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Satellite Remote Sensing of Lake Ice Meltout Patterns Near Barrow, Alaska

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Satellite Remote Sensing of Lake Ice Meltout Patterns Near Barrow, Alaska

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Abstract

The Arctic Coastal Plain of northern Alaska is covered with thousands of lakes developed atop continuous permafrost. Lake ice meltout patterns near Barrow were evaluated across the historical record using available cloud-free Landsat imagery. An Iterative Self-Organizing Data clustering algorithm was used to differentiate ice from water during meltout, and was overlain on lake boundary vectors to calculate the percentage of ice covering each lake in the ~11,000 km² study area near Barrow. This processing technique was carried out for seven available scenes in early July across the ~35-year Landsat record. Analysis of these scenes reveals that lakes exhibit interannual variability in the general timing of lake ice meltout. A consistent regional spatial pattern is apparent in all years, with a general decrease in the percentage of ice coverage on lakes further south reflecting the regional climatic pattern. The ice cover on many lakes near the coast persists for a longer period due to the influence of onshore winds and cooler temperatures, possibly enhanced by coastal fog. Some lakes consistently experience earlier or delayed meltout compared to lakes nearby owing to lake-specific factors that may include lake depth, basin geometry, or water salinity. This study illustrates the utility of satellite-derived lake ice observations as a regional climatic indicator.

Key words: Alaska, lakes, lake ice, NDWI, permafrost, remote sensing
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Preface

The Arctic Coastal Plain (ACP) of northern Alaska is dominated by thousands of lakes and drained lake basins. This area stretches from the east-west trending Brooks Range foothills in the south to the Beaufort and Chukchi Seas to the north (Wahrhaftig, 1957) and is completely within the zone of continuous permafrost where ground ice reaches depths of 300-600 meters (Brown, 2001). This area is characterized by topography that gradually slopes upward from the Seas to Foothills with very low relief near the coast and increasing relief further inland. The region can be defined as a dry, polar climate (at 70º to 71.5º N latitude) due to low amounts of total annual precipitation (~30 cm), with long, cold winters, and short, relatively cool summers. Despite the lack of considerable precipitation, the frozen ground conditions prevent infiltration of water to depths resulting in significant overland flow and surface water coverage. Almost 50% of the landscape is covered with lakes and scars left from drained lake basins (Hinkel et al., 2003; Frohn et al., 2005). There are no trees in the landscape as a result of soil temperature and ground ice conditions, but tundra vegetation is present in the form of sedges and grasses, moss, willows, blueberry and cloudberry bushes, and other herbaceous shrubs. Significant fluctuations of insolation occur annually in this area with ice and snow dominating the landscape for most of the year, but then rapidly disappearing with the onset of summer. These seasonal freeze and thaw processes give way to a variety of thermokarst landforms and play a dominant role in landscape modification. Lakes in this region are critical natural resources for local communities, are often used for industrial operations, and provide habitat for a diverse collection of waterfowl, macroinvertebrate, and biotic communities.

The geomorphic processes occurring in this region of northern Alaska are not easily interpreted due to the regional geologic history and the powerful effects of cryoturbation. The
Western ACP (WACP) was subjected to several marine transgressions and regressions (Dinter et al., 1990), resulting in various levels of coastal shorelines that are identifiable across the Coastal Plain. These various stages of sea-level fluctuation can be reconstructed using exclusive features, such as breaks in the slope, dune features, and wave-cut scarps. Ocean waters inundated the land during interglacial stages with reported evidence of a pre-Illinoian interglacial high around 23-29 m asl that dates about 175 kya (Lewellen, 1972; Sellman et al., 1975; Hopkins, 1982). Another prominent shoreline has been identified around 7m asl and is said to be associated with the Simpsonian transgression around 58-75 kya. These clear boundaries divide the WACP into a series of age-related surfaces, with the younger Outer Coastal Plain (OCP) to the north and the older Inner Coastal Plain (ICP) to the south. They can be further characterized by marine sediments associated with the OCP and fine grained aeolian sand sediments with the ICP (Williams et al., 1978; Hinkel et al., 2005). A subset of Inner Coastal Plain coincides with an extensive Pleistocene sand sea discussed by Carter (1981) and characterized by large stabilized dunes oriented parallel to the prevailing winds. The orientation of the dune features indicates that the wind regime has not changed considerably over the recent past. The dunes create topographic ridges on the order of ~50 m, and have likely been affected by thermokarst processes since their formation.

The morphology of lakes within this region has been the subject of scientific investigation since the 1950s, and of lately has gained considerable attention; however, the natural processes by which these lakes evolve are not yet fully understood. This is likely due to their dynamic nature, complex history, and vast areal coverage. Lakes that exhibit similar qualities can be found in other areas of the circum-Arctic (e.g. northern Siberia), but the conditions under which other lakes form may be different than from those on the WACP of
northern Alaska. The result of such variance in lake characteristics is due to a number of factors, of which not all may be yet defined. Attempts to break down the geographic variation of lake characteristics is needed in order to better understand the morphological factors involved in the lake basin development and evolution processes. The initial question in deciphering the lake evolution process stems from the origin of such features; whether the lake basin formed by thawing the underlying permafrost resulting in subsidence and depression of the surface layer, or if the lake is a result of water flowing down an elevation gradient and infilling natural topographic lows within the terrain. Moreover, lake basins with unique shapes discussed below may be the result of both processes operating over time.

Physical and thermal weathering largely drives the expansion, erosion, and subsidence processes occurring within this region. Such a suite of factors may be acting independently of or in conjunction with each other. Additionally, some factors are known to be influential over large geographic regions, while others are confined to smaller areal extents. These micro-physiographic regions, defined by their geographic location, regional relief, geologic history, surficial sediments, ground-ice content, and/or meteorological and climatological regimes play an important role in determining which processes are influential and which processes are less significant.

Lakes in this region exhibit a strong orientation ~10-20° west of north (Sellman et al., 1975). Lakes tend to gain this elliptical shape through a set of wind-induced circulation cells, intersecting the lake perpendicular to its orientation (Cabot, 1947; Carson and Hussey, 1962). These circulation currents gain velocity as they migrate towards to ends of the lake and preferentially erode the ice-rich permafrost. Carson (2001) noted the extensive season for lake shoreline erosion, beginning before the lake is completely ice-free and not ending until lake ice
becomes shorefast (a period of <5 months). In addition to modifying the surface extent of lakes, these circulation cells are transferred to the vertical column of the water body, and thereby physically disturbing lake bottom sediments and shaping lake bottom topography. Ground subsidence also takes place in addition to the thermo-mechanical erosion, but the degree of influence of this factor depends largely on the volume of ice in the substrate (Livingstone et al., 1958; Hussey and Michelson, 1966, Pelletier, 2005).

