

LAKE TEMPERATURE AND ICE COVER REGIMES IN THE ALASKAN SUBARCTIC AND ARCTIC: INTEGRATED MONITORING, REMOTE SENSING, AND MODELING¹

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ABSTRACT: Lake surface regimes are fundamental attributes of lake ecosystems and their interaction with the land and atmosphere. High latitudes may be particularly sensitive to climate change, however, adequate baselines for these lakes are often lacking. In this study, we couple monitoring, remote sensing, and modeling techniques to generate baseline datasets of lake surface temperature and ice cover in the Alaskan Subarctic and Arctic. No detectable trends were observed during this study period, but a number of interesting patterns were noted among lakes and between regions. The largest Arctic lake was relatively unresponsive to air temperature, while the largest Subarctic lake was very responsive likely because it is fed by glacial runoff. Mean late summer water temperatures were higher than air temperatures with differences ranging from 1.7 to 5.4°C in Subarctic lakes and from 2.4 to 3.2°C in Arctic lakes. The warmest mean summer water temperature in both regions was in 2004, with the exception of Subarctic glacially fed lake that was highest in 2005. Ice-out timing had high coherence within regions and years, typically occurring in late May in Subarctic and in early-July in Arctic lakes. Ice-on timing was more dependent on lake size and depth, often varying among lakes within a region. Such analyses provide an important baseline of lake surface regimes at a time when there is increasing interest in high-latitude water ecosystems and resources during an uncertain climate future.

(KEY TERMS: lakes; Arctic; Subarctic; water temperature; ice cover; Alaska.)

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INTRODUCTION

The phase of water (i.e., liquid or solid) and its corresponding temperature regime are fundamental attributes of lake ecosystems that directly control energy balance and exchange with the atmosphere

through thermal stratification, evaporation, and ice cover (Bonan, 1995; Doran *et al.*, 1996; Gao and Stefan, 2004; Boehrer *et al.*, 2008), and strongly influences dissolved gas concentrations, biological productivity, and phenology (McDonald *et al.*, 1996; Crump *et al.*, 2003; Mooji *et al.*, 2008). In permafrost environments, lake temperature and ice thickness

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additionally influence thaw bulb (talik) development, thermal erosion, and ultimately surface permafrost degradation (Carson and Hussey, 1962; Sellmann *et al.*, 1975; West and Plug, 2008). Though surface and depth-integrated lake temperatures are basic measurements of limnological analysis, such data are often patchy in time, making retrospective characterization and inter-site comparison of lake physical properties and regimes challenging (Kettle *et al.*, 2004; Oesch *et al.*, 2005). Yet an understanding of these basic lake physical regimes and how they vary seasonally and interannually is needed to evaluate changes in biological habitat, water quality, and ecosystem processes in high-latitude regions (Livingstone *et al.*, 1958; Sibley *et al.*, 2008). There is much interest in reconstructing and predicting lake temperature and ice cover regimes because of observed and projected warming associated with Arctic climate change (Rouse *et al.*, 1997; Livingstone and Adrian, 2009), particularly in the context of observed divergent responses between lakes in continuous and discontinuous permafrost landscapes (Smith *et al.*, 2005; Riordan *et al.*, 2006). Other forcing phenomena of interest include: (1) perturbations to regional weather, such as cloud cover, storm activity, and drought and flooding; (2) regional and global climate cycles, such as El Niño/La Niña, and the Pacific Decadal and North Atlantic oscillations (Serr-eze *et al.*, 2003); and (3) large volcanic eruptions (Briffa *et al.*, 1998). High latitude, lake-rich ecosystems are common (Smith *et al.*, 2007; Arp and Jones, 2009), and given rapid climate change in these regions (Hinzman *et al.*, 2005; Richter-Menge *et al.*, 2008), better development and testing of practical and integrated methods for reconstructing and predicting lake regimes is warranted.

Several designated index lakes in the Arctic have been studied long-term to describe open-water thermal regimes, providing important datasets to understand physical and biological responses to climate [e.g., Toolik Lake in Alaska (McDonald *et al.*, 1996; Crump *et al.*, 2003)]. Yet the vast diversity of lake types and vast expanse of understudied lake districts at high latitudes (Arp and Jones, 2009) leaves much uncertainty in our understanding of basic aspects of lake thermal behavior in Arctic and Subarctic regions (Edmundson and Mazumder, 2002; Kettle *et al.*, 2004; Oesch *et al.*, 2005). Synoptic surveys provide insight into lake responses to variation in landscape position and morphology and when repeated may reveal trends in response to climate. Continuous monitoring efforts, however, clearly show that hourly and daily variability in lake surface temperature (T_{LS}) make comparison of point measurements among years impractical (Hann *et al.*, 1992; Kettle *et al.*, 2004). The continually improving precision, battery

life, and economy of thermistors has greatly increased opportunities for continuous monitoring of lake temperature regimes (Porter *et al.*, 2005), but still can lack spatial and retrospective coverage pertaining to questions of contemporary limnological interest (Kettle *et al.*, 2004). This deficiency suggests the need for retrospective modeling of T_{LS} that can be driven by existing meteorological data (Kettle *et al.*, 2004) and calibrated to and validated by remotely sensed data (Reinart and Reinhold, 2008). Similarly, understanding of lake ice thickness and ice cover regimes could be greatly improved and made accessible to scientists and natural resource managers by a practical integration of *in situ* and remotely sensed measurements with empirical modeling (Bilello, 1964; Jeffries *et al.*, 1996; White *et al.*, 2008; Livingstone and Adrian, 2009). Considering ice cover is essential for studying and managing high-latitude lakes since, although most data are collected when the surface is liquid, lakes are frozen for the major part of the year. Understanding ice dynamics is becoming increasingly important in the Arctic, where winter water-use for ice roads and other industrial and municipal uses is mostly from lakes (Sibley *et al.*, 2008; Jones *et al.*, 2009).

Physical energy-balance models of lake temperature and lake ice are improving (Stefan *et al.*, 1998; Gao and Stefan, 2004), yet basic empirical research has shown that air temperature (T_a) can explain much of this variation in both water temperature (Kettle *et al.*, 2004) and the growth and decay of lake ice (Bilello, 1964; Duguay *et al.*, 2006; Livingstone and Adrian, 2009). This basic relationship can be used to develop simple surface regime models for individual lakes, where both meteorological and lake data exist. Remotely sensed measurements of lake temperature and ice cover offer an important opportunity to extend surface regime models in space and time. The satellite-based sensors Moderate Resolution Imaging Spectroradiometer (MODIS) and Advanced Very High Resolution Radiometer have been used successfully to monitor sea surface temperature (SST) and T_{LS} (Oesch *et al.*, 2005; Reinart and Reinhold, 2008) and can be used to track ice formation and break up for large lakes during cloud-free conditions (Reinart and Parn, 2006). Additionally, synthetic aperture radar (SAR) has also been used to monitor lake ice dynamics and is minimally affected by cloud cover and lack of sunlight during the winter (Jeffries *et al.*, 1996; White *et al.*, 2008). These remote sensing approaches present promising opportunities for monitoring many lakes across large regions, when adequate field validation data exist. Coupling remotely sensed and *in situ* temperature monitoring and ice cover measurements with models driven by meteorological data presents an effective and realistic

approach for constructing baseline datasets. Developing such baseline datasets of T_{LS} and ice cover for a region and the variety of lakes therein, opens important opportunities for retrospective and predictive analysis of responses to climate and other mechanisms of lake change.

