PERENNIAL SPRINGS IN THE CANADIAN HIGH ARCTIC: ANALOGUES OF MARTIAN HYDROTHERMAL SYSTEMS

by

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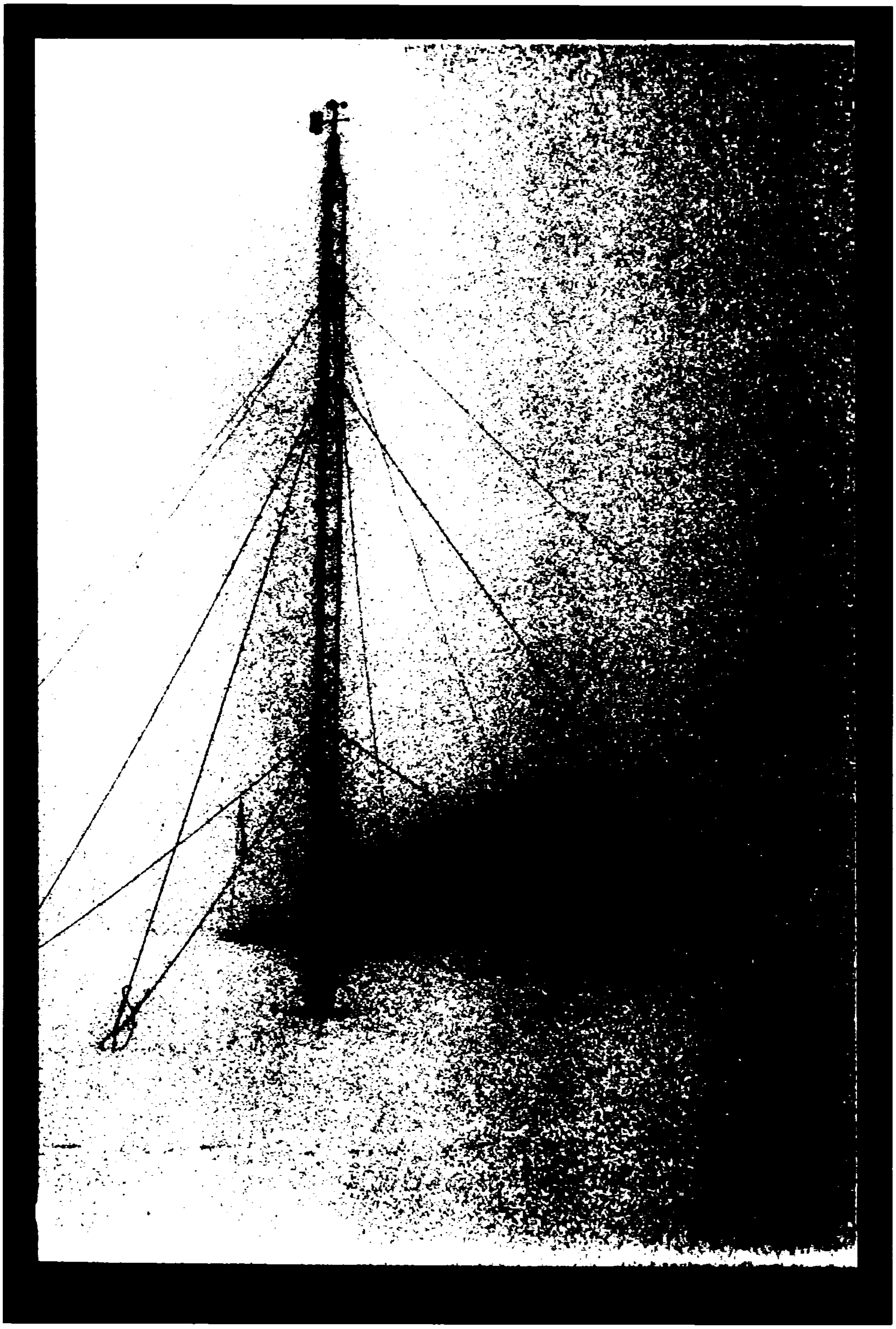

I dedicate this thesis to Maria Andersen whose strength, love, and support have helped focus my mind and increase my resolve. I hope she will always seek knowledge and find new ways to explore the wonderful world(s) around her.

ABSTRACT

The search for extraterrestrial life begins with understanding how life developed and evolved on our own planet. Earth's polar regions provide a unique setting conducive to developing the methodologies and techniques that will be needed to find new life forms either as living organisms or as some biological signal in a sedimentary record.

The highest latitude perennial spring system in a region of thick, continuous permafrost resides in the Arctic archipelago of northern Canada. At Expedition Fiord on Axel Heiberg Island there are two sets of springs that provide useful analogs to liquid water habitats that may have existed on Mars. The springs occur in a region with a mean annual air temperature of -15.5° C. Spring flow rates and discharge temperatures are constant throughout the year. Filamentous bacteria, biofilms and mineral precipitates occur in association with the emergent, anoxic brine flowing from the springs.

Preliminary data on the microbial composition of the spring water has been obtained by applying the culture-independent approach. The majority (76%) of the fifty-five environtaxa showed high sequence similarity to *Thiomicrospira* species (sulfur-oxidizing organism). Other sequences show high similarity to sulfate-reducing members of the delta (*Desulfocapsa* sp.) and epsilon (*Sulfurospirillum* sp.) proteobacterium groups. A single sequence was found to have 99% sequence

similarity to species of the genus *Haloanaerobium* a group of low G+C Gram positive, anaerobic, halophiles (Rainey et al 1995).

A combined flow and thermal model of the Axel Heiberg springs has demonstrated how such springs are able to persist throughout the year despite temperatures that fall to below -50°C during the winter darkness. Dissolved gases emanating from the springs provide evidence of the origin of the water for the springs as well as placing constraints upon the residence time. The gas is composed primarily of N_2 with relative concentrations of Ar, Kr, and Xe almost identical to air. No O_2 is detectable and Ne is 60% of air values. We believe that ~50% of this gas originates from the direct release of air by nearby alpine glaciers and local ice sheets into groundwater that infiltrates sub-ice sedimentary deposits.

