Polar Lakes and Rivers

Limnology of Arctic and Antarctic Aquatic Ecosystems

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CHAPTER 2

Origin and geomorphology of lakes in the polar regions

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Outline

A characteristic and often dominant feature of many polar landscapes is the great diversity and abundance of their standing surface waters. The focus in this chapter is on lakes and ponds. Antarctica has little surface water by comparison with the Arctic, but many saline lakes exist in the ice-free oasis areas, and freshwater ponds and lakes are abundant in the maritime and peripheral Antarctic regions. The aim of this chapter is to provide a brief introduction to the different origins, distinguishing features, and landscape controls that result in the extraordinary diversity of lakes and ponds that exist in both polar regions. The main emphasis is on the description of the geological and geomorphological processes involved in the formation and modification (natural change) of these high-latitude lake ecosystems. Throughout this review, we have drawn on examples from both the Arctic and Antarctic to emphasize the differences and similarities that exist between north and south polar lake ecosystems.

2.1 Introduction

In both polar regions, permanently frozen soils (permafrost) exert a strong influence on catchment properties such as hydrological processes and geochemical interactions (Lamoureux and Gilbert 2004). Similarly, in both regions snow and ice cover are major controls on the structure and functioning of aquatic ecosystems. In this review, we delimit the northern polar zone, or Arctic, using the southern limit of continuous permafrost, whereas the extent of the southern polar zone, or Antarctica, is defined here by the northern limit of pack ice.

High levels of solar radiation reaching high latitudes in spring result in rapid snowmelt. Runoff in late spring typically comprises 80–90% of the yearly total in the Arctic and lasts only a few (2–3) weeks or even less (Marsh 1990). This pulse is not as prevalent in the Antarctic, as snow cover is less, and much of the snow sublimates or recharges soil moisture rather than running off into streams. Infiltration of the Arctic pulse of water is limited initially by frozen soil and later by permafrost, perennially frozen ground that stays at or below 0°C for at least two consecutive summers (Woo and Gregor 1992). Permafrost may reach depths of 600–1000 m in the coldest areas of the Arctic, becoming discontinuous and patchy (sporadic) in Sub-Arctic regions (Stonehouse 1989). Permafrost is also prevalent in the Antarctic, including the Dry Valleys of the Trans-Antarctic Mountains in south Victoria Land. Cartwright *et al.* (1974) and Decker and Bucher (1977) inferred regional thicknesses of permafrost approaching 1000 m based on local temperatures. However, permafrost is absent below much of the East Antarctic Ice Sheet due to the pressure of the overlying ice which initiates basal melting (Bockheim and Tarnocai 1998). In the McMurdo Dry Valleys of Antarctica, most of the water feeding lakes comes from melting glaciers and follows similar stream paths year after year. Permafrost in the Dry Valleys is relatively thin, presumably because of the presence of the lakes themselves. McGinnis *et al.* (1973) and Cartwright *et al.* (1974) found areas of completely unfrozen sediment beneath thermally stratified saline lakes such as Lake Vanda or Don Juan Pond because of the freezing-point depression due to abundant salts. As permafrost seals the subsoil, spring meltwater may flow over land and enter rivers, or accumulate into the many wetlands, ponds, and lakes characteristic of low-relief tundra environments. Summer sources of water include late or perennial snow patches, glaciers, rain, melting of the upper (active) layer of permafrost, as well as some cases of groundwater discharge (Rydén 1981; Van Everdingen 1990). Groundwater levels and distribution within polar regions are, in general, greatly influenced by bedrock geology, soil thickness, and permafrost layers. Permafrost can control the amount of physical space in the ground available to groundwater as well as its movement within drainage systems.

In the northern hemisphere, ice-sheet advance during the Last Glacial Maximum (LGM) produced a lake-rich postglacial landscape. Using a geographic information system (GIS)-based statistical comparison of the locations of about 200 000 lakes in the northern hemisphere (sized 0.1–50 km^2, northwards of latitude 45°N) with data from global databases on topography, permafrost, peatlands, and LGM glaciation, Smith *et al.* (2007) revealed the apparent importance of LGM glaciation history and the presence of permafrost as determinants of lake abundance and distribution at high latitudes. According to their study, lake density (i.e. the number of lakes per 1000 km^2) in formerly glaciated terrain is more than four times that of non-glaciated terrain, with lake densities and area fractions being on average approximately 300–350% greater in glaciated (compared with unglaciated) terrain, and approximately 100–170% greater in permafrost-influenced (compared with permafrost-free) terrain. The presence of peatlands is associated with an additional approximately 40–80% increase in lake density, and generally large Arctic lakes are most abundant in formerly glaciated, permafrost peatlands (≈14.4 lakes/1000 km^2) and least abundant in unglaciated, permafrost-free terrain (≈1.2 lakes/1000 km^2).

Lithology and geological setting also exert a strong influence on polar lakes. In many cases, faults and fractures control the location of lakes and the characteristics of basins and drainage networks. Many lakes are located in glacially overdeepened, tectonically controlled sites, whereas others are located in craters or in tectonic troughs. Lakes most strongly influenced by fractures are generally of more elongate shape and usually aligned parallel to each other, for example on Byers Peninsula, Livingston Island, Antarctica (López-Martinez *et al.* 1996). Igneous rocks often show abundant *'roches moutonnées'* (elongate rock hills) and the development of 'rock bars' and 'riegels' that may act as dams for lakes. Differences in geological substrates can also affect the extent of rock weathering and the chemical composition of soil water that ultimately discharges into lakes. There are many significant differences in lake and pond characteristics in various parts of the Arctic, which partly reflect gradients of geology, climate, and vegetation, as well as local impacts and other important factors.

2.2 Lake origins

2.2.1 Wetlands

Wetlands and saturated soils are characteristic features of the Arctic because moisture received from rain and snowmelt is retained in the active layer above the permafrost barrier. Additionally, recent glaciated landscapes often have poorly developed drainage patterns that result in frequent ponded water. Due to the generally higher levels of precipitation at lower latitudes, wetlands are more common in the Low Arctic than in the High Arctic. The size of Arctic wetlands ranges from small strips to very extensive plains such as the West Siberian Lowland and the Hudson Bay Lowlands. Wetlands may co-exist with tundra ponds or lakes of various sizes, since both wetlands and lakes occupy areas with abundant storage. Wetlands tend to occur locally in high concentrations in the form of lowland polygon bogs and fens, peat-mounted bogs, snowpatch fens, tundra pools, or as flood-plain marshes and swamps. More detailed descriptions of these and other Arctic wetland types are provided in Woo

(2000). In polar desert regions there is less snow for meltwater to accumulate; however, ponds and lakes are still present due to the surface ponding of water by the permafrost. In some Arctic regions, winter snow is plentiful and ponds and shallow lakes occupy large areas, often forming networks of static waterways. On ice-scoured uplands and coastal flats the landscape may actually be dominated by ponds and lakes. Coastal plains often feature long, shallow lakes that are separated and aligned by raised beaches and occupy more than 90% of the terrain (Figure 2.1). Although a variety of definitions exist for distinguishing lakes from ponds, we define a pond here as a freshwater basin that is sufficiently shallow so that it freezes totally to the bottom in winter, whereas a lake always has liquid water in its basin.