In this paper, we selected a set of three lakes in both the Subarctic and Arctic of Alaska with characteristics suitable for integrated surface regime monitoring and modeling and subsequent comparison of regimes by lake attributes, primarily lake bathymetry, drainage area, and regional and landscape setting. This approach required relatively short-term *in situ* monitoring datasets (less than two years) partly overlapping with adjacent longer meteorological station records (eight years). Additionally, one lake in each region was selected with sufficiently large surface area (greater than one km width) for remotely sensed measurements of T_{LS} for summer periods between 2004 and 2008. We used this integrated monitoring and modeling approach to hindcast longer-term, baseline records of T_{LS} and ice cover duration for this set of contrasting lakes. Our primary research objective is to characterize seasonal behavior and interannual variability among lakes, look for any trends over this decadal-scale period, and to compare and contrast two lake-rich regions of Alaska, where lake responses to climate change are expected to be strong and potentially divergent (Smith *et al.*, 2005; Riordan *et al.*, 2006).

STUDY AREA

The two study areas used in this analysis, the Teshekpuk Lake Special Area (TLSA) in Arctic Alaska and Denali National Park and Preserve (DENA) in Subarctic Alaska, are separated by approximately 8° of latitude and up to 600 m elevation (Figure 1). These study areas were selected because of a contrast in climate and physiography and because of closely located meteorological stations with records extending back to 2001 or longer. Additionally, at least one lake in each area was of sufficient size for MODIS derived T_{LS} measurements.

Subarctic Study Area

The DENA study area is located on the northern Alaskan Range toeslope spanning a sharp elevational gradient from the mountainous Kantishna Hills (>1,000 masl) at the maximum late Pleistocene glacial extent and just below the active Muldrow Glacier

to lowland outwash plains created by the glacially fed McKinley and Foraker Rivers. The northern half of DENA is part of the Minchumina Lake District, where thermokarsting and oxbow formation are the dominant lake origin processes (Arp and Jones, 2009). The surficial geology of DENA is highly varied and was importantly shaped by fluvial, eolian, and glacial processes and deposits (Karlstrom, 1964). Permafrost in this area ranges from discontinuous to sporadic or absent at higher elevations or adjacent large river courses.

As part of the U.S. Geological Survey-National Park Service Water Quality Partnership Program, a study of Wonder Lake (WON) was initiated in 2006 and extended to Lakes Minchumina (MIN) and Chilchukabena (CHIL) in 2007. Several meteorological stations were established and are maintained by DENA and the Central Alaska Inventory and Monitoring Network including the Wonder Lake Ranger and McKinley River stations (<http://www.wrcc.dri.edu/denali/>). Long-term National Weather Service stations are located at the village airports of Lake Minchumina on the NW side of MIN and in the Telida, southwest of MIN (Figure 1).

Our Subarctic study lakes are extremely varied in landscape setting, bathymetry, source-waters, and origin. WON is deep, >70 m maximum depth and mean depth of 40 m, and originates in a glacially scoured and moraine-dammed basin and with a relatively small drainage area (Table 1). MIN has a maximum depth of 12 m and mean depth of 9 m and is fed most directly by the glacial Foraker River. CHIL is large and very shallow (<5 m depth) and occurs in a wide valley between low elevation hills opening to where the three rivers join to form the Kantishna River (Table 1). The origins of MIN and CHIL are less certain, but likely relate to their positions between the downslope edge of large glacial outwash plains set against low elevation hills that collect and divert regional surface and groundwater.

Arctic Study Area

The northern TLSA is about midway between Barrow and Prudhoe Bay on the Arctic Coastal Plain (ACP) of northern Alaska along the Beaufort Sea coast (Figure 1). The surficial geology is characterized by organic-rich, glacio-marine silty sediments cemented as permafrost that is continuous and ice-rich (Brown, 1968; Sellmann *et al.*, 1975). The degradation of this near-surface, ice-rich permafrost has resulted in the formation of numerous oriented thermokarst lakes and equally numerous drained lake basins (Hinkel *et al.*, 2003). The northern TLSA is centrally located within the ACP Lake District with a

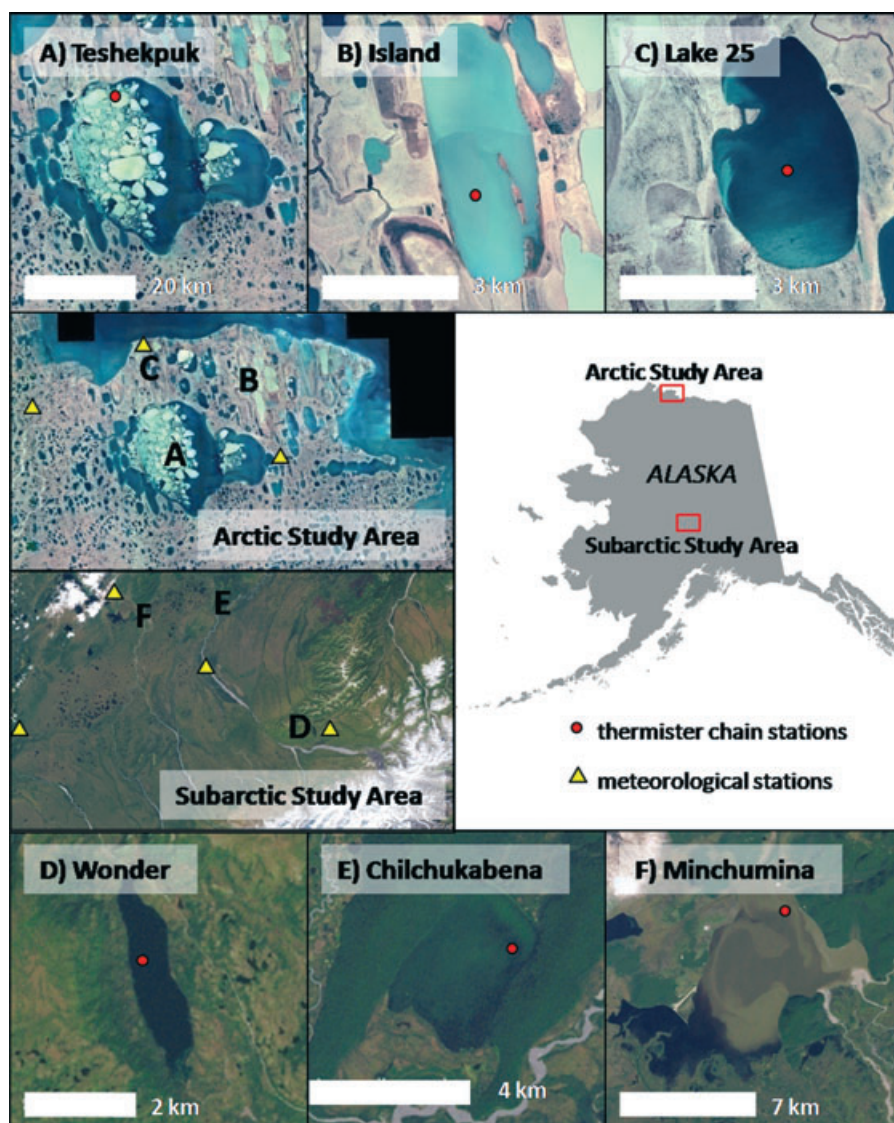


FIGURE 1. Nested Maps Showing the Arctic and Subarctic Study Area Locations in Alaska, Locations of Study Lakes Within Each Study Area, and Color Infrared Photography (CIR) of Each Arctic Study Lake (A-C) and IKONOS Satellite Imagery of Each Subarctic Study Lake (D-F). Locations of lake temperature and meteorological stations are indicated.

lake-area extent of 17% (Arp and Jones, 2009). A U.S. Geological Survey permafrost active-layer monitoring network with associated meteorological stations has been developing here since 1998 and provides high spatial and temporal resolution climate data for the TLSA (<https://my.usgs.gov/akcm/>) (Figure 1).