RÉSUMÉ

La vie est connue sur notre planète pour trouver place dans les confins les plus reculés. Si de l'eau à l'état liquide est présente, et ce, même en quantités infimes, il est grandement probable qu'une forme de vie, quelque soit-elle, se soit adaptée pour tirer profit des condition environnantes. L'avènement des méthodes moléculaires sans culture pour l'identification des microorganismes a révélé à la science un nombre stupéfiant de nouveaux organismes. De nouveaux mondes à l'intérieur de notre système solaire sont d'ores et déjà explorés avec l'espoir qu'un jour un cousin cosmique de la vie terrienne soit trouvé au sein du permafrost martien ou dans un des océans couverts de glace d'une lune jovienne.

Le plus septentrional des systèmes de sources pérennes recensés dans des régions de pergélisol continu et profond se trouve dans l'archipel arctique du nord du Canada. A Expedition Fiord sur l'île Axel Heiberg, deux ensembles de sources pérennes représentent des analogues représentatifs de ce que pourraient avoir été des habitats caractérisés par de l'eau liquide sur Mars. Les sources se trouvent dans une région dont la température moyenne annuelle est de -15,5°C et dont le sous-sol est caractérisé par un pergélisol atteignant des profondeurs de 400 à 600 mètre. Les taux d'écoulement des sources et les températures du débit sont constant tout au long de l'année. Des bactéries filamenteuses, des biofilms ainsi que des précipités minéraux sont présents là où émerge une saumure anoxique provenant des sources. Cet environnement de très basses températures est significatif des systèmes

hydrothermaux observés en présence de profond pergélisol., ainsi qu'auraient pu l'être ceux de Mars en temps plus clément.

Les données préliminaires portant sur la composition microbienne de l'eau de la source ont été obtenues à l'aide d'une méthode sans culture. La majorité (76%) des cinquante-cinq environtaxa a montré une grande similarité des séquences à l'espèce *Thiomicrospira* (organisme sulfure-oxydant). D'autres séquences ont montré une grande similarité aux membres sulfate-réducteurs des groupes proteobacterium delta (espèce *Desulfocapsa)* et epsilon (espèce *Sulfurospirillum*). Une seule séquence montra une similarité de 99% de la séquence aux espèces du genus Haloanaerobium, un groupe d'halophiles anaérobiques et positifs à faible G+C Gram (Rainey et al 1995).

Un modèle combinant le comportement thermique et les propriétés d'écoulement des sources pérennes a pu démontrer de quelle manière ces sources peuvent subsister au long de l'année, et ce par des températures qui tombent régulièrement sous les – 50 °C pendant les mois d'obscurité. Des gaz dissous émanant des sources permettent d'identifier l'origine de ces eaux ainsi que de déterminer les contraintes relatives à leur temps d'emprisonnement. Le gaz est composé principalement de N_2 ainsi que de concentrations relatives d'Ar, de Kr et de Xe quasi-similaires à celles de l'air. Nous pensons qu'environ 50 % de ce gaz tire son origine de l'éjection d'air par les glaciers alpins et calottes de glace avoisinants dans l'eau souterraine qui infiltre les dépôts sédimentaires sous-glaciaires.

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There are many people I would like to thank for supporting my adventures in learning throughout the years. I am particularly indebted to Chris McKay, George Simmons, Bob Wharton, Penny Firth, John Rummel, Linda Billings, Sylvia Earle, Phil Ballou, Ragnhild Landheim, for their interest, guidance and help over the years.

In the field I had assistance by Chris Omelon, Greg De Pascale, Nicole Couture, Ian Hawes, Ellie Bell, Lyle Whyte, Nancy Perreault, Roger Knowles and Fred Rainey. I am indebted to them for their help with fieldwork, numerous discussions and camaraderie while in the high arctic.

Thanks to Chris Romanek who provided elemental analysis at the Savannah River Ecology Lab, and to Fred Rainey, Norm Pace and Carrine Blank for providing molecular phylogenetic analyses of samples I collected in the high arctic. Lynn Gillespie and Laurie Consaul of the Canadian Museum of Nature in Ottawa provided identifications for plants collected at the spring sites.

I also would like to acknowledge the Canadian Polar Continental Shelf Project for its superb logistical support in the Canadian High Arctic and NASA's Exobiology Program for research support during the course of this project.

Most of all, I am indebted for life to my advisor, Wayne Pollard for his friendship, intellectual guidance and relentless support during our tenure together. I am constantly amazed at how much I have learned from him, particularly while in the field, but more importantly he has shown me how much more there is to understand and experience in the far north. My thanks also to the other members of my committee, Drs. Nigel Roulet, Hojatollah Vali and George Wenzel for their input and comments on this thesis. In addition to educating me about the finer points of electron microscopy, Dr. Vali and Kelly Sears have always made me feel at home at the McGill Facility for Electron Microscopy Research. I am indebted to both for encouraging my participation in Canadian Astrobiology and helping with my research. I also would like to acknowledge the support I have received from the faculty, staff and students within the Department of Geography at McGill. This has been a wonderful place to carry out these studies and the warm and productive environment within this department played an important role in my studies.

To Sasha, my wife, our daughter Masha, my parents, and my sister Dixie: I cannot even begin to tell you how grateful I am for the unremitting emotional support you have provided. Knowing that I could always count on you was critical to the completion of this body of work.

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STATEMENT OF MANUSCRIPT FORMAT

Candidates have the option of including, as part of the thesis, the text of one or more papers submitted, or to be submitted, for publication, or the clearly duplicated text (not the reprints) of one or more published papers. These texts must conform to the "Guidelines for Thesis Preparation" with respect to font size, line spacing and margin sizes and must be bound together as an integral part of the thesis.