Various types and shapes of lake basins have been generally identified and classified in the literature as flat-bottomed (Dallimore et al., 2000; Hopkins, 1949; Burn, 2002; Kozlenko and Jeffries, 2000), bowl-shaped (Vardy et al., 1997; Johnston and Brown, 1964), or deep-centered with wide, but shallow littoral shelves (Burn, 2002; Carson and Hussey, 1962; Murton, 1996; Kozlenko and Jeffries, 2000). Winston et al. (2008) report of basins that exhibit similar characteristics as the classes above, but also included lakes with discontinuous basins, lakes with a series of littoral terraces at various depths, and lakes that contain deep pools that are located near or adjacent to the lake perimeter. These various types of basins often co-occur on the Inner Coastal Plain and spatial patterns of lake bottom features seem to be limited. However, an inventory of lake bottom topography is not complete enough to make this division for the entire WACP.

Many researchers have used lake surface characteristics to classify lakes on the Arctic Coastal Plain. Using remote sensing imagery, Sellman et al. (1975) classified lakes for the entire ACP based on lake surface extent, length and orientation of elongate axis, and the density of lake coverage. In this analysis, Sellman classified six different lake units and created a map whereby the lake units are spread discontinuously across the ACP. Sellman excluded lakes that occupy river valleys and floodplains, as lakes in these areas are likely to be the result of different
morphological factors affecting basin development. Moreover, Sellman classifies basins into three classes based on their depth, consisting of shallow lakes and ponds (<1 m), lakes of intermediate depths (~1-2 m), and lakes in excess of 2 meters, but omits these bathymetric characteristics when developing his six-unit scheme. Hinkel et al. (2005), divided lakes into six distinct regions based on the quaternary history of the landscape and categorized lakes by size, length/width ratio, orientation, degree of basin development, and lake density. Conclusions from this analysis indicate that shape metrics are statistically different across space, with lake shape metrics exhibiting more variation as surfaces get older. This study demonstrates a geographic variation of dominant landscape processes that are influential across the WACP such as lake orientation due to onshore prevailing winds from the Beaufort Sea, while highlighting localized factors affecting shape characteristics (e.g. statistical similarity of lakes in ICP subregions ‘a’, ‘b’, ‘c’, which differ from lakes in ICP subregion ‘d’). Through this analysis, Hinkel et al. (2005) reveal the difficulties in identifying spatial patterns due to the various processes and conditions that exist over such a large geographic area and across time.

Many studies note the depth of winter lake ice cover in the high Arctic, which has been observed to be between 1.6 and 2 m (Brewer, 1958; Burn, 2002; Jones et al., 2008), with recent observations on the lower end and varying as a result of geographic location, annual air temperature, and seasonal snow depth. Maximize winter lake ice thickness plays a critical role is basin morphology. Some lakes are shallow such the overwinter ice becomes bedfast, or freezes completely to the bottom. Thermal erosion and thermokarst subsidence are restricted in these lakes, but are accelerated in lakes that do not completely freeze to the bottom. Lakes that retain liquid water throughout the winter disturb the ground thermal regime and result in morphology of the lake basin. Lake bottom temperatures above the 0° C isotherm transfer heat to lake bottom
sediments and thaw the underlying permafrost, creating a zone of perennially thawed sediment known as talik (French, 2007). Lakes deeper than the maximum winter ice cover are also identified for extraction of liquid water for industrial oil and gas explorations during winter (Jones et al., 2008; Jones et al., 2009).

Burn (2002) investigated temperature profiles and permafrost conditions in cores beneath lakes on Richards Island in the western Arctic coast, Canada and discovered taliks penetrating the entire zone of continuous permafrost. Burn (2002) also notes the rapid warming of near surface water over top of the littoral shelves (5.3°C over six days) of lakes brought about by spring conditions and accelerating as the albedo drops with the reduction of ice. Simulating lake ice growth for various types of basins and understanding the relationship of lake ice and basin bathymetry have earned more attention of recently (Williams et al., 2004; Hirose et al., 2008; Jones et al., 2008) as such deep lakes are an important resource for civil and industrial uses, as well as providing a overwintering habitat for aquatic biota (Sellman et al., 1975; Mellor, 1982, 1994; Chambers et al., 2008). In his analysis at Richards Island, Burn (2002) also measured depths of the active layer beneath the shallow littoral terraces of one lake. He found an average thawed sediment thickness of ~140 cm approximately 20 m from shore, and determined that active layer thickness below the littoral shelf is similar to the measured active layer thickness of the adjacent terrestrial environment. Burn (2002) also took a series of 27.5 m deep core samples from the littoral terrace perpendicular to the lake shore and observed a near-vertical boundary of frozen ground at the terrace edge, although the permafrost beneath the terrace shelf reportedly low in ‘excess’ ice content. These observations of a near vertical drop off are consistent with the findings of Winston et al. (2008), where through the use of GPS-assisted sonar soundings, noticed similar basin and slope features near the edge of the littoral shelf. Results from this
study reveal that several shelves were estimated at up to 50° slopes, exceeding the angle of repose for sand under hydrostatic pressure (Stegner and Wesfried, 1999). Therefore, such sediments must be bonded by permafrost at the basin/lake water boundary. Furthermore, Burn’s (2002) findings agree with Mackay’s (1992) estimation of permafrost not occurring in water that is deeper than 2/3 of the maximum winter ice thickness. West and Plug (2007) describe the morphology of lakes through a time-dependent model of heat transfer and subsidence for various quantities of ground ice content to estimate growth rates and talik development. They conclude that lakes with deep-centered pools (~20 m) take approximately 3,000-5,000 years to form. However, their field measurements to validate the model are taken from the Seward Peninsula and the Yukon Coastal Plain and may not be indicative of conditions spanning the Arctic Coastal Plain.

In addition to the lakes of this region, the drained basins that lakes previously occupied are also dominant features of the landscape, accounting for ~25% of the land area (Frohn et al., 2005). Hinkel et al. (2003) used a sample of sediment cores from within drained lake basins to discover a relationship with basin age and accumulation of organic material. Using stages of vegetation succession, Hinkel et al. identify young, medium, old, and ancient basins (14C ages of <50, 50-300, 300-2000, 2000-5500, respectively). This study identifies a significant portion of the evolutionary cycle of lakes, but also highlights the difficulty of reconstructing the ecological record due to the effects of cryoturbation and damage to near-surface sediments. Other research efforts look to describe the movement and dominance of lakes and drained lake basins across the landscape and suggest that these features are part of a thaw lake cycle (Hopkins, 1949; Carson and Hussey, 1962; Britton, 1966; Billings and Peterson, 1980), while recent reports contend that lake evolution is much more complex across the larger WACP and may not be a cyclic as
previously thought (Jorgenson and Shur, 2007). These differing reports may be the result of the spatial variation in field sites, and observed conditions may not be continuous across the entire WACP.

Lakes in general are very transient features, but can provide a continuous stratigraphic record to be used for paleoecological reconstruction (Anderson, et al., 2007). Their existing record contains significant amounts of information, with most of the ecosystem’s representatives applying their signature to the lake document. Using available techniques, lake-bottom sediment cores may be carried out to determine changes in depositional rates and particle sizes, as well as variance over space. Identifying such changes may reveal information about the sources of the material and the capacity at which it was transported. Therefore, an in-depth examination of the aquatic setting can provide details to help explain lake basin development over the recent past.