Our Arctic study lakes all occur near sea level and are relatively shallow, but with differing size, shape, and orientation (Figure 1). Teshekpuk (TESH) is the third largest lake in Alaska at 430 km², but is very shallow, likely <7 m deep, and is of uncertain origin. Island (ISL) is oriented north to south with a maximum depth >2.5 m and a mean depth of ~1.5 m based on observation of bedfast ice distribution. Boat (BOAT) is round in shape and closest to the coast

(<2 km) with much of its area >2 m depth and bedfast ice only along its shoreline (Table 1). Both ISL and BOAT are thermokarst lakes. High turbidity from suspended sediment, expressed by secchi depth, is common, but variable, in many TLSA lakes including these studied (Table 1).

METHODS

Field Data Collection and Monitoring

Thermistor chains were constructed of braided polymer rope, anchored with concrete weights or

TABLE 1. Study Lake Characteristics Relevant to Temperature and Ice Cover.

Lake	Lat	Long	Elev	DA	SA	D_{mean}	D_{max}	SD	DOC	SC
Teshekpuk	70.6	153.6	1	—	430	3	6	1.6	-	260
Island	70.8	153.1	2	39	17.4	1.5	2.5	0.2	-	233
Boat	70.8	153.8	4	13	5.9	2	3	0.4	-	629
Wonder	63.5	150.9	616	17	2.6	40	76	9.8	2.0	195
Chilchukabena	63.8	151.5	191	35	8.3	3	4	0.7	7.7	100
Minchumina	63.9	152.2	196	2,940	51	9	12	2.1	7.5	234

Notes: Lat, latitude in degrees North; Long, longitude in degrees West; Elev, elevation above mean sea level in m; DA, approximate drainage area in km²; SA, lake surface area in km²; D_{mean} , mean depth in m; D_{max} , maximum depth in m; SD, secchi depth measured in late summer in m; DOC, dissolved organic carbon in mg/l; SC, specific conductivity in $\mu\text{S}/\text{cm}$.

sand bags, and held vertically in the water column with a mid-depth and near-surface buoys made of closed cell foam. Temperature was recorded hourly with thermistors (HOBO U22, $\pm 0.2^\circ\text{C}$) placed at ten depths in WON, five in MIN, and three in CHIL. Sensors placed at 1.0 m depth were used to measure hourly T_{LS} and reported in this study, while other lake temperature data for Subarctic sites are located at the National Water Information System (<http://waterdata.usgs.gov/ak/nwis/qw>). All Subarctic thermistor chains were installed through the ice in early April 2007. For Arctic thermistor chains in lakes ISL and BOAT, temperature sensors (HOBO U12, $\pm 0.2^\circ\text{C}$) were placed at 0.5 m depth in April 2008 through the ice – thus near surface sensors were frozen into the ice column until thaw. The TESH thermistor chain is of the same design as used at Subarctic sites and was installed in early September 2007. One thermistor chain was installed at each lake based on combining central locality with avoidance of ice drift and potential movement of sensors during break up. Exact locations of each sensor string are indicated in Figure 1. Even with this cautious placement, the MIN thermistor chain was never recovered after downloading instruments in late August 2007, which we believe was due to movement of large ice pans during break up.

Before placement, each sensor was set in a water tank for one week and the temperature was varied between <1 and $>20^\circ\text{C}$ (monitored with an independent thermometer) to assure accuracy. For Subarctic sites, thermistor chains were downloaded and reset three times at WON and once at CHIL and MIN before removal and 0.5 m vertical temperature profiles were recorded for each site at this time with a multi-probe sonde (YSI 6600, Yellow Springs, Ohio). For Arctic sites, thermistor chains were downloaded and surface temperatures (0.5 m depth) were measured in late August 2008 and redeployed. Hourly T_{LS} records were checked for drift, erroneous measurements, and correspondence with independent field point measurements.

Temperature and Lake Ice Modeling

We estimated mean daily T_{LS} for each study lake using a simple empirical model that requires mean daily T_{a} and theoretical clear sky solar radiation (TCSR, W/m^2) using the multiple linear regression equation (Kettle *et al.*, 2004) (Equation 1):

$$T_{\text{LS}} = a + bf(T_{\text{a}}) + c\text{TCSR} \quad (1)$$

where, a , b , and c are constants (reported in Table 2) determined from the general linear model and f is the exponential smoothing functions developed by relating measurements of T_{LS} to T_{a} , as (Equations 2-4):

$$\frac{dT_{\text{LS}}}{dt} = k(T_{\text{a}} - T_{\text{LS}}) \quad (2)$$

$$\alpha = (1 - e^{-k\Delta t}) \quad (3)$$

$$f(T_{\text{a},t}) = \alpha T_{\text{a},t} + (1 - \alpha)f(T_{\text{a},t-\Delta t}) \quad (4)$$

where, t is time in days ($\Delta t = 1$), k is the heat exchange coefficient, and α is the smoothing parameter in $^\circ\text{C}^{-1}$. There are a number of approaches for estimating TCSR (Annear and Wells, 2007), we selected the Bird and Hulstrom (1981) method because measurements of radiation were not available at all sites as needed for calibration of other TCSR methods. In this method, TCSR is computed as instantaneous hourly values (W/m^2) using a constant atmospheric albedo of 0.08 as suggested by Bird and Hulstrom (1981) and we averaged TCSR for the solar day for regression analysis. The T_{LS} model assumes that major components driving variability can be accounted for by T_{a} and TCSR, an assumption supported for Arctic environments (Dingman, 1972). This

TABLE 2. Lake Surface Temperature (T_{LS}) Monitoring, Modeling, and MODIS Values for Study Lakes.

Lake	Monitoring Period	Weather Stations	Model Coefficients and Results						MODIS	
			α	a	b	c	R^2	RMSE	Scenes	RMSE
Teshekpuk	September 2007-September 2008	Ikpikpuk	0.21	2.98	1.09	-0.005	0.43	0.15	5/22	0.29/0.30
		E.Teshekpuk								
Island	April 2008-September 2008	E.Teshekpuk	0.38	4.16	1.43	-0.007	0.77	0.50	-	-
25	April 2008-September 2008	Drew Point	0.49	4.00	1.47	-0.006	0.50	0.40	-	-
Wonder	April 2007-August 2008	Wonder	0.04	0.80	1.13	0.002	0.94	0.10	-	-
Chilchukabena	April 2007-July 2008	McKinley Minchumina Telida	0.05	-8.31	1.17	0.002	0.95	0.12	-	-
Minchumina	April 2007-August 2007	Minchumina	0.07	-10.77	1.64	0.005	0.97	0.06	8/32	0.75/0.51

Notes: Weather stations are those used to model T_{LS} ; α = smooth coefficient in (Equation 3), a = intercept, b = slope for temperature, and c = slope for theoretical clear sky solar radiation all in (Equation 1), RMSE, root mean squared error; MODIS scenes are the number of dates from 2007/2004-2008, RMSE is for 2007 compared to sensor values and 2004-2008 compared to modeled values.

model approach has also been used successfully for high-latitude lakes in Greenland (Kettle *et al.*, 2004) and alpine lakes in Europe (Livingstone *et al.*, 1999). Meteorological station measurements of T_a used for each lake are given in Table 2.

Lake ice growth and thickness (Z_{ice}) was estimated using a modified Stefan solution (Equation 5):

$$Z_{ice} = \alpha' \sqrt{AFDD} \quad (5)$$

where, α' is the ice growth coefficient and AFDD is accumulated daily freezing degree day in $^{\circ}\text{C}$ (Stefan, 1890; USACE, 2004). Suggested values of α' for lake ice range from 2.4 for snow-covered lakes to 2.7 for snow-free lakes (USACE, 2004). We have little consistent data on lake snow cover and depth for our study lakes, so we estimated α' by fitting Equation 5 to field measurements of ice thickness collected in April 2007 at Subarctic sites and April 2008 for Arctic sites. Summation of AFDD starting each fall was based on three consecutive days of mean daily T_a below freezing and initiated on the third day; anomalous early season cold spells were not used. This day was also designated as the start of lake ice cover, termed ice-on and defined at lake-scales as >50% ice-covered surface.

