tundra greening of the Alaskan Arctic in recent decades as documented with AVHRR satellite data at 1- to 8-km pixel resolution (Jia et al. 2003; Verbyla 2008; Bunn et al. 2007; Bhatt et al. 2010). Do the different landscapes also have different rates of vegetation change under the prevailing climate? We might expect change to occur more rapidly on disturbed sites and on younger landscapes with warmer, more nutrient-rich soils. Tundra disturbances—for example, landslides, thermokarst, fire, roadside areas, and off-road

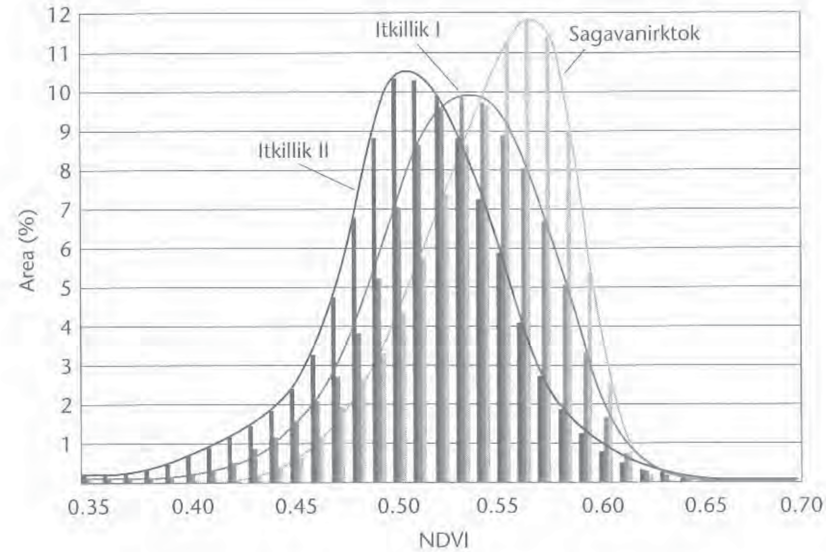


Figure 3.10 Distribution of NDVI on three glacial units in the Toolik Lake region. Curves represent smoothed data for 45 NDVI classes (based on Munger et al. 2008).

vehicle trails—often lead to enhanced plant growth and are likely to be areas of increased productivity and enhanced NDVI (Walker et al. 2009).

An analysis of NDVI changes in the upper Kuparuk region using 30-m resolution Landsat data permitted a closer examination of the parts of the region where significant changes in NDVI have occurred (Raynolds et al. 2010). The most rapid and significant ($p > 0.05$) changes occurred in areas of anthropogenic disturbance such as roadsides and revegetated gravel pads (21% increase). Compared to the Sagavanirktok-age surfaces, areas on the Itkillik surfaces with significant changes in peak-NDVI had larger NDVI increases, and the areas of NDVI increase covered a larger percentage of the vegetated glacial drift and outwash surfaces. The changes on the Itkillik-age surfaces were distributed widely across the surfaces, whereas significant changes on the Sagavanirktok-age surface were concentrated in a few areas, possibly areas of local buried ice or thawing permafrost or other large disturbances. These results suggest that NDVI and biomass changes in the future will also be larger and more widespread on younger surfaces and in areas of disturbance.

Landscape Paludification

Paludification is the long-term accumulation of organic matter (peat) that leads to increased soil moisture and water logging of previously dry landscapes (Gorham et al. 2007). The process has been described extensively in forested landscapes and also has been invoked to describe landscape-evolution in the low Arctic (Walker and Walker 1996). Paleoecological studies from lakes on the Itkillik II and Sagavanirktok-age surfaces near Toolik Lake indicate that during the early part of the Holocene both surfaces likely had tundra with many prostrate dwarf shrub and species indicative of drier, nonpaludified conditions (e.g., *Equisetum, Thalictrum,* Rosaceae, *Encalypta, Selaginella*) (Walker and Walker 1996; Mann et al. 2002; Oswald et al. 2003; see chapter 4). Tundra with plants typically found in acidic tussock tundra (e.g., *Rubus chamaemorus, Sphagnum* spp., Ericales, *Betula nana, Polygonum bistorta*) increased on the older surfaces between 10,000 and 7,500 years ago.

Mosses are critical to this paludification process. The advent of the mosses changes the soil hydrology, soil thermal properties, and soil chemistry, which results in acidic mires in colluvial basins and tussock tundra and extensive water-track development on hillslopes. Observations of the moss carpet at Prudhoe Bay, Sagwon, and Toolik Lake indicate that before *Sphagnum* becomes established, other mosses cause the initial trend toward wetter and more acidic conditions. Early colonizing species include small mosses such as *Encalypta* spp., *Ceratodon purpureus, Distichium* spp., and *Ditrichum flexicaule.* These are followed by larger branching moss species, such as *Tomentypnum nitens, Aulacomnium turgidum,* and *Rhytidium rugosum,* which can develop thick moss mats that insulate the soil, trap moisture, reduce the active-layer thickness, and promote the process of paludification. Once the soils are continuously wet, peat buildup and pH reduction occur, permitting the spread of *Sphagnum* spp., *Dicranum* spp., ericaceous shrubs, and other acidophilus species.

Based on the present-day contrasts between the glacial surfaces, we can deduce that enhanced peat formation and reduced thickness of summer thaw layers on the older surfaces led to restricted drainage, a general acidification of the soils, and the introduction of thick moss carpets to hillslopes (Walker and Walker 1996). The landscapes gradually became less diverse with more tussock tundra, more shrub tundra (primarily in water tracks), fewer lakes, less moist and dry nonacidic tundra, fewer snowbeds, and fewer rich fens.

The patterns of vegetation on present-day landscapes that span multiple glaciations do not, however, represent a continuous successional sequence spanning >100,000 years. Cores taken along foothill toposequences near the Mesa archeological site 150 km west of Toolik Lake show that peat accumulation began near the beginning of the Holocene, slowed during the cooling of the Younger Dryas (12,800–11,500 yr BP), and then resumed by about 8,500 yr BP when the organic surface horizons probably had approached their present wide distribution (Mann et al. 2002; see chapter 4 for the local paleoenvironmental interpretation). A critical factor affecting the different rates of soil and vegetation development on the different-aged glacial surfaces during the Holocene was the accumulation of weathered clays and wind-blown, glacially derived silt (loess) on the older landscapes, which promoted higher soil moisture retention, colder soils, more mosses, thicker organic soil horizons, and greater overall vegetation cover.

Relevance to Arctic Ecology

Much of the LTER terrestrial research at Toolik Lake occurs on the south and east sides of the lake on glacial drift surfaces of Itkillik I age (>55,000 yr BP).

Research occurs on the west and north sides of Toolik Lake on drift and outwash deposits of Itkillik II age (<25,000 yr BP; Figure 3.3, bottom). The 1 × 1 km research grid at Toolik Lake (small red square in Figure 3.6) is mainly on an Itkillik I glacial moraine, but includes areas of Itkillik II outwash on the east side of the map. More detailed maps of the Toolik Lake area are contained in Walker and Maier (2008) and the online Toolik-Arctic Geobotanical Atlas (http://www.arcticatlas.org/; Walker et al. 2009).

LTER research is also conducted on surfaces of Sagavanirktok age (>125,000 yr BP) in the upper Imnavait Creek watershed (Reynolds and Tenhunen 1996) (Figure 3.1; Figure 3.3, top). The 1 × 1 km research grid at Imnavait Creek (small black rectangle in Figure 3.6) spans two broad Sagavanirktok-age lateral moraines with a stream and colluvial basin in between the moraines. The landforms and vegetation at Imnavait Creek were mapped as part of Department of Energy R4D studies, which readers should consult for more detailed vegetation and surficial geomorphology maps of this area (Walker and Walker 1996).

The relationship between soil pH and vegetation patterns has been described in association with other types of long-term disturbance such as wind-blown loess at Prudhoe Bay (Walker and Everett 1991), cryoturbation associated with small-scale patterned-ground features in northern Alaska (Kade et al. 2005), and landslides on the Yamal Peninsula, Russia (Walker et al. 2009). A study of ecosystem properties of acidic and nonacidic tundra was conducted at a major soil pH boundary near Sagwon, Alaska (Walker et al. 1998). Compared to acidic tundra at the same location, nonacidic tundra had less gross photosynthesis, respiration, leaf area index, NDVI, average canopy heights, moss cover, and shrub cover; and much greater evapotranspiration, soil heat flux, active layer depths, and cover of nonsorted circles. More recent studies on different-aged glacial surfaces near Toolik have noted that litter decomposition, soil respiration, dissolved organic C production, and net N mineralization are much greater in acidic tundra than in nonacidic tundra (Hobbie et al. 2002; Hobbie and Gough 2004). Both soil pH and calcium ion concentrations appear to affect microbial respiration and dissolved organic carbon dynamics of tundra and different-aged surfaces, and both need to be considered in models of tundra biogeochemistry (Whittinghill and Hobbie 2011).

Plant diversity is also higher in areas of moist nonacidic tundra (Walker et al. 1998; Gough et al. 2000). Kade et al. (2005) showed the southern variant of nonacidic tundra, *Dryado integrifoliae–Caricetum bigelowii* var. *Lupinus arcticus*, has very high diversity within this plant community (alpha diversity) with an average of 30.8 vascular-plant species per 1-m^2 study plot, compared to 13.8 species per plot in the acidic *Sphagno-Eriophoretum* plant community. A total of 155 species, including vascular plants, lichens, and bryophytes, were recorded in the nonacidic plant association. Numerous basiphilous species in this association have distributions restricted to Beringia or western North American (e.g., *Claytonia bostockii*, *Lagotis glauca*, *Novosieversia glacialis*, *Parrya nudicaulis* ssp. *septentrionalis*, *Potentilla biflora*, and *Saussurea angustifolia*).

Furthermore, plant community diversity (beta diversity) is higher within nonacidic tundra landscapes because of the abundance of nonsorted circles that have a different plant association (*Junco biglumis–Dryadetum integrifoliae*; Kade et al.

2005), which also has high vascular-plant species diversity (20 species per 1-m^2 plot). This greatly increases the amount of species and microhabitat diversity within nonacidic-tundra landscapes (Walker et al. 2011). Exceptional diversity has also been noted in the equivalent vicariate plant association, *Carici arctisibiricae–Hylocomietum alaskani*, on the Taimyr Peninsula in Russia (Matveyeva 1998).

Although the connection between wildlife and nonacidic tundra has apparently not been studied in Alaska, there are numerous characteristics that likely make it important to a wide range of animal species (Walker et al. 2001). The higher floral diversity and greater diversity of microhabitats undoubtedly also affects the local diversity of microbes and invertebrates. Our aerial and ground observations of wildlife during many years of vegetation mapping indicate that many mammals, including ground squirrels, caribou, muskoxen, wolves, and grizzly bears are more commonly found in areas that are rich in nonacidic habitats. The ecosystem properties of moist nonacidic tundra are analogous to those of the hypothesized "mammoth steppe" or steppe tundra of glacial Beringia (Guthrie 1990). Compared to acidic tussock tundra, nonacidic tundra has firm, well-drained, deeply thawed, nutrient-rich soils; high diversity of plant species and habitats; and plants low in secondary protective compounds. Plant-tissue calcium is also much higher in nonacidic tundra (Walker et al. 2001; Hobbie and Gough 2002) and could be a factor affecting wildlife patterns.

Within the southernmost tundras (bioclimate subzone E), nonacidic tundra is not common except on carbonate-rich tills or bedrock, river floodplains, and on late-Pleistocene glacial surfaces. Such areas are likely to be especially important to wildlife. Areas with a juxtaposition of acidic and nonacidic tundra may be particularly valuable to some species, such as caribou, that use different parts of the landscape during their annual migrations as plants change their phenological development.

Conclusion

Most studies that have related vegetation succession to glacial history have taken place on surfaces recently exposed by the retreat of glaciers (e.g., Chapin et al. 1994). The terrain around Toolik Lake offers a unique opportunity to study landscape evolution on glacial surfaces that span much longer periods of time. The analyses at Toolik Lake showed that glacial surfaces all have characteristic vegetation, geomorphology, and patterns of plant production and changes in productivity. Over time, plant succession trends toward peaty, wetter upland surfaces and infilling of lakes in lowland sites. The highest NDVI values occur on those portions of the landscape with abundant shrubs, such as water tracks, on moderate slopes, and on older glacial surfaces. Greening as indicated by NDVI during the period of available satellite imagery occurred heterogeneously across the landscape, with the most rapid change occurring in areas of recent disturbance and on relatively young glacial surfaces. The oldest surfaces show little change except in isolated areas that may be undergoing thawing permafrost. Acidic and nonacidic areas that are characteristic of relatively old versus young landscapes have very different ecosystem properties that are important to aquatic ecosystems (e.g., Kling et al. 1992, 2000) and wildlife, and to their response to climate change.

The surfaces discussed in this chapter span the second (younger) half of the Quaternary. Further north there are glacial deposits from early-Pleistocene and Tertiary times (Table 3.1; Hamilton 1986). Aerial photos and initial ground surveys of these areas indicate that these surfaces have even fewer lakes, more subdued topography, more abundant and larger water tracks, more abundant shrub cover associated with the water tracks, less dry heath vegetation on hillcrests, and even less diverse landscapes than the Sagavanirktok-age surfaces of this study. Thus, each different-aged surface spanning the entire Quaternary period in northern Alaska has distinctive assemblages of periglacial features and vegetation that are legacies of their geomorphic histories. These differences are clear and striking and can be quantified by means of maps derived from aerial photographs and satellite-derived data. Such differences are important with respect to water chemistry of streams and transport of materials into river systems and the general ecology of these regions.

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10

Ecological Consequences of Present and Future Changes in Arctic Alaska

John E. Hobbie
George W. Kling

Introduction

In this chapter we bring together the evidence for changes in the ecology of an arctic Alaska site caused by climate changes, and predict the ecological changes by 2100. Some of these changes are the direct effects of a changing climate on physical, chemical, and biological processes, while others are indirect effects resulting from a number of ecological processes and interactions. One example of an indirect effect is the argument that an increase in plant biomass was likely not the direct effect of warmer soils on plant growth but was the result of a longer period of unfrozen soils on the total microbial activity. The microbial activity converts organically bound soil nutrients to forms usable by plants. Most of the evidence summarized here has been presented in the previous chapters of this book.

The Arctic LTER site at Toolik (the Toolik Field Station of the University of Alaska, Fairbanks) was chosen to represent a well-vegetated, low-arctic landscape including tundra, lakes, and streams. Data have been collected there since 1975 when the road supporting the Trans-Alaska Pipeline was completed. The rivers and lakes near Toolik drain into the Kuparuk River, which flows north to the Arctic Ocean.

In interpreting this long record of research, and in comparing predictions of change in this northern site with other landscapes in different climates, several key factors must be kept in mind: First, the climate at latitude 69° N is cold; summers are short and cool, with continuous daylight for most of the summer season and with continuous darkness in winter. The mean annual air temperature at Toolik is cold enough (–8.5°C) that the entire region is underlain by continuous permafrost. Second, there is a strong north-south climate gradient of ~3.5°C in the mean annual temperature of the Kuparuk River Basin (KRB) from the cool coastal plain bordering the Arctic Ocean to the warmer northern foothills some 170 km to the south where Toolik is located. Third, the site is quite isolated and in many ways pristine, with minimal direct impact of human activities both in the past and at present. Effects of human activities are limited to road dust and recreational and subsistence fishing and hunting. All wastes from the Toolik Field Station itself are trucked to Prudhoe Bay or Fairbanks. Streams and lakes in the area receive no nutrient enrichment, and even atmospheric nitrogen deposition is minimal. The original animal communities of caribou, bear, wolves, foxes, arctic grayling, arctic char, and lake trout are largely intact.

Alterations in Climate Drivers of Change: Effects in Northern Alaska

Air Temperatures

Based on the 22-year record (1989–2010), Toolik annual air temperature averages –8.5°C with a range from –10.6 to –5.8 °C (see Table 2.2). Over this period there is no statistically significant warming trend. One reason for this lack of a trend, despite significant warming at other areas in northern Alaska (Hinzman et al. 2005), is the particular period of record at Toolik. As noted in chapter 2, this lack of a warming trend is largely an artifact of this period of record and is entirely consistent with the lack of a warming trend at Barrow on the coast during exactly the same period. Barrow does, however, show a long-term warming trend over its 110-year record. For the 70 years beginning in 1940, the annual average air temperature at Barrow increased by a total of 2.0°C (Hinzman et al. 2005). At the scale of the entire North Slope, the near-surface air temperatures have warmed more than 3°C over the past 60 years (Shulski and Wendler 2007).

The lack of a significant warming trend in mean air temperatures at Toolik during the past 22 years contrasts with evidence of cumulative environmental changes near Toolik. One change linked directly to warming of the air is an increase in the ground temperatures at a depth of 20 m in a deep permafrost borehole located about 20 km south of Toolik (Romanovsky et al. 2011). At this depth the annual changes are damped out and most of the temperature increases (0.8°C over 20 years) are likely caused by warming air temperatures and perhaps changing snow cover (Figure 2.15). Another indicator of change at the North Slope scale is the continued shrinkage of all the Brooks Range glaciers, including one only 30 km from Toolik (Hinzman et al. 2005). Other cumulative effects, such as an increase in soil weathering linked to a deepening of the active layer and an increase in height and canopy density of vascular plants, are also linked to warming, but indirectly (see details later in this chapter).

Precipitation, Water Balance, and River Discharge

A review of the Arctic as an integrated system (Hinzman et al. in press) reports that pan-Arctic precipitation has increased by about 5% since the 1950s. Despite this increase, which also occurred in northern Alaska, Oechel et al. (2000) found that the summer water balance (precipitation minus potential evapotranspiration) has decreased since 1960 in North Slope villages because of longer and warmer summers. In addition, Muskett and Romanovsky (2011) report that runoff has also decreased overall on the Alaskan arctic coastal plain and foothills. The complete water balance is more complicated, however, and recent analyses indicate that groundwater storage on the coastal plain and foothills, probably in the unfrozen "talik" zones beneath lakes and streams, has increased from 1999 to 2009 (Muskett and Romanovsky 2011). At the same time, there has been a reduction in surface water as thousands of shallow lakes have dried (Smith et al. 2005). These apparently contrasting observations may in fact be consistent, considering that the surface active-layer depth of unfrozen ground has been increasing in many areas of the Arctic (Brown et al. 2000; Liu et al. 2003), which would lead to a redistribution of surface water to groundwater.

The data from Toolik (chapter 2) show no statistically significant trends in either annual amounts of precipitation from 1989 to 2010 or in summer and winter amounts. The range of annual precipitation was 201–462 mm and the average annual precipitation was 312 mm. During this period, 60% of the precipitation fell during summer months (June through August). Over the KRB, mean rainfall ranged from <50 mm yr^{-1} along the arctic coast to >240 mm in the foothills and Brooks Range.

The best record available for the long-term trends in Kuparuk River discharge is from the USGS gauge at the river mouth (chapter 7). The total annual runoff has not changed significantly. However, there has been a shift in the timing of the spring runoff, which prior to 1980 mainly occurred in June (Figure 7.36). After 1990 the runoff in May became significant and now dominates the spring runoff, although the total spring-runoff volume has not changed. Despite the shift in timing of the entire Kuparuk River discharge, there has been no trend in the timing of the snowmelt at Toolik. The fall runoff (September and October) of the entire Kuparuk River has increased in recent years, but it is still a small proportion of the total. The shift in runoff to earlier and later months has lengthened the flow season, and there has been a marked increase in summer high-flow events. This longer flow season with no increase in spring runoff, and potentially higher temperatures, leaves the river susceptible to short-term droughts. These droughts have become more frequent in recent years and have the potential to negatively impact the biota by drying the river channel, reducing available habitat areas, and impeding migration of the grayling population to and from their overwintering sites (details in chapter 7).

Arctic Sea Ice

The September ice extent of the Arctic Ocean has decreased by 11% per decade since 1979 (Serreze et al. 2009). One estimate is that the summer ocean will be ice-free by 2050, although the newest models predict an even more rapid loss of ice (Wang and Overland 2012). This reduced summer ice allows additional solar heat to be absorbed into the top 20 m of the ocean. The resulting heat is slowly released to the atmosphere during the following autumn, which increases atmospheric temperatures. The September-to-November deviation in air temperature in the Toolik area (compared to the temperatures for the same months in the period 1979–2007 is 1-2°C (Serreze et al. 2009). It is not yet determined if the major disturbance of reduction of sea ice in the summer has increased the number of convective storms in the KRB and in the area around Toolik. There are unpublished data showing an increase in the number of lightning strikes in arctic Alaska over the past decade, but a connection with a change in number of storms and an apparent increase in number of wildfires is not yet established.

Future Climates

Based on modeling exercises using mean output from the five general circulation models described in chapter 2, it seems likely that the future climate throughout the KRB will be warmer and wetter, with more snow than presently occurs. By 2099 the annual mean temperature in the KRB, compared with the present, is projected to be ~0.5°C warmer in the south and up to 3°C warmer in the north. For the same period, the precipitation is projected to be ~1.5 times the amount of current precipitation in the southern end of the basin and ~3 times the amount in the northern end (Figures 2.18 and 2.19). This large change in the precipitation of the future would greatly affect both stream flow and soil moisture.

However, it is well known that accurate measurement of precipitation is difficult because of wind and blowing snow, to mention only two problems. It is also well known that the global climate models are not very good at modeling precipitation. Hinzman et al. (in press) also discuss the difficulties of projecting future precipitation. They give one scenario in which permafrost degradation could lead to surface drying, and this would result in less precipitation because ~25% of high-latitude summer rainfall comes from recycled evapotranspiration (according to Serreze and Etringer 2003).

Physical Responses to Change in Northern Alaska

Active-Layer Thickness

As the climate warms, we expect that the thickness of the active layer near Toolik (i.e., the annual depth of thaw of the ground surface) will increase. This is the principal layer where the cycling of C, N, and P controls availability of nutrients and sequestration of C. The expectation of change is based on correlations of the thickness of the active layer and the square root of the degree days of thawing (Shiklomanov et al. 2010), an indicator of the total heating over the summer. For example, in the drained-lake landscape at Barrow, the 1995–2009 data showed a mean of 36 cm of summer thawing (range 14–62 cm). At Toolik, Figure 6.14 shows that for the period 1990–2011 the mean active-layer thickness in the tussock tundra was 40 cm, with a range of 28 to 51 cm. There was no statistically significant change of the active-layer thickness at Toolik over this period. No comparison of degree days of thawing and the thickness of the active layer has been made for Toolik.

In the Toolik region there is, however, strong evidence for an increase in the thickness of the active layer in at least some portion of the catchment, based on changes in the chemistry of Toolik Lake and its inlet streams. The alkalinity of the lake has increased from ~350 to 600 μEq L^{-1} from 1975 to 2007 (Figure 6.14). The most reasonable explanation is that there has been an increase in weathering of previously frozen glacial till during this time. This material is either in a thin layer that has thawed at the bottom of the active layer, or in layers under the water tracks, rivers, and lakes where there is deepening of the thawed layer caused mainly by changes in soil-water temperature or flow. As described in chapter 6, the changing ratio of strontium isotopes in stream water supports this weathering hypothesis to explain the changes over time in the chemistry of Toolik Lake.

Permafrost Warming

The best evidence for warming in the region near Toolik is the continuous warming of the permafrost at a borehole near Galbraith Lake, some 20 km south of Toolik. This is one of five deep boreholes along the Dalton Highway from Prudhoe Bay (West Dock) south to Galbraith Lake at the edge of the Brooks Range (Romanovsky et al. 2011). The temperatures in these boreholes are measured annually at a depth of 20 m, the depth where annual fluctuations are damped out. At the Arctic Ocean at West Dock, the present temperature is –8°C; at Happy Valley, halfway between the coast and Toolik, it is –4.5°C; and at Galbraith Lake, next to the Brooks Range, it is –5°C. The temperature has risen in all five boreholes; at Galbraith, the rise was 0.8°C from 1992 to 2011 (Figure 2.15). Some of this rise was due to an increase in snow on the ground, which insulates the ground from the very cold air of winter (Stieglitz et al. 2003). Some of the rise was due to an increase in the total amount of heat reaching the deep soil (see the preceding discussion of active-layer thickness). At the present rates of warming, there will still be permafrost through 2100. However, because permafrost soils contain tremendous stores of carbon that, when thawed, may be released to the atmosphere as greenhouse gases, there is an increasing debate on whether permafrost warming will feedback to lower latitudes and accelerate global warming (Serreze et al. 2009).

Thermokarst Formation

Collapse of the land surface when permafrost thaws is termed "thermokarst formation." This is a normal landscape process wherever permafrost contains a significant amount of ice (chapter 7). When the ice melts, the ground may collapse, and water moving over the surface may form a gully; or the soil may become saturated with water, and whole hillsides fail or slide downhill. Sometimes a piece of the shore may slump into a lake. On the South Slope of the Brooks Range, where the air temperature is warmer, it is plain that many more fresh thermokarsts have formed than on the North Slope (Figure 7.31). Evidence from the past, a sediment core from Lake NE14 just northwest of Toolik, shows that at least for ~5,500 yr BP there has been a series of thermokarst failures in the basin that left a recognizable mineral signature in the sediments; this indicates that recent thermal erosion of permafrost resulting in thermokarst disturbances has happened in the past in the Toolik area (F. S. Hu, personal communication). However, for the region around Toolik Lake, it is difficult to be completely sure that the rate of thermokarst formation has increased in recent decades. The thermokarst features are difficult to see on satellite pictures, and aerial photos from the past are rare in this region. Evidence for a recent increase in thermokarst formation comes from Gooseff et al. (2009), who used remotely sensed imagery on a helicopter transect to conclude that there were a number of new thermokarst features near Toolik.

Thermokarst failures can have impacts on terrestrial and aquatic systems. For example, over a period of two to three years, a single thermokarst gully carried 18 times more sediment than would normally be delivered by the Kuparuk River basin south of the road crossing near Toolik (~143 km^2) over the same period. Ammonium and total P in the disturbed streams are sometimes ten times the normal stream concentrations (Figure 7.35). Downstream rivers and lakes will be affected. Given the warming predicted by the end of this century, permafrost degradation and thermokarst formation will undoubtedly increase in frequency and become a major disturbance of the tundra landscape.

Wildfire

Fire has not been a normal part of disturbances on the North Slope. Until recently the assumption was that the tundra was too moist to sustain fire. But in 2007 a lightning storm sparked an intense fire that burned for two months and covered an area of 1039 km^2 northwest of Toolik during a period of low rainfall (chapter 6; Jones et al. 2009). This was the largest wildfire known to have occurred on the North Slope of Alaska. At the fire site, the albedo was reduced and the net radiation increased. As a result, in 2008–2010 the soils at the fire site were warmer and the depth of thaw greater than in unburned areas (Rocha et al. 2011). The change in surface energy balance contributed to the formation of new thermokarst features in the burned area. The recovering vegetation is mostly the same as before the fire, except for the absence of mosses and lichens, and except in those areas where the fire burned down to mineral soil. Here there are many seedlings of *Eriophorum* and forbs, indicating a remarkable ability of the tundra to rebound from this novel and severe disturbance.

Even though the area affected by the fire was only a small part of the entire North Slope, the carbon balance of the entire region was affected when this fire released >2 Tg of carbon to the atmosphere, mostly due to combustion of the upper few centimeters of soil organic matter. The amount of C released was equal to about half the annual net C sequestration of the entire North Slope region (chapter 5). Future increases in wildfire frequency, severity, and area burned have the potential to dominate the slower, climate-driven rates of regional changes. Disturbance due to wildfire will likely become another major disturbance of the tundra landscape in the same way as wildfire dominates large-scale disturbance in the boreal forest.

Microbial Response to Warming in Northern Alaska

The makeup of the soil and water microbial communities of the Toolik region is strongly shaped by the vegetation (Wallenstein et al. 2007) in the soils and the source of the organic carbon of the flowing waters (chapter 6). Within one landscape, such as MAT, there are even different microbial communities associated with tussocks and with shrubs (Judd et al. 2006). Most of the microbial data is about bacteria, but measurements of the carbon stocks reveal that the biomass of fungi is 50–500 times greater than that of bacteria in the tussock tundra (Vignette 5.2).

Response of Microbial Processes to Warming

Microbes in the laboratory and in nature respond to warming by increases in metabolic activity, including respiration. In surface waters, the maximum activity was found at both 12°C and 20°C (Adams et al. 2010). Surprisingly, the maximum activity for bulk soils is at 20–30°C in the Arctic and sub-Arctic (Nadelhoffer et al. 1991; Rinnan et al. 2011). Almost all soil measurements are of the total microbial respiration, but Pietikäinen et al. (2005) were able to separately measure growth of bacteria and growth of fungi in temperate soils incubated at a wide range of temperatures. Optimum growth occurred at 25–30°C, but below ~10°C the fungal activity was much higher than bacterial activity. At Toolik, the fungal biomass in the soil is many-fold greater than the bacterial biomass, but it is well known that microbial biomass does not necessarily equate with microbial activity. Certainly more research is needed to answer this interesting question of whether fungal or bacterial activity dominates in the Toolik region.

Beneath the snow cover, bacteria in soils continue to respire even when temperatures reach –5°C or even –10°C, because a thin layer of unfrozen water still covers soil-particle surfaces (chapter 5; Mikan et al. 2002). This under-snow respiration accounts for ~20% of annual respiration at Toolik (Schimel et al. 2006).

At the LTER site and at other sites in northern Alaska where nitrogen strongly limits plant growth, there is a continuing slow increase in plant growth that indicates the microbial mineralization of organic nitrogen, the major pool in soil, to N forms available to plants. One type of evidence comes from satellite views (Jia et al. 2002; Verbyla 2008) that show an increase over decades in the annual maximum NDVI (normalized difference vegetation index), an indication of leaf biomass. Another type of evidence of increase comes from a detailed analysis of plant growth in a large number of plots sampled four times over 20 years at Toolik and Imnavait Creek (Vignette 5.5), and from warming and fertilization experiments discussed in chapter 5. The plot samples and the satellite views agree with the slow changes in plant growth in the warming experiments (Figure 5.6) and with the results of fertilization experiments; all indicate increases in plant growth that are likely to be primarily related to the increase in the availability of N to plants. One of the first publications to point this out was Chapin et al. (1995), reporting Toolik data.

The observations discussed above, the change in NDVI recorded by satellite and the change in plant growth at Toolik over time, indicate that plants are responding to an increase in microbial activity caused by warming. This increase is not reflected in the measures of the mean annual temperature (Figure 2.4) at Toolik. A better way is to examine the total amount of heating during a summer; for example, Shiklomanov et al. (2010) found a correlation between the thickness of the active layer at Barrow each summer and the square root of the degree days of thawing. Applying the same approach, Jia et al. (2003) found that on the North Slope the NDVI-based greenness index of maximum plant biomass derived from satellite images correlated well with a summer warming index. This index was the sum of monthly mean air temperatures greater than 0°C. This summer warming index of total heating has not been calculated for Toolik, but it may well be a better way than the average annual temperature to describe possible warming effects on microbial process in the soil.

Effect of Major Disturbances on Microbes: Warming, Thermokarst, Active-Layer Deepening, and Fire

There is little documentation of effects of major arctic disturbances on microbes. Certainly the direct effect of warming on increased microbial activity, already discussed, is by far the major effect. Another effect is the potential for high concentrations of dissolved organic matter (DOM) to appear in a stream or a lake downstream from a thermokarst event. This DOM is undoubtedly decomposed by stream and lake microbes, but there is little information on rates (see chapter 6). Further, Walter et al. (2006) described the movement of particulate carbon into Siberian lakes when lake edges collapsed; concentrations were so high that the lakes became anaerobic and methane was released. Schuur et al. (2008) point out that the organic material released from thawing permafrost and processed *in situ* in soils is only slowly decomposed because most of the organic matter is of low quality; it has already been partially decomposed before it was incorporated into permafrost. However, if the previously frozen soil C is brought to the surface during a thermokarst failure, the exposure to ultraviolet (UV) light can increase the lability to bacteria (see Vignette 6.1). Wildfire leads to warmer, drier soils and increased thermokarst disturbance, all of which could lead to higher rates of microbial activity.

Ecosystem Responses Acting Through Microbes

When microbes make more nutrients available as a result of warming, the effects may extend beyond increased plant growth. As explained in chapter 5, long-term fertilizer and warming experiments consistently lead to higher terrestrial biomass, primary production, and changes in species composition. In MAT, the most common type of tundra in northern Alaska, the experiments indicate that deciduous shrubs should become more dominant in a warmer climate with higher nutrient availability. Experiments in other plant communities (e.g., moist nonacidic tundra, heath, and wet sedge) do not show this proliferation of shrubs (Hobbie et al. 2005). Yet, pairs of photographs taken 30–40 years apart do support the prediction of an increase in shrubbiness, although the careful monitoring at Toolik Lake and Imnavait Creek (Vignette 5.5) indicate a general increase in plant height and canopy density with little change in relative abundance of shrubs. Modeling calculations indicate that a regional shift to shrubs could have a greater effect on regional climate than would global warming. The whole system response would be a positive feedback (Figure 5.15) as shrub canopies reduce the springtime albedo of the snow cover, absorb solar radiation, and warm the air significantly. An increase in shrubs would also trap more blowing snow in the winter and in this way insulate the soil from very cold winter temperatures more effectively. Microbial activity would increase as warmer soils in the wintertime would allow nutrient mineralization to proceed through much or all of the year. However, a negative feedback also occurs in the summer when the denser canopies intercept more incoming solar radiation, thereby reducing the amount of solar radiation that reaches the soil surface. In this way the shrubs cause cooler soil temperatures during the summer. It is not yet known whether the positive or the negative feedback from increased arctic shrubs will be most important for microbes.

The LTER warming experiments produced an unexpected ecosystem effect on the ectomycorrhizal fungal community structure and function (Vignette 5.3; Deslippe et al. 2011). After 18 years of warming, the vegetation had shifted to the shrub *Betula nana*, and the shift was accompanied by changes in the symbiotic fungal species. These newly dominant species were able to mine the nitrogen in resistant organic matter, and their hyphae were able to explore far out from the roots. Evidently the nitrogen in labile organic matter of the soil lying close to the roots in the control plots had been depleted by microbial activity during the warming experiment. In the same warming experiments, Deslippe et al. (2012) found that the bacteria *Actinobacter* increased as did the fungi *Russula* ssp., *Cortinarius* ssp., and Helotiales.

The data on aquatic bacterial response to warmer temperatures also point to higher activity with warmer temperatures, but there are interesting nonlinearities in the response. There appear to be two major groups of aquatic bacteria in the Toolik area, adapted to two different temperature optima, one at ~12 °C and another at ~20 °C (Vignette 6.2). The result is that depending on the initial, average temperature of the system, warming may either increase or decrease microbial community activity. The concentrations and "quality" of DOC modify this response to temperature, and as described in chapter 6 the amounts and timing of terrestrial carbon inputs to surface waters exert a strong control on microbial activity and community structure. Although there are no arctic experiments on effects of warming on concentrations or quality of dissolved carbon, the results from Adams et al. (2010) indicate that both carbon limitation and temperature limitation of microbes occurs, apparently at different times of the summer season. Thus the response of aquatic microbes to future climate changes will depend on the matrix of interactions between landscape water balance and its effects on hydrological export of terrestrial DOC and also on inherent physiological responses of microbes to temperature.

Will the mineralization of organic N and P compounds in the soil result in increases in concentration of nutrients in soil water and eventually in streams and lakes? The experimental treatments of fertilization in streams and lakes have proven that both N and P nutrients limit primary production, but there is no evidence that this is already happening in undisturbed waters. For example, the concentration of water column chlorophyll in Toolik and other lakes, an indicator of algal growth, has not changed appreciably since 1985 (Figure 8.18).

Changes in Microbial Communities and Processes by 2100

In predicting changes by 2100, we begin with several assumptions: (1) that the types and rates of changes that have already occurred will continue; (2) that the predictions of changes in annual mean temperature from models for Toolik are correct; and (3) that the drastic changes of fire and thermokarst will increase in frequency but will not dominate the landscape. The slow increase in N mineralization, likely caused by an increase in the total degree days above 0°C, will continue. The predicted increase in the mean annual air temperatures will also add to the mineralization. However, as discussed in chapter 2, at Toolik the increase in mean annual temperature by 2100 is predicted to be less than 1°C. The net result will likely be a small but continual increase in the abundance of shrubs and their symbiotic mycorrhizal fungi. There are no long-term experiments that allow more than these general predictions. Changes in microbial communities will be linked to changes in vegetation communities through mutual feedbacks and interactions of both communities with climate. Presumably the microbial communities would become more similar to those existing in the boreal forest to the south of the Brooks Range. Chapter 6 suggests that terrestrial environments serve as reservoirs of microbial diversity for surface waters and that patterns of aquatic microbial diversity are structured by initial inoculation from upslope habitats. In this case, future environmental changes on land will likely impact the diversity and function of aquatic bacteria. Overall, there will be only slight changes in the landscape by 2100. Permafrost will still exist and the vegetation will be essentially the same as in 2012 unless, of course, the frequency of these now-rare events increases much more than expected.

Vegetation Responses to Change in Northern Alaska

Direct and Indirect Response of Vegetation to Change

All the evidence from the LTER research presented in chapter 5 confirms the conclusion by F. S. Chapin (1983), p. 47 that "temperature limits the rate at which resources become available but that temperature is not a strong direct limitation to plant growth in the Arctic." Thus, when plant photosynthesis is measured over the short term at different temperatures, there is a positive temperature effect. When plant growth was measured in experimental warming plots at Toolik, there was a positive response the first year, but it soon became evident that the plant growth had become limited by nutrients. Chapter 5 states that the limited responses to greenhouse warming that were observed in tussock and wet sedge tundras (Figure 5.6) are interpreted as resulting from a relatively slow increase in soil nutrient availability in response to modest temperature increases. A similar result was obtained from warming experiments at 20 arctic sites in open-topped chambers as a part of the International Tundra Experiment; initial increases in individual plant growth in the first two years of treatment were frequently not sustained over longer periods (chapter 5).