The thesis must be more than a collection of manuscripts. All components must be integrated into a cohesive unit with a logical progression from one chapter to the next. In order to ensure that the thesis has continuity, connecting texts that provide logical bridges preceding and following each manuscript are mandatory.

The thesis must conform to all other requirements of the "Guidelines for Thesis Preparation" in addition to the manuscripts. The thesis must include: a table of contents; a brief abstract in both English and French; an introduction which clearly states the rational and objectives of the research; a comprehensive review of the literature (in addition to that covered in the introduction to each paper); a final conclusion and summary; and a thorough bibliography.

Additional material must be provided (e.g., in appendices) in sufficient detail to allow a clear and precise judgment to be made of the importance and originality of the research reported in the thesis.

In general, when co-authored papers are included in a thesis the candidate must have made a substantial contribution to all papers included in the thesis. In addition, the candidate is required to make an explicit statement in the thesis as to who contributed to such work and to what extent. This statement should appear in a single section entitled "Contributions of Authors" as a preface to the thesis. The supervisor must attest to the accuracy of this statement at the doctoral oral defense. Since the task of the examiners is made more difficult in these cases, it is in the candidate's interest to clearly specify the responsibilities of all the authors of the co-authored papers.

STATEMENT REGARDING THE ROLE OF CO-AUTHORS

One of the manuscripts included in this thesis has been published in a peer-reviewed journal and one has been submitted for publication: Andersen, D.T., W.H. Pollard, C.P. McKay, and J. Heldmann, Cold springs in permafrost on Earth and Mars, ***Journal of Geophysical Research***, *107* (3), 4, 2002: (Chapter 3), Andersen, D.T., C.P. McKay, W.H. Pollard, and B. Hudson, Composition and Source of Gas Bubbling from a Saline High Arctic Spring, ***Journal of Geophysical Research***, Submitted, 2004: (Chapter 4). Although the manuscript is neither submitted or in press, co-authors are listed for Chapter 5 in order to adequately acknowledge their collaboration and input in anticipation of future publication of these results. The published and submitted manuscripts are reproduced here in full, in a format that is consistent with the rest of the thesis.

In accordance with McGill University Faculty of Graduate Studies and Research guidelines, I declare here that the research presented in these papers is work of my conception and execution undertaken with the input and or assistance of the listed co-authors. The role of the listed co-authors in the production of these papers is elaborated below.

Chapter 3: Cold Springs in Permafrost on Earth and Mars

Dr. Christopher P. McKay (Space Science Division, NASA's Ames Research Center, Moffett Field, CA) assisted in the development of the FORTRAN code that was used to model the thermal system at Expedition Fiord, Axel Heiberg Island. Dr. McKay provided technical instruction in the computer software and the VAX/VMS computer that was used to run the code. Dr. Wayne Pollard provided technical expertise and guidance with respect to the permafrost conditions found at Axel Heiberg. Both co-authors provided intellectual comments on drafts of the manuscript prior to submission to JGR Planets. Dr. Jen Heldmann a Ph. D. student at the University of Colorado at the time this manuscript was being prepared provided intellectual input during the preparation of the manuscript. She also helped with some of the fieldwork associated with the paper particularly with meteorological data.

Chapter 4: Composition and Source of Gas Bubbling from a Saline High Arctic Spring

Dr. McKay and I have authored (and co-authored) papers that have dealt with the nature of dissolved gases in perennially ice-covered lakes in the Antarctic and this paper evolved from discussions we had about this topic over the years. Data analysis was performed at NASA Ames using FORTRAN code previously developed by Dr. McKay. Dr. Hudson, at the Noble Gas Lab, Lawrence Livermore National Lab performed sample analysis of some samples using techniques and capabilities that are available at this state of the art facility. He also provided insight with respect to noble gas analysis in groundwater in general. Dr. Pollard provided assistance with sampling the gases at the springs and intellectual comments on drafts of the manuscript.

Chapter 5: Perennial Springs in the Canadian High Arctic: Analogues of Martian Hydrothermal System

This chapter evolved primarily from discussions with Hojatollah Vali, Lyle Whyte, Carrine Blank, and Fred Rainey. Dr. Vali aided in the use of scanning electron microscopy as a means to better understand the interaction of microorganisms and biomineralization. Drs. Whyte and Rainey and Blank provided field assistance and analytical support during the execution of this phase of the research. Dr. Pollard's expertise in low temperature settings helped shape the research and he provided intellectual support throughout this effort.

ORIGINAL CONTRIBUTION TO KNOWLEDGE

The work presented in this thesis makes original contributions in the disciplines of geography, planetary science, Arctic hydrology, and microbial ecology. Studies of perennial springs in regions of thick, continuous permafrost are rare due to the few examples that exist. This study represents the first major effort undertaken to understand the physical, chemical and biological attributes of the northernmost spring system in the Canadian High Arctic.

Chapter 3 examines the thermal regime associated with the perennial springs occurring at Axel Heiberg Island and uses the interpreted data to model conditions that may have been present on the planet Mars. A combined flow and thermal model of the subsurface flow associated with the springs at Expedition Fiord using the locally measured geothermal gradient was developed to explain how such springs sustain themselves in regions of thick permafrost on Earth and possibly on Mars.

Chapter 4 presents data that support the hypothesis that the source of the spring water is a combination of sub-glacial flow and water stored in local ice-dammed alpine lakes situated several hundred meters above the springs. Measurements of the dissolved gases in the spring outflows and the alpine lakes suggest that approximately 50% of the water is derived from each of these sources. These data are the first reported for dissolved noble gases in subsurface waters from the high arctic.

Chapter 5 presents a detailed analysis of the chemical and physical components of the spring ecosystem and provides the first data regarding the biological components. Using culture independent methods. phylogenetic analysis of environtaxa provides insight into the phylogenetic and physiological diversity of the organisms present in the spring water. Petrographic analysis and micropalentological examinations of mineral precipitates were used to investigate possible biomineralization and preservation.These data are novel and provide the first glimpse of the diverse microbial communities associated with the spring environment. The rarity of these springs cannot be over emphasized and as a part of this study, other sites were located on Axel Heiberg that will warrant further study in the future.