Lake ice decay was estimated from the Bilello equation (Bilello, 1964):

$$Z_{ice} = Z_{max} - \alpha'' ATDD \quad (6)$$

where, α'' is the ice decay coefficient, Z_{max} is the maximum ice thickness in cm from Equation 5 and ATDD is accumulated daily thawing degree days in $^{\circ}\text{C}$. Here, empirical estimates of α'' range from 0.2 to 2.0 from long-term ice decay studies conducted in Alaska and Canada (Bilello, 1964). We estimated α'' for each lake based on our best estimate of lake ice-out timing (i.e., $Z_{ice} = 0$) using both lake thermister chain data and

MODIS observations for one study year (2007 for Subarctic lakes and 2008 for Arctic lakes). Ice-out timing is defined as <50% ice-covered surface. The summation of ATDD was initiated the day following Z_{max} for each year. We note here that for identification of both ice-on and ice-out timing using *in situ* sensors, we were unable to confirm our ice-cover criteria in all cases, but relied on visual observations of two lakes (MIN and WON) along with remotely sensed MODIS observations detailed in the next section.

Remote Sensing of Temperature and Ice Cover

We used MODIS sensors aboard NASA Terra and Aqua satellites to reconstruct T_{LS} for the largest lakes, TESH and MIN. These satellites acquire data for 36 channels from 0.415 to 14.235 μm in MODIS bands 31 and 32 (11-12 μm , respectively) of the thermal-infrared spectrum, which are suited for determination of skin-surface temperatures in both ocean and freshwater bodies that are of sufficient size for imaging given the 1 km resolution of the MODIS SST (Reinart and Reinhold, 2008) and land-surface temperature products (Crosman and Horel, 2009).

Our analysis and methodology of the MODIS/Terra Level 2 SST products (MOD28L2) follow the more detailed description provided by Reinart and Reinhold (2008). In brief, a manual preselection of cloud-free images over the two study regions was conducted for 2004-2008. The imagery was acquired from NASA's MODIS web, ocean color archive (<http://oceancolor.gsfc.nasa.gov/>) and processed using the Windows Image Manager software (Mati Kahru; WimSoft, Ottawa, Ontario, Canada). In all, 22 scenes for the Arctic study lakes and 34 scenes for the Subarctic study lakes were processed.

Ice cover for our study lakes was monitored to a minimal extent with the SST product, however, a more focused analysis was conducted with the MODIS Rapid Response System (RSS) for our Arctic and

Subarctic study lakes (<http://rapidfire.sci.gsfc.nasa.gov>). A time-series of spaceborne SAR imagery for the Arctic study lakes available over the period also was used to verify observations in the MODIS imagery. Although MODIS is not ideally suited for these types of analysis due to large gaps in available cloud-free imagery to assess ice cover regimes, it was the best available time-series data for both study areas.

RESULTS AND DISCUSSION

Lake Temperature and Ice Regimes in 2007 and 2008

Due to the remoteness and logistical challenges of working at these sites, ice thickness measurements were collected only once during mid-spring, when ice growth was near its maximum. At Subarctic lakes, ice thickness measurements were 73 cm at WON, 110 cm at CHIL, and 90 cm at MIN during thermistor installation around April 1, 2007. At Arctic lakes, ice thickness measurements were 162 cm at TESH and ISL and 137 cm at BOAT on around April 5, 2008. These measurements were used to calibrate an ice growth model based on AFDD, which suggested that maximum ice thickness was reached at the Subarctic lakes several days after measurement, April 3–5, 2007, and not until late May in 2008 at Arctic lakes (Figure 2). The same modeling approach was compared to multiple measurements through the winter in a lake near Barrow, Alaska showing a very close fit to observations, particularly during the late winter and early spring (Jones *et al.*, 2009). The point when ice growth switches to ice decay can be very gradual or abrupt depending on weather patterns and hydrologic responses (Gould and Jeffries, 2005). At lake monitoring stations, this was signaled by a steady rise in water or ice temperature (Figure 3), as has been demonstrated at higher spatial resolution in Subarctic lakes (Gould and Jeffries, 2005).

Ice decay is shown to be primarily a function of ice thickness relative to lake volume driven by and most sensitive to T_a (Gao and Stefan, 2004), such that we expected ice-out to occur earlier in Subarctic and deeper lakes. Ice-out at these lakes was signaled by several days of rising lake temperatures up to 6°C followed by a sharp drop to <4°C and then an irregular increase in temperatures after ice-out (Figure 3). Besides a change in temperature, the shift from a regular diel temperature oscillation under ice cover, to a very irregular diel temperature oscillation due to wind stress and mixing without ice cover, was important for detecting ice-out timing. At Subarctic lakes, ice-out occurred first at lower elevation CHIL

and MIN on approximately May 20 and a week later at WON on May 27 according to sensor measurements. Ice-out at WON and MIN also was verified by independent field observations (J. Caulfield with NPS and P. Green, a volunteer observer for NOAA, personal communications), as well as by MODIS imagery for all three sites, providing greater confidence in using thermistor data to detect ice-out. Still, we note here that ice-out on these lakes occurs over several days to weeks during which time large ice pans freely move across the lake surface (Woo and Heron, 1989; Reinart and Parn, 2006). The estimated duration of ice decay for these Subarctic lakes ranged from 40 days at WON to 46 days at MIN. At Arctic lakes, ice-out occurred on June 25 at ISL, July 3 at BOAT, and July 7 at TESH based on a combination of thermistor data and MODIS imagery. At TESH, the station located near the North shoreline appeared ice free on June 30 (Figure 3D), but the very large area of this lake typically melts out over several weeks or longer. The estimated duration of ice decay for these Arctic lakes ranged from 31 days at ISL, to 43 days at TESH and BOAT, which are deeper (Table 1).

At Arctic lakes, T_{LS} rose very rapidly following ice-out, while Subarctic lake T_{LS} increased more gradually – the difference likely relating to ice-out timing relative to peak solar radiation. The three Subarctic lakes had comparable T_{LS} records from ice-out in late May 2007 until late August 2007. During this comparable period, T_{LS} averaged 13.6°C at WON, 18.8°C at CHIL, and 16.8°C at MIN, which likely corresponds to the much greater depth and higher elevation of WON. Mean T_a for the same period were 12.3°C at WON and 15.9°C at MIN – about 1–2°C lower than T_{LS} . A similar pattern of warmer lake water was described for MIN in 1979 with mean daily T_{LS} peaking at 19°C in late July and early August (Kodama *et al.*, 1983). Reports from European alpine lakes suggest even warmer T_{LS} , up to 7°C higher than T_a (Livingstone *et al.*, 1999). For Arctic lakes, average T_{LS} after ice-out until late August was 8.1°C at TESH, 8.9°C at ISL, and 8.7°C at BOAT. Peak mean daily T_{LS} for these lakes was 10.8°C at TESH on July 27, 14.1°C at ISL on July 10, and 13.1°C at BOAT on July 13 (Figure 3). During this monitoring period in 2008, a strong storm system impacted this area of the coast (C. Arp, unpublished data) beginning on July 31 causing a two-day drop in T_{LS} ranging from 2.0°C at TESH to 3.7°C at ISL with T_{LS} never recovering to pre-storm levels. Such abrupt shifts in T_{LS} suggest the importance of using continuous monitoring or modeling approaches or both rather than point measurements when assessing lake physical conditions (Kettle *et al.*, 2004).