In the warming and fertilizer experiments in MAT at Toolik (chapter 5), the response of *Betula nana* is striking as this species comes to dominate in every plot. Graminoid plants also respond, but *Betula* and other shrubs have the ability to branch rapidly and grow taller as biomass accumulates. In this community, *Betula* soon shades out other plants except for a few shade-tolerant species like *Rubus chamaemorus*. In the warming experiments in the nonacidic tundra, where *Betula* and other taller shrubs are rare or absent, there is a general increase in abundance of all plant functional types, resulting in an overall increase in biomass and productivity similar in magnitude to that of MAT (Hobbie et al. 2005). In other types of tundra (such as dry heaths), wet sedge, grasses, or sedges are favored by fertilization and warming over at least the first 10–15 years of the experiments (e.g., Gough et al. 2012). But even in these experimental sites where fertilizer treatments have been maintained for over 20 years, the relative abundance of *Betula* appears to be increasing (G. Shaver, personal observation).

Effect of Major Disturbances: Wildfire and Thermokarst Formation

The response of tundra vegetation to wildfire depends strongly on the severity of the burn. The types of response have been well described by Racine et al. (1987) for the North Slope plants after a major fire on the Seward Peninsula (northwestern Alaska) and another fire on the Noatak River on the southern side of the Brooks Range. Lichens and mosses are intolerant and will not survive even the lightest of fires, which typically also consume most or all of the aboveground leaves and shoots. Even after severe burning, which may consume ten or more centimeters of surface organic matter, *Eriophorum* will resprout as the plant rhizomes are protected by the dense, wet tussock. Also, sedges like *Carex* spp. and shrubs such as *Betula*, *Ledum*, and *Vaccinium* will resprout from belowground plant parts if charring is not too deep. A few species, in particular some of the grasses, may survive as seeds. At the other extreme, fires that consume the tussocks and most of the soil organic matter will completely destroy the vegetation communities. However, recovery is rapid: within five to six years after a 1977 fire, the total vascular plant cover reached 50% to 100% of an unburned control. The graminoids often flourish after a fire that removes competing shrubs.

Thermokarst formation will certainly be a major future disturbance of tundra; the effect on streams and lakes of more sediments and nutrients has already been mentioned. A major effect on terrestrial ecosystems will be through rapid colonization of disturbed surfaces by shrubs. An increase in shrubs, discussed in detail in chapter 5, will have impacts on such things as the amount of trapped snow and thus on soil temperatures, on the regional climate, and on animal food and shelter (see the preceding discussion).

Unknown Aspects

As explained in chapter 5, plot-scale and regional-scale models of ecosystem function exist for the Arctic; however, the intermediate scale, which is the scale of hillslopes and small catchments including spatial interactions among neighboring patches of different ecosystem types, is lacking (Rastetter et al. 2004). Temporal modeling is needed to predict ecosystem changes in the 20- to 100-year future. Spatial and temporal scale models are also needed for scaling plot-scale experiments to regions and the pan-Arctic. These longer-term and coarser-scale models will also have to include additional processes that have been relatively little studied at Toolik, such as seed dispersal, germination, and seedling establishment, which are currently rare and slow in undisturbed tundra. Disturbances such as fire and thermokarst may create opportunities for following the time course of disturbance-recovery processes that are poorly understood in the Arctic.

The changes have many implications for the ecological system. For example, in the two types of tussock tundras that have been followed in detail (Vignette 5.5) the increased vascular plant growth reduced the moss and lichen cover in the moister vegetation at Toolik but had no effect on the bryophytes in a mixture of moist and dry tundra at Imnavait Creek. The eventual loss of lichen cover would eliminate an important winter food for caribou in the Toolik region. Studies from other arctic tundra sites suggest that small herbivores are also likely to be impacted by increasing woody shrub dominance as voles, lemmings (Oksanen et al. 2008), ground squirrels (Karels et al. 2000), and marmots all depend very heavily on sedges, grasses, and forbs—plants that are shaded out by increasing dominance of taller woody shrubs. In addition, migratory songbirds that breed on the tundra during the summer months have varied nesting habitat requirements, with some species nesting only in low-stature, graminoid-dominated tussock tundra (i.e., Lapland longspur) and others nesting only at the base of or within tall *Betula nana* or *Salix* spp. shrubs (i.e., Gambel's white-crowned sparrow, redpolls, and American robins) (Boelman, Gough, and Wingfield, unpublished data). Tundra canopies dominated by woody shrubs harbor significantly more arthropod biomass than graminoid dominated communities (Boelman, Gough, and Wingfield, unpublished data), suggesting that increased shrub cover could impact tundra trophic dynamics by altering the base of the food web.

Lakes too would be affected by a change in vegetation to shrubs and trees, which leads to a change in the chemistry of the particulate and dissolved organic carbon entering lakes from the surrounding soils. This is more than speculation because it is known that lakes south of the tree line in Canada contain high concentrations of CDOM (chromophoric dissolved organic matter), which on the one hand absorb light and thus limit the amount of photosynthesis in lakes, but on the other hand protect algae and bacteria from UV radiation (Vincent and Hobbie 2000).

Changes by Year 2100

The slow and subtle changes in the vegetation now in place near Toolik Lake certainly indicate a trend toward a future increase in shrubs. These changes are still small and took decades to develop (Figure 5.V5). The timetable for the change is unknown; therefore, all that can be said right now is that by 2100 the tundra will look a lot more like land on the South Slope of the Brooks Range than at the present time. There will be more and taller shrubs. Trees, most likely the white spruce now found just south of Atigun Pass, will move north at some point. By 2100 the productivity of the tundra will be higher, and the amount of organic matter accumulation in plants and soils will continually increase. Cycling of the nutrients nitrogen and phosphorus will be more rapid.

Stream Ecology Response to Changes in Northern Alaska

Changes in the River Environment

The environmental factors controlling the ecology of the Kuparuk River have remained unchanged over the study period (the mid-1970s to present; see chapter 7). It is an oligotrophic stream with low amounts of algal growth and low amounts of input of organic matter from its watershed. The stream temperature in summer ranges from 7°C to 13°C. Inorganic nutrients are either close to the limits of detectability (soluble reactive phosphate and ammonium are <0.05 μmol L^{-1}) or very low (nitrate plus nitrite 0.1 to 10 μmol L^{-1}). Oxygen is always close to its saturation amount. The one exception to the above lack of change is the McClelland et al. (2007) suggestion that warmer and wetter summers have increased the concentrations and export of inorganic nitrogen in the form of NO_3 from the upper Kuparuk catchment, especially in the last 15 years (Figure 7.37). The reason for this change is unknown, but a good possibility is that there is more microbial nitrification taking place.

The total flow of the Kuparuk River and other streams (chapter 7) has not appreciably changed over the study period (data for the whole Kuparuk River spans 1971 to present). There has been a shift to earlier flow in the spring; most of the spring runoff now occurs in May instead of June, and some flow continues late in the fall. Despite the shift in timing, which spreads the flow over a longer period, the total flow is unchanged. In addition the summer precipitation appears to be more variable in recent years than it was earlier, which increases the probability of drought and of very low stream flows. In times of low flow, the water temperatures are warmer than in times of high flow (Figure 7.7).

The conclusion of chapter 7 is that over the study period, climate or other environmental changes do not appear to have contributed significantly to changes in the stream biota. But there is an effect of temperature on the growth and survival of adult grayling and young-of-the-year (YOY) fish (e.g., Figure 7.24). Each summer differed in the average stream discharge and in the temperature of the water. It was found that adults grew more in cool summers with high discharge, while YOY grew best in warm summers with low flow (Deegan et al. 1999). This makes ecological sense as well because in warm summers the metabolism of adults is raised, and they need more food. The need for more food raises an important question—if stream fish need more food in the future when water temperatures are higher, will insect productivity increase enough to meet this need?

Effects of Major Disturbance

Twice during the 37 years of observations, in 1999 and 2002, the rivers near Toolik flooded during the summer months. In 2002, for example, an August snowfall was followed by rain. The combined precipitation and snow melt caused a tremendous flood in the Kuparuk estimated at >100 $m^3 s^{-1}$, while normal summer flow is 1–5 $m^3 s^{-1}$. The usual flooding at the time of the spring runoff does not do much damage to the streambed because much of the river still contains ice and the streambed is frozen. This summer flood, however, was unique in its intensity. The beds of many streams were completely transformed and eroded, with tremendous effects on the stream insect biota, the basis of the food web for the arctic grayling. The water in Toolik Lake became turbid because of the particulate matter that entered the lake.

A summer drought is another major disruption with important biotic consequences. There have been major drought events in 2005, 2007, 2009, and 2011. During droughts the flow in the Kuparuk River may stop or disappear beneath the stream bed in certain stretches of the river, and grayling are restricted to pools. During the extreme drought in the summer of 2011, the Kuparuk River near Toolik stopped flowing completely from mid-August until mid-September. The ecological consequence was that the arctic grayling could not make their usual end-of-summer migration to the lake in the headwaters of the Kuparuk River where they spend the winter. Luckily, a mid-September rain allowed several thousand grayling to reach the lake.

Response of the Whole Stream System

As noted above, environmental changes occurring in the stream, such as warming, discharge, or chemistry have not changed the stream biota. An increase in the rate of thermokarst formation would add nutrients and sediments to streams; nutrients would increase primary and secondary productivity, but sediments could shade the epilithon and reduce algal productivity. The result of these contrasting effects on the growth of grayling is uncertain. Thermokarst formation would also harm grayling survival if added sediments changed the circulation of oxygen-rich waters in the streambeds, reducing the oxygen necessary for survival of grayling eggs.

As discussed in the section on response of microbial processes to warming, the resultant increase in the availability of nutrients to terrestrial plants appears to be already happening. Eventually some of the increased amounts of nutrients cycling in the soil will move into streams, but there is no clear indication yet of changes in nutrient concentrations in streams (but cf. Figure 7.37). Another exception is described by Hobbie et al. (1999), where a small stream flowing near one of the seven gravel mines in the Toolik Lake catchment supplies 5% of the water entering Toolik Lake but 35% of the phosphate. The underlying permafrost has thawed exposing several meters of previously frozen glacial till to weathering that releases alkalinity (Figure 6.14) and phosphate. A change in the flowpath of groundwater would have the same effect. This phenomenon of increased weathering of previously frozen till is likely happening at a slow rate throughout this region, but the phosphate is being taken up by plants on land and not released into the streams or it is being released and taken up by P-limited stream organisms while the alkalinity is released and because it is more biologically conservative, building up in the lake.

The extensive research based on the nearly 30 years of adding low levels of phosphate to a stretch of the Kuparuk River is summarized in chapter 7. Its aim was to determine the controls on the river biota and river productivity by fertilizing at a low level for the summer months. There was an immediate response by the diatoms, followed by increases in biomass and productivity of all the parts of the food web including the top predator, the arctic grayling. After nearly a decade of treatment, we were surprised to find that a genus of moss (*Hygrohypnum*) was rapidly covering the rocks of the river in the phosphorus-fertilized reach. There is no way of telling whether or not this dramatic shift in the dominant primary producer will happen in the future because of the possibility that all the phosphorus would be removed from the soil water and from small streams by the plants before higher concentrations reached the river.

Changes by Year 2100

The ecology of the upper Kuparuk River is not expected to change dramatically during the remainder of the 21st century. The rivers and adjacent streams will continue to be oligotrophic with low nutrients and low primary productivity. Diatoms on the surface of the rocks and insect larvae will be unchanged by the probable small changes in the water temperature and the indirect effects of higher rates of nutrient cycling in the tundra soils. Even though the precipitation is expected to increase significantly, increases in flow will be within the tolerance of the present organisms.

There are, however, two types of disturbances that may change this scenario and alter the survival of the arctic grayling, the top predator. These, thermokarst formation and drought, have already occurred to a limited extent. A thermokarst, formed on the nearby Toolik River, demonstrated that high amounts of nitrogen and phosphorus as well as eroding soil are added to the river. The soil eroding from a thermokarst would strongly affect the spawning of grayling. If this disturbance came to affect a number of streams, the grayling populations could be threatened. The droughts, which are becoming more frequent, are a larger threat to grayling, however, because if one occurred during the late summer period when the grayling absolutely had to migrate to the overwintering lake, most of the population would be wiped out. Presumably some of the YOY might survive in a few springs or deep holes, but would a population survive the loss of all the adults? Thus, if the late summer droughts became more frequent and lasted longer, it is possible that the grayling, the only species of fish in the river and adjacent streams, could be extinct by 2100. It is also true that populations of grayling are found in lakes near Toolik. Are these genetically different from those found in streams? Would these adapt to stream life?

Lake Ecology Response to Changes in Northern Alaska

Disturbances and Changes in Lakes

The overall conclusion of chapter 8 on lakes was that there were few observed changes or trends in the physics, chemistry, and biology of lakes in the Toolik region during the 37 years of study, from 1975 to the present. Those changes we did observe were driven by changes in disturbance caused by fire, by thermal erosion of permafrost and increased weathering, or by thermokarst formation, rather than by the direct impact of temperature changes.

The ice cover in Toolik Lake typically lasts until mid-June; between 2000 and 2009, the date of the last ice in the lake ranged from June 8 to 1 July. One reason for variation in the date is the snow cover in the late spring. A late-lasting snow cover with a high albedo protects the ice cover; but when the ice is wind-blown and without a snow cover, the melt begins earlier. This variation in ice-out dates, which affects the timing of summer temperatures and stratification in Toolik Lake, and the lack of warmer summer air temperatures (chapter 2) results in no statistically significant warming of lake temperatures. For example, in Figure 8.17 the mean annual epilimnion temperature in Toolik in July from 1975 to 2007 shows no warming.

Direct Response of Lake Organisms to Temperature Change

Algae, the basis of the lake food web, do not show a species shift over the range of temperatures in Toolik and other lakes. The same species of algae are found beneath the ice and when the water temperatures reach 15°C later in the summer (O'Brien et al. 1997). Algae do increase their photosynthesis when the temperature is increased in laboratory experiments. However, as described in chapter 8, algae in natural systems respond much more to changes in light and to changes in nutrient concentrations than to changes in temperature. In fact, primary productivity is strongly limited by low concentrations of both nitrogen and phosphorus (Whalen et al. 2008). In Toolik and other nearby lakes, there is a brief burst of primary production after snowmelt and ice off, but nutrients are soon used up and the productivity falls to a low value. There is no evidence that so far the loading of nutrients into Toolik Lake has changed. The lake chlorophyll content, an indicator of algal productivity, has not changed appreciably from 1985 to the present (Figure 8.18).

Chapter 8 details several temperature effects on animals from lakes near Toolik. A study of effects of increased temperatures on zooplankton from Toolik and other lakes revealed that *Daphnia middendoffiana*, the most common species, is near its thermal maximum in shallow ponds. This species is adapted to thrive in arctic conditions; when warming occurs, it will be replaced by invasive species, probably other *Daphnia*, either from nearby lakes or from lower latitudes.

Natural populations of three species of fish from lakes near Toolik also show an effect from warming. These fish (lake trout, arctic grayling, and arctic char) showed a decrease in their fitness or condition as estimated from the length-weight relationship. In warm summers when epilimnetic temperatures exceeded 15°C for extended periods, a decrease in the slope of the length-weight relationship occurred for all three fish (Figure 8.19). A rise in the epilimnetic temperature would likely result in dramatic declines in fitness. The shallow nature of many of the lakes coupled with the lack of food resources below the epilimnion may narrow the window of suitable habitat and make it difficult for these fish populations to persist.

Effects of Major Disturbances

Both climate warming and fire will increase thermokarst activity in northern Alaska but further data are needed before it is certain that this has happened on the North Slope in the same way as it has already happened on the South Slope. As described by Bowden et al. (2008), these events greatly increase the solute transport to streams and lakes, especially nitrogen and phosphorus (Figure 7.35). In addition, the clarity of recipient lakes is reduced for many years (Figure 7.34). There is no detailed information yet describing the effect on a specific lake except anecdotal information on starving lake trout found in a lake where a thermokarst slump had occurred (M. McDonald, personal communication) and the observation that the clarity of Lake NE14, which was impacted by a thermokarst failure on its shoreline, has been increasing over time since the initial failure made the lake turbid (G. Kling, personal observation).

Response of the Whole Lake System

An initial study on the effects of climate on fish using a bioenergetics simulation model showed that increased lake temperatures would greatly reduce the ability of lake trout to thrive in arctic lakes (McDonald et al. 1996). Lake trout prefer water near 10°C with oxygen concentrations above 6 mg L^{-1} and show stress at temperature above 15°C. The lack of abundant deep-water habitat in most lakes of the region means that usable summer habitat for lake trout would decrease by 30% if epilimnetic temperatures rise by 2°C (Hobbie et al. 1999). This reduction of habitat is exacerbated if nutrient loading to the lakes increases phytoplankton productivity, resulting in decreases in hypolimnetic oxygen concentrations, or if ground water inflows increase, leading to chemical stratification which reduces mixing and also contributes to increased anoxia.

Lakes in Northern Alaska in 2100

If the projections of an increase in air temperature by 0.2°C –0.5°C are correct, then there will be no dramatic change in the ecology of the lakes of the Toolik region by 2100. This is not enough of a change in temperature to increase appreciably the amount of nutrients reaching the lakes or to change the summer temperatures of the lakes to an important degree. However, it is likely that increasing temperatures will further erode the thermal stability of permafrost in the catchments or in the talik beneath lakes, which will lead to continued alterations in lake chemistry (Figure 6.14). Exactly how much the input of nutrients and turbidity from thermokarst formation will change is unknown, but this disturbance, rare at the present time, may increase from climate warming and from fire. It is clear that the most vulnerable part of the lake ecosystem is the fish. Lake trout in particular will be the first to show an effect of climate warming, and this may well happen by 2100.

Summary

Past and Present Changes at Toolik: Ecological Consequences

1. The climate of the whole North Slope has warmed by more than 3°C over the past 60 years. Despite this regional warming, the climate record at the Toolik Field Station shows no statistically significant warming trend in the annual average air temperatures. This could be largely an artifact of the rather brief Toolik record (1989–2010).
2. Significant changes have accumulated over time in the glaciers, soils, streams, lakes, and tundra near Toolik that indicate warming is occurring. These changes include:
   a. Shrinking of mountain glaciers within 30 km of Toolik;
   b. Warming of the permafrost at a depth of 20 m in the soil by 0.8°C over 20 years;
   c. Increasing of vegetation canopy height, leaf area, and "greenness" (satellite view). These are probably related to a change in aboveground biomass and primary production. Based on experimental results, we conclude that there has been an increase in microbial decomposition that releases more nitrogen than previously from the large storehouse in soil organic matter. This change in the release could be caused by an increase in temperatures in the soil, by a longer growing season (in this example the period of time when the microbes are active), or by both.
   d. Changing chemistry in streams and a doubling of the alkalinity in lakes. The most reasonable explanation is that warming has led to increased weathering of previously frozen glacial till either by increasing the thickness of the active layer or by increasing the size of the thaw bulb under streams and lakes, which exposes more mineral soil.
3. In aquatic and terrestrial ecosystems the direct effects of temperature on plants and animals are small, or changes are not sustainable without input of nutrients; direct responses to temperature only will be slow. The associated nutrient responses depend on the temperature-caused changes in nutrient turnover by microbes or on changes related to nutrient inputs from disturbances.
4. Climate-related changes in disturbances including fire but especially those related to permafrost may be more important than the direct effects of climate warming.
5. The response of aquatic ecosystems to warming is linked closely to the response of terrestrial ecosystems through runoff of nutrients, sediments, and organic matter. These factors are very slow to change and are well buffered so that increases in nutrients in runoff are very small; as a result,

the biology of streams and lakes has shown little response to warming, and no drastic change is expected at the century scale. The one exception is the evidence that unusually warm summers have had a negative impact on fish, particularly the arctic grayling and lake trout.

6. The effect of changing seasonality on arctic biology is not discussed in this book; it is under study by several projects and is therefore a work in progress. Topics include the timing of bird arrival and life-history stages, the emergence and peak abundance of their insect prey, the phenology of plant flowering and the timing of emergence of their insect pollinators, the changes in timing of stream flow and fish migration, and the late-season and winter nutrient mineralization and the timing of nutrient uptake by plants.

Future Changes at Toolik: Predictions of Ecological Consequences

Climate models indicate a warmer and wetter environment for northern Alaska by 2100. The warming effects on the terrestrial ecology near Toolik by the end of the century will result in more biomass of grasses and shrubs. The aquatic systems will be somewhat more productive but mostly unchanged. The large, top-of-the-food-web fish are the most vulnerable part of the ecosystem, and some species may disappear. Finally, the warmer and wetter climate will lead to an increase in a major disturbance, thermokarst, caused by the melting of ice in permafrost and resulting in slippage of sections of hillsides, compaction of land, and transport of sediments and nutrients to streams and lakes. The prediction of a wetter environment is under discussion and argument in the scientific community and awaits better understanding and models.

The ecological information collected by the Arctic LTER and associated projects has focused on effects of warming and related changes on communities and species. It is more difficult to predict the ecosystem effects. Yet, it is well known that species-composition changes in terrestrial and aquatic ecosystems have important feedback (or feed-forward) effects that must be considered in making long-term predictions. These effects may cascade through the trophic structure, through changes in carbon and nutrient turnover, habitat structure, food-resource availability and quality, biodiversity, or in the physics of energy balance in ecosystems. As concerns rise that thawing permafrost soils may release greenhouse gases to the atmosphere and accelerate global warming (Serreze et al. 2009), we suggest that it is the ecological interactions in tundra ecosystems which will govern how a changing Arctic system affects the climate at lower latitudes in this century.

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REVIEW

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Climate change and the permafrost carbon feedback

E. A. G. Schuur[1,2], A. D. McGuire[3], C. Schädel[1,2], G. Grosse[4], J. W. Harden[5], D. J. Hayes[6], G. Hugelius[7], C. D. Koven[8], P. Kuhry[7], D. M. Lawrence[9], S. M. Natali[10], D. Olefeldt[11,12], V. E. Romanovsky[13,14], K. Schaefer[15], M. R. Turetsky[11], C. C. Treat[16] & J. E. Vonk[17]

Large quantities of organic carbon are stored in frozen soils (permafrost) within Arctic and sub-Arctic regions. A warming climate can induce environmental changes that accelerate the microbial breakdown of organic carbon and the release of the greenhouse gases carbon dioxide and methane. This feedback can accelerate climate change, but the magnitude and timing of greenhouse gas emission from these regions and their impact on climate change remain uncertain. Here we find that current evidence suggests a gradual and prolonged release of greenhouse gas emissions in a warming climate and present a research strategy with which to target poorly understood aspects of permafrost carbon dynamics.

In high-latitude regions of Earth, temperatures have risen 0.6 °C per decade over the last 30 years, twice as fast as the global average[1]. This is causing normally frozen ground to thaw[2–4], exposing substantial quantities of organic carbon to decomposition by soil microbes. This permafrost carbon is the remnant of plants and animals accumulated in perennially frozen soil over thousands of years, and the permafrost region contains twice as much carbon as there is currently in the atmosphere[5,6]. Conversion of just a fraction of this frozen carbon pool into the greenhouse gases carbon dioxide (CO_2) and methane (CH_4) and their release into the atmosphere could increase the rate of future climate change[7]. Climate warming as a result of human activities causes northern regions to emit additional greenhouse gases to the atmosphere, representing a feedback that will probably make climate change happen faster than is currently projected by Earth System models. The critical question centres on how fast this process will occur, and recent publications differ in their outlook on this issue. Abrupt releases of CH_4 forecast to cause trillions of dollars of economic damage to global society[8] contrast with predictions of slower, sustained greenhouse gas release that, although substantial, would give society more time to adapt[1,9]. This range of viewpoints is due in part to the wide uncertainty surrounding processes that are only now being quantified in these remote regions.

Here we provide an overview of new insights from a multi-year synthesis of data with the aim of constraining our current understanding of the permafrost carbon feedback to climate, and providing a framework for developing research initiatives in the permafrost region[10,11]. We begin by reviewing new research, much of it published since the Intergovernmental Panel on Climate Change (IPCC)'s Fifth Assessment Report (AR5)[1], on the size of the carbon pool stored in the permafrost region. Synthesis research has enlarged the number of observations in the permafrost region soil carbon pool database tenfold[12], and confirms that tremendous quantities of carbon accumulated deep in permafrost soils are widespread[5,6]. We then discuss new long-term laboratory incubations of these permafrost soils that reveal that a substantial fraction of this material can be mineralized by microbes and converted to CO_2 and CH_4 on timescales of years to decades, which would contribute to near-term climate warming. Initial estimates of greenhouse gas release point towards the potential for substantial emissions of carbon from permafrost in a warmer world, but these could still be underestimates. Field observations reveal that abrupt thaw processes are common in northern landscapes, but our review shows that mechanisms that speed thawing of frozen ground and release of permafrost carbon are entirely absent from the large-scale models used to predict the rate of climate change.

Bringing together this wealth of new observations, we propose that greenhouse gas emissions from warming permafrost are likely to occur at a magnitude similar to other historically important biospheric carbon sources (such as land-use change) but that will be only a fraction of current fossil-fuel emissions. At the proposed rates, the observed and projected emissions of CH_4 and CO_2 from thawing permafrost are unlikely to cause abrupt climate change over a period of a few years to a decade. Instead, permafrost carbon emissions are likely to be felt over decades to centuries as northern regions warm, making climate change happen faster than we would expect on the basis of projected emissions from human activities alone. This improved knowledge of the magnitude and timing of permafrost carbon emissions based on the synthesis of existing data needs to be integrated into policy decisions about the management of carbon in a warming world, but at the same time may help temper the worst fears about the impact of carbon emissions from warming northern high-latitude regions.

Permafrost carbon pool

The first studies that brought widespread attention to permafrost carbon estimated that almost 1,700 billion tons of organic carbon were stored in terrestrial soils in the northern permafrost zone[6,7,13]. The recognition of this vast pool stored in Arctic and sub-Arctic regions was in part due to substantial carbon stored at depth (>1 m) in permafrost, below the traditional zone of soil carbon accounting[14]. Deeper carbon measurements were initially rare, and it was not even possible to quantify the uncertainty for the permafrost carbon pool size estimate. However, important new syntheses continue to report large quantities of deep carbon preserved in permafrost at many previously unsampled locations, and that a substantial fraction of this deep

[1]Center for Ecosystem Science and Society and Department of Biological Sciences, Northern Arizona University, Flagstaff, Arizona 86011, USA. [2]Department of Biology, University of Florida, Gainesville, Florida 32611, USA. [3]US Geological Survey, Alaska Cooperative Fish and Wildlife Research Unit, University of Alaska Fairbanks, Alaska 99775, USA. [4]Alfred Wegener Institute Helmholtz Centre for Polar and Marine Research, 14473 Potsdam, Germany. [5]US Geological Survey, Menlo Park, California 94025, USA. [6]Climate Change Science Institute and Environmental Sciences Division, Oak Ridge National Laboratory, Oak Ridge, Tennessee 37831, USA. [7]Department of Physical Geography, Stockholm University, 10691 Stockholm, Sweden. [8]Earth Sciences Division, Lawrence Berkeley National Laboratory, Berkeley, California 94720, USA. [9]National Center for Atmospheric Research, Boulder, Colorado 80305, USA. [10]Woods Hole Research Center, Falmouth, Massachusetts 02540, USA. [11]Department of Integrative Biology, University of Guelph, Guelph, Ontario N1G 2W1, Canada. [12]Department of Renewable Resources, University of Alberta, Edmonton, Alberta T6G 2H1, Canada. [13]Geophysical Institute, University of Alaska Fairbanks, Fairbanks, Alaska 99775, USA. [14]Tyumen State Oil and Gas University, Tyumen, Tyumen Oblast 625000, Russia. [15]National Snow and Ice Data Center, Boulder, Colorado 80309, USA. [16]Earth Systems Research Center, Institute for the Study of Earth, Oceans and Space, University of New Hampshire, Durham, New Hampshire 03824, USA. [17]Department of Earth Sciences, Utrecht University, 3584 CD Utrecht, The Netherlands.

permafrost carbon is susceptible to future thaw[15]. The permafrost carbon pool is now thought to comprise organic carbon in the top 3 m of surface soil, carbon in deposits deeper than 3 m (including those within the yedoma region, an area of deep sediment deposits that cover unglaciated parts of Siberia and Alaska[16–18]), as well as carbon within permafrost that formed on land during glacial periods but that is now found on shallow submarine shelves in the Arctic. Recent research has expanded our knowledge considerably while at the same time highlighting remaining gaps in our understanding of this vulnerable carbon pool[19].

Surface carbon

The new northern permafrost zone carbon inventory reports the surface permafrost carbon pool (0–3 m) to be 1,035 ± 150 Pg carbon (mean ± 95% confidence interval, CI)[12,20] (where 1 Pg = 1 billion tons) (Fig. 1a). This estimate supported the original studies while improving precision by increasing the number of deeper (>1 m) sampling locations tenfold. This surface permafrost carbon pool is substantial. The rest of Earth's biomes, excluding the Arctic and boreal regions, are thought to contain 2,050 Pg carbon in the surface 3 m of soil[21]. Even though these northern regions account for only 15% of global soil area, the 0–3 m global soil carbon pool is increased by 50% when fully accounting for the carbon stored deeper in permafrost zone soil profiles.

Deep carbon in yedoma

Processes that accumulate carbon deep into permafrost soils do not stop at 3 m depth, and our previously limited understanding of those deep carbon deposits (>3 m depth) has been improved. In particular, several new estimates have emerged for carbon that accumulated during, and since, the last Ice Age in the yedoma region in Siberia and Alaska[16–18]. These new data support previous findings of relatively high carbon concentrations in permafrost soil at depth, but revised the understanding of total carbon stock by improving the estimates of spatial extent, type of deposit, sediment depth, and ground ice content. These deep, perennially frozen sediments are particularly ice-rich, where ice occupies 50%–80% of the ground volume[22,23]. Although this excess ice does not alter soil carbon concentration, it affects the total carbon inventory contained in a particular volume of soil, decreasing carbon stocks per unit soil volume by 22%–50% compared to previous estimates[24]. Because of the continued difficulty of measuring total ground ice content and total sediment depth, carbon pool estimates for the yedoma region still range by twofold even as new data from this region have accumulated. This region is now thought to contain between 210 ± 70 Pg carbon (ref. 16) and 456 ± 45 Pg carbon (ref. 18), still supporting the original accounts of several hundred billion tons of carbon stored deep in the permafrost even when recalculated with new observations.

Deep carbon outside the yedoma region

While new measurements of deep carbon have been largely focused on the 1.2 million square kilometres of the yedoma region in recent years, other areas in the northern permafrost zone with thick loose sedimentary material may also contain substantial organic carbon pools in permafrost (Fig. 1b). The major Arctic river deltas are now thought to contain 91 ± 39 Pg carbon (95% CI)[12], while carbon contained in the approximately 5 million square kilometres of thick (>5–10 m) sediments overlying bedrock outside the yedoma and river delta regions remain largely unknown. Taking the spatial extent of these poorly known permafrost areas, along with an estimated thickness in the tens of metres (similar to that of yedoma), and average carbon content of a few deep borehole soil samples, there could be an additional deep permafrost carbon pool of 350–465 Pg C outside the yedoma region (calculated using a depth interval of 3–10 m and carbon content of 11–14 kg C m^{-3}, which accounts for ground ice[25]).

Subsea permafrost carbon

Much of the inventory until this point has focused on terrestrial ecosystems where permafrost is currently sustained by cold winter air temperatures. But permafrost also exists below Arctic Ocean continental shelves, in particular the East Siberian Arctic Shelf, the largest and shallowest shelf on Earth. This permafrost is an extension of the terrestrial permafrost that existed during the last Ice Age, but became submerged when sea level rose during the late Pleistocene–Holocene transition, and at the beginning of the Holocene epoch. The shallow shelf area exposed as dry land in the area around Alaska and Siberia during the last Ice Age (<125 m current ocean depth), at almost 3 million square kilometres, is about 2.5 times the size of the current terrestrial yedoma region[16,26]. But the quantity of organic permafrost carbon stored beneath the sea floor is even more poorly quantified than on land and could be lower than it once was[27,28]. Subsea permafrost as a whole has been slowly degrading over thousands of years as relatively warm ocean water has warmed the newly submerged sea floor. Frozen sediments are thickest near the shore, where submergence with seawater occurred more recently than on the outer shelf, which is now underlain by discontinuous, patchy permafrost[29,30]. During this time of thaw, organic carbon was mineralized by microbes within the sediment in low-oxygen conditions that promote the formation of CH_4, reducing the pool of permafrost carbon remaining under the sea.

Taken together, the known pool of terrestrial permafrost carbon in the northern permafrost zone is 1,330–1,580 Pg carbon, accounting for surface carbon as well as deep carbon in the yedoma region and river deltas, with the potential for ~400 Pg carbon in other deep terrestrial permafrost sediments that, along with an additional quantity of subsea permafrost carbon, still remains largely unquantified.

Carbon decomposability

Permafrost carbon stocks provide the basis for greenhouse gas release to the atmosphere, but the rate at which this can happen is also controlled by the overall decomposability of organic carbon. Conceptual models and initial data on decomposability suggested that a portion of permafrost carbon is susceptible to rapid breakdown upon thaw[13,31]. But it has not been clear to what degree this could be sustained on the decade-to-century timescale of climate change, or what degree of variation exists within soils across the vast landscape of the permafrost zone. New research has confirmed that initial rates of permafrost carbon loss are potentially high, but continued observation reported declines in carbon loss rates over time, which might be expected as more labile carbon pools are exhausted[32]. This has highlighted the need for long-term observation under controlled conditions to estimate the potential decomposability of permafrost carbon. New data from a 12-year incubation of permafrost soil from Greenland showed that 50%–75% of the initial carbon was lost by microbial decomposition under aerobic and continuously unfrozen laboratory conditions over that time frame[33]. This experiment, of unprecedented length for permafrost soils compared to typical incubations that might be only weeks to months long[34,35], was then extended geographically in a new synthesis of long-term (>1 year) permafrost zone soil incubations. Soils from across the permafrost region showed similarly high potential for microbial degradation of organic carbon upon thaw in the laboratory, with a wider range of decade-long losses projected to be 1%–76% (Fig. 2a) under laboratory conditions[36].

A major cause of landscape-scale variation in decomposability across soils was linked to the carbon to nitrogen ratio of the organic matter, with higher values leading to more greenhouse gas release. This simple metric (the carbon to nitrogen ratio) is in part illustrated by grouping soils as organic (>20% C) with mean decade-long losses of 17%–34% (lower-to-upper 97.5% CI) and mineral (<20% C) with mean decade-long losses of 6%–13% (Fig. 2a). The metric takes into account the ability of microbes to process permafrost carbon for metabolism by breaking down organic carbon for energy, and to grow by acquiring nutrients such as nitrogen released during the decomposition process. Because carbon and nitrogen are often measured in soil surveys, maps of permafrost carbon pools can then be combined with the findings from laboratory incubations to project potential carbon emission estimates across the permafrost region to determine which regions could be emission hotspots in a warming climate. The location of such potential emission hotspots is expected to be affected by both the total pool of permafrost carbon and the potential for that carbon to be broken

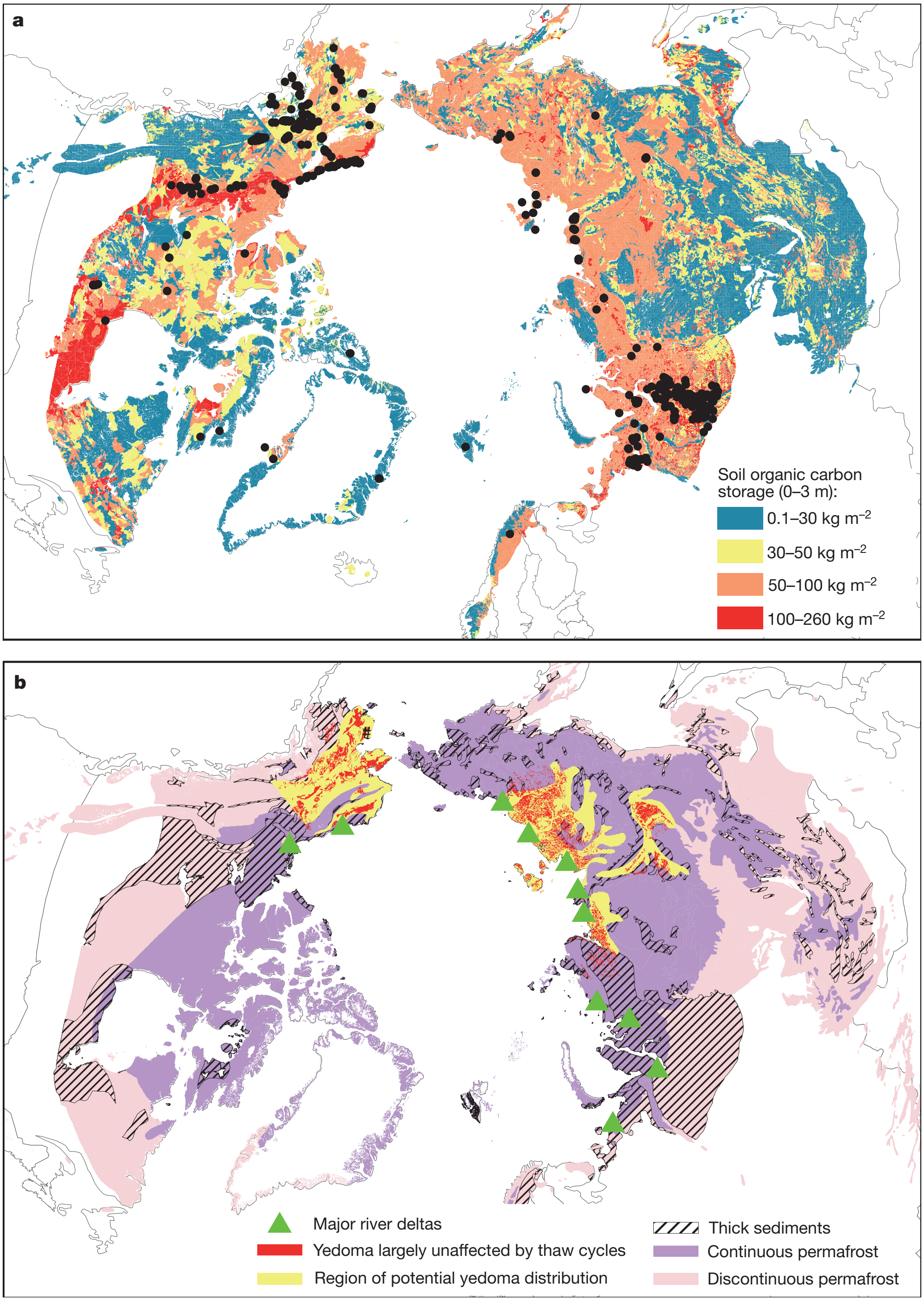


Figure 1 | Soil organic carbon maps. **a**, Soil organic carbon pool (kg C m^{-2}) contained in the 0–3 m depth interval of the northern circumpolar permafrost zone[12]. Points show field site locations for 0–3 m depth carbon inventory measurements; field sites with 1 m carbon inventory measurements number in the thousands and are too numerous to show. **b**, Deep permafrost carbon pools (>3 m), including the location of major permafrost-affected river deltas (green triangles), the extent of the yedoma region previously used to estimate the carbon content of these deposits[13] (yellow), the current extent of yedoma region soils largely unaffected by thaw-lake cycles that alter the original carbon content[17] (red), and the extent of thick sediments overlying bedrock (black hashed). Yedoma regions are generally also thick sediments. The base map layer shows permafrost distribution with continuous regions to the north having permafrost everywhere (>90%), and discontinuous regions further south having permafrost in some, but not all, locations (<90%)[96].

down by microbes after thaw as controlled by the energy and nutrients contained within the organic matter.

