

RESEARCH ARTICLE

10.1002/2016JG003729

Key Points:

- Extensive survey of 617 lakes in Alaska
- The physical and chemical properties of lakes varied by lake origin, parent material, and permafrost characteristics
- Vulnerability of lake ecosystems to climate change will vary across lake basin types, as mediated by ground ice content and subsurface hydrology

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Citation:

Larsen, A. S., J. A. O'Donnell, J. H. Schmidt, H. J. Kristenson, and D. K. Swanson (2017), Physical and chemical characteristics of lakes across heterogeneous landscapes in arctic and subarctic Alaska, *J. Geophys. Res. Biogeosci.*, 122, 989–1008, doi:10.1002/2016JG003729.

Received 23 NOV 2016

Accepted 5 APR 2017

Accepted article online 18 APR 2017

Published online 30 APR 2017

Physical and chemical characteristics of lakes across heterogeneous landscapes in arctic and subarctic Alaska

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Abstract Lakes are an important component of high-latitude regions, providing habitat for fish and wildlife and playing a critical role in biogeochemical and global carbon cycles. High-latitude lakes are sensitive to climate change, in part due to their development within permafrost soils. Considerable heterogeneity exists across arctic and subarctic landscapes, yet little is known about how this landscape variability influences chemical and physical attributes of lakes. We investigated the physical and chemical limnology of 617 lakes in Alaska's boreal forest and boreal-arctic transition zone. We categorized lakes into 10 basin types based on parent material, topography, genesis, and permafrost characteristics. Physical parameters varied across lake basin types, with the deepest lakes occurring in ice-poor glacial deposits and ice-rich terrain, while the shallowest lakes were observed in floodplain deposits and coastal lowlands. Dissolved inorganic nitrogen (N) and phosphorous (P) concentrations were generally low across all landscapes, whereas total N and P were highest in lakes underlain by ice-rich Pleistocene loess. Total N and P concentrations were significantly correlated with chlorophyll *a*, indicating a possible colimitation of primary productivity in these systems. Base cation concentrations helped elucidate lake basin hydrology and the relative influence of shallow versus deep groundwater inputs to surface water. Using these results, we developed a simple conceptual model for each lake and landscape type based on differences in physical and chemical parameters. Overall, we expect that the vulnerability of lake ecosystems to climate change will vary across lake basin types and will be mediated by spatial patterns in permafrost characteristics and subsurface hydrology.

1. Introduction

Lakes and ponds cover nearly 5.9×10^7 ha in northern high-latitude regions (above 45.5°N latitude) [Smith et al., 2007]. These waterbodies play an important role in both providing habitat for fish and wildlife and influencing regional and global biogeochemical cycles [Abnizova et al., 2012; Wik et al., 2016]. Recent warming in arctic and subarctic regions has altered lake ecosystems and basins through increasing lake water temperature [Hinzman et al., 2005], permafrost thaw [Jorgenson et al., 2006; Romanovsky et al., 2010], thermal erosion [Lantz and Kokelj, 2008], and shifting hydrology and chemistry [Lewis et al., 2015]. These climate-driven factors have contributed to shifts in lake biological communities [Michelutti et al., 2005; Smol et al., 2005] and ecosystem function [Wrona et al., 2006]. Despite many advances in understanding arctic and subarctic lake ecosystems over the past decade [Vonk et al., 2015], considerable uncertainty remains regarding the trajectory of lakes across space and time. Arctic and subarctic lakes drain heterogeneous landscapes that may vary in their vulnerability to climate drivers and disturbance. Spatial variability in lake basin properties will likely determine the timing, magnitude, and nature of climate impacts to lake ecosystems.

Spatial variability in permafrost is an important factor driving lake distribution and dynamics in high-latitude regions. Permafrost extent (continuous versus discontinuous versus isolated/sporadic zones) varies with climate and latitude, edaphic characteristics, vegetation, and landscape position [Jorgenson et al., 2008b] and can influence the geographic distribution of lakes [Smith et al., 2007] and lake area trends [Smith et al., 2005; Yoshikawa and Hinzman, 2003]. For instance, Smith et al. [2005] observed a long-term increase in lake area in the continuous permafrost zone and a decrease in the discontinuous permafrost zone in part because of variable thickness and spatial extent of permafrost. Spatial variability in ground ice content is also an important factor driving lake formation. Thermokarst, or the thawing of ice-rich soils followed by ground

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subsidence, is responsible for the formation of lakes across the northern permafrost region [Jones *et al.*, 2011; Jorgenson *et al.*, 2007]. Active thermokarst development along lake margins and subsequent bank erosion can alter lake chemistry [Kokelj *et al.*, 2005, 2009]. Permafrost configuration and thermal state can also influence lake basin hydrology, determining both subsurface inflows and possible drainage pathways [Jepsen *et al.*, 2013; Yoshikawa and Hinzman, 2003]. Further, permafrost-driven controls on hydrology can influence the chemical composition of surface waters [Frey and McClelland, 2009; O'Donnell *et al.*, 2012]. Recent warming trends have caused warming and thawing of permafrost across the Arctic [Romanovsky *et al.*, 2010], and widespread loss of permafrost is projected over the next century [Lawrence *et al.*, 2008; Marchenko and Romanovsky, 2007; Panda *et al.*, 2014a, 2014b, 2016]. Thermokarst lakes are expected to increase in areal extent in response to thawing of ice-rich permafrost [Olefeldt *et al.*, 2016], yet little is known about basic limnological characteristics of these lakes across different landscape types.

Numerous limnological inventories have been conducted to document physical and chemical conditions found in Alaskan lakes [Eilers *et al.*, 1993; Gregory-Eaves *et al.*, 2000; Heglund and Jones, 2003; Kling *et al.*, 1992; LaPerriere, 2003; LaPerriere and Jones, 2002; LaPerriere *et al.*, 2003; LaPerriere and Simpson, 2003; Lewis *et al.*, 2015; O'Brien *et al.*, 1997; Roach *et al.*, 2011]. In general, these synoptic studies document a high degree of spatial variability among lake characteristics across heterogeneous landscapes, as well as among lakes within close proximity. These studies were largely confined to a small assemblage of lakes, within a narrow geographic range, capturing only a small portion of the overall landscape variation. This approach, though informative for a particular region, sheds little light on the large landscape- and regional-scale variation that exists. Given the diversity and spatial heterogeneity of high-latitude ecosystems, it is challenging to make generalizations about lake structure and function or possible outcomes for lake ecosystems under projected climate warming. In order to accurately interpret and predict how Alaska's lake ecosystems will respond to climate change, it is necessary to identify and explain spatial patterns in lake dynamics and possible landscape controls on these patterns. An important step forward is to establish better linkages between surface water and lake basin properties, as determined by permafrost configuration, ground ice content, geologic composition, and thermal state. These permafrost characteristics strongly influence lake chemistry and are actively changing in response to climate and other disturbances.

To better understand landscape controls on physical and chemical limnology of arctic and subarctic lakes, we studied lakes over a large spatial extent throughout six National Park Service (NPS) units distributed across central and northern Alaska between 2006 and 2013. We sampled 617 lakes, recording a suite of physical and chemical characteristics. We then used these values to classify lakes based on physiography (e.g., upland versus lowland), lake origin (e.g., thermokarst, glacial, and floodplain), permafrost extent, and ground ice content (e.g., ice rich versus ice poor) and identify possible controls on spatial variation in the lake characteristics. Using field observations, lake chemistry, and mapping techniques, we also developed basic hydrologic-biogeochemical conceptual models for each lake class. We used these models to assess heterogeneity in lake basin structure and function across arctic and subarctic Alaska, and in turn, the potential impacts climate and disturbance may have on the diverse array of lakes on this complex landscape.

2. Materials and Methods

2.1. Study Area

The NPS units comprising our study area included Bering Land Bridge National Preserve, Noatak National Preserve, Kobuk Valley National Park, Yukon-Charley Rivers National Preserve, Denali National Park and Preserve, and Wrangell-St. Elias National Park and Preserve. These land management units are located in northern and interior Alaska and span a latitude range of 61–68°N and longitude range of 141–153°W (Figure 1). They cover 13.3×10^7 hectares and include approximately 4.4×10^4 individual lakes (Table 1) covering 20.9×10^4 ha (Table 1). The lakes within these units span 13 ecoregions [Nowacki *et al.*, 2001], from the Alaska and Chugach Mountain Ranges north across the intermountain plateau and into the Brooks Range. The parks located in interior Alaska span three climate zones (central interior, northeast interior, and southeast interior) between the cordillera [Bieniek *et al.*, 2012]. These zones have continental climates with cold, dry winters and warm summers. Three parks lie in northwestern Alaska, one of which, Bering Land Bridge, borders the Bering and Chukchi Seas and has a strong coastal influence, with a moderate temperature regime including cooler summers and milder winters (Table 1). Since the mid-1900s, the park units in our study

