Summary of Changes in the Active Layer and Permafrost in the Arctic

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Permafrost and active layer conditions have experienced significant changes in the past decades in the Arctic. Permafrost temperature changes can be used as a sensitive indicator of climate change and long-term changes in the surface energy balance (Lachenbruch and Marshall, 1986). Very small changes in surface climate conditions can produce significant changes in permafrost temperatures. Monitoring of permafrost thermal regime can go back as early as in 1940s in northern Alaska. Permafrost surface temperature has in general increased about 2 to 4°C over the last 50-100 years on the North Slope of Alaska (Lachenbruch and Marshall, 1986) although at some sites there was little warming or even cooling trend. Warming of air temperatures in the late 1880s and early 1900s in North America may have preceded warming of the permafrost. Based on data from Barrow, Alaska, changes in air temperature alone since the early 1920s could not explain the observed warming of the permafrost, implicating changes in snow cover or perhaps an earlier warming (Zhang and Osterkamp, 1993). Recent measurements show that permafrost on the Alaskan Arctic Coastal Plain and Foothills warmed about 3°C since the late 1980s (Clow and Urban, 2002). Long-term monitoring of permafrost temperatures through deep boreholes in a north/south transect across the North Slope reveals variable warming over the last twenty-five years. Permafrost temperatures along the Alaskan Arctic Coast increased about 2 to 3°C and about 0.5 to 1.5°C in the discontinuous permafrost regions south of the Brooks Range since the mid 1980s (Osterkamp, 2003).

Permafrost temperature data collected since the mid 1980s between depths of 10 and 20 m from monitoring sites south of Norman Wells in the Mackenzie valley indicates that a general warming of permafrost has occurred with the larger increases observed at the more northerly sites (Smith et al. in press). No significant trend of permafrost temperature change is observed in the southern Mackenzie valley, where permafrost is thin (less than 10 to 15 m thick) and warmer than –0.3°C (Smith et al. in press, Couture et al., 2003). The absence of a trend is likely due to the absorption of latent heat required for phase change (e.g. Riseborough, 1990). Similar results are reported for warm permafrost in the southern Yukon Territory (Burn 1998; Haeberli and Burn, 2002). In the central Mackenzie valley, an increase in temperatures of up to 0.03°C/year has been observed in permafrost with mean annual near surface temperatures near –1°C and thickness of less than 50 m (Smith et al. in press). Kershaw (2003) reports an increase in shallow (1.5 m) mean annual permafrost temperatures between 1990 and 2000 at some palsa sites to the west of the Mackenzie valley, in the Macmillan/Caribou Pass region of the Northwest Territories. In the northern Mackenzie basin, in colder (-7°C) and thicker permafrost, temperatures at a depth of 28 m increased by up to 0.1°C per year in the 1990s (Smith et al. in press, Couture et al., 2003). Warming of permafrost since the 1960s and 1970s, in the Mackenzie Delta region has also been observed by Burn (2002). This warming in the central and northern Mackenzie region is consistent with the general increase in air temperature observed since the 1970s.
Warming of permafrost is also observed in the Canadian eastern and high Arctic but this appears to have mainly occurred in the late 1990s. At the high Arctic observatory at Alert, Nunavut, a warming of 0.15°C per year at a depth of 15 m occurred between 1995 and 2001 and warming of about 0.06°C per year has occurred since 1996 at a depth of about 30 m (Smith et al. 2003). Shallow permafrost temperatures (2.5 m) increased by 1°C between 1994 and 2000 at a high Arctic site at Lake Hazen (Broll et al., 2003). Cooling of permafrost was observed from the late 1980s to the early 1990s at a depth of 5 m at Iqaluit in the eastern Arctic. This cooling however, was followed by warming of 0.4°C per year between 1993 and 2000 (Smith et al., in press). This trend is similar to that observed in Northern Quebec, where cooling of permafrost was observed between the mid 1980s and mid 1990s at a depth of 10 m (Allard et al. 1995) which was followed by warming beginning in 1996 (Allard et al., 2002; Brown et al. 2000).