The inherent range of permafrost carbon decomposability across soil types also intersects with environmental conditions, and aerobic decomposition is only part of the story for northern ecosystems. While temperature control over decomposition is implicit when considering permafrost thaw, this region is characterized by widespread lakes, wetlands, and soils waterlogged as a result of surface drainage restricted by underlying permafrost. The lack of oxygen in saturated anaerobic soils and sediments presents another key control over emissions from newly thawed permafrost carbon. Comparing the results from the aerobic permafrost soil incubation synthesis[36] with those from another circumpolar synthesis of anaerobic soil incubations[37]

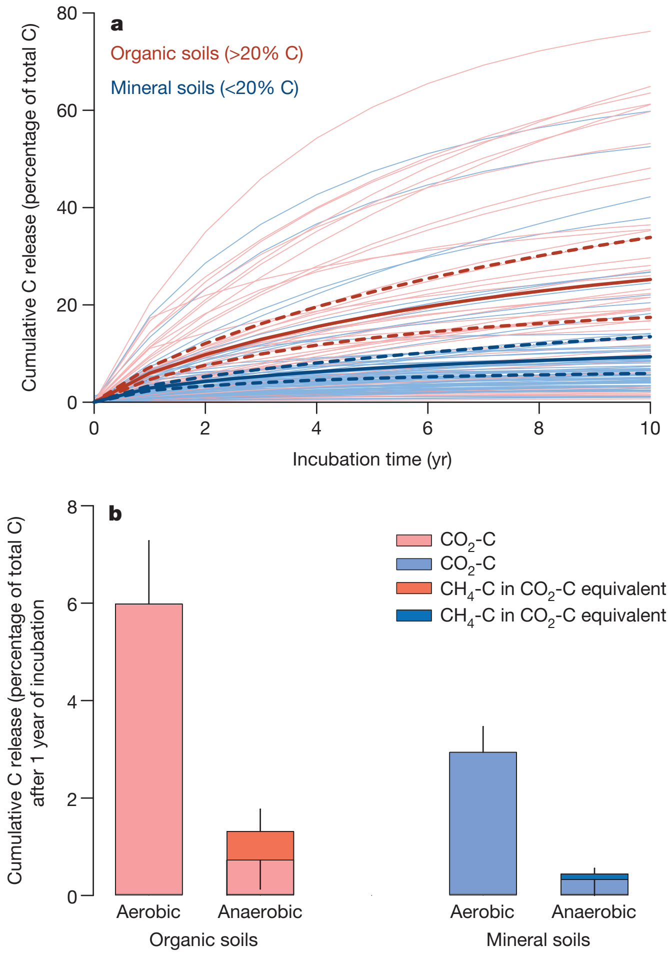


Figure 2 | Potential cumulative carbon release. Data are given as a percentage of initial carbon. **a**, Cumulative carbon release after ten years of aerobic incubation at a constant temperature of 5 °C. Thick solid lines are averages for organic (red, $N = 43$) and mineral soils (blue, $N = 78$) and thin solid lines represent individual soils to show the response of individual soils. Dotted lines are the averages of the 97.5% CI for each soil type. **b**, Cumulative carbon release after one year of aerobic and anaerobic incubations (at 5 °C). Darker colours represent cumulative CH_4-carbon calculated as CO_2-carbon equivalent (for anaerobic soils) on a 100-year timescale according to ref. 38. Positive error bars are upper 97.5% CI for CO_2-carbon and negative error bars are lower 97.5% CI for CH_4-carbon. $N = 28$ for organic soils and $N = 25$ for mineral soils in anaerobic incubations. Aerobic cumulative carbon release is redrawn from ref. 36 and anaerobic cumulative carbon release is calculated based on ref. 37.

shows that cumulative carbon emissions, over an equal one-year incubation time frame, are, on average, 78%–85% lower than those from aerobic soils (Fig. 2b). Specialized microbes release CH_4 along with CO_2 in these environments, and the more potent (that is, it affects climate change more powerfully) greenhouse gas CH_4 in the atmosphere can partially offset a decreased decomposition rate. While mean quantities of CH_4 are 3% (in mineral soils) to 7% (in organic soils) that of CO_2 emitted from anaerobic incubations (by weight of carbon), these mean CH_4 values represent 25% (in mineral soil) to 45% (in organic soil) of the overall potential impact on climate over a 100-year timescale when accounting for CH_4 (ref. 38). Across the mosaic of ecosystems in the permafrost region, controlled laboratory observations brought together here imply that, in spite of the more potent greenhouse gas CH_4, a unit of newly thawed permafrost carbon could have a greater impact on climate over a century if it thaws and decomposes within a drier, aerobic soil as compared to an equivalent amount of carbon within a waterlogged soil or sediment.

Controlled laboratory work is critical for identifying the key mechanisms for potential greenhouse gas release from permafrost carbon, but some important processes are difficult to address with incubation experiments. For example, CH_4 generated from permafrost carbon can be oxidized in aerobic soil layers above the water table and released to the atmosphere as CO_2 instead. This effect can be modified by vegetation, for example, sedge stems acting as pipes provide a pathway for CH_4 to avoid oxidation and to escape to the atmosphere[39]. A synthesis of field CH_4 emission rates showed that sedge-dominated sites had emission rates 2–5 times higher[40], due in part to sedges allowing the physical escape of CH_4, as well as providing more decomposable carbon to the microbial community[41,42]. But even with sedges, it is likely that CH_4 oxidation as a whole would decrease the warming impact of permafrost carbon decomposing in a waterlogged environment compared to what was measured from a laboratory potential. Incubation results, while needing to be interpreted carefully, are useful for scaling the potential of permafrost soils to release greenhouse gases upon thaw, and also for helping to quantify the fraction of soil carbon that is likely to remain relatively inert within the soil after thaw.

Projecting change

A number of ecosystem and Earth system models have incorporated a first approximation of global permafrost carbon dynamics. Recent key improvements include the physical representation of permafrost soil thermodynamics and the role of environmental controls, in particular the soil freeze/thaw state, on decomposition of organic carbon[43–45]. These improved models, which specifically address processes known to be important in permafrost ecosystems but that were missing from earlier model representations, have been key for forecasting the potential release of permafrost carbon with warming, and the impact this would have on the rate of climate change. Model scenarios show potential carbon release from the permafrost zone in the range 37–174 Pg carbon by 2100 under the current climate warming trajectory (Representative Concentration Pathway RCP 8.5), with an average across models of 92 ± 17 Pg carbon (mean ± s.e.) (Fig. 3)[45–52]. Furthermore, thawing permafrost carbon is forecasted to impact global climate for centuries, with models, on average, estimating that 59% of total permafrost carbon emissions will occur after 2100. While carbon releases over these time frames are understandably uncertain, they illustrate the momentum of a warming climate that thaws near-surface permafrost, causing a cascading release of greenhouse gases as microbes slowly decompose newly thawed permafrost carbon. At the scale of these models not all differentiated between CO_2 and CH_4 loss, but expert assessment, a method for surveying expert knowledge, placed CH_4 losses at about 2.3% of total future emissions from the permafrost zone[53,54]. This has the effect, in the expert assessment, of increasing the warming potential of released carbon by 35%–48% when accounting for the more potent greenhouse gas CH_4 over a 100-year timescale.

Within the wide uncertainty of forecasts, some broader patterns are just beginning to emerge. Models vary widely when predicting the current pool of permafrost carbon, which is the source of future carbon emissions in a warmer world. The model average permafrost carbon pool size was estimated at 771 ± 100 Pg carbon (mean ± s.e.), about half as much as the measurement-based estimate, potentially related in part to the fact that models mostly represented carbon to only 3 m depth. A smaller modelled carbon pool could, in principle, constrain forecasted carbon emissions. Normalizing the emissions estimates from the dynamic models by their initial permafrost carbon pool size, 15% ± 3% (mean ± s.e.) of the initial pool was expected to be lost as greenhouse gas emissions by 2100[55]. This decrease in the permafrost carbon pool is similar, but somewhat higher, than the 7%–11% (95% CI) loss predicted by experts[53,54], and the relatively constant fraction across model estimates does hint at the importance of pool size in constraining carbon emissions. However, sensitivity to both modelled Arctic climate change, as well as the responses of soil temperature, moisture and carbon dynamics, are important controls over emissions predictions within these complex models, not pool size alone[44,56,57]. Full diagnosis of the important parameters that regulate the permafrost carbon feedback is not currently possible from the small number of modelling studies that exist, but the estimates do seem to converge on a vulnerable fraction of permafrost carbon that seems to be in line with other approaches.

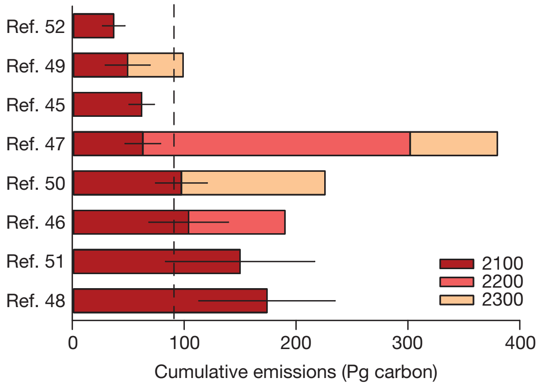


Figure 3 | Model estimates of potential cumulative carbon release from thawing permafrost by 2100, 2200, and 2300. All estimates except those of refs 50 and 46 are based on RCP 8.5 or its equivalent in the AR4 (ref. 97), the A2 scenario. Error bars show uncertainties for each estimate that are based on an ensemble of simulations assuming different warming rates for each scenario and different amounts of initial frozen carbon in permafrost. The vertical dashed line shows the mean of all models under the current warming trajectory by 2100.

These dynamic models also simultaneously assess the countering influence of plant carbon uptake, which may in part offset permafrost carbon release. Warmer temperatures, longer growing seasons, elevated CO_2, and increased nutrients released from decomposing organic carbon may all stimulate plant growth[58]. New carbon can be stored in larger plant biomass or deposited into surface soils[59]. A previous generation of Earth system models that did not include permafrost carbon mechanisms but did simulate changes in plant carbon uptake estimated that the vegetation carbon pool could increase by 17 ± 8 Pg carbon by 2100, with increased plant growth also contributing to new soil carbon accumulation of similar magnitude[60]. The models reviewed here that do include permafrost carbon mechanisms (as well as many of the mechanisms that stimulate plant growth that were used in the previous generation of models) generally indicate that increased plant carbon uptake will more than offset soil carbon emissions from the permafrost region for several decades as climate becomes warmer[45,46,48]. Over longer timescales and with continued warming, however, microbial release of carbon overwhelms the capacity for plant carbon uptake, leading to net carbon emissions from permafrost ecosystems to the atmosphere. Modelled carbon emissions projected under various warming scenarios translate into a range of 0.13–0.27 °C additional global warming by 2100 and up to 0.42 °C by 2300, but currently remain one of the least constrained biospheric feedbacks to climate[1].

Abrupt permafrost thaw

Recent progress towards predicting change in permafrost carbon dynamics focuses mostly on gradual top-down thawing of permafrost. However, increasing evidence from the permafrost zone suggests that abrupt permafrost thaw may be the norm for many parts of the Arctic landscape[17,18,61,62] (Fig. 4). Abrupt permafrost thaw occurs when warming melts ground ice, causing the land surface to collapse into the volume previously occupied by ice. This process, called thermokarst, alters surface hydrology. Water is attracted towards collapse areas, and pooling or flowing water in turn causes more localized thawing and even mass erosion. Owing to these localized feedbacks that can thaw through tens of metres of permafrost across a hillslope within only a few years, permafrost thaw occurs much more rapidly than would be predicted from changes in air temperature alone. This raises the question of whether key complexity is missing from large-scale model projections that are based on first approximations of permafrost dynamics.

Abrupt thaw occurs only at point locations but often causes much deeper permafrost thaw to occur more rapidly. This is in contrast to top-down thawing, which occurs across the entire landscape but affects only the permafrost surface. New regional research is beginning to reveal that a large fraction of permafrost carbon is vulnerable to abrupt thaw. For example, since the end of the last Ice Age, thermokarst thaw-lake cycles have affected 70% of the yedoma permafrost deposits in Siberian lowlands[17]. These cycles occur when abrupt permafrost thaw forms lakes that can drain over time, allowing sediments and carbon to refreeze into permafrost, while elsewhere new thaw lakes form and repeat this cyclic process (Fig. 4a, c). Abrupt thaw in upland regions, where water does not generally pool and form lakes, often creates gullies and slump features that can erode permafrost carbon into streams, rivers and lakes (Fig. 4b, d). These thaw features can also be widespread but are not as well recognized as are thaw lakes; over 7,500 upland thaw features were mapped within a 1,700-square-kilometre foothill region of Alaskan tundra[49]. Studies such as these illustrate a widespread influence of abrupt thaw in both upland and lowland permafrost landscapes, even though they do not provide a chronology of change.

Climate change is expected to increase the initiation and expansion of abrupt thaw features, potentially changing the rate of this historic disturbance cycle[62–65]. Wetland expansion due to abrupt thaw has affected 10% of a peatland landscape in northwestern Canada since the 1970s, with the fastest expansion occurring in the past decade[66]. Landscape lake cover is also affected by abrupt thaw, with net change being the sum of both lake expansion and drainage. The area of small open-water features around Prudhoe Bay on the Alaskan tundra has doubled since 1990 (ref. 67). In northwestern Alaska, lake initiation has increased since 1950, while lake expansion rates remained steady[68]. In general, landscape lake cover is currently believed to be stable or increasing within the continuous permafrost zone, whereas there is a tendency for lake drainage and vegetation infilling to dominate over lake expansion in the discontinuous permafrost zone[68–72].

Abrupt thaw influences carbon emissions to the atmosphere by exposing previously frozen carbon to microbial processes, and also by altering the hydrology that is critical for determining the balance of CO_2 and CH_4 emissions. Some of the highest CH_4 emissions in the permafrost region have been observed in lakes and wetlands formed through abrupt thaw[40,73]. At the same time, accumulation of new carbon under anaerobic conditions in peat[74] and in lake sediments[18] can be greater than permafrost carbon losses, at least in some ecosystems. In this way, anaerobic environments replace freezing temperatures as a mechanism for soil carbon stabilization, keeping greenhouse gas emissions lower than they would otherwise be[75]. In contrast, abrupt thaw processes in other landscapes clearly accelerate carbon loss. Drained lakes and lowered water tables will expose previously waterlogged carbon to microbial decomposition in aerobic conditions with relatively higher rates of carbon emissions. Also, lateral movement of permafrost carbon by leaching or erosion into lakes, rivers and the ocean[76–78] can increase loss, as carbon may be more readily mineralized through microbial and photochemical processes after mobilization[79,80]. How carbon cycling at the landscape scale will change under a warming climate will depend critically on how much of the landscape becomes wetter or drier, a question difficult to answer. It is clear that abrupt thaw is an important mechanism of rapid permafrost degradation, with widespread but varying influences on hydrology and carbon cycling. Yet abrupt thaw is not included in large-scale models, suggesting that important landscape transformations are not currently being considered in forecasts of permafrost carbon–climate feedbacks. This is in part due to the fact that we do not know at this stage what the relative importance of abrupt to gradual thaw across the landscape is likely to be.

Subsea carbon emissions

A majority of the observations and all of the modelling to date has focused on potential emissions from permafrost carbon on land. This is in part because subsea permafrost is buffered from recent climate change by the overlying ocean, and because ocean incursion at the end of the Ice Age has already been thawing and potentially reducing the pool of permafrost carbon under the sea. However, aside from organic carbon stored in permafrost, the sea bed underlying Arctic shelves also accumulated fossil CH_4 stored either as free CH_4 gas or as clathrates (CH_4-ice lattices that are stable at pressures and temperatures found at depth in this region). Layers of permafrost may serve as a physical barrier to the release of this CH_4 gas from the sediment into the water column and eventually the atmosphere. These

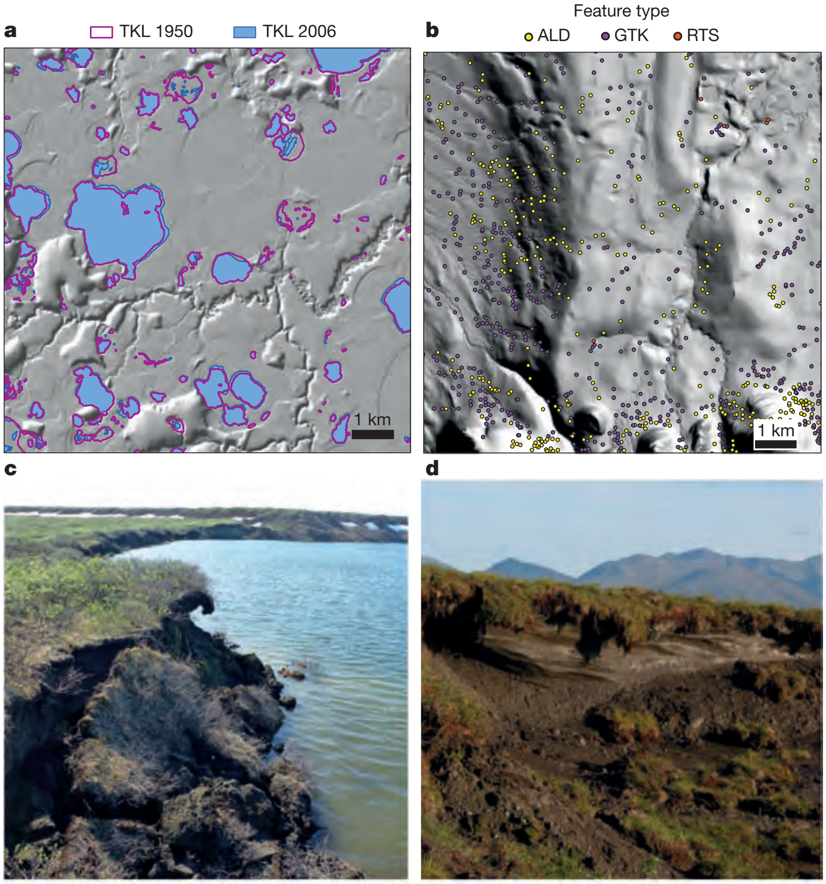


Figure 4 | Abundance of abrupt thaw features in lowland and upland settings in Alaska. Left panels (**a**, **c**) show thermokarst lake (TKL) abundance, expansion, and drainage on the Seward Peninsula, Northwest Alaska, between 1950 and 2006[68], with collapsing permafrost banks (photo credit G.G.). Right panels (**b**, **d**) show extensive distribution of ground collapse and erosion features (ALD, active layer detachment slide; RTS, retrogressive thaw slump; GTK, thermal erosion gullies) in upland tundra in a hill slope region in Northwest Alaska[61], and thawing icy soils in a retrogressive thaw slump (photo credit E.A.G.S.).

shallow shelves are also depositional areas for carbon from the erosion of coastal permafrost carbon and from inland permafrost carbon transported by Arctic rivers[81]. Together, these processes form ocean hotspots that are documented sources of high CH_4 emissions to the atmosphere[82,83], similar to hotspots formed in Arctic lakes on land[58]. New quantification has estimated that 17 Tg of CH_4 per year (where 1 Pg = 1,000 Tg) is emitted from the East Siberian Arctic Shelf after accounting for both diffusive and point-source bubble emissions[83]. Although this amount represents an increase from what was previously estimated for this region[27], this is probably because of improved observations of these emissions that may have been persistent over the thousands of years of land submergence. Climate warming, sea-ice decline, and increasing storminess have been linked to a 2.1 °C increase in bottom water (<10 m depth) temperature since the mid-1980s in this region[84]. Degradation of subsea permafrost from above by climate warming, and also from below by ongoing geothermal heat, will tend to increase new pathways between CH_4 storage areas deeper in the sediments and the sea floor[30]. But it is not known whether meaningful increases in CH_4 emissions via these processes could occur within this century, or whether they are more likely to manifest over a century or over millennia[84]. What is clear is that it would take thousands of years of CH_4 emissions at the current rate to release the same quantity of CH_4 (50 Pg) that was used in a modelled ten-year pulse to forecast tremendous global economic damage as a result of Arctic carbon release[8], making catastrophic impacts such as those appear highly unlikely[85–87].

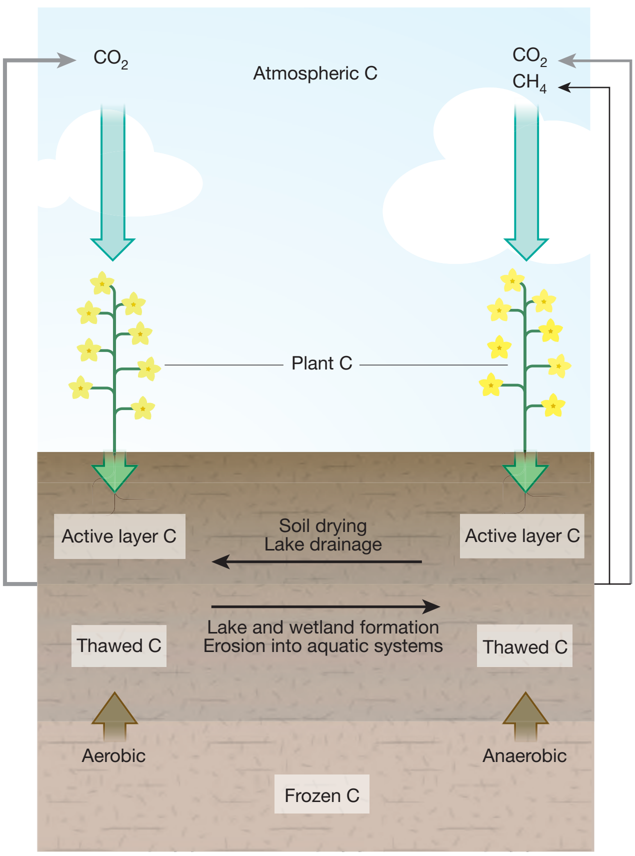
Permafrost and the global carbon cycle

Carbon pools in permafrost regions represent a large reservoir vulnerable to change in a warming climate. While some of this carbon will continue to persist in soils and sediments over the long term, our understanding that a substantial fraction of this pool is susceptible to microbial breakdown once thawed has been verified at the landscape scale (Box 1 and the Box 1 Figure). The exponential nature of microbial decomposition and CO_2 and CH_4 release over time means that the initial decades after thaw will be the most important for greenhouse gas release from any particular unit of thawed soil. Our expert judgement is that estimates made by independent approaches, including laboratory incubations, dynamic models, and expert assessment, seem to be converging on ~5%–15% of the terrestrial permafrost carbon pool being vulnerable to release in the form of greenhouse gases during this century under the current warming trajectory, with CO_2-carbon comprising the majority of the release. There is uncertainty, but the vulnerable fraction does not appear to be twice as high or half as much as 5%–15%, based on this analysis. Ten per cent of the known terrestrial permafrost carbon pool is equivalent to ~130–160 Pg carbon. That amount, if released primarily in the form of CO_2 at a constant rate over a century, would make it similar in magnitude to other historically important biospheric sources, such as land-use change (0.9 ± 0.5 Pg carbon per year; 2003–2012 average), but far less than fossil-fuel emissions[88] (9.7 ± 0.5 Pg carbon per year in 2012). Considering CH_4 as a fraction of permafrost carbon release would increase the warming impact of these emissions. At these rates, the observed and projected emissions of CO_2 and CH_4 from thawing permafrost are unlikely to occur at a speed that could cause abrupt climate change over a period of a few years to a decade[1,9]. A large pulse release of permafrost carbon on this timescale could cause climate change that would incur catastrophic costs to society[8], but there is little evidence from either current observations or model projections to support such a large and rapid pulse. Instead, permafrost carbon emissions are likely to occur over

BOX 1

Permafrost carbon feedback to climate change

As shown in the Box 1 Figure, carbon stored frozen in permafrost, once thawed, can enter ecosystems that have either predominantly aerobic (oxygen present) or predominantly anaerobic (oxygen limited) soil conditions. Across the permafrost region, there is a gradient of water saturation that ranges from mostly aerobic upland ecosystems to mostly anaerobic lowland lakes and wetlands. In aerobic soils, CO_2 is released by microbial decomposition of soil organic carbon, whereas both CO_2 and CH_4 are released from anaerobic soils and sediments. Microbial breakdown of soil organic carbon can happen in the surface active layer, which thaws each summer and refreezes in the winter, and in the subsurface as newly thawed carbon becomes available for decomposition after it has emerged from the perennially frozen pool. The decomposability of soil organic carbon varies across the landscape depending in part on the plant inputs as well as the soil environment, and also with depth in the soil profile. The landscape mosaic of water saturation is also affected by permafrost thaw. Gradual and abrupt thaw processes such as top-down thawing of permafrost (increasing the thickness of the active layer) and lake draining can expose more carbon to aerobic conditions. Alternatively, abrupt thaw processes can create wetter anaerobic conditions as the ground surface subsides, attracting local water. Carbon can also be mobilized by erosion or by leaching from upland soils into aquatic systems or sediments. Plant carbon uptake can be stored in increased plant biomass or deposited in the surface soils, which in part can offset losses from soils.



Box 1 Figure | Key features regulating the permafrost carbon feedback to climate from new, synthesized observations.

decades and centuries as the permafrost region warms, making climate change happen even faster than we project on the basis of emissions from human activities alone. Because of momentum in the system and the continued warming and thawing of permafrost, permafrost carbon emissions are likely not only during this century but also beyond. Although never likely to overshadow emissions from fossil fuel, each additional ton of carbon released from the permafrost region to the atmosphere will probably incur additional costs to society.

Next steps for model–data integration

The Earth system models analysed for the IPCC AR5[1] did not include permafrost carbon emissions, and there is a need for the next assessment to make substantive progress analysing this climate feedback. It is clear, even among models that are currently capable of simulating permafrost carbon emissions, that improvements are needed to the simulations of the physical and biological processes that control the dynamics of permafrost distribution and soil thermal regime[43,44,57]. The initial model projections we review here are based on a range of different model formulations, many of which are known to lack key structural features. Critical next steps that are being achieved by the research community include a permafrost carbon model intercomparison using standard driving variables to improve model formulations and conceptualization. Initial intercomparison results point towards several key structural features that should be implemented by models attempting to forecast permafrost carbon emissions. These include explicitly defining the vertical distribution of carbon in permafrost soils to account for the way atmospheric warming at the surface propagates through the soil, causing permafrost thaw and carbon decomposition at depth. Additionally, many large-scale models do not distinguish CH_4 versus CO_2 release and project only total carbon emissions. This partitioning depends on explicitly describing the interactions between permafrost thaw and surface hydrology and is critical to produce credible projections of the effect of permafrost carbon on climate. A first-order issue is whether the terrestrial landscape in the permafrost region, already interspersed with thaw lakes, wetlands and waterlogged soils, becomes wetter or drier in a warmer world[89]. Lastly, new modelling formulations for describing abrupt thaw are being developed. These are needed to understand how gradual warming from the surface, occurring across the entire landscape as currently modelled, compares to hotspots on the landscape where permafrost undergoes catastrophic ground collapse and rapid thaw. These issues go beyond temperature sensitivity alone and are at the forefront of current ecosystem model development and research.

Models are useful tools for making projections, but need to use observations more effectively for benchmarking and parameterization. Current models show a wide range of results when compared against benchmark data sets of permafrost soil temperatures[44], soil carbon stocks[90], and high-latitude carbon fluxes[91], emphasizing the high uncertainty in these projections. Now, new data sets on decomposability (reviewed here) are available and should be used to parameterize key aspects of model carbon feedbacks. The databases on decomposability however, remain two orders of magnitude smaller than surface (<1 m) carbon pool data sets. Increasing the number of laboratory incubations will help to constrain uncertainty regarding the potential for permafrost carbon to remain stable under different environmental conditions and will allow researchers to understand which controls over decomposition are most important for the slow turnover pools that comprise a large fraction of the total permafrost carbon pool. At the same time, further work is required to quantify the permafrost carbon pool itself better. Despite substantial recent progress, remote regions such as the Canadian High Arctic, central Siberia, and the subsea continental shelves remain poorly represented, with very few data points deeper than 1 m. Other data sets synthesizing field observations of CH_4 emissions and CO_2 exchange provide process-level understanding available for model validation as well[40,91–93]. Model–data fusion using these newly created databases from both laboratory and field observations is urgently needed to evaluate which models can credibly represent the permafrost region and thus help reduce the uncertainty in forecasting the permafrost carbon feedback.

High-latitude warming and the emission of permafrost carbon remains a likely global carbon cycle feedback to climate change. The sheer size of these frozen carbon pools and the rapid changes observed in the permafrost region warrant focused attention on these remote landscapes. The observations and modelling steps outlined here will help in forecasting future change. At the same time, it is imperative to continue developing effective observation networks, including remote sensing capability[94], to adequately quantify real-time CO_2 and CH_4 emissions from permafrost regions[95]. While increased permafrost carbon emissions in a warming climate are more likely to be gradual and sustained rather than abrupt and massive, such observation networks are needed to detect the potential emissions predicted here, and also to provide early warning of phenomena and potential surprises we do not yet fully appreciate or understand. The combination of robust observations with appropriate modelling tools for forecasting change is essential to properly evaluate permafrost carbon sources. The quantification of carbon sources in addition to those that are a direct result of human activity is necessary when developing and evaluating climate change mitigation policies.

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Energy and trace-gas fluxes across a soil pH boundary in the Arctic

D. A. Walker*, N. A. Auerbach*, J. G. Bockheim†, F. S. Chapin III‡, W. Eugster‡, J. Y. King§, J. P. McFadden‡, G. J. Michaelson‖, F. E. Nelson¶, W. C. Oechel#, C. L. Ping‖, W. S. Reeburg§, S. Regli§, N. I. Shiklomanov¶ & G. L. Vourlitis#

* *Tundra Ecosystem Analysis and Mapping Laboratory, Institute of Arctic and Alpine Research, University of Colorado, Boulder, Colorado 80309-0450, USA*
† *Soils Department, University of Wisconsin, 1525 Observatory Drive, Madison, Wisconsin 53706, USA*
‡ *Department of Integrative Biology, University of California Berkeley, Berkeley, California 94720, USA*
§ *Department of Earth System Science, University of California Irvine, 205 Physical Sciences, Irvine, California 92697-3100, USA*
‖ *Agricultural and Forestry Exploratory Station, University of Alaska Fairbanks, 533 E. Fireweed, Palmer, Arkansas 99645, USA*
¶ *Department of Geography, University of Delaware, Newark, Delaware 19716, USA*
\# *Global Change Research Group, Department of Biology, San Diego State University, San Diego, California 92182, USA*

Studies and models of trace-gas flux in the Arctic consider temperature and moisture to be the dominant controls over land–atmosphere exchange[1,2], with little attention having been paid to the effects of different substrates. Likewise, current Arctic vegetation maps for models of vegetation change recognize one or two tundra types[3,4] and do not portray the extensive regions with different soils within the Arctic. Here we show that rapid changes to ecosystem processes (such as photosynthesis and respiration) that are related to changes in climate and land usage will be superimposed upon and modulated by differences in substrate pH. A sharp soil pH boundary along the northern front of the Arctic Foothills in Alaska separates non-acidic (pH > 6.5) ecosystems to the north from predominantly acidic (pH < 5.5) ecosystems to the south. Moist non-acidic tundra has greater heat flux, deeper summer thaw (active layer), is less of a carbon sink, and is a smaller source of methane than moist acidic tundra.

In 1995 and 1996, we studied the ecosystem properties on either side of a prominent pH boundary within the Kuparuk River basin (KRB) in Alaska, the primary study area of the Arctic System Science Land–Atmosphere–Ice Interactions (ARCSS–LAII) Flux Study[5] (Fig. 1). We characterized moist non-acidic tundra (MNT) and moist acidic tundra (MAT) ecosystems at two intensive study sites about 7 km apart on either side of the boundary (Fig. 1b, sites 3 and 4). We also collected soil and vegetation data from numerous other MNT and MAT sites within the KRB during an accuracy assessment of the landcover map in Fig. 1b (ref. 6). This adds to earlier information from Toolik Lake, Happy Valley and Prudhoe Bay, Alaska[7–11].

The vegetation and soil properties on either side of the boundary are similar to those described for MNT and MAT in other studies[8,12]. Site 3 has MNT with 36% cover of non-sorted circles[13]. The non-sorted circles are partly vegetated patches of highly frost-active soils that are about 1–2 m in diameter and spaced at intervals of 2–3 m; bare soil covers about 4% of site 3. The vegetation community between the circles is *Dryado integrifoliae-Caricetum bielowii*[8], which is dominated by non-tussock sedges (*Carex bigelowii*, *C. membranacea* and *Eriophorum triste*), prostrate shrubs (*Dryas integrifolia*, *Salix arctica*, *S. reticulata* and *Arctuous rubra*) and minerotrophic mosses (*Tomentypnum nitens*, *Hylocomium splendens* and *Ditrichum flexicaule*). Soils of MNT have a broken organic layer over a dark-coloured A horizon (a mineral horizon containing organic-matter accumulation) with high base saturation, over a gleyed C horizon (a subsoil mineral horizon relatively unaffected by soil-formation processes except for the presence of grey colours resulting from poor drainage and reduction of iron)[11,14]. All soil horizons have consistently high pH (>6.5) and are highly frost stirred (cryoturbated).

Site 4 is covered by tussock tundra (*Sphagno-Eriophoretum*[8]) with few (<1% cover) non-sorted circles. This vegetation type is dominated by dwarf shrubs (*Betula nana*, *Ledum palustre* ssp. *decumbens*, *Salix planifolia pulchra*), tussock sedges (*Eriophorum vaginatum*) and acidophilous mosses (*Sphagnum* spp., *Aulocomnium* spp., *Polytrichum* spp. and *Dicranum* spp.). Soils of MAT have a thick continuous organic horizon over gleyed subsoil material and contain cryoturbated organic material in the lower part. Both sites 3 and 4 are on silty loess deposits[11]. Soil pH of MAT sites tends to increase with depth from about 4.0 at the surface to 6.5 in the frozen C horizons.

The pH boundary extends at least 300 km to the east and west of the study area[15,16]. Loess blankets much of the Arctic Coastal Plain and Arctic Foothills, and both MAT and MNT occur on these extensive deposits, so it is difficult to explain the sharp vegetation boundary solely by differences in surface deposits[17]. The boundary may be partly due to a stronger winter Arctic climate north of the topographic barrier of the Arctic Foothills[18]. A colder, windier climate with shallower snowpack would promote the formation of non-sorted circles[12] and cause the continual stirring of non-acidic subsoils to the surface[11,14]. The abundance of non-sorted circles and relatively low shrub biomass (85 versus 202 g m^{-2}) north of the boundary results in the greyer tones on the false-colour infrared image (Fig. 1a). Lower shrub biomass, lower leaf-area index (LAI) and lower normalized difference vegetation index (NDVI) of MNT at site 3 is consistent with previous studies[9,19] (Table 1).

South of the boundary, MNT is found only in relatively small areas on limestone bedrock and in naturally disturbed systems, such as river floodplains, snowbeds, windy hill crests and recently glaciated areas. In most of the Arctic Foothills, vegetation succession and peat formation (paludification) during the Holocene have converted formerly dry vegetation on mineral-rich loess and till deposits to MAT. Paludification is enhanced toward the south as a result of increased temperature and precipitation. Mosses, particularly *Sphagnum*, are important to this conversion. It is abundant in MAT but not MNT, and has numerous unique properties that strongly promote waterlogging and cold acidic soils[20–23].

The vegetation and soil differences between MAT and MNT have important consequences for land–atmosphere exchanges. Site 3 (MNT) had 28% more soil heat flux during 10 days of observation and 54% deeper end-of-summer thaw than site 4 (MAT). Summer thaws of MNT are consistently deeper than those of MAT throughout the KRB, despite MNT being dominant in the northern, colder portion of the study area[24], because the MNT has shorter, more open plant canopies (less shading by vascular-plant leaf area), less continuous moss cover and thinner organic horizons (Table 1). In a related study, evapotranspiration, soil heat flux and sensible heat flux (heat exchange between the atmosphere and the land surface) showed a similar relationship with net radiation at two acidic tundra sites (sites 4 and 6; Fig. 1b) despite latitudinal and elevation differences in climate, indicating that the energy budgets are more strongly correlated with vegetation type than with climate[25].