2.2.2 Ice-dependent lakes

Lakes form also on poorly drained ice, next to glaciers and ice caps, on ice sheets, and between ice sheets and ice-filled moraines (Figure 2.2). Some are long-lived, fed by annual influxes of meltwater, while others are temporary systems that are flooded during the summer melt and liable to sudden drainage through rifts in the ice and other causes.

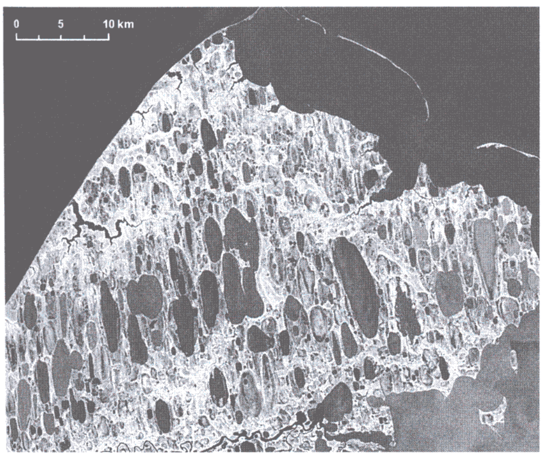

Figure 2.1 Landsat 7 image of thermokarst thaw lakes located on the northern coastal plain near Barrow, Alaska. The image was acquired on 30 August 2000. Source: Kenneth Hinkel and Benjamin Jones.

Ice-dammed lakes are more common in the Arctic than elsewhere and in Greenland all the larger lakes are of this type. Glaciers or ice dams have formed some of the lakes in Iceland (e.g. Grænalón), and such lakes are occasionally emptied beneath the damming ice, sometimes resulting in *jökulhlaups* (Icelandic for 'glacier-burst' or catastrophic drainage of lakes). Ice-dammed systems develop next to glacier fronts in mountainous terrain or besides ice shelves (epiglacial) and in depressions on glaciers, ice sheets, and ice shelves (supraglacial; e.g. Ward Hunt Ice Shelf in the Canadian High Arctic, McMurdo Ice Shelf in Antarctica; see Vincent 1988 and Plates 2 and 3). Epiglacial lakes are common in Antarctica given that, in most cases, the source of water that sustains the lakes is from glaciers. They can persist for many years with frequent changes in water level and morphology resulting from glacial movements and changing meltwater inputs (Hodgson *et al.* 2004), and are also prone to periodic draining (Mackay and Løken 1974; Smith *et al.* 2005). Supraglacial lakes are often ephemeral, forming during the summer melt. These systems have been proposed as potential refugia for microbiota in the controversial Snowball Earth hypothesis when the Earth underwent extreme freeze–thaw events between approximately 750 million and 580 million years ago (Vincent *et al.* 2000).

Another class of ice-dependent water bodies is 'epishelf lakes'. These are a form of proglacial lakes in which floating ice shelves dam freshwater runoff. Epishelf lakes are hydrologically connected to the ocean and therefore are tidally forced. Some epishelf lakes have saline bottom waters while some (e.g. Doran *et al.* 2000) are fresh throughout. In stratified epishelf lakes, inflows of meltwater are impounded by glacial ice, and the density difference between fresh and salt water prevents the mixing of the runoff with the marine water below. Many examples of epishelf lakes are known from Antarctica, including Moutonnée and Ablation lakes on the coast of Alexander Island, in the Schirmacher Oasis and Bunger Hills (reviewed in Hodgson *et al.* 2004). Beaver Lake located in the Radok Lake area and associated with the Amery Ice Shelf (Prince Charles Mountains, Antarctica) is especially interesting, because it is probably the world's largest epishelf lake (McKelvey and

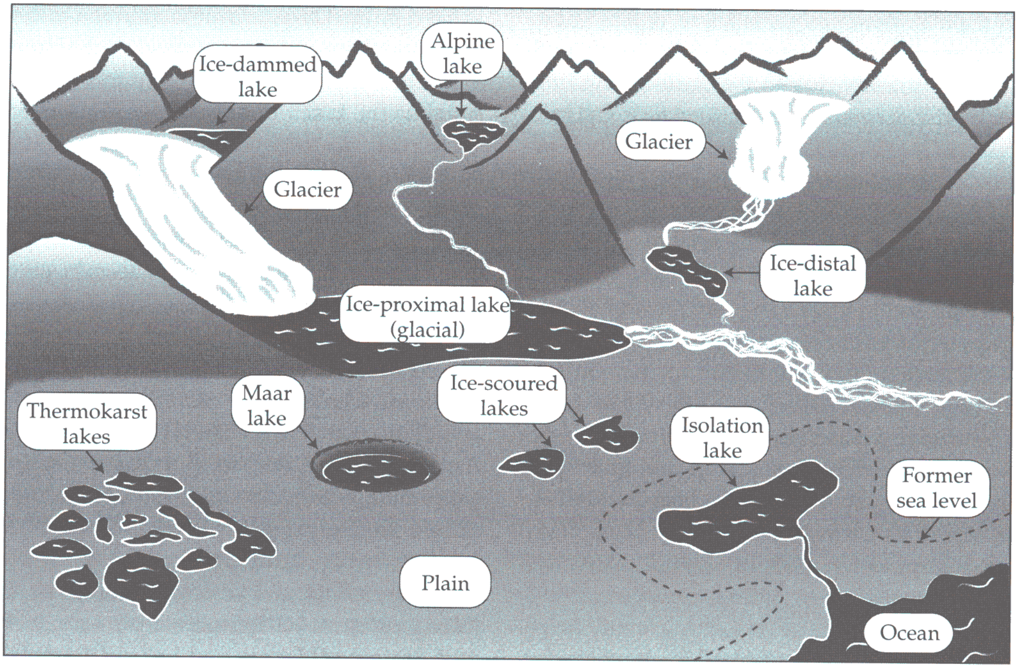

Figure 2.2 Some of the diverse lake types that can be found in the polar regions.