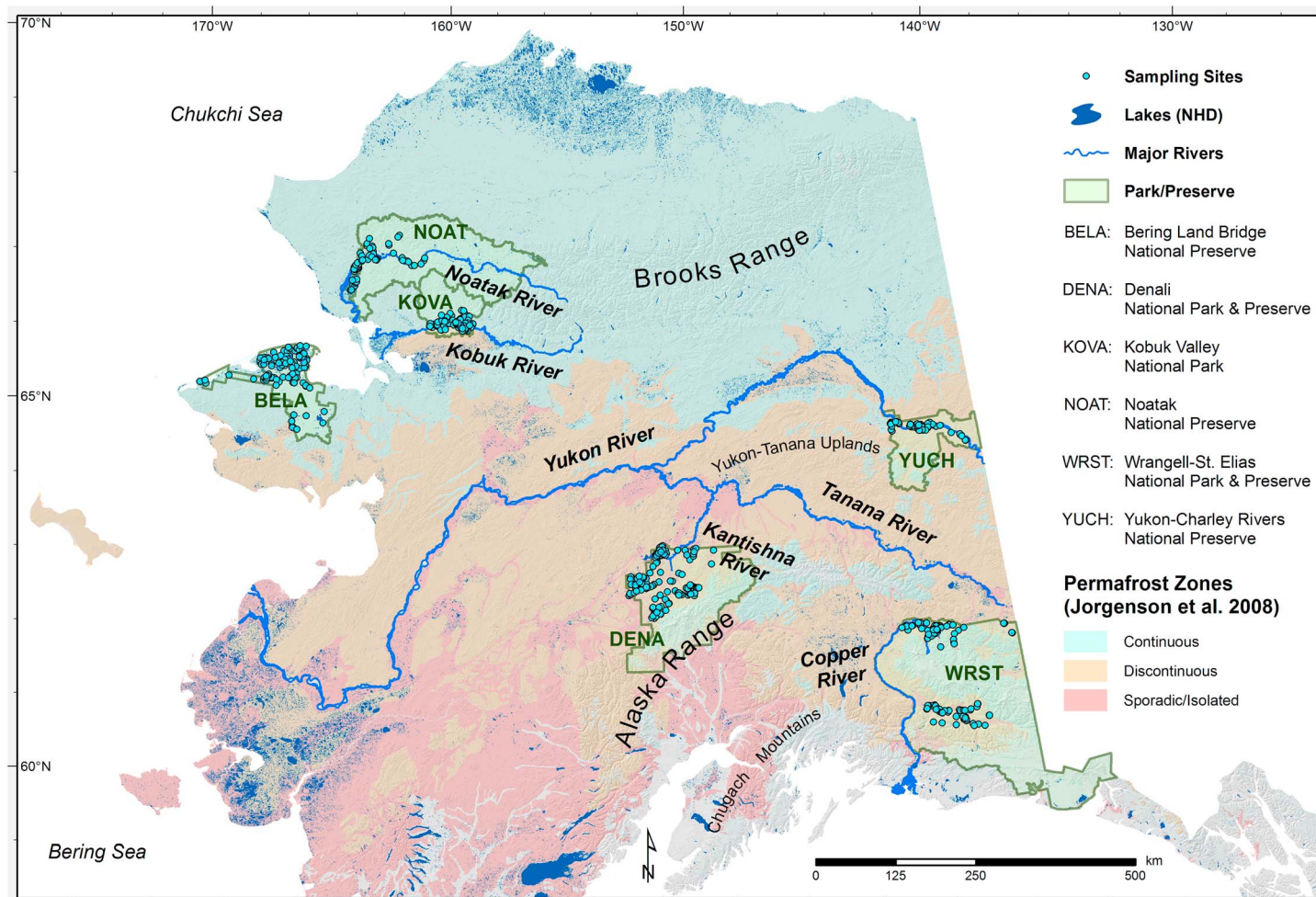


Figure 1. Map of lake basins surveyed across six National Park units in Alaska superimposed on the permafrost zones found in Alaska: sporadic/isolated (underlying <50% of land area), discontinuous (50–90%), and continuous (> 90%) [Jorgenson *et al.*, 2008c]. Lake and river features are from the National Hydrography Dataset [USGS, 2015].

have warmed about 1.4°C on average (Table 2), and the interior parks have undergone a 45% increase in growing season length [Wendler and Shulski, 2009]. These regions are all expected to become warmer and wetter throughout the century (Table 2).

2.2. Environmental Setting

In the modern Alaska landscape, lakes cover approximately 2.1×10^7 ha [Hinkel *et al.*, 2012; Milner *et al.*, 1997]. They are distributed unevenly and are highly heterogeneous physically (surface area, depth, and

Table 1. Number of Lakes Sampled (N), Total Land Surface Area, Total Number of Lakes, and Lake Surface Area Found in Bering Land Bridge (BELA), Denali (DENA), Noatak (NOAT), Wrangell-St. Elias (WRST), and Yukon-Charley Rivers (YUCH), Estimated in GIS Using the National Hydrography Dataset [USGS, 2015]^a

	N	Land Area (km ²)	Lake Number	Lake Area (km ²)	Dominant Permafrost
BELA	113	1.1×10^4	1.1×10^4	5.4×10^2	Continuous
DENA	156	2.4×10^4	1.1×10^4	2.8×10^2	Discontinuous
KOVA	100	7.1×10^3	2.5×10^3	59.5	Continuous
NOAT	52	2.7×10^4	6.5×10^3	3.2×10^2	Continuous
WRST	126	5.3×10^4	1.0×10^4	4.0×10^2	Discontinuous
YUCH	69	1.0×10^4	8.7×10^2	16.8	Discontinuous

^aDominant permafrost extent determined for each park from Jorgenson *et al.* [2008c].

Table 2. Summary of Mean Annual Air Temperature (MAAT; °C), and Mean Annual Precipitation (MAP; mm) for Bering Land Bridge (BELA), Denali (DENA), Noatak (NOAT), Wrangell-St. Elias (WRST) and Yukon-Charley Rivers (YUCH) Calculated From Climate Station Data Collected at a Remote Automated Weather Station (RAWS) Within Each Park Unit or a Nearby Airport [Applied Climate Information System, 2017]^a

	Historic 1950–1959		Measured 2001–2016		Projected 2050–2059		Projected 2090–2099	
	MAAT	MAP	MAAT	MAP	MAAT	MAP	MAAT	MAP
BELA(Kotzebue Airport)	−5.5	246	−4.3	283	−1.5	303	0.5	328
DENA (Lake Minchumina RAWS)	−3.6	307	−1.3	308 ^b	0.2	378	1.7	405
KOVA (Kavet Creek RAWS)	−5.1	359	−3.7	386 ^b	−1.3	407	0.6	439
NOAT (Kelly River RAWS)	−6.0	313	−3.8	413	−1.9	396	0.2	432
WRST (Gulkana Airport)	−1.6	337	−1.9	278	0.4	386	1.8	426
YUCH (Eagle Airport)	−4.7	336	−3.2	288	−1.5	359	0.0	394

^aModels from the Scenarios Network for Alaska and Arctic Planning (SNAP) were used to generate historic and projected values [Scenarios Network for Alaska and Arctic Planning, 2017a, 2017b, 2017c, 2017d].

^bPrecipitation data were not available for sampling areas in DENA and KOVA; the values included are from the SNAP models for 2000–2010 [Scenarios Network for Alaska and Arctic Planning, 2017c].

temperature) as well as chemically, reflecting the spatial variation in landscape characteristics and climate. The vast majority of lakes in Alaska occur within 20 lake districts [Arp and Jones, 2009] containing large concentrations of lakes of similar geologic origin.

Glacial, volcanic, coastal, and fluvial activities are the primary geologic processes that drive lake formation in Alaska, and these physical drivers, combined with cryogenic processes, yield the diverse array of lakes found throughout the state. During the Last Glacial Maximum, glaciers covered nearly half of Alaska, and complex geomorphological processes that occurred during and following these glaciations have resulted in a myriad of lakes [Arp and Jones, 2009]. Aeolian deposits of sand and silt cover unglaciated areas below 300–450 m elevation, with thick deposits of silt concentrated in the central and western portions of the study area [Pewe, 1975]. Fine-grained sediment within these tracts contain syngenetic permafrost, which formed by the rise of the permafrost table as sediment was deposited, and these deposits are very ice rich and commonly referred to as yedoma [Schirmer et al., 2013]. Many lakes formed in depressions within this silt-capped terrain. Extensive Pleistocene sand dunes occur as well and are concentrated on the northwestern flanks of the Alaska Range near the Kantishna River [Collins, 1985] and south of the western Brooks Range along the Kobuk River [Pewe, 1975]. Oxbow lakes formed within vast fluvial deposits along all of the major rivers including the Copper, Kantishna, Yukon, Kobuk, and Noatak. Small areas of ice-rich, glacio-lacustrine deposits occur in the Noatak [Hamilton, 2010] and Copper River valleys [Jorgenson et al., 2008a; Wiedmer et al., 2010]. Volcanic ash is common throughout the entire study area, and fine-grained sand and coarse volcanic material are present throughout much of the northern Seward Peninsula. Marine deposits are limited to the northern section of the Seward Peninsula and the mouth of the Noatak River.

Permafrost is a dominant feature throughout this environment and varies in extent among the parks (Figure 1 and Table 1). Permafrost with high ice content is common throughout central and western Alaska and is most prevalent on the Seward Peninsula [Jorgenson et al., 2008b; Kanevskiy et al., 2011]. Isolated patches of ice-rich permafrost also persist in small areas throughout the rest of the study region, including deep loess deposits on terraces and lower side slopes of hills bordering the Yukon and Kobuk Rivers and on the northern flanks of the Alaska Range. Permafrost thaw during the Holocene was responsible for the formation of thousands of thermokarst lakes throughout the study area, and thermokarst lakes represent a major proportion of the lakes sampled in this study [Burn and Smith, 1990]. Warming over the past century has led to a dramatic increase in permafrost degradation throughout northern Alaska [Jorgenson et al., 2006]. Continued warming is projected to thaw 16–24% of Alaska's remaining permafrost by the end of the 21st century [Pastick et al., 2015], and mean decadal ground temperatures are expected to exceed zero in all the parks except Noatak [Panda et al., 2014a, 2014b, 2016].

2.3. Study Design

We identified all lakes (>1 ha < 50 ha) within each NPS unit using the U.S. Geological Survey's National Hydrography Dataset [US Geological Survey (USGS), 2015] and remotely sensed imagery, establishing the

population of lakes from which we sampled. We used a general randomized tessellation stratified (GRTS) sampling approach [Stevens and Olsen, 2004] to select a spatially balanced sample of lakes from the population within each NPS unit. A total of 617 lakes were sampled for this study, and between 80 and 160 lakes were sampled per park unit depending on the size and density of lakes in that unit. In a few cases, lakes were sampled that were greater than 50 ha or less than 1 ha in surface area, due to considerations of accessibility or importance on the landscape. Some lakes selected in the GRTS process were not possible to sample due to lack of accessibility. We also selected one to three easily accessible lakes in each park to be sampled annually; data from these lakes are also included in the study. All measurements and analyses in this study were conducted in conjunction with the NPS Inventory and Monitoring Program for the Arctic and Central Alaska Network [Lawler *et al.*, 2009; MacCluskie *et al.*, 2005]. Details of lake sampling procedures and protocols can be accessed online (<https://irma.nps.gov/DataStore/Reference/Profile/2170408>). In section 2.4, we briefly summarize our field sampling approach and laboratory analyses.

