15. PALEOLIMNOLOGY OF EXTREME COLD TERRESTRIAL AND EXTRATERRESTRIAL ENVIRONMENTS

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Introduction

The study of environments that exist near planetary extremes has become a research area of considerable contemporary importance. Our understanding of life's environmental limits has been largely modified as discoveries and evidence of functional ecosystems are reported in places such as the deep ocean, hot springs, deep beneath the Antarctic ice cap, and deep within the Earth. These discoveries have led to a separate field of interdisciplinary study focused on life in extreme environments. A working definition of an extreme environment is one where physical and chemical conditions approach or exceed the perceived tolerances for life (Benchley et al. 1998). Some organisms can thrive under these conditions and not survive in more clement environments. A key impetus for studying extreme environments is to understand the origin and evolution of life itself. Life originated on our planet when a select set of environmental conditions allowed molecules to organize themselves into self-replicating entities. Defining the boundaries of those conditions is essential to our understanding of life's origins on our planet and other extraterrestrial bodies.

Liquid water is essential for all life on Earth, including life that exists in extreme environments. Therefore, the study of inland waters (limnology) is by definition an essential element in the study of life in extreme environments. In temperate regions, extreme aquatic environments can be found mostly as a result of hypersalinity (e.g., salt lakes) and heat (hot springs). In polar regions, extreme aquatic environments are created by cold hypersaline water bodies, thick permanent lake ice covers, and the combined effect of high pressure, aphotic, and oligotrophic conditions in subglacial lakes.

In this chapter, we consider these polar extreme limnological environments, and in particular the historical record preserved by them. We also provide a discussion of the connection between these polar extreme environments and water bodies that may exist elsewhere in our solar system, notably Mars and Europa.

Perennially ice-covered lakes

A number of lakes exist in the polar regions that maintain their ice cover throughout the year. The list in Tables 1 and 2 is not meant to be exhaustive as there are undoubtedly numerous lakes not accounted for here, particularly glacier-contact lakes at high altitudes.

Impact of perennial ice covers

Perennial ice covers have numerous effects on the underlying water column including: (a) reduction of light penetration, (b) reduction of gas exchange between lake water and

Table 1. Documented perennially ice-covered lakes in the Arctic.

Location	Ice (m)	lat.	long.	Selected References
1. Angiussaq Lake, Greenland[1]	2-2.5	77º00'N	66º08'W	Barnes 1960
2. Pond at Rundfjeld, central Ellesmere Island[2]	5.5	78º05'N	81º04'W	Blake 1989
3. Ziegler Island, Franz Joseph Land[3]	< 2.4	81º00'N	56º02'E	Panzenbock et al. 2000
4. Lake A, northern Ellesmere Island[4]	2	82º08'N	78º00'W	Belzile et al. 2001; Gibson et al. 2002
5. Ward Hunt Lake, Ward Hunt Island	4	83º01'N	74º02'W	Villeneuve et al. 2001; R.A. Wharton, pers. comm.

[1]At least one other smaller lake nearby also reported to be perennial
[2]Pond was frozen to bed in winter. ^{14}C date indicates lake has not thawed in > 5000 yr
[3]Single unnamed lake in the centre of the island
[4]Lake C nearby believed perennially ice-covered but not studied

the atmosphere, (c) reduction in water column mixing, and (d) alteration of sedimentation pathways (Wharton et al. 1989; Gibson et al. 2002). Although all of these effects are important in defining the modern ecosystem of the lakes, and therefore the character of sedimentary deposits, the alteration of sediment pathways is likely the most important to paleolimnology. When a perennial ice cover is present, sediments can either be blown across the lakes or get deposited on the surface. In either case, sedimentation within the lake is altered. Sediments that get trapped in the ice will be warmed by the sun and melt into the ice to a depth that is largely related to the sediment grain diameter (Hendy 2000a). The depth to which any sized sediment can sink through this ice is limited, and layers of sediment collect at the dynamic equilibrium between particles melting downward and the ice cover moving upward through ablation and ice growth. This process forms sediment layers and associated lenses or inclusions of liquid water in the ice cover at approximately 2 m beneath the surface (Fritsen et al. 1998; Priscu et al. 1998). Beyond this level in a thick ice cover, sediment has to make its way through cracks in the ice, resulting in heterogeneous ridges and mounds of sediment on the bottom (Figure 1). Additionally, different lakes and different areas of lakes have varied rates of sediment accumulation on the ice, so that this process is not consistent across all lakes (Adams et al. 1998).

Arguably the most studied perennial ice-covered lakes occur in the McMurdo Dry Valleys of Antarctica (77-78ºS, 160-164ºE). These lakes occur along the valley bottoms and are mostly closed basins. Groundwater exchange in the lakes is believed to be minimal, but remains largely uninvestigated. A previous review of the paleolimnology of these lakes is given in Doran et al. (1994a), and they are also treated in Hodgson et al. (this volume). The McMurdo Dry Valleys comprise a cold desert ecosystem, with mean annual temperatures ranging between -14.8 to -30ºC, and annual precipitation (all as snow) usually < 100 mm/yr with as little as 7 mm recorded in a single year (Doran et al. 2002a). Dry valley lake levels and lake ice thicknesses are constantly fluctuating in

Table 2. Documented perennially ice-covered lakes in the Antarctic.

Location	Ice (m)	lat.	long.	Selected References
1. Bunger Hills, various[1]	< 3	66º03'S	100º08'E	Doran et al. 1996, 2000
2. Frames Mountains, MacRobertson Land, various	3.5-4+	67º08'S	62º08'E	Pickard and Adamson 1983; Chambers et al. 1986
3. Vestfold Hills, various	2.5-5+	68º05'S	78º00'E	Pickard and Adamson 1983
4. Larsemann Hills, various[2]	?-2.2-?	69º03'S	76º00'E	Gillieson et al. 1990; D.A. Hodgson, pers. comm.; J. Burgess, pers. comm.
5. Schirmacher Hills, various[3]	1.6- > 3	70º08'S	11º07'E	Bormann and Fritzsche 1995; Gore 1997
6. Beaver Lake, MacRobertson Land	3-4	70º08'S	68º03'E	Laybourn-Parry et al. 2001
7. Ablation Valley[4]	2.5-4.5	70º08'S	68º04'W	Heywood 1977
8. Central Dronning Maud Land, various[5]	2.2-3.8	71º04'S	13º04'E	Bormann and Fritzsche 1995; Loopman et al. 1986; Wand et al. 1997
9. Citadel Bastion Lake, Alexander Island	3.7	72º00'S	68º03'W	D.A. Hodgson, pers. comm.
10. McMurdo Dry Valleys[6]	3-19	77º00'S	163º00'E	Wharton et al. 1989; Chinn 1993; Doran et al. 1994a, 2002b,c
11. Lake Wilson	4	79º08'S	159º06'E	Webster et al. 1996

[1]Only ice contact lakes at edge of oasis (mostly epishelf)
[2]Lake Ferris (LH11) and small lake to the north of Lake Oskar (LH18), L22, L25 (lake names from Gillieson et al. 1990, but ice observations unpublished)
[3]Ice thickness data limited. Perennially ice-covered lakes are all epishelf (at least 6 lakes)
[4]3 proglacial lakes: Ablation, Moutonée, and another unnamed
[5]Lakes Untersee, Obersee and Burevestniksee
[6]Appoximately 20 permanently ice-covered lakes and ponds (Doran et al. 1994)

Figure 1. Sand mound on the bottom of Lake Hoare in the McMurdo Dry Valleys. The leg of a sediment trap can be seen in the top of the image.

Figure 2. Perched deltas on Crescent Stream in the McMurdo Dry Valleys. Person standing on distant delta for scale. Perennially ice-covered Lake Fryxell is in the background.