Stephenson 1990; Laybourn-Parry *et al.* 2001). Epishelf lakes are becoming increasingly rare in the Arctic, as one of the few existing systems in Disraeli Fiord was recently destroyed by the break-up of the Ward Hunt Ice Shelf (Mueller *et al.* 2003), and another in Ayles Fiord by the collapse of the Ayles Ice Shelf. In these cases, the freshwater layers escaped and drained into the Arctic Ocean. Only one epishelf lake has been confirmed to still exist in the Arctic, in Milne Fiord. Epishelf lakes can be thought of as windows in a proglacial estuary, where the position of the window along the estuary gradient and the depth of the lake dictate lake chemistry. A review of the formation and dynamics of Antarctic lake ecosystems (including epishelf, epiglacial, subglacial, and supraglacial lakes) and their paleolimnology is provided by Hodgson *et al.* (2004).

2.2.3 Postglacial lakes

Glacial erosion and deposition by former continental and local ice masses have left behind a vast number of lakes in the polar and temperate regions of the world. The irregularity of residual landscapes caused by glacial processes has resulted in a diverse range of lakes, many of which still exist due to the relatively young age of these landscapes (Benn and Evans 1998). In polar regions that have not been subjected to late Pleistocene glaciation, most lakes are of non-glacial origin and are described in other sections of this chapter.

Differential substrate erosion by ice is caused by large-scale patterns of ice flow, as well as by localized flow controlled by topographic relief and the resistance of substrates to erosion. Hence, in areas that have been subject to pronounced ice flow, scouring and erosion of bedrock and surficial sediments show broad-scale linear alignment or radial patterns. This is particularly true where basal ice is at melting temperature and meltwater is available to further increase the entrainment and erosion of subglacial sediments (Benn and Evans 1998). Subject to pre-existing topography, persistent ice flow will differentially erode surfaces and generate quasi-linear troughs, as well as more irregular scour

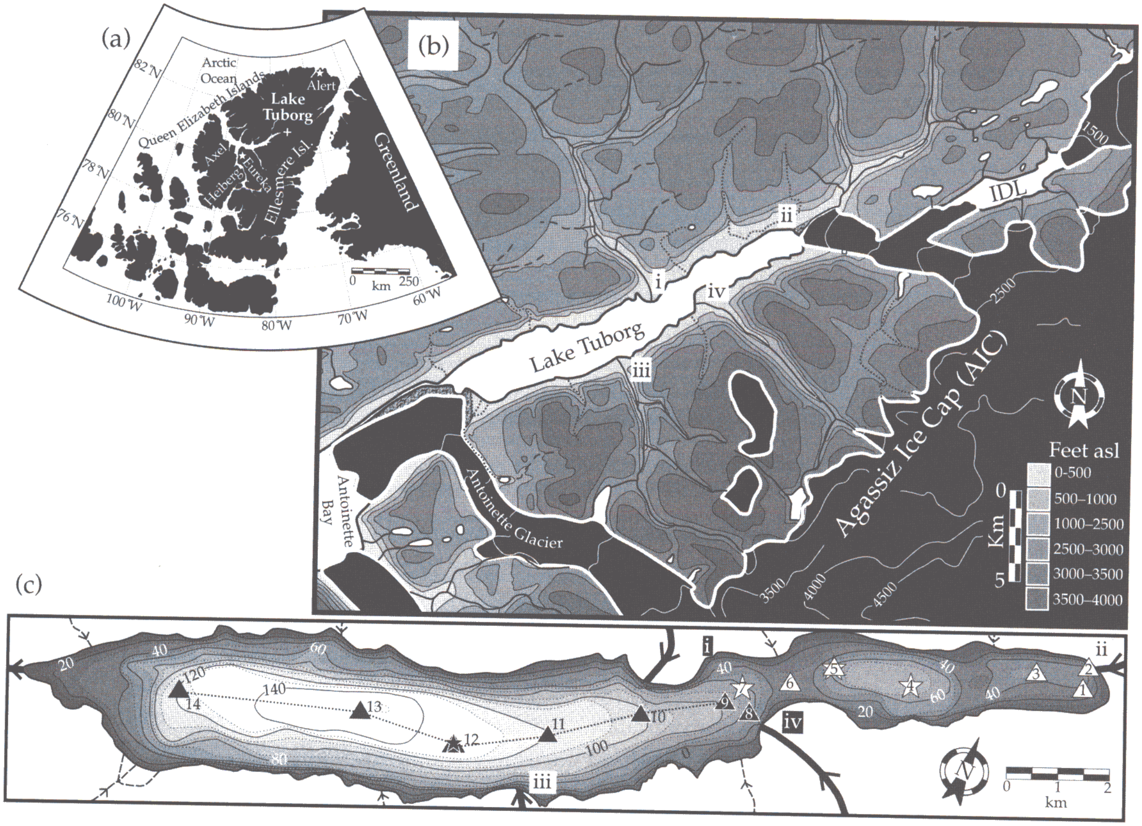

Figure 2.3 Bathymetry of glacially eroded, fiord-like Lake Tuborg on Ellesmere Island, Nunavut, Canada. Solid bathymetric contours are shown at every 20 m. asl, above sea level. From Lewis *et al.* (2007).

depressions. In mountainous terrain, glacial erosion is further constrained by pre-existing topography and focuses erosion in the valley bottom and walls, producing the U-shaped cross-section typical of glacial valleys. As erosion by ice is not constrained longitudinally to a fixed downstream base level, as is the case with fluvial processes, localized scour and erosion generate closed depressions that subsequently form lakes, as exemplified by Lake Tuborg on Ellesmere Island in the Canadian High Arctic (80°59′N, 75°33′W; Figure 2.3). Reduced erosion along glacier margins, together with the frequent deposition of morainal material, further generate sedimentary dams and form lakes behind former ice margins. In this regard, lakes formed by glacial scour that fill valleys are the terrestrial equivalents of fiords that are common features in many formerly glaciated coastal areas.

In regions with resistant bedrock, the differential mechanical strength of the available lithology may result in localized and selective erosion of less-resistant units. Although this may further reinforce pre-existing topography, the result can also be a complex array of subglacial depressions that form lake basins. The resultant landscape can be covered with a large number of lakes that have diverse shapes and morphometries. Excellent examples of these types of lakes are found on the large Precambrian shields of northern Canada and Scandinavia (Figure 2.4).