2.4. Field Sampling and Laboratory Analyses

We collected water samples from 617 lakes during a single site visit to each lake during the open water season (June to September) between 2006 and 2013. Lake depths were measured throughout the lake using a Speedtech 5 portable depth sounder (Speedtech, Great Falls, VA, USA). Spatial data for each depth measurement was generally collected using a Trimble GPS unit, but when GPS locations were not available depth measurement locations were approximated on a sketched map of the lake. Lake elevation (ELEV) was generally determined from spatial data collected with either a Trimble or Garmin GPS receiver or was estimated based on digital elevation models and/or topographic maps. Lake surface area (LSA) was calculated using shoreline polygons digitized using the best available imagery (satellite images or aerial photography) and on-site photographs. Shoreline development index (SDI) approximates the amount of shoreline that is available in a lake for development of littoral vegetation, an important biotic indicator, and was calculated by dividing the shoreline length from each of the digitized shoreline polygons by the circumference of a circle with an area equal to that of the polygon [Hutchinson, 1957].

Surface water temperature and pH were measured in situ 50 cm beneath the surface at one to three locations within the lake basin, using a Yellow Springs Instrument (YSI) QS multiparameter sonde (YSI Incorporated, Yellow Springs, OH, USA). The water clarity at each site was characterized by measuring the Secchi depth (SECCHI), using a 20 cm Secchi disk. Water samples were collected in 2 L and 500 mL bottles from the deepest area of each lake at approximately 50 cm below the water surface. The 500 mL unfiltered water sample was frozen for analysis of specific conductivity (SPCOND), total nitrogen (TN), and total phosphorus (TP). Part of the 2 L sample was filtered through prerinsed Geotech cellulose acetate or Pall Metrical 0.45 μm membrane filters, and the filtrate was frozen for analysis of nutrients (total dissolved nitrogen [TDN], ammonia [NH_3], and nitrate + nitrite [NO_3^-] orthophosphate [PO_4^{3-}]), base cations (calcium [Ca^{2+}], magnesium [Mg^{2+}], potassium [K^+], and sodium [Na^+]), and major anions (sulfate [SO_4^{2-}] and chloride [Cl^-]) or refrigerated for analysis of dissolved organic carbon (DOC) and silicon (Si). Water samples collected from lakes sampled in 2011 and 2012 were also filtered for determination of total dissolved nitrogen (TDN). To measure chlorophyll concentrations, 600 mL of the unfiltered water sample was filtered through a Gelman A/E 0.7 μm glass fiber filter, and the filter was wrapped in aluminum foil and stored frozen in the dark until extraction for the determination of chlorophyll *a* and phaeophytin. Raw chlorophyll *a* concentrations were corrected by subtracting phaeophytin to yield corrected chlorophyll *a* values (CHLAC).

Chlorophyll samples were analyzed at University of Alaska, Fairbanks using the Turner Designs Method for determination of chlorophyll [Arar and Collins, 1997]. All other samples were analyzed at the Cooperative Chemical Analytical Laboratory at the College of Forestry at Oregon State University using APHA Standard Methods for the Examination of Water and Wastewater [Eaton *et al.*, 2005]. A Technicon Autoanalyzer II was used to measure the concentration of all nutrient species and Si concentrations. Anion concentrations were measured on a Dionex 1500 ion chromatograph. Alkalinity and pH were measured using a ManTech PC-Titrate Auto Titrator System. A Shimadzu AA-7000 was used to measure base cation concentrations.

2.5. Lake Basin Classification

We used a three-step, semiquantitative process to classify lakes (Figure 2). In the first step of the process, each lake was attributed in the geographic information system (GIS) with all known hydrologic, physiographic,

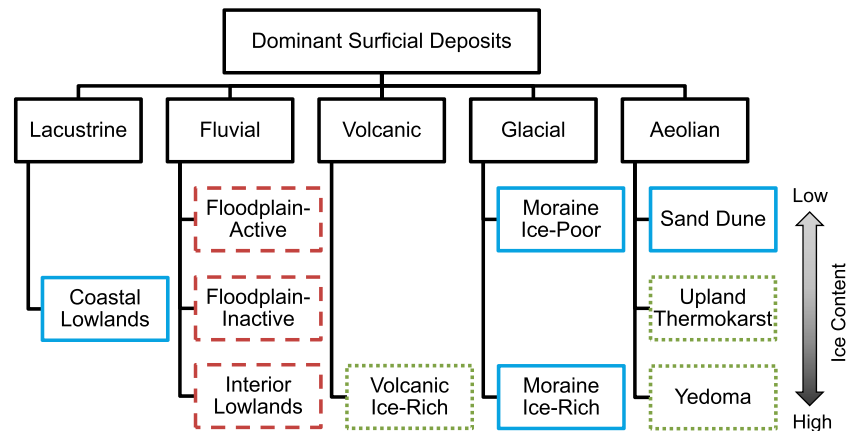


Figure 2. Distribution of lake basin types in the five dominant surficial deposits identified in the subsection analysis. The arrow on the right indicates the variation in ice content from low ice content (at the top) to high ice content (on the bottom). Lake basin types outlined in solid blue indicate lakes dominated by precipitation, those outlined in dashed red indicate potential deep groundwater, and those outlined in dotted green indicate shallow groundwater inputs.

lithologic, and edaphic information based on subsection maps produced for each NPS unit [Jorgenson, 2001; Jorgenson et al., 2008a; Swanson, 2001a, 2001b]. Subsection mapping is a landscape-level mapping process that delineates unique ecological units at a spatial scale that integrates biotic and abiotic characteristics of the environment, as well as meaningful boundaries that generally coincide with major geologic or geomorphologic features [Bailey, 1996]. Subsection maps are park specific, mapped by different authors, and are regionally named, but there is substantial overlap of ecosystem types among the different park units.

In the second step of the process, we combined subsections across park units to reflect shared edaphic characteristics, lake origin, and geomorphologic features. Sample lakes were located in one of five major surficial deposits: fluvial, glacial, marine, aeolian, and volcanic. Using detailed soil descriptions provided with the subsection maps, combined with expert opinion, we derived 10 lake basin types that were characteristic of each of these surficial deposits, yet reflected lake genesis, soil composition, permafrost distribution, and ice content (Figure 2). We identified five different thermokarst lake basin types: Coastal Lowland, Interior Lowland, Upland Thermokarst, Volcanic Ice-Rich, and Yedoma. Coastal Lowland lakes occur near the Chukchi Sea coast in the Bering Land Bridge and Noatak National Preserves, in the continuous permafrost zone, on nearly level plains covered almost completely by current and drained thermokarst lakes and separated by ice wedge polygon terrain. Interior Lowland lakes are found on low-lying plains composed of alluvial or lacustrine deposits on flat plains within old river terraces at inland locations. Upland Thermokarst lakes are situated on gently sloping bedrock-cored hills capped with thick deposits of loess that contain ice wedges interspersed with syngenetic permafrost. The lake basins of this type occupy topographic lows that are only partly due to thermokarst. Volcanic Ice-Rich and Yedoma lake basins both contain large syngenetic ice wedges and occupy deep basins that are mostly of thermokarst origin [Clark and Duffy, 2006; Swanson, 2001a, 2001b]. Volcanic Ice-Rich lake basins in this study are present only in Bering Land Bridge and differ from Yedoma lake basins in that the ice-rich loess is capped with coarse- to fine-grained volcanoclastic materials [Jorgenson, 2001]. The lakes comprising the remaining five lake basin types occur in glacial deposits (Moraine Ice-Poor and Moraine Ice-Rich), ice-poor alluvium (Floodplain-Active and Floodplain-Inactive), or Aeolian sand (Sand Dunes). Moraine Ice-Rich lakes were in glacio-lacustrine deposits [Hamilton, 2010; Jorgenson et al., 2008a], while Moraine Ice-Poor lakes were in coarse-grained glacial till [Clark and Duffy, 2006; Hamilton, 2010].

In the third step of the process, detailed subsection [Jorgenson, 2001; Jorgenson et al., 2008a; Swanson, 2001a, 2001b] and detailed soils units [Clark and Duffy, 2006] were used to assign lake basin type. Once assigned to a lake basin type, lake assignments were validated by reviewing site photographs and notes collected during site visits. Several lakes situated along map unit margins were adjusted to reflect site characteristics and when possible validated using supplementary maps. Maps indicating the distribution of sand sheets [Collins, 1985; Swanson, 2001a] and glacio-lacustrine deposits [Hamilton, 2010; Jorgenson et al., 2008a] were helpful in this process. If on-site photographs, notes, or supplementary maps demarked different soils, lake genesis or water source, or suggested higher or lower ice content, lake basins were reassigned as appropriate.