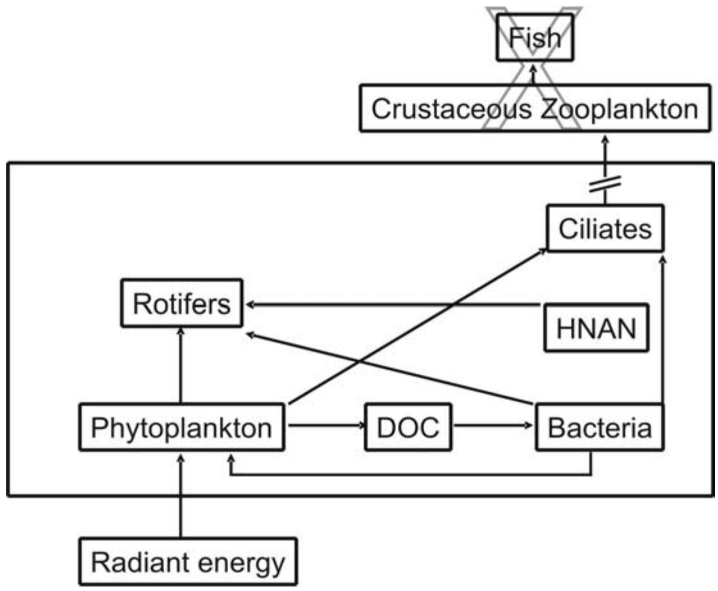

Figure 3. The food web in McMurdo Dry Valley lakes. HNAN is heterotrophic nanoflagellates. DOC is dissolved organic carbon. The X covers organisms not found in this ecosystem.

response to changing climate, primarily summer temperature (Chinn 1993; Fountain et al. 1999; Doran et al. 2002b). The current lakes have been suggested to be remnants of much larger glacial lakes that occupied the valleys during the last glacial maximum (LGM) (Stuiver et al. 1981; Hall and Denton 2000a; Hendy 2000b). Sediments from these lakes stranded on the sides of the valleys (Figure 2) provide useful information on the paleoclimate and the history of the West Antarctic Ice Sheet (Stuiver et al. 1981; Hall and Denton 2000b). However, this sediment record is sporadic, and in Taylor Valley is near 8700 yr BP; the radiocarbon (^{14}C age) of the youngest paleolake deposit found (Hall and Denton 2000a). This suggests that lake-levels have been mostly near or below present levels since that time, and that detailed information about lake history during the Holocene must be retrieved from the lake water or sediments.

The water columns of the McMurdo Dry Valley lakes are dominated by protists (few metazoans have been observed) and the benthos lacks burrowing fauna (Figure 3). There is no sediment bioturbation in the traditional sense of the word. However, because the lakes have elevated levels of dissolved gases, conditions within the shallow zones of the lakes cause microbial mats growing on the bottom to become buoyant and start to float off the bottom as “lift-off” mats (Parker et al. 1982). In areas where this occurs, disruption of the sediment surface can be significant (Figure 4). After rising through the water column, lift-off mats can become frozen in the base of the lake ice and eventually migrate up to the surface of the ice due to ablation from the top and freezing onto the bottom of the floating ice cover. Lift-off of mats located along the shallow water perimeters of the lakes (Wharton et al. 1989) may be an important process in the carbon cycle of these lakes (e.g., Parker et al. 1982). Priscu et al. (1998)

showed that communities seeded by surface aeolian deposition actually live within the ice cover producing organic matter within the ice that eventually reaches the lake bottom. Underwater SCUBA observations have shown that lift-off does not always occur within the shallow "lift-off zone". We have made the further observation that lift-off is frequent in areas where there is significant sediment rainout from the ice cover. It therefore appears that depth and disturbance are requirements for benthic mat lift-off in these lakes.

Figure 4. Photo of microbial mat lifting off of the bottom of Lake Hoare. Picture top to bottom is approximately 0.5 m.

Paleolimnology from sediments in perennially ice-covered lakes

Lyons et al. (1985) analysed an archived core taken from Lake Vanda (77º53'S, 161º55'E) during the Dry Valley Drilling Project (DVDP). Preservation of the core stratigraphy was poor, but they were able to identify three desiccation events represented by peaks in total salt, $CaCO_3$, organic carbon, and biogenic silica.

Lawrence and Hendy (1985) extracted 14 cores from Lake Fryxell (77º61'S, 163º18'E), the longest of which was just over 1 m. Based on two ^{14}C dates of carbonates, they concluded that the age of the core was in excess of 21 Ka BP. However, carbon reservoir results of Doran et al. (1999) indicated that these ages can only be considered as absolute maxima, and the closest date to the surface of the core was at ca. 0.5 m, so that a surface correction cannot be applied. Lawrence and Hendy (1985) used carbonate phase and $\delta^{18}O$ changes to infer past salinity changes. Squyres et al. (1991) used a 1.5 m spaced 3 row by 3 column grid of short (< 40 cm) cores to show the heterogeneity of sedimentation in Lake Hoare (77º63'S, 162º85'E), and tried to define a perennially ice-covered lake facies. Doran et al. (1994b) analysed the $\delta^{13}C$ of buried organic mats and carbonates, $CaCO_3$ and organic matter (OM) content, and diatom species shifts in a ca. 30 cm core from Lake Hoare. They showed that sediment characteristics vary markedly over the short length of this core, but carbonates are consistently calcite, unlike the sediments of Lake Fryxell which contain alternating sequences of calcite and aragonite. Spaulding et al. (1997) were able to show zonation of diatoms in modern Lake Hoare surface sediments, but were unable to use that information to draw any firm paleoenvironmental conclusions from a 30 cm core of the same lake. Doran et al. (1999) used ^{14}C dates in a Lake Hoare sediment core to infer a sediment accumulation rate of 0.015 cm yr^{-1}. Bishop et al. (2001) did a comparative study between a 38 cm sediment core from a deep anoxic zone in Lake Hoare and a 47.5 cm core from a shallower oxic region of the lake. They showed that organic matter $\delta^{13}C$ and $\delta^{15}N$ trends provide a more complex history for the anoxic region sediments, and that biogenic pyrite found in the core from the anoxic zone is associated with depleted $\delta^{34}S$ values and high organic C values (Figure 5).

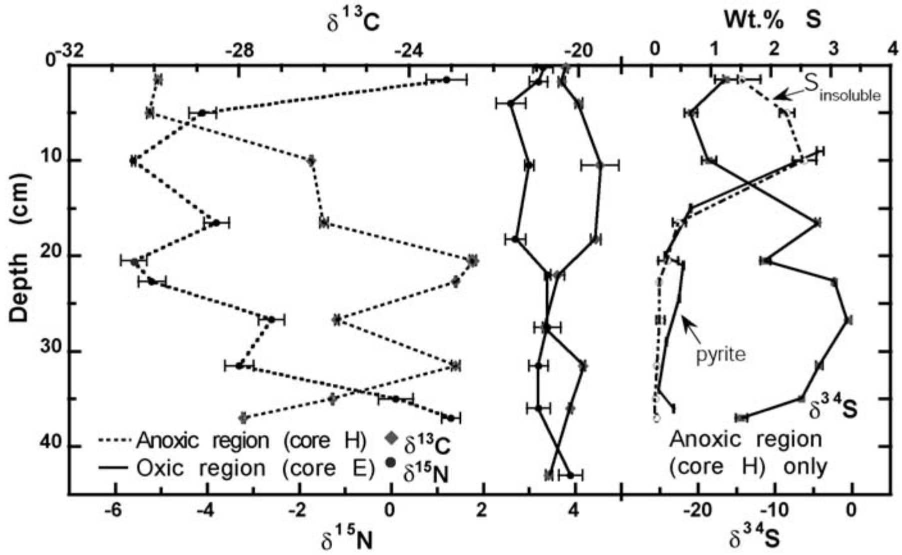

Figure 5. Lake Hoare core profiles for isotopes. $\delta^{13}C$ and $\delta^{15}N$ trends for oxic and anoxic region cores; $\delta^{34}S$ trends for the anoxic core; and percent by weight sulphur (Wt.% S) due to pyrite from Mössbauer Spectroscopy measurements and due to insoluble S (mostly pyrite) from elemental analyzer measurements (from Bishop et al. 2001).

Paleolimnology from perennially ice-covered water columns

Due to the isolation from wind mixing and strong chemical stratification in perennially ice-covered lakes, water chemistry can provide abundant paleolimnological information, a fact that was first recognized and used by Wilson (1964). Many of these lakes have evaporated to near dryness in the past, leading to the concentration of solutes and the precipitation of various salts during times of drier and/or colder conditions. As the climate ameliorated, glacier melt flowed into the lake basins on top of the concentrated monimolimnion, leaving the meromictic conditions that are observed today. This is the case for Lake Vanda, in Wright Valley (Wilson 1964), and Lake Bonney (77º71'S, 162º41'E) (Hendy et al. 1979; Matsubaya et al. 1979) and Lake Fryxell (Lyons et al. 1998a) in Taylor Valley. There is now ample evidence to suggest that these lakes have undergone numerous water-level fluctuations in response to subtle climate changes at least throughout the Holocene (Friedman et al. 1995; Hall et al. 2001), and perhaps as far back as 300 Ka BP (Hendy 2000b).