Glacial processes and the resultant glaciogenic sediments provide a wide range of settings for the ultimate formation of depressions and lakes, mainly through the emplacement of low-permeability tills and kettling. Thick glaciofluvial deposits along the margin of ice can generate sedimentary dams

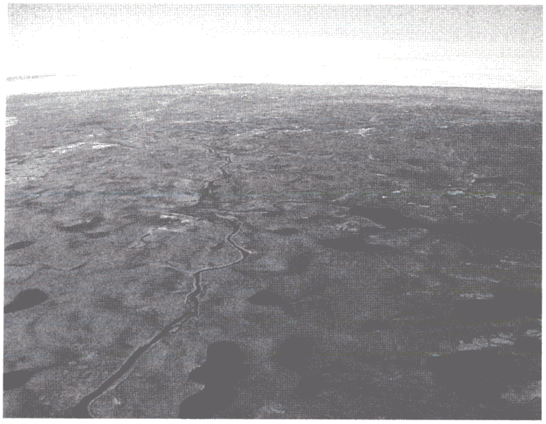

Figure 2.4 Canadian Shield lakes on the Boothia Peninsula (Nunavut) near the northern coast of Canada. Photograph: S. Lamoureux.

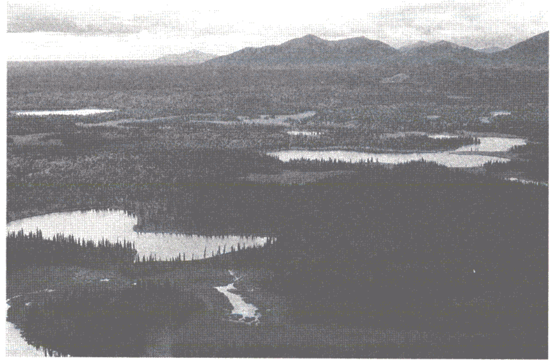

Figure 2.5 Kame lakes at the edge of Ahklun Mountains, southwestern Alaska, USA. Photograph: S. Lamoureux.

on tributary valleys or cross pre-existing drainage patterns on low-relief landscapes. The results are frequent settings for the ultimate formation of lake basins. Similarly, sediments can accumulate around stagnant ice margins and blocks. Referred to as kame deposits, the subsequent ice melt generates depressions that can become lakes with irregular morphology (Figure 2.5).

The proglacial environment is highly conducive to the formation of lakes through high rates of sediment transport and deposition. Proglacial river systems may rapidly form during ice advance, maximum stand, or retreat. Sediments deposited at the mouths of valleys may block pre-existing drainage and form lakes, in much the same way that a landslide or other mass movement may block drainage. In some instances, sediment transported from a tributary valley may form a fan or delta into another lake, effectively separating the former lake into two. In the case of coastal environments, progradation of sediments may isolate marine embayments and result in the ultimate formation of a freshwater or meromictic lake (see Section 2.2.5 below).

Finally, the impact of glaciation on landscapes typically lasts long after ice melts. After ice ablates, the barren landscape is especially susceptible to erosion. The wide variety of glacial sediment deposits are typically reworked extensively during the postglacial period, resulting in increased erosion, sediment yield, and drainage changes. Rivers exhibit increased sediment loads that can fill shallow lake basins and form new fans and deltas into larger water bodies and damming of downstream valleys. The result is relatively rapid alteration of initial lakes and the formation of new lakes. These processes in the deglaciated environment were defined as paraglacial by Church and Ryder (1972) and have been recognized subsequently as important in postglacial landscape evolution in many formerly glaciated regions. In some cases, evidence for continued paraglacial influences has been documented in the Canadian Cordillera more than 10 000 years after deglaciation (Church and Slaymaker 1989).

2.2.4 Thermokarst lakes and ponds

Thermokarst ponds and lakes (also called thaw lakes) are common features of Arctic landscapes, developing in depressions that result from the thawing of permafrost (Figure 2.6; Plates 8 and 9). Most of these water bodies are shallow (<1 m depth), and they are widespread in lowland areas of western and northern Alaska, Canada, and Siberia. For example, thaw lakes comprise approximately 90% of the lakes in the Russian permafrost zone (Walter *et al.* 2006) and are widespread across much of the northern slope of Alaska (Hinkel *et al.* 2005) and the coastal lowlands of Canada (Côté and Burn 2002).

Figure 2.6 Forest-tundra landscape with abundant thermokarst lakes and ponds near treeline in northern Québec. Photograph: I. Laurion (see also Plates 8 and 9).

The process of thermokarst formation is triggered by the degradation of ice wedges, followed by the subsidence of the surface and the presence of ice beneath the sunken area, which leads to the formation of ponds. The pond then accelerates thermokarst formation because the heat released from its water body during winter promotes the formation of a talik (an unfrozen zone in permafrost that is located under a lake or deep pond that results from heat release from the overlying water column, thereby preventing formation of ground frost during winter) below its basin. Eventually, the reduction of permafrost can lead to complete drainage of the pond and subsequent dry-up. Depending on atmospheric conditions, palsas may form after refreezing of the surface and permafrost may form again. Residual depressions and cross-cutting relationships of thaw lakes suggest that they are relatively short-lived features and are sensitive to underlying bedrock, surficial sediment, and climate conditions (Hinkel *et al.* 2005), including predominant wind directions (Côté and Burn 2002).

Studies into the dynamics and evolution of these widespread thaw lakes in the Northern Hemisphere receive great attention as they contribute to a better understanding of their role in the global atmospheric methane budget, a potent greenhouse gas with highest concentrations between latitudes 65° and 70°N (Hinkel *et al.* 2003; IPCC 2007). Walter *et al.* (2006) linked a 58% increase in lake methane emissions during recent decades to the expansion of thaw lakes in northern Siberia, demonstrating the importance of this feedback to climate warming (Smith *et al.* 2005).