Site 4 also had about twice the gross photosynthesis and three times the respiration of the MNT site, as well as a greater net carbon gain, during the same 10-day measurement period (Table 1), despite the close proximity of the two sites and nearly identical temperature, net radiation and relative humidity[25]. These results are consistent with CO_2-flux data from two other sites (11 and 21; Fig. 1b) during the same period in 1995. Site 11 (MAT) had similar summer climate and CO_2 flux to that at site 4, whereas site 21 (MNT) had a lower flux than site 3, probably owing to the colder early summer climate near the coast[26]. Integrated fluxes from sites 11 and 21

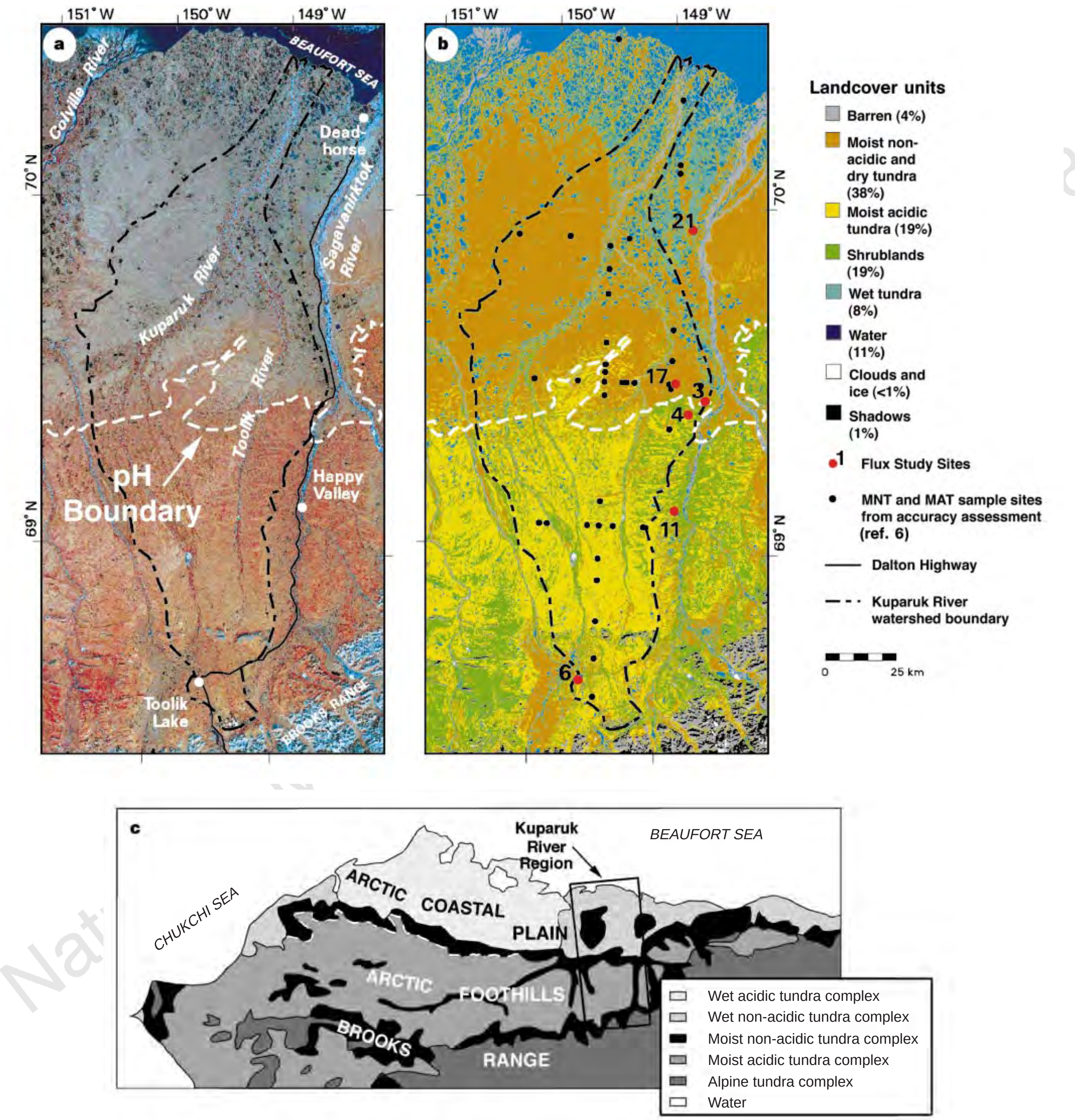

Figure 1 Maps of the study area. **a**, Landsat MSS false-colour infrared mosaic of the Kuparuk River Basin (dashed line), northern Alaska. A broad band of foothills occurs between the mountains and the Arctic Coastal Plain with many lakes. The pH boundary separating the redder tones to the south from the greyer tones to the north corresponds to soils that are acidic and non-acidic, respectively. Landsat data are courtesy of the US Geological Survey Alaska Data Center. **b**, Land-cover map. Red circles indicate intensive study sites. Black dots are other sites visited during the accuracy assessment of the map[6]. **c**, The map shows a generalized distribution of acidic and non-acidic vegetation types in northern Alaska, prepared from an integration of information from several sources including AVHRR (Advanced Very-High Resolution) satellite images, soil maps, vegetation maps and surface-geology maps. The location of the pH boundary west of the Colville River (white dashed line) is less distinct and unstudied.

throughout the summer of 1995 showed that the MAT site was a much greater carbon sink than the MNT site (55.2 versus 27.6 g C m^{-2} per season). In 1996, an even larger difference was observed between site 11 and site 17, an MNT site close to site 3 that has a summer climate very similar to sites 3 and 11. Site 11 (MAT) gained 52.5 g C m^{-2} per season compared with 3.3 g C m^{-2} per season at site 17 (MNT) (Table 1). Taken together, our data demonstrate a consistent spatial and temporal pattern of a much larger carbon sink in MAT than in MNT. Methane flux showed a pattern opposite to that of CO_2, with the wetter, more anaerobic soils of MAT effluxing over six times the methane of MNT (Table 1).

Greater carbon accumulation in the vegetation has also led to twice as much organic carbon in both the active layer and the permafrost in the soils at site 4 than site 3 (Table 1). The basal ^{14}C date from the frozen C horizon at site 4 was 8,500 years BP compared to 12,500 years BP in the same horizon at site 3, demonstrating the

Table 1 Comparison of ecosystem properties of MNT and MAT at sites in the Kuparuk River Basin

	Sites 3 and 4			Other sites			
Ecosystem property	MNT	MAT	Significance	MNT	MAT	Significance	Reference
Soil							
pH of top mineral horizon	7.6 [1]	5.5 [1]	n.a.	7.0 ± 0.16 [20]	5.3 ± 0.13 [10]	***	Ref. 14
				6.3 ± 0.1 [14]	4.6 ± 0.1 [33]		Ref. 8†
O-horizon thickness (cm)	9 ± 1 [71]	15 ± 1 [71]	**	11 ± 1.9 [21]	21 ± 1.8 [15]	***	This study‡
Soil moisture of top mineral horizon ($cm^3 cm^{-3}$, Jul 95)	0.37 [1]	0.40 [1]	n.a.				This study
Bare soil (% cover)	4.4 ± 1.6 [6]	0.2 ± 0.0 [6]	***	8 ± 1 [140]	1 ± 0.2 [121]	***	This study§
Vegetation							
Height of plant canopy (cm)	3.9 ± 0.3 [340]	6.5 ± 0.4 [340]	***				This study
Leaf area index	0.50 ± 0.03 [66]	0.84 ± 0.05 [66]	***	0.57 ± 0.06 [7]	0.81 ± 0.08 [11]	**	Ref. 19
NDVI (MSS)	0.23 [1]	0.32 [1]	n.a.	0.28 ± 0.00 [4×10^6]	0.41 ± 0.00 [2×10^6]	n.a.	This study‖
NDVI (Hand-held)				0.62 ± 0.02 [7]	0.71 ± 0.01 [11]	**	Ref. 19
Moss cover (%)	65 ± 4 [12]	79 ± 4 [12]	**				This study
Above ground biomass ($g m^{-2}$)							
Shrubs	85 ± 18 [10]	202 ± 22 [10]	***	127 ± 19 [7]	270 ± 19 [11]	***	Ref. 19
Graminoids	124 ± 12 [10]	112 ± 15 [10]	n.s.	118 ± 22 [7]	118 ± 24 [11]	n.s.	Ref. 19
Forbs	40 ± 22 [10]	10 ± 2 [10]	***	12 ± 4 [7]	12 ± 2 [11]	n.s.	Ref. 19
Mosses, lichens, litter	504	460	n.a.	221 ± 85 [7]	207 ± 33 [11]	n.s.	Ref. 19
Total	753 ± 60 [10]	784 ± 139 [10]	n.s.	447 ± 23 [7]	607 ± 27 [11]	***	Ref. 19
Energy and trace-gas flux							
Soil heat flux (19–30 Jun 1995, $MJ m^{-2} d^{-1}$)	1.39 ± 0.21 [331]	1.09 ± 0.16 [275]	***				This study
Thaw depth (cm)	57 ± 1 [71]	37 ± 1 [71]	***	52 ± 2 [20]	39 ± 2 [14]	***	This study§
				57 ± 5 [14]	36 ± 3 [33]	***	Ref. 8
Evapotranspiration (19–30 Jun 1995, $mm d^{-1}$)	1.16 ± 0.17 [331]	1.06 ± 0.16 [275]	n.a.				This study
10-d gross primary production (19–30 Jun 1995 $g CO_2$-$C m^{-2} d^{-1}$)	0.94 ± 0.14 [331]	1.82 ± 0.27 [275]	n.a.				This study
10-d net CO_2 uptake ($g CO_2$-$C m^{-2} d^{-1}$)	0.67 ± 0.10 [331]	0.95 ± 0.27 [275]	n.a.	0.27 ± 0.41 [12]	1.02 ± 0.33 [12]	n.a.	This study¶
10-d respiration loss ($g CO_2$-$C m^{-2} d^{-1}$)	0.27 ± 0.04 [331]	0.87 ± 0.13 [275]	n.a.				This study
1995 net CO_2 uptake ($g CO_2$-$C m^{-2}$ per season)				27.6 [77]	55.2 [90]	n.a.	This study#
1996 net CO_2 uptake ($g CO_2$-$C m^{-2}$ per season)				3.3 [31]	52.5 [73]	n.a.	This study☆
Methane emission ($mg CH_4 cm^{-2} yr^{-1}$)				69 ± 33 [12]	449 ± 301 [15]	*	This study††
Soil organic carbon ($kg C m^{-3}$)	40 [1]	88 [1]	n.a.	56 ± 5 [5]	44 ± 11 [6]	n.s.	Ref. 11
				55 ± 5 [16]	49 ± 4 [7]	n.s.	Ref. 10

Standard error of the mean and number of samples [in brackets] are given for most variables. Probability of significance in all cases was based on two-sample *t*-test. Significance levels: $*P \leq 0.1$; $**P \leq 0.05$; $***P \leq 0.01$; n.s., non-significant; n.a., non-applicable.
† Data are from 47 permanent plots in the Toolik Lake region.
‡ Measurements at 36 random points within the Kuparuk River basin during accuracy assessment of the land-cover map.
§ Estimates obtained from aerial surveys at 361 sites within the Kuparuk River basin during accuracy assessment of the land-cover map.
‖ Mean MSS NDVI values for the land-cover map.
¶ Sites 11 (MAT) and 24 (MNT).
# Sites 11 (MAT) and 24 (MNT). Time intervals for measurements: site 11, 1 Jun–31 Aug 1995; site 24, 16 Jun–31 Aug 1995.
☆ Sites 11 (MAT) and 17 (MNT). Time intervals for measurements: site 11, 6 Jun–31 Aug 1996; site 17, 15 Jun–19 Aug 1996.
†† Methane measurements at 37 plots at Toolik Lake region, Happy Valley and Deadhorse.

faster accumulation rate at the acidic site. Other soil data from MAT and MNT sites throughout the basin do not show a similar consistent trend of more carbon in the MAT soils (Table 1), presumably because there is a wide diversity of genetic environments, including MNT fen and fluvial sites, and MAT sites on a variety of surface ages. Other studies, however, show that MNT soils have consistently lower C:N ratios (15 versus 20)[11], greater microbial activity and more highly decomposed organic fraction[27]. The relative winter CO_2 flux rates of MNT and MAT remain unresolved[28].

Extrapolations of trace-gas fluxes and soil carbon based solely on numbers from the more extensively studied MAT, as has been done in all previous high-latitude extrapolations, results in large errors. For example, in the map area of Fig. 1, this would overestimate gross photosynthesis by at least 35%, respiration by 140%, net CO_2 uptake by at least 15%, and methane flux by 140%. Similar, but more diffuse, pH boundaries separate worldwide zonal tundra types. MNT corresponds to the sedge-dominated 'typical tundra' of Russian authors, whereas MAT corresponds to shrubby 'southern tundra'[12]. Over century to millennium time scales, we expect that zonal soil pH boundaries will shift northwards in response to climate warming, deeper winter snowpacks and reduction of loess sources. Regions with declining soil pH will show a decrease in soil heat flux and large increases in methane flux and carbon storage in the plant canopy. □

Methods

Site selection. In the summers of 1995 and 1996, several sites were monitored to characterize the energy and trace-gas fluxes of arctic tundra[5]. Sites 3, 17 and 24 have MNT and sites 4, 6 and 11 have MAT vegetation. To compare MNT and MAT fluxes under nearly identical summer climates, we chose two sites about 7 km apart (sites 3 and 4; Fig. 1b) on opposite sides of the pH boundary on hilltops (224 and 332 m, respectively) with similar topography. The sites were accessible by helicopter from Happy Valley and were as similar and homogeneous as possible, and were far enough from the Dalton Highway to eliminate the effects of road dust.

Soils. Percentages of soil types and O-horizon thickness at sites 2 and 4 were determined from 71 random points at each site. Soil classification is according to the Gelisol Order in US soil taxonomy[29]. The pH values at all sites are from surface samples collected at 39 random MNT and 24 MAT points within the KRB basin. The mean O-horizon thickness for the same sites was determined from 10 samples at each site. Bare-soil values for the KRB are visual estimates from 261 sites visited during accuracy assessment of the land-cover map[6]. Soil organic carbon was analysed on acid-treated samples using a Leco CHN-1000 analyser[11].

Vegetation. The land-cover map of the KRB was derived from a mosaic of Landsat Multispectral Scanner (MSS) images provided by the USGS EROS Alaska Data Center[6]. The height of the plant canopy was determined from 340 random points each at sites 3 and 4. Percentage cover of moss and bare soil was determined from four 50-m and two 70-m line transects at each site. Vascular plant leaf-area index was measured with a LI-COR PCA 2000 plant canopy analyser[19] at 66 points at each site. Biomass is the mean dry mass of 10 random 20 × 50-cm clip harvest plots within sites 3 and 4. The normalized difference vegetation index (NDVI) at sites 3 and 4 was determined from the pixel of the Landsat MSS image centred on the sample sites. Mean NDVI for MNT and MAT within the KRB was calculated from the total set of pixels in each class in Fig. 1.

Flux measurements. Short-term flux measurements were made simultaneously at sites 3 and 4 from 19 Jun to 29 Jun 1995, using four heat-flux plates and four temperature probes. Evapotranspiration and CO_2 flux were measured using the eddy-covariance method with an Applied Technologies sonic anemometer and LI-COR 6262 infrared gas analyser mounted on 2-m towers[25]. The mean and standard error for energy flux, gross primary production and evapotranspiration at sites 3 and 4 were calculated on the basis of 30-min averages. CO_2 fluxes at sites 11, 17 and 21 were determined using eddy-covariance methods and 2.5-m towers[26]. Mean values and standard errors at these sites were calculated using the daily mean CO_2 fluxes. The daily methane fluxes were integrated over the thaw period to obtain annual emission. Winter methane fluxes were assumed to be zero. CH_4 flux was measured during the thaw season, Jun–Aug, at 27 MNT and MAT sites along the Dalton Highway in 1996 using a static chamber method[30]. Air samples were taken over periods of 30–45 min and were analysed on a gas chromatograph equipped with a flame ionization detector.

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Correspondence and requests for materials should be addressed to D.A.W. (e-mail: swalker@taimyr.colorado.edu).

Complementarity and the use of indicator groups for reserve selection in Uganda

Peter C. Howard*†, Paolo Viskanic*†, Tim R. B. Davenport*†, Fred W. Kigenyi*, Michael Baltzer*†, Chris J. Dickinson*†, Jeremiah S. Lwanga*†, Roger A. Matthews*† & Andrew Balmford‡

* *Forest Department, PO Box 1752, Kampala, Uganda*
‡ *Department of Animal and Plant Sciences, The University of Sheffield, Sheffield S10 2TN, UK*

A major obstacle to conserving tropical biodiversity is the lack of information as to where efforts should be concentrated. One potential solution is to focus on readily assessed indicator groups, whose distribution predicts the overall importance of the biodiversity of candidate areas[1,2]. Here we test this idea, using the most extensive data set on patterns of diversity assembled so far for any part of the tropics. As in studies of temperate regions[2–8], we found little spatial congruence in the species richness of woody plants, large moths, butterflies, birds and small mammals across 50 Ugandan forests. Despite this lack of congruence, sets of priority forests selected using data on single taxa only often captured species richness in other groups with the same efficiency as using information on all taxa at once. This is because efficient conservation networks incorporate not only species-rich sites, but also those whose biotas best complement those of other areas[9–11]. In Uganda, different taxa exhibit similar biogeography, so priority forests for one taxon collectively represent the important forest types for other taxa as well. Our results highlight the need, when evaluating potential indicators for reserve selection, to consider cross-taxon congruence in complementarity as well as species richness.

By containing elements of both East African savannas and Central African rain forests, Uganda boasts more species for its size than almost any other country in Africa[12]. Much of this diversity is restricted to 15,000 km^2 of forest reserves (which also contain non-forest habitats) under the jurisdiction of the Uganda Forest Department[13]. The aim of a five-year inventory of the woody plants, large moths (saturnids and sphingids), butterflies, birds, and small mammals (rodents and insectivores) of all of the principal forest reserves was to provide information to the government regarding a plan to protect ~3,000 km^2 (20%) of the remaining forest estate as a strict nature reserve[14,15]. Forests were surveyed in proportion to their area (see Methods). In total, nearly 100 man-years of survey effort yielded records of 2,452 species.

Constraints on funding and expertise mean that surveys of this magnitude will rarely be undertaken elsewhere in the tropics. However, the size and taxonomic breadth of the Uganda data set mean that it provides an exceptional opportunity to test ways in which future priority-setting exercises could be conducted more quickly and at lower cost. Here we focus on one widely proposed short cut to establishing priorities for biodiversity conservation, and determine whether survey data on just one or two putative indicator groups can identify robust reserve networks capable of conserving biodiversity as a whole[1,2].

† Present addresses: Wildlife Department, PO Box M239, Accra, Ghana (P.C.H.); Food and Agriculture Organisation, PO Box 521, Kampala, Uganda (P.V.); World Wide Fund for Nature-Cameroon, PO Box 6776, Yaounde, Cameroon (T.R.B.D.); Groundtruth International Ltd, 4 Duck Lane, Oundle, Northants PE8 4DY, UK (M.B.); Wildlife Conservation Division, Royal Forest Department, Phaholoythin Road, Chatuchak, Bangkok 10900, Thailand (C.J.D.); Makerere University Biological Field Station, PO Box 409, Fort Portal, Uganda (J.S.L.); and Countryside Council for Wales, Park Street, Newtown, Powys SY16 1RD, UK (R.A.M.).

VEGETATION AND FLORISTICS OF PINGOS, CENTRAL ARCTIC COASTAL PLAIN, ALASKA

by

MARILYN DREW WALKER

B.A. University of Colorado, 1978

M.A. University of Colorado, 1980

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CHAPTER I

INTRODUCTION

> We had paddled and floated most of the day, which at this latitude was 24 hours long, but the wind from the north had made our progress very slow. Our arms and backs ached from paddling and poling the heavily loaded canvas folboats, which frequently had to be pulled off shoals and gravel bars.... During the afternoon as I surveyed the landscape to the north from the river bank, I noticed a dome-like structure on the horizon but couldn't really believe there was anything that high out there.... It was nearly 2 A.M. and the tundra was still brilliantly lit by the arctic sun when we rounded a bank of the meandering Toolik River, and picked out our camp site on the terrace bank six feet above the river. I jumped onto the bank and there not more than 600 yards away was a large conical mound that looked to me like one of the pyramids of Egypt. It rose at least 100 feet above the tundra and had a base of 1000 feet, which made it almost like a mountain on the flat, prairie-like landscape. It was a pingo. [Koranda 1970, pg. 18]

The paragraph quoted above, describing J.J. Koranda's first encounter with a pingo, demonstrates the intrigue with which people first view pingos, and illustrates the personal fascination that led to this dissertation. The word 'pingo' was borrowed from the Inuit language by Porsild (1938) to represent a specific type of ice-cored mound found in arctic regions. Early explorers along the Alaskan and Canadian coasts noted and described these features (Richardson 1851; Schrader 1904; Leffingwell 1919).

Pingos are described by Mackay (1979) as "...ice-cored hills which are typically conical in shape and can grow and persist only in permafrost." They may reach as high as 50 m and obtain a diameter over 1 km (Embleton and King 1968; D.A. Walker et al. 1985), although in the area studied most are less than 10 m high. Pingos are genetically separated from other types of mounds by the presence of an ice core, which can range from ice-rich sediment to nearly pure ice (Mackay 1979; Pissart 1983).

Objectives

The primary objective of this dissertation is to characterize the vegetation and associated soils of the pingos of a region within the central Arctic Coastal Plain of Alaska (Fig. 1). This is approached with five goals: 1) classifying the vegetation in a form that can be related to other vegetation assemblages, 2) relating environmental gradients to the pingo vegetation, 3) determining the floristic affinities of the pingo flora and comparing this with the regional flora, 4) determining if the number of species on pingos represents an equilibrium constrained by pingo size (area), and 5) characterizing the successional sequences on these pingos.

There are several reasons why the pingos are of ecological interest. Koranda (1970) appears to be the first to mention that pingos would be excellent sites for plant ecological studies. He noted that pingo vegetation is particularly diverse due to the variety of distinct habitats, which are related to differences in slope, aspect, effects of wind, disturbance by animals, and deposition of snow. Perring (1959, p. 447) envisioned the ideal situation in which to study soils and vegetation of any biome as:

> ...Two or more **isolated hemispherical hills** .... made of the same parent materials similarly oriented, undisturbed by burning or ploughing and grazed at the same intensity ...: isolated so that no disturbing local climate would upset the picture, and with strata horizontal so that drainage would be similar on all aspects. If soil development had commenced about the same time ...the soil and vegetation should be comparable in both areas. In practice hemispherical hills are not very frequent and in the field it is necessary to use samples from scattered sites in an attempt to synthesize the ideal. [emphasis my own]

The northern Alaska pingos represent a condition close to this ideal, with similar parent material, topography, and climate. They are essentially small ecosystems. There are few studies that directly examine the ecological effects of slope and aspect in the Arctic (Ritchie 1984), as relatively few situations exist where this can be easily

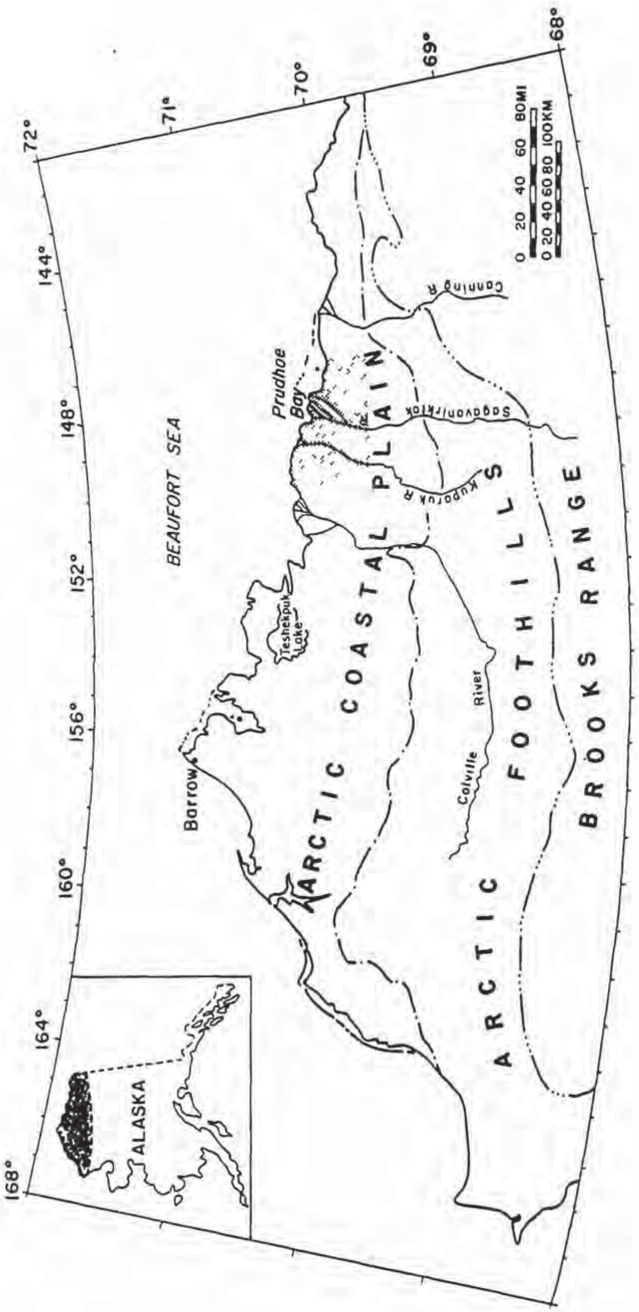

Figure 1. Location of the study on the Alaskan Arctic Coastal Plain. The shaded area represents the approximate boundary of the study region. Physiographic provinces are according to Wahrhaftig (1965).

considered, although directional patterns have been reported, for example, by Polunin (1948) and Webber (1971).

A compelling reason to study the ecology of these particular pingos at this time is that most have been impacted to some degree by the oil development in the Prudhoe Bay and Kuparuk regions that is now expanding across much of northeastern Alaska. Because the area is primarily a wetland, the management focus has been primarily on maintaining the wetland integrity and functional values, with essentially no attention to the dry environments of pingos. At least one pingo has been destroyed completely by construction activities (D.A. Walker et al. 1986), and others have had their entire vegetation cover removed. Most within the oil field are littered with surveyors' trash and have obvious vehicle tracks on at least one slope. Several pingos are regularly used as sites for radio towers (Fig. 2).

Questions and Predictions

The central question of this study was, "Are there unique elements of the pingo vegetation, and if so, what are they, and why do they exist?" Pingo vegetation has never been fully described. Koranda (1970), Koranda and Evans (1975), D.A. Walker et al. (1985), and Everett (1983a,b) have stated that pingos are unique features in this landscape. The work of Mackay (1979) in the Tuktoyaktuk Peninsula, N.W.T., Canada, has demonstrated the opportunities for understanding periglacial geomorphic processes through the study of pingos. To date, however, no one has illustrated just what it is about the vegetation and ecology of pingos that makes them so interesting. This general question is too broad to be approached with a single prediction, so a series of three secondary questions have also been asked: 1) "Are there rare species or communities present on the pingos, and if so, what are their elements?" 2) "Do the pingos function as biogeographic islands?" and 3) "Is there a

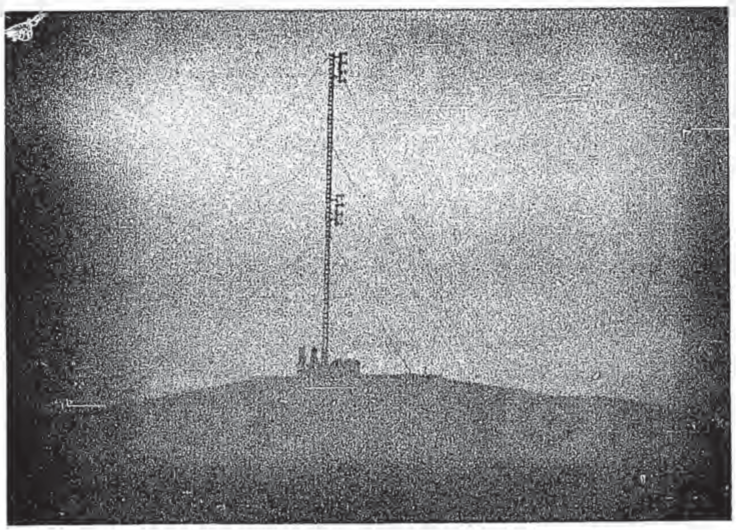

Figure 2. Pingo no. 5 (Prudhoe Mound) with radio tower on top. This pingo is within the Prudhoe Bay oil field, and most pingos within the oil field have been affected by the development to some degree.

successional vegetation sequence on the pingos that is related to their age and geomorphic development?"

Question 1: Rare Elements

During the last full glacial[1], the Alaskan vegetation was much different than today, and large expanses of grass-dominated tundra ('steppe-tundra') may have been present (Hopkins et al. 1982). Many authors have hypothesized that if there were such environments then they should be present today in areas with appropriate habitat, i.e., in isolated refugia. The Brooks Range, river bluffs in interior Alaska, and scattered sites in Siberia have been demonstrated to have steppe-like plant assemblages that may represent relicts of this type (Yurtsev 1982; Murray et al. 1983; Cooper in prep.). Some of the pingos may be very old stable sites that were present during this full glacial period (Rawlinson 1984a; D.A. Walker et al. 1985). If so, they may support relicts of this steppe vegetation. The pingos are essentially the only well-drained sites on the coastal plain. It has been presumed that steppe elements are missing from this area due to a lack of habitat, but the dry pingo slopes could potentially support steppe assemblages. Rarity here is defined as not being present regionally, except at these sites, although the species or community may be abundant elsewhere.

Question 2: Pingos as Islands

Both Koranda (1970) and D.A. Walker (1985a) stated that the Arctic Coastal Plain pingos are island-like. Walker and Acevedo's (1987) Landsat classification of the Beechey Point quadrangle shows that 53% of the terrestrial portion of the map is either standing water or wet tundra, while only 39% is moist or

[1]Hopkins' (1982) Duvanny Yar, approximately 12 to 30 ka; there were no glaciers within the region of this study at any time during the Pleistocene.

dry tundra. They were not able to separate moist and dry tundra, but the dry area is estimated to be a very small percentage of this 39%. At Prudhoe Bay, which includes some of the wetter portions of the Beechey Point Quadrangle, less than 1% is dry tundra (D.A. Walker 1985a). The pingos are conspicuous, dry, high points within this landscape, and between the pingos there are only scattered areas that are also dry and could serve as source areas for dispersal of plants to the pingos. Dry sites other than pingos are river bluffs and the tops of high-centered polygons. Thus, the pingos are certainly 'island-like'. They are isolated dry areas surrounded by much wetter habitat, much of it standing or open water, and in this sense they are one of the better terrestrial analogs of oceanic islands. It is hypothesized that because pingos are isolated dry sites they will have species-area relationships similar to islands.

Question 3: Successional Sequence

This question is approached with three specific hypotheses. The first is that because the pingos are in equivalent substrates, the environmental gradients on different pingos will be the same, and there will therefore be a change over time toward a characteristic pingo flora. If this is true, then the floras of oldest pingos will be more similar than floras of younger pingos. This would not predict that a steady state had necessarily been reached, but that within the given time scale there is a point at which the vegetation remains stable for some long period of time.

The second hypothesis is that species composition on the youngest sites is less dependent on site factors than on the oldest sites, because initial community composition is a function of chance events. Margalef (1963, 1968) proposed this idea, and Christensen and Peet (1984) tested and supported the hypothesis in a deciduous forest. It remains to be tested in an arctic environment.

Finally, it is hypothesized that because at this latitude there are more species at their northernmost limit than at their southernmost limit, diversity will

develop more slowly on the cold sites (north slopes and ENE sides) than on the warm sites (south slopes and summits) and will maintain a lower level on the cold sites. Forty-six percent of the Prudhoe Bay flora reaches its northernmost limit in this region (D.A. Walker 1985a). The cold sites, therefore, have a more depauperate flora, and there are fewer species capable of colonizing these sites. Auclair and Goff (1971) found different patterns of diversity in a temperate forest successional sequence on mesic and xeric sites. They stated that in intermediate portions of environmental gradients competition would be most intense, and therefore diversity should be lessened. They also stated that in high-stress environments the dominant 'climax' species may also be the pioneers. This is very similar to Svoboda and Henry's (1987) model for the high arctic. Reiners et al. (1970) found an initial rapid rise in diversity at Glacier Bay, followed by a leveling off, with a maximum stand age of 1500 years.

CHAPTER II

BACKGROUND

Pingos

Mechanism of Formation

Porsild (1938) was the first to propose the theory that the pingos of northern Alaska and northwestern Canada:

> ...were formed by local upheaval due to expansion following the progressive downward freezing of a body or lens of water of semi-fluid mud or silt enclosed between bedrock and the frozen surface soil, much in the way in which the cork of a bottle is pushed up by the expansion of the water when freezing.

Porsild's theory was essentially correct, although it is now thought that it is expulsion rather than expansion that causes the upheaval (Mackay 1979). Another type of pingo has also been recognized that forms under pressure due to an upslope water source. These two types of pingos have been called "closed-system" and "open-system" respectively, but Mackay (1979) suggested that the terms hydrostatic and hydraulic pingos are more appropriate. Certain pingos that would be classified as the closed-system type might not be completely closed, as they could be connected to an open talik (an unfrozen water-saturated zone in permafrost). The two types are not exclusive, and gradations between them occur when local conditions such that both types of water sources are available. The pingos in this study are all presumed to be the hydrostatic type, because there is no obvious water source necessary for the formation of hydraulic pingos.

Hydrostatic pingos generally form in drained thaw-lake basins following drainage of lakes greater than 2 m deep (Fig. 3). Lakes of this depth do not freeze to the bottom in winter, which creates a deep talik beneath them. Following drainage, permafrost aggrades into the talik from all sides. As freezing progresses, water is expelled from the pore spaces of sandy or gravelly materials into the remaining unfrozen area, where pressure builds, eventually becoming great enough to push up the overlying sediments and form a pingo (Mackay 1979; Everett 1980a). Any talik that refreezes could potentially form a pingo, as under an abandoned stream channel.

This expulsion of water into a mass in the center of the pingo gives rise to injection ice, which results whenever a mass of injected water freezes within sediments (Pissart 1983). Mackay (1979) described three types of ice that are possible in pingos: pore ice, segregated ice, and intrusion ice. The conditions necessary for formation of these various ice types are a result of the pressure difference between the ice and water phases and also the type of sediments present at the ice-water interface. Some authors have emphasized ice origin as a critical factor in the classification of pingos and in separating pingos from other types of mounds, such as palsas (Pissart 1983), but Mackay (1979) stressed that it is the mound form that identifies a pingo as such and not the origin of ice within. He emphasized that the core may range from icy sediment to pure ice, and that rather than trying to classify ice-cored mounds into discrete groups, a continuum exists going from flat ground with massive sill ice to pingos with intrusion ice to pingos with segregated ice to flat areas with icy sediment. The pingos of northern Alaska and northwestern Canada are limited to regions of sand or gravel (Shumskii 1959; Mackay 1962, 1966, 1968, 1973, 1978, 1979; Carter and Galloway 1979; Müller 1962; D.A. Walker et al. 1985), but in other parts of the world pingos have grown in bedrock (e.g. Balkwill et al. 1974).

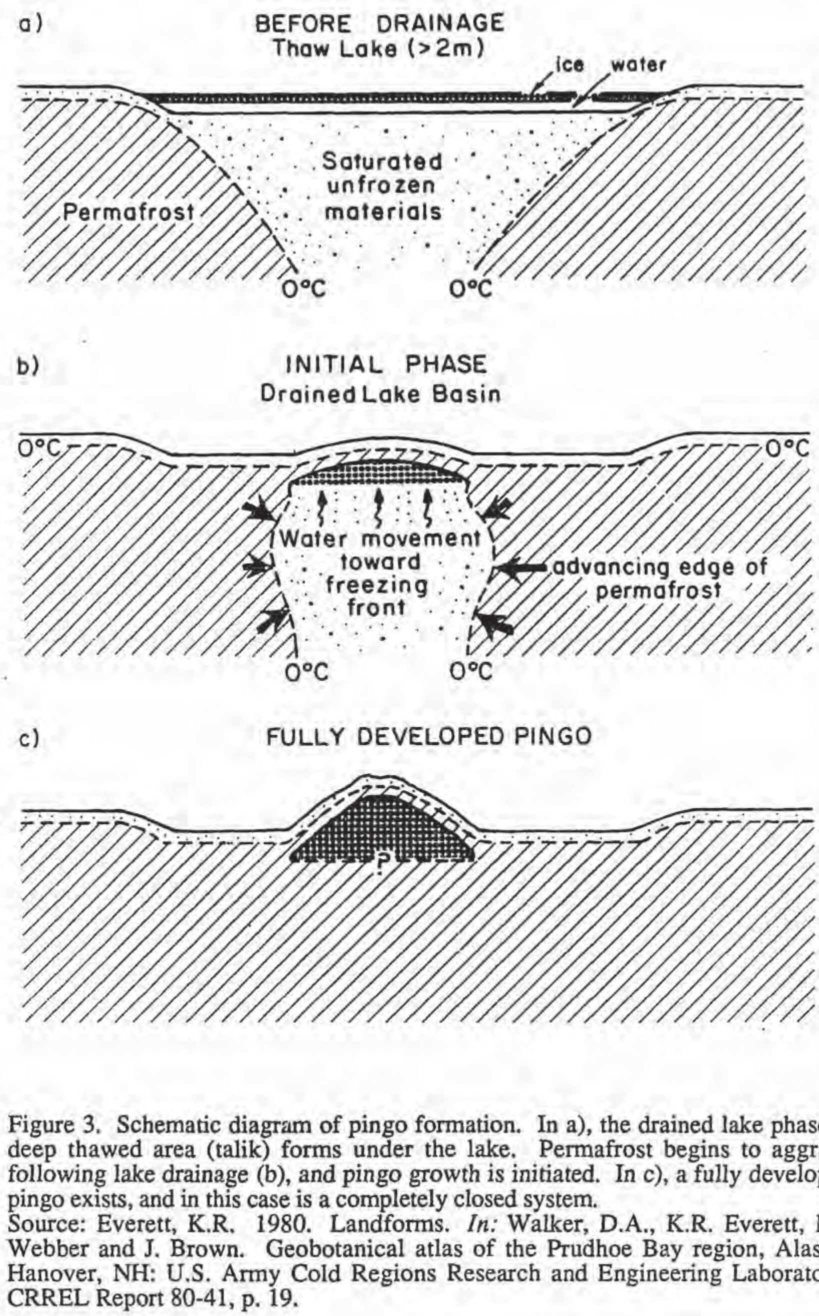

Figure 3. Schematic diagram of pingo formation. In a), the drained lake phase, a deep thawed area (talik) forms under the lake. Permafrost begins to aggrade following lake drainage (b), and pingo growth is initiated. In c), a fully developed pingo exists, and in this case is a completely closed system.
Source: Everett, K.R. 1980. Landforms. *In:* Walker, D.A., K.R. Everett, P.J. Webber and J. Brown. Geobotanical atlas of the Prudhoe Bay region, Alaska. Hanover, NH: U.S. Army Cold Regions Research and Engineering Laboratory, CRREL Report 80-41, p. 19.