CHAPTER 1: PERENNIAL SPRINGS IN THE CANADIAN HIGH ARCTIC: ANALOGUES OF HYDROTHERMAL SYSTEMS ON MARS

1.1 Introduction

Water is present throughout the universe occurring in the gaseous phase within the interstellar medium, and as low temperature, amorphous ice accreted upon interstellar dust grains (Committee on Planetary Biology and Chemical Evolution 1990). Liquid water appears to be much less common, and to date the only place liquid water has been observed is on our own planet, Earth. There is strong evidence that liquid water exists elsewhere in our solar system beneath the thick ice-cover of several Jovian moons, and perhaps, beneath the thick permafrost of Mars (McCord et al. 1998, McCord et al. 2001, Baker 2001). Water, with its unique chemical and thermodynamic properties provides a medium within which life has originated, evolved and thrived on Earth – but only in its liquid form. Ongoing efforts to discover, catalog, and understand life on Earth have yielded results that provide insight not only about the history of life on Earth but also about the range of habitats and extreme environments where life may be found beyond Earth. Studies of life on Earth residing in extreme environments are helping to define the search for life elsewhere in the universe. Characteristics of extreme terrestrial environments may include physical/chemical properties such as desiccation, high salinity, high radiation, low light, high pressure, high and low temperatures, extreme pH, and high or low oxygen tension. During the last twenty years investigations of terrestrial ecosystems around the planet have yielded an astonishing number of newly recognized microorganisms living in habitats in almost every conceivable place (Rothschild and Mancinelli 2001). Despite enormous metabolic diversity and other clever adaptations used by the organisms that inhabit these niches, one fundamental requirement they all have in common is the necessity of liquid water which is required by cellular machinery to carry out metabolic and reproductive functions. Thus, the search for life is by default a search for liquid water.

A long-term, multinational commitment to the exploration of Mars is now underway (Squyres et al. 2004, Kerr 2004). Currently there are three spacecraft in orbit and two rovers operating on the surface of Mars with the primary goal of understanding the history and presence of water on Mars. While these and previous missions have provided a wealth of information, future missions designed to detect life, past or present, will rely upon the lessons learned from examples of terrestrial microbial life, to design, build and test the methods and tools that will be needed on Mars.

Earth's polar regions provide relevant analogues to ecosystems which may have existed on an earlier, more clement planet Mars (Andersen et al. 1992, McKay 1993). In regions of thick, continuous permafrost, the behavior of water is qualitatively different, its presence occurring primarily in the mineral phase as ice. Melting occurs principally as a result of seasonal and geothermal warming, and perennial liquid water environments are uncommon. While there is ample evidence that liquid water once existed at or near the surface of Mars, current climate models suggest that mean annual temperatures on early Mars remained below 0°C and permafrost has dominated the planet throughout its history (Segura et al. 2002). Liquid water, being the non-negotiable prerequisite for the origin and evolution of living systems would have been required on Mars for life to have evolved there. These aqueous environments would have been constrained by similar physical and chemical processes unique to polar settings on Earth. The Canadian High Arctic, particularly Axel Heiberg and Ellesmere Islands offer a number of analogous features found in association with the low temperature climate regime and the resulting thick, continuous permafrost.

1.2 Research Objectives

The primary purpose of this study was to conduct a multidisciplinary research effort to examine the geomorphic, physical-chemical, and biological aspects of perennial springs and associated mineral precipitates located near Expedition Fiord, Axel Heiberg Island, Nunavut. Three guiding hypotheses were developed to focus the work: (1) the thermal regime of the perennial springs is controlled by the local geothermal gradient, not local magmatic hotspots; (2) the source of the water for the springs is local meteoric water derived from either sub-glacial melt, local alpine lakes, or a combination of both; (3) prokaryotic communities associated with the springs are diverse, contribute to the formation of the mineral deposits, and are preserved within the sedimentary record. A secondary goal was to apply the results from this work to conceptual models of environmental conditions that may have been present on early Mars with the express purpose to further future exploration of that planet.

NASA's Exobiology program, which seeks to understand the origin, evolution and distribution of life in the universe, utilizes Earth analogues (particularly temperate hydrothermal systems such as those found in Yellowstone National Park, undersea vents and perennially ice-covered lakes in Antarctica) as part of ongoing research directed towards understanding the biological history of our solar system. Studies of subsurface hydrology and associated mineral and biological assemblages in high latitude regions with thick, continuous permafrost are rare, and studies of the perennial springs located at Axel Heiberg have provided an opportunity to make significant, original contributions to knowledge of these perennially cold environments.

The springs at Expedition Fiord are very unusual with respect to their location in a thick permafrost setting. As such, a number of fundamental questions need to be answered:

1. What are the physical processes that maintain perennial flow despite mean annual temperatures that are well below freezing and the presence of thick (~600m) permafrost?

2. What are the sources of water for the springs?

3. What constraints can be placed upon the age of the emergent spring water?

4. What microbial assemblages are associated with the springs, and are biological signatures incorporated into the mineral precipitates as a fossil record?

5. Do these spring environments provide a relevant analogue to ancient thermal spring sites on early Mars?

From these questions, the following objectives were developed:

1. Using spring flow and outflow temperature records in conjunction with observations of regional geology and available subsurface thermal histories, develop a thermodynamic model to explain perennial flow through the permafrost.

2. Obtain samples and measurements to determine the origin of the spring water considering sub-glacial melt and/or large alpine lakes as potential sources.

3. Constrain the age of the emergent spring water.

4. Determine the major groups of microorganisms associated with the springs using culture independent methods.

5. Determine the extent of preservation of biological signals in the mineral precipitates.

6. Determine the extent and distribution of perennial springs in the Canadian High Arctic.

7. Develop models of this permafrost spring system that can be extrapolated to Mars in order to aid in site selection and experiment development for future Mars missions.