2.2.5 Coastal uplift systems

Glacioisostatic rebound or uplift is the process by which the Earth's crust, once depressed by the weight of overlying ice, begins to rise following the retreat of continental ice sheets. In many polar regions, this rebound is still occurring today, although at a much slower rate than immediately following glacial retreat. In areas such as eastern Hudson Bay (Canada), for example, the landscape continues to rise at a rate of about 1 m per century (Allard and Tremblay 1983). New lakes are thus still being formed as depressed, scoured land emerges from the sea (Fulford-Smith and Sikes 1996; Lamoureux 1999; Saulnier-Talbot *et al.* 2003). Hence, lakes closer to present-day sea level are often younger than those located at higher elevations further inland. For example, coastal Nicolay Lake in the Canadian High Arctic has effectively been a lake for only approximately 500 years. Glacioisostatic submergence of the land to over 100 m or more below current sea level effectively moved coastlines inland by tens of kilometers for much of the Holocene. As a result, the Nicolay Lake basin remained marine until the late Holocene, and freshening and fluvial processes that deliver sediment and related terrestrial materials were captured by upstream lake basins until approximately 500 years before present (Lamoureux 1999). Recent research has demonstrated that this type of lake evolution may follow a number of divergent paths, and results in a range of lake types ranging from fully fresh to hypersaline (Van Hove *et al.* 2006). In many instances, glacioisostatic rebound results in the formation of meromictic saline lakes that become stratified for part of the year (monomictic) or are permanently stratified (meromictic). Sea water that is trapped in the lake during the isolation process is frequently preserved in a dense saline hypolimnion and the resultant stable water column may last for thousands of years, or indefinitely. Numerous examples of coastal meromictic lakes have been documented in the polar regions and are of particular interest due to their unique limnology and sedimentary records (reviewed in Pienitz *et al.* 2004). Long-term preservation of meromixis appears to be dependent on a number of factors, including lake morphometry, ice cover, and the relative loading of freshwater to the lake from the surrounding catchment. Meromictic lakes have a surface freshwater layer that mixes with the wind (mixolimnion), but permanently ice-covered Antarctic lakes remain stratified throughout their water column due to a combination of salt and the protection offered by the ice cover, and therefore the term meromictic may not properly apply.

2.2.6 Meteoritic impact crater lakes

Approximately 174 meteoritic impact craters have been identified so far on the Earth's surface (Earth Impact Database, Planetary and Space Science Centre: www.unb.ca/passc/ImpactDatabase/index.html), and include the Arctic sites El'Gygytgyn (67°50′N, 172°E) and Popigai (71°30′N, 111°E), both in northern Russia, and Lac à l'Eau Claire (Clearwater Lake) in northern Québec (Plate 9). A striking example of a meteoritic impact crater lake in North America is Pingualuk Crater Lake. This crater is located at 61°N, 74°W in the northernmost part of the Ungava Peninsula in northern Québec (Nunavik, Canada). The crater was formed from the impact of a meteorite that entered Earth's atmosphere approximately 1.4 million years ago, as determined by Ar/Ar dating of the impactites collected at the site (Grieve *et al.* 1989). Pingualuk is situated close to the area where the inland ice masses reached maximum thickness during the last (Wisconsinan) glaciation. The crater is believed to have formed a subglacial lake basin under a dome of the Laurentide Ice Sheet with its waters remaining liquid below the ice due to the pressure of the overlying ice cap and to geothermal heating from below, thereby representing conditions similar to those found in modern subglacial systems in Antarctica (e.g. Lake Vostok; see Section 2.3.3). The crater is a circular depression about 400 m deep and 3.4 km in diameter, hosting a lake presently 267 m deep with no surface outflow (Figure 2.7). The deep sediment infill of the Pingualuk Crater Lake, which escaped scouring and erosion from ice flow (Figure 2.8), promises to yield an uninterrupted, approximately 1.4-million-year Arctic paleoclimate record of several interglacial–glacial cycles. These unique long sediment records may offer unique insights into long-term climatic and environmental change in high-latitude regions, thereby having the potential to significantly enhance our understanding of past climate dynamics in the Arctic (Briner *et al.* 2007; see also Chapter 3 in this volume).

Similarly, Lake El'Gygytgyn in northeastern Siberia, an impact crater created approximately 3.6 million years ago, is the only site with a continuous terrestrial Arctic climate record that covers several interglacial periods (Nowaczyk *et al.* 2002; Brigham-Grette *et al.* 2007) as its lake basin was created in the center of what was to become Beringia, the largest contiguous landscape in the Arctic to have escaped northern-hemisphere glaciation. An overview of papers detailing the remarkable paleolimnological record of the last 250 000 years from El'Gygytgyn Crater Lake is available in Brigham-Grette *et al.* (2007).

2.2.7 Volcanic lakes

One important type of volcanic lake are maar lakes (Figure 2.2), which are shallow, broad, low-rimmed craters formed during powerful explosive eruptions involving magma–water (phreatomagmatic) interaction. The accumulation of material ejected from craters contributes to the formation of their surrounding rims. A few maar lake systems exist in explosion craters on Iceland (Sigurdsson and Sparks 1978), whereas they are quite widespread in Alaska. Zagoskin Lake and all other major lakes on St. Michael Island (Alaska) are maar crater lakes (Ager 2003; Muhs *et al.* 2003). Ranging from 4 to 8 km in diameter, the four Espenberg Maars (Whitefish Maar, Devil Mountain Maar, and North and South Killeak Maars) on the northern Seward Peninsula just south of the Arctic Circle in northwestern Alaska have been described as the largest known maar craters on Earth (Beget *et al.* 1996). They were formed by a series of Pleistocene basaltic eruptions through thick (approximately 100 m) permafrost,