As mentioned earlier, lakes in this region are heavily affected by wind-induced currents, which lead to roughened surface waters and subsequent energy that is distributed throughout the water column. The three general types of basins defined in the current literature of the WACP include flat-bottom, bowl shaped, and deep centered with both broad and narrow littoral shelves, as mentioned above. Hypsographic relationships can define the varying types of basins by assessing their relative depth to relative area ratios, yielding information on the basin shape and water volume.

Consideration of the depositional and sedimentary processes within the lake may also contribute to the understanding of basin development. Shallow lakes (<3 m) with gently sloping margins will have a different distribution of sediments than lakes with littoral shelves and deep (~20 m) basins. For shallow lakes, the entire lake bed can be considered a depositional environment with mixing and re-suspension of material continuously across the basin. Whereas,
differential deposition zones exist in deep-centered basins and once particles are deposited below the critical depositional boundary depth, they are restricted from mixing and re-suspension (Kalff, 2002). The resuspension of material is therefore a function of both turbulence and particle size. Using ground-penetrating radar, Moorman et al. (1997) discovered sediment characteristics for lakes in the Northwest Territory, Canada based on the reflection signals of various materials throughout the lake bed, with clear distinctions between lacustrine sediment accumulations and diamicton (these lakes are characterized by aeolian deposits). Additionally, Moorman et al. discovered a spatial variability of sediment thickness, noting that the coarser grained materials were found on the sloping margins of the basin and aeolian sediment accumulation maximized in the deepest area of the basin. These findings are largely linked to the critical depositional boundary depth explained by Kalff (2002), which suggests a primary zone of sediment erosion characterized by high turbulence and a zone of sediment accumulation distinguished by low turbulence.

In addition to mixing initiated by wind, the relative densities of differing water temperatures also produce internal seiching, although the energy velocities are not as significant as those kick-started by wind. These thermal structures are responsible for vertical stratification within the lake and more so define the thermocline, or line of abrupt thermal change. Several studies show the varying types of basins under this parameter, which include monomictic lakes, dimictic lakes, isothermal (for shallow, wind-exposed lakes; Brewer, 1958; figure 5 in Burn, 2002), horizontal uniformity during summer and vertical stratification during winter, (figure 6b in Burn, 2002), and lakes that exhibit fine scale variability in the vertical temperature gradient (Bradford, 2007).
Closed-system isolated lakes are good candidates to study depositional environments. In this case, sediments are either local to the lake basin based on quaternary history or result of autochtonous production, are washed in from overland flow (allochthonous production), have an aeolian source, or are broken off of the shoreline by thermomechanical erosion. This results in a net gain of material into the lake. Opposing factors such as subsidence of the lake bed and wave surges that deposit sediments on shore which are then carried away by wind, result in a net loss of material. Quantifying these inputs and outputs may provide further insight to accumulation rates, total net accumulation, and other morphological conditions over time. Open system lakes in this area are more dynamic and much harder to model in this detail. However, techniques have been developed to assess lake characteristics along a series of connecting lakes, or lake chains, and were applied to a series of lakes in northern Alaska, and other areas of North America (Soranno et al., 1999). This method considers the mass balance of open-system lakes that are both spatially connected and functionally connected. Through their analysis, Soranno et al. discovered several relationships along the lake chain, such as increasing nutrients (total phosphorous and total nitrogen) and algal variables for lakes further down the chain. This study notes the significance of these interlinked communal ecosystems and attempts to describe their relationship across space. An analysis of this sort has yet to be conducted for lakes on the Coastal Plain of northern Alaska. However, analyzing lakes across a geographic gradient may provide clues to the morphological processes, as inter-lake connections play an increasing role in landscape modification.

Aquatic biota can also be used as inferences during morphological analysis. Other studies have shown that bioindicators exhibit significant relationships, such as the inverse relationship between lake productivity and basin depth (Kalff, 2002). Kalff (2002) also states
that in shallow transparent lakes, basin morphometry and sediment composition will determine where biological activity occurs. For example, benthic plants hold an ideal depth for efficient production, but may vary based on species, basin morphometry, and/or geographic position. Another factor affecting the presence of plant communities is lake age. During previous summer field seasons, our team has observed benthic vegetation in several lakes in two distinct areas of the Coastal Plain. Areas where we did not observe any vegetation were situated within the Outer Coastal Plain, where lakes are both shallower (<2.5 meters) and more turbid. During a formal presentation at the American Geophysical Union’s annual meeting, researcher Katey Walter Anthony discussed the importance and relationship between the presence of benthic vegetation and the apparent efflux of methane gas from these lakes (Walter Anthony et al., 2009). Advancements in remote sensing techniques allow the detection of such features and their distribution across the landscape in clear waters. However, further research is needed in order to identify such lakes with the presence of such benthic species.

A number of studies have focused on multi-decadal, annual, and inter-annual changes in the areal extent of lakes using recent and historical airborne and satellite imagery. Most recently, Plug et al. (2008) analyzed lake changes on the Tuktoyaktuk Peninsula, western Canadian Arctic from 1978 – 2001 using Landsat imagery and observed that surface extent of lakes in this region may change up to 4% in one year and observed a maximum variability of 14% between all years sampled. The authors of this particular study did not discover trends in lake change over the data record, but did discover a significant relationship between total areal coverage and cumulative precipitation in the 12-months prior to image acquisition. Jones et al. (2008) also discovered the significant correlation between lake area and precipitation prior to image acquisition in the eastern portion of the WACP and attributed the interannual variability of lake surface extent to
the precipitation pattern. Smith et al. (2005) used satellite-derived data between 1973 and 1998 to measure changes in lake surface extent for thousands of lakes across a 515,000 km² region of Siberia and discovered a 6% decline in total lake surface extent for the study area, but revealed a 12% increase in total surface extent in the zone of continuous permafrost. Results from this study highlight the force of thermokarst action as a lake modification mechanism. Duguay and Lafleur (1999) used Landsat data from 1986-1999 for the Old Crow Flats, Yukon Territory and found both increases and decreases in individual lake surface extent, but concluded that overall lake surface extent change was negligible. Riordan et al. (2006) used remote imagery from 1950’s to 2002 to observe changes in lake surface extent across various subregions in Alaska, but focused on only a small portion of the Western Arctic Coastal Plain in northern Alaska. Jones (2006) and Hinkel et al. (2007) did a more thorough analysis of lake change in northern Alaska for the Barrow Peninsula using multi-temporal imagery between mid-1970s and 2002. These studies revealed that ~50 lakes had completely or partially drained within the ~25 year time period. Additional historical aerial photo analysis from 1949-1955 were used to identify a series of lakes that drained within the 5 year period. Hinkel et al. (2007) use a series of proxies (remote imagery, Indigenous Knowledge, and general landscape morphology) to discusses the processes resulting in lake drainage, which are identifiable in some areas as natural drainage processes (e.g. stream meandering and lake embankment spillway) and attributed events in other areas to be human-triggered drainages.