Distribution

Pingos have been described from areas of continuous or discontinuous permafrost in Canada and Greenland (Fraser 1956; Pihlainen et al. 1956; Stager 1956; Craig 1959; Müller 1959; Robitaille 1961; Mackay 1962, 1963a,b, 1966, 1972, 1977a,b, 1979, 1981, 1983; Cruickshank and Colhoun 1965; Pissart and French 1976, 1977; Tarnocai and Netterville 1976; Vernon and Hughes 1976; Hughes 1969; Hughes et al. 1972; Brown and Péwé 1973; Balkwill et al. 1974; French 1975, 1976; Péwé 1975; French and Dutkiewicz 1976; Bennike 1983). They are also known from Siberia (Bobov 1960; Evseev 1976; Yurtsev 1982), Spitsbergen (Åhman 1973; Svensson 1976), Mongolia (Rotnicki and Babinski 1977; Kowalkowski 1978), and the Tibetan Plateau (K.T. Cheng, cited in Mackay 1979, pg. 6). Most recently, pingos have been described from Antarctica (Pickard 1983).

Pingos occur in most areas of the Alaskan Coastal Plain where thaw lakes and sandy or gravelly sediments are present (Porsild 1938; Burns 1964; Koranda 1970; Carter and Galloway 1979; Ferrians 1983; Rawlinson 1984a; D.A. Walker et al. 1985). Carter and Galloway (1979) mapped the pingos of the National Petroleum Reserve - Alaska (NPR-A). Hydraulic pingos are found throughout central Alaska in the zone of discontinuous permafrost, and in valleys of the northern Brooks Range (Holmes et al. 1966, 1968; Hamilton and Obi 1982).

The distribution of pingos in the Prudhoe Bay and Kuparuk oil fields was described and mapped by D.A. Walker et al. (1985; Fig. 4). at a scale of 1:63,360, and they described two distinct geomorphic types of pingos in this area. One is a steep-sided type with small basal diameter, steep slopes, and found in drained lake basins (Fig. 5). The other type has gentle slopes and broad base (Fig. 6). The terms 'steep-sided' and 'broad-based' when used in this dissertation refer to these types as defined by D.A. Walker et al. (1985).

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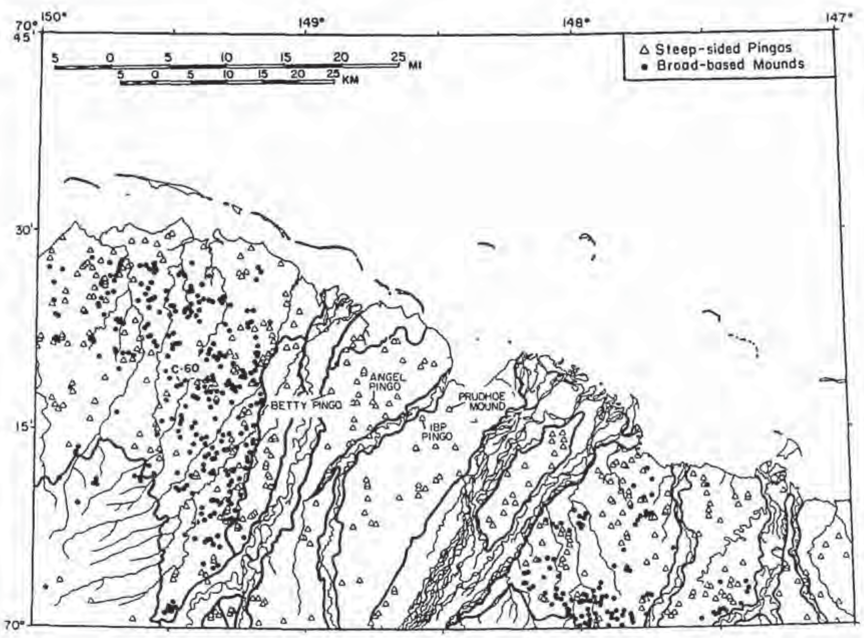

Figure 4. Distribution of pingos on the USGS 1:250,000 Beechey Point Quadrangle. Source: D.A. Walker, M.D. Walker, K.R. Everett and P.J. Webber. 1985. Pingos of the Prudhoe Bay region, Alaska. *Arctic and Alpine Research*, 17:333.

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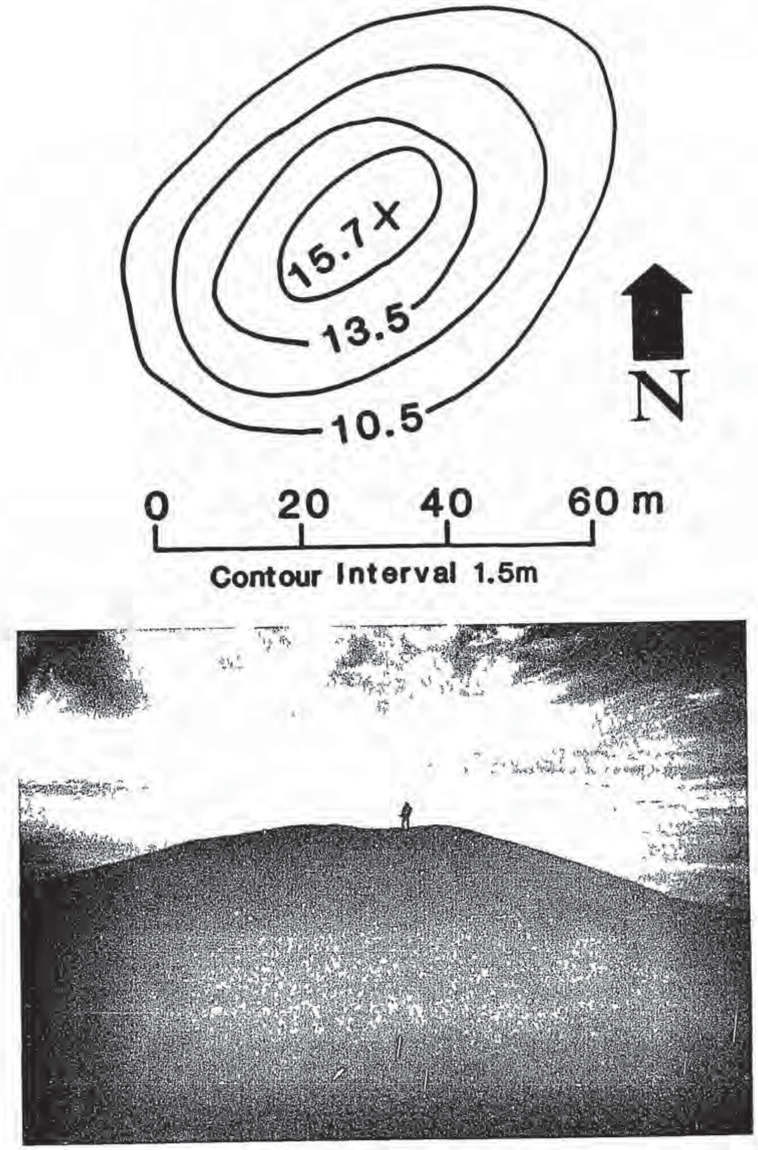
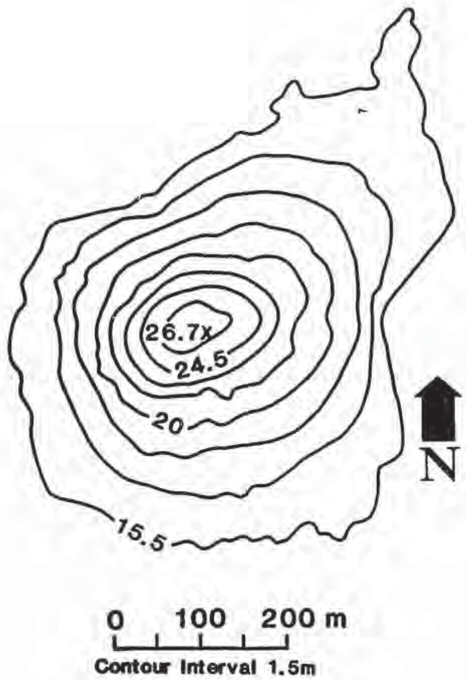


Fig. 5. A typical 'steep-sided' pingo (Pingo 1, Flower) at Prudhoe Bay, in plan view (a) and facing north looking at the south slope of the pingo (b). This pingo is 7 m high and 85 m in diameter. Slopes are as steep as 30°. Topographic data adapted from Air Photo Tech (1979).



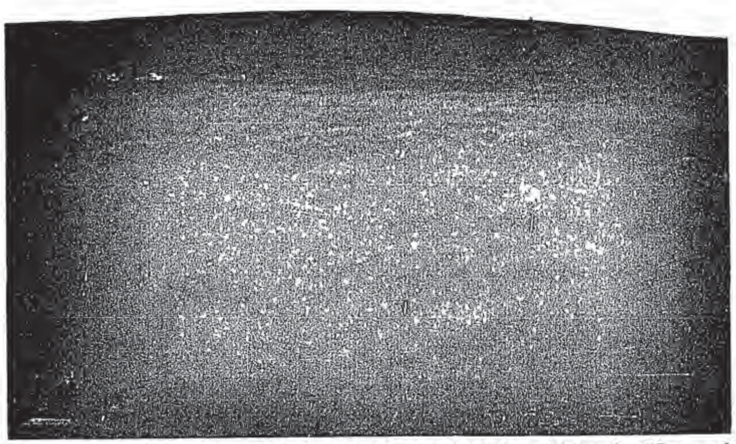


Fig. 6. A typical 'broad-based' pingo (Pingo 35, Pingok) in the Kuparuk area, in plan view (a) and facing north looking at the south slope of the pingo from its base(b). This pingo is 13 m high and 350 m in diameter. Slopes range from 5 to 15°. Topographic data adapted from Air Photo Tech 1979.

Regional Setting

The region of Alaska north of the Brooks Range crest and extending from the Chukchi Sea on the west to the Canadian border has been called the Arctic Slope or North Slope by various authors. The area of this study lies entirely within the Arctic Coastal Plain physiographic province of Wahrhaftig (1965). The nearest alpine areas are the Brooks Range, which is the northwestern extension of the Rocky Mountain Cordillera, and the geologically distinct Richardson Mountains in northwestern Canada.

Prudhoe Bay is located on the northern coast of Alaska, at latitude 70°N, longitude 148°W. This study covers the area within a 70 km radius of Prudhoe Bay, to the east, west, and south of the region defined by D.A. Walker (1985a) (Fig. 1). It includes major portions of the USGS 1:250,000 Beechey Point and Sagavanirktok quadrangles.

Regional Landscape Units

D.A. Walker and Acevedo (1987) divided the region within the Beechey Point and Sagavanirktok quadrangles into four landscape units: 1) flat thaw-lake plains, 2) gently rolling thaw-lake plains, 3) hills, and 4) river floodplains. All areas include some floodplain units (Fig. 7). The landscape units of Walker and Acevedo (1987) differ somewhat from the definition of Forman and Godron (1986). In order to keep this distinction clear, the term 'landscape units' is used when referring specifically to the regional units defined by Walker and Acevedo, and the term 'study areas' is used when referring to the landscapes as divided for this study.

The flat thaw-lake plains represent an ancient floodplain surface between the Kuparuk and Sagavanirktok Rivers. This was a glaciofluvial outwash plain during the melting of the Brooks Range glaciers 8,000 to 10,000 BP (Rawlinson 1984b;

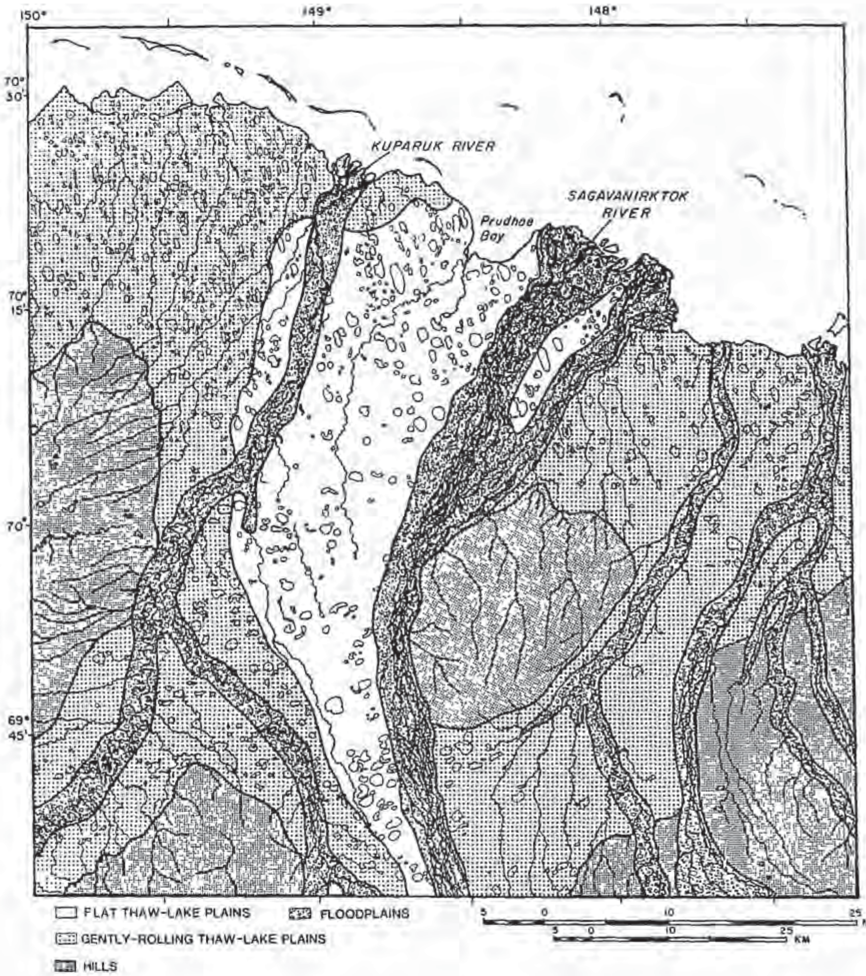


Figure 7. Landscape units of the study region. Units are from D.A. Walker and Acevedo (1987).

D.A. Walker and Acevedo 1987). The exact age of the surface is not known, but the oldest available radiocarbon date on basal peat of 9,330 ± 150 BP represents a minimum for initiation of peat formation in the region (Everett 1980b; D.A. Walker and Acevedo 1987). The gently rolling thaw-lake plains are an older surface of unknown age.

Landscape Elements

Most of the region's terrain is flat, and the predominant landscape elements are related to the presence of continuous permafrost up to 600 m deep (Everett 1980b). D.A. Walker et al. (1986) subdivided landscape elements into two types: 1) landforms, which are large landscape units that may contain within them one or more surface forms, and 2) surface forms, which are smaller-scale units. The landscape is dominated by lakes, which occupy 25-30% of the surface, and most lakes have a long axis orientation of N15°W (perpendicular to the primary wind vector) (Black and Barksdale 1949, Everett 1980a). Most regional landscape elements are a result of two related processes, ice-wedge formation and the thaw lake cycle.

A theory of ice-wedge formation based on thermal contraction cracking is generally accepted as the best explanation for this phenomenon (Leffingwell 1915, 1919; Lachenbruch 1959, 1966). Cracks form in the ground during the winter due to low temperatures, either at the surface or within the permafrost (Mackay 1984). Water flows into these cracks in the springtime, freezes, this process is repeated over the years, and an ice wedge develops. The intersection of the cracks results in a polygonal ground pattern expressed on the surface. The ice wedge displaces soil around the edge of the polygons, resulting in a raised rim.

A cyclical process of pond and lake formation, followed by subsequent lake drainage and reestablishment of ice wedges, has been called the 'thaw-lake cycle', and this process shapes most of the region's landscape (Hopkins 1949; Britton 1957;

Everett 1980a; and others). The cycle is a result of natural thermokarst processes, going from ice-wedge polygons to growing lakes to drained lake basins. Pingos may form during the drained lake stage, and are therefore considered part of the the thaw-lake cycle. Landforms, other than pingos, that are associated directly with the thaw-lake cycle include low-centered polygons, high-centered polygons, and strangmoor/disjunct polygon rims. An important element of the landscape that is not a result of ice wedges or thaw lakes is the large, braided rivers that cross through the region. These have their headwaters in the foothills and mountain valleys of the Brooks Range, and as landscape corridors they have had a major influence on the regional vegetation and wildlife.

Climate

Regional climate is characterized by long, cold winters and short, cool summers (D.A. Walker 1980). Mean annual temperature at the Prudhoe Bay (Deadhorse) airport is -13°C. Mean annual precipitation is quite low, around 25 cm annually. There is a maritime influence along the coast that leads to different summer climates between the coastal and inland areas of this study (Haugen and Brown 1980). The southern extent of the maritime influence is unknown, but is best expressed as a north to south gradient. Summer conditions at the coast are predominantly cloudy, moist, cool, and windy with temperatures within a few degrees of freezing, while clear skies and more variable wind speed and direction are prevalent inland.

Winter climate is probably more uniform across the area, although most available data are from Prudhoe Bay. Monthly means for January through March are around -30°C, and the sun is down for 49 days (Gavin 1973; D.A. Walker 1985a).

Precipitation near the coast is frequent in the summer, but total amounts are small. Away from the coast precipitation events become less frequent but produce

more moisture, so that the net result is approximately equal amounts across the region (Kane and Carlson 1973; Dingman et al. 1980; Haugen and Brown 1980).

Geology

The area lies within the broad geographical area called Beringia that includes the portions of northeastern Asia and northwestern North America that lay outside the Plio-Pleistocene continental ice sheets (Fig. 8) (Hopkins 1967, 1982). Five glacial sequences are recognized in the Brooks Range; none of these reached as far north as the study region. From oldest to youngest they are: 1) Gunsight Mountain (Tertiary), 2) Anaktuvuk River (early Pleistocene), 3) Sagavanirktok River (middle Pleistocene), 4) Itkillik Phase I (early Wisconsin), and Phase II (late Wisconsin), and 5) Fan Mountain neoglacial (Hamilton 1982, 1983, 1986; Hamilton and Hopkins 1982). The Gunsight Mountain drift is the furthest north, and reaches its northern extent approximately 35 km south of the southernmost pingo sampled.

Surficial geology is dominated by unconsolidated late Cenozoic fluvial, glaciofluvial, eolian, and lacustrine sediments collectively called the Gubik formation, but contained within this unit are a number of distinct and different aged units (Smith and Mertie 1930; O'Sullivan 1961; Black 1964; Rawlinson 1984b; Brigham 1985). There are also a few scattered areas of marine sediments.

The flat thaw-lake plains are a combination of alluvium and glaciofluvial outwash, with a few isolated coastal occurrences of Flaxman marine deposits and Sangamon age sand and gravel deposits (Rawlinson 1984b, 1986a,b,c,d,e; Hickmott 1986a,b). A surface layer of loess and peat 0.5 to 1 m deep overlies most of this surface (Everett 1980b; Rawlinson 1984b). The area's drainage history is complex, but it is evident that the Putuligayuk River was once the primary channel for the Sagavanirktok River and perhaps also the Kuparuk River (Rawlinson 1984b). A sandy gravel outwash 3 to 5 m deep is of Birch and Duvanny Yar age (see Chapter III

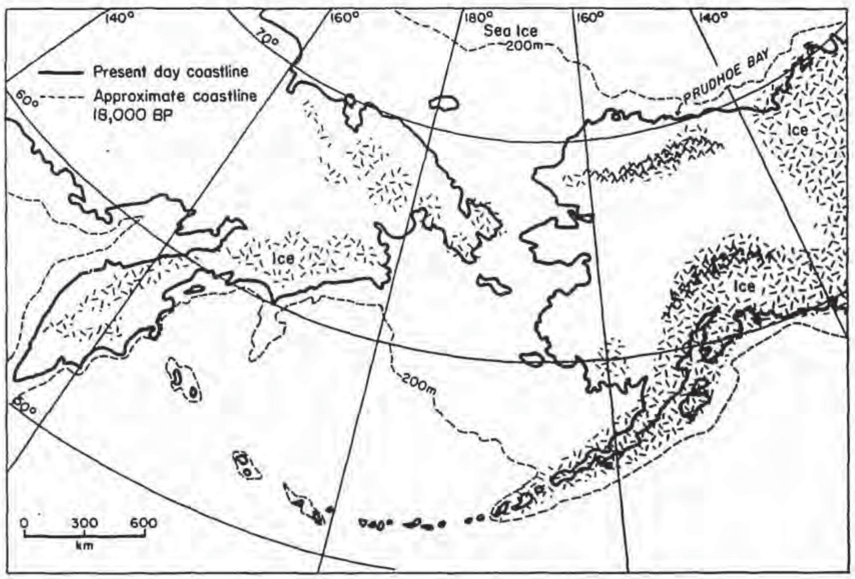


Figure 8. Hypothetical reconstruction of Beringia 18,000 BP. Continental ice sheets and mountain glaciers are shown with the stippled pattern (adapted from Barry 1982).

for dates) and has been termed the Putuligayuk outwash by Rawlinson (1984b). A layer of sandy gravel alluvium approximately 7 m thick (the Putuligayuk alluvium) underlies the Putuligayuk outwash (Rawlinson 1984b).

The gently rolling thaw-lake plains are covered by a layer of peat and are similar to the flat thaw-lake plains, but have more relief and represent older surfaces. Rawlinson (1984b) described a sequence of three terraces from the Colville River eastward toward the Kuparuk River. These same terraces are repeated on the gently rolling thaw-lake plains east of the Sagavanirktok River. The region of this study that is within the gently rolling thaw-lake plains is to the east of the oldest of these terraces, Colville terrace I. This surface is a combination of Ugnuravik sand and gravel, which are speculated to be Sangamon age (Rawlinson 1984b).

The major areas of hills, Franklin Bluffs and the White Hills, represent isolated Tertiary surfaces of the Sagavanirktok formation (Payne et al. 1951). Floodplains throughout the region consist of Holocene gravelly sand and silt; peat has formed in less active areas.

Soils

Four U.S.D.A. soil orders are represented regionally: Histosols, Entisols, Inceptisols, and Mollisols (Everett 1975, 1980c; Everett and Parkinson 1977; Parkinson 1978; D.A. Walker 1985a). U.S. Soil Conservation Service maps are not available for this region, and soil mapping has been done as part of geobotanical mapping within the oil field (Everett 1980c; D.A. Walker et al. 1986; and others). The primary environmental gradients controlling regional soils are drainage, disturbance, age, and deposition of calcareous loess.

Soils of wet tundra are either Pergelic Cryofibrists or Pergelic Cryohemists, or in regions where calcareous eolian loess is important Histic Pergelic Cryaquepts are found (Everett and Parkinson 1977). In this case, environmental conditions at the site would normally lead to a Histosol, but the input of eolian mineral material dilutes the surface horizon such that it is classified as mineral rather than organic.

Soils of moist sites are Pergelic Cryaquolls, Pergelic Ruptic Aqueptic Cryaquolls, or Pergelic Cryosaprists. The primary difference between the first two types is related to frost disturbance, with Pergelic Ruptic Aqueptic Cryaquolls found primarily in frost scar terrain, often in close association with Pergelic Cryaquolls. Pergelic Cryaquolls may be found on any relatively undisturbed upland surface with moderate drainage. Pergelic Cryosaprists are distinguished from Pergelic Cryaquolls by the presence of an organic surface horizon (Everett and Parkinson 1977; Everett 1980c).

Pergelic Cryorthents and Pergelic Cryopsamments are soils of active alluvium and sand dunes, respectively. These Entisols have little or no differentiation of horizons; they represent either young surfaces or chronic disturbances (Everett 1980c; D.A. Walker 1985a). In the absence of disturbance, soils of well-drained sites, including most pingos, are either Pergelic Cryoborolls or Calcic Pergelic Cryoborolls. Dry soils throughout the region have abundant free carbonates present due to carbonate-rich parent materials, but the concentration of this carbonate within the soil subhorizons is a function of time, leading to a Calcic Pergelic Cryoboroll (Parkinson 1978).

Arctic Vegetation

The literature of the vegetation of the circumpolar Arctic is vast. Polunin estimated in 1955 there there were nearly 10,000 titles in the field of arctic botany, and the number must now be many times that. Much of this literature is in Russian and German and has not been translated into English. The synthesis work of Aleksandrova (1980) opened up much of this foreign literature, and provided a much

needed overview. Studies of large regions include the work of Böcher (1938) in Greenland, Bliss (1977) on Devon Island, and Spetzman (1959) in Alaska.

Studies of Alaskan Coastal Plain vegetation have been concentrated mainly around Barrow and the area within the National Petroleum Reserve - Alaska (NPR-A) (Cantlon 1961; Britton 1967; Komárková and Webber 1976, 1978, 1980a; Brown et al. 1980; Ebersole 1985). Other North Slope vegetation studies, outside of the immediate Prudhoe Bay region, include Hanson (1951, 1953), Johnson et al. (1966), Anderson (1974), Dean and Chesemore (1974), Young (1974), Racine (1976), Racine and Anderson (1979), and D.A. Walker (1985b; Walker and Acevedo 1987). These cover the Beechey Point and Sagavanirktok Quadrangles, the Noatak and Kobuk river valleys, the Seward Peninsula, the Cape Thompson area, Kuskokwim Flats, and Eagle Summit. Studies of alpine areas include Jordal (1951), Lambert (1968), Batten (1977), and Cooper (1983, 1986). Gill's (1971) Mackenzie River Delta study is also pertinent.

Very little work had been done in the Prudhoe Bay region prior to its selection as a secondary study site for the U.S. Tundra Biome program under the auspices of the International Biological Program (IBP) in the early 1970's (Brown 1975). Much of what has been done is directly related to impacts from the oil development and was funded by oil companies as required to meet permitting and environmental impact assessment needs. The vegetation within the Prudhoe Bay region was described by Brown (1975) and D.A. Walker (1985a; Walker et al. 1980). Detailed geobotanical maps of much of the Prudhoe Bay oil field have been made at a scale 1:6,000 (D.A. Walker and Webber 1980a, D.A. Walker et al. 1986). Koranda's (1960) work on Franklin Bluffs, a relict Tertiary surface southeast of Prudhoe Bay, was one of the first detailed studies in this area, but it concentrated on the bluffs, which are atypical. The work of Komárková and Webber (1980b), Murray et al. (1980), and Walker and Webber (1980b) along the Dalton Highway, described the vegetation and flora along a north to south transect from Prudhoe Bay to the Yukon River. Viereck and Dyrness (1980; Viereck et al. 1986) have synthesized the classification systems in use in Alaska.

Pingo vegetation has been briefly covered in a number of studies. D.A. Walker (1985a) described the major communities on the pingos within the Prudhoe Bay region, and Ito (1978) described the vegetation of several pingos in the vicinity of Tuktoyaktuk, N.W.T., Canada. Burns (1964) described several small pingos in the Yukon-Kuskokwim river delta region, but it is unclear whether or not these were truly pingos or some other type of ice-cored mound such as a palsa. There are also references to pingo vegetation in Mackay (1973, 1976) and Andreev and Perfilyev (1975).

Vegetation of the Study Region

Aleksandrova (1980) divided the Arctic into two regions, the tundra region and the polar deserts. Using her classification, the region of this study is within the Alaska subprovince of the Chukotka-Alaska province of the subarctic tundra sub-region of the tundra region, but is rather near the northern border between the arctic and subarctic tundras, and contains elements of both. Thus, she relates the vegetation of this region more closely to Siberia than to most of the North American Arctic, but she separates it at the subprovince level due to a number of differences, including the species that form the northern tree limit. All of Alaska, except the southeastern panhandle, the extreme northwest of Canada west of the Mackenzie River, and the Chukotka region of Siberia are collectively known as Beringia (Yurtsev 1974a; Hopkins et al. 1982). The non-mountainous portions of Beringia were not glaciated during the Pleistocene, and the continents were connected via a land bridge because of lowered sea levels (Hopkins 1967; Fig. 8). The mesic tundra throughout this area

is similar, dominated by *Eriophorum vaginatum* and *Carex lugens* tussock tundra, which are absent east of the Mackenzie River (Aleksandrova 1980).

The Prudhoe Bay vegetation differs from most of the coastal plain by the rarity of *Eriophorum vaginatum*. D.A. Walker (1985a) described the stand types at Prudhoe Bay. Dry areas are dominated by *Dryas integrifolia*, and *Carex rupestris*, *Oxytropis nigrescens*, *Saxifraga oppositifolia*, and *Lecanora epibryon* are also common. Plant cover in dry sites is often incomplete, and crustose lichens may be abundant. Moist tundra in mesic uplands is dominated by *Eriophorum angustifolium*, *Dryas integrifolia*, *Tomenthypnum nitens*, and *Carex aquatilis*. Fruticose lichens are abundant in all but the wettest sites. Wet tundra is dominated by *Carex aquatilis*, and these sites have few dicots present. *Carex aquatilis* and *Arctophila fulva* are the predominant emergent species.

The study region extends beyond the oil field and includes large expanses of tussock sedge, dwarf shrub tundra dominated by *Eriophorum vaginatum*, *Carex lugens*, and species of *Salix* and other dwarf shrubs.

Environmental Gradients

The relationship of vegetation to the landscape and its environment has been an important part of modern vegetation science from its beginnings in Europe in the latter part of the last century. It was Ramensky (1924, 1930), Gleason (1926), and Lenoble (1927), however, who first argued that classification into discrete units was not a natural model, but that vegetation varies continuously across a landscape. This idea of gradients was developed in North America by Whittaker (1948, 1951, 1967, 1973), Curtis and others (Curtis and McIntosh 1951; Brown and Curtis 1952), Ellenberg (1950, 1952), and Perring (1958). It was considered a radical departure from the strict European school of classification and was not readily accepted by most phytosociologists (Shimwell 1971). In the past several decades, however, classification and gradient analysis have come to be recognized as complementary techniques that when used together can give the best understanding of vegetation patterns and their causes (Waring and Major 1964).

It has been fairly recently that gradient analyses have been regularly done in conjunction with descriptive studies, as there were few reliable and consistent methods available. For this reason there are few such studies in the arctic. Webber (1971) was the first study that concentrated on techniques of gradient analysis in arctic regions. He cites as earlier studies Summerhayes and Elton (1928), Hansen (1930), Seidenfaden and Sørenson (1937), Böcher (1954), Aleksandrova (1960), Beschel (1963), Raup (1965), and Johnson et al. (1966), but none of these were really gradient analyses as the term is applied today. Cantlon (1961) first pointed out the need for such studies in Alaska, and D.A. Walker (1985a) followed his model of considering the vegetation of Prudhoe Bay in terms of three scales of environmental gradients (micro-, meso-, and macroscale). Other studies that have directly considered gradients include Bliss (1956), Gill (1971), Webber (1978; Webber et al. 1980), Cooper (1983), Odasz (1983), and Ebersole (1985).

The field of landscape ecology, which is still emerging in North America (Naveh and Lieberman 1984; Forman and Godron 1986), takes gradient analysis another step, from community-level gradients to landscape-level gradients. This field has been active in Europe for some time (Troll 1950, 1968; Neef 1963, 1967; Haase 1964; Schmithusen 1964, 1967; Bobek and Schmithusen 1967). It has come into popular usage in North America relatively recently (e.g. Komárková 1976). This type of approach is useful in northern Alaska, where there are large, unbroken expanses of tundra. The present study views vegetation at the level of the landscape, as well as on the smaller scale of the community. M.D. Walker and D.A. Walker (1985) examined macroscale gradients on the USGS 1:250,000 Sagavanirktok Quadrangle in relation to Landsat multispectral scanner data and color infrared

photography; this is one of the first studies to use indirect gradient analysis techniques at the landscape scale.

Floristics

Floristics and vegetation are closely related topics, but whereas vegetation studies concentrate on regional patterns and communities, floristics is the study of the flora and its history. The regional flora contains the building blocks for communities. The work of Hultén (1937, 1958, 1962, 1963a,b, 1968), Löve and Löve (1963), Johnson and Packer (1967), Young (1971, 1976, 1982), Murray (1978, 1979, 1980; Murray et al. 1983), Yurtsev (1962, 1963, 1972a,b, 1974a,b, 1978, 1982), D.A. Walker (1985a), and Cooper (in prep.) have all dealt with floristics directly.

The arctic contains only about 600 indigenous species, and about 200 of these are circumpolar (Löve 1959). The Bering land bridge served as a migration route between Asia and North America for many species during the Pleistocene. Because of this connection, and also because Alaska was isolated to the east by the continental ice sheet, the Alaskan flora is a combination of circumpolar and Beringian species, including Alaskan endemics (Hultén 1968). The floristic link between Alaska and Asia was first described by Hultén (1937), and the more recent work by Yurtsev (1982), and Murray (1980; Murray et al. 1983) has continued to elucidate this connection and its role in the formation of the modern vegetation.

Because of its importance to these trans-continental migrants, the Pleistocene environment of the land bridge and Beringia in general has generated considerable interest (Hopkins 1967; Hopkins et al. 1982). Many workers have envisioned that much of this area was a 'steppe-tundra' (Hibbert 1982). The first use of this term or concept was from German and Russian paleontologists, who were trying to imagine an environment that could have supported a large ungulate fauna. Nehring (1890) discovered these Pleistocene vertebrate faunas in central Europe, and he wrote of a "steppe climate with an arctic tinge to it" (Nehring 1895). Tugarinov (1929), a Russian paleontologist, further developed the concept, and made the suggestion that the mammalian fossil associations found in northern Asia, Europe, and Alaska represented actual faunal assemblages, and were not just together due to some type of depositional coincidence. This is a matter of considerable debate today, as there has been little stratigraphic control in the major paleontological finds (Matthews 1982).

The important point is that the history of the steppe-tundra concept was based on the existence of large, grazing animals now associated with grassland and shrub steppes, rather than on direct evidence from the pollen or plant macrofossil records. Following the development of the concept, however, independent evidence from Europe, Russian, and North American palynologists developed that supported the idea. When the Soviet picture of a cold, dry, environment was first presented in the U.S. in 1965 (Giterman and Golubeva 1967), it was very similar to the independent conclusions drawn by Colinvaux (1964) from Imuruk Lake cores. Earlier, Livingstone (1955, 1957) recognized three pollen zones from two lakes in northern Alaska, an herb zone, dominated by grasses, sedges, and species of *Artemisia*, a birch zone, and an alder zone, the most recent. These three zones have held up remarkably well in other studies (Anderson 1982; Nelson 1982a; Brubaker et al. 1983; Baker 1984; Wilson 1984), and the herb zone has been postulated to represent this steppe-tundra type. Yurtsev (1982) described the modern pollen rain from Wrangel Island, which he considers a good modern example of a northern steppe-tundra type, and it is quite similar to many of the herb-zone pollen spectra, with high amounts of grasses, *Artemisia* spp., and *Selaginella sibirica*. Ritchie (1984; Cwynar and Ritchie 1980) has argued that there is no concrete evidence for these steppe-tundra assemblages, and that the herb zone probably represents a fellfield type of vegetation.

Island Biogeography

Since the publication of MacArthur and Wilson's theory of island biogeography in 1967, there have been a multitude of studies attempting to verify or refute the theory or fit specific data to it. This theory states that the number of species on an island represents an equilibrium between extinction rates and colonization rates. Colonization rates depend on distance from source areas; extinction rates depend on island size. The theory predicts that the number of species on an island, S, is exponentially related to the area of the island:

$$S = cA^z. \tag{1}$$

Data have been published for plants and animals, on oceanic islands and archipelagoes (Terborgh 1973; Diamond et al. 1976; Linhart 1980; Nilsson and Nilsson 1978, 1982, 1983) as well as terrestrial habitat patches (Culver 1970; Vuilleumier 1970; Brown 1971; Cook 1974; Johnson 1975; Behle 1978; Crowe 1979; Riebesell 1982; Murray et al. 1983). No arctic data have ever been applied to the theory, although Young (1982) recognized its significance in relationship to the steppe-tundra question, because if modern steppe-tundras are relict, they should have non-equilibrium biogeography.

The theory leads to three specific predictions: 1) that turnover of species occurs, 2) that the species-area curve will be steepest for the most isolated islands and steeper on islands than on equivalent mainland areas, and 3) that a dynamic equilibrium is in fact operating (Williamson 1981). The value of z in equation (1) represents the slope of the line defined by the species-area equation, and most attention has centered on this aspect. Gould (1979), however, pointed out that when the slopes of two equations are equal, the ratio of their intercepts, c, represents the relative species richness of different regions of similar size. Some data suggest that isolation lowers the value of c, rather than having a direct effect on z (Slud 1976).