Ice formation was indicated using lake thermistors by a steady decline with irregular diel oscillation to a

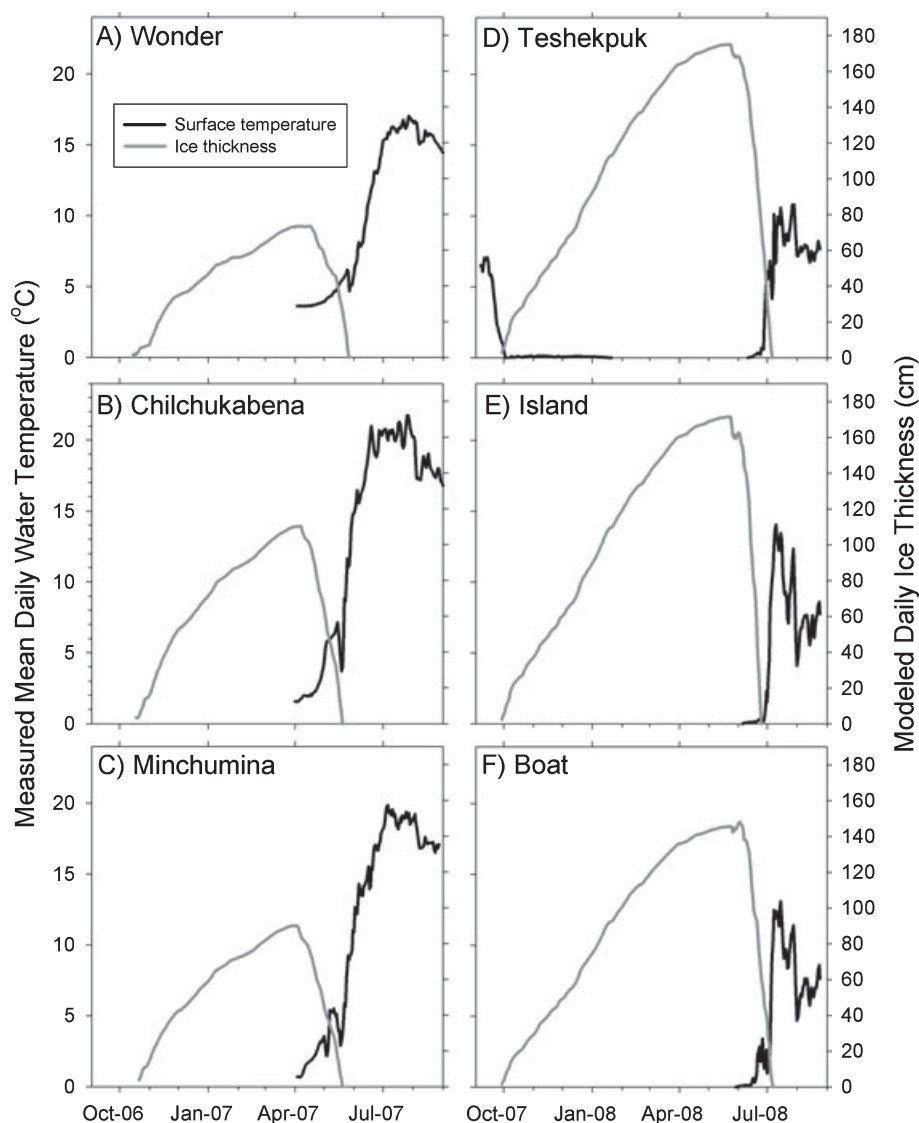


FIGURE 2. Lake Temperature (measured) and Ice Thickness (modeled) Regimes for Subarctic Lakes (A-C) From 2006 to 2007 and Arctic Lakes (D-F) From 2007 to 2008.

temperature approaching 0°C, and depending on sensor depth, to a point at which stable temperature and progressively more regular diel oscillation suggested isolation of the water column from wind stress (Figure 3). However, this ice formation period is known to occur episodically as frazil ice forms and is moved by wind or collects snowflakes that sink or freeze to crystals (Maher and Malm, 2003). MODIS observations helped us bound and verify thermistor data signatures of ice development, however, future work should include visual field observations or use of time lapse cameras. At Subarctic lakes, ice formed much earlier at very shallow CHIL on October 16 than at very deep WON on November 2. In both cases, however, thermistor data suggested that ice formation took about one week (Figure 3B). At TESH,

ice formation appeared more abrupt based on steady temperatures near 0°C starting on October 3 and development of thermal stratification within four days (Figure 3C). Other modeling work suggests that ice formation is more difficult to predict than ice-out (Gao and Stefan, 2004; Reinart and Parn, 2006), such that further development of such monitoring strategies that can be used to improve our understanding of the ice formation process is warranted.

Though these comparisons are incomplete in temporal overlap, as is common in similar studies (Kettle *et al.*, 2004), they provide a picture of how these lake regimes operate and vary within and among geographic regions. Challenging access coupled with the harsh and dynamic nature of many Subarctic and Arctic lakes make the use of autonomous sensor

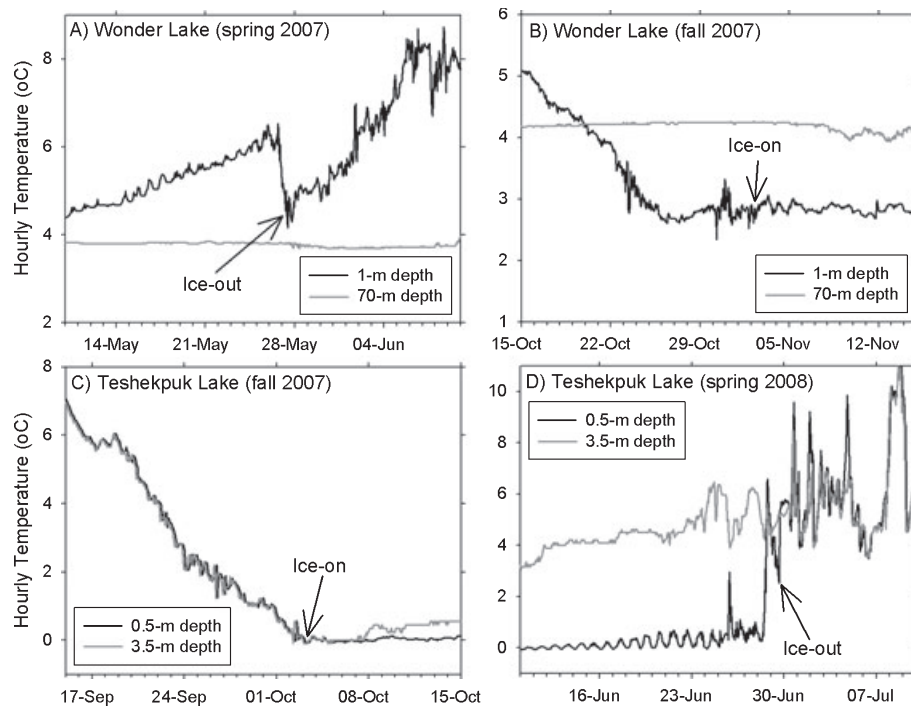


FIGURE 3. Examples of Water Temperature Patterns Indicating Ice-Off (A and D) and Ice-On (B and C) for Two Study Lakes.

networks advantageous for characterizing ecosystem processes because instruments can be deployed and retrieved as logistical and environmental conditions permit (Kettle *et al.*, 2004). However, even well planned and financed lake monitoring networks can have spatial or temporal gaps in datasets, making development of comparable baselines difficult among independent research and monitoring programs. Our study, presents such an example, where we collected temperature and ice thickness data for independent projects with differing field schedules, making direct comparison between Subarctic and Arctic sites impossible (Figure 2, Table 2). Additionally, one thermistor chain, MIN, was lost after five months of data collection – not an unusual occurrence in lake or river monitoring programs without excessive sensor redundancy. Next, we present results from an extension of these short monitoring baselines using a simple model driven by climate data and validated using remote sensing data to extend baseline datasets and make them comparable.