By using simple diffusion models of conservative solute species such as Cl^- coupled with stable isotope measurements of $\delta^{18}O$ and δD in lake waters, the timing of the beginning of the last fill event (or termination of the last desiccation event) can be determined. These model calculations indicated that for Lake Vanda and Lake Fryxell the filling of the lakes began ca. 1000 years BP (Wilson 1964; Lyons et al. 1998a). This time is generally associated with numerous other warmer climatic proxies all along the Victoria Land coast line (e.g., Lyons et al. 1998a). In addition, Lake Wilson at 80ºS in southern Victoria Land shows a similar timing of last refill, beginning ca. 1000 years ago (Webster et al. 1996). Thus, it appears that the climatic change that occurred at this time covered the entire length of the Victoria Land coast to at least 80ºS. Matsubaya et al. (1979) estimated that the drawdown event in the east lobe of Lake Bonney ended at 2600 years BP. Contrary to the work of Hendy et al. (1977), they suggested that the west lobe of Lake Bonney is only ca. 6000 years old, and represents the dissolution of previously deposited salt with Taylor Glacier melt. Conversely, Lake Hoare, another major lake in Taylor Valley, demonstrates no signs of previous cryo-concentration based on both total dissolved solids and stable isotopic evidence. Lyons et al. (1998a) speculated that Lake Hoare was a relatively new feature, possibly produced as the last refill event began ca. 1000 years ago. As the climate warmed and the Canada Glacier advanced, it separated the Lake Hoare basin from the Lake Fryxell basin to the east (Lyons et al. 2000). Consequently, both ^{36}Cl and $\delta^{37}Cl$ profiles have supported the notion that Lake Hoare is indeed "young" and the chemistry of the lake is due solely to modern glacier melt input (Lyons et al. 1998b, 1999). This modern age also compares very well to sediment analysis suggesting the bottom of Lake Hoare was subaerially exposed prior to ca. 1000 years BP (Doran et al. 1999).

The use of density or chemical stratification of antarctic lakes as paleoclimate indicators, in a more general sense, has been described eloquently by Gibson and Burton (1996). The production of "paleoepilimnia" from changes in water balance in antarctic lakes from season-to-season and even year-to-year are important indicators of climatic variation in antarctic coastal regions from the Vestfold Hills at ca. 68º30'S to Lake Wilson at ca. 80ºS.

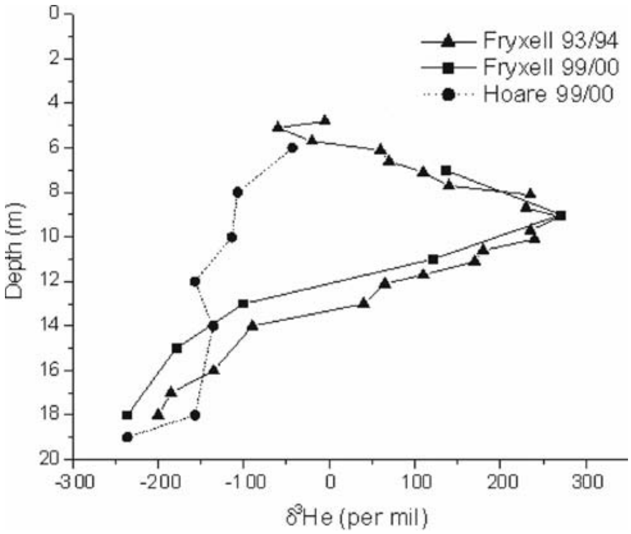

Figure 6. Profiles of δ^3He in Lake Hoare and Lake Fryxell in the McMurdo Dry Valleys.

In some cases, more recent (< 50 yr) lake conditions can be recorded by variations in ^{3}H and helium isotopes. Initial work by Hood et al. (1998) utilized this approach to determine both the short time scale physical dynamics and the longer term water balance variation in Lake Fryxell. Because ^{3}H decays to ^{3}He, ^{3}H/^{3}He ages of lake water can be measured as initially outlined by Torgersen et al. (1977). Hood et al. (1998) determined that the top 7 m of Lake Fryxell water was less than 25 years old during the 1993-1994 field season. More recent data obtained during the 1999-2000 field season estimated the top 5 m of Lake Fryxell to be less than 18 years, and the top 7 meters to be less than 50 years old. Prior to the collection of water by Hood et al. (1998), Lake Fryxell had been rising at a very rapid rate (Chinn 1993). However, in the ensuing six years between these two data sets, the lake level had actually decreased due to a substantial decrease in glacier meltwater inflow (Doran et al. 2002b). This decrease in lake-level actually allows for older, deeper water to come closer to the ice-water interface, thereby explaining our increase in ^{3}H/^{3}He ages compared to the earlier work of Hood et al. (1998).

Helium isotopic measurements are also very useful in understanding past hydrologic events occurring within these lakes. In Figure 6 we have plotted the δ^3He values for

$$\delta^3\text{He} = [(\text{Rm}/\text{Ra})-1] \times 1000 \quad (1)$$

both Lake Fryxell and Lake Hoare where Rm is the measured ^{3}He/^{4}He in the sample and Ra is the value in water in equilibrium with the atmosphere. The Lake Fryxell profile is very similar to what previously had been published by Hood et al. (1998). The change of slope of the profile at ca. 9 m is similar to that of conservative solutes such as Cl^-, and suggests a strong diffusional gradient of ^{4}He from the sediment/water interface, as first suggested by Hood et al. (1998). The surface waters more closely resemble waters in equilibrium with the atmosphere with additional input of tritiogenic ^{3}He.

Perennially ice-sealed lakes

The perennially ice-covered lakes discussed above (i.e., lakes whose permanent ice covers overlie a well established liquid water column) form seasonal moats and have porous ice covers that influence the hydrology, physical structure and biological activity of the underlying liquid water column. However, some antarctic lakes have sufficiently thick ice covers that summer meltwater flows over the permanent ice rather than under it, so that the lake ice grows from the surface up (while still growing at the bottom when the heat flux allows it) and accumulates over time if meltwater overflow exceeds ablation. Previous authors believed these lakes were frozen to their beds and referred to them as ice block lakes (e.g., Chinn 1993). Shallower lakes are likely frozen to their beds, and such a lake (5.45 m deep) has been reported in the Canadian High Arctic (Blake 1989). Using ground-penetrating radar (GPR), at least two of these so-called ice-

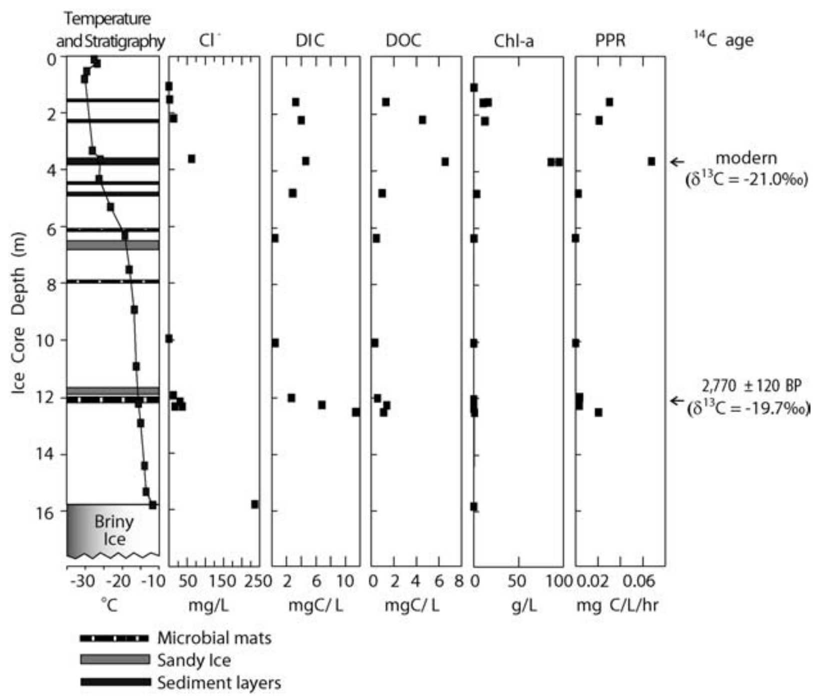

Figure 7. Physical and chemical properties of a Lake Vida ice core extracted in October 1996. Black horizons on the stratigraphy plot represent sediment layers, gray horizons are sandy ice, and vertically banded horizons contain microbial mats. Profiles are shown for chloride (Cl^-), dissolved inorganic carbon (DIC), dissolved organic carbon (DOC), chlorophyll-a (Chl-a), and primary productivity (PPR). The temperature profile plotted was taken at the time of the core extraction (from Doran et al. 2003).

Figure 8. Photograph of ca. 1 m of water trapped on the surface of the 19 m Lake Vida ice cover during the anomalously warm 2001/2002 austral summer. Photo was taken by T. Nylen on 18 January 2002 from the Lake Vida meteorological station at the west end of the lake.

block lakes in the Dry Valleys show evidence of thick ice covers overlying saturated brine that is not allowed to communicate with the atmosphere. Lake House (77º07'S, 161º04'E) in Pearse Valley and Lake Vida (77º38'S, 161º95'E) in Victoria Valley have ice thicknesses of 11 m and 19 m, respectively. The origin, evolution, and maintenance of ice-sealed Lake Vida has been described by Doran et al. (2003).

Because of the way perennially ice-sealed lakes grow over time, it is very likely that clastic sedimentation is essentially halted when the lake becomes hydrologically sealed from inflowing water. From this point until the lake becomes unsealed, we anticipate that the only sedimentation occurring is through precipitation of salts. Growing ice cover will cause salts to be deposited; thinning ice cover will cause salts to be dissolved. Therefore, we do not envision that lake bottom sediments during the sealed mode will be very useful repositories of paleoenvironmental information, with the exception of changes in salt chemistry. However, if cores of sufficient length can be obtained, the timing and extent of the shifts between sealed and unsealed modes should be detectable through parameters as simple as clastic sediment load.