The WACP has been a breeding ground for scientific invetsigation across the academic spectrum, focusing on the smallest details, such as the relationship between the snowy owl and ground lemmings, the planktonic and macroinvertebrate communities within these lakes and ponds, and vegetation succession both across the tundra landscape and within drained lake
basins. These studies will ultimately provide a better understanding of the dynamic interactions in this changing arctic environment, but a thorough understanding necessitates cross-disciplinary cooperation and collaboration in order to sufficiently capture environmental, climatological, biological, and social interactions. It is my expectation that many field scientists who experience this unique setting first hand often leave with more questions than answers. The research presented in this paper is intended to both provide a small amount of detail towards a holistic understanding of the climatological and environmental processes, and their relationship to biological and social communities in this area, as well as facilitate a cross-disciplinary call to action for researchers and the broader permafrost, arctic, and scientific communities.
Introduction

The Western Arctic Coastal Plain (WACP) of northern Alaska is a vast, remote, and lake-rich region situated north of the Arctic Circle between approximately 68° and 72° North latitude. These lakes are developed atop continuous permafrost that reaches to depths of up to 400 m (Brown, 2001). Many of these lakes exhibit an elliptical shape with long axis oriented 10°-20° west of north, or roughly perpendicular to the prevailing wind direction (Carson and Hussey, 1962; Sellman et al., 1975; Hinkel et al., 2003). These lakes are formed by thawing ice-rich permafrost, causing land surface subsidence and enhancing depressions that are favorable for surface water accumulation. Initial sites tend to be ponds in the flat terrain of the coastal plain, interdunal troughs in stabilized sand seas, and depressions in floodplains and deltas (Jorgensen and Shur, 2007). Analysis of satellite imagery demonstrates that lakes and drained lake basins cover roughly 46% of the WACP landscape (Sellman et al., 1975; Frohn et al., 2005; Hinkel et al., 2005). Lakes in this region are especially crucial as they provide resources for local indigenous subsistence communities, are used as a water source for municipal and industrial operations, and provide habitat for a diverse collection of waterfowl, macroinvertebrate, and biotic communities.

The region is characterized by strong seasonal patterns of insolation. Snow and ice envelope the landscape during the long, cold arctic winter, but rapidly disappears with the onset of summer. The timing of spring snowmelt in northern Alaska brings an abrupt and dramatic change to the landscape as the ground cover transitions from the highly reflective properties of snow and ice to the highly absorptive properties of water and vegetation. The effect of this
seasonal transition has considerable impacts on the regional energy budget and geomorphological processes.

Recent evidence suggests that high-latitude environments such as the WACP have experienced increasing air temperatures and are projected to warm further in the near future (Serreze et al., 2000; Solomon et al., 2007). The observed decrease in summer sea ice extent across the Arctic Ocean, as well as locally in the Chuckchi and Bering Seas fringing the WACP, will likely impact terrestrial temperature and precipitation patterns (Cosimo et al., 2008). The expected degradation of the upper permafrost over the next several decades (Chapin et al., 2005; Lawrence and Slater, 2005; Jorgensen et al., 2006), coupled with changes in the precipitation regime and especially winter snow cover, will likely trigger hydrological and ground thermal changes that will impact lake water levels and susceptibility to drainage.

Arctic lakes have also been identified as a globally significant source of atmospheric methane (Kling et al., 1991; Walter et al., 2007a). Climate warming projections suggest that CH₄ efflux from lakes will increase and impact global carbon balances (Walter et al., 2007b; Serreze et al., 2009). A report on methane emissions from lakes in northern Alaska by Phelps et al. (1998) links high methane efflux rates during springtime ice breakup, with measured amounts during initial breakup equal to or exceeding that released during the entire open water period. According to Phelps et al., (1998), the springtime pulse results from overwinter biogenic methane production, and the accumulation of gas in water and ice beneath the ice sheet. Phelps et al. (1989) suggest that the overall contribution of Arctic methane efflux to the global atmospheric methane pool may be underestimated due to this occurrence. The Pan Arctic Lake Ice Methane Monitoring Network, an initiative begun during the International Polar Year, is focused on comprehensive monitoring of methane ebulation from individual lakes, and applying
measurements to regional scales based on remotely sensed data. Roughly one-quarter of the lakes on Earth are located in Arctic Eurasia and North America (Lehner and Döll, 2004), with lakes in the WACP covering 20% (7,000 km²) of the land area (Frohn et al., 2005). Therefore, a better understanding of this changing seasonal land surface change is crucial.

The objectives of this research are to use the multispectral satellite imagery archive to analyze springtime lake ice coverage in an area that includes the Barrow Peninsula and extends southward across the Arctic Coastal Plain. Available springtime Landsat scenes from 1975 to 2006 are used to construct a multiyear record of springtime lake ice conditions for >1,700 lakes. The purpose of this study is to map regional and local patterns in the timing of lake ice meltout, and identify causative factors.

**Background**

Lake ice has shown great utility as a proxy metric in assessing long-term climate records. Historical lake ice freeze-up and meltout dates are often used an indicator of regional climate trends over the observation period. A report by Magnuson et al. (2000) revealed statistical trends for delayed freeze-up dates and earlier meltout dates for a set of major lakes in the Northern Hemisphere over the past ~150 years. Their study concluded that linear trends of delayed freeze-up (5.8 days) and earlier meltout (6.5 days) is equivalent to a temperature increase of 1.2 °C over 100 years. They also identified a significant relationship between lake ice freeze-up and meltout dates, and the air temperature in the month or two preceding the event.

It is well known, however, that air temperature is not the only control on lake ice thickness or meltout dates. Snow cover thickness, for example, influences the timing of meltout through its insulating effects and control on ice growth or decay. A study published on the
degradation of lake, river, and sea ice in the Canadian and Alaskan Arctic (Bilello, 1980) revealed a correlation between yearly cumulative freezing degree-days and the decay rate of lake ice. More recently, a study conducted by Zhang and Jeffries (2000) in northern Alaska indicate that freezing degree-days are not as well correlated with seasonal ice thickness ($r^2 = .53$) as is a snow cover index that incorporates the thickness and duration of snow cover ($r^2 = .82$). Modeled results from Zhang and Jeffries (2000) suggest that lake ice thickness has not shown any direct trend in northern Alaska since 1947, but rather reflects interannual variation in maximum lake ice thickness.

Temperature and snow often interact in less direct ways to increase the net ice thickness. During periods of rapid cooling, thermal contraction of the ice cover triggers cracking and allows lake water to flood the snow-covered surface, yielding a snow ice layer above the conglomeration ice. A similar process can occur during an extensive thaw when snow meltwater flows across the frozen ground to the lake basins and causes flooding and snow ice formation. Field measurements at numerous lakes near Barrow reveal that a majority of lakes augered in spring (April) have a thick layer of snow ice. This additional accumulation of ice may impact meltout patterns, but the process is not continuous throughout the course of a season and may be episodic.