1.3 Study Area

Axel Heiberg Island (Fig 1) resides within the Sverdrup Basin, a pericratonic sedimentary trough encompassing roughly 313,000 km^2 and located within the Arctic Archipelago (Pollard *et al.* 1999). The axis of the basin strikes northeasterly from the Sabine Peninsula of Melville Island to the northwestern region of Ellesmere Island. Axel Heiberg is geologically complex consisting of folded and faulted sedimentary rocks ranging from Triassic to Tertiary in age. Late Paleozoic evaporites locally intrude the overlying sedimentary/clastic rocks in a series of exposed piercement structures that tend to occur mainly in line with the basin axis (Stephenson et al. 1992).

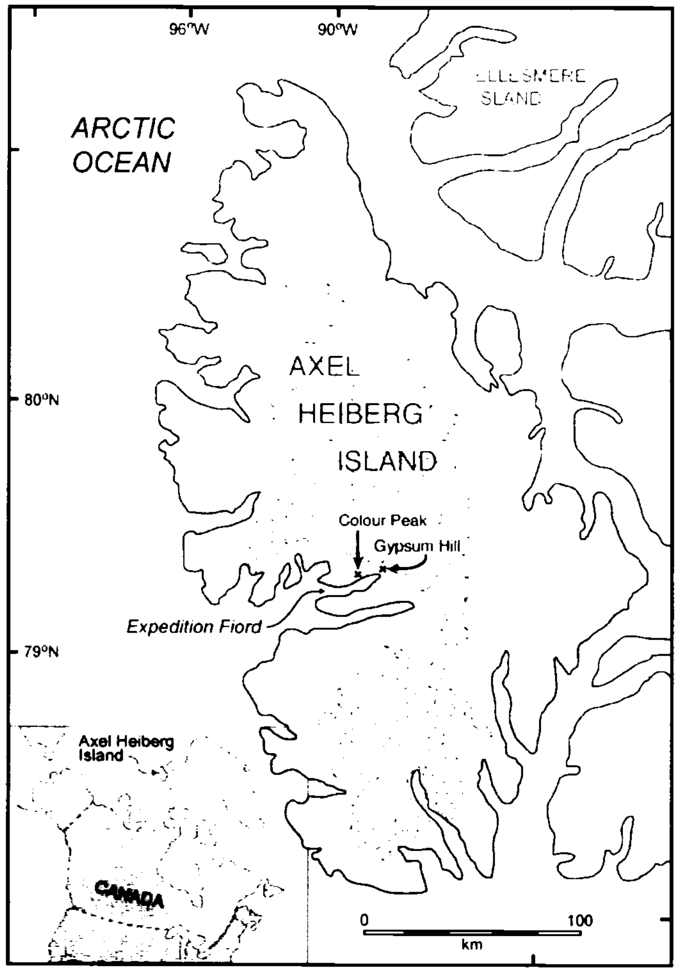


Figure 1.1 Location map of Expedition Fiord, Axel Heiberg Island, Nunavut Canada.

The intrusions of the evaporites are typically associated with early Tertiary (Innuitian) orogenic activity and are mapped as Tertiary (Otto Fiord Formation) by Thornsteinsson (1971). However the emplacement of the anhydrite structures at Expedition Fiord took place during an earlier tectonic phase prior to the mid Cretaceous (Embry and Osadetz 1988, Stephenson et al. 1992). Quaternary stream, deltaic, marine and glacial sediments cover the valley floors and form raised shorelines.

There is considerable relief near the head of Expedition Fiord with peaks approximately 2000 m ASL that decrease toward the fiord mouth. Breached anticlines produce asymmetrical ridges characterized by steep (70-80°) scarp faces and 25-35° dip slopes. In some locations, piercement structures create more regular and symmetrical hill and mountain features. Weathering of gypsum and anhydrite outcrops has given the diapirs very ragged, serrated profiles. Resistant volcanic sills and dikes have differentially eroded leading to steep slopes mantled with coarse angular talus.

Glaciers currently cover 30-35% of Axel Heiberg Island, including the Stacie and Müller (McGill) Ice Caps and numerous outlet and valley glaciers (Figure 1.2). Small ice caps and isolated cirque and valley glaciers are widespread. At Expedition Fiord the White and Thompson glaciers converge roughly 10 km up valley from the head of the fiord, while the terminus of Crusoe glacier lies only a few kilometers west of the Expedition River Valley.

1.4 Climate and permafrost

The north and northwest arctic archipelago is characterized by polar desert conditions displaying very cold, dry winters and cool summers with maximum precipitation occurring in July (Maxwell 1982). Intermittent climate records for the Expedition Fiord locale are available for the last 30 years, with a more complete record for the last 12 years. A long-term meteorological station is also located at Eureka (79°59'N, 85°56'W, 10 m ASL) on Ellesmere Island, approximately 120 km

NE of Expedition Fiord. Long-term averages for Eureka are characterized by a mean annual air temperature of -19.7°C and mean monthly temperatures for January and July of -36.1°C and +5.4°C, respectively (extreme winter minima of -55°C are not uncommon). Inland, July air temperatures on Ellesmere Island frequently reach 20°C and nival streams rapidly recede after mid-June and are often dry by mid-July. Precipitation at Eureka is lower than measured at Expedition Fiord due to a rain shadow effect caused by the mountains on Axel Heiberg Island (Edlund and Alt 1989).

Figure 1.2 ASTER Image of the Expedition Fiord Region, Axel Heiberg

Early climatological research at Expedition Fiord suggests a mean annual accumulation of 371 mm (water equivalent) on the Müller Ice Cap (Müller 1963). Year-round data from an automatic weather station at Colour Lake (elevation 64 m ASL) has been collected since 1992 indicates a mean annual temperature of –15.5° C with approximately 451 thawing degree-days and 6084 freezing degree days during a typical year (Andersen et al. this study, Doran et al. 1996). Tables 1.1-1.3 list the mean average temperatures, thawing and freezing index for each month during the years 1998-2002.