Coupled Modeling and Remote Sensing of Temperature and Ice Cover from 2004 to 2008

Because of the proximity of established meteorological stations to each study lake, we developed simple empirical models of T_{LS} based on measured temperature data from 2007 in Subarctic lakes and 2008 in Arctic lakes. This approach has been used

successfully for lakes in Greenland (Kettle *et al.*, 2004) and alpine Switzerland (Livingstone *et al.*, 1999) and we were thus interested in trying this method in low elevation Arctic lakes and low to high elevation Subarctic lakes in Alaska. It is recognized, however, that using such empirical models may limit analysis beyond the data range (Kettle *et al.*, 2004) and thus we compared modeled values of T_{LS} to measurements from MODIS derived skin-surface temperatures extending back to 2004.

The first step in relating T_a to T_{LS} was determining a smoothing coefficient α that describes how rapidly T_{LS} responded to the differential between T_{LS} and T_a as in Equations 2 and 3. Alpha can range from 0 to 1, with 1 representing an immediate response on a daily time step. Lakes in both regions showed moderated responses to T_a , but to a much lesser extent in Arctic lakes than in the Subarctic (Table 2). The two smaller Arctic lakes had α values of 0.38–0.49, while the much larger TESH was 0.21. Interestingly all three Subarctic lakes had similar and low α values – 0.04 at WON to 0.07 at MIN (Table 2), even though the range in mean depths among these sites varied greatly (Table 1). For 14 Greenland lakes, α values ranged from 0.09 in a relatively large deep lake to 0.37 in a relatively small shallow lake (Kettle *et al.*, 2004). The T_a coefficient, b , in Equation 1 describes T_a sensitivity and our values ranged from 1.09 at TESH to 1.64 at MIN, whereas Kettle *et al.* (2004) reports b values <1 for most lakes. Because of the large size of MIN and large flux of

glacial meltwater during the summer, it seems likely that T_a is actually responding to T_{LS} as has been demonstrated by earlier research (Kodama *et al.*, 1983). The models for all Subarctic lakes explained a high proportion of variability, $R^2 > 0.93$, indicating a relatively strong and stable relationship between T_a and T_{LS} . Arctic lake T_{LS} were less closely related to T_a , particularly at TESH ($R^2 = 0.43$, Table 2). This likely relates to higher variability common in weather patterns and associated wind directions on the Alaskan ACP – variable wind stress also was noted as a confounding factor when modeling Greenland lakes (Kettle *et al.*, 2004). Another notable difference in T_{LS} models between regions was that the TCSR coefficient, c , was positive for Subarctic lakes, as would be expected, but negative for Arctic lakes (Table 3). This difference likely relates to the occurrence of ice-out of Arctic lakes after peak solar radiation in 2008, as happens in most years, while Subarctic lakes had been ice free for nearly a month before this peak in 2007. Coastal fog conditions are also common along the ACP through the summer, strongly attenuating solar radiation and potentially limiting the usefulness of TCSR in our model near the Arctic coastline (Annear and Wells, 2007). Additionally, we used a different TCSR model (Bird and Hulstrom, 1981) than in Kettle *et al.* (2004) based on more recent analysis and model comparison by Annear and Wells (2007), which may limit direct comparison of model coefficient in this case.

Overall model performance was evaluated for the calibration year by the root mean squared error (RMSE), which ranged from 0.1°C at WON and MIN to 0.5°C at ISL (Table 2). Comparison of observed and predicted T_{LS} values through the summer at Subarctic lakes showed small and consistent differences for MIN (Figure 4A) and CHIL, while predicted values for WON became increasingly close to observed values in the late summer. At TESH, predicted T_{LS} was consistently higher than observed T_{LS} during July and lower than observed during August (Table 4 and Figure 4B). For the other two smaller Arctic lakes, the opposite pattern was observed during July and August. This differing

pattern of model performance among these lakes likely relates to how TESH responded to weather patterns and because of its large size, likely moderating strong shifts in T_a .

To see if we could reasonably extend this empirical model beyond the development year, we compared MODIS derived T_{LS} measurements made at two larger lakes in each region, MIN in the Subarctic and TESH in the Arctic, to measured and modeled T_{LS} for 2007 and 2008, respectively. MODIS skin temperatures measured using a satellite radiometer were compared to bulk temperatures measured at 0.5–1.0 m depth, which depending on wind and mixing conditions may be similar or very different (Reinart and Reinhold, 2008). At MIN, all MODIS derived values were within 1.3°C of thermistor measurements and showed little bias ($\Sigma \text{Predicted} - \text{Observed}$) of 0.4°C, except a value that was 5°C higher than recorded on May 23 at 12:45, two days after ice-out (Figure 4A). At TESH, MODIS derived values were within 0.2°C of thermistor measurements and showed no bias (0°C), except a value that was 1.7°C lower than recorded on July 12 at 22:42, 13 days after local ice-out and 6 days after total lake ice-out (Figure 4B). Similar accuracy and precision has been demonstrated for two lakes in Sweden using MODIS measurements (Reinart and Reinhold, 2008). In addition to cloud interference, another major limitation to MODIS for many applications is the relatively coarse resolution (1 km) of the SST product that only allowed us to measure T_{LS} on the two largest lakes in this study. Observations from these very large lakes, however, presented an additional opportunity to compare mean T_{LS} for each image to the variation in temperature across the entire lake surface, where other work using MODIS measurements has described considerable spatial variation (Reinart and Reinhold, 2008). Both lakes have heterogeneous lake bathymetry with smaller shallow subbasins, coupled with multiple inflowing river systems that influence temperatures and mixing (Figure 1). This heterogeneity presented an opportunity to better understand to what extent T_{LS} can vary spatially, providing important information about lake hydrodynamics and

TABLE 3. Summary of the Timing of Lake Ice-Out and Ice-On Estimates From *In Situ* Sensors, Numerical Models, and Observation From MODIS Satellite.

Lake	Year	Ice-Out Timing			Ice-On Timing		
		Sensor	Modeled	MODIS	Sensor	Modeled	MODIS
Teshekpuk	2008	June 30	July 6	June 29 to July 8	October 3	October 1	October 5 to October 11
Island	2008	July 1	June 25	June 23 to July 3	-	September 27	September 20 to October 5
Boat	2008	July 3	July 3	June 29 to July 4	-	October 1	September 20 to October 5
Wonder	2007	May 27	May 27	May 23 to May 29	November 3	October 5	October 13 to October 14
Chilchukabena	2007	May 21	May 20	May 18 to May 23	October 9	October 5	October 10 to October 11
Minchumina	2007	May 21	May 20	May 18 to May 23	-	October 3	October 13 to October 14

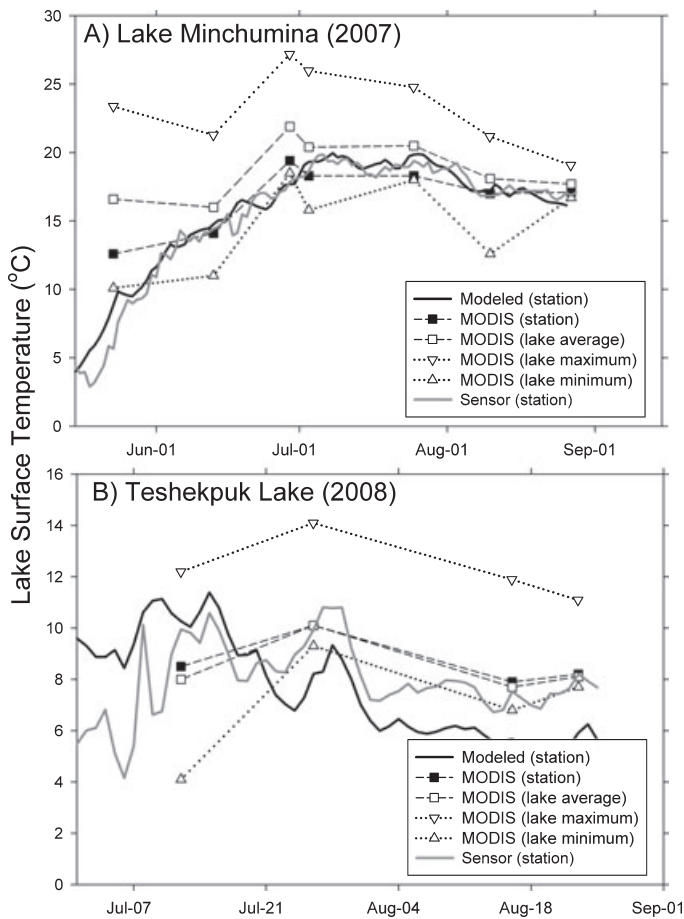


FIGURE 4. Comparisons of *In Situ* Measured, Remotely Sensed, and Modeled Lake Surface Temperatures for Lake Minchumina in 2007 (A) and Teshekpuk Lake in 2008 (B).