The ice cover in perennially ice-sealed lakes contains important paleoenvironmental information. Doran et al. (2003) identified numerous layers of sediment and microbial mat (Figure 7) throughout the top 16 m of the Lake Vida ice cover (see also Fritsen and Priscu 1998). These layers are believed to represent sediment deposition and associated microbial mat growth during exceptionally warm summers as turbid water collects on top of the ice cover, such as during the 2001/2002 austral summer (Figure 8). Sediment deposited during these warm summers can be trapped in the liquid water layer near the surface of the ice cover in a manner similar to what happens with windblown material in the thinner ice covers. These sediments and associated microbes then become trapped within the ice cover as the liquid water freezes in winter. We believe productivity is associated with this substrate not only during the open water events, but can also occur during colder years when liquid water does not accumulate on the surface. During the latter, sediment particles would absorb solar radiation forming liquid water inclusions, similar to those observed in thin-ice lakes (Fritsen et al. 1998; Priscu et al. 1998). Radiocarbon dating of organic sediment layers in the ice indicates that the lake has been in the sealed mode for at least 2800 ^{14}C yrs BP. Microbial mats throughout the ice cover are viable when thawed in the laboratory (Fritsen and Priscu 1998; Doran et al. 2003).

Subglacial lakes

Water may pond below glaciers on various scales ranging from mm to km in area. Larger accumulations of water can form permanent subglacial lakes: (1) in deep glacier bed depressions, commonly below ice domes, where the depression must be deeper than the surfaces of equipotential contours above it; and (2) if the glacier bed is flat, but where a basin on the ice surface is surrounded by a ring of thicker ice and subglacial water becomes trapped below the basin area due to equipotential gradients (Nye 1976). A further classic mechanism of subglacial lake formation is from high geothermal heat flow being concentrated in one subglacial area causing local basal melting, such as in Iceland today at subglacial volcanic vents (e.g., Björnsson 1975).

Although the unequivocal presence of subglacial lake deposits in the rock record is rare, inferences of their existence under paleo-ice sheets is commonly invoked to explain subglacially produced forms that are thought to have required large volumes of water to be released catastrophically. In Antarctica there are extensive dendritic channel systems that form the Labyrinth in Upper Wright Valley (77º55’S, 160º83’E), which have been suggested to have formed under more extensive outlet glaciers during such a catastrophic release of subglacial lake water (Sugden et al. 1991; Denton et al. 1993). In addition to the fossil lakes inferred to have caused the Labyrinth, there are at least 77 existing subglacial lakes beneath the East Antarctic Ice Sheet (EAIS) (Siegert et al. 1996).

Airborne radar mapping in the 1970s first identified the presence of lakes beneath the EAIS (Siegert et al. 1996); however, more recent mapping has identified at least another 15 lakes in the region of Dome C, some of which may exceed 40 km in length (Tabacco et al. 2002). Lakes in the region of Dome C seem to exist in what may be considered a “Lake District” that may be connected hydraulically. Lakes larger than about 5 km in the longest dimension generally have a surface slope of < 0.5 m km^{-1} indicative of ice floating on liquid making them identifiable from satellite images. However, care must be taken with interpreting every low slope area as a lake until the depth of subglacial water is known (e.g., Tikku et al. 2001). Subglacial lakes remain liquid as a result of the melting point depression caused by the weight of the overlying ice, insulating ice cover, and heat flux from the Earth (Kapitsa et al. 1996; Siegert and Kwok 2000; Siegert et al. 2001). Lake Vostok (78º05'S, 107ºE) (Figure 9), by far the largest of these lakes, is about 280 km long and 50 km wide and > 1000 m deep at its deepest point. The lake lies under 3750 m of ice on its south end and 4150 m on the north end within the East Antarctic Precambrian craton. The difference in ice thickness can produce barotropic flow in the lake (Wüest and Carmack 2000) and differential regions where lake water melts and freezes (accretes) to the bottom of the overlying ice sheet (Siegert et al. 2001).

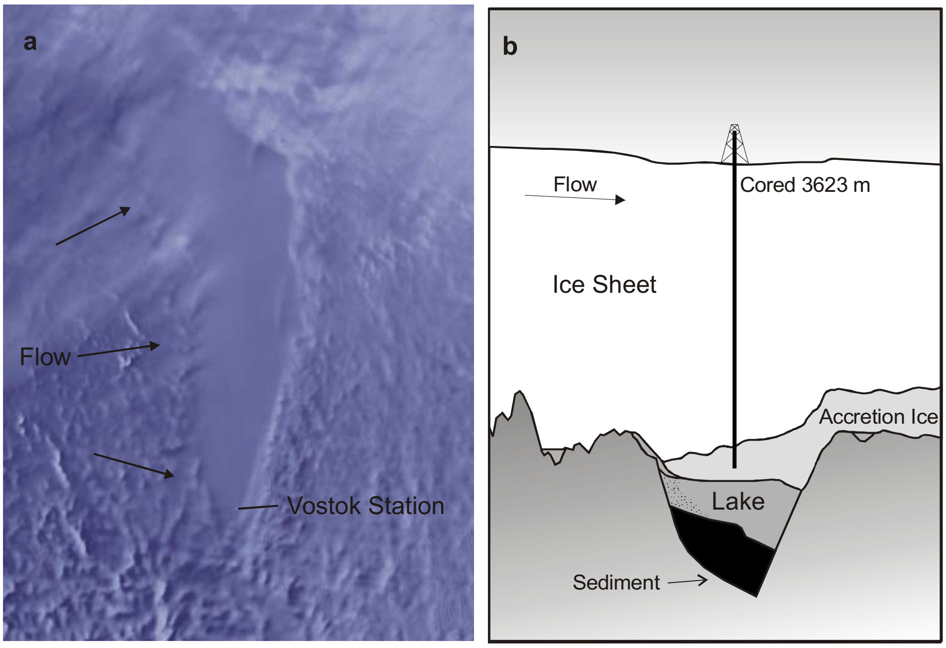

Figure 9. (a) RADARSAT image of the glacier ice surface in the region of subglacial Lake Vostok in eastern Antarctica. The flat floating ice surface can clearly be seen which outlines the lake area ca. 3.5 km below. The lake is 280 km long and 50-60 km wide. Arrows mark general direction of glacial flow at the surface (from Siegert et al. 2001). (b) cross-section of the ice sheet at Lake Vostok showing the general configuration of glacier ice, accretion ice, bedrock, and lake. The ice melts as it flows into the lake from the left, and refreezes forming the accretion ice to the right. The depth of the water column beneath the Vostok borehole is 670 m and average thickness of sediments is thought to be ca. 300 m.

Tectonic setting of East Antarctic subglacial lakes

Most of the EAIS subglacial lakes appear to be deep basins associated with ice divides. However, major questions remain as to the geological origin of the deep basins beneath regions near the ice divides. The most likely geological alternatives are (e.g., Dalziel

1998): (1) intracratonic rifts associated with extensional processes, (2) rifts from continental collision, (3) hot spot or mantle plume-driven depressions, (4) glacial scour eroding older feature, or (5) meteor impacts. Lake Vostok lies within the Vostok subglacial highlands, which are connected to the Gamburtsev Subglacial Mountains; the origin of both the mountains and lake is currently uncertain. The most recent interpretation from new airborne geophysical surveys is one of a reactivated suture zone where a thrust block overlies an old continental margin (Studinger et al. 2001). The lake is a younger, smaller extensional basin set in a sedimentary succession overlying the thrust block, thought to have been created during a later, weak reactivation phase in the tectonics. Based on helium (He) isotope ratios in accretion ice, present-day tectonic activity within the rift basin seems unlikely (Jean-Baptiste et al. 2001), though recent seismic activity in the area has been recorded (R. Bell, pers. comm., 2001). It should be noted that the He isotope evidence is from accreted ice, which reflects only the upper portion of the lake. Deeper waters may have a different He isotope signature.

Inferred lake processes

Physical processes

Some inferences about the total physical character of Lake Vostok have been made from geophysical remotely sensed data, refrozen lake water recovered from ice cores and from modeling. However comprehensive process modeling may only be done after the documentation of the physical characteristics of the lake upon lake entry.

Freezing is thought to be occurring across the entire southern section of the lake and the accreted ice is being exported from the lake at the base of the ice sheet (Bell et al. 2001). Export of accreted ice is used to indicate that fresh water must be replenishing the lake from an upstream subglacial water catchment. The residence time of the water in the lake has been estimated to range from 5000 to 125,000 years (Kapitsa et al. 1996; Jean-Baptiste et al. 2001). Mayer and Siegert (2000) have suggested a residence time of about 100,000 years based on estimates of annual meltwater mixing within Lake Vostok. The lowest value was based on helium isotopes measured in lake water accreted onto the bottom of the ice sheet (Jean-Baptiste et al. 2001). However, the ice measured in this study may have accreted in a shallow embayment isolated from the main lake basin (Bell et al. 2001) and probably underestimates the actual residence time. Isotope records, crystal growth rates and ice flow modeling suggest that the basal glacier ice overlying Lake Vostok could be as old as 1,000,000 years (Siegert et al. 2001), marking the maximum possible age of the youngest lake water.