MacArthur and Wilson did not make specific predictions about the value of z, but they did present empirical data suggesting that for equilibrium island populations it should fall between .24 and .35, and for mainland sites, or subsamples within an island, it should be between .12 and .17. Slopes outside these ranges have been interpreted as representing nonequilibrium conditions. Nonequilibrium would be expected in areas where islands have formed from isolation of a previously extensive habitat, such as mountaintops that were isolated by climatic change during the Holocene, or oceanic islands in the Bering Strait that were continuous with the exposed continental shelf during the Pleistocene (Brown 1971; Riebesell 1982; Young 1982). In these situations, extinction rates are expected to exceed colonization rates, as these areas have gone from part of the mainland to islands, and it takes longer to reach an equilibrium from too many species than from too few species for a given area (MacArthur and Wilson 1967). Once a viable population is established, it will tend to persist. MacArthur and Wilson (1967) hypothesized that on mainlands the lower z values are a result of the presence of small populations of many species in an area that belong to a larger, viable population. On an island these small populations could not persist. High colonization rates, as compared to extinction rates, should lead to decreased values of z, as colonization rate is more dependent on isolation than it is on area.

Predictions concerning turnover and equilibrium have been difficult to treat. The prediction of species turnover does not address the time scales involved. Simberloff (1976) suggested that this effect is likely to be seen only in very small systems, while others (*e.g.* Gilbert 1980) have claimed that because turnover has not been shown in most cases, the theory is nullified. It has also been shown that observed 'turnover' often represents sampling error (Lynch and Johnson 1974; Simberloff 1976; Nilsson and Nilsson 1982). It is true that this prediction cannot be invalidated, but this does not make the theory intractable. Colonization and

subsequent extinction of some species has been well documented in successional studies (e.g. Crocker and Major 1955, and many others). Arctic plants are relatively long-lived (Billings 1973), and so it would be difficult to definitively demonstrate turnover for a given area in the Arctic. For the present study it is assumed that turnover has occurred at some point in time.

Succession

Because the pingos may well be some of the oldest stable sites on the coastal plain, they are logical sites on which to study succession. Within the last decade studies of plant succession have concentrated on mechanisms driving succession and the evolutionary consequences of succession (*e.g.* Pickett 1976; Peet and Christensen 1980; Christensen and Peet 1984; Tilman 1986). In tandem with these experimental studies have come a series of models of succession that attempt to define the process as a generalized one driven by similar forces in all environments (*e.g.* Connell and Slatyer 1977; Tilman 1985). Clements' (1916) original concept was that all recently exposed surfaces will eventually become inhabited by plants, and that this process of going from barren ground to plant cover is highly deterministic and has a pre-determined end-point based on existing environmental factors at the site. This concept of the *climatic climax* contains within it the concept of *convergence*, that is, change within a community to a particular stand type, the climax type. This concept was inherent in the earliest successional descriptions (Cowles 1899; Cooper 1916; Clements 1916, 1928).

Since these earliest works, two critical questions have emerged from successional studies in many different environments: 1) whether or not there is convergence toward a characteristic climatic climax type, and 2) the importance of site factors or characteristics as opposed to chance immigration in determining both the initial and final species composition (Margalef 1963, 1968; Connell and Slatyer 1977;

Matthews 1979a,b; Christensen and Peet 1984; Matthews and Whittaker 1987). Margalef (1963, 1968) and Egler (1954, 1975) both recognized that initial species composition is likely to be controlled by the chance arrival of species at the sites, although they predicted entirely different outcomes from this initially apparently random assemblage. Matthews and Whittaker (1987) concluded that species composition differences along environmental gradients increase rather than decrease over time, and Pineda et al. (1981) and Christensen and Peet (1984) came to a similar conclusion.

Arctic Studies

There have been rather numerous descriptive studies of succession in northern Alaska and other regions of the Arctic, many associated with river alluvium (Polunin 1936; Spetzman 1951; Bliss and Cantlon 1957; Svoboda 1977; Peterson and Billings 1978, 1980; Billings and Peterson 1980; L.R. Walker and Chapin 1986; L.R. Walker et al. 1986; Cargill and Chapin 1987; Svoboda and Henry 1987). Churchill and Hanson (1958) reviewed concepts of arctic succession. Peterson and Billings' work related Coastal Plain succession to natural geomorphic cycles (the thaw-lake cycle in particular), and the pingos are part of the same cycle.

Svoboda and Henry (1987) have presented a model for succession in high arctic environments. It addresses the problem that in the most extreme environments, where plants are at the very limits of their tolerance range, succession as it it generally understood does not seem to occur. This model predicts that in marginal environments, competition is of little importance, as few species are able to exist at all, and these may not persist over any period of time. Savile (1960) also wrote of decreased competition in the High Arctic, and Griggs (1934) wrote that arctic environments are generally ruderal or weedy. Vascular plant communities in such settings will not develop high enough cover to build up a substantial amount of soil

organics. Thus, the site remains in an early sere for a long period of time, perhaps indefinitely.

Taking Svoboda and Henry's work one step further, one can envision a gradient from these most extreme environments, at the very limit of plant growth, to some point farther south where successional processes, as they have generally been described, are acting. At this more southerly point, competition is controlling most species' distributions, not environmental stresses. The study region probably falls somewhere in the middle of this gradient, and so is an interesting area in which to study successional processes.

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Cumulative geoecological effects of 62 years of infrastructure and climate change in ice-rich permafrost landscapes, Prudhoe Bay Oilfield, Alaska

MARTHA K. RAYNOLDS[1], DONALD A. WALKER[1], KENNETH J. AMBROSIUS[2], JERRY BROWN[3], KAYE R. EVERETT[4]†, MIKHAIL KANEVSKIY[5], GARY P. KOFINAS[6], VLADIMIR E. ROMANOVSKY[7,8], YURI SHUR[5] and PATRICK J. WEBBER[9]

[1]*Institute of Arctic Biology, University of Alaska Fairbanks, Fairbanks, AK 99775, USA,* [2]*Aerometric Geospatial Solutions, Anchorage, AK 99501, USA,* [3]*P.O. Box 7, Woods Hole, MA 02543, USA,* [4]*Byrd Polar Research Center, Ohio State University, Columbus, OH 43210, USA,* [5]*Department of Civil & Environmental Engineering, University of Alaska Fairbanks, Fairbanks, AK 99775, USA,* [6]*School of Natural Resources and Agricultural Sciences, University of Alaska Fairbanks, Fairbanks, AK 99775, USA,* [7]*Geophysical Institute, University of Alaska Fairbanks, Fairbanks, AK 99775, USA,* [8]*Earth Cryosphere Institute SB RAS, Box 1230, Tyumen, 625000, Russia,* [9]*P.O. Box 1230, Ranchos de Taos, NM 87557, USA*

Abstract

Many areas of the Arctic are simultaneously affected by rapid climate change and rapid industrial development. These areas are likely to increase in number and size as sea ice melts and abundant Arctic natural resources become more accessible. Documenting the changes that have already occurred is essential to inform management approaches to minimize the impacts of future activities. Here, we determine the cumulative geoecological effects of 62 years (1949 2011) of infrastructure- and climate-related changes in the Prudhoe Bay Oilfield, the oldest and most extensive industrial complex in the Arctic, and an area with extensive ice-rich permafrost that is extraordinarily sensitive to climate change. We demonstrate that thermokarst has recently affected broad areas of the entire region, and that a sudden increase in the area affected began shortly after 1990 corresponding to a rapid rise in regional summer air temperatures and related permafrost temperatures. We also present a conceptual model that describes how infrastructure-related factors, including road dust and roadside flooding are contributing to more extensive thermokarst in areas adjacent to roads and gravel pads. We mapped the historical infrastructure changes for the Alaska North Slope oilfields for 10 dates from the initial oil discovery in 1968 2011. By 2010, over 34% of the intensively mapped area was affected by oil development. In addition, between 1990 and 2001, coincident with strong atmospheric warming during the 1990s, 19% of the remaining natural landscapes (excluding areas covered by infrastructure, lakes and river floodplains) exhibited expansion of thermokarst features resulting in more abundant small ponds, greater microrelief, more active lakeshore erosion and increased landscape and habitat heterogeneity. This transition to a new geoecological regime will have impacts to wildlife habitat, local residents and industry.

Keywords: Arctic, climate change, cumulative impacts, geoecological mapping, ice rich permafrost, ice wedge polygons, infra structure, photo interpretation, thermokarst, tundra

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Introduction

Oil and gas exploration and extraction are occurring in ice-rich permafrost (IRP) areas of Alaska, Canada, and Russia, and it is inevitable that more extensive networks of infrastructure than presently exist will be required to extract the resources of these areas (AMAP, 2010). These will be constructed against a backdrop of rapid climate change, rapid technological changes, and unpredictable social-ecological changes (Truett & Johnson, 2000; Orians *et al.*, 2003; ACIA, 2005; AMAP, 2010; Krupnik *et al.*, 2011; Kofinas *et al.*, 2013). Documenting the history of these developments as they occur will aid local communities, researchers, land managers, industry, and policy makers in developing adaptive approaches to plan for and respond to future changes (AMAP, 2010; Streever *et al.*, 2011).

Correspondence: Martha K. Raynolds, 311 Irving, Institute of Arctic Biology, University of Alaska Fairbanks, Fairbanks, AK 99775, USA, tel. + 907 474 6720, fax + 907 474 7666, e mail: mkraynolds@alaska.edu
†Deceased.

The Prudhoe Bay Oilfield

The Prudhoe Bay Oilfield (PBO) in northern Alaska was the first developed oilfield in the Arctic, and is the largest in the United States. It is an extremely important asset to the United States and the state of Alaska, containing 16% of US proven reserves of oil and gas (Alaska Oil & Gas Association, 2012; USEIA, 2012). In 2012, taxes on the oil from the Northern Alaska oilfields accounted for over 90% of the Alaska state budget (Alaska Oil & Gas Association, 2012). The PBO is located on the Beaufort Sea coast, halfway between the Canadian border and Point Barrow, a region that was remote and roadless prior to the discovery of oil in March 1968. An extensive infrastructure network quickly grew following the oil discovery, resulting in development within an approximately 2600 km^2 area (Fig. 1a).

Oilfield engineering evolved rapidly in response to the IRP conditions encountered (Gilders & Cronin, 2000; Orians *et al.*, 2003). Gravel construction pads over 2-m thick were used to insulate the frozen tundra. Since 1995, new technologies, including much closer spacing of the well heads, directional drilling to reach deposits up to 6.4 km from the drilling sites, and reinjection of drilling fluids into the geological formations to eliminate the need for reserve pits, considerably reduced the size of gravel pads for drill sites in newer oilfields. The use of winter ice-roads and roadless access to drill sites have further reduced the footprint of modern oilfields (Gilders & Cronin, 2000; AMAP, 2010; Streever *et al.*, 2011).

The early phase of PBO development stimulated geoecological and permafrost research in the region by the Tundra Biome investigations of the International

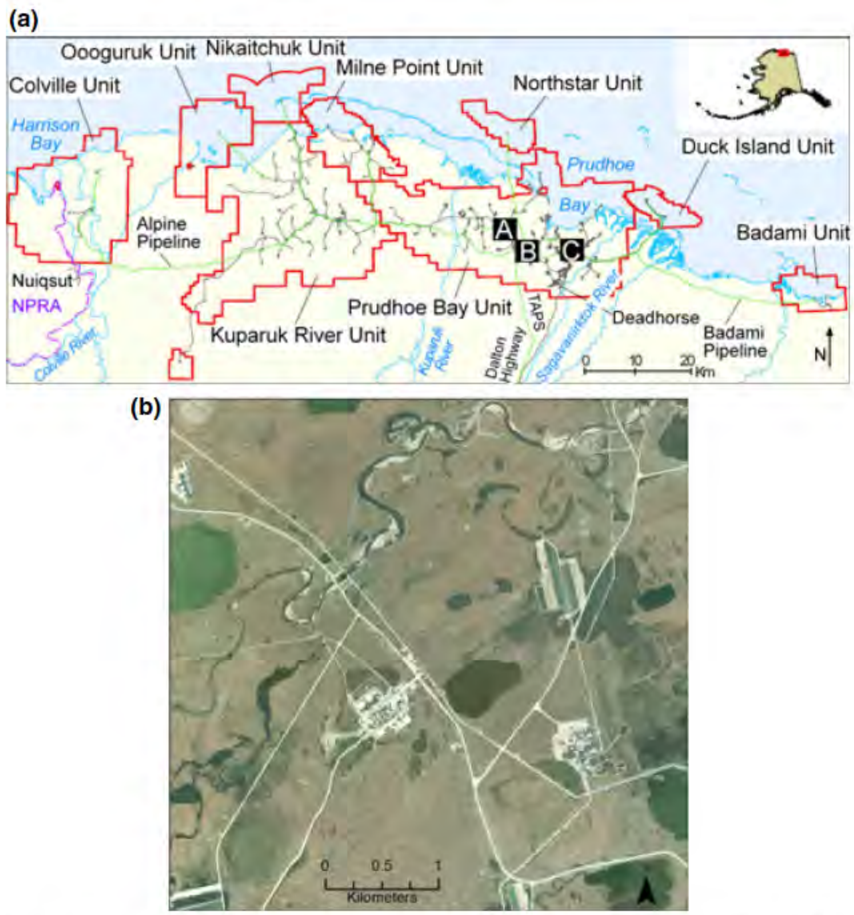

Fig. 1 Alaska North Slope production units that comprise the Prudhoe Bay Oilfield referred to in this article. (a) Major oil production units (red boundaries) and extent of infrastructure. Gray lines are gravel roads, airstrips, and construction pads. Green lines are major pipelines. Small black squares A, B, and C are locations of the detailed 20 km^2 map areas of this study. Note the Inupiat village of Nuiqsut at left in the National Petroleum Reserve Alaska, the Trans Alaska Pipeline System and Dalton Highway that link the oilfields to southern Alaska. *Inset*: Location within Alaska. Map courtesy of Aerometric, Inc. and BP Alaska, Inc. (b) Map B, showing the meandering Putuligayuk River and its floodplain, thaw lakes, drained lake basins, and primary surfaces between lakes. The infrastructure consists of a network of roads, drill sites, and processing facilities. The width of the area shown is 4.7 km. See Fig. 5 for geoecological and historical change maps of this area. Imagery courtesy BP Exploration (Alaska), Inc.

Biological Programme (Brown, 1975, 1980; Walker *et al.*, 1980); and the Circumpolar Active-Layer Monitoring Program (Brown *et al.*, 2000). Although the oil industry states that the density and extent of the infrastructure of the PBO will likely not be replicated in new oilfields, we focused our studies here because it has the longest history of development and scientific research, and is the only area with detailed time-series of geoecological and historical-change maps that span the complete history of the field. Many of the types of landscape change seen here, such as those associated with roads and climate change, will occur in other IRP areas.

IRP terrain and thermokarst

The PBO region is a showcase of periglacial landforms, including thaw lakes, pingos, meandering beaded streams, and many types of patterned ground such as nonsorted circles, and ice-wedge polygons indicative of IRP (Everett, 1980a). Within this landscape, ground-ice formation and thawing occurs in a hierarchy of time and space scales ranging from daily needle-ice formation to annual processes associated with frost-heave features (e.g., frost boils) to centuries and millennia involved with ice-wedge-polygon and thaw-lake formation (Walker, 2000). These processes create a mosaic of exceptionally dynamic landforms, soils, and vegetation that are susceptible to abrupt changes, especially if the features are disturbed by mechanical means or by rapid climate change (Lawson *et al.*, 1978; Lawson, 1982; Komárková & McKendrick, 1988; Walker, 1996, 1997; Callaghan *et al.*, 2005; Shur & Jorgenson, 2007; Shur & Osterkamp, 2007; Grosse *et al.*, 2011).

Permafrost is ground (soil or rock, in most cases including ice) in which a temperature below 0 °C exists for two or more years (Van Everdingen, 1998). At Prudhoe Bay, permafrost is continuous and extends to a depth of 660 m (Gold & Lachenbruch, 1973). The *active layer*, the layer of soil near the surface that thaws annually, varies in thickness from about 0.25 m in peaty soils near the coast to over 2 m on some south-facing gravelly slopes and averages about 0.5 m (Everett, 1980b). IRP has a high percentage of *excess ice*, where the ice in the ground exceeds the total pore volume that the ground would have under unfrozen conditions (Van Everdingen, 1998). Although IRP has high load-bearing capacity if its thermal stability is maintained, it has none if the excess ice melts. Nearly all of the developed and developing oil and gas fields in arctic Alaska, Canada, and Russia are in regions with extensive IRP.

Much of the excess ice in the Prudhoe Bay permafrost is in the form of ice-wedges, which occupy on average 11% of the total volume of the upper 2 3 m of permafrost for all sites studied along the Beaufort coast, but which can reach 30% in some areas of the Arctic Coastal Plain of Alaska (Everett, 1980b; Kanevskiy *et al.*, 2013). The average total volumetric ice content, including the wedge ice, segregated ice, and pore ice, exceeds 80% on all terrain units studied except in sand dunes and deltas (Kanevskiy *et al.*, 2013).

If the insulative vegetated mat over ice wedges is disturbed or if other causes introduce heat to the ice-wedges (e.g., standing or flowing water), thawing of the ice will result in settling of the surface and creation of *thermokarst terrain* (Fig. 2) (Van Everdingen, 1998). Although thermokarst is not a form of karst, the subsidence and collapse associated with thermokarst terrain has some analogies to karst topography (French, 1976). Considerable research has been devoted to defining the hazards of thermokarst to structures (Nelson *et al.*, 2002) and in the development of engineering solutions to avoid thermokarst formation (US Arctic Research Commission Permafrost Task Force, 2003).

Within the PBO, most of the extremely IRP occurs within approximately 2 m of the surface, in frozen, organic-rich silts that overlie more stable alluvial sands

Fig. 2 Flooding and thermokarst along roads at Prudhoe Bay, AK. (a) Typical roadside environment showing flooded troughs and resulting thermokarst terrain. Heavy road dust killed vegetation on the centers of ice wedge polygons near the road. (b) Thermokarst with active erosion of resulting high centered polygons and widening of polygon troughs, near a gravel road. (c) Aerial photograph of flooding of ice wedge polygon troughs due to blocked drainage. Previous low centered ice wedge polygons were converted to high centered polygons. *Photos:* D.A. Walker, 1 July 2013.

and gravels (Everett, 1980b; Rawlinson, 1993) (Fig. SA1). See Supporting Information, Appendix S1, for further explanation of IRP in the PBO with photographs of segregated ice and a cross section of a 100-m trench illustrating the types and amount of ice in the substrate (Fig. SA5).

Scattered small *thermokarst pits* were mapped during the baseline geobotanical mapping in the PBO (Walker *et al.*, 1980). These small near-circular ponds commonly overlie the intersections of ice wedges, most commonly on *primary surfaces* (old landscapes unaltered by thaw-lake or river processes) (Sellmann *et al.*, 1975), wherever there has been time for large ice-wedges to form (Fig. SA6a). These pits have been identified in the Russian literature as the first step in thermokarst development (Shur & Osterkamp, 2007). Until recently, the thermokarst pits in natural landscapes at PBO appeared to be fairly stable because most of the pits noted in the 1970s were also visible on the 1949 aerial photographs and showed little change through 1990 (Fig. 3). Recent abrupt changes in thermokarst pit size and density were documented in the Fish Creek region, about 40 km west of the PBO. These changes were thought to be a response to a period of warm summer temperatures during the 1990s (Jorgenson *et al.*, 2006). As discussed below, our study documents a similar trend in thermokarst in the PBO.

Cumulative effects of oil development

The effects of oil and gas activities take different forms in different parts of the Arctic, where environmental and social conditions vary. Summaries of some of the cumulative effects have been addressed in Russia (Forbes *et al.*, 2009; Walker *et al.*, 2011; Kumpula *et al.*, 2012), Alaska (Walker *et al.*, 1987; Orians *et al.*, 2003), and globally (AMAP, 2010). The cumulative effects of oil development were first studied at Prudhoe Bay in the 1980s with mapping that quantified the extent of direct and indirect landscape effects of oilfield infrastructure (Walker *et al.*, 1986a,b, 1987). *Direct effects* were defined as physical changes that were planned in advance, such as roads, gravel pads, and gravel mines. The unplanned *indirect effects* were more difficult to anticipate (Walker *et al.*, 1987; Shur, 1988). They often occurred in areas adjacent to infrastructure and

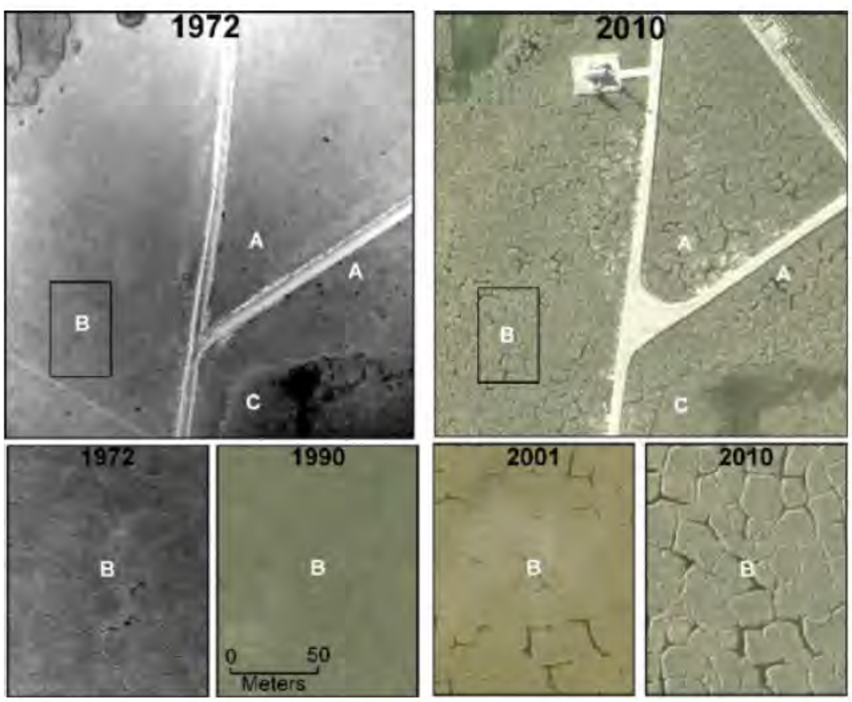

Fig. 3 Changes between 1972 and 2010 near the main road junction in map B. Note changes in infrastructure, including small new gravel pad and building near the top of the photo, a new pipeline and gravel road on the upper right side, and an expanded road inter section. Also note extensive changes in the character of the ice wedge polygons with many more flooded ice wedge troughs between polygons in 2010. Areas near A and B have extensive ice wedge polygons. Numerous scattered small thermokarst pits (small dark ponds) are visible near A in 1972, with little change in 1990. Areas A and B and show marked increases in thermokarst in 2001 and even more by 2010. Area C is in a drained thaw lake basin with ice poor soils and shows little evidence of new thermokarst. Changes in area A were mapped as infrastructure related because of the proximity of the road and the pipeline that have significantly altered the hydrology of this area. Area B is not clearly affected by infrastructure.

included roadside dust, infrastructure-related flooding, off-road vehicle traffic, and thawing of near-surface permafrost (Walker *et al.*, 1987) (Appendix S3).

A 2003 US National Research Council (NRC) study of the cumulative effects of oil development on Alaska's North Slope (Orians *et al.*, 2003) included a time-series inventory of the total extent of infrastructure on the North Slope (Ambrosius, 2003). Although the NRC study recognized that climate change would likely have numerous effects to sea ice and Arctic ecosystems, the report concluded that climate change would not seriously affect oil and gas activities on the North Slope (Orians *et al.*, 2003). This was based largely on the assumption that cold, continuous permafrost, such as that found in the PBO, is robust and not likely to thaw even if the permafrost temperatures were raised several degrees. Since the NRC study, several regional studies have pointed to terrain and vegetation changes related to climate change (Sturm *et al.*, 2001; Jia *et al.*, 2003; Bhatt *et al.*, 2010; Myers-Smith *et al.*, 2011; Epstein *et al.*, 2012; Tape *et al.*, 2012), including recent thawing of the near-surface permafrost (Jorgenson *et al.*, 2006).

Here, we update both the regional assessment of total infrastructure extent from the NRC report (Orians *et al.*, 2003) and the analysis within three 20-km^2 areas previously analyzed in 1983 (Fig. 1) (Walker *et al.*, 1987). This article addresses the questions 'How have oilfield infrastructure and climate change affected the IRP landscapes of the PBO over 62-year of observations?' And 'How have the initial geoecological conditions in the region affected the changes?'

Materials and methods

Total North Slope oilfield infrastructure footprint

BP Exploration (Alaska) Inc. maintains a time series of aerial photographs and a set of topographic base maps (map scale 1 : 6000) of the PBO. Aerial photos taken in 1968 were used to define the areas prior to most development. Infrastructure changes were added from each successive analysis year (1973, 1977, 1983, 1988, 1994, 2001, 2006 2007, 2010, and 2011), creat ing CAD files for calculating incremental changes for the area shown in Fig. 1a. The files contained the areas covered by gravel facilities, roads, and mine excavations. Indirect effects, exploration facilities, riverbed gravel extraction, and other impacted areas not covered by gravel facilities were not included on these maps. Calculations were performed using ARC View software in an Alaska State Plane, zone 4, NAD27 projection (Appendix S2).

Integrated geoecological and historical change maps

A detailed mapping approach was used to examine both direct and indirect landscape changes within three 20 km^2 areas (A, B, C, example shown in Fig. 1b). The set of aerial photo missions used for the analysis included the years 1949, 1968, 1970, 1972, 1973, 1977, 1979, 1983, 1990, 2001, and 2010 (Table SC1). The methods used in the first analysis of cumula tive effects of oil development (Walker *et al.*, 1986a,b) were modified for this update to take advantage of new advances such as heads up digitizing, GIS database formatting and improved infrastructure maps of the region. The database con tained polygons coded with nine geoecological attributes: dominant vegetation, secondary vegetation, tertiary vegeta tion; percentage open water; landform; dominant surface form, secondary surface form; dominant soil, and secondary soil (Table SC2). Secondary and tertiary variables were mapped if they covered more than 30% of a map polygon. Eighteen infrastructure related change attributes, and six non infrastructure related change attributes were also mapped (Table SC2).

Results

North Slope oilfield infrastructure footprint

Results for the entire oilfield are presented by number of infrastructure items, length of linear infrastructure items, and area covered by facilities (Fig. 4a c respectively). As of 2011, there were 127 production pads, 25 facility pads, 145 support pads (power stations, camps staging areas, etc.), 103 exploration sites, 13 offshore exploration islands, 7 offshore production islands, 9 airstrips, 4 exploration airstrips, 2037 culverts, 27 bridges, 50 caribou crossings, and one active landfill. The number of these infrastructure items increased rapidly between 1968 and 1983 and more slowly since then. The number of exploration pads decreased slightly after 2001, but increased again after 2007 (Fig. 4a and Table SB1).

The road network consisted of 669 km of gravel roads, 154 km of abandoned peat roads, 12 km of causeways, 96 km of abandoned tractor trails, and 54 km of exploration roads with thin gravel or tundra scars. Similar to the number of facilities, the total length of roads increased rapidly until 1988 and then leveled off at 931 km (Figs 1a and 4b). The 790-km pipeline network includes groups of parallel pipelines elevated 1 2 m above the tundra surface on vertical supports. Pipeline corridors included anywhere from 1 to 21 closely spaced parallel pipelines with diameters up to 60 cm. The length of major powerlines with towers totaled 541 km (Table SB1).

The total oilfield infrastructure covered 7429 ha of the North Slope by 2011, mainly consisting of 2345 ha of gravel pads, 2737 ha of gravel mines, and 1255 ha of gravel roads and causeways (Fig. 4c). Impacted areas also included airstrips (125 ha), offshore gravel pads and islands (82 ha), exploration sites (290 ha),

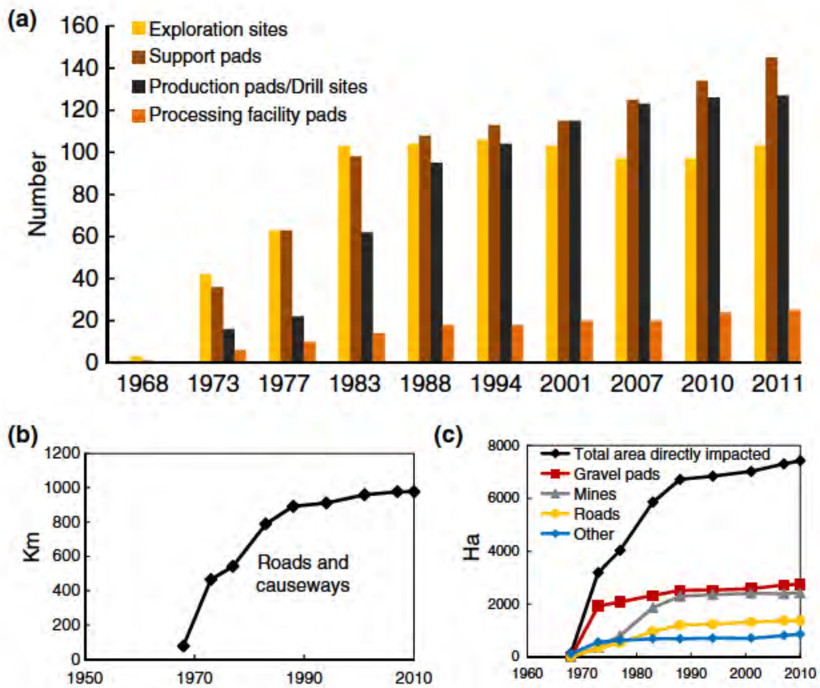

Fig. 4 History of infrastructure on the North Slope oilfields 1968 2011 (excluding the Dalton Highway and Trans Alaska Pipeline System). (a) Number of infrastructure items, (b) total length of roads (km), (c) directly impacted area (ha). Data courtesy of Aerometric, Inc. and BP Exploration (Alaska), Inc.

exploration airstrips (20 ha), peat roads (209 ha), tractor trails/scars (104 ha), exploration roads (72 ha), and areas where pads have been removed and are in the process of recovery (190 ha). Most of the direct effects to the landscape occurred within 18 years of the initial discovery of oil, reaching 6722 ha by 1988 (Fig. 4c). Since 2001, the oil development has expanded westward, increasing the infrastructure area to 7429 ha (Table SB1).

Direct and indirect effects of infrastructure in maps A, B, and C

Thematic maps for area B show the dominant vegetation, landforms, and surface forms prior to development (Fig. 5). The landform map is overlaid with the major infrastructure-related changes (Fig. 5c). Area B is of special interest because it was the focus of the International Biological Programme Tundra Biome studies and several other scientific studies at Prudhoe Bay (Brown, 1975; Everett & Parkinson, 1977; McKendrick, 1987, 1991; Walker *et al.*, 1987; Walker & Everett, 1991) and contains Pump Station 1, the start of the Trans-Alaska Pipeline System. Thematic maps for all three map areas (A, B, and C) portray surface forms, landforms, dominant vegetation, soils, percent water, total infrastructure-related effects, and total noninfrastructure-related effects (Fig. SC4 SC6).

Time series of maps portraying infrastructure-related changes (Fig. SC7 SC9) showed that the progression of area impacted by direct effects on maps A, B, and C (Fig. 6a) was similar to that in the larger North Slope area (Fig. 4), except that this area was the first to be developed and construction leveled off within 15 years (by 1983 instead of by 1988 as was the case for the regional infrastructure) and declined some afterwards as a few sections of roads were removed and revegetated and some areas of gravel mining in rivers were no longer detectable. The total area of direct effects in 2010 on the three detailed maps was 919 ha (14.6%), ranging from 12.2% in map A to 19.3% in map C (Table SC5). Gravel pads covered the most area (438 ha), followed by excavations (257 ha), roads (136 ha), and pipelines (79 ha) (Fig. 6a and Table SC5).

The total area of indirect effects of oilfield development exceeded the direct effects by 1977, and showed an almost linear rate of increase of about 23 ha year^{-1} in the most recent twenty years (1990 2010), resulting in a total of 1794 ha (28.6% of the area of the three maps), about double the area of the direct effects (919 ha) (Fig. 6a and Table SC5). By 2010, indirect effects included 701 ha of flooding, 367 ha of infrastructure-related thermokarst, 332 ha of gravel and

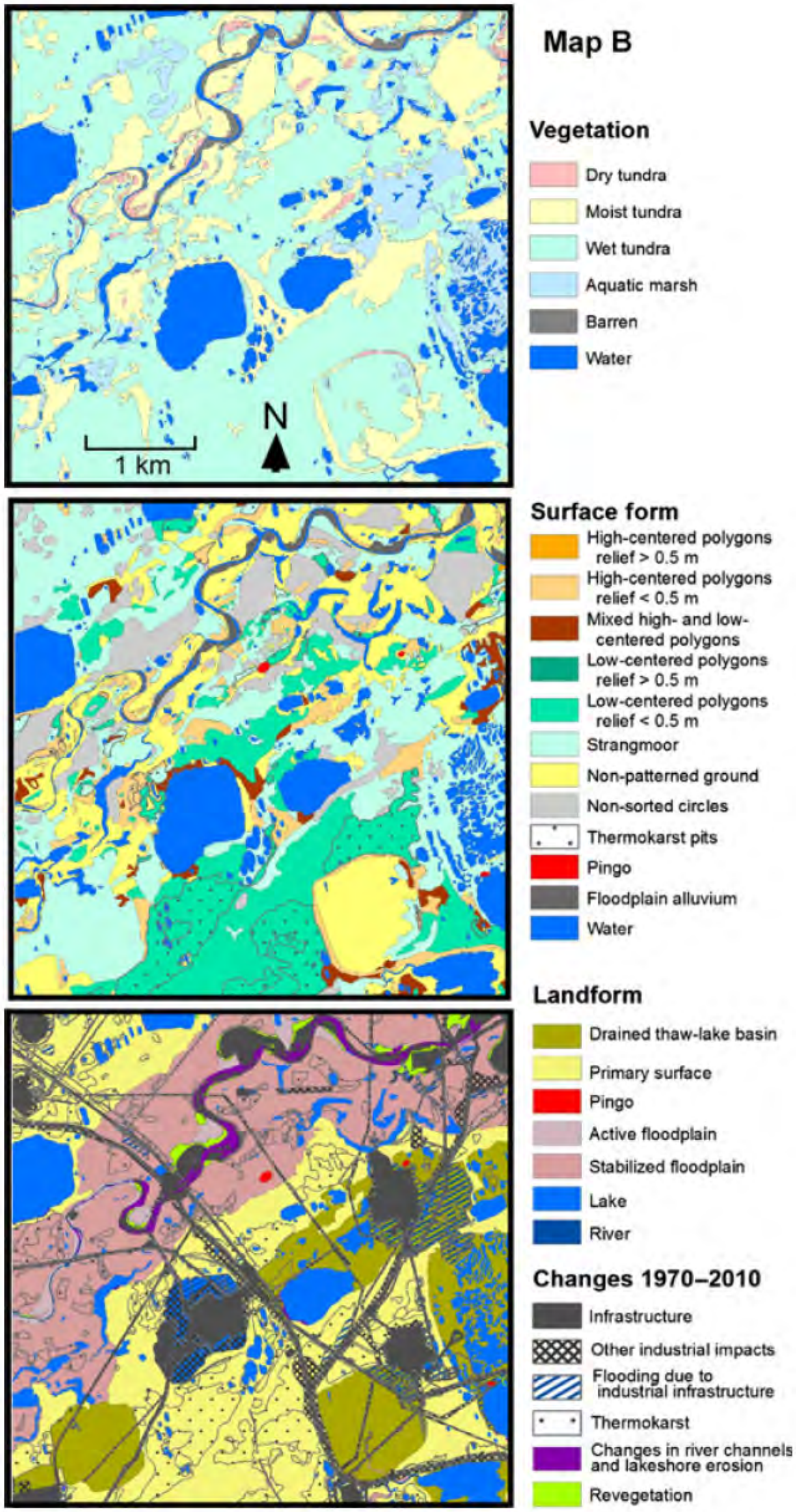

Fig. 5 Dominant vegetation types, surface forms, and landforms with mapped changes (1968 2010) for 20 km^2 area of the Prudhoe Bay oilfield (map B on Fig. 1). See Fig. SC4 S C6 for full sets of thematic maps for maps A, B, and C.

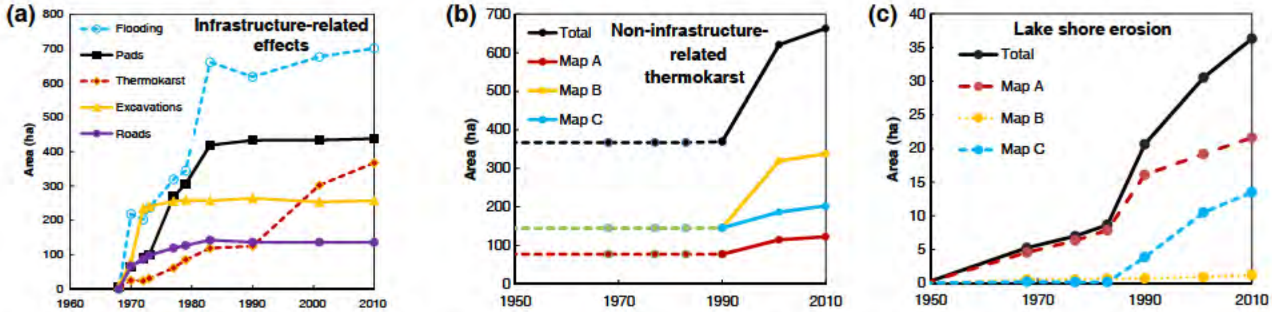

Fig. 6 History of changes (1949 2010) in three 20 km^2 mapped areas within Prudhoe Bay Oilfield, North Slope, Alaska: (a) history of most common infrastructure related effects direct effects (solid lines) and indirect effects (dashed lines); (b) history of noninfrastructure related thermokarst and (c) history of lake shore erosion for maps A, B, and C and for the total mapped area (60 km^2). See Fig. SC7 S C9 for maps of these changes.

debris adjacent to roads and pads, 291 ha of off-road vehicle tracks, and 34 ha of road dust (Table SC5). The extent of road dust and vehicle trails is underrepresented in the data because they were difficult to detect at the scale of mapping. Furthermore, these factors were most often mapped as secondary or tertiary effects in areas where flooding and thermokarst occurred and thus do not show up as the dominant factors, which are summarized here.