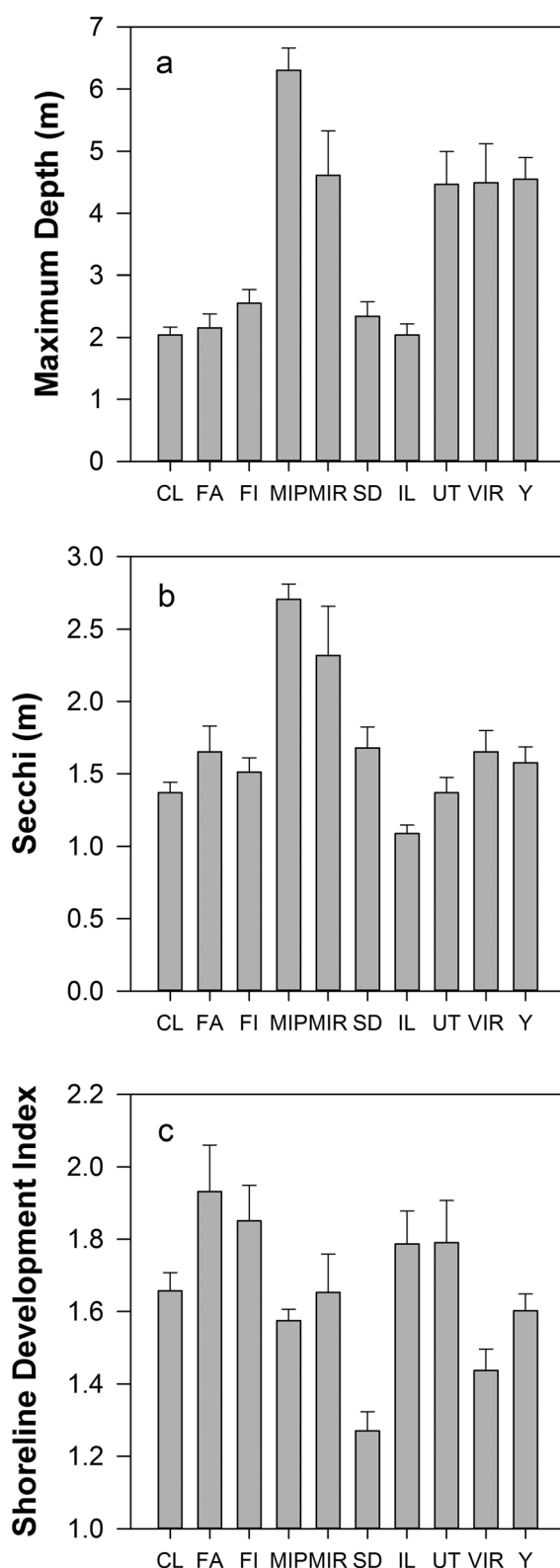


Figure 3. Maximum lake depth (m), Secchi (m), and shoreline development index (mean + SD) measured in Coastal Lowlands (CL), Floodplain-Active (FA), Floodplain-Inactive (FI), Moraine Ice-Poor (MIP), Moraine Ice-Rich (MIR), Sand Dune (SD), Interior Lowland (IL), Upland Thermokarst (UT), Volcanic Ice-Rich (VIR), and Yedoma (Y) lake basin.

2.6. Data Analysis

All measured physical and chemical attributes were summarized by lake basin type and were used to identify broad differences among types. Data were nonnormally distributed, so we used a Kruskal-Wallis rank sum test to test for differences in physical and chemical parameters among lake basin types. The pairwise Wilcoxon test was used to test for differences between individual lake types. Simple linear regression was used to examine relationships between nutrient and chlorophyll *a* concentrations.

NO_3^- and PO_4^{3-} concentrations were frequently below detection limits, and nondetect values were replaced with half the detection limit [Croghan and Egeghy, 2003]. This method is preferential to removing all values below the detection limit, a practice that overestimates mean concentrations. Because so many values were below detection limits in this study, standard deviation among lake basin types was reduced.

3. Results

3.1. Lake Physical Characteristics

We documented broad spatial variation in lake physical characteristics among lake basin types. Lake surface area (LSA) averaged 16.5 ± 32.5 ha, ranging from 0.5 to 326 ha across all lakes. Coastal Lowland, Sand Dune, and Volcanic Ice-Rich lake basins were significantly larger than all other lake basin types (Kruskal-Wallis test; $H = 55.88$, $P < 0.001$). Maximum lake depth (Z_{max}) averaged 4.08 ± 0.15 m across all lakes, ranging from 0.3 to 30.6 m. Most lakes (93%) were relatively shallow with Z_{max} less than 10 m (Figure 3a). Z_{max} differed significantly among lake basin types ($H = 176.45$, $P < 0.001$) and was strongly correlated with elevation ($r = 0.413$, $P < 0.001$). Three depth categories emerged across lake basin types. Moraine Ice-Poor basins above 600 m elevation were significantly deeper than the other lake basin types (Figure 3a). The shallowest lakes, ranging from 2.04 to 2.34 m, were in lowland landscapes and

Table 3. Mean Specific Conductivity (SPCOND $\mu\text{S}/\text{cm}$), pH, and Ion Concentrations (mg/L) for Each Lake Basin Type

	<i>N</i>	SPCOND	pH	SO_4^{2-}	Cl^-	Ca^{2+}	Mg^{2+}	Na^+	K^+	Si
Coastal Lowland	50	64.2	7.7	0.14	7.43	4.46	2.61	4.29	0.46	0.25
Floodplain-Active	29	164.6	8.1	12.90	0.90	21.61	9.02	2.79	1.02	1.36
Floodplain-Inactive	47	116.7	7.8	3.75	1.24	12.79	5.48	2.32	0.88	1.47
Moraine Ice-Poor	178	88.9	8.0	1.75	0.39	12.81	3.84	2.22	0.85	1.39
Moraine Ice-Rich	23	91.9	8.4	0.68	0.67	12.20	4.79	1.85	0.60	0.99
Sand Dune	63	39.0	7.4	0.09	0.39	6.96	1.30	1.23	0.42	0.48
Interior Lowland	67	159.7	7.5	14.11	1.52	17.20	9.86	4.21	0.73	0.85
Upland Thermokarst	26	34.3	6.6	0.39	0.40	3.90	1.74	1.71	0.57	0.74
Volcanic Ice-Rich	31	118.3	8.4	0.45	12.13	8.12	5.73	8.38	1.24	0.63
Yedoma	101	143.0	8.0	5.20	5.77	14.74	8.11	5.50	0.74	0.75

included Coastal Lowland, Floodplain-Active, Floodplain-Inactive, Sand Dune, and Interior Lowland basins (Figure 3a). Large proportions of these shallow lakes were found in thermokarst depressions in areas with low to moderate ice content. Lake depth appeared to be correlated with ground ice content, which determines the potential for ground surface subsidence. Lakes in ice-rich terrain (Moraine Ice-Rich, Upland Thermokarst, Volcanic Ice-Rich, and Yedoma) were also largely formed in thermokarst depressions, although Zmax was a meter deeper on average than lowland lake basins underlain by low to moderate ice content.

Across all lake basin types, SECCHI averaged 1.8 ± 1.2 m and was visible on the lake bottom in 45% of the lakes. Lakes found in glacial moraines, both Ice-Poor and Ice-Rich, had significantly higher SECCHI values ($H = 137.16$, $P < 0.001$; Figure 3b) than other lake basin types, with SECCHI averaging 2.7 ± 1.4 m and 2.3 ± 1.5 m, respectively. SECCHI was lowest in Interior Lowland lake basins, averaging 1.08 ± 0.4 m (Figure 3b).

Lake shoreline complexity, as indicated by SDI, varied in shape from nearly circular ($\text{SDI} = 1.04$) to highly convoluted ($\text{SDI} = 5.3$). Active and Inactive Floodplain lakes had significantly higher SDI values (Figure 3c) than the other lake basin types ($H = 55.35$, $P < 0.001$), averaging 2.01 ± 0.66 and 1.90 ± 0.65 . Lakes in these basin types tended to be long and narrow. Upland thermokarst and interior lowland basins also had relatively high SDI values, averaging 1.88 ± 0.59 and 1.78 ± 0.74 , respectively (Figure 3c). These types tended to have convoluted shorelines in cases where multiple lakes had merged, or when the lake was located in a complex basin surrounded by small hills. Sand Dune and Volcanic Ice-Rich lakes had the lowest SDI values and most circular lake structure of all the lake types, averaging 1.43 ± 0.42 and 1.43 ± 0.33 , respectively.

3.2. Lake Chemistry: pH, Conductivity, and Conservative Solutes

Surface water pH averaged 7.81 ± 0.88 across all lakes, ranging from mildly acidic (min = 5.2) to alkaline (max = 9.6). Upland Thermokarst lakes were significantly ($H = 88.15$, $P < 0.001$) more acidic than all other lake basin types, with pH averaging 6.6 ± 0.83 (Table 3). The most alkaline lakes were in Moraine Ice-Rich, Volcanic Ice-Rich, and Floodplain-Active basin types.

Specific conductivity (SPCOND) averaged 103 ± 161.24 $\mu\text{S}/\text{cm}$ across all lakes, ranging from 3.9 $\mu\text{S}/\text{cm}$ in a Moraine Ice-Poor lake basin to 1844 $\mu\text{S}/\text{cm}$ in a Yedoma lake basin. Mean SPCOND values in Sand Dune and Upland Thermokarst lake basins were generally low (< 50 $\mu\text{S}/\text{cm}$; Table 3) and were lower than all other lake basin types ($H = 95.29$, $P < 0.001$). Floodplain-Active, Interior Lowland, and Yedoma basins had the highest mean SPCOND (averaging 155.73 ± 249.33 , collectively; Table 3). SPCOND exceeded 500 $\mu\text{S}/\text{cm}$ in only 15 lakes, 10 of which exceeded 800 $\mu\text{S}/\text{cm}$. These lakes were geographically restricted to the Yukon River Valley, and occurred in Yedoma, Interior Lowland, and both Floodplain-Active and Inactive lake basin types, contributing to the high variance observed in SPCOND in those lake basin types.

Ca^{2+} concentrations were highly variable across lakes, averaging 12.1 ± 16.05 mg/L and ranging from 0.41 to 146.45 mg/L. In most lakes, Ca^{2+} was the dominant cation and was highly correlated with Mg^{2+} ($r = 0.75$, $n = 615$, and $p = 0.01$) (Table 3). Floodplain-Active lakes had the highest Ca^{2+} concentrations, averaging 21.61 ± 20.53 mg/L, whereas Upland Thermokarst lakes had the lowest Ca^{2+} concentrations, averaging 3.90 ± 4.58 mg/L (Table 3 and Figure 4). Upland Thermokarst and Sand Dune lake basin types had significantly less Ca^{2+} ($H = 85.54$, $P < 0.001$) and Mg^{2+} ($H = 98.15$, $P < 0.001$) than all other lake types. Most lake basin types were Ca-Mg- HCO_3 type waters, with the exception of Volcanic Ice-Rich