The ice above Lake Vostok has been cored to a record depth of 3623 m, stopping ca. 120 m above the surface of the lake. The upper 3500 m of glacial ice represents a 420,000 year environmental record covering four complete ice age climate cycles; ice below 3500 m is thought to represent refrozen lake water accreted to the bottom of the glacial ice (Jouzel et al. 1999; Petit et al. 1999; Priscu et al. 1999; Siegert et al. 2000).

Chemical and biological processes

Abyzov et al. (1998) provided evidence that a wide range of microbes (bacteria, yeasts, fungi and microalgae) exist in the Vostok glacial ice at depths ranging from 1500 to

2750 m. The concentration of microbes, some viable, was correlated with the density of mineral particles in the ice, implying that they were deposited in the snow mainly during glacial periods when the flux of dust and the wind speed were greatest. Recent studies of accretion ice from 3590 m (Priscu et al. 1999) and 3603 m (Karl et al. 1999) have shown the presence of microbes at densities of 10^2 to 10^4 cells ml^{-1}, and the development of metabolic activity in the presence of liquid water. Bacterial 16S rDNA revealed low diversity in the gene population (Priscu et al. 1999; Christner et al. 2001). The phylotypes were closely related to extant members of the alpha- and beta-Proteobacteria and the Actinomycetes. Actinomycetes were also observed in overlying glacial ice (Abyzov et al. 1998) implying that the biological seed for the accretion ice (and presumably for Lake Vostok) may have arisen from airborne particulates (e.g., Marshall 1996) deposited on the surface of the ice sheet ca. 500,000 years BP. The microbes (Figure 10) were released into the lake and subsequently refrozen to the overlying glacial ice following downward migration and melt at the glacial grounding point. Alternatively, the microbes could be remnants from an ancient Lake Vostok that existed before permanent glacial ice cover (ca. 15 million years ago). Importantly, the microbes identified within the accreted ice represent assemblages that exist near the ice/water interface and may have recently been released from the overlying glacial ice. Presumably, a completely different assemblage exists in the deeper waters and in association with the sediments, particularly if the water column is vertically stratified (Wüest and Carmack 2000; Siegert et al. 2001). Because the Lake Vostok ecosystem receives no solar radiation, all biochemical activity within the lake must depend upon chemically mediated oxidation-reduction reactions. The supply of oxidants and reductants are presumably derived from the overlying ice sheet and may result from geothermal activity, although the latter has yet to be confirmed (Jean-Baptiste et al. 2001). The geochemistry of Lake Vostok was estimated from the overlying accretion ice by Priscu et al. (1999) and Souchez et al. (2000). These estimates indicate that the waters of Lake Vostok should contain adequate quantities of dissolved organic carbon, anions and cations to provide a system that is thermodynamically favourable to support life.

Lake sediment record

Although planning is underway, Lake Vostok has yet to be penetrated. Many uncorroborated ideas exist concerning the history of Lake Vostok. Perhaps the best evidence of the paleolimnology of Lake Vostok can be found in the sediments. Russian seismic data have been used to estimate that the sediment succession in the lake is approximately 300 m thick near Vostok Station (Popkov et al. 1998). The lake sediments could contain an unparalleled record of antarctic paleoenvironmental information, extending well beyond the limit of ice core records. In fact, meltwater and sediment currently entering the lake comes from the base of the ice sheet so, in essence, the sediment record should pick up where the ice cores leave off.

To understand the possibilities of what the lake may contain, a brief evaluation of what is currently known about the ice sheet is useful. Glaciological models of the development of the antarctic ice sheet most commonly have ice building on highlands and progressively expanding into larger ice masses until a full continental-

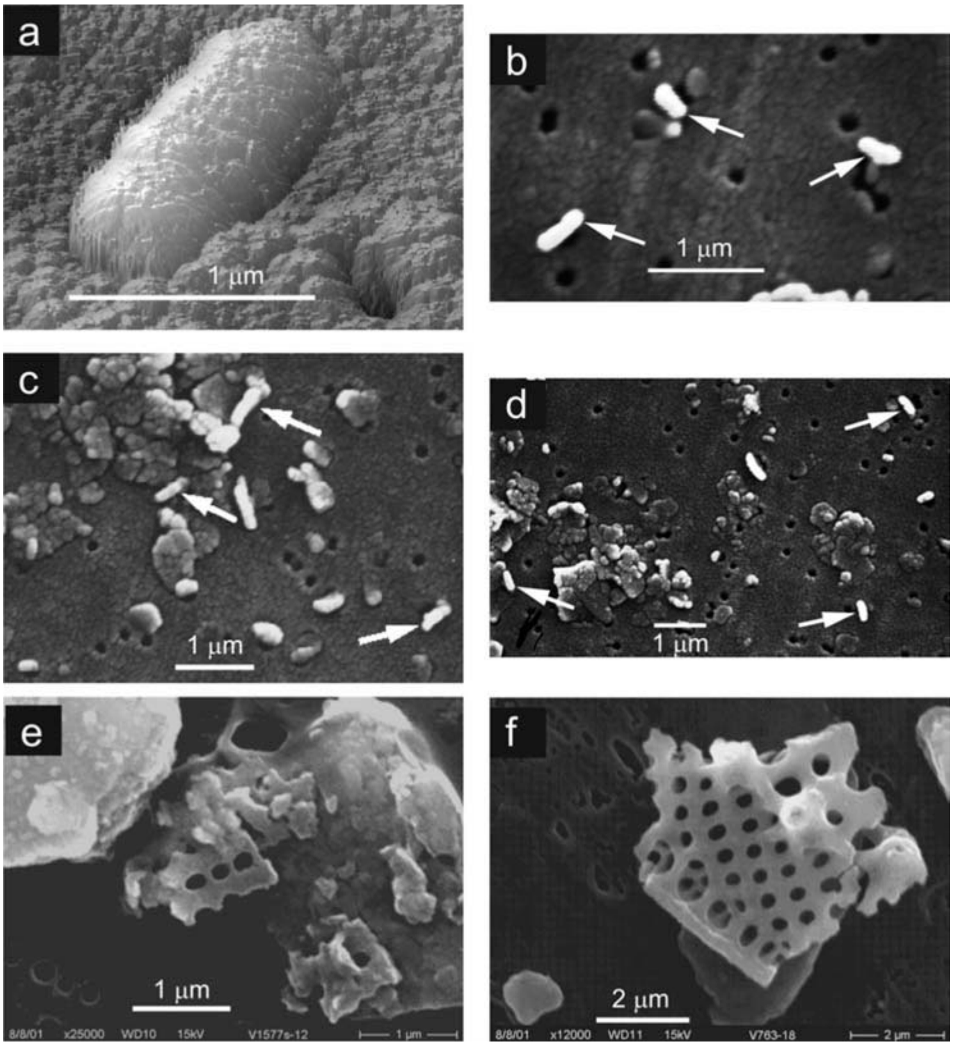

Figure 10. Atomic Force micrograph (a) and scanning electron microscope images (b-f) of particulate matter from accretion ice located 3590 metres below surface (mbs) (a-d) and from the overlying glacial ice (e = 1577 mbs; f = 763 mbs). Accretion ice images show bacterial cells (arrows) associated with organic debris (a-d); glacial ice images show diatom fragments and associated debris (e, f).

encompassing ice sheet forms (e.g., Huybrechts 1993; Budd et al. 1994; DeConto and Pollard 2001). The glaciological models are consistent with distal proxy records of oxygen isotopes (Zachos et al. 2001) and Mg/Ca ratios (Lear et al. 2001) derived from deep sea foraminifera. These proxy records of inferred ocean temperatures and ice volumes indicate that, prior to about 38 Myr ago, glaciers were either on mountains or highlands, or that small intermittent ice sheets grew on the continent. Antarctica then entered a phase of larger, more permanent ice sheets, although their volumes still changed temporally and they probably reached their maximum size in Pleistocene times. These scenarios are also in agreement with those inferred from direct glacial deposits in long geological drill cores through glacio-marine successions in Prydz Bay (Hambrey et al. 1991; O'Brien et al. 2000) and McMurdo Sound (Hambrey et al. 1989; Powell et al. 2000, 2002).