Flooding developed quickly as roads and pipelines spread across the flat PBO landscape and dammed the flow of runoff waters during the spring melt season. The extent of flooding nearly leveled off after 1983 at over 650 ha of maps A, B, and C (Fig. 6a). By 2010, the area of infrastructure-related flooding ranged from 25.1% of map A (an extremely flat area with many thaw lakes) to 2.0% of map C (a relatively well-drained area close to the Sagavanirktok River).

Thermokarst, the second most extensive indirect effect, began developing in roadside areas soon after the roads were built (Fig. 2). Visible on the aerial photographs as increased standing water in polygon troughs and subsidence of polygon edges, thermokarst initially covered rather small areas, but expanded at a linear rate over the entire history of the field (9.2 ha year^{-1}, r^2 0.96). After 42 years, within maps A, B, and C, 367 ha of tundra near infrastructure had thermokarst (Fig. 6a). Road dust in high concentrations adjacent to the more heavily traveled roads kills much of the vegetation (Fig. 2a), especially the low-growing mosses and lichens, decreasing the insulative value of the vegetation, and greatly increasing the active-layer thickness (ALTs) and susceptibility of the tundra to thermokarst (Walker & Everett, 1987). In winter, snow drifts develop along both sides of the elevated road berms, resulting in warmer winter soil temperatures near the roads, which increases ALTs and thermokarst in areas adjacent to infrastructure. Road dust in the snow reduces its albedo and leads to early snow melt next to the roads (Benson *et al.*, 1975), increased roadside flooding, and deeper ALTs (Walker & Everett, 1987).

Changes not related to infrastructure

Numerous changes that cannot be attributed to oilfield development also occurred (Fig. 6b, c), including thermokarst far from facilities, erosion of lake shorelines, and erosion, deposition, and revegetation of river bars and banks along the Putuligayuk and Sagavanirktok Rivers. The total area affected by these changes by 2010 was 687 ha (11% of the three mapped areas, Table SC5). While erosional and depositional changes in the rivers mostly compensated for each other, trends for thermokarst (Fig. 6b) and lakeshore erosion (Fig. 6c) were unidirectional.

Lakeshore erosion totaled 36 ha, 0.6% of the mapped area by 2010 (Fig. 6c). Lakeshore erosion showed an abrupt increase in recent years, as was seen with noninfrastructure-related thermokarst. There was a slow, steady increase in lakeshore erosion between 1949 and 1983, reaching 8.6 ha on maps A, B, and C; then to 36 ha from 1983 to 2010, more than a fourfold increase in 27 years (Fig 6c).

The most extensive noninfrastructure-related change was thermokarst, visible as increased standing water in the troughs of ice-wedge polygons (Figs 2 and 3), which covered 503 ha by 2010, 8% of the three mapped areas (Table SC5 and Fig. SC7h S-C9h). Thermokarst in areas distant from infrastructure increased from 367 ha in 1968 to 663 ha in 2010 (Fig. 6b). The area of thermokarst did not noticeably increase between the late 1960s and 1990 on any of the maps (e.g. Fig. 3), but increased 1.8-fold between 1990 and 2010.

By 2010, noninfrastructure-related thermokarst occurred on 19.1% of the areas unaffected by development where thermokarst potentially could occur (i.e., excluding lakes and active floodplains). The landforms most affected by noninfrastructure-related thermokarst were primary surfaces between thaw lakes (residual surfaces not affected by thaw-lake processes), and stabilized river floodplains (31.3% and 16.2%, respectively, Fig. 7, Table SC6). Drained thaw lakes showed little increase in thermokarst features. Surface forms showed a wide range of responses. Low-centered polygons with <0.5 m elevation contrast between the center and the rim, and areas with mixed high- and low-centered polygons were most affected by thermokarst (46.1% and 45.7%, respectively), whereas only 26.0% of high-centered polygons with <0.5 m center-trough contrast had enhanced thermokarst (Fig. 7). Of the area without industrial effects and mapped as having thermokarst pits in 1968, 54% showed increased thermokarst by 2010. Wet tundra and moist tundra were the vegetation types showing the most nonindustrial thermokarst (23.9% and 16.7%, respectively); no other vegetation type had over 5% affected by increased thermokarst. Wet, patterned-ground soil associations showed the most increase in thermokarst (37.6%, Fig. 7 and Table SC6). The total area where increased thermokarst was detected, including infrastructure-related and non-infrastructure-related thermokarst was 870 ha (13.9% of the mapped area).

Discussion

Relationship of thermokarst to air and soil temperatures, ALT, and precipitation

The documented regional increase in thermokarst is most likely due to a long-term upward trend in summer temperatures and to the exceptionally warm summers of 1989 and 1998 (Fig. 8). Summer air temperatures as indicated by the summer warmth index (SWI sum of monthly mean air temperatures above 0 °C, or thawing degree months [°C mo]) increased about 5 °C mo over the 1970 2012 period of record at the Deadhorse airport (Fig. 8). Similar trends are seen in the long-term records from Barrow and Umiat (Jia *et al.*, 2003). The highest recorded SWI at Deadhorse occurred in 1989 and 2012 (30.7 °C mo, 30.5 °C mo), and the third highest was in 1998 (27.5 °C mo). The average SWI for the period 1970 to 1999 was 19 °C mo. The 1989 and 1998 SWI peaks at Deadhorse are similar to the magnitude of thawing-degree-day peaks at the Kuparuk Airport, 100 km west of the study area (Jorgenson *et al.*, 2006).

The mean annual air temperature (MAAT), mean annual permafrost temperature at the upper surface of the permafrost ($MAPT_S$), and at 20-m depth ($MAPT_{20}$),

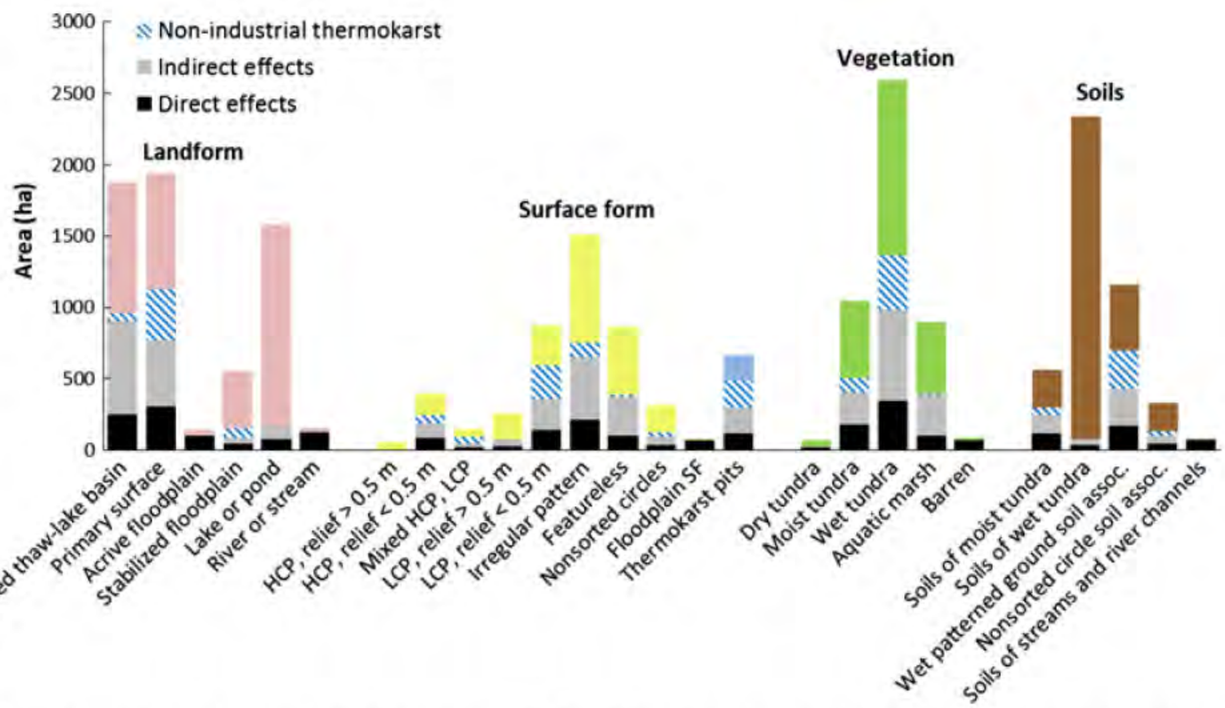

Fig. 7 Total areas of geoecological units >50 ha within maps A, B, and C, area of each unit with direct (black) and indirect (gray) indus trial impacts, with nonindustrial thermokarst (blue diagonal stripes) and remaining unchanged portion (includes area of floodplain gravel erosion and deposition). Thermokarst pits include areas mapped with this code as dominant or secondary surface form. HCP, high centered polygons; LCP, low centered polygons; SF, surface form. See Supporting Information Table C6 for tabular summary.

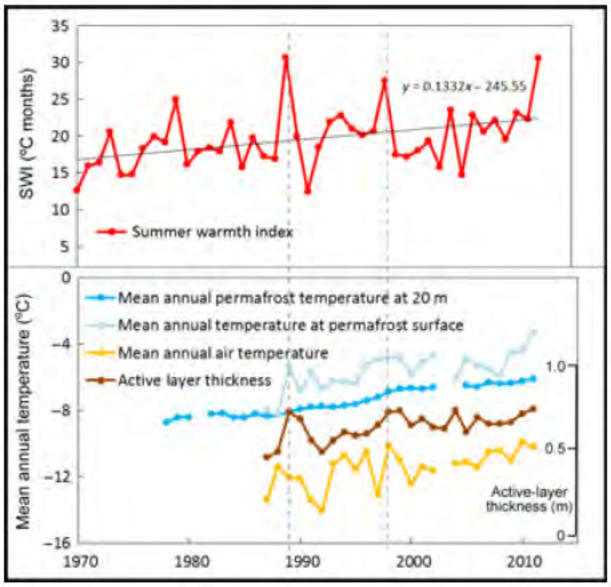

Fig. 8 *Top*: Trend of summer air temperature as indicated by the summer warmth index (SWI Sum of the monthly mean temperatures above freezing, °C mo). Data are from the Western Regional Climate Center, Prudhoe (1970 1986) and Deadhorse (1987 2012). *Bottom:* Mean annual air temperature at 2 m height (MAAT, orange line), mean annual permafrost temperature at 20 m depth ($MAPT_{20}$, dark blue line), mean annual temperature at the upper surface of permafrost ($MAPT_s$, light blue line); and active layer thickness (ALT, brown line and scale on right). All data are from Osterkamp and Romanovsky Deadhorse station. The climate station and maps (A, B, and C) are within 15 km of each other. Active layer depth was measured by interpolation of soil temperature data from several depths for 1987 1996 and by using a metal probe for 1997 2011. Note the corresponding peaks in SWI, ALT, and $MAPT_s$ in the extreme warm summers of 1989 and 1998 (gray dashed lines). Trend lines are as follows: SWI 0.1332 year − 245.55; $MAPT_{20}$ 0.082 year − 170.9; MA PT_s 0.1089 year − 223.37; MAAT 0.0918 year − 194.92; AL T 0.0665 year − 141.85.

and ALT all increased markedly (Fig. 8). The period of most rapid increase in all these variables was during the 1990s. The rate of increase declined from 2001 to 2010, corresponding to a period of moderate air temperatures. Records of ALT from CALM grids at West Dock and Deadhorse also showed maximum thaw depths in 1998, followed by less deep summer thaws in the following decade (Shiklomanov *et al.*, 2012).

Increases in ALT can trigger thawing of ground ice in the upper permafrost, including the top parts of the ice wedges, which can result in the formation of water-filled thermokarst troughs above ice wedges. However, the expansion of thermokarst pits into a network of flooded ice-wedge-polygon troughs as documented in this study may also be related to thermal feedbacks associated with flooding of the troughs. We find no indication that the increase in surface water visible in the ice-wedge polygon troughs in the recent aerial photographs from 1990, 2001, and 2010 was due to higher water tables or to exceptionally wet years. No change in lake levels was detected on the aerial photographs. Furthermore, precipitation records over the 1991 2000 and 2001 2010 period averaged 10.6 and 9.9 cm year^{-1}, respectively, with values for the years of the aerial photography at 7.4 cm in 1990, 7.4 cm in 2001, and 3.9 cm in 2010 (WRCC, 2012). Modeled Prudhoe Bay Region ALTs for 1992 2000 showed an exceptionally deep active layer in 1998, and a trend of subsidence of the ground surface caused by the melting of soil ice (Liu *et al.*, 2012). The similarity of thermokarst patterns in the PBO with those detected west of the oilfield (Jorgenson *et al.*, 2006) indicates that regional climate change is the most likely cause of thermokarst.

Relationship of thermokarst to mapped geoecological features

The occurrence of thermokarst is not uniformly distributed across the landscape and was to some extent predictable based on the earlier geoecological mapping. Areas mapped on the 1968 imagery as 'primary surfaces', 'low-centered ice-wedge polygons', 'wet sedge, moss tundra', 'wet patterned-ground soil association', and 'thermokarst pits' were the most susceptible to further thermokarst. Most of the thermokarst is occurring on surfaces that have not experienced recent lake drainage or reworking of riparian sediments, and are old enough to have accumulated large volumes of excess ice in the form of ice wedges and segregated ice. High-centered ice-wedge polygons experienced less change than low-centered polygons, most likely because these areas have experienced previous thawing and subsidence of ice wedges. Accumulation of organic and mineral matter in the troughs probably helped to protect these ice-wedges from additional thawing (Jorgenson *et al.*, 2006), making these areas less susceptible to the present-day climate change.

A planning process for siting new infrastructure that recognizes the susceptibility of different landforms, vegetation types, and soils to flooding and thermokarst would help to minimize these effects.

A conceptual model of infrastructure-related and climate-change-related thermokarst

Thermokarst development is a complex process that includes numerous positive and negative feedbacks. Thermokarst in areas of IRP with near-surface ice wedges can follow two distinctly different scenarios (Fig. 9). Both scenarios start with partial thawing of the upper ice wedges and formation of small shallow ponds (thermokarst pits) in the troughs over ice wedges and especially at ice-wedge intersections. This initial stage of thermokarst development is triggered by an increase in the ALTs caused by higher than normal air temperatures, flooding, or destruction of vegetation.

The first (stable or reversible) scenario (Fig. 9, blue arrows) is often observed in a natural environment and is described in part by Jorgenson *et al.* (2006). This scenario is possible when the ice wedges are affected by thermokarst, while the polygon centers remain relatively stable because of an undisturbed insulative mat of vegetation and soils. An increase in the ALTs in the polygon centers usually results only in moderate surface subsidence due to partial thawing. Over time, mineral and organic matter accumulate in the troughs from erosion of the trough margins and increased plant productivity due to a combination of deeper thaw,

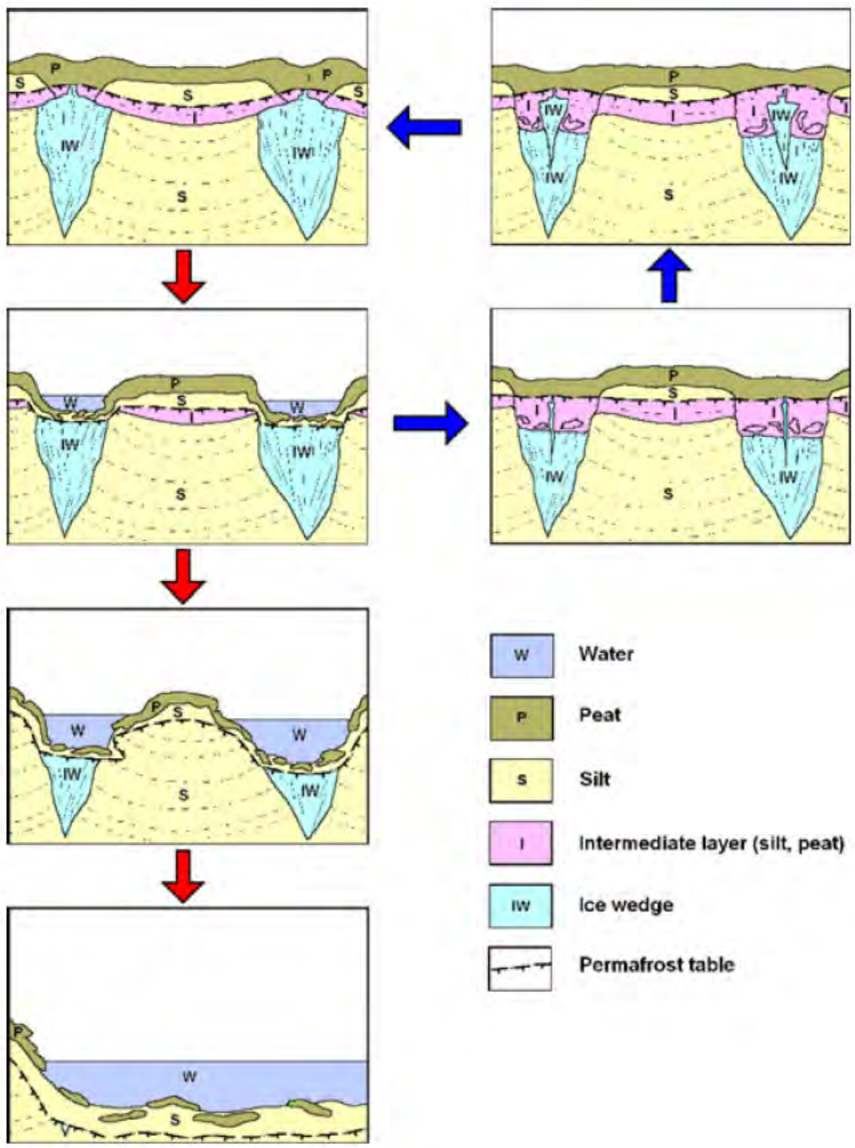

Fig. 9 Two possible thermokarst scenarios associated with ice wedge polygon terrains. A stable or reversible process (blue arrows) is often observed in natural environments. The centers of the polygons remain stable because the protective insulative mat of vegetation and organic soils remains undisturbed. An unstable or irreversible pathway (red arrows) leads to larger water bodies and lakes if the central parts of the polygons experience thaw settlement. This can occur when the thermal insulating properties of the vegetation and organic soils are reduced due to disturbance, accumulation of road dust, or infrastructure related flooding, resulting in increases in the thickness of the active layer.

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release of nutrients and wetter soils. The mineral and organic matter insulates the trough bottoms and slows the thawing of ice wedges, eventually stabilizing the landscape in a new configuration with higher centers and deeper troughs. In areas with cold climates, a new generation of ice wedges may start to form. These wedges penetrate into the previous generation of wedges that were truncated by thermokarst (Fig. 9, two right boxes with blue arrows). If the climate remains favorable for ice wedge growth, the melted wedge-ice eventually reforms and returns the landscape to its original condition.

The second (unstable or irreversible) scenario (Fig. 9, red arrows) usually occurs where there is disturbance to the central part of the polygons, as often occurs in areas in close proximity to infrastructure. This scenario is more severe because the reduction in the protective organic layer can lead to thawing of the upper permafrost in the polygon center, rapid erosion of the edges of the polygon, and subsidence of the entire polygon. Further thermokarst development results in continuing ground subsidence, and to the formation of a shallow thermokarst pond above the polygons. This leads to accelerated thermokarst and relatively fast degradation of ice-rich soils under the pond.

Consequences to ecological systems

The areas that have been affected by extensive thermokarst in the PBO exhibit major ecological changes. No detailed plot-based studies of vegetation changes related to the transformed landscapes were available for this report, but a photographic survey of the roadside areas in map B in summer 2013 showed that numerous areas that were previously low-centered polygons as late as 1983 have been converted to well-drained high-centered polygons. The redistribution of water on the tundra surface has changed the plant communities. The vegetation in the centers of the polygons has been converted from a wet sedge, moss tundra to either a moist sedge, prostrate dwarf-shrub, moss tundra or, in more extreme situations particularly near heavily traveled roads where there has been continuous input of dust for the past 40 years, to a dry prostrate dwarf-shrub, grass, forb tundra (Fig. 2) (Walker, 1985). Many low-centered polygon troughs that previously had wet sedge, moss tundra are now ponds with up to a meter of water with either no vegetation or with aquatic sedges.

The thermokarst terrain is more topographically complex than the initial condition. The changes in hydrology and vegetation undoubtedly affect the distribution and abundance of a wide variety of organisms including insects, shorebirds, waterfowl, small mammals such as voles and lemmings , and could in turn affect the patterns of use by prey species (Batzli & Jung, 1980; Brown *et al.*, 2007). Increases in surface water may increase the habitat of some waterfowl species harvested by village residents of the region (Ward *et al.*, 2005). The implications of thermokarst on the habitat and distribution of fish species, such as the three-spined stickleback (*Gasterosteus aculeatus*) and broad whitefish (*Coregonus nasus*), are not well-understood, nor are the overall effects on lakes and outputs to streams and rivers. Surface water distribution and extent, and connectivity among water bodies influence fish abundance by affecting their access to seasonally important overwintering, spawning, and rearing habitats (M. Wipfli, personal communications). The altered hydrology associated with widespread thermokarst formation also has implications for tundra CO_2 and methane exchanges (Schuur *et al.*, 2009; Sturtevant *et al.*, 2012).

Consequences to engineered and social systems

The negative consequences of thermokarst to infrastructure are well-known and extensively documented (US Arctic Research Commission Permafrost Task Force, 2003; Streletskiy *et al.*, 2012). The additional maintenance and replacement costs due to climate change on public infrastructure in Alaska is estimated at $3.6 6.1 billion through 2030, but it is difficult to isolate the projected costs of thermokarst from other climate change effects (Larsen *et al.*, 2008). Within the PBO, thermokarst affects rehabilitation efforts at sites where gravel has been removed (e.g. former gravel pad or road) and trenches where cables and pipelines are buried, resulting in subsidence that greatly exacerbates the cost of rehabilitation (Streever, 2012).

Situations similar to those in the PBO also occur in villages. High-resolution satellite images show that thermokarst has become extensive within the village road network at Nuiqsut, west of the PBO. Deeper thermokarst (up to 2 m) occurs in the sandy eolian deposits west of the Colville River (Lawson *et al.*, 1978), and much deeper thermokarst (up to 5.5 m) occurs in the thick, silty, organic-rich, and extremely ice-rich *yedoma* deposits of the northern Arctic Foothills (Lawson, 1983; Carter, 1988; Kanevskiy *et al.*, 2011; Shur *et al.*, 2012). Apart from current rough projections of costs related to relocation of Alaskan villages facing erosion problems, we are unaware of any cost estimates for private or industrial infrastructure on the North Slope related to thawing permafrost. Future effects are difficult to predict especially when combined with simultaneous rapid changes to climate, political, and socio-economic factors, and oil drilling technology.

Value of long-term studies of a rapidly changing ecosystem

When the PBO studies began in the 1970s, none of the now-senior authors who were involved foresaw the possibility of the rapid transitions that are occurring now. For over 20 years, the areas that were not affected by oilfield infrastructure showed little change. Based on the mapped information and current air and permafrost temperature trends, starting in 1990 we are witnessing landscape changes that will have major implications for much of the Arctic Coastal Plain. The conceptual model of thermokarst formation presented here and the description of the characteristics of areas most vulnerable to thermokarst will help in the development of predictive models of how thermokarst spreads in different climate-change and infrastructure scenarios.

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Supporting Information

Additional Supporting Information may be found in the online version of this article:

Appendix S1. Ice rich permafrost at Prudhoe Bay. By Y. Shur, M. Kanevskiy, V.E. Romanovsky, K.R. Everett, J. Brown, and D.A. Walker.

Appendix S2. Calculation of impacts of oilfield development, North Slope Alaska, by K.J. Ambrosius.

Appendix S3. Integrated Geoecological and Historical Change Mapping: history, methods, maps, and summary information, by D.A. Walker, M.K. Raynolds, P.J. Webber and J. Brown.

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Environmental Change and Potential Impacts: Applied Research Priorities for Alaska's North Slope

by B. Streever, R. Suydam, J.F. Payne, R. Shuchman, R.P. Angliss, G. Balogh, J. Brown, J. Grunblatt, S. Guyer, D.L. Kane, J.J. Kelley, G. Kofinas, D.R. Lassuy, W. Loya, P. Martin, S.E. Moore, W.S. Pegau, C. Rea, D.J. Reed, T. Sformo, M. Sturm, J.J. Taylor, T. Viavant, D. Williams and D. Yokel

INTRODUCTION

The North Slope of Alaska is the vast area north of the crest of the Brooks Range (Fig. 1). Its land base encompasses 231 000 km² (89 000 mi²), an area roughly the size of Minnesota, most of which is wetland habitat underlain by permafrost and part of which contains the largest operating oil fields in the United States. The nearshore and offshore waters of the Chukchi and Beaufort seas add another 295 000 km² (114 000 mi²) and hold what may be the largest undeveloped oil reserves remaining in the United States. The region is home to an abundant and diverse array of fish, wildlife, and plants, resources that support the vibrant subsistence culture of about 6000 Iñupiat Eskimos. The caribou herds that summer on the North Slope are an important food resource for Iñupiat communities, as are some native plants, bowhead whales, beluga whales, four species of ice seals, and walruses living in the Beaufort and Chukchi seas. Further, Alaska's North Slope is at the forefront of global climate change, with an increase in mean annual temperature of about 1˚C per decade in Barrow, Alaska (ACRC, 2008).

Federal, state, and local agencies manage the biotic and abiotic resources of the North Slope to maintain fish and wildlife populations and their habitats while also allowing energy development. The laws and regulations applied by government agencies managing the North Slope are rigorous, complex, and often controversial.

Appropriate management requires information that can be gained only through applied research. We provide a brief history of applied research on the North Slope, introduce the North Slope Science Initiative (NSSI) as an organization tasked with improving the coordination of science across the region, and posit applied science priorities that are essential for successful and informed management.

HISTORY OF NORTH SLOPE APPLIED SCIENCE

The earliest attempts to understand the North Slope region undoubtedly occurred when Iñupiat people and their predecessors shared information immediately relevant to their survival (Chance, 1990). Much later, a tradition of science grew from work initially undertaken during mapping expeditions, such as that of Rochfort Maguire and Dr. John Simpson during their sojourn near Barrow from 1852 to 1854 (Maguire, 1988). This tradition grew during the First International Polar Year in 1882–83 and through Diamond Jenness's anthropological studies during the 1913 Karluk expedition (Jenness, 1957). The presence of oil seeps led to establishment of the Naval Petroleum Reserve Number 4 in 1923. Exploratory drilling for oil and gas in the Alaskan Arctic started during World War II, and the Office of Naval Research established what would eventually become the Naval Arctic Research Laboratory (NARL) in Barrow in 1947 (Reed, 1958; Norton, 2001a). By 1948, a Scientific Advisory Board had been established for NARL, and nine research projects were underway, including work sponsored by multiple government agencies (Schindler, 2001). Over the decades, Barrow became a center for research activity, including ice island research, field studies across the North Slope, and the establishment in 1970 of the International Biological Programme's Tundra Biome project funded by the National Science Foundation (NSF). Following transfer of NARL to the Ukpeagvik Iñupiat Corporation, the Barrow Environmental Observatory (BEO) was established and subsequently zoned as a scientific research district. The Barrow Arctic Science Consortium was established in 1995 to promote science in the region, integrate scientists with the local community, and assist with management of the BEO.

To the southeast, adjacent to the TransAlaska Pipeline, in 1975 the National Science Foundation and the University of Alaska established the Toolik Field Station, which has hosted an Arctic Tundra Long-Term Ecological Research Program for freshwater and terrestrial field studies since the late 1980s. To the east of the pipeline lies the Arctic National Wildlife Refuge, where wildlife and wilderness studies began in the 1950s and continue into the present.

The discovery of economically recoverable oil in 1968 about 240 km (150 mi) east of Barrow and the subsequent development of oilfields spawned efforts to collect baseline data and to assess environmental impacts. Research

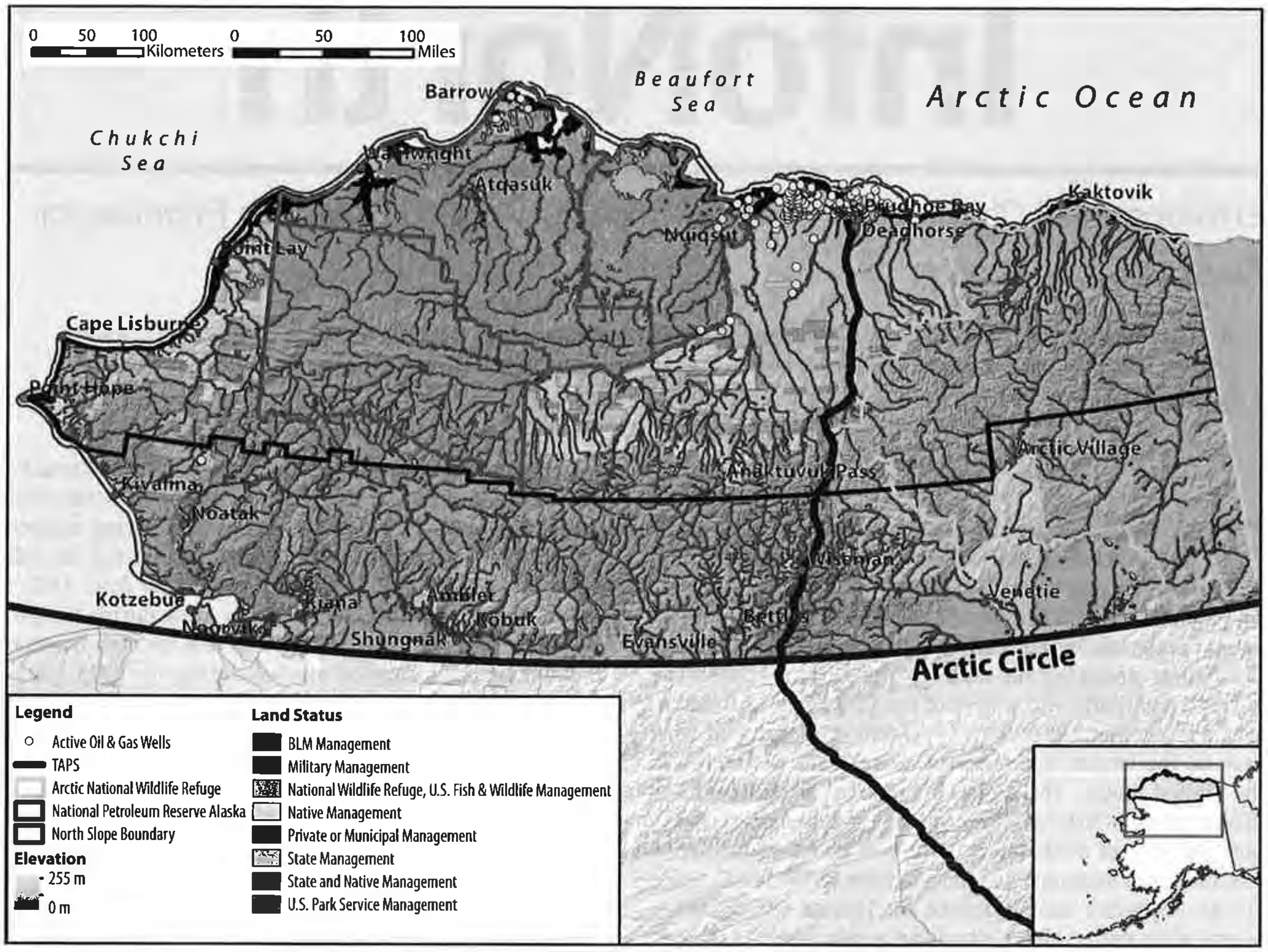

FIG. 1. The North Slope of Alaska.

was supported by various government agencies, private companies, and nonprofit organizations, but communication between these groups was often limited. Nevertheless, attempts to integrate science across disciplines occurred. For example, in the 1970s the U.S. Department of the Interior and Department of Commerce collaborated on the Outer Continental Shelf Environmental Assessment Program, an integrated marine and coastal field research program relevant to management needs (NOAA, 1978). Over time, this work evolved into the ongoing Environmental Studies Program of the Bureau of Ocean Energy Management, Enforcement, and Regulation (formerly the Minerals Management Service) and led to the production of multiple long-term data sets in coastal oceanography, biology, and social systems.

A number of book-length reviews have summarized work relevant to managers (e.g., Truett and Johnson, 2000; Norton, 2001b; NRC, 2003). These publications and other efforts promoted the benefits of an integrated, cross-disciplinary approach to science in terrestrial and marine environments.

In the last decade, several efforts to enhance coordination of applied, management-oriented Arctic science were initiated. In 2004, the Alaska Ocean Observing System began with a mission of improving the ability to detect change in marine ecosystems. In 2009, the Department of the Interior initiated an Arctic Landscape Conservation Cooperative as well as an Alaska Climate Science Center. In 2010, the National Oceanic and Atmospheric Administration proposed a National Climate Service that will include an Arctic section. Additional programs include the interagency Study of Environmental Arctic Change and the National Science Foundation's Arctic Observing Network.

All of these groups are tasked, to some degree, with fostering cooperative and intergovernmental approaches to the scientific understanding of North Slope ecosystems. The role of each of the current efforts is not clearly delineated, but the groups are working together and attempting to share data and information tracking systems, as well as striving to limit duplication of effort and to advance relevant science in the interest of best management practices.

THE NORTH SLOPE SCIENCE INITIATIVE AND THE ISSUE PAPERS

Recognizing the need for enhanced coordination of applied science, federal, state, and local governments collectively formed the North Slope Science Initiative (NSSI) in 2001. The NSSI was formally authorized under the Energy Policy Act of 2005 (Section 348), with a broad legislative mandate to implement efforts to coordinate applied science needs relevant to resource managers on the North Slope. Its membership comprises 14 management entities (see Appendix).

The organizational structure of the NSSI allows for direct interaction between an oversight group staffed by high-level agency executives, an internal advisory group staffed by experienced agency personnel, and an external advisory group staffed by Iñupiat elders and scientists from universities, nonprofit organizations, and industry. This external advisory group, called the Science Technical Advisory Panel (STAP), is a 15-member committee established under the Federal Advisory Committee Act, making it independent of direct agency supervision. The NSSI issues an annual report to Congress through the Department of the Interior (www.northslope.org).

Soon after its formation, the NSSI asked the STAP to summarize issues important to North Slope management (Table 1). Broad topics were identified by agency executives, and questions and specific issues related to each broad topic were developed by experienced agency regulators and scientists. Working through an iterative process that combined input from agencies with information and opinions from external subject-matter experts, the STAP developed the issue papers. The first 13 of these issue papers were released to the public in late 2009 (see http://www.northslope.org/).

PRIORITIES FOR NORTH SLOPE APPLIED RESEARCH

Priorities for Individual Issues

Each of the issue papers provided recommendations for future applied research likely to be relevant to managers, but the papers were written independently of one another. After reviewing the issue papers, the NSSI Oversight Group asked the STAP to develop a prioritized list for future applied research and to assess how various issues might be related to one another.

As an initial step toward prioritizing applied research, the STAP collectively and by consensus assigned each issue (with the exception of "weather and climate," which was addressed separately) to one of three "state of knowledge" categories:

- issues that are reasonably well understood and for which research is sufficient to address most current management questions;
- issues that are less well understood and require additional research and monitoring to address management questions; and
- issues that are poorly understood and require substantial additional research and monitoring to address management questions.

Importantly, most of the issues are interdependent to some degree. For example, changes in active layer thickness above permafrost will likely result in changes to hydrology, which in turn will affect vegetation, and through vegetation, caribou and some bird populations. A conceptual model was developed displaying these issues and their interconnectedness (Fig. 2).

In addition, the amount of time needed to generate meaningful results was estimated. For example, meaningful results from restoration experiments, permafrost studies, and assessment of vegetation change will require at least 10 years because of the slow growth of plants and the slow response of permafrost. To consider "time to meaningful results," the STAP collectively and by consensus estimated the number of years (in 5-year increments to a maximum of 20 years) likely needed to move a topic from "requires substantial additional research" to "requires additional research," or from "requires additional research" to "research is sufficient" (Fig. 2).

Because some forms of research are much more expensive than others, "state of knowledge" and "time to meaningful results" categorizations should not be interpreted as suggesting funding levels. For example, "vegetation change" and "migratory birds" were both categorized as requiring additional research, but at least some aspects of vegetation change can be studied using remote sensing techniques with limited field validation, whereas migratory bird studies require substantial field efforts. In addition, a categorization of "research is sufficient" was not meant to justify a reduction in funding. Even the best researched issues require an ongoing investment in monitoring. After considering information on relationships between issues, the state of knowledge for each issue, and the estimates of time to meaningful results, the STAP prioritized the top three most pressing applied research topics for each issue (Table 1).

Overarching Priorities

Throughout development of the issue papers and during prioritization of applied research topics, five broadly applicable overarching priorities emerged: (1) systematic assessment of the range of potential development scenarios for 20 years into the future in a manner that will contribute to refinement of specific research priorities; (2) systematic assessment of the range of potential climate scenarios for 20 years into the future in a manner that will contribute to refinement of specific research priorities; (3) enhanced and well-organized collection of climate and weather data across the North Slope in a manner that will facilitate

TABLE 1. The North Slope Science Initiative issues and associated "top three" applied science priorities identified by the Science and Technical Advisory Panel.