Permafrost depth has not been measured at Expedition Fiord, however, a permafrost thickness of > 400 m was documented in an exploration well at Mokka Fiord on the east side of Axel Heiberg Island, roughly 60 km from Expedition Fiord (Taylor and Judge 1976). Other exploration wells in the area reveal that permafrost is generally between 400-600 m thick. Permafrost features include extensive polygonal ice wedge development in unconsolidated fluvial and colluvial deposits at lower elevations. Wet areas develop

Table 1.1 Mean Average Temp 1998-2002: -15.50 °C

	1998	1999	2000	2001	2002
January	-31.96	-32.6	-31.16	-32.42	-33.4
February	-34.38	-27.86	-23.83	-33.76	-35.24
March	-28.48	-17.89	-15.94	-30.02	-31.02
April	-16.65	-4.68	-4.88	-20.8	-25.07
May	-5.47	5.17	6.82	-10.13	-6.26
June	5.39	9.55	6.64	2.3	1.98
July	7.36	0.29	-2.64	5.23	4.42
August	2.86	-9.68	-13.88	3.34	4.08
September	0.1	-25.34	-23.89	-5.92	-2.61
October	-15.27	-31.69	-25.7	-20.3	-10.07
November	-23.55	-35.03	-24.48	-23.53	-26.93
December	-33.33	-35.58	-22.1	-26.28	-24.08
Mean	-14.45	-17.11	-14.59	-16.02	-15.35

Table 1.2 Thawing index

	1998	1999	2000	2001	2002
January	0	0	0	0	0
February	0	0	0	0	0
March	0	0	0	0	0
April	1.07	10	17.65	0	0
May	7.79	160.68	211.4	2.7	4.65
June	162.13	286.55	207.19	92.62	83.51
July	228.66	57.35	20.62	163.09	135.61
August	91.36	0.29	0	114.09	128.17
September	36.51	0	0	7.89	13.34
October	0	0	0	0	8.96
November	0	0	0	0	0
December	0	0	0	0	0
Total	527.52	514.87	456.86	380.39	374.24

Table 1.3 Freezing index

	1998	1999	2000	2001	2002
January	-990.7	-1010.57	-965.81	-1004.93	-1035.31
February	-962.77	-780.15	-690.93	-945.34	-986.78
March	-882.79	-554.66	-494.18	-930.57	-961.77
April	-500.69	-150.44	-164.06	-624.09	-752.15
May	-176.43	-0.46	-0.07	-316.71	-198.68
June	-0.53	0	-8.12	-23.67	-25.12
July	-0.36	-48.27	-102.42	-1.03	-1.19
August	-2.8	-300.44	-430.37	-10.69	-1.72
September	-33.6	-760.32	-716.74	-185.55	-91.62
October	-473.34	-982.51	-796.72	-629.34	-321.25
November	-706.5	-1050.9	-734.33	-705.98	-807.98
December	-1033.28	-1102.86	-685.04	-814.7	-746.61
Total	-5763.79	-6741.58	-5788.79	-6192.6	-5930.18

tussock microtopography while poorly sorted circles and stripes characterize midslopes. Two small pingos exist in outwash gravels near Middle Fiord while various frost and icing blisters and localized icings and icing mounds develop in the Expedition River floodplain. Massive ground ice is widespread in the Eureka Sound Lowlands (Pollard 1991*b*) but is limited to ice-cored moraines at

Expedition Fiord. Numerous active layer detachment slides and nivation hollows are present along hill-slopes in the region.

1.5 Gypsum Hill and Colour Peak Springs

The two study areas are located approximately 11 km apart along the northern boundary of the Expedition River floodplain and Expedition Fiord. Both spring systems are associated with anhydrite/gypsum diapirs and emerge from the base of the piercement structures. The McGill Arctic Research Station (MARS), the logistical base of operations for this project, is located on the western side of Gypsum Hill at an elevation of approximately 200 m.

The Gypsum Hill spring site (Figure 1.3) consists of approximately forty springs and seeps on the northeast side of Expedition River, discharging along a band nearly 300 m long and 30 m wide between 10-20 m ASL. The springs are concentrated at the break in slope where bouldery colluvial materials overlap sandy outwash. The surface around the springs is littered by large boulders from both the till and exposed anhydrite. The area immediately surrounding the springs consists of small mounds separated by shallow gullies.

There are outflows occurring in Expedition River; however thick ice and snow in the winter and high stream flows during the summer have prevented determination of their exact locations. Pollard et al. (1998, 1999) estimated the total discharge of the Gypsum Hill springs to be approximately 10-15 l/s.

The Colour Peak springs (Figures 1.4 and 1.5) are located on the south-facing slope of Colour Peak at an approximate elevation of 100 m ASL, emerging from the top of the slope along a line nearly 400 m long. These springs are grouped into 3 distinct topographically controlled areas with 20 vents discharging directly into Expedition Fiord three hundred meters down slope. Unlike the springs at Gypsum Hill, the Colour Peak springs form distinct troughs, pipes, and sheet flows with well-developed travertine and other mineral precipitates. A lateral moraine mantles the slopes immediately above the springs while sandy beach and alluvial fan deposits mantle the slopes along the edge of the fiord. The surface in the immediate vicinity of the springs is a combination of weathered bedrock and a grey silty mud covered by grey and black precipitates. The lowest spring is nearly 15 m ASL. There is no evidence of discharge in the fiord, although it is conceivable that spring flow may be occurring beneath the water line.