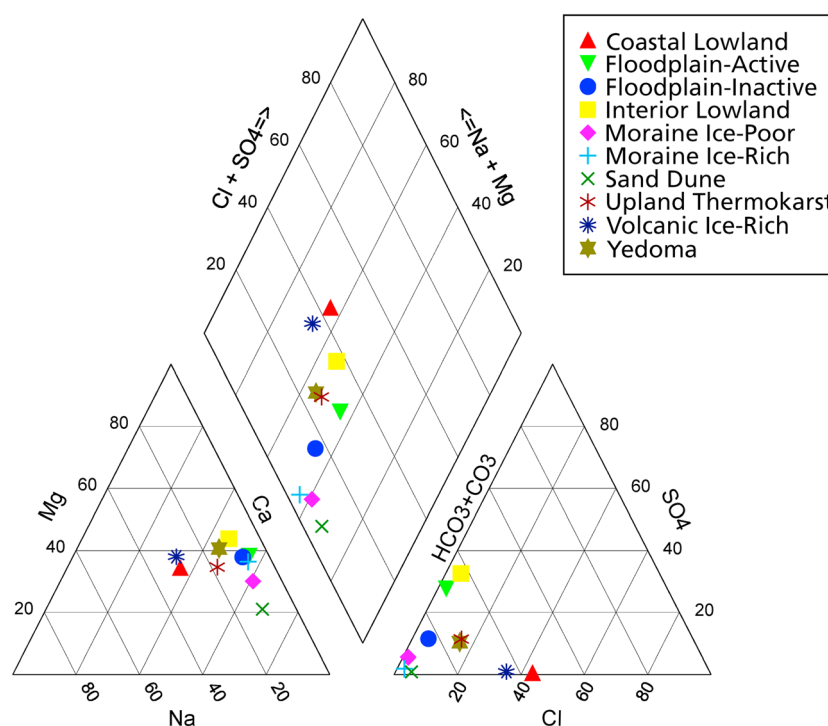


Figure 4. Piper diagram showing dominant water types using mean cation and anion concentration (% meq/L) from each lake basin type sampled in Alaskan parks. The lower left ternary plot shows cation concentrations (sodium [Na], calcium [Ca], and magnesium [Mg]) and the lower right ternary plot show anion concentrations (sulfate [SO₄], carbonate [CO₃], bicarbonate [HCO₃], and chloride [Cl]). The diamond plot illustrates the combined cation and anion concentrations for each lake basin type.

and Coastal Lowland lakes, which had a higher concentration of both Na⁺ and Cl[−] than all other lake types ($H = 108.36$, $P < 0.001$; $H = 230.53$, $P < 0.001$, respectively) (Figure 4). K⁺ concentrations were typically low within all lake types, averaging 0.75 ± 0.72 mg/L. K⁺ was highest ($H = 59.97$, $P < 0.001$) in Volcanic Ice-Rich and Floodplain-Active lakes, averaging 1.24 ± 0.72 and 1.02 ± 1.16 mg/L, respectively (Table 3). Across all lakes, Si concentration averaged 0.97 ± 1.30 mg/L but varied substantially among lake basin types ($H = 100.07$, $P < 0.001$). Concentrations were highest in Moraine Ice-Poor lakes and Active and Inactive Floodplain lakes and lowest in Coastal Lowland sites (Table 3 and Figure 4). SO₄^{2−} concentrations were significantly different among basin types ($H = 104.70$, $P < 0.001$; Figure 4). Similar to patterns observed for SPCOND, SO₄^{2−} concentrations were highest at sites within the Yukon River Valley and included Floodplain-Active, Interior Lowland, and Yedoma lake basin types.

3.3. Lake Chemistry: Nutrients, Carbon, and Chlorophyll

TN concentrations averaged 679 ± 695 μg N/L (range = 120–12450 μg N/L) across all study lakes, with the highest concentrations in Yedoma lakes (1033 ± 1059 μg N/L) and lowest in Moraine Ice-Poor lakes (522 ± 304 μg N/L; Figure 5). Lake basin types concentrated along the Yukon River Valley (Yedoma, Interior Lowland, and Floodplain-Inactive) had significantly higher TN concentrations than the same lake basin types in other regions ($H = 83.77$, $P < 0.001$). We calculated dissolved organic nitrogen (DON) by subtracting dissolved inorganic nitrogen sources, NO₃[−] and NH₃, from TDN. DON was generally the dominant species of N and represented 95% of TN across all lakes. Mean DON concentrations were highest in Interior Lowland (737 ± 331 μg N/L) and Yedoma (732 ± 553 μg N/L) lake basin types, and lowest in Sand Dune (409 ± 309 μg N/L) lake basins. Dissolved inorganic N concentrations, NO₃[−] + NH₃, were frequently at or below analytical detection limits and comprised only a small fraction (<5%) of the TN. NO₃[−] concentration averaged 4 ± 17 μg N/L (range = 1 to 352 μg N/L), and NH₃ concentration was 32 ± 24 μg N/L (range = 1 to 4995 μg N/L) across all study lakes. Although NO₃[−] concentrations were markedly higher in Floodplain-Active lakes (Figure 5), we did not detect statistical differences because NO₃[−] concentrations were

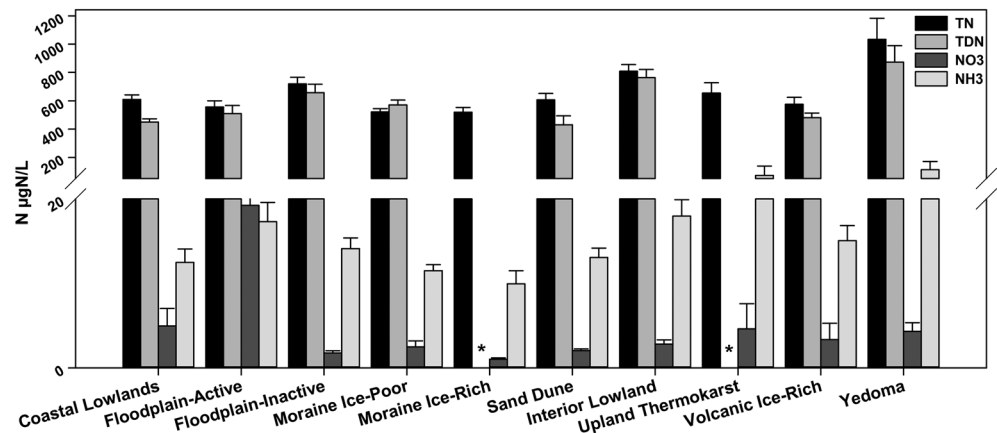


Figure 5. Mean (+SD) total nitrogen and dissolved nitrogen concentrations (total dissolved nitrogen, nitrate, and ammonia) observed in the 10 lake basin types. The asterisk symbol indicates lake basin types where TDN data were unavailable.

extremely variable, with concentrations ranging from 0.5 $\mu\text{g N/L}$ (one half the detection limit) to 352 $\mu\text{g N/L}$. Yedoma lakes had the highest NH_3 concentration, with a mean value of $113 \pm 585 \mu\text{g N/L}$, which was nearly an order of magnitude higher than other lake basin types.

TP concentration averaged $21 \pm 45 \mu\text{g P/L}$ across all lake basin types, ranging from 2 to 799 $\mu\text{g P/L}$ (Figure 6). Mean TP concentrations were higher in Yedoma lakes than all other lake basin types (Figure 6), while Moraine Ice-Poor lakes had lower concentrations of TP than the other lake basin types ($H = 110.71$, $P < 0.001$). We observed considerably more spatial variability in TP concentration in Yedoma and Upland Thermokarst lakes (Figure 6), whereas TP concentrations were considerably less variable across individual lakes within the other lake basin types. PO_4^{3-} concentrations were generally low, with values below analytical detection limits in 164 lakes. Across all lakes, PO_4^{3-} concentrations averaged $3 \pm 20 \mu\text{g P/L}$, and values ranged from a low of 1 to a high of 426 $\mu\text{g P/L}$. Lake basin PO_4^{3-} was highly correlated with TP ($r = 0.93$, $n = 10$, and $p < 0.05$), Yedoma lakes having the highest values, and Moraine Ice-Poor lakes the lowest (Figure 6).

DOC concentrations averaged $13.6 \pm 8.7 \text{ mg C/L}$ (range = 0.74–61.13 mg C/L) across all lakes. Mean DOC concentrations were higher ($H = 113.91$, $P < 0.001$) in Floodplain-Inactive, Interior Lowland, Upland Thermokarst, and Yedoma lakes ($>15 \text{ mg C/L}$) than the other lake basin types (Figure 7). The ratio of DOC:TDN ranged from a low of 4.39 to a high of 54.66, with an average ratio of 23.84 overall. DOC:TDN was lowest in Sand Dune lakes and highest in Interior Lowland basins. The relative differences between DOC and DOC:TDN are greatest in

Moraine Ice-Poor basins and smallest in Yedoma lake basins (Figure 7).

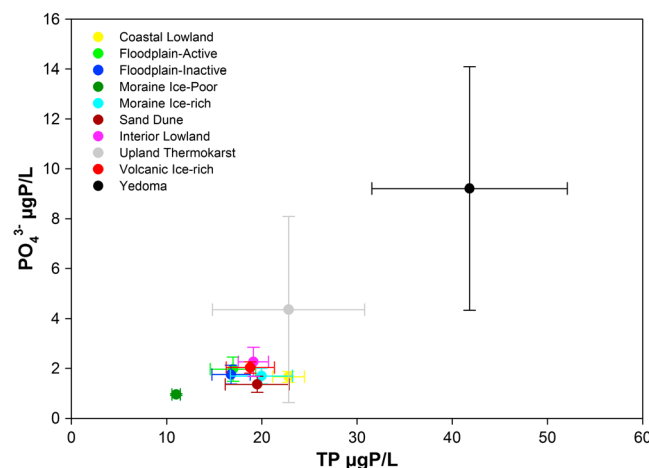


Figure 6. Relationship between mean orthophosphate concentration (PO_4^{3-}) and total phosphorus (TP), with bidirectional error bars indicating standard error of the mean for each of the 10 lake basin types sampled in Alaska parks.