Satellite remote sensing has been used in various ways to monitor lake ice conditions. Maslanik and Barry (1987) used coarse resolution Advanced Very High Resolution Radiometer (AVHRR) imagery with nearly daily repeat coverage to identify ice meltout dates for a set of large lakes in Finland and Canada. Latifovic and Pouliot (2007) also used AVHRR to determine statistical trends of delayed freeze-up and earlier meltout dates for 36 Canadian lakes. Both studies note the difficulty in ascertaining satellite derived freeze-up dates due to lower sun
elevations and complications induced by the surrounding snow cover. Advantages of remote sensing systems include broad spatial coverage providing the ability to monitor many lakes, an archived set of observations of past conditions, and return coverage of the satellite to monitor temporal change and dynamics. Some drawbacks in using optical remote sensing data to make lake ice observations include the presence of clouds, long periods of seasonal darkness at high latitudes, insufficient satellite return time to observe ice dynamics at the appropriate time scale, and limited availability of historical high-resolution imagery for monitoring smaller lakes. Active sensing systems such as Synthetic Aperture Radar (SAR) can address some of these issues and have been used to monitor the growth of ice on lakes throughout the arctic winter (Mellor, 1982). However, data availability is more restrictive.

Remote sensing studies of the lake-rich region of northern Alaska also offer insight into land surfaces processes at the landscape scale to better assess regional climate and associated limnological processes. A study by Zhang et al. (2000) examined patterns of surface albedo across the Coastal Plain of northern Alaska with AVHRR imagery, and concluded that spring snowmelt occurs progressively from the foothills of the Brooks Range in the south towards the Arctic Coast. Similarly, Eastman et al. (2009) used MODIS data to estimate land surface temperature for the period 2001-2008, and reported a significant warming trend in northern Alaska during the onset of spring and into summer, while winter temperatures remained unchanged. Both studies used coarse resolution images (250 m to 1.1 km spatial resolution), which cannot adequately resolve most Arctic lakes nor map lake ice.

Methodology

Selecting Archived Images
The central focus of this paper is to evaluate the percentage of ice covering each lake, hereafter referred to as ice cover ratio, during spring meltout. This procedure is repeated for each year during which a viable satellite image is available. Several processing steps were applied to Landsat MSS, TM, and ETM+ imagery in order to calculate ice cover ratios on a lake-by-lake basis. A semi-automated procedure was implemented to efficiently process imagery for this large and lake-rich area. Images were selected and downloaded from the USGS’s Earth Resources and Observation Science (EROS) data center through the Global Visualization Viewer (GLOVIS). Since this analysis is concerned with lake ice conditions during meltout, all cloud-free scenes were examined that covered the western Arctic Coastal Plain beginning at early snowmelt (typically late May), during lake ice meltout (early June through early July), and into the ice-free season (mid-late July). Final scenes were selected on the basis of several criteria: (1) the scene exhibits a range of lakes in various stages of meltout in order to capture regional patterns; (2) near-coincidence of imagery acquisition dates (typically early July); and (3) overlapping aerial coverage centered on the Barrow Peninsula and regions immediately south.

As listed in Table 1, a total of seven scenes satisfied these criteria, with several supplementary images capturing the change in lake ice conditions over the course of meltout within a single year. The primary limitation encountered when comparing Landsat scenes both within a single year and across the entire ~35 year record is the 16-day revisit time of the Landsat satellite. Coupled with the frequently overcast weather conditions, these factors restrict the ability to capture ice meltout since the process occurs fairly rapidly. However, the subsequent analysis makes use of all available datasets and addresses these issues.

Image processing of these scenes is conducted in two steps: 1) extracting the lake shoreline and estimating lake area and lake ice coverage for each lake in every satellite scene,
and 2) calculating the lake specific ice cover ratio for all lakes exceeding 10 ha. All images were processed in ENVI 4.5 software to map lakes and extract lake ice, and the vector-based files of lake and lake ice extents were imported into ArcGIS 9.3 for spatial analysis.

**Ice Cover Ratio Derivation and Accuracy Assessment**

The boundaries of each lake were identified and defined using the Normalized Difference Water Index (NDWI = (Green - NIR) / (Green + NIR)) (McFeeters, 1996). This method uses the absorptive properties of water and ice, and reflective properties of vegetation from near-infrared wavelengths, to define the border between open water or ice and terrestrial land or vegetation. Simultaneously, the procedure enhances the boundary using the reflected response of water in the visible green wavelength. The segmentation of NDWI yields a polygon that represents the lake shoreline.

Lake ice coverage was determined from the images using an Iterative Self-Organizing Data (ISOData) unsupervised classification algorithm. An unsupervised approach was chosen due to the varying spectral response of ice that relates to differing ice conditions, such as the presence of snow or snow ice, ponded water atop the ice, and/or candel ice. However, the spectral properties of ice are sufficiently distinct from water to optically differentiate between the two materials. Lake shorelines extracted using the NDWI method discussed above were added to the classification as a mask to exclude everything in the image except water bodies. This step was incorporated to both reduce the processing time and to improve spectral identification of lake ice pixels. A simple binary classification scheme was then used to identify either the presence or absence of ice for each pixel within the lake. Image data used as input into the classification algorithm included (i) the three raw Landsat bands in the visible wavelengths, (ii)
the first Minimum Noise Fraction (MNF) band which calculates and separates image noise from all bands while enhancing image coherency and improving processing (Green et al., 1988; Boardman and Kruse, 1994), (iii) a mean co-occurrence texture filter (Haralick et al., 1973), (iv) and a histogram stretch of the blue band for contrast enhancement.

The accuracy of the ice/water classification was assessed by visually interpreting a set of randomly assigned points for each scene since the time-scene dependency and transitive nature of lake ice prevents us from making true ground-truthing measurements and confirming our results. We therefore determined if each randomly selected point contained ice or water. The number of test pixels required for accuracy assessment was adopted from the Binomial Probability theorem. To achieve an accuracy of 95% at an allowable error of 5%, a total of 380 verification points are required (Jensen, 2005). We conservatively used 500 points, where an approximately equivalent number of points were used as ground-truth data for each of the two classes (ice and water). The task of individual point interpretation is time consuming, and was initially applied to only one Landsat MSS scene (6 July 1975) and one Landsat TM scene (1 July 2001). Overall classification accuracy was 95% for 6 July 1975 and 98% for 1 July 2001, with Kappa coefficients of 0.91 and 0.95, respectively.

Qualitative assessment was also carried out for all seven scenes to build confidence in the method; this was done by overlaying the extracted lake ice coverage polygons on the original Landsat scene, and visually scanning the image for any errors. One source of misclassification occasionally occurred along the ice/water boundary. These border pixels likely exhibit a mixed reflectance response, or indicate ice in the final stage of ice degradation (and presumed to be completely thawed within a short time period). These rare occurrences were no more than a few pixels wide and are unlikely to have an impact on the calculation of ice cover ratios.
Additionally, late-lying snow banks along lake margins were occasionally classified as ice, and this was observed far more frequently in the south of the study area where relief is greater and snowdrifts deeper. These snow banks are easily identifiable and were manually removed from all the images prior to calculation of the ice cover ratio. The accuracy assessment results suggest that the processing methods used for classification are acceptable for the different satellite sensing systems and therefore eliminate the need to conduct similar accuracy assessment for every processed scene.