Coastal records indicate that early phases of glaciation were relatively 'warm and wet', with temperate glaciers (Powell et al. 2002). The drill core records also indicate that glaciation on Antarctica became progressively colder through Oligocene to Pleistocene times going through subpolar/polythermal to full polar glaciers (Powell et al. 2002). However, even the younger periods of glaciation appear to have had a range of glacial activity, especially at ice sheet margins, which may have influenced interior ice flow (cf., Powell 1981; Hambrey et al. 1989; Webb et al. 1996; Wilson et al. 1998; Hambrey and McKelvey 2000; Miller and Mabin 2000). During periods of purported temperate ice (e.g., Hambrey and McKelvey 2000), glacial dynamics would have been altered and the interior ice sheet drawn down, dramatically changing drainage divides in some of the areas of the current subglacial lakes. It has also been shown recently that the Miocene ice sheets were being driven by Milankovich forcing in a similar way to Pleistocene glaciers (Naish et al. 2001), adding evidence to the likelihood of changes in mass balance and dynamics through time.

The sedimentary record contained in subglacial lakes may contain evidence that can be used to decipher early basin histories. If the basins formed before the formation of the ice sheet, then their sediment fill may contain continental interior syntectonic sediments which may help define the timing and style of tectonism by lake and/or volcanic deposits. However, given the type of history just described, it is likely that glaciers on interior highlands early in their history were temperate and very dynamic; the type of glaciers that flow down valleys and erode low areas, such as lake basins. The Lake Vostok basin, if it existed at the time, may well have been eroded over several to tens of millions of years of dynamic glaciation. Consequently, the likelihood of having a pre-glacial record in the lake basins is probably quite low, unless the sediments were somehow protected. One mechanism for sediment protection may be the formation of a perennially ice-sealed lake prior to glacial over-ride. As the local snowline descended to near lake-level, an ice-sealed lake could have formed providing the protection of a thick lake ice cover. Whether such conditions could have existed during the temperate transitional phase from the prior "greenhouse" period is unknown.

There may well be records of important ice sheet and climatic fluctuations through Lake Vostok's history retained in the sediment record. These records would be of the interior of the ice sheet rather than of its edges, as have been recovered thus far from other sources. A complication of inferring paleoclimate from Lake Vostok sediments is that any atmospheric input to the lake is delayed by hundreds of thousands of years, and possibly up to 1,000,000 years which is the expected maximum age of basal ice (Siegert et al. 2001). Furthermore, the length of the delay will vary with internal ice sheet dynamics. Sedimentation will be a mix of atmospherically deposited, glacier-scoured, and authigenic sediments, all from processes on very different time scales. Separating out the signals will be difficult, but certainly performing particle-size specific analysis will play a key role as we would expect the atmospheric signal to be carried into the lake through gases and fine-grained particles. It is tantalizing to contemplate what those 300 m of sediment in Lake Vostok basin could contain, but will remain speculation until it is recovered.

In addition to the climatic and geological history, the record contained in the Vostok basin sediments may also reveal information on past geochemical processes, macrobiological and microbiological communities and paleoclimate. The sediments could also contain a record of extra-terrestrial material capture. Micrometeorites within Vostok accretion ice have been observed (Priscu and Mogk, unpublished data) making it likely that the sediments contain a large number of meteorites, micrometeorites and cosmic dust (e.g., interplanetary dust particles and cometary debris). Hence, sediments in the lake basin should offer an extraordinary opportunity to measure extra-terrestrial flux over millions of years. The previously unrecognized 100,000 year periodicity in the Earth's orbital inclination has been suggested to influence the accretion rate of extra-terrestrial material. Assuming that the Earth's orbital inclination is related to the climate record, measurements of extra-terrestrial material in the Vostok sediments could provide an unparalleled record of climate change that could only be provided in this unique setting.

It is interesting to speculate about the accumulation of clathrates (air hydrates) in the sediments of Lake Vostok. Clathrates are preserved in the glacier ice entering the lake and should be stable as the ice passes the pressure melting point. Air hydrate has a equilibrium density of between 0.980 to 1.025 g cm^{-3} (Uchida and Hondoh 2000) and Lake Vostok is estimated to have an average water density of 1.016 ± 0.001 g cm^{-3} (Wüest and Carmack 2000). Therefore we might expect some hydrates to float and some to sink. Lipenkov and Istomin (2001) point out that the proposed circulation in Lake Vostok would be enough to keep hydrates with a density of 1.050 g cm^{-3} and up to 200 μm in diameter in suspension in the water column. Nevertheless, the fact that it is possible to have air hydrate heavier than the surrounding water holds out the potential to have some fraction of air hydrate accumulate in the sediments and provide a sedimentary record of atmospheric gas. We would also expect natural gas hydrates to accumulate *in situ* if conditions are appropriate.

The current location of the Vostok borehole (the hole left after extraction of the Vostok ice core) is not ideal for limnological and paleolimnological studies. The borehole is very close to the southeastern edge of the lake. If sampling is restricted to one hole, it should be over the deepest part of the lake, as is standard practice in paleolimnology. The cross-section in Figure 9 is based on recent sediment profiling and suggests that sediment redistribution (slumping) has occurred over time. This information can also be used to choose the best sampling location with respect to sediments. The closer the sampling is to the west side of the lake, the more of a subglacial melt/scour record will be obtained, whereas moving towards the east may provide more of a "whole lake" signal.

Exopaleolimnology

Most available data implies that water in its liquid form is currently rare in our solar system, with the only confirmed reserves residing on Earth. There are indications that liquid water may be present beneath thick ice covers of the Jovian moon Europa (Kargel et al. 2000; Kivelson et al. 2000) and there is considerable evidence that Mars may have had liquid water at or near the surface throughout its climatic history (Malin and Edgett 2000a,b; Baker 2001). Operationally, the search for life beyond Earth can be regarded as a search for liquid water, since all life as we know it depends on liquid water. It would be of enormous interest to both the exobiological and paleolimnological communities to retrieve samples from old lacustrine deposits on Mars. The sedimentary environments of ancient fluvial features on Mars may provide the best opportunity for discovering the first evidence of life beyond Earth. An earlier review on the topic of paleolakes on Mars was provided by Wharton et al. (1995).

Evidence of fluvial features on Mars

The Mariner 9 orbiter provided the first indications that the Martian surface had been modified by water, unveiling what appeared to be fluvial features (Figure 11a). Perhaps with these initial observations of aquatic environments on another planet, the discipline of exopaleolimnology was born. The case for water on Mars, past and present, has been building since then through a series of orbiters and landers. Under the present climate regime, liquid water is not stable at the surface because of the low atmospheric pressure and cold temperatures. Features attributable to periglacial processes suggest that water may now be hidden beneath the surface as deposits of massive ground ice. The observation that fluvial features on Mars were localized, possibly restricted to regions of geothermal activity, and the difficulty of constructing self-consistent CO_2 greenhouse models for Mars has led to the theory that early Mars, although comparatively warmer and wetter than today, was quite cold and that the fluvial features formed in association with a cold climate regime. There are two major classes of fluvial features on Mars: valley networks and outflow channels.

Valley networks

Valley networks (Figure 11b) often resemble terrestrial runoff channels. Estimates by Baker et al. (1992) place the formation age of most valley systems at 3.8-3.9 Gyr. There are a few examples of younger valley networks residing in what may be Hesperian or Amazonian units (for definition of Martian epochs see Doran et al. in press). The most notable are located on the flanks of the volcano Alba Patera which provide important paleoclimatic clues to more recent fluvial activity (Gulick and Baker 1989). Most common are valleys that are short, with steep gullies on the slopes of large craters in the ancient highlands. The valley walls are usually steep talus slopes and the gullies often terminate abruptly at the flat crater floors. Another characteristic common to most valley networks is that tributaries often have blunt, theatre-headed terminations. The formation of the valleys has generated a great deal of debate with most arguments focusing on the role of fluvial erosion versus other processes such as liquid CO_2, ice,

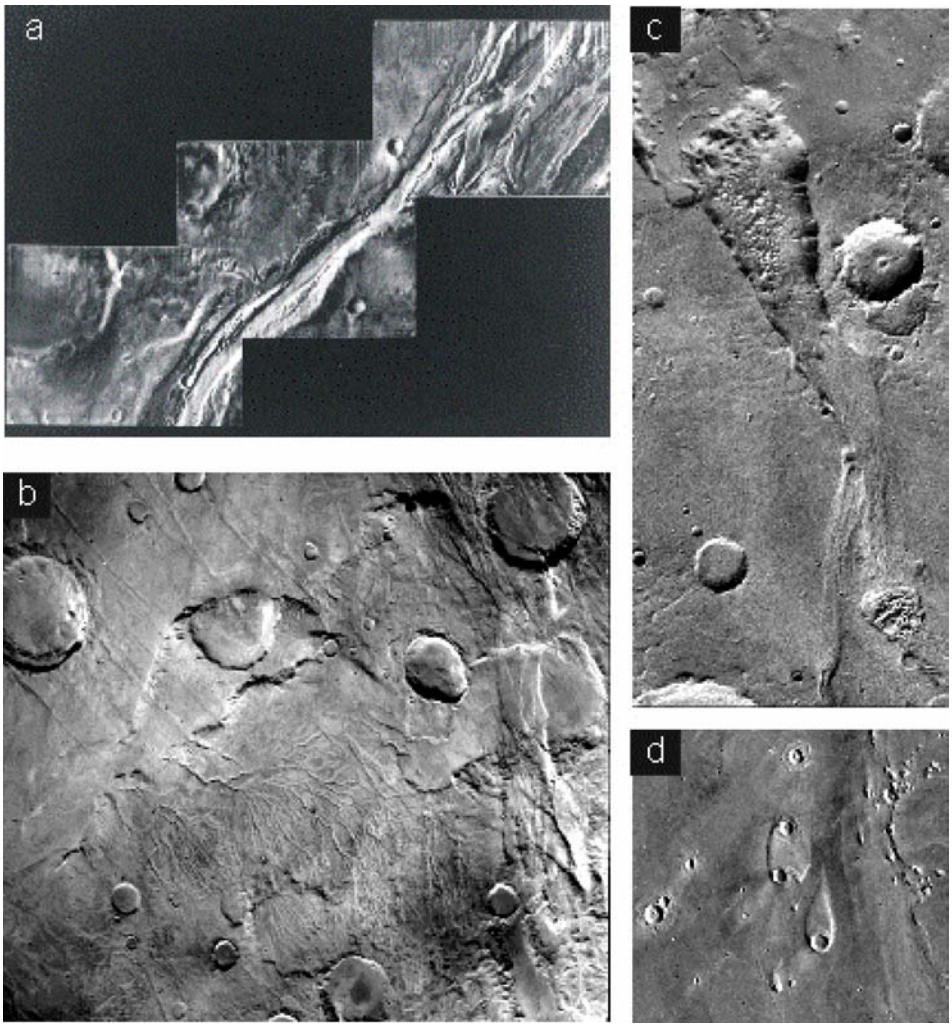

Figure 11. Mars images showing evidence of water activity. (a) Mariner 9 mosaic of about 1/3 (ca. 120 km) of the Amazonis-Memnonia channel centred at 7ºS, 151ºW. This mosaic (Mariner 9 revolution 458, pictures 12499650, 12499650, 12499790) was among the first published evidence of past water on Mars (Sagan et al. 1973). (b-d) Viking images of (b) Dense drainage network in the Southern Highlands. The image is about 250 km across and centred at 48ºS, 98ºW (Viking image 63A09). (c) Outflow channel emerging from chaotic terrain. The image is about 140 km across and centred at 1ºS, 43ºW (Viking image P-16983). (d) "Islands" near Chryse Planitia. Image is centred at 21ºN, 31ºW (Viking image 211-4987).

lava, the role of groundwater sapping versus surface runoff, and climatological constraints. Nanedi Vallis, an 800 km long valley with only a few short tributaries, is a good example of a valley that appears to have been cut by slow erosion of running water. Its sinuosity suggests it was not a flood channel but the lack of tributaries implies that it was fed largely by groundwater rather than runoff. Aharonson et al. (2002) suggest that this feature and many other networks and outflow channels on Mars were formed by groundwater sapping.

Outflow Channels

Outflow channels are enormous features generally attributed to catastrophic flooding initiating from subsurface sources (Baker 2001). They can be over 150 km wide and over 2000 km long. While their initiation points are fairly distinct, they often just fade into obscurity at their downstream ends. Source regions are associated with what has been termed “chaotic terrain” (Figure 11c), a complex topography formed as the result of the removal of subsurface material and widespread collapse of topography.

Typical bedforms of outflow channels include longitudinal grooves, teardrop-shaped islands (Figure 11d) and horseshoe-shaped escarpments. Outflow channels on Mars have morphological features very similar to those of the Channeled Scablands of eastern Washington State in the U.S (Baker 2001). On Earth, approximately 18,000 years ago, Lake Missoula (47º05’N, 114º05’W), a large ice-dammed lake ($2x10^{12}$ m^3 and up to 600 m in depth), burst through its dam releasing a tremendous volume of water (up to 120 m deep) down the regional slopes of the Columbia Plateau. Peak drainage is estimated to have been ca. 10^7 m^3 s^{-1} and complete drainage occurred within a few days leaving the surface heavily scarred with large channels and other morphologic features similar to those found associated with the outflow channels on Mars. The episodic catastrophic floods on Mars probably had discharge rates that may have ranged as much as 100 times higher than the largest flood events that occurred on Earth. According to Carr (1996) most of these events seem to have been caused by the sudden release of groundwater under high artesian pressure trapped below thick permafrost. Several plausible mechanisms were proposed by Carr (1996) to account for the sudden discharge of water to the surface including volcanic activity, faulting and meteoritic impacts. The collapsed remains of the surface at the initiation site, forming the characteristic chaotic terrain, are testament to the magnitude of these voluminous discharge events. Because of their enormous discharge volumes and concomitant latent heat energy, they could, even under current climatic conditions, have flown for vast distances before being halted after boiling off to form ice crystals and water vapour. For this reason, they do not provide a great deal of useful climatic information. Nevertheless, by estimating the volume of water necessary for producing the erosional features that are observed, a lower limit may be placed on the total inventory of water on the planet. Carr (1996) reports that this lower limit is equivalent to 40 m of water spread over the whole planet. These megafloods, which most certainly would have had an impact on the martian climate, occurred mainly in the Hesparian and Hesparian/Amazonian and possibly even in recent (10 Myr) history (Burr et al. 2002).

Springs

Emergent groundwater and perhaps glaciers (Lucchitta 1982) appear to have played an important role in shaping the landscape throughout Martian history. Recent springs may have fed streams that formed the valley networks. Images obtained by the Mars Global Surveyor (MGS) spacecraft of small-scale debris flow gullies suggest spring activity during the last several million years (Malin and Edgett 2000a).

Hydrothermal convection driven by magmatic activity or impact melt may have provided a mechanism of replenishing water in a Martian aquifer. It is likely that springs have formed early in Mars' history as a result of volcanism and meteoritic impacts. Dissolved salts are likely to be present and may enhance the persistence of liquid water environments by depressing the freezing point. On Earth, highly mineralized brines are found in subarctic Canadian Shield wells and in high latitude springs of the Canadian High Arctic (Pollard et al. 1999; Andersen et al. 2002). Brines flowing onto the surface form large icings and deposit salts. The Thermal Emission Spectrometer (TES) instrument on the Mars Global Surveyor (MGS) mission has discovered an accumulation of crystalline hematite in a sedimentary rock formation that covers an area approximately 350 by 350-750 km in the Sinus Meridiani region. One possibility is that the hematite formed by precipitation from aqueous fluids under either ambient or hydrothermal conditions. The deposits, which may have been exposed by wind, provide additional evidence for long-term stability of near-surface water on early Mars (Christiansen et al. 2001). Given the nature of the site, it would not be unexpected to find other mineral precipitates as well. Sites such as this will most assuredly be targeted for exploration as part of the search for evidence of past life.

Evidence for standing water

On Mars, three main types of lakes have been identified – those that formed in areas of convergent drainage by valley networks in the heavily cratered uplands; those that formed within the canyons; and those that formed at the terminus of large outflow events, most notably where the circum-Chryse and Elysium outflow channels terminate (Carr 1996). These putative lakes appear to have formed by a number of mechanisms including meteoritic impact, groundwater seepage and catastrophic flooding. Lakes associated with the valley networks in the ancient, heavily cratered uplands, may contain sedimentary records of the events that led to the evolution of life on Mars since they date to 3.8-4.0 Gyr BP. McKay and Stoker (1989) point out that, since the earliest record of life on Earth has been obscured by erosion and plate tectonics, the best repository of information about the origins of life may actually reside on Mars. The Valles Marineris canyon system may have been flooded with water throughout much of Mars' history (Carr 1996). Box canyons such as Hebes Chasma appear to have thick sequences of fine grained sedimentary material with no visible source regions. One explanation for this is that the materials are carbonates that were deposited in standing bodies of water (Nedell et al. 1987; McKay and Nedell 1988). McKay and Davis (1991) attempted to calculate the time these lakes would exist using a climate model developed by Pollack et al. (1987). They concluded that, as long as a source of meltwater entered into the lake to offset ablation of ice at the surface, the lakes could have exist for several hundred million years or more. The same authors employed calculations that explain the