TABLE 4. Summary of Lake Surface Temperature (°C) Numerically Modeled and Measured From MODIS at the Sensor Monitoring Station, and Measured From MODIS for Each Entire Lake.

	Modeled	MODIS				
Month	Station	Station	Mean	Minimum	Maximum	<i>n</i>
Teshekpuk						
July	8.2	7.6	6.4	4.2	12.3	11
August	9.4	10.2	10.0	9.2	14.0	6
September	6.1	5.7	5.5	4.4	7.4	4
Minchumina						
June	13.2	14.6	15.8	13.0	20.2	10
July	18.7	18.0	19.3	17.0	23.2	8
August	14.6	15.7	16.4	14.5	18.6	6
September	10.9	12.9	12.8	11.4	13.9	4

Note: *n* = number of image measurements per month during the period of 2004–2008.

energy balance, as well as the importance of station placement for lake monitoring programs. At TESH, station values were very similar to average values for the entire lake in 2008 (Figure 4B) and similarly for

the period from 2004 to 2007 (Table 4). At MIN, station values were consistently lower than the whole lake average in 2008 (Figure 4A) and also from 2004 to 2007 (Table 4) likely due to the position of our thermistor chain in the deep basin of the lake receiving discharge more directly from the glacial Foraker River (Figure 1).

This comparison suggests a close and generally unbiased agreement among thermistor-measured, modeled, and MODIS derived T_{LS} for both Subarctic and Arctic lakes, giving us confidence to extend our analysis further and make some longer-term retrospective comparisons of T_{LS} , along with ice-cover timing, among lakes and between regions in Alaska. We also suggest that this coupling of *in situ* monitoring, remote sensing, and modeling of T_{LS} provides a reasonable and practical framework for monitoring and analysis of lake temperature patterns across a diversity of high-latitude ecosystems. A similar approach was used in Greenland, which took this a step further by estimating model parameters and developing a general lake T_{LS} model based mainly on lake morphometry to estimate T_{LS} regimes for lakes without empirical datasets (Kettle *et al.*, 2004). Our contribution here is extending the time scale of T_{LS} analysis among two distinct high-latitude lake-rich regions, while also including ice cover duration – which is greater than half the year in both regions studied.

Subarctic and Arctic Lake Temperature and Ice Regimes from 2001 to 2008

Based on these models and measurements of T_{LS} and ice cover, we developed estimates of the complete baselines for lake surface regimes for six lakes in Subarctic and Arctic Alaska for an eight-year period. Comparison of these lake regimes allows for the analysis of temporal coherence, which is the degree of correlation or synchronicity in ecological responses among lakes within and between regions (Kling *et al.*, 2000; Carpenter *et al.*, 2007). We also looked for potential trends due to climate oscillations or directional change, albeit a relatively short period for such analysis.

We expected very different thermal regimes among our three Subarctic lakes because of differences in elevation and landscape position, source waters and hydrologic fluxes, and depths (Table 1). Estimated maximum ice thickness for this period ranged from 71 cm at WON to 102 cm at CHIL with the overall greatest ice growth occurring in 2006 and 2007 and the least in 2008. Despite some variation in ice thickness among lakes and years, estimated ice-out timing appeared to be relatively similar with average occurrence on May 26 and ranging from as early as

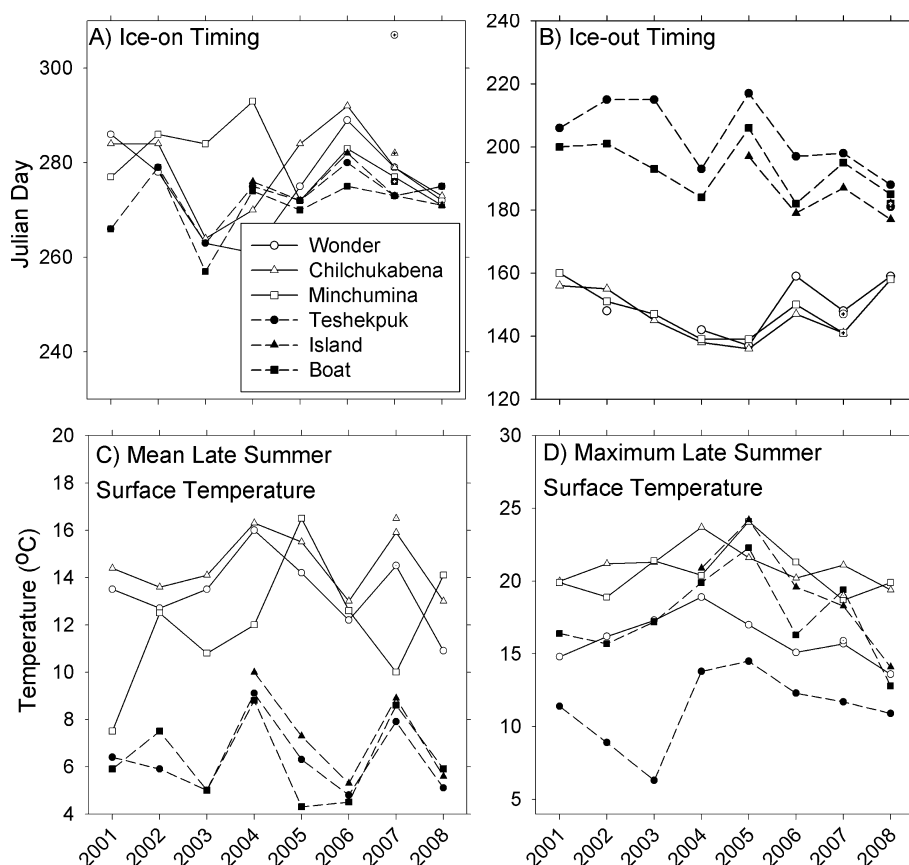


FIGURE 5. Interannual Patterns in Modeled Ice-Out (A) and Ice-On (B) Timing and Late Summer (August 15 to September 15) Mean (A) and Maximum Surface Temperature of Study Lakes From 2001 to 2008. Measured values are indicated by a dot or cross for each site symbol.

May 15 in 2005 to as late as June 8 in 2001 (Figure 5A) with high regional coherence among lakes ($r = 0.77$). These results follow more extensive work in the upper Midwest and southern Canada suggesting coherent timing of ice-out that can be readily predicted from T_a data (Gao and Stefan, 2004). For T_{LS} , interannual patterns averaged for late summer (August 15 to September 15) were very similar between WON and CHIL ($r = 0.94$) with the lowest values in 2008 and highest values in 2004, but MIN showed very different interannual patterns than the other Subarctic lakes (Figure 5A). AT MIN, late summer mean T_{LS} was lowest during the warm summer of 2007 and highest during the moderate summer of 2005 with no correlations to WON ($r = -0.19$) and CHIL ($r = -0.08$). This very different thermal response is likely related to the important glacial source-water contribution to MIN with increasing discharge of cool meltwater during warmer years and decreasing discharge during cool years. A more extensive survey of lake temperature regimes also distinguished lower water temperatures and lack of stratification in Alaskan glacially fed lakes (Edmundson and Mazumder, 2002). Still, WON and CHIL are

at very different elevations and landscape positions with completely different bathymetry and source waters, yet followed very similar interannual T_{LS} patterns. Estimates of ice-on timing among these lakes also suggested that MIN diverged from the other Subarctic lakes. At WON, however, modeled ice-on was about a month earlier than measured most likely because of the much greater heat storage that was not accounted for in our model – a well established modifier of lake ice formation (Gao and Stefan, 2004).