Issue	Priority 1	Priority 2	Priority 3
Weather and climate	• Inventory and assess existing meteorological stations and perform gap analysis	• Pool resources from multiple funding entities to install and maintain new stations to fill gaps	• Develop a database that integrates output from stations with existing national archives
Changing sea ice conditions	• Collect sea ice data at spatial and temporal scales relevant to users and modelers	• Study the fate and effects of oil spills	• Study oil spill response in broken ice conditions
Coastal salinization	• Investigate the effect of increased salinity on vegetation	• Develop models of coastal salinization	• Understand the impact on tundra of ice roads built with saline water
Coastal and riverine erosion	• Inventory and make broadly available all coastal imagery	• Generate accurate and ground-truthed baseline maps for selected areas	• Instrument the coastline with wind and wave sensors
Increasing marine activity	• Understand future scenarios of marine activities	• Develop standard methods of impact assessment, especially underwater sound measurement methods	• Increase broad availability of existing data
Fire regime	• Monitor recovery following tundra fires	• Complete land-cover mapping to facilitate understanding of change	• Evaluate fire return intervals
Contaminants	• Monitor levels to detect change in air, water, soil, and biota	• Evaluate toxicity levels and monitor contamination in subsistence resources	• Improve understanding of fate and effects, especially from discharges to broken ice
Hydrology and lake drying	• Develop a stream gauge network complemented by meteorological stations	• Develop remote-sensing technologies to facilitate mapping	• Use local knowledge in planning and assessment studies
Permafrost (including active layer)	• Increase permafrost monitoring on representative landscapes	• Develop remote-sensing technologies to facilitate mapping	• Inventory existing data and improve its availability
Vegetation change	• Expand monitoring for vegetation change	• Inventory and evaluate existing vegetation plot data	• Complete the North Slope land cover map
Caribou	• Inventory data, improve availability, and improve coordination of future data collection	• Develop understanding of seasonal range use and harvests (subsistence and sport)	• Improve communications between researchers, managers, and stakeholders
Migratory birds	• Improve monitoring before, during, and after development	• Inventory key data and improve availability	• Improve understanding of impacts from spills, especially in broken ice and ice leads
Marine mammals and their prey	• Increase knowledge of marine mammals, their prey, habitat use, impacts, and harvest, with emphasis on listed species	• Increase long-term studies that integrate information on marine mammals, their prey, and the environment	• Understand the cumulative effects from human activities, including underwater sound
Ecological restoration	• Develop a systematic long-term research program, recognizing time needed to obtain results	• Develop seeding methods using sedges commonly found on the North Slope	• Develop a clear understanding of rehabilitation trajectory during at least 20 years of growth
Fisheries	• Develop an understanding of subsistence use in past and present.	• Develop a single, accessible database on local fish abundance and distribution	• Implement long-term studies on fish, their habitat, and their prey capable of differentiating between changes from natural and anthropogenic causes
Social impacts	• Coordinate and review all research involving North Slope residents as human subjects	• Improve methods for inclusion of local and traditional knowledge in monitoring and research of social and ecological systems	• Implement systematic studies of the implications of future oil and gas development activities on North Slope communities livelihoods and well being

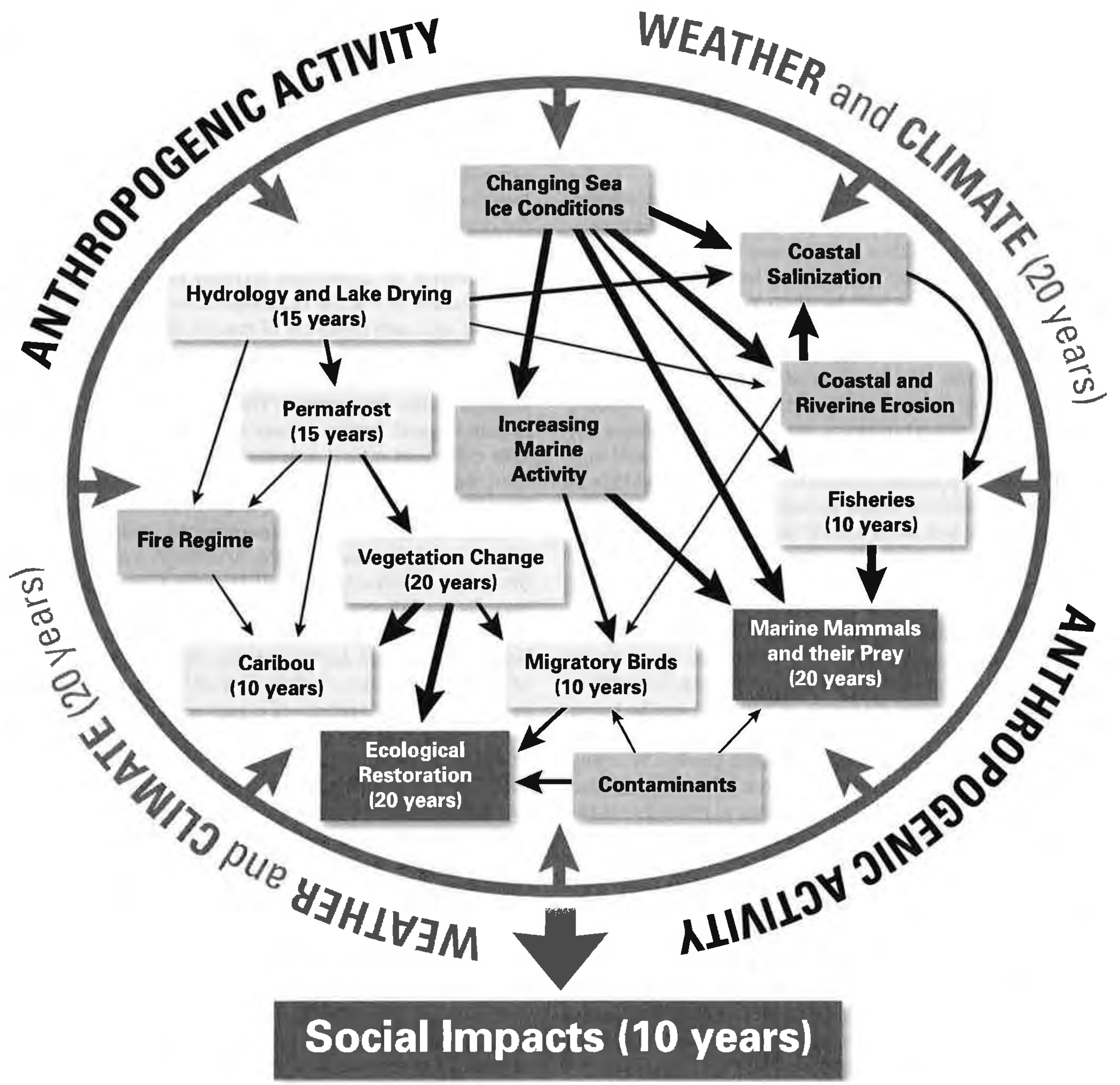

FIG. 2. Sixteen issues or research topics relevant to the management of the North Slope and their influence on one another. All are potentially affected by climate change and anthropogenic activity (i.e., development). Social impacts, which affect both local people and the intrinsic value of intact ecosystems to people well removed from the Arctic, are influenced by all the other issues. The strength of relationships is suggested by the thickness of arrows. Green represents topics for which research is sufficient to satisfy most management questions; yellow topics are less well understood and require additional research support; and topics in red are poorly understood and require substantial additional research. For topics in yellow and red, parentheses show the estimated time needed (assuming reasonable funding support for research) to move a topic up to the next knowledge level.

improved regional climate modeling, verification of climate models, and application of data in research projects; (4) regional coordination of existing long-term monitoring projects; and (5) renewed and systematic efforts to improve communication among managers, residents, and scientists through initiation of frequent "place-based" workshops.

Potential Development Scenarios: An understanding of the estimated size, location, and intensity of plausible development activities in the foreseeable future, defined here as the next 20 years, is important for prioritizing and implementing temporally and spatially appropriate research and monitoring. Because of many uncertainties, projecting future development scenarios will need to encompass a range of possibilities, from the "least" to the "most" new development. Both onshore and offshore development should be considered, and energy development, commercial shipping through ice-free routes, tourism, mining, commercial fishing, road construction, military activities, and other forms of development should be included.

Three realities must be addressed when considering development scenarios. First, because of changing economic conditions and the age of the two largest North Slope oilfields, future scenarios based on linear projections of past development rates will be of no value because future developments will not employ the same designs used in the past or follow the same progression. Second, while no one entity has the expertise needed to responsibly consider development scenarios on its own, by bringing together expertise from the oil industry, the regulatory community, the nonprofit community, the Iñupiat community, and others, it should be possible to consider a range of development scenarios responsibly. Third, an initial projection of a range of development scenarios should not be viewed as a static model; instead, it should be systematically revised every three to five years to optimize its usefulness in the planning of applied research.

Potential Climate Scenarios: While it is clear that the Arctic is warming, there is likely to be fine-scale spatial and temporal variation in this warming pattern that will be important to managing resources or activities on the North Slope (Martin et al., 2009). Setting science priorities properly will require downscaling of climate models in a way that facilitates understanding of potential ecological and physical impacts at various spatial scales of interest to managers (e.g., at the scale of watersheds, not continents) in the next 20 years.

It is not enough to downscale models that produce only average temperatures and precipitation. The spatial and temporal variability in temperature and precipitation, plus likely changes in wind directions and speeds, summer rains, snowpack thickness and water content, timing of freeze-thaw events, erosion, and other dynamic environmental parameters need to be modeled in order to optimize applied research prioritization.

Resources such as the circumpolar Arctic Climate Impact Assessment (ACIA, 2004) and North Slope Specific Wildlife Response to Environmental Arctic Change ("WildREACH") (Martin et al., 2009) have been useful, but climate change science cannot yet offer firm projections at local and sub-regional scales across the North Slope. A systematic review of advances in climate modeling is needed, as well as discussion of how modeling results may provide useful information about potential changes or impacts likely to be experienced by fish, wildlife, and habitats. Such a review should occur every three to five years as a way of ensuring that North Slope applied research provides the most relevant and recent information to resource managers and decision makers.

Climate and Weather Data: Meteorological data collected at adequate spatial and temporal scales are necessary for the development and validation of models underlying climate scenarios. However, the existing meteorological network on the North Slope of Alaska is haphazard at best. Individual stations are operated by myriad groups and agencies, are often short-lived, frequently use dissimilar instrumentation, and are generally at low elevations along the coast (in villages or in oilfields). There is a need to better distribute stations, which will require installation and maintenance of unmanned stations in extreme environments. In addition to the challenge of the harsh environment itself, a successful network of meteorological stations must address the costs of access, provision of power for real-time transmission of data and images, and potential wildlife damage to the instruments. There is also a need to install meteorological stations where they can complement other data collection efforts assessing variables such as stream flow, snowpack conditions, active-layer thickness, permafrost thermal state, gas fluxes, and wildlife movements.

One way forward, as recommended in the NSSI issue paper on weather and climate, is through a staged process involving (a) inventory of all stations currently in place, regardless of their capabilities; (b) assessment of the flexibility of design in existing stations to determine if modifications in design and deployment are possible; (c) canvassing of various end users to define clearly what information is needed; (d) development of a gap analysis to understand exactly what data or information is missing; and (e) pooling of resources to support an integrated network. This process should be overseen by a small working group of data collectors and end users. Because cost has been the main obstacle to development and maintenance of a spatially distributed meteorological network, it is imperative to have the participation, cooperation, and collaboration of all land- and water-management organizations.

Coordination of Long-term Monitoring: Despite the broad availability of a number of long-term monitoring reports, many of the NSSI issue papers recognized the need for additional long-term monitoring. Long-term monitoring—defined here as monitoring that has occurred for at least 10 years and is likely to be continued through the foreseeable future—requires exceptional commitment on the part of funding organizations. Monitoring must account for a highly variable environment, a warming climate, and anthropogenic stressors that affect the rates and pathways through which many components of the Arctic ecosystem interact. The involvement of North Slope communities in ecological monitoring through residents' observations and understanding of change, in partnership with scientists, may provide a useful way to achieve stronger integration and a richer understanding of emergent conditions.

In many cases, two or more monitoring programs assessing the same variables may use different methods that make comparisons difficult or impossible. For instance, plant surveys using quadrats produce different results than plant surveys using line transects, so that comparison of apparently similar summary statistics may be problematic. While adoption of standard protocols may seem beneficial, it does not acknowledge the underlying reasons for different protocols, such as differing research objectives or logistical constraints. Therefore, where possible, the means of comparing results from data collected using different methods should be developed. Similarly, on any one project, protocols may change over time, making comparisons across time difficult or impossible. As an example, changes over time in quadrat size or changes in plant identification skills make assessment of ecological changes difficult. The degree to which methods change over time must be understood, and if necessary, a means of allowing comparisons across time must be developed.

Moving beyond individual variables, the absence of integration hinders understanding of cause and effect. For example, failure to coordinate across topics and across temporal and spatial scales makes it impossible to correlate factors such as rainfall and grazing. Although to date no single report has summarized the key results of long-term monitoring projects from across the North Slope, reports such as Neff (2010) and Douglas et al. (2002) suggest the value of a coordinated effort and the possibility of data integration.

Improving Communication among Managers, Residents, and Scientists: Information relevant to North Slope management agencies is multidisciplinary, and collectively the amount of information available is, by any standard, overwhelming. As a result, it may be tempting for specialists to work within their discipline, in relative isolation from other disciplines. However, a clear need exists for sharing information among disciplines in a way that makes it accessible to resource managers and local residents. Furthermore, successful sharing of information among managers, residents, and scientists requires communication that is dependent on trust relationships across cultural boundaries.

One tool for improved communication could be the broad use of tracking sheets describing proposed and ongoing studies and monitoring projects. Another is a one-stop information exchange, such as the NSSI Data Catalog and Project Tracking System (http://www.northslope.org).

Other approaches are needed to enhance oral communication. The annual Alaska Marine Science Symposium and the Western Arctic Caribou Herd Working Group meeting provide good examples of sharing information, but many other issues could benefit from enhanced information sharing. One approach is initiation of smaller place-based conferences or workshops—that is, events that bring together researchers, managers, and stakeholders with different backgrounds and different areas of expertise to encourage communication across specialties, such as the recent "Science, Natural Resources, and Subsistence in Alaska's Arctic Lands and Waters" meeting held in March 2011 in Barrow, Alaska.

CONCLUDING REMARKS AND A WAY FORWARD

Research is, in part, an entrepreneurial endeavor, with proposals competing for often scarce resources on the basis of intellectual merit. However, research on applied problems—including problems related to management of the North Slope—progresses most rapidly when resources are strategically deployed to enable cooperation, collaboration, and coherent development of relevant information. Research proposals assessed on the basis of carefully considered management needs are most likely to provide results that are of immediate value to managers.

Coordination of research should not be equated with control of research. The role of coordination is to help managers and local residents understand what applied research can realistically offer, to help scientists understand what managers and local residents need, and overall to reduce unwanted or unneeded redundancy while advancing complementary efforts. The suggestions outlined here may seem obvious when laid out in a systematic manner and in the context of the NSSI issue papers. However, the current reality of prioritization and funding of scientific research on the North Slope and the degree to which it is useful to managers and local residents suggest that what seems obvious in retrospect may not be obvious at all. If acted upon, the suggestions proposed here will lead to a step change in the way applied science is done on the North Slope of Alaska and, importantly, will dramatically increase the value of this science.

DEDICATION

This paper is dedicated to the late Iñupiat elders Arnold Brower, Sr., and Warren Matumeak, both past members of the North Slope Science Initiative's Science Technical Advisory Panel.

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APPENDIX: NORTH SLOPE SCIENCE INITIATIVE MEMBER AGENCIES AND ORGANIZATIONS.

- Alaska Department of Fish and Game
- Alaska Department of Natural Resources
- Arctic Research Commission
- Arctic Slope Regional Corporation
- Bureau of Land Management
- Bureau of Ocean Energy, Management, Regulation, and Enforcement (previously Minerals Management Service)
- National Oceanic and Atmospheric Administration, National Marine Fisheries Service
- National Oceanic and Atmospheric Administration, National Weather Service
- National Oceanic and Atmospheric Administration, National Climate Service (proposed)
- National Park Service
- North Slope Borough
- U.S. Department of Energy
- U.S. Fish and Wildlife Service
- U.S. Geological Survey

Bill Streever is with BP Exploration (Alaska) Inc., Anchorage, Alaska; R. Suydam and T. Sformo are with the North Slope Borough Department of Wildlife Management in Barrow, Alaska; J.F. Payne, S. Guyer, and J.J. Taylor are with the North Slope Science Initiative in Anchorage; R. Shuchman is at Michigan Tech Research Institute, Ann Arbor, Michigan; R.P. Angliss and S.E. Moore are at the National Marine Mammal Laboratory, Alaska Fisheries Science Center, National Marine Fisheries Service, Seattle, Washington; Greg Balogh is with the Arctic Landscape Conservation Cooperative, U.S. Fish and Wildlife Service (USFWS), Anchorage; J. Brown lives in Woods Hole, Massachusetts; J. Grunblatt, D.L. Kane, J.J. Kelley, and G. Kofinas are with the University of Alaska – Fairbanks; D.R. Lassuy is with the USFWS, Anchorage; W. Loya is with the Wilderness Society, Anchorage; P. Martin is with the Arctic Landscape Conservation Cooperative, USFWS, Fairbanks; W.S. Pegau is with the Oil Spill Recovery Institute in Cordova, Alaska; C. Rea works for ConocoPhillips, Anchorage; D.J. Reed is with the Alaska Department of Fish and Game (ADFG), Nome, Alaska; M. Sturm is with USA-CRREL-Alaska, Ft. Wainwright, Alaska; T. Viavant is with ADFG in Fairbanks; D. Williams is with the Bureau of Ocean Energy Management, Anchorage; and D. Yokel is with the Bureau of Land Management's Arctic Field Office in Fairbanks. The corresponding author is Bill Streever: Bill.Streever@bp.com.

Science and Emotion, on Ice: The Role of Science on Alaska's North Slope

BILL STREEVER

In a room full of people gathered to consider the cumulative effects of development on Alaska's North Slope, a man captures the entire complex situation in just 40 words. "You have to look at the data closely," he says, "and think about the science, but when you get up to the North Slope, you'll hear those caribou go thundering past, and you'll get this gut feeling that you just can't ignore." There it is, in a nutshell: the juxtaposition of technical information on the one hand, and the unavoidable presence of emotion on the other.

No one on the North Slope—either within the oil industry or outside it—denies the importance of emotion in the decisionmaking process. In talking to technical experts in Fairbanks, Anchorage, and Barrow, and in the oil fields themselves, it seems that the frustration does not stem from inclusion of emotion in the decisionmaking process. Instead, it comes from the indiscriminate mixing of science with emotion and the failure to separate the two. It is a mix that has to some degree polarized the scientific community. It is a mix that has led to an us–them dichotomy and a "with-us-or-against-us" attitude. It is a mix that has ended professional relationships, contributed to early retirements, and, in at least one case, torpedoed an otherwise perfectly viable romance.

Ted Rockwell, who regulates North Slope development for the Environmental Protection Agency, points out that both the personal ownership of ideas and the sense of accomplishment that accompanies the development of ideas can become an emotional issue that clouds objectivity. Lloyd Fanter, a veteran of the US Army Corps of Engineers who works in the environmental regulatory arena, says that some industry representatives become so attached to their ideas about development that they refuse to consider alternatives suggested by agencies and the public. "At some point," he says, "people lose perspective. My job is to bring balance to the process. I work toward environmental integrity that is founded on technical data—data from all sources, including government scientists and industry scientists."

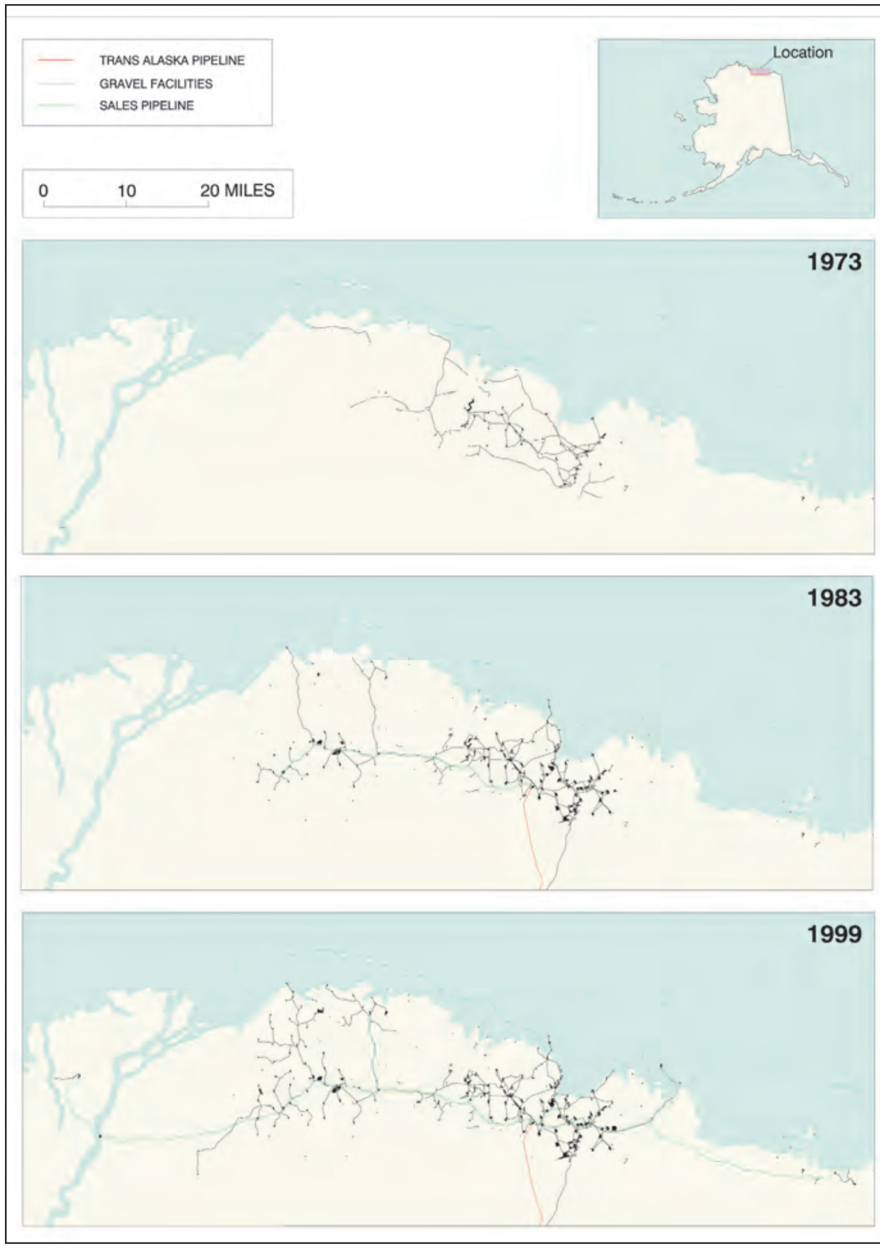
Dave Trudgen has managed British Petroleum's (BP) environmental research program on the North Slope for the past 2 years. He shows me a list of projects, with general categories (lake recharge and restoration) mixed with animals (caribou, polar bears, grizzly bears, ringed seals, marine fish, freshwater fish, eiders, snow geese, arctic fox, shorebirds, and more) (Truett and Johnson 2000). "BP," he tells me, "spends somewhere between \$5 million and \$10 million each year on environmental research." The results of the research contribute to improved environmental stewardship on the North Slope. They are used to guide the placement, design, and operation of oil industry infrastructure. In many cases, the results are also used to respond to concerns raised by nonprofit organizations and government agencies. But often, the emotional baggage that comes with the topic of North Slope development cannot be overcome by research. "The issues that are most driven by emotion," Trudgen says, "are caribou, whales, and what people are starting to call arctic sprawl. And the emotion is on both sides. Some people become attached to ideas and don't want to let them go, or they let ideas become a personality issue. Some people in the industry believe we have solved the environmental problems, and they're sick of spending money on concerns that can never be resolved. Some people who are against industry believe that development is wrong, and they'll look through the data until

Bill Streever has worked in academics, government laboratories, and the private sector and is involved with numerous nonprofit conservation organizations. He is known primarily for his work in wetland restoration, and he has published in many fields, ranging from invertebrate ecology to environmental economics. His latest book, *Saving Louisiana? The Battle for Coastal Wetlands*, was released in October 2001. In December 2000, he began work as British Petroleum's environmental studies leader in Alaska.

they find something supporting that position. Most people are somewhere in between the extremes—leaning one way or the other, but somewhere in between."

Arctic sprawl

Arctic sprawl—a phrase reputedly coined by Bruce Babbitt, former secretary of the interior in the Clinton administration, to describe the increasing size of the North Slope oil fields—cannot be denied. The first real oil strike came in 1968, in Prudhoe Bay, after more than 40 years of exploration. In 1977, the Prudhoe Bay field came on line, delivering oil through the Trans-Alaska Pipeline to Valdez and from there to markets in the lower forty-eight. In 1981, the Kuparuk field came on line, followed by Milne Point, Lisburne, and Endicott in 1985, 1986, and 1987, respectively (BP and Phillips Alaska 2001). By 2001, 18 fields were sending oil south. By and large, the spread of exploration has been to the west, toward Barrow, and almost exclusively on state-owned lands. But there has also been movement east, toward the Arctic National Wildlife Refuge, and offshore, into the Beaufort Sea. The most recent exploration has moved onto federally owned land in the National Petroleum Reserve, land originally set aside by President Warren Harding in 1923 because of the belief, even then, in its potential for oil production (Coates 1993).



The gravel footprint in North Slope oil fields has changed over the last three decades, as shown in the maps above.

The oil industry builds facilities on layers of gravel, up to two meters thick, that insulate the ground. Without gravel, underlying permafrost would melt, leaving sinkholes on the surface. In an Anchorage conference room, red and green lines cross a map projection—red for gravel and green for pipelines. Prudhoe Bay sits at the center of the map, at ground zero and almost completely hidden under an exploding star of red and green. "The oil fields are growing," states the presenter. "We're letting them bury the American Arctic."

But even in Prudhoe Bay, the oldest and most developed field, there remains a great deal of tundra—less than 3 percent of the Prudhoe Bay landscape has been covered by gravel. What the map shows at Prudhoe Bay—tundra buried by roads and pipelines—can be interpreted as an artifact of scale; as drawn, the roads and pipelines would translate to a width of 300 meters on the tundra. But if they were drawn any finer, in keeping with the map scale, they would be hairlines, barely visible. The width of the lines, coupled with a strong reaction to lost wilderness, might suggest a larger impact than actually exists. However, the lines, despite their width, may not capture indirect impacts. Ted Rockwell, in talking about the footprint, likes to distinguish between what he calls the "actual footprint," the "effective footprint," and the "perceived footprint." The map exaggerates the actual footprint, but the effective footprint may affect wildlife for some distance beyond the lines on the map, and the footprint as perceived by some individuals may cover an even greater area.

Steve Taylor, a long-term environmental manager and advisor to the oil industry, explains arctic sprawl in terms of an evolution of technology. "People look at Prudhoe Bay," he says, "and they don't realize that what they are seeing is 1970s technology. Look at the Kuparuk field. Prudhoe Bay has 50 gravel production pads, and Kuparuk has 49. But Prudhoe Bay's footprint is two-and-a-half times bigger than Kuparuk's. The industry learned from Prudhoe Bay. If we built Prudhoe Bay today, it would be less than one-third of its size. But this is an emotional issue. You look at a map, and the industry is growing."

The Kuparuk field came into production just 4 years after Prudhoe Bay. Since then, gravel footprints have become even smaller (Gilders and Cronin 2000). In the early days, gravel pads included impoundments, called "reserve pits," for storage of drilling mud and other liquids, but since 1987 these waste liquids have been injected below ground into stable geological formations. Drilling equipment has been redesigned to allow close spacing of wells. Directional drilling, which can allow access to oil reservoirs more than 5 miles from a gravel pad, reduces the need for large numbers of pads. Some of the new fields are roadless developments, meaning that they are not connected to Prudhoe Bay by gravel roads. The pipelines running to Prudhoe Bay are built and maintained from temporary ice roads during the long winter season, and personnel and supplies move in and out by air during the warmer months.

While it is true that gravel covers well under one-quarter of 1 percent of the North Slope—less than 45 out of a total of 230,000 square kilometers—it is also true that the gravel is concentrated in certain areas, such as Deadhorse, a staging area for the oil fields. In addition, the gravel leaves long, thin lines across the landscape in the form of roads that may have impacts beyond those captured in measurements of the area covered. Dave Yokel, a wildlife biologist who works for the federal government, echoes Ted Rockwell's concerns. "How spread out is the gravel?" he asks. "Isn't it really a line of gravel, or a network of gravel lines? And what does that mean in terms of environmental impact? Pipelines and power lines add to the footprint. And there is the issue of habitat fragmentation."

There have been attempts to reverse arctic sprawl through restoration of abandoned gravel pads and roads (e.g., McKendrick 1991, Jorgenson and Joyce 1994). Jay McKendrick, a retired professor and a consultant to industry, has worked for two decades to develop restoration methods that can be applied when the oil runs out. He has seeded and fertilized dozens of sites scattered across the North Slope, and on one pad in the Prudhoe Bay field, he maintains a complex experiment, now more than 10 years old. In many cases, restored sites are similar in appearance to natural areas. But real progress with restoration, McKendrick believes, is not a technical issue. "We need to decide what we want out here," he says. "The instinct is to say that we should put the land back to what it was, back to tundra. We may be able to do that in some cases, with enough time and money. But is that really what we want?"

Everyone, it seems, has an opinion. For some, restoration to a tundra landscape will not suffice; development has occurred, and even if every sign of its presence is removed, it has converted wilderness to something less than wilderness. Development has changed natural tundra to palimpsest, creating a film of human history that can never be erased. For others, gravel pads and gravel roads provide habitats that are otherwise rare on the North Slope—they point to data showing that caribou use gravel pads and roads for insect relief, and they see the expense and risk of removing every sign of development not only as a waste of money, but as poor wildlife stewardship.

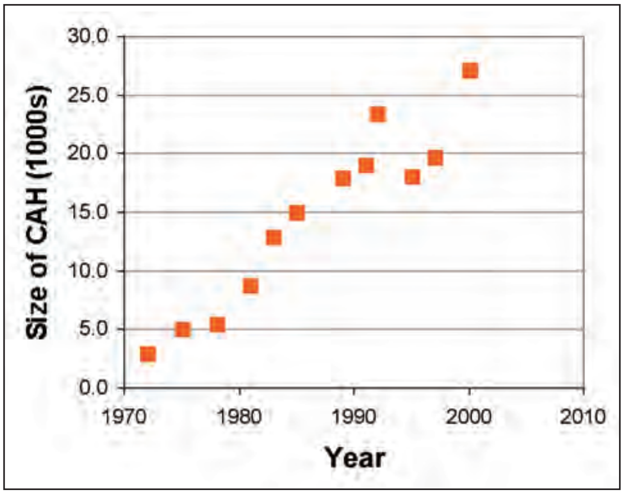
Caribou

Long before development became a reality on the North Slope, caribou became a flagship species for the Arctic. Collins and Summer, writing in a 1953 Sierra Club article, described caribou migrations and then said, "Now we know what it must have been like to see the buffalo herds in the old days" (quoted in Coates 1993, p. 97). Today, similar statements are often heard about the Arctic National Wildlife Refuge's Porcupine caribou herd.

Jay McKendrick examines **Descuriania sophoides** *(Northern tasty mustard) plants established on an abandoned gravel pad. Photograph by David Predeger, © BP Exploration (Alaska).*

Caribou are a common sight in North Slope oil fields. Photograph by John Warden, © BP Exploration (Alaska).

In February 2001, Ken Whitten, a retired biologist who spent 25 years studying caribou for the Alaska Department of Fish and Game, published an opinion piece in the *Anchorage Daily News* warning that development of the Arctic National Wildlife Refuge would jeopardize the Porcupine caribou herd (Whitten 2001). His piece, it might be argued, called for application of the precautionary principle, because data showed that cows with calves avoid development and pipelines may deflect caribou movement. Matt Cronin, a molecular geneticist and wildlife biologist who has been working as a consultant to the government and industry in Alaska for 15 years, responded to Whitten's editorial (Cronin 2001). Cronin listed, as bullet points, what he believes to be the objective facts of oil industry effects on caribou. He pointed out that the Central Arctic Herd has grown since development took place—from 5000 animals in the mid-1970s to 27,000 animals today. Further, he pointed out that caribou density and calf production are as high in oil fields as they are in undeveloped areas, and that caribou do use and travel through developed areas. In Cronin's rebuttal, also published in the *Anchorage Daily News*, the irritation cannot be missed: "Whitten," Cronin wrote, "was selective in presenting information."



The size of the Central Arctic herd (CAH) of caribou has increased since 1972. Data are from censuses and estimates by the Alaska Department of Fish and Game, US Fish and Wildlife Service, and Bureau of Land Management.

When I mention these editorials to Dave Yokel, the wildlife biologist, he says that both Whitten and Cronin are respectable scientists, despite the divergent viewpoints. "The growth of the Central Arctic Herd is an undeniable fact," Yokel adds. "But we have to look at that in context. Other North Slope herds have also grown dramatically in the past 20 years. The real question is, 'How much did the Central Arctic herd grow relative to how much it would have grown in the absence of industry?' And we can't answer that. I attend meetings, and people around me will have answers to tough questions, like the question about caribou. They'll make statements with certainty. But I just don't know. An unbiased look at the data leaves more questions than answers."

Whales

Several hundred people gather in Hopson Middle School's gymnasium for the Arctic Economic Development Summit. The gymnasium itself could be from almost anywhere—a team logo is painted on one wall, a scoreboard is mounted on another wall, and basketball hoops hang from the ceiling. But this is not "almost anywhere"; this is Barrow, Alaska, the largest permanent settlement on the North Slope and home to some 4000 residents, mostly Inupiaq natives. Outside, the temperature is –20 degrees Fahrenheit. The talk here is about the future, and, repeatedly, the importance of today's children in tomorrow's world. "For a lot of us," one speaker says, "subsistence is our only way of life. Development should not be at the expense of our subsistence way of life."

The speaker refers to the hunting life style that has evolved over thousands of years and that continues to provide both physical and cultural sustenance to the community. The mayor, who is also a whaling captain, adds to this sentiment: "We from the North Slope Borough," he says, "have fostered development offshore and onshore, but we have protected our subsistence way of life first and foremost. Priority number one is protecting our way of life."

Shown here is a whaler and his umiaq (traditional skin boat) on the ice pack near Point Hope. Photograph © BP Exploration (Alaska).

Emotions run high on both sides where industrial development is concerned. On the one hand, the oil industry provides a source of income. On the other hand, people worry about the survival of their culture. A large part of this culture is tied to the spring and fall bowhead whale hunts. It is little wonder, then, that activities that could harm whales evoke strong responses.

While emotions run high, there is an interest in science and scientific research that cannot be missed. Tom Albert, a scientist employed by the North Slope Borough and a recent recipient of a retirement award presented by the Barrow community, has pointed out that North Slope residents are interested in the outcomes of environmental studies, perhaps more so than residents anywhere else in the United States (Albert 2001). But this interest in and appreciation of science has not been without problems. In the 1970s, a "scientific" census undertaken by the National Marine Fisheries Service estimated the bowhead whale population at about 1300 animals (Tillman 1980). In response, the International Whaling Commission prohibited subsistence hunting of bowheads in 1978, writing, "from a biological point of view the only safe course is to reduce the kill of bowhead whales from the Bering Sea stock to zero" (International Whaling Commission 1979, p. 48). The native community did not believe the census results. The census was based on visual observations of whales swimming through open water leads near shore. However, native hunters knew that bowheads swam long distances under ice and used leads further offshore, well away from census observers. In a written statement, one group of respected hunters blamed scientists for what they saw as unreasonable interference with their way of life: "There are a lot of bowheads out there that the scientists aren't counting. Many are out in the ice and therefore are not seen when they pass by Barrow. As a result of poor counting, the scientific community helps put these unfair quotas upon us" (Albert 2001).

Later, under the guidance of Tom Albert and others, improved census methods were developed. The improved methods coupled traditional knowledge of whale behavior with scientific methods, which combined hydrophone and visual observations with complex statistical methods. In one instance, observers saw only three whales, but hydrophones tracked 130 whales under the ice. Population estimates were revised upward, to 4417 in 1985, 7200 in 1987, 7800 in 1988, and to 8200 in 1996 (Raftery and Zeh 1998). As a result, the International Whaling Commission backed down from the prohibition on subsistence hunting.

The residents of Barrow have not forgotten this experience. To some degree, it left an atmosphere of distrust that hindered sharing of traditional knowledge with scientists. This distrust was overcome in part by the efforts of the late Harry Brower, Sr., a native whaling captain and an employee of the Naval Arctic Research Laboratory, who helped convince other whalers that good science could help their cause and that scientists would benefit from their traditional knowledge (Albert 2001).

Today, natives are concerned about the impact of oil development on whales (BP 2001). The problem, or at least one of the problems, is noise. The whalers are concerned that construction noise from offshore development may push the migrating whales further out to sea, making it harder to hunt the whales. Seismic surveys, which use arrays of air guns to send sound pulses into the ground in search of oil-bearing formations, are known to displace whales from their normal migration patterns by tens of kilometers. Seismic surveys are now timed to avoid impacts on whales. Currently, the first major offshore development—the Northstar project—is under construction in the Beaufort Sea. Hydrophones are deployed to quantify sound from construction and to determine the distance at which construction noise is lost in the ambient noise of the Beaufort Sea. While North Slope residents await results of this study with interest, they remain skeptical after their experience with the early whale census. It is not unheard of for scientists engaged in public meetings to be called liars or worse, and those scientists working in this arena have two choices: leave or grow thick skin.

A role for science

"Decisions about the oil industry on the North Slope are not driven by science," Steve Taylor says. "But science plays a role. It contributes to the decisionmaking process."

Lloyd Fanter adds to this theme: "Would industry protect the environment in the absence of agency and public scrutiny? And would the agencies have been successful at pushing for changed practices in the absence of data? Science drives conditions set in permits, but it is public interest and public values that drive science forward on the North Slope. It is an emotional attachment by millions of people to something they have never seen that lets the agencies insist on a scientific approach to managing for environmental integrity on the North Slope."

Is it possible to filter out the emotion from the data? "All humans have biases," Dave Yokel says. "Occasionally, someone will claim to be entirely objective. That's just baloney."

In thinking about science and its role in decisionmaking on the North Slope of Alaska, the words of Paul Feyerabend, the scientific historian and philosopher, come to mind: "Scientific institutions are not 'objective'; neither they nor their products confront people like a rock, or a star. They often merge with other traditions, are affected by them, affect them in turn" (Feyerabend 1995, p. 143). And Feyerabend's words remind me of the man in the conference room: "You have to look at the data closely," the man had said, "and think about the science, but when you get up to the North Slope, you'll hear those caribou go thundering past, and you'll get this gut feeling that you just can't ignore."

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