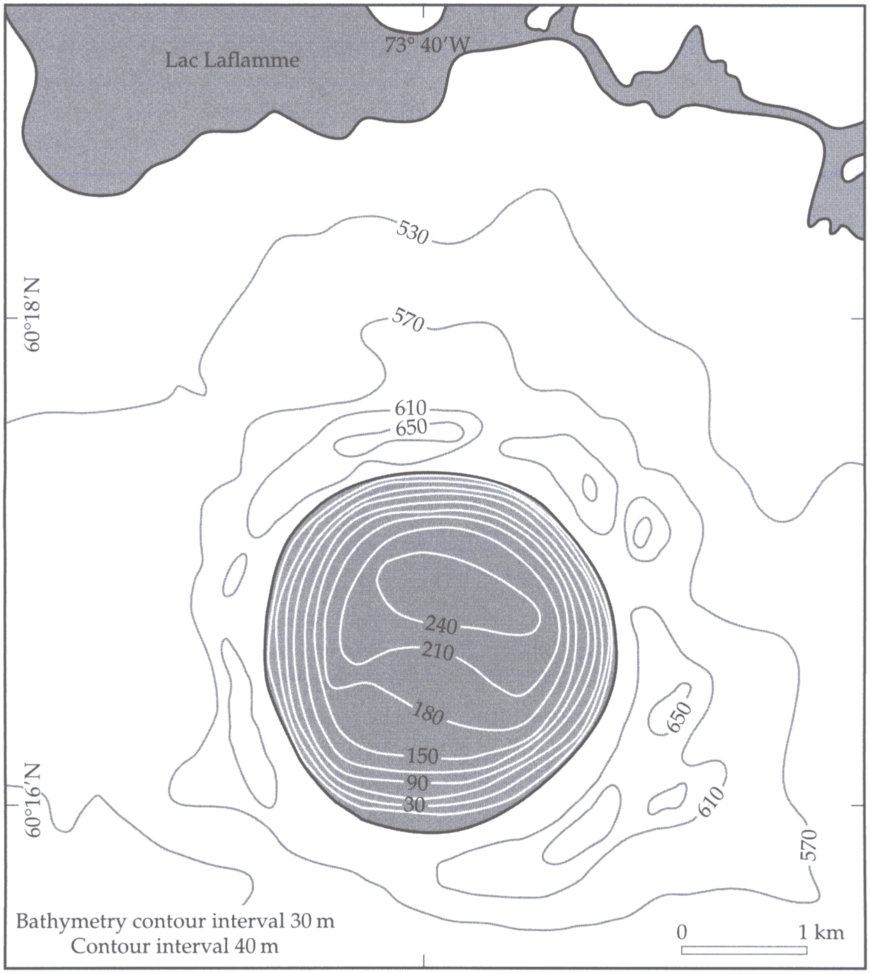

Figure 2.7 Location of Pingualuk Crater Lake within Pingualuit impact structure, Parc National des Pingualuit, Nunavik (northern Québec, Canada). The topographic and bathymetric contour intervals shown are 40 and 30 m, respectively. Pingualuk is a closed lake basin, without any inlet or outlet.

excavated as much as 300 m into older lithologies. There are also volcanic crater lakes (not maars) in the Ingakslugwat Hills in the south-central Yukon Delta (Ager 1982). Other crater lakes (some of which may be maars) are located on Nunivak and St. Lawrence Islands in southwestern Alaska, whereas some sizable lakes in calderas can be found in southwestern Alaska (e.g. Aniakchak crater on the Alaska Peninsula, approximately 3600 years before present). Craters differ from calderas both in size and origin. Craters are much smaller features than calderas and are typically defined as being less than 1 km in diameter. Although both craters and calderas are most often associated with explosive eruptions, craters are typically formed by the explosive ejection of material in and surrounding the upper part of the conduit, rather than by collapse. Steep-walled pit craters, in contrast, often found on shield volcanoes, are more passive features formed when magma drains from a fissure, leaving overlying lava flows unsupported. Multiple explosive eruptions can form overlapping or nested craters, and adjacent craters may reflect localized areas of eruption along fissures, as seen in rows of small

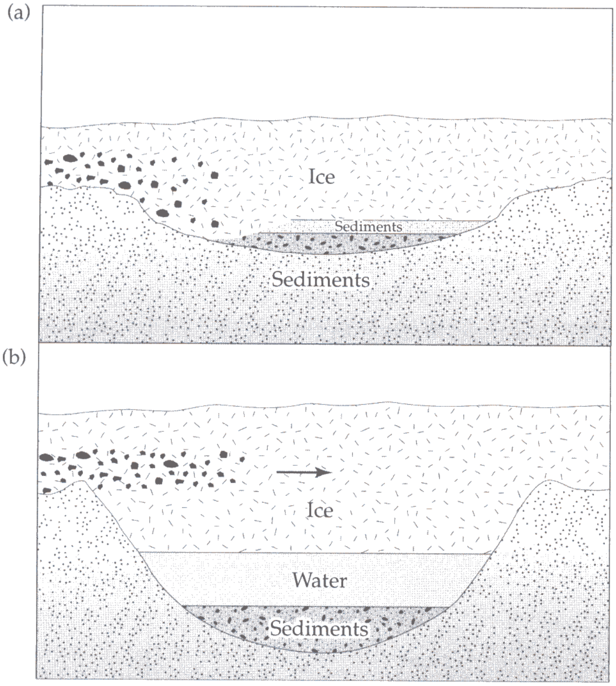

Figure 2.8 (a) Glacial erosion of bottom sediment deposits in shallow lake basins within formerly glaciated Arctic regions. (b) The situation at Lake Pingualuk (northern Québec, Canada), where lake sediments have escaped glacial erosion due to the extremely deep and steep-walled crater basin, perhaps analogous to Antarctic subglacial lakes (see Chapter 7).

crater (maar) lakes in Iceland that extend for tens of kilometers.

2.2.8 Karst systems

Lakes of karstic origin are relatively rare in high latitudes due to the scarcity of limestone and other rocks suitable for karst development in many locations, in addition to the pervasive permafrost conditions that restrict the infiltration of water into the vadose (unsaturated soil) zone and the absence of effective groundwater systems. The discovery of mineral springs in polar settings (Andersen *et al.* 2002) and groundwater drainage in permafrost locations like the Ungava Peninsula, northern Québec (Lauriol and Gray 1990) suggest that the role of groundwater in permafrost regions may be more extensive than commonly thought and that these regions hold the potential for karst systems. This potential has been realized to some extent by studies carried out on Svalbard (Norway) where karst systems in Paleozoic carbonate and evaporate bedrock maintain open taliks and several karst lakes have been produced (Salvigsen and Elgersma 1985). Further evidence of karst systems that occurred during hypsithermal (early to mid-Holocene) conditions in the northern Yukon (Clark *et al.* 2004) suggest that karst processes may have contributed to lake formation in the past, although no systems have been specifically recognized to date.

2.2.9 Tectonic lakes

Generally rare in the polar regions, tectonic lake basins tend to be old systems compared to the relatively young age of most high-latitude lakes. Although the origins of Icelandic lakes are diverse, a large number are of tectonic origin, including the two largest Icelandic lakes, Thingvallavatn (84 km^2) and Thorisvatn (70 km^2). Subglacial systems such as Lake Vostok in Antarctica may also be considered tectonic lakes (Studinger *et al.* 2003; see below).

2.2.10 Lakes of other origins

The flood-plains bordering large Arctic rivers are characterized by meandering or anastomotic streams, oxbow lakes and bayous, marshes, or stagnant pools (Plate 8). Flood-plain lakes that have been cut off from river channels are seasonally flooded with river water (usually in the spring when the river overflows its banks) and can be found in great numbers within the deltas of the Mackenzie and Lena rivers in northwestern Canada and Siberia, respectively. Small, crescent-shaped lakes called oxbow lakes can form in river valleys as the result of meandering. The slow-moving river forms a sinuous shape as the outer side of bends are eroded away more rapidly than the inner side. Eventually a horseshoe bend is formed and the river cuts through the narrow neck. This new passage then forms the main passage for the river and the ends of the bend become silted up, thus forming a bow-shaped lake.

The well-known icelandic Lake Mývatn (37 km^2) is a basin formed in a collapsed lava flow. Other Icelandic lakes have been formed by rock slides or glacial deposits. Lagoon lakes are common on the sandy shores, Hóp (45 km^2) being the largest of this type and the fourth largest lake in Iceland.

There are also artificial or human-made lakes or reservoirs within the Arctic that have been created along Arctic rivers for the purpose of hydroelectric power generation. In Canada and Russia, some of these reservoirs are quite large, for example the La Grande complex in northern Québec (9900 km^2 of flooded surface area) and the Krasnoyarskoye More ('Krasnoyarsk Sea') reservoir on the Yenisey River in Siberia (2000 km^2).

2.3 Geographical regions

2.3.1 The circumpolar Arctic

Repeated glaciations during the **Pleistocene** epoch resulted in glaciers and vast inland ice sheets covering much of the eastern and central North American and western Eurasian Arctic regions. These large ice masses carved out the land as they advanced over it, gouging out topsoil and broken rock. The many structural depressions left behind filled with water once climate warming caused glaciers to retreat, forming lakes of glacial origin. A characteristic feature of the Arctic, therefore, is the large numbers of lakes and ponds. In the Province of Murmansk in Russia, there are more than 100 000 lakes, the largest of which is Lake Imandra with an area of 812 km^2 and a maximum depth of 67 m (NEFCO 1995). Iceland, Sweden, and Finland have approximately 2.7, 5.2, and 5.8% of their territory occupied by lakes, respectively, whereas only about 0.5% of Alaska is covered by fresh water (CAFF 1994). In the Canadian Arctic north of latitude 60°N, approximately 18–20% of the landscape is covered by surface waters (Prowse and Ommanney 1990). The Canadian Arctic contains more than 28 large lakes in excess of 600 km^2 in surface area, including Great Bear (31 326 km^2) and Great Slave (28 568 km^2) lakes located on the mainland (Mackay and Løken 1974). In particular, there are innumerable small lakes in the Arctic (also called ponds). Although a variety of definitions exist for distinguishing lakes from ponds, the latter are generally defined as sites that are sufficiently shallow for their water column to freeze entirely to the bottom in winter, whereas lakes always preserve a lens of liquid water within their basin (see above). Due to the extreme harshness of High-Arctic winter climates, ice thicknesses of between 4 and 5 m are possible, and thus by this definition sites shallower than about 5 m could be classified as ponds. However, most Arctic ponds are much shallower, and are often less than 2 m in depth. Despite their shallow depths, these ponds often dominate Arctic landscapes and are important habitats for waterfowl and other animals.

2.3.2 Coastal Antarctic lakes

The Antarctic ice cap is a dome of ice exceeding 4 km thickness in places, with thinning towards the continental edges. In these ice-marginal areas, some ice-free areas occur that have been variably named oases, dry areas, and dry valleys. Ice-free areas in Antarctica only comprise 0.35% of the continent (Fox and Cooper 1994). Much of this dry land is in the form of nunataks formed by the Trans-Antarctic Mountains. In this review, only the coastal dry areas are of relevance, as most nunataks are at elevations that are too high for the occurrence of ice melt and lake formation.

The majority of Antarctica's lakes are found in coastal oases such as the McMurdo Dry Valleys, Vestfold Hills, Larseman Hills, Bunger Hills, Schirmacher Oasis, and Syowa Oasis (Plates 4 and 5; see Chapter 1 of this volume for descriptions). The Antarctic Peninsula is also home to many ice-free areas, some containing lakes (Hodgson *et al.* 2004). Many coastal Antarctic lakes have been formed as a direct result of isostatic rebound after postglacial retreat of ice (e.g. Zwartz *et al.* 1998; Hodgson *et al.* 2001; Verleyen *et al.* 2004). Coastal lakes can evolve into either closed (no outflow) or open basins (with outflow), and both types are common in Antarctic oases, as well as on the Antarctic Peninsula and the maritime islands (Hodgson *et al.* 2004). Trapped sea water can be flushed out with fresh water (e.g. Pickard *et al.* 1986), but closed basin lakes typically become saline due to the evaporative concentration of salts over time. Antarctic saline lakes are among the saltiest water bodies on Earth, with lakes in the Vestfold Hills and the McMurdo Dry Valleys having salinities ranging from 4 to 235 $g\,l^{-1}$ (Spigel and Priscu 1998; Gibson 1999). Lake Vida, which today occupies the center of Victoria Valley, is the largest modern lake in the Dry Valleys and was previously thought to be frozen to its bed (Calkin and Bull 1967; Chinn 1993). However, Doran *et al.* (2003) provided evidence for an extensive brine body beneath its 19-m-thick ice cover sealed from the atmosphere for at least 2800 ^{14}C years. The NaCl brine remains below –10°C year-round and its salinity exceeds seven times that of sea water, thereby representing a unique new type of lake: an ice-sealed lake.

Lake basins formed in scoured rock by glacial erosion and ice-cap retreat are common in the maritime Antarctic islands such as Signy Island, King George Island, and Livingston Island. Another type of Antarctic lake forms where former marine embayments or fiords are dammed by advancing ice shelves and isolated from the sea. In these lakes, the marine water has been replaced over a period of time by glacial meltwaters. Where the ice shelf forms a complete seal, these lakes will be similar to the saline lakes discussed above. Where an incomplete seal is formed and a hydrological connection to the sea persists under the ice shelf, an epishelf lake is formed. Lake Untersee in East Antarctica is one of the largest epiglacial lakes on the continent (Wand and Perlt 1999). White Smoke Lake in the Bunger Hills, almost 50% of which is bordered by glacier ice and also an epishelf lake, has maintained its rough dimensions for at least 3000 years (Doran *et al.* 2000).

Coastal lakes of lagoon origin are very common in the lower beaches of maritime Antarctica. They have gently sloping edges and are either subcircular or elongate in plan (Jones *et al.* 1993; Cuchí *et al.* 2004), and their salinities tend to be high because of intermittent exchange with sea waters.