For our set of Arctic lakes, we generally expected that TESH, with its very large size and multiple inflows, would make its behavior unique from other lakes in the area that are predominantly oriented thermokarst lakes. Otherwise, these Arctic lakes are within a relatively small area, all near sea level, and are very shallow without stratifying in most summers. Estimated maximum ice thickness for this period averaged much lower at BOAT, 150 cm, compared to 173 cm at ISL and 177 cm at TESH, because of slightly higher winter T_a near the coast. The thickest annual ice maximum averaged among these lakes was in 2006 and the thinnest annual ice maximum was in 2003 with no trend observed toward thinning

or thickening lake ice. Using similar methods to estimate ice growth for the Alaskan ACP over a longer time period, 1985–2007, a slight trend toward decreasing and more interannually variable ice thickness has been described (Jones *et al.*, 2009). Patterns of estimated ice-off time were very coherent among lakes during this period ($r = 0.91$) and consistently earliest at ISL and latest at TESH with average ice-out dates of 3 and 21 July, respectively. A weak decreasing trend of three days/year in ice-out timing was noted at TESH ($R^2 = 0.39$, $p < 0.10$) albeit for a short period of analysis (eight years), while any trend in ice-out timing at other Arctic lakes were not apparent (Figure 5B). Using a much more spatially and temporally comprehensive dataset, Duguay *et al.* (2006) did find a clear trend toward earlier ice-out timing throughout Arctic and Subarctic Canada. Interannual patterns in T_{LS} averaged for late summer were very similar among all lakes ($r = 0.87$) with the lowest values in 2006 and highest values in 2004. Late summer daily maximum T_{LS} were consistently much higher at BOAT and ISL averaging 18.5°C compared with TESH averaging 11.2°C. Peak mean daily T_{LS} were all estimated to occur in 2005 and ranged from 14.5°C at TESH to 24.2°C at ISL both on August 13 (Figure 5B). Estimates of ice-on timing among these lakes also suggested a high degree of coherence among ice-on dates ($r = 0.75$) with average occurrence at all lakes on September 28 and ranging from as early as September 13 at BOAT in 2003 to as late as October 8 at ISL (Figure 5B). Based on MODIS observations and theoretical behavior of the much larger heat storage in TESH, it is likely that this lake develops a permanent ice cover later than modeled estimates (Table 3).

Some relevant comparisons of interest can be made between lake surface cover regimes between these geographically distant and climatically distinct regions in Alaska with respect to magnitude, timing, and temporal coherence. Ice was on average 81 cm thicker on Arctic than Subarctic lakes at its winter maximum with little regional coherence ($r = +0.33$), which in the case of our ice growth estimates is only a function of similar winter T_a patterns. The difference in ice-out timing between regions averaged 49 days earlier in Subarctic than Arctic lakes. Late summer T_{LS} was 7.0°C higher for Subarctic lakes – the greatest difference in 2005 and smallest difference in 2007 (Figure 5A) with some coherence between these regions ($r = +0.57$) and a higher amount when the larger lakes MIN and TESH were not considered ($r = +0.73$). Mean late summer (August 15 to September 15) T_{LS} were always higher than T_a for the same period with differences ranging from 1.7 to 5.4°C in Subarctic lakes and from 2.4 to 3.2°C in Arctic lakes (Figure 6). Ice-on timing

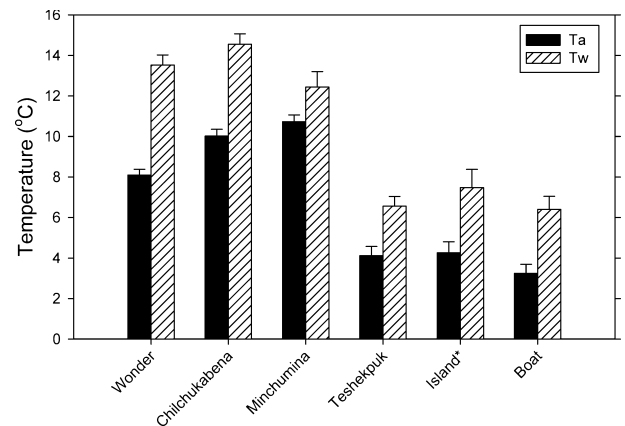


FIGURE 6. Mean Late Summer (August 15 to September 15) Lake Surface and Air Temperatures for the Period From 2001 to 2008 Compared Among Each Study Lake With Bars Indicating Standard Errors.

appeared to be very similar among these regions averaging only six days later at Subarctic lakes with slight coherence ($r = 0.38$). However, given some of our uncertainty in predicting the ice-on process for WON and TESH due to their very large water volume and heat storage, we suggest that the timing in this important lake event among regions is unrelated. Research to better understand the ice initiation process and predict how it varies among lakes would help greatly in modeling ice cover regimes for high-latitude lakes.

SUMMARY AND CONCLUSIONS

In this study, we developed an eight-year baseline of surface temperature and ice cover regimes for a set of Subarctic and Arctic lakes, which allowed us to analyze patterns of inter-annual and inter-regional variability and look for potential trends due to a changing climate. In both regions, late summer water temperatures were consistently higher than air temperatures, ranging from 2 to 5°C greater in Subarctic lakes and from 2 to 3°C greater in Arctic lakes; emphasizing the important role that lakes play in heat storage across high-latitude regions. Considerable interannual variability and temporal coherence in surface temperature was noted among lakes with highest mean values occurring in 2004 for both Subarctic, 16°C, and Arctic, 10°C, lakes and near lowest mean values occurring in 2008 for Subarctic, 12°C, and Arctic, 5°C, lakes. An exception to this coherent pattern of behavior was a Subarctic glacially fed lake that had highest temperatures in 2005 and lowest temperatures in 2001 with similar deviations from

other lakes in ice cover timing as well, suggesting that variation in source-waters among lakes may be more important than setting or size in driving surface behavior. Otherwise ice-out timing had high coherence within regions and years, typically occurring in late May for Subarctic lakes and in early-July for Arctic lakes. Ice-on timing was more dependent on lake size and depth, often varying widely among lakes within a region, though it is also more difficult to predict. A key feature of this research in describing high-latitude lake regimes is our treatment of ice cover duration, which our analysis suggests lasted on average for 64% of the year for Subarctic lakes and 79% of the year for Arctic lakes – a long and critical period of time when little scientific work is traditionally conducted. Analysis of how this ice-cover period may be changing among lakes and regions represents an important topic for future research.

Understanding full and multiple years of lake surface regimes through an integrated monitoring, remote sensing, and modeling approach represents a simple, but powerful case study demonstrating how combining these methods can provide a comprehensive baseline for high-latitude lakes, whereas use of each alone has limitations. For example, maintaining long-term lake monitoring programs are logistically challenging and costly in places like the roadless ACP or Interior Alaska. Remote sensing of lake surface conditions offers a means for making landscape-scale measurements and comparisons for points in time, but often limits understanding regimes or processes. With sufficient input data, lake models can predict continuous temporal behavior, but with much uncertainty without datasets for model calibration and validation. Our approach uses a relatively short, one year, *in situ* lake monitoring records to calibrate longer, eight year, models of lake surface behavior that are validated with remote sensing – effectively capitalizing on the advantages presented by each method alone. Such an integrated approach to lake research can provide a basic, but critical baseline of lake surface regimes and interannual variability at a time when this information is increasingly valuable for improving both scientific and management efforts of water ecosystems and resources during an uncertain climate future.

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