CHLAC, a frequently used metric of phytoplankton biomass, ranged from 0.10 to 89.5 $\mu\text{g/L}$, with an overall mean of $3.2 \pm 5.9 \mu\text{g/L}$. CHLAC was highest in Yedoma lake basins, averaging $5.16 \pm 12.22 \mu\text{g/L}$, and lowest in Moraine Ice-Poor basins, averaging $1.86 \pm 1.61 \mu\text{g/L}$. Aquatic production, as indicated by CHLAC, was significantly lower ($H = 71.25$, $P < 0.001$) in Floodplain and Moraine lake basin types compared to all other lake basin types and was highly correlated with both TN ($y = 0.0044x + 0.2786$; $R^2 = 0.55$, $P = 0.014$; Figure 8a) and TP ($y = 0.0985x + 1.1374$; $R^2 = 0.71$, $P = 0.002$; Figure 8b). The TN:TP ratio

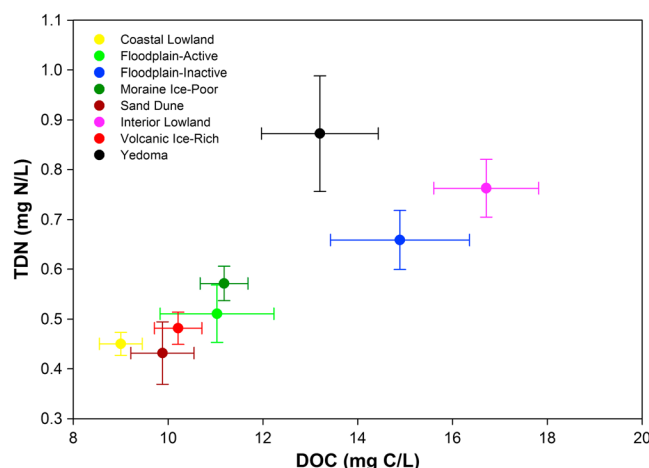


Figure 7. Relationship between mean total dissolved nitrogen concentration (TDN) and dissolved organic carbon (DOC), with bidirectional error bars indicating standard error of the mean for 8 of the 10 lake basin types sampled in Alaska parks. TDN data were not acquired in 2008, when lakes in Upland Thermokarst and Moraine Ice-Poor basins were sampled.

which are typically situated on recently deglaciated Quaternary deposits at high elevations that largely support alpine vegetation and well-drained soils. Conversely, the physical structure of lakes formed by thermokarst processes was dependent on ground ice content. The shallowest lakes with the shallowest photic zone were observed in the Interior Lowland basins, which are typically situated in low-elevation catchments in discontinuous to continuous permafrost with moderate ice content, while the deepest thermokarst lakes were found in ice-rich syngenetic permafrost.

4.2. Spatial Patterns in Lake Chemistry and Implications

Conservative solutes have frequently been used to understand hydrologic processes within lake watersheds, including subsurface flow paths, hydrologic mixing of source waters, and evaporative enrichment [Koch *et al.*, 2014; Lewis *et al.*, 2015], and we used these well-established linkages to interpret the patterns observed in our chemical data. Most lakes in this study had low specific conductivity values and base cation concentrations, suggesting a strong dependence on precipitation and snowmelt for recharge as opposed to deep groundwater flow paths. Some lakes, however, were located in landscapes with high potential for surface-groundwater interaction and had comparatively high specific conductivity (i.e., Floodplain-Active and Floodplain-Inactive, and Interior Lowland basins). These groundwater-dominated lake basins were

was highly variable across all lakes and ranged from 6 to 198. The lowest TN:TP ratio was found in Coastal Lowland lake basins, and the highest was found in Floodplain-Inactive lake basins.

4. Discussion

4.1. Spatial Patterns in Physical Properties of Lakes

Our findings indicate that the physical properties of lakes (e.g., size, depth, and depth of photic zone) were strongly influenced by geologic origin, edaphic properties, and permafrost characteristics of the watershed. For instance, the deepest and clearest lakes were generally observed in Moraine Ice-Poor basins,

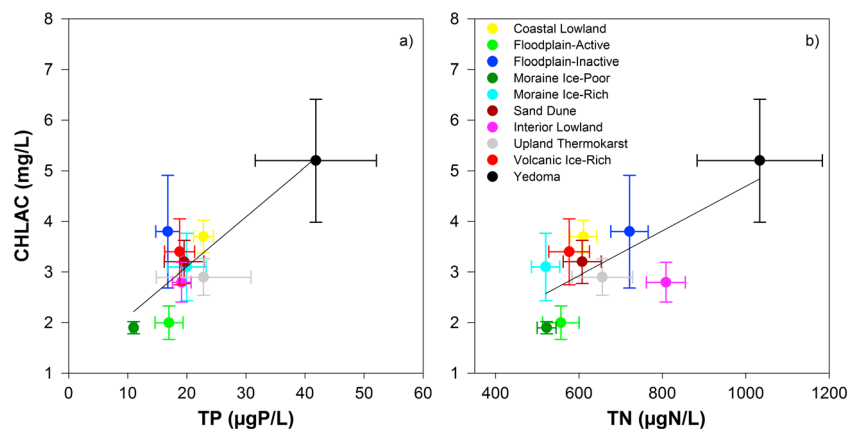


Figure 8. Linear regression between mean chlorophyll *a* concentration (CHLAC) and (a) mean total phosphorus (TP) and (b) mean total nitrogen concentration (TN). Both plots include bidirectional error bars that show the standard error of the mean for each of the 10 lake basin types sampled in Alaska parks.

concentrated in discontinuous permafrost underlain by unfrozen gravels, where surface water and groundwater are connected via unfrozen conduits [Michel and Van Everdingen, 1994]. Lakes underlain with more continuous permafrost generally had lower specific conductivity, with ice-rich permafrost acting as a barrier to groundwater upwelling [Walvoord and Kurylyk, 2016]. Specific conductivity and solute concentrations were especially high in a small subset of lakes in the Yukon River basin (YRB), likely due to the combined effects of hydrologic inputs from deeper flow paths [Walvoord et al., 2012], the relatively warm and dry climate [Monahan and Fisichelli, 2014] that favors evapoconcentration [Lewis et al., 2015], and solute inputs from active thermokarst [Kokelj et al., 2009; Walvoord et al., 2012]. A wide variety of other processes can influence solute concentrations, including atmospheric inputs, evaporation, erosion, and biological processes, although a lack of source water data limited our ability to determine the specific contributions these processes make.

Lakes with high Cl^- concentrations were largely confined to the coastal lowlands of the Seward Peninsula or the Noatak River lowlands, both of which are in close proximity to the Bering Sea. These are generally thermokarst lakes situated on continuous permafrost; the lake bottoms are underlain by closed taliks [Parsekian et al., 2013], restricting groundwater inputs and limiting recharge to precipitation and runoff with low levels of total dissolved solids [Hopkins, 1949; Jorgenson, 2001]. In addition, this area has a high lake density, the lakes are near sea level, and they tend to have small catchments, further limiting ionic inputs. Other studies in Alaska corroborate the relationship between Cl^- concentrations and the distance from coastline [Eilers et al., 1993; Kling et al., 1992; Pienitz et al., 1997].

4.3. Spatial Patterns in Lake Nutrients, Carbon, and Productivity

The trophic status of lakes in this study spans a large range of conditions, from ultraoligotrophic ($< 10 \mu\text{g P/L}$) to hyper-eutrophic ($> 100 \mu\text{g P/L}$) [Nürnberg, 1996]. Yedoma lakes tended to have the highest N and P concentrations, likely due to the high concentrations of N and P in ice-rich Pleistocene loess that can be leached upon thaw [Ewing et al., 2015], combined with active bank erosion and sedimentation. Active and inactive thaw slumps were frequently observed along Yedoma and Upland Thermokarst lake margins throughout the study area. These slumps were often within recent fire perimeters [BLM Alaska Fire Service, 2016]. Moraine Ice-Poor lakes were characterized by very low N and P concentrations, reflecting nutrient-poor soils with minimal organic matter accumulation at the ground surface. N and P are the primary nutrients limiting phytoplankton productivity in many arctic and subarctic lakes [Levine and Whalen, 2001].

In our study, spatial variation in lake DOC concentrations was tied to wetland extent and soil C storage in the surrounding basin, a relationship documented in prior studies on stream and lake ecosystems [Harms et al., 2016; Hope et al., 1996; O'Donnell et al., 2016]. Poorly drained soils underlain by permafrost facilitate the accumulation and stabilization of thick organic-soil horizons, which store large amounts of organic C and function as a source of DOC to aquatic ecosystems [Wickland et al., 2007]. Since DOC is an important source of nutrients to aquatic ecosystems, low DOC inputs, in addition to low DIN and P inputs, may have served as an additional constraint on lake productivity. Lakes with low DOC concentrations were often draining catchments with low soil C stocks [Hugelius et al., 2014], such as Sand Dune, Volcanic Ice-Rich, Moraine Ice-Poor, and Moraine Ice-Rich lake basin types. Upland Thermokarst and Interior Lowland basins have extensive wetland habitats that function as a DOC source for lakes. Runoff to these lake basins was largely confined to the shallow organic-rich soil found in the seasonal active layer, resulting in low pH and high DOC concentrations. Upland Thermokarst and Yedoma lakes are underlain by ice-rich syngenetic permafrost that stores large amounts of organic C, which can be mobilized as DOC and released to lakes upon thaw [Vonk et al., 2015; O'Donnell et al., 2014]. DOC tended to be lowest in areas with either low C stocks or where groundwater discharge to lakes was primarily via deep groundwater flow paths. Floodplain-Active lakes presumably have the shortest water retention time of all lake basin types and also had some of the lowest DOC concentrations, suggesting dilution of the lake DOC pool by low-C river water. Moraine (both Ice-Rich and Ice-Poor), Sand Dune, and Volcanic Ice-Rich lake basins have relatively thin organic horizons, well-drained soil, and watersheds that support low-profile vegetation and have correspondingly low DOC concentrations.

4.4. Conceptual Models of Lake Basin Hydrology and Chemistry

Our findings suggest permafrost extent (i.e., continuous versus discontinuous zones) influenced lake chemistry by mediating groundwater discharge to lakes, surface water-groundwater interactions, and erosion.

Using these observations we developed simple conceptual models for each lake basin type based on hypothesized lake basin hydrology and ecosystem processes (Figure 9). The conceptual models were largely based on our understanding of permafrost and soil characteristics, including soil properties, ground ice content, and thermal state. These factors govern subsurface hydrology in arctic and subarctic watersheds [Koch *et al.*, 2013; Walvoord *et al.*, 2012] and are known to influence the chemical composition of surface waters [O'Donnell *et al.*, 2014, 2016].