Icing pastes and surface efflorescence of salts are present in and around the spring outflows and along flow paths at Colour Peak and Gypsum Hill. High salinity in the soils surrounding the springs appears to have inhibited the growth of the local plant flora (Lynn Gilespi, personal communication) with the alkali grasses of the genus *Puccinellia* being found closest to the springs. The moderately salty ground in the vicinity of the springs supports a plant community distinct from the surrounding tundra (for example Fleabane and Foxtail were not observed at the spring locations, but are commonly found in the surrounding tundra). Table 1.4 lists the plants collected and identified at the two spring sites.

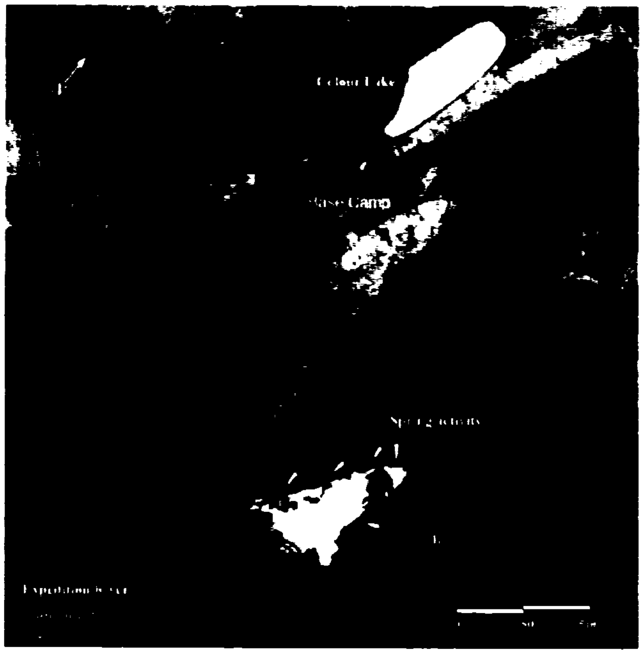


Figure 1.3 Gypsum Hill study area showing location of springs, base camp, and Colour Lake; note the size and shape of the spring icing. Airphoto A30860-139 © 1973 Her Majesty the Queen in Right of Canada, reproduced from the collection of the National Air Photo Library with permission of Natural Resources Canada.

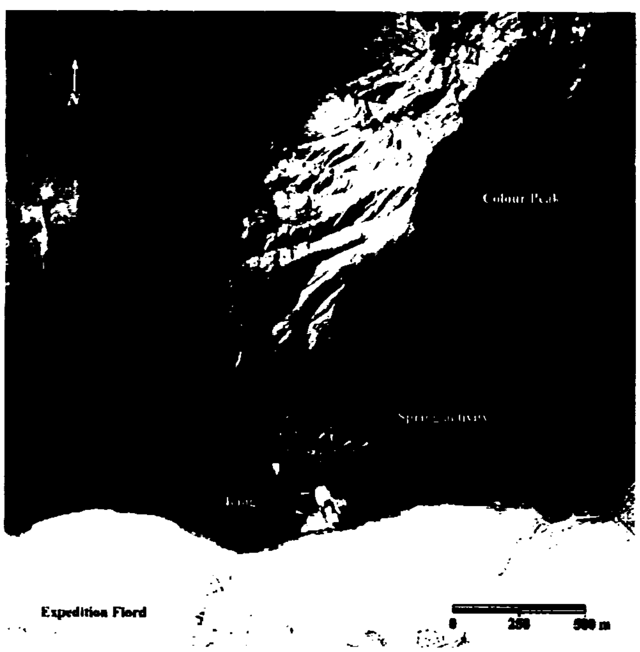


Figure 1.4 Colour Peak study area showing location of springs adjacent to Expedition Fiord; note smaller icing associated with spring discharge. Airphoto A30860-133 © 1973 Her Majesty the Queen in Right of Canada, reproduced from the collection of the National Air Photo Library with permission of Natural Resources Canada.