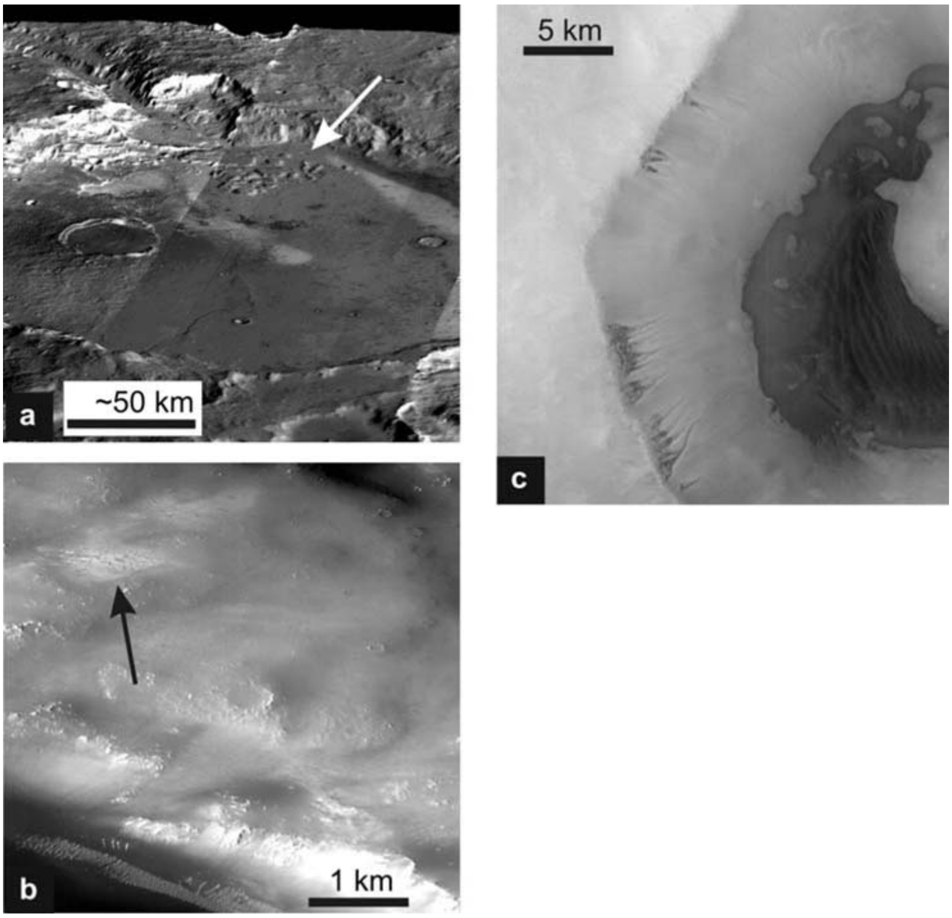

Figure 12. Lakes on Mars? (a) Mosaic of daytime thermal infrared images (obtained by Mars Odyssey) draped over altimetry data (obtained by Mars Global Surveyor). Whiter shades are warmer temperatures and are sun-lighting effects. An ancient river following the channel in the top left is purported to have flown into a crater lake. The arrow points to what are believed to be deltaic deposits left where the river collapsed at lake level (cf., Figure 2). This image mosaic covers an area approximately 180 kilometres on each side centred near 14ºS, 175ºE, looking toward the south. Image credit: NASA/JPL/Arizona State University, PIA04260. (b) Image obtained during Mars Global Surveyor (MGS) aerobraking manoeuvres on October 18, 1997 at 15:42 PST, centred at 5º05'S, 340º07'W shows a feature hypothesized to be a playa deposit (beyond the black arrow). (c) Image taken December 29, 1997 by MGS at 1:19 p.m. PST, centred at 65º01'S, 15º01'W showing evidence of groundwater seepage from a crater rim. It is unclear what the darker deposit at the base of the crater is, but it may have been an area of pooling and evaporite formation in association with the seeps. Images in (b) and (c) provided by Malin Space Science Systems/NASA.

physics that allows similar perennially ice-covered lakes to flourish with microbial communities in the Antarctic. If water was episodically present on Mars, it may have been possible that life survived to more recent times. Newsom et al. (1996), using a similar approach to McKay and Davis (1991) and taking into consideration the effects of the impact melt, geothermal energy released from the uplifted impact basement and the latent heat of freezing, found that the formation of large ice-covered, impact crater lakes > 65km in diameter on Mars would persist for thousands of years even under contemporary climatic conditions. The estimate of the lifespan of lakes on Mars would be extended if the lakes shift into a perennially ice-sealed mode.

Cabrol et al. (1998, 1999) have argued that impact craters such as Gusev (Figure 12a) and Gale may have provided an oasis for life from the Noachian-Hesparian boundary (2.5-3.8 Gyr BP) and lasted for up to 2 Gyr. Using high-resolution Viking images to locate possible lacustrine features, they propose that Gale experienced a number of aqueous environments that transitioned from earlier warmer and wetter, to cold and ice-covered waters. In this case, the sedimentary record in these environments would be important to investigate for evidence of past life. Cabrol et al. (1999) identified 179 impact crater lakes using Viking images and attempted to document their distribution, types and ages. The main implication of the study is that the 179 lakes observed most likely represent a fraction of the total number of lakes in impact craters. With new high resolution images being returned by MGS, the case for past standing water bodies on Mars is building (e.g., Figures 12b and 12c).

Although not definitively under the realm of paleolimnology, we should point out the related evidence of past oceans on the surface of Mars. Parker et al. (1989, 1993) identified two contacts near the southern boundary of the northern plains and interpreted them to be shoreline features of a previous polar ocean. Baker et al. (1991) proposed that the oceans formed as a result of volcanism which triggered catastrophic flooding releasing water into the topographically lower northern plains. Head et al. (1999) have tested the original hypothesis of Parker et al. (1989) by using high-resolution altimeter data from the Mars Orbital Laser Altimeter (MOLA). They have reported their measurements to be in agreement with the observed shorelines made previously by Parker et al. (1989). Additionally, they report that their measurements revealed two major basins within the plains and that previously mapped features associated with ground ice (polygonal cracking and lobate ejecta craters) show a high degree of correlation with these basins, indicating that water may have been present at these locations in previous times. While the MOLA data seem to support the ocean theory, Malin and Edgett (1999) reported they could not find images in support of shoreline features using the MGS camera at high resolution. It may be that the processes that normally form shorelines on Earth were not operating in a similar fashion on Mars. The lack of a large moon results in solar tides that are much weaker. If the ocean was ice-covered, it is not clear how definitive the shoreline features would be.

Summary

In this chapter we have attempted to review the present state of knowledge concerning limnology and paleolimnology in extremely cold environments both on Earth and beyond. The main impact of colder temperatures is to reduce the liquid water volume in that ecosystem. For instance, we can follow the progression of a closed-basin lake as temperature decreases. The first phase of lake evolution with colder temperatures will be to have occasional summers where the ice cover remains on the lake. If the temperature continues to drop, a perennial ice cover with a hydrologically connected seasonal moat will form (i.e., there is communication in the summer between the surface stream flow and under ice water column). Further cooling will cause the ice cover to become sufficiently thick that summer melting can not create a hydrologically connected seasonal moat, so that all summer melt flows onto the surface of the ice cover, and accumulates in layers (i.e., there is now ice growth at the top and bottom of the permanent ice). When the lake shifts into this perennially ice-sealed mode, summer meltwater input is proportional to ice growth because all meltwater flowing onto the surface freezes. At any time in this evolution, if long-term summer meltwater input does not exceed annual ablation from the lake, the lake will diminish and eventually dry up.

If we follow the resulting sedimentation through the above scenario, we will reduce clastic sedimentation in the lake over time, and eventually eliminate it. Increasing the thickness of a perennial ice cover decreases the ability of wind-blown sediments to reach the water column. When the lake shifts into the perennially ice-sealed mode, there may initially be some sediment trapped in the basal ice, but over time the sediment will be purged as the ice melts and refreezes due to the small changes in thermal gradient that occur. Eventually, all significant amounts of sediment will be exhausted and clastic sedimentation will cease. At this point, we could only expect to record chemical and biological sedimentation. So the best record of this type of shift in lake conditions may simply be the amount and character of clastic sedimentation over time. Certainly salt chemistry and thermodynamics become increasingly important to paleolimnologists as a lake passes through these stages.

We have proposed that severe climatic deteriorations could shift a perennially ice-covered lake into a perennially ice-sealed lake, and eventually the lakes would disappear if ablation exceeded meltwater input. Evidence of this sequence of events may be in the geological record in relation to severe glacial periods even in temperate regions (e.g., Hoffman et al. 1998). The formation of perennially ice-sealed lakes in periglacial regions may also provide a mechanism to protect pre-glacial lake sediments as glaciers advance. This may have been the case for a pre-glacial Lake Vostok.

These extreme aquatic environments also provide interesting analogs for conditions on Mars in the distant and possibly even recent past. There is strong evidence of a lacustrine history on Mars, but we can not be certain of how warm the planet was in the past. We can, however, be reasonably certain that the planet was warm enough to have perennially ice-covered or ice-sealed lakes, as we have shown these to be cold end-member lakes. Hence, the study of the paleolimnology of these cold extreme environments may provide guidance of where to look and what to look for during future life detection missions to Mars (e.g., Doran et al. 1998). A similar argument can be made for the study of subglacial lakes like Lake Vostok, which may be analogs for subglacial lakes on Mars and the proposed global ice-covered ocean of Europa.

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