The lake shorelines and the lake ice coverage determined using the above methods were saved as polygon shapefiles and imported into ArcGIS 9.3. All data layers were set to the same reference datum and projection (WGS-84, UTM Zone 5 North). The estimation of lake surface area and lake ice area was conducted using standard software tools. Lake-specific ice cover ratios were then calculated by dividing the lake ice area by the total lake area for each scene listed in Table 1. An example is shown as Figure 1.

**Standardization of Study Area**

The study area extent was standardized to ensure that the same lakes were evaluated each year. First, a standardized study area was derived based on the maximum amount of overlap for all available scenes, resulting in an area of ~11,200 km². All lakes located within major floodplains were excluded from the analysis due to their dissimilar morphometry and behavior (Sellman *et al.*, 1975; Jorgensen and Shur, 2007). The number and total area of lakes in the standardized study area should be similar in all years, but differences are known to result when comparing datasets with sensors of differing spatial resolution (Table 2). There was also computational confusion due to the high lake density. For example, in some cases several
unconnected lakes were classified as one lake because of their extreme proximity, and such errors were resolved manually. Furthermore, some lakes have completely or partially drained or coalesced over the course of the record (Hinkel et al., 2007). For these reasons, there is a slight disparity in the number of lakes and total lake area between scenes, as shown in Table 2. Therefore, a historical dataset using the coarser resolution (6 July 1975) was used to match lake features in all other datasets and then was repeated again using a scene representing more modern conditions (1 July 2001). Figure 2 shows the lakes in the standardized study area used in this analysis, while Table 2 presents the descriptive statistics for all seven years.

**Spatial Interpolation**

Lake centroids were extracted from lake boundaries using the center of gravity. The ice cover ratio for each lake centroid was interpolated onto a 1 km grid using a natural neighbor algorithm. For visualization purposes, a continuous surface of the ice cover ratios was created and contoured using a standard scale for all years. The resulting ice cover ratio maps for six of the seven years are presented in Figure 3. Owing to space limitations, the ice cover ratio map for 1992 is not shown, but mimics the pattern presented in 2006.

**Results and Discussion**

**Interannual Variation In Ice Cover Ratios**

The first impression of the ice cover ratio maps in Figure 3 is their dissimilarity, even though all maps represent the ice cover condition in the first week of July. Distinguishable differences are also noticed across years in the total amount of ice covering lakes within the study area, and in the mean ice cover ratio for each year (Table 2). The ice-cover ratio map for
1984 is very similar to 1975, while 1992 replicates the patterns shown in 2006. Ice cover ratios in 1976 is intermediate between 1986 and 2006. Interpretation of these data demonstrate that lakes do not melt out at approximately the same time every year, but instead demonstrate inter-annual variability in the timing of lake ice meltout over the study area.

Regional Spatial Pattern Of The Ice Cover Ratio

Similarities between scenes are apparent in the overall spatial pattern within the study area. In general, there is a northwest to southeast trending boundary that separates higher and lower ice cover ratios; this is especially noticeable in the 7 July 1986, 1 July 2001 and 18 July 1976 scenes. The boundary is not as pronounced in the other images when most lakes are still ice covered (6 July 1975 and 1 July 1984) or when most lakes are nearly ice-free such as 7 July 2006 and 8 July 1992 (not shown). However, the spatial pattern of the ice cover ratio maps demonstrates the impact of regional climate. Lakes further south experience earlier meltout, much like the snow cover (Zhang et al., 2000).

Spatial patterns in other scenes, covering different spatial extents or acquired on varying dates, also exhibited this regional trend. Due to the necessity of subsetting images to a standardized area, the general patterns shown in Figure 3 do not highlight the magnitude of the trend. Extension of the ice cover ratio calculations further south and into the foothills of the Brooks Range show an even more widespread distribution of lakes that are completely ice-free.

The Maritime Influence

In all scenes, there is a pattern of higher ice cover ratios in the northeast and lower ice cover ratios in the southwest. Lakes that retain ice longer into summer are concentrated along
the northeastern coast, southwest of Admiralty Bay, and near clusters of large lakes such as those observed near the center of the Barrow Peninsula. These patterns reflect the influence of cold, onshore winds from the Arctic Ocean. Climate normals from Barrow show strong (11 knots) and persistent winds from the east and eastnortheast in the spring months (NCDC, 2009), with cloudy conditions dominating. The ocean surface of ice or water is cold relative to the snow-free terrestrial surface. This results in a cool maritime effect that extends 25–40 km inland from the Beaufort Sea coastline (Haugen and Brown, 1980; Hinkel et al., 2004). Note that the pattern is not nearly as strong along the Chukchi Sea side of the Barrow Peninsula, since the winds here tend to be offshore.

An array of temperature monitoring stations was installed in the summer of 2002 along a coast-to-inland transect. As shown in Figure 2, stations are spaced every ~20 km beginning near Barrow and stretch 100 km inland to the village of Atqasuk. Average monthly values, calculated from hourly measurements, demonstrate significantly warmer spring and summer temperatures at sites further inland (Figure 4), while sites on the coast remain relatively cool. At inland sites, thawing degree days in June are greater by a factor of 2-3 while average July air temperatures are 3-6° C warmer than coastal locations. Conversely, average winter (February) temperatures are slightly warmer (1-2° C) near the coast. These data suggest that lake ice preservation is enhanced by a maritime effect along the windward coast.

**Local And Lake-specific Factors**

During a field campaign in early summer 2009, it was observed from the air that several lakes on the Barrow Peninsula were completely free of ice, while ice in surrounding and nearby lakes remained largely intact. The historical archive of satellite images was explored for scenes
acquired earlier in the meltout period to see if this pattern is consistent, and it was determined that several lakes consistently experience complete meltout earlier in the season. Such lakes are characterized by a variety of sizes in surface extent and depth. Two such lakes are partially drained (Jones, 2006) and demonstrate a large seasonal reduction in surface area during summer; this implies that the lakes are very shallow. Lakes typically experience meltout first near the shore to create a moat of open water, so this pattern of shallow water meltout is consistent. However, other early meltout lakes are deeper than the maximum winter ice thickness (Jeffries et al., 1996).

Conversely, a few lakes tend to retain their ice cover longer than nearby lakes. These tend to be larger lakes near the Beaufort Sea coast, as well as farther north on the Barrow Peninsula. A lake-specific change in meltout pattern is also noticed after one lake coalesced with a larger lake complex. From these observations we expect that dominant lake-specific factors impacting early meltout may include basin geometry, lake depth, surface extent, and possibly water chemistry.

**Dynamics Of Ice Cover Decay**

Most studies using lake ice as a proxy for climatic interpretation use the exact timing of initial ice formation and meltout within a specific lake, and analyze associated statistical trends over a multiyear record (e.g., Lake Baikal or Lake Mendota in Magnuson et al., 2000). However, the timing of ice cover formation and disappearance in this study area is not possible given the coarse temporal frequency dictated by the 16-day satellite revisit time. This research attempts to identify regional and local controls on ice degradation over a large study area and to make use of lake ice as a proxy for regional climatic differences.
The rate of meltout for these lakes in this area cannot be specifically addressed with Landsat imagery. The shorter return time and more opportunistic image acquisition characteristics of active sensing systems such as SAR are better suited to this task. However, a few cases of clear sky conditions and overlapping adjacent Landsat scenes provide some insight on regional differences in lake ice dynamics.

During spring 1986, two different Landsat scenes (different path/row) with spatial extents falling within the standardized study area were acquired seven days apart; 30 June and 7 July. A mean center was calculated using the lake centroids and the ice cover ratio as a weighting factor, and the weighted standard deviational ellipses were calculated to describe the directional distribution of the ice cover ratios (Figure 5). Over the 7-day period, the weighted mean center migrated ~25 km to the northeast and the total area of ice covering the lakes was reduced 48%. Concurrently, the ellipse became more elongated, with the major axis becoming parallel to the Beaufort Sea shoreline. This suggests that ice degradation was enhanced in the interior, while ice was preferentially preserved along the windward coast.

An identical analysis was conducted for images east of the Barrow peninsula and acquired 16 days apart in 1984. This geographic area lies outside of the standardized area described in this paper, and thus not listed in Table 2. However, the spatial trends evident in these images are very similar to the patterns described above (and in Figure 5), where regional differences in ice cover during meltout reflect the maritime influence.

Discussion

Long-term approximate freezing and thawing indices were calculated using mean monthly air temperature data collected at Barrow. Freezing degree days during January,
February, and March were compared with thawing degree days in June throughout the historical record to place the years evaluated in this study in the longer-term context. As shown in Figure 6, both 1984 and 2006 experienced warmer-than-normal June temperatures, while winter 1984 was somewhat colder. To a large degree, however, the years analyzed here are distinctly non-anomalous.

Based on basic heat flow theory, results from numerous studies, and the analysis presented here, it is clear that air temperature is a primary control on the lake ice meltout date. Other factors can also have an impact. The timing, duration, and accumulated thickness of snow influences ice thickness owing to its insulating properties, but there are no spatially distributed snow thickness databases for the period of record. Snow depth data collected at Barrow or even at several monitoring sites is not applicable to the greater study area because it does not take into account the presence of strong winds and drifting snow, and is likely not representative of conditions in the ubiquitous topographic depressions such as lakes and drained lake basins. Episodic melting events can also result in net increase of ice thickness, while episodic cooling events result in contraction and cracking of ice allowing water to seep to the surface to form snow ice. Additionally, variations in lake water chemistry may also be a contributing factor during ice meltout, but intensive and detailed measurements are needed to confirm such results.

Zhang and Jeffries (2000) discussed the interannual variation in lake ice thickness in northern Alaska using field data collected near Barrow and temperature and snow depth data collected from the National Weather Station in Barrow. Additionally, Jones et al. (2009) modeled lake ice thickness using air temperature and lake ice thickness measurements from Barrow, as well as measurements from Drew Point and Fish Creek (sites approximately 100 km east of Barrow and along the coast). They concluded that there were not considerable
differences in modeled ice thickness between the sites, but their model inputs of air temperature come from stations that are situated along the coast.

Remotely sensed data, supplemented with measured and modeled maximum lake ice thickness, has also been applied in this area of the WACP to estimate general lake bathymetry (Mellor, 1994; Morris et al., 1995; Jeffries et al., 1996). However, such studies assumed that seasonal lake ice degradation and temperature regimes are similar for lakes across a coast to inland transect, suggesting that the timing of ice formation and degradation for these lakes is similar. Based on the results from this study, in situ measurements of ice thickness from lakes near Barrow or at other coastal locations may not be representative of thousands of inland lakes that occupy the WACP.

The maritime effect discussed here may have implications beyond the timing and dynamics of ice growth and decay. Based on currently published studies and evidence from many field seasons in different regions of the WACP, there appears to be a general spatial pattern in lake depth, where lakes further inland tend to be deeper while lakes near the coast are relatively shallow (Sellman et al., 1975; Mellor, 1984; Jeffries et al., 1996). Such differences are associated with physiographic provinces and can primarily be explained by the topography, relief, surficial geology and ground ice content (Wahrhaftig, 1965; Carter, 1981). Here, we suggest that the maritime gradient might also influence lake basin processes and evolution. For instance, a longer duration of snow-free tundra and open-water period makes inland lakes more susceptible to wave action and shoreline erosion.

Preliminary data support the contention that water in lakes further inland tend to be warmer. Measurements collected during early summer from lakes on the Barrow Peninsula suggest that regional water temperatures are very similar once the ice cover has completely
disappeared, and that the water column is nearly isothermal owing to mechanical mixing by wind, waves and currents. The mean July 2009 temperature of a lake near Barrow was 6.9°C, with ice meltout occurring around 8 July. By contrast, a lake near Atqasuk ~100 km inland had a mean July temperature of 10.2°C and was ice free throughout (J. Lenters, pers. comm.). Warmer water temperatures caused by a longer ice-free season may lead to deeper taliks in sediments beneath the lakes, with consequent ground subsidence and deeper basins.

Furthermore, it is safe to assume that lakes are not the only landscape elements to be affected by this coastal climatic gradient. As illustrated earlier, localized differences in winter and summer temperature regimes, as well as differences in cloud and fog formation and snow-cover depth, are largely influenced by location and proximity to the coast. Such climatic regimes also have larger effects on the duration of season as it relates to vegetation cover, hydrologic processes, and active layer thickness. Most research has been conducted in the vicinity of Barrow, which is often assumed to represent the larger Arctic Coastal Plain. Data collected from this coastal site are often used as input in both arctic and global climate models, and don’t account for the geographic variation described here and elsewhere. We suggest that future research activities in all climate-related fields should explore and incorporate the effects of these first-order controls, and be more geographically integrative in order to accurately extrapolate results across the larger Western Arctic Coastal Plain.

**Conclusions**

Specific conclusions from this analysis include (1) pronounced interannual differences in the timing of lake ice meltout, likely in response to variations in solar forcing, air temperature, snow depth, lake size, and lake ice thickness, (2) a consistent spatial pattern in the timing of
springtime lake ice meltout, dictated largely by latitude and proximity to windward coasts, and (3) a small group of lakes that consistently melt out earlier or later than neighboring lakes in response to local conditions.

General results from this study indicate a rapid and significant alteration of reflective properties of land surface with snow melt. Such findings have greater implications for geomorphological processes, regional energy and hydrologic budgets, and for observing and measuring methane efflux from lakes. A better understanding of the nature of arctic lakes during seasonal transition periods is especially important for a more complete assessment of the Arctic and global carbon and energy budgets, the sustainability of high arctic villages and industrial operations, and for monitoring the diversity and productivity of biotic communities. The methodology applied in this analysis is neither region-specific nor sensor-specific, and can be used to explore regional climate variations as expressed in the lake ice record in other arctic and/or coastal locations. Coupled with a more intensive program of lake ice measurement, as well as continual monitoring of lake ice growth and decay processes using, for example, time-lapse photography, may yield additional insights into landscape processes across this extensive region.