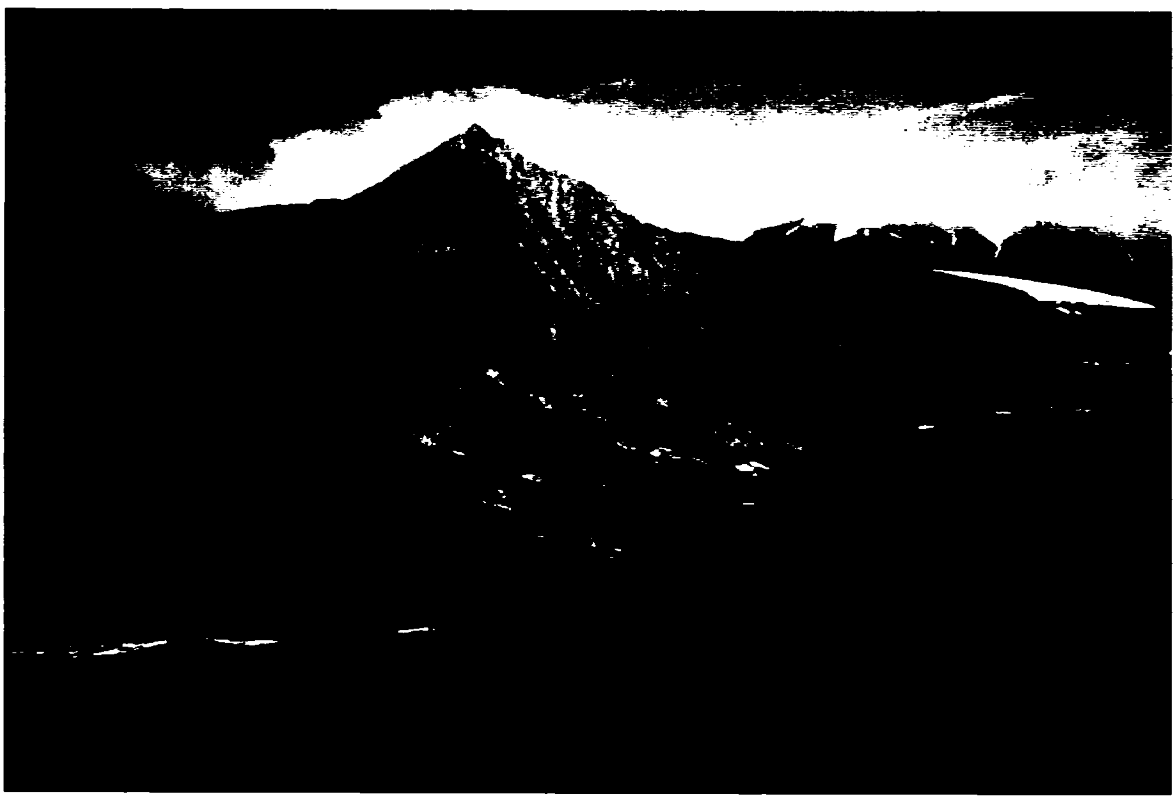


Figure 1.5. Perennial springs at Colour Peak, Axel Heiberg Island, Nunavut.

Table 1.4 Plants observed and collected at Gypsum Hill and Colour Peak springs.

Species	Common name	Plant family	Colour Peak			Gypsum Hill	
			1 Colour Peak (top)	2 Colour Peak (bottom)	3 Colour Peak (bottom)	1 Gypsum Hill springs	2 Gypsum Hill Spring area
Alopecurus alpinus	foxtail	Poaceae (grass family)			x		
Braya purpurescens	braya	Brassicaceae (mustard family)					x
Cerastium alpinum	mouse-ear chickweed	Caryophyllaceae (pink family)					x
Cochlearia officinalis	scurvy-grass	Brassicaceae (mustard family)					x
Draba oblongata	draba	Brassicaceae (mustard family)					x
Equisetum scirpoides	horsetail	Equisetaceae (horsetail family)	x				
Erigeron eriocephalus	fleabane	Asteraceae (aster family)			x		
Festuca edlundiae	fescue	Poaceae (grass family)			x		x
Minuartia rubella[1]	sandwort	Caryophyllaceae (pink family)					x
Papaver radicatum	arctic poppy	Pappaveraceae (poppy family)			x		x
Poa hartzii	Hartz's spear-grass	Poaceae (grass family)	x				x
Potentilla pulchella	cinquefoil	Rosaceae (rose family)	x		x		x
Puccinellia andersonii	alkali grass	Poaceae (grass family)		x	x	x	x
Puccinellia angustata	alkali grass	Poaceae (grass family)		x	x	x	x
Puccinellia phryganodes	alkali grass	Poaceae (grass family)			x		
Salix arctica	arctic willow	Salicaceae (willow family)	x		x		x
Saxifraga oppostitifolia	purple saxifrage	Saxifragaceae (saxifrage family)					x
Silene involucrata subsp. *involucrata*[2]	arctic bladder-campion	Caryophyllaceae (pink family)	x		x		x
Silene uralensis subsp. *arctica*[3]	purple bladder-campion	Caryophyllaceae (pink family)			x		
Stellaria longipes	star chickweed	Caryophyllaceae (pink family)			x		x
Taraxacum phrymatocarpon	arctic dandelion	Asteraceae (aster family)	x				x
Unidentified	moss	**********			x		

1 name in Porsild: *Arenaria rubella*
2 name in Porsild: *Melandrium affine*
3 name in Porsild: *Melandrium apetalum*

1.6 Other sites of interest

The two spring sites at Expedition Fiord have been the primary focus of this investigation; however there are other sites of interest that have been either identified via the literature or discovered during the course of this study. A list of features and their locations (including the primary study sites) is given in Table 1.5. Reconnaissance flights have revealed additional evidence of subsurface flow in the form of perennial springs, icings, and pingos. It is not clear if all of these sites exhibit perennial flow such as the springs at Expedition Fiord, but it does indicate that other examples of subsurface hydrology occur in the region.

Table 1.5 Location Index for known springs, icings and pingos

Feature	Latitude	Longitude
Gypsum Hill Springs	N79°22.192'	W090°43.939'
Colour Peak Springs	N79°22.866'	W091°16.270'
Stolz Diapir Springs[2]	N79°05.343'	W87°02.228'
Junction Diapir Spring[1]	N79°03.736'	W090°41.893'
Lost Hammer Spring[1]	N79°04.285'	W090°12.755'
Skaere Fiord Glacial Outflow[1]	N78°56.702'	W088° 17.666'
Borup Fiord Glacial Outflow [2]	N81°01.660'	W081°38.484'
Middle Fiord Pingo 1	N 79°43.653'	W94°13.794'
Middle Fiord Pingo 2	N79°47.154'	W94°09.343'
Icing 1 North West Axel[1]	N 80°22.749'	W93°57.584'
Icing 2 West Axel[1]	N79°58.966'	W93°57.539'
Icing 3 Strand Fiord[1]	N79°17.517'	W090°06.314'

[1]located during this study
[2]located via previous literature

1.7 Methods

Sampling and analytical procedures used in the field and laboratory during this project are described in detail below.

Climate

Climatic conditions at Expedition Fiord were monitored during this project at two principal sites. Meteorological data were collected at Colour Lake with a Campbell Scientific CR10 datalogger emplaced in 1992 by Doran et al. (1996). The station at Colour Lake initiates measurements every 30 seconds and records averages every 30 minutes. A model 05013 RM Young wind monitor (wind speed and direction) and LiCor LI200SZ (solar flux) are affixed to the mast ~2.5 m above ground, and a model 207 Phys-Chem relative humidity-temperature sensor housed in a radiation shield is attached to the mast ~2.2 m above the ground. Ground temperature is measured with two Campbell Scientific 107B temperature thermistors buried at depths of 2 cm and 34 cm. Data from the station were collected twice a year (spring, summer) providing a continuous record for the years 1998-2002. A second Campbell Scientific CR10 datalogger was placed next to the largest spring outflow at Colour Peak to monