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Hydro-climatic and lake change patterns in Arctic permafrost and non-permafrost areas

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SUMMARY

This paper investigates patterns of lake-area and hydro-climatic change in Arctic river basins, and possible influence of permafrost change reflected in such patterns. A salient change pattern, emerging across all investigated basins in both permafrost and non-permafrost areas, is an opposite change direction in runoff (*R*) from that in precipitation (*P*). To explain this change contrast, an increase (decrease) in relative water-balance constrained evapotranspiration ET_{wb}/P is required where *R* decreases (increases). Increasing temporal variability of daily river discharge (*sdQ*) is found in all basins with spatially extensive lake decrease, which also exhibit decrease in ET_{wb}/P. Clear indication of basin-wide permafrost thaw is found in only one basin, and is possible in two more, but unlikely in the largest of the total four investigated permafrost basins.

1. Introduction

Arctic change in relation to the hydrological cycle has been frequently reported during the last decade (Hinzman et al., 2005). Some of the hydrological effects of Arctic climate change are clearly observable, such as decline in annual snow cover (Serreze et al., 2000), retreat or melting of glaciers and ice caps which contributes to increased melt water flow (Dyurgerov et al., 2010), and changes in lake number and area (Smith et al., 2005; Riordan et al., 2006; Carroll et al., 2011; Jones et al., 2011; Chen et al., 2014; Karlsson et al., 2014). River discharge has also received increased attention in the Arctic, since its change may enable interpretation of other large-scale changes in the terrestrial hydrological cycle by integrating precipitation, evapotranspiration and water storage change over basin scales. Observed changes in Arctic river discharge during recent decades include: increased mean discharge from rivers in Russia (White et al., 2007), decline in mean discharge from North American rivers with an exception from glacial fed rivers in Alaska, and changes in seasonal discharge (including a shift to earlier spring discharge in Russia) (Shiklomanov et al., 2007). Some of the observed changes in discharge are attributed to climate change (e.g., atmospheric drivers and permafrost change), while others are attributed to other anthropogenic influences (e.g., reservoir regulations) (Yang et al., 2004).

Although some effects are observable, other consequences of Arctic climate change are not so readily apparent, including permafrost degradation that is expected to become widespread with a warming climate, causing dramatic changes in the Arctic terrestrial system. However, hydrological changes in permafrost regions have been found useful also for interpretation of permafrost change. Specifically, changes in ice-rich permafrost and active layer depth directly affect the subsurface water storage and as such can be expected to also affect river discharge (Kane, 1997; Yang et al., 2002; Yoshikawa and Hinzman, 2003). Generally, watersheds with a high percentage of permafrost coverage have low subsurface storage capacity and thus a low winter base flow (minimum flow), and a high spring or summer peak flow (maximum flow) (Kane, 1997). Previous studies have indicated a decreasing trend in intra-annual runoff variability (quantified by decreasing maximum and increasing minimum flow) as a hydrological signal of permafrost degradation, attributed to increasing active layer thickness and deeper and longer flow pathways to surface water discharge (Ye et al., 2009; Frampton et al., 2011). Changes in maximum flow

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are also consistent with changes in cold-season precipitation (November–April), as peak flow is primarily snowmelt dominated (Shiklomanov et al., 2007). Furthermore, also other factors may influence peak flow, such as faster snowmelt caused by increased air temperature in spring, increased cold-season base flow and/or increased meltwater flow from basins that contain glaciers (Shiklomanov et al., 2007). Minimum flow is therefore believed to be more robust than maximum flow as an indicator of permafrost change, since subsurface processes almost exclusively control minimum flows (Smakhtin, 2001), in contrast to maximum flows which are governed by a combination of several hydrological processes (Rennermalm and Wood, 2010). Specifically, it is argued that minimum flow can be used as an approximation of base flow, or groundwater flow, which is sensitive to change in permafrost conditions. Frey and McClelland (2009) and Rowland et al. (2010) have shown that permafrost degradation may cause transition from a surface water-dominated to a groundwater-dominated hydrological system. Increasing trends in cold-season minimum flow is also observed in most of the Arctic, suggesting widespread changes in subsurface hydrology (Rennermalm and Wood, 2010).

Furthermore, thermokarst lakes depend on permafrost, with lakes forming where melting of ground-ice and surface settlement initiate ponding. Eventually, the lakes may also drain if lateral erosion (expansion) and continued permafrost thawing follows the initial ponding. Lateral erosion may yield bank overflow and, in some cases, open talik formation may connect the surface water with the groundwater system (Hinkel et al., 2003; Jorgenson and Shur, 2007; Huissteden et al., 2011). Thermokarst lake changes have been observed in northern Canada, Siberia, Alaska, and in Scandinavia (Fig. 1), where lake drainage usually occurs over timescales of years to decades (Marsh et al., 2009). However, the main driving mechanism of changes in lake surface area is not always clear, as lake-surface area may change due to several processes, including (a) changes in precipitation (Plug et al., 2008; Tarasenko, 2013), (b) changes in evapotranspiration (Riordan et al., 2006; Labrecque et al., 2009), (c) lateral drainage caused by shoreline erosion (Marsh et al., 2009; Smith et al., 2005; Jones et al., 2011), (d) internal drainage when underlying permafrost thaws and open taliks appear (Yoshikawa and Hinzman, 2003; Smith et al., 2005; Karlsson et al., 2012, 2014), (e) ice-jam flooding (Chen et al., 2014), or (f) terrestrialization, including encroachment of floating mat vegetation and basin infilling of organic matter and sediment (Roach et al., 2011; Sannel and Kuhry, 2011).