Future Work and Concluding Remarks

While this study utilizes several remote sensing and GIS tools with direct geomorphological and climatological applicability, it does not quite capture the dynamics of change over time. As a direct geographic and remote sensing study, we have been able to achieve an understanding of the dynamics of lake ice coverage over space to a certain degree, and can now direct future efforts more efficiently. However, the available data included in this
study is not sufficient enough for statistical trend analysis of springtime lake ice conditions over the study period. The historic archive of remote sensing products intends to provide such details, but often inhibited from weather conditions or gaps in data coverage due to satellite sensor ownership. However, this study has not reached the finish line. Instead, it is hoped that the inquiry of springtime lake ice conditions will be reviewed on an annual basis using these methods. Moreover, the research design used in this analysis will hopefully spark more investigations, where similar methods can be used in other areas. The study discussed here is innovative for two reasons; 1) calculating the percentage of ice covering a single lake, and evaluating relative percentages of ice for thousands of lakes in the same region and across years 2) and using lake ice coverage as a proxy for regional climatic regimes. Such qualities have shown their usefulness here and should undoubtedly be extended to study areas elsewhere, such as the Seward Peninsula or northern Siberia.

However, there are a few notes to researchers should this type of analysis be initiated in the future. Since, hindsight is always 20/20, I will make these reservations now. The first deals with the processing of imagery in order to derive the ice cover ratios. The method used in this analysis includes a pixel-by-pixel spectral classification for both ice and water. Then a machine learning algorithm was used in order to maximize spectral heterogeneity between classes, while minimizing heterogeneity within classes. This type of hard classification technique has been shown to work well, but theoretically does not capture the dynamics of real world objects, like lakes, for instance. Recent advancements in digital image processing have placed more confidence in the object-oriented algorithm, which goes beyond the spectral properties of individual pixels to include the neighborhood view and classification of objects based on shape indices. In the current scientific literature, the object-oriented segmentation approach to the
satellite remote sensing classification of lakes in the Arctic Coastal Plain of northern AK is credited with being the most accurate mapping approach to date (Frohn et al., 2005). The study conducted by Frohn et al. also utilize the available inventory of Landsat imagery, but still achieve a high level of accuracy. At the time I began this project, I did not have the skills in order to perform a similar assessment using the segmentation approach. However, with some training now, and a greater appreciation for such a method, this paper would greatly benefit from this type of classification approach, in the identification of both lake area and lake ice area. Some errors related to the pixel classification were noted in the methodology section of the article, but would likely be reduced or eliminated using this technique.

Another drawback to this paper, also mentioned in the paragraph above, relates to the ability to sufficiently capture the dynamics of springtime lake ice meltout. The Landsat image archive evaluated in this study was chosen due to 1) its free and accessible electronic archive and 2) its historical archive of yearly images dating back to 1975. Despite Landsat’s continual coverage over ~35 years, scenes from every year were not accessible. Some of these gaps are due to image ownership and licensing issues, with other gaps attributed to weather conditions and infrequent repeat coverage from the satellite. Analysis using this imagery simply provides discrete representations of ice dynamics. Other remotely sensed data sources may be better suited to fit this approach. For instance, synthetic aperture radar (SAR) has the ability to acquire imagery in all weather conditions, including day and night. This dataset provides a more continuous coverage of images (1 every few days), providing a more complete time-series analysis of a dynamic set of processes. SAR is widely used in remote sensing studies of ice environments, especially related to lakes in the WACP (Mellor, Jeffries, Hirose, etc.). The main limitations to SAR imagery are the availability of both the data as well as software packages to
process the data. However, since this project was funded under a National Science Foundation 
grant, we have been given access to a limited number of scenes. Acquiring such images for 
further analysis would benefit this study in several aspects. First, using currently developed 
techniques, it would provide a baseline study with a simple binary classification of lakes that 
have bottomfast ice and lakes that are deeper than maximum winter ice cover. Such a map may 
then provide useful information on the causative factors of early- and late-ice-off lakes, or detect 
any correlation between timing of meltout and lake depth. Secondly, the frequent repeat time of 
the satellite offers enough resolution to observe the ice decay processes which occur over short 
periods of time, whereas Landsat cannot capture such dynamics. Then, by gathering rates of 
decay, and how such rates vary both spatially and temporally, information of much greater detail 
may be derived for model inputs. Results of such a study may significantly enhance models of 
springtime lake ice decay. Once such models become valid, a most-useful product of such 
research would be the prediction of ice-off dates on a lake-specific basis, given depth, latitude, 
and spatially varying climatic controls. Such information would greatly supplement the 
scientific community as it relates to the monitoring of the hydrologic and biologic systems, 
geomorphic evolution of landforms, and enhanced industrial activity. The importance of 
methane efflux from these lakes during springtime meltout has already been emphasized, and 
would also greatly benefit from such additional information.

In conclusion, I hope this article reaches the broader scientific community so to 
emphasize the importance of geographic variation, especially in model upscaling, and the 
advantages of time series analysis in process-based geomorphic surveys. In addition, I hope that 
this paper assumes greater responsibility than just implementing a more spatially intensive 
sampling scheme, but rather to understand the importance of system integration in dynamic
environmental modeling. In the end, such models are only as good as the data that is collected and used for input.

Acknowledgements

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**Table 1:** Primary and ancillary Landsat imagery used in this study.
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<th>Total ice area (km²)</th>
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</tbody>
</table>

**Table 2:** Descriptive statistics of lake ice cover ratios for the given years and standardized study area. Also shown are the degree days of frost (DDF) for January, February, and March, as well as the degree days of thaw (DDT) for June.
Figure 1: Original Landsat TM image on left (8 July 1992) and resulting extracted lake boundaries and lake ice coverage (white) map on right. Also shown are several lake-specific ice cover ratios.
Figure 2: Standardized study area for the multitemporal ice cover ratio analysis (lakes in dark grey). Location of temperature stations, operational since 2002, are also shown as stars.
Figure 3: Continuous surface of spring time ice cover ratios for four years of the standardized study area overlain on Landsat color composites for each year. The spatial pattern in the other three years, excluded due to space limitation, mimic those presented.
Figure 4: Summary temperature data collected along north-south transect at sites shown in Figure 2; (a) June thawing degree days for five years, and mean monthly temperatures and average daily range (bar) for three years during (b) July and (c) February.
Figure 5: Migration pattern of ice meltout over a 7-day period in 1986 using weighted mean center and standard deviational ellipse (SDE) for both dates.
Figure 6: Time series showing degree days of frost during winter (January, February, and March) and degree days of thaw in June using data from National Weather Service station at Barrow. Years included in the analysis are identified by arrow, with median and mean values shown. The dotted line represents the average over the 60-year period.