2.3.3 Antarctic and Arctic subglacial lakes

Under certain circumstances, liquid water masses can be maintained beneath glaciers and ice caps, such as the Antarctic ice sheet. Subglacial water forms under thick ice sheets because geothermal heat is trapped by the insulating effect of the ice. The pressure of the overlying ice also depresses the freezing point by a few degrees. Water then pools in subglacial depressions to form lakes. Lakes are detectable mainly through aerial radar surveys which have now detected at least 145 individual lakes beneath the Antarctic ice sheet (Siegert *et al.* 2005). New evidence (Wingham *et al.* 2006) in Antarctica suggests that these lakes are interconnected, creating vast subglacial drainage systems. Recent studies show that several subglacial lake systems may have existed beneath portions of the Laurentide Ice Sheet in North America near the last glacial (Wisconsinan) maximum, such as subglacial Lake McGregor in south-central Alberta (Munro-Stasiuk 2003) and possibly Lake Pingualuk in northern Québec (see Section 2.2.6).

The largest and most studied subglacial lake is Lake Vostok, which lies beneath the Russian Vostok Station, in Antarctica (Plate 1). Lake Vostok has been suggested to be in a basin created by tectonic thrusts (Studinger *et al.* 2003), but this interpretation remains controversial because it and the majority of Antarctic subglacial lakes are developed on basement rocks that became part of the stable craton in the Precambrian and there have been few opportunities for more recent faulting to create basins in which subglacial lakes can form. Whether there was a preglacial Lake Vostok is also controversial. For a lake to survive from preglacial times into the present, a fairly specialized set of circumstances would have to occur, but models for this have been put forth (Pattyn *et al.* 2004) and arguments made against the concept (Siegert 2004). Without direct evidence, all we can say about the history of formation of subglacial lakes is that they began forming when ice achieved sufficient thickness to cause the glacial bed to be above the pressure melting point. A full discussion of subglacial lakes can be found in Chapter 7.

2.4 Effects of landscape evolution and climate change on polar lakes

Landscapes continue to evolve, and aquatic ecosystems including lakes and ponds are transient features of the polar regions that can experience rapid changes, from initial formation until the eventual basin filling by abiotic and biotic sediments. Climate change has the potential to affect lake evolution through a variety of processes, especially in the polar regions where even small changes in temperature can have profound impacts on landscape properties such as snowpack, permafrost, glacial melt, and hydrological inputs, as well as soil and vegetation stability. These complex controls are exemplified by the Holocene development of small lakes near Toolik Lake on the north slope of the Brooks Range, Alaska (Hobbie 1980, 1984; O'Brien *et al.* 1997). Glacial deposits with different composition and permeability resulted in the establishment of different tundra vegetation communities. Gradual evolution of the postglacial landscape, coupled with vegetation succession and Holocene climate changes, all combined to generate substantially different lake-watershed systems (Oswald *et al.* 2003).

The climate-induced dynamics of alpine glaciers and ice sheets exert a strong control on landscape evolution and geomorphological processes that determine the abundance, distribution, and form of lake basins. Climate change can modify erosional patterns and landscape morphology in the catchments of lakes and ponds, for example through changes in evaporation/precipitation patterns, water-induced erosion and the thawing of permafrost. In turn, these changes will affect the transport of sediments from the surrounding terrestrial landscape to lakes and therefore the extent and rate of infilling. The glacioisostatic readjustment of formerly glaciated land masses also has the potential to radically alter the influence of marine waters on coastal lakes and lagoons at high latitudes (see above).

Ice-bound lakes are especially sensitive to small variations in climate. In the Canadian High Arctic, for example, the extensive ice shelves that once dammed the northern fiords of Ellesmere Island (e.g. Ayles, Markham, M'Clintock, Ward Hunt) experienced considerable contraction, fracturing, and ultimately break-up during the twentieth century (Vincent *et al.* 2001). This has resulted in the erosion and loss of many ice-dammed epishelf lakes (see above), and most recently the freshwater layer of Disraeli Fiord was completely drained away as a result of the break-up of the Ward Hunt Ice Shelf in 2002 (Mueller *et al.* 2003). Likewise, the catastrophic drainage events or *jökulhlaups* of ice-dammed lakes are likely to become more extensive and frequent in the polar regions (Lewis *et al.* 2007).

Recent so-called change-detection studies show that climate warming is already having profound effects on thermokarst landscapes in the northern hemisphere, with lake expansion occurring, especially in continuous permafrost regions (Osterkamp *et al.* 2000; Christensen *et al.* 2004; Payette *et al.* 2004; Smith *et al.* 2005), but shrinkage or disappearance in regions with discontinuous, sporadic, and isolated permafrost (Yoshikawa and Hinzman 2003; Riordan *et al.* 2006). In the former case, thawing permafrost triggers thermokarsting and associated lake growth, whereas in the latter scenario it enhances water infiltration to the subsurface and underlying groundwater systems through taliks. Thermokarst lakes are thus very fragile systems (sentinels) that are affected by climate change and associated permafrost dynamics in high-latitude regions. Because of this, research into the response of these freshwater bodies to accelerated permafrost thaw and their contribution to greenhouse gas emissions (e.g. release of methane and carbon dioxide to the atmosphere via ebullition; enhanced methanogenesis in sediments through higher lake productivity) is at the forefront of recent and ongoing scientific efforts in the circumpolar Arctic (e.g. Zimov *et al.* 2001; Christensen *et al.* 2004; Walter *et al.* 2006).

2.5 Conclusions

A full understanding of lake formation and distribution throughout the polar regions requires consideration of many factors including geology, topographic relief, glacial history, climate and periglacial processes, substrate permeability, permafrost properties, peatland distribution, groundwater movement, and talik depths, to name just a few. Formerly glaciated lowland environments possess the greatest abundance of lakes in the northern hemisphere, whereas widespread permafrost can result in the persistence of lakes by presenting a barrier to water infiltration to the subsurface. In the Arctic, the spatial and temporal distribution of Arctic lake systems is changing rapidly due to amplified climate warming and permafrost thaw, with anticipated profound impacts on the hydro-ecological processes and the quality and availability of fresh waters in these landscapes (ACIA 2005; IPCC 2007; Plate 16). A key issue in the context of these changes is whether lakes and ponds that often dominate Arctic landscapes will act as sources or sinks of greenhouse gases. With the exception of the Peninsula region, the Antarctic continent has been less impacted by global warming as yet (Doran *et al.* 2002; Thompson and Solomon 2002), but models show (Shindell and Schmidt 2004) that this should change in the near future and the Antarctic lakes and surrounding regions will undergo the same change that the Arctic lakes are experiencing.

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