Proximity of study lakes to rivers was an important geographic factor that strongly influenced lake chemistry, likely through gradients in flooding frequency and surface water-groundwater interactions. Lake basins situated in low-lying landscapes along river corridors, such as Floodplain-Active, Floodplain-Inactive, and Interior Lowland, had high-specific conductivity and solute concentrations reflecting considerable groundwater discharge to lakes through highly conductive alluvial aquifers (Figure 9). High NO_3^- concentrations observed in Floodplain-Active basins suggest that flooding or groundwater discharge occurs routinely in these basins (Figure 9e). Prior work in interior Alaska has shown the importance of fluvial processes in delivering N to floodplain ecosystems [Lisuzzo *et al.*, 2008; Clilverd *et al.*, 2008]. High DOC concentrations commonly observed in Floodplain-Inactive and Interior Lowland likely reflect DOC delivery from riparian wetlands, and hydrologic import of NO_3^- is reduced in these less frequently flooded basins.

Precipitation and shallow regional groundwater also played important roles in determining lake chemistry and are probably the primary means of recharge for the remaining lake basin types (Sand Dune, Moraine Ice-Poor, Moraine Ice-Rich, Coastal Lowland, Upland Thermokarst Volcanic Ice-Rich, and Yedoma lakes; Figure 9). These lake basins are situated in continuous permafrost with moderate ice content, or on glacial moraines or sand dunes, and had comparatively low specific conductivity and DOC concentrations.

Thermal erosion appears to facilitate the delivery of nutrients, C, and other solutes to thermokarst lake basins as documented in other northern lake basins underlain by thawing permafrost [Kokelj *et al.*, 2005], while lake basin types perched on porous, stable surfaces tended to have low nutrient, DOC, and base cation concentrations, likely due to minimal erosion. Nutrient and solute concentrations in these basins varied with ground ice content and composition of the parent material. Recent work by O'Donnell *et al.* [2016] also showed that parent material and ground ice content are critical controls on riverine DOC concentration. The most nutrient- and solute-rich lake basins were found in syngenetic permafrost (i.e., Yedoma; Figure 9i), which stores large quantities of labile organic matter in frozen deposits that can be rapidly mineralized upon thaw [Spencer *et al.*, 2015]. This thawed organic matter may serve as an important source of nutrients (N, P) to lake ecosystems in these basins. In comparison, Volcanic Ice-Rich lake basins had lower nutrient and solute concentrations, suggesting that the low-carbon, igneous material (volcanic ash and cinders) blanketing the Yedoma (Figure 9h) in this region prevented the underlying yedoma from entering the lake basin during thermokarst events but did allow the influx of potassium-rich igneous material. Similar processes likely took place in Moraine Ice-Rich basins (Figure 9c), where thermokarst leads to the reworking of carbon-poor glacial till. Deposition of this material appears to liberate few nutrients or solutes, but concentrations in these basins were comparatively higher than those observed in Moraine Ice-Poor lake basins in similar deposits that did not undergo thermokarst activity.

Lake catchment characteristics (i.e., wetland extent, active layer thickness, and lake density) strongly influenced lake chemistry. Wetland extent and active layer thickness were particularly important in determining DOC concentration and pH. While all of the thermokarst lake basins were affected by wetland extent to some degree, it was most apparent in lakes in the southern portion of the study area where the active layer was thicker [Panda *et al.*, 2014a, 2014b, 2016] and the lakes were surrounded by large wetlands. These lakes had high DOC concentrations and low pH relative to their counterparts, which is consistent with findings that DOC concentration in Arctic rivers is positively correlated to wetland extent [Harms *et al.*, 2016; O'Donnell *et al.*, 2016]. Furthermore, spatial variation in active layer thickness controls DOC concentration in Arctic Rivers in northern Alaska [O'Donnell *et al.*, 2016] and likely influences lake chemistry as well. Lake water DOC and pH were also influenced by lake density. Low lake density combined with extensive wetlands and acidic organic-rich soils resulted in lakes with low pH and high DOC concentrations (Upland Thermokarst; Figure 9f), while lakes in similar settings with high lake density and small catchments with less wetland extent (Coastal Lowland; Figure 9g) had circumneutral pH with moderate DOC concentrations.

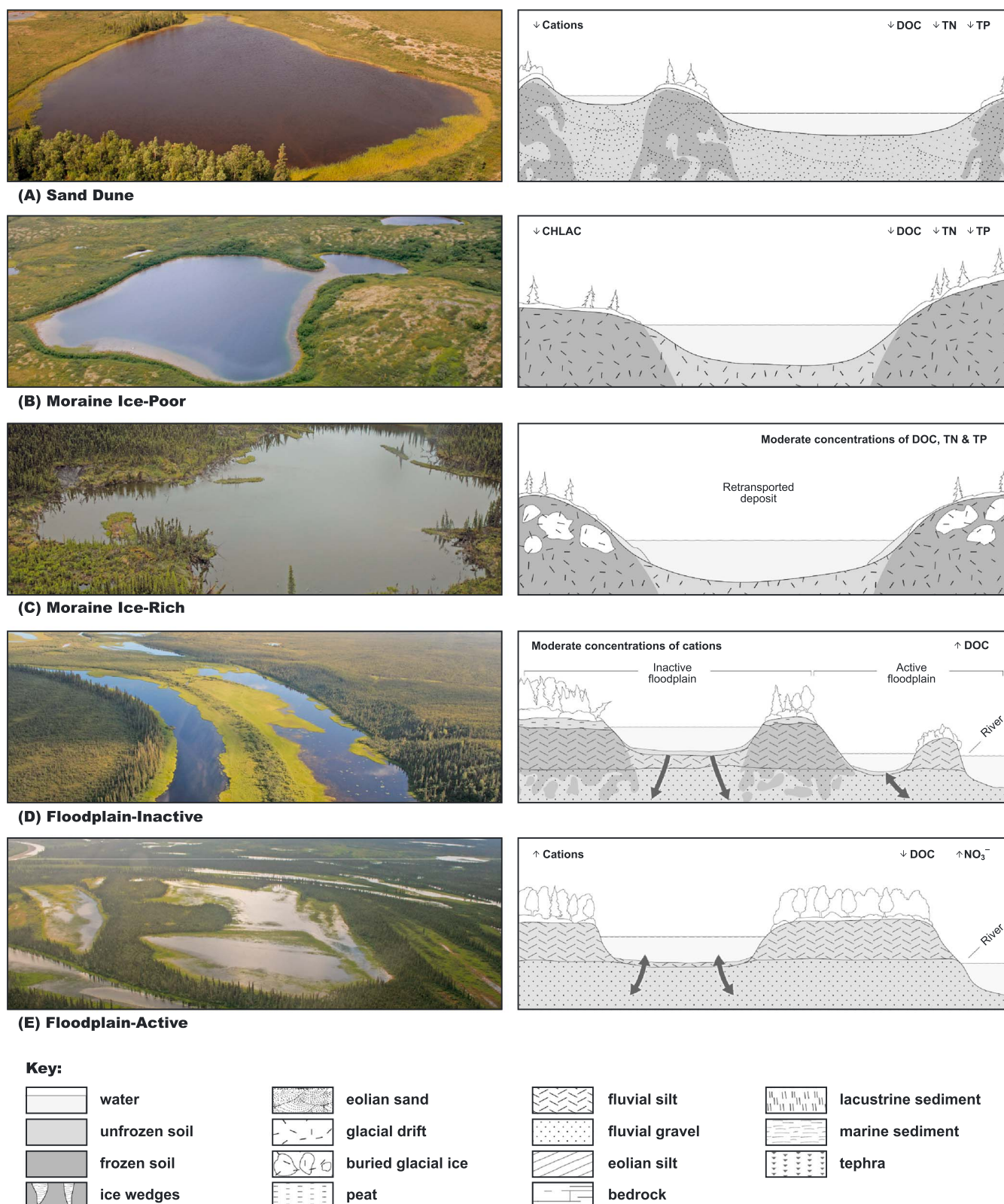


Figure 9. Conceptual model illustrating the dominant sediment types and ground ice content common to the 10 different lake basin types, overlaid with the key chemical signatures observed in each lake basin. Arrows indicate if the concentration of the chemical constituent is high (up arrow) or low (down arrow) relative to other lake basin types. Lakes are ordered on each page from low (top) to high (bottom) specific conductivity. Groundwater flow paths are indicated on lake basins with significant groundwater exchange.