Although connections between permafrost degradation and hydrological changes in thermokarst landscapes have been established (Yoshikawa and Hinzman, 2003; Karlsson et al., 2012), the linkages between permafrost degradation, thermokarst-lake changes (initiation, expansion, drainage) and changes in hydrological fluxes are still not fully understood. Nevertheless, the high abundance of lakes in the Arctic is largely dependent on the presence of permafrost, acting as an aquiclude that limits exchange between surface water and groundwater systems. This dependence implies that observations of changes in thermokarst lake size and number may offer an observation window to ongoing permafrost change, based on relevant interpretation of how permafrost thaw, thickening of the active layer, thermokarst processes and associated lake changes combine with observable hydro-climatic changes.

This paper aims at identifying possible characteristic patterns of lake-area change and corresponding hydro-climatic changes in Arctic permafrost and non-permafrost areas, and assessing possible influence of permafrost change in such patterns. This aim is pursued by synthesizing observations of hydro-climatic changes and large-scale lake-area changes (with both increasing and decreasing trends) in drainage basins of varying permafrost coverage (from continuous to no permafrost). More specifically, we investigate here reported changes in total lake area over time, based on previous published literature, and use available monthly precipitation and temperature data, along with daily discharge data records from drainage basins that overlap the areas of reported lake change observation, in order to assess concurrent change patterns.

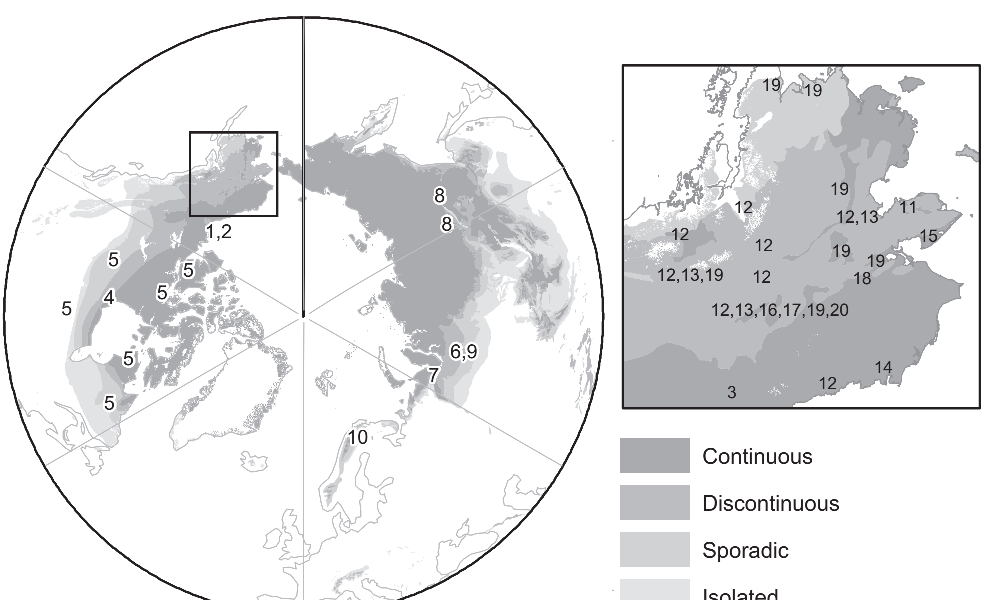

Fig. 1. Pan-Arctic study regions where thermokarst lake changes have been observed: 1, Tuktoyuktuk Peninsula, Canada (Plug et al., 2008); 2, Tuktoyuktuk Peninsula (Marsh et al., 2009); 3, Old Crow Basin (Labrecque et al., 2009); 4, Hudson Bay Lowlands (Sannel and Kuhry, 2011); 5, Canada 50–70°N (Carroll et al., 2011); 6, Northwestern Siberia (Smith et al., 2005); 7, Rogovaya (Sannel and Kuhry, 2011); 8, Central Yakutia (Tarasenko, 2013); 9, Nadym and Pur River Basins (Karlsson et al., 2012, 2014); 10, Tavvavouma (Sannel and Kuhry, 2011); 11, Council (Yoshikawa and Hinzman, 2003); 12, Arctic Coastal Plains and boreal regions in interior Alaska (Riordan et al., 2006); 13, Alaskan boreal forest (Roach et al., 2011); 14, Arctic Coastal Plain (Hinkel et al., 2007); 15, Northern Seward Peninsula (Jones et al., 2011); 16, Yukon River, central Alaska (Rover et al., 2012); 17, Yukon Flats (Jepsen et al., 2013); 18, Kobuk Valley (Necsoiu et al., 2013); 19, Alaska (Roach et al., 2013); 20, Yukon Flats (Chen et al., 2014).