Thermokarst topography forms as ice-rich permafrost thaws, either naturally or anthropogenically, and the ground surface subides into the resulting voids. The important and dynamic processes involved in thermokarsting include thaw, ponding, surface and subsurface drainage, surface subsidence and related erosion. These processes are capable of rapid and extensive modification of the landscape and preventing or controlling anthropogenic thermokarsting is a major challenge for northern development. Extensive thermokarsting has been discovered near Council, Alaska, U.S.A. (Hinzman et al., 2001; Yoshikawa and Hinzman, 2003). Investigation of the physical factors that influence thermokarst pond formation indicates that in warming climate, the permafrost will degrade and these ponds will drain. In response to some imposed disturbance, such as a tundra fire or climatic warming, massive ice permafrost may differentially thaw, creating irregular surface topography. Depressions forming on the surface soon form ponds, accelerating subsurface thaw through lower albedo and additional heat advected into the pond through runoff. In time, a talik (a layer of unfrozen soil above the permafrost and below the seasonally frozen soil) may form below such ponds as the depth of water becomes greater than the amount that can refreeze during the winter. If the talik grows to a size that completely penetrates the underlying soil or connects to a subsurface layer that allows continued drainage, the pond may then begin to drain.

The active layer is that portion of the soil above permafrost that seasonally experiences thawing and freezing and plays an important role in cold regions because of most ecological, hydrological, biogeochemical, and pedogenic activity takes place within it (Himzman et al., 1991; Kane et al., 1991). Changes in active layer thickness are influenced by many factors, including surface temperature, physical and thermal properties of the surface cover and substrate, soil moisture, and duration and thickness of snow cover (Brown et al., 2000; Frauenfeld et al., 2004; Zhang et al., 2005). The inter-annual variation of thaw depth at a site is quite large and consequently, utilizing depth of thaw as an indicator of climatic change may be quite difficult as one would be looking for the response to a subtle change amidst large annual variations. When the other conditions remain constant, changes in active layer thickness could be expected to increase in response to the warming of climate, especially summer air temperature.
Long-term monitoring of active layer has been conducted over the past several decades in Russia. By the early 1990s, there were about 25 stations, each containing 8-10 plots and 20-30 boreholes to depth 10-15 m for measuring ground temperatures (Pavlov, 1996). Measurements of soil temperature in permafrost have been carried out in the former Soviet Union from more than 30 stations, most of them started in the 1950s but a few was as early as in the 1930s. Over the period 1956-1990, the active layer exhibited a statistically significant deepening by about 20 cm (Frauenfeld et al., 2004). Changes in air temperature, thawing index, and snow depth are responsible for the increase in active layer thickness.

The Circumpolar Active Layer Monitoring (CALM) program was developed in the 1990s and currently incorporates more than 100 sites worldwide (Brown et al., 2000). CALM is designed to observe the response of the active layer and near-surface permafrost to climate change. The results from northern high-latitude sites demonstrate substantial inter-annual and inter-decadal fluctuations in active layer thickness. The active layer responds consistently to forcing by air temperature on an inter-annual basis. During the mid- to late-1990s in Alaska and northwestern Canada, maximum and minimum thaw depth was observed in 1998 and in 2000, corresponding to the warmest and coolest summers, respectively. Evidence of increase in active layer thickness, thaw subsidence, and development of thermokarst are observed, indicating degradation of warmer permafrost (Brown et al., 2000).

The primary control on local hydrological processes in northern regions is dictated by the presence or absence of permafrost, but is also influenced by the thickness of the active layer and the total thickness of the underlying permafrost. As permafrost becomes thinner or decreases in areal extent, the interaction of surface and sub-permafrost ground water processes becomes more important (Woo, 1986). The inability of soil moisture to infiltrate to deeper groundwater zones due to ice rich permafrost maintains very wet soils in arctic regions. However, in the slightly warmer regions of the subarctic, the permafrost is thinner or discontinuous. In permafrost-free areas, surface soils can be quite dry as infiltration is not restricted, impacting ecosystem dynamics, fire frequency and latent and sensible heat fluxes. Other hydrologic processes impacted by degrading permafrost include increased winter stream flows, decreased summer peak flows, changes in stream water chemistry, and other fluvial geomorphologic processes (McNamara et al., 1999). Hydrologic changes witnessed among study sites include drying of thermokarst ponds, increased active layer thickness, increasing importance of groundwater in the local water balance and differences in the surface energy balance. By far, the most significant changes occur in response to changing permafrost extent or thickness. As permafrost becomes thinner, the sub-permafrost groundwater becomes more important, either by contributing groundwater to streamflow, or allowing surface water to drain. Thickening of the active layer and melting of ice-rich permafrost in the Russian Arctic drainage basin may have already contributing, in part, to the increased river runoff (Zhang et al., 2005).