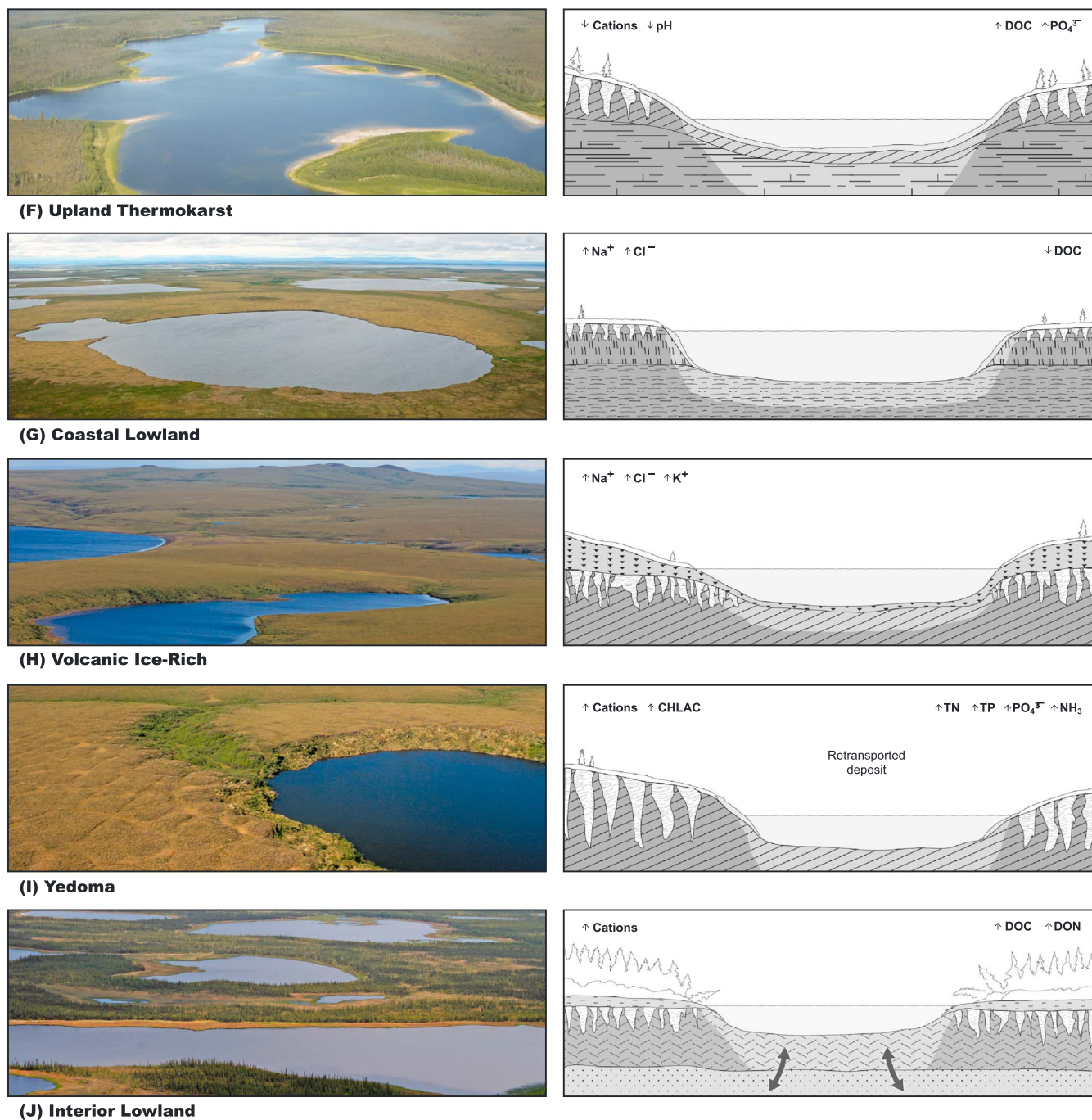


Figure 9. (continued)

Some solute concentrations were also influenced by coastal proximity, such as lakes along the coastal portion of northwestern Alaska (Coastal Lowlands; Figure 9g and Volcanic Ice-rich; Figure 9h). These lakes had elevated Cl^- concentrations relative to all other lake basins illustrating the strong influence of marine source water on lake chemistry.

4.5. Potential Climate Implications

In response to rising air temperatures across the study region, mean decadal ground temperatures are projected to rise above freezing in the southern portion of the study area by 2050 and be above freezing in all our study areas, except a few lakes in the northernmost region, by 2090 [Panda *et al.*, 2014a, 2014b, 2016].

This warming will drive widespread loss of permafrost, increasing active layer thickness, and resulting in extensive thermokarst development. The mode of thermokarst development will vary depending upon ground ice content, among other factors [Jorgenson and Osterkamp, 2005], and may take the form of degradation of ice wedges and other ground ice [Liljedahl et al., 2016, Jorgenson et al., 2006], and thawing of sublacustrine permafrost [Michel and Van Everdingen, 1994; Walvoord et al., 2012].

Permafrost thaw via active layer thickening and formation of closed taliks is expected to be widespread in the coming century [Slater and Lawrence, 2013; Panda et al., 2014a, 2014b, 2016] and will likely affect lake chemistry in all lake basins underlain by permafrost to some extent. In lake basins dominated by poorly drained organic-rich soils (i.e., Interior Lowland, Upland Thermokarst, and Coastal Lowlands), water will generally percolate deeper into the soil as active layer thickness increases, reducing runoff and increasing its contact with mineral soils. One consequence of this shift in subsurface hydrology could be reduced lake surface area and depth or increased NO_3^- and PO_4^{3-} solute delivery to lake basins due to greater contact with mineral soil [MacLean et al., 1999].

Climate-driven increases in thermokarst development [Jorgenson et al., 2006] will likely be most pronounced in lake basins underlain by ice-rich permafrost (i.e., Yedoma, Upland Thermokarst, and Moraine Ice-Rich). Potential consequences include both lake expansion (formation of new lakes and lateral expansion of existing lakes) and drainage by lateral breaching of permafrost barriers as well as vertical drainage, and increased sedimentation, nutrient, and solute delivery due to thermokarst bank erosion. Ice-rich permafrost is abundant in large areas of northwestern Alaska and limited portions of interior Alaska [Kanevskiy et al., 2011]. Permafrost has been stable since the early Holocene, a period of widespread thermokarst and lake expansion [Burn and Smith, 1990, Edwards et al., 2016], recent research and field observations suggest that this equilibrium has stopped and thermokarst activity is on the rise [Smith et al., 2005; Jones et al., 2011]. Projected increases in mean annual air temperature and precipitation (Table 2) could destabilize permafrost across the region, resulting in widespread thermokarst similar to that present during the early Holocene. Erosion-induced consequences for lake ecosystems may include eutrophication and surface water browning in lakes surrounded by fine-grained sediments rich in organic nutrients and solutes (i.e., Yedoma and Upland Thermokarst). Lake basins situated in coarse-grained, nutrient-poor sediments (i.e., Volcanic Ice-Rich and Moraine Ice-Poor) may be more resilient to climate and disturbance associated with permafrost thaw.

While many studies have documented the role of permafrost in governing subsurface hydrology [Walvoord and Kurylyk, 2016, Walvoord et al., 2012, Yoshikawa and Hinzman, 2003] little is known about the specific hydrologic regimes for the lake basin types presented in our study. However, we know that the presence of ice-rich permafrost exerts strong controls on water flux and groundwater flow paths by preventing the vertical movement of water through the soil, thus preventing the hydrologic exchange between surface water and deep groundwater. Warming ground temperatures can lead to sublacustrine permafrost thaw, resulting in lake drainage via open taliks [Yoshikawa and Hinzman, 2003] or increased exchange between lake water and groundwater [Walvoord et al., 2012], depending on hydraulic gradients. However, the importance of sublacustrine permafrost thaw in a warming climate is limited by the fact that many lakes in our study areas probably already have a sublacustrine talik through the permafrost. Ground temperature modeling by Smith [1976] showed that, under equilibrium conditions in the Mackenzie Delta region of Canada, taliks through the permafrost should be present under any water body more than about 60 m wide that does not freeze to the bottom. Permafrost temperatures in Smith's study area (-4°C) are comparable to the coldest of our study areas (BELA and CAKR), suggesting that most of our stable lakes already have through taliks. However, Smith's [1976] modeling also showed that centuries are required for a talik under a waterbody to melt through thick (tens of meters) permafrost; more recent simulations show that taliks can form more quickly [Wellman et al., 2013], in a matter of decades in systems with near-surface groundwater [Rowland et al., 2011]. Thus, permafrost is probably present but degrading under many young thermokarst lakes in our study area, and continuing permafrost degradation could in the future open a talik that allows exchange of water through the lake bottom. Recent research by Arp et al. [2016] suggests that shallow lakes in the northernmost part of Alaska have already passed a critical thaw threshold.

In discontinuous permafrost, taliks provide a pathway for groundwater influx, and as taliks expand due to permafrost degradation, more lake ecosystems will be affected by groundwater recharge and discharge. Lake basins with high hydraulic connectivity (i.e., coarse-grained sediments) have the highest potential for

groundwater interactions (i.e., Floodplain-Inactive and Interior Lowland). Increased groundwater influx to thermokarst lake basins, via sublacustrine conduits, combined with warming and drying may lead to increased solute and NO_3^- delivery to lowland lake basins, particularly those basins in floodplains with sufficient hydraulic connectivity to allow repeated groundwater recharge. Warm conditions combined with high evaporative potential and limited precipitation could lead to widespread evapoconcentration of solutes, as has already been observed in the Yukon River basin [Lewis *et al.*, 2015]. The YRB is a region of the subarctic experiencing extreme warm and dry conditions [Monahan and Fisichelli, 2014] where lake surface area has already declined [Riordan *et al.*, 2006].

Lake basins lacking deep groundwater connections, or those recharged by local groundwater sources not dependent on permafrost (e.g., basins in glacial moraines and sand sheets), might be less affected by shifts in permafrost due to the high hydraulic conductivity of coarse-grained soil prevalent in these lake basin types. Lakes on ice-poor moraines and sand dunes may be less affected by warming-related permafrost changes, because they are not subject to thermokarst and their flow paths to groundwater are not strongly dependent on permafrost. Less dramatic shifts in water chemistry and lake volume are possible as a result of warming-related changes in the water balance of their catchments.

5. Summary and Conclusions

This study captures a wide range of landscape-scale variation among lakes in Alaska and provides an important baseline for the region, as well as a context for other limnological surveys that have been conducted throughout the Arctic and Subarctic. Furthermore, it demonstrates the strong association that exists between geomorphology and lake morphometry and chemistry. Using these associations, we developed a landscape classification scheme for lake basins based on parent material (geology and soils), topography, and ground ice content of near-surface permafrost. We then used this classification approach to assess spatial variability in lake physical and chemical properties. The two factors that most affected the physical structure of lakes were lake genesis and ground ice content, while modern lake chemistry was a function of source water input and erosion and sedimentation associated with thermokarst. Numerous climate-related impacts are expected in arctic and subarctic lake ecosystems in the coming years [Vonk *et al.*, 2015], and the lake classifications we developed are useful in making predictions regarding the effects of climate warming on lakes in the Arctic and Subarctic. The physical and chemical response of lakes to projected changes will depend on how climatic changes affect precipitation, temperature, evaporation rates, and permafrost degradation across northern latitudes. Although there is great uncertainty in how climate change will affect an individual lake, the physical and chemical differences observed among the various lake basin types described here illustrate potential pathways for lake change across the subarctic and arctic landscape.

Acknowledgments

We would like to thank the numerous technicians, volunteers, and pilots affiliated with the National Park Service who assisted in data collection for this research. In particular, we thank M. C. MacCluskie and J.P. Lawler for supporting this research and L.E. Weaver for graphic design. We also thank Cam Jones and Kathryn Motter with the Cooperative Chemical Analytical Laboratory, in the College of Forestry at Oregon State University for laboratory support. Funds for this work were provided by multiple entities within the United States National Park Service including the Central Alaska and Arctic Network Inventory and Monitoring Programs, Yukon-Charley Rivers National Preserve, and Gates of the Arctic National Park and Preserve. All supporting data are available on the National Park Service's data portal (<https://irma.nps.gov/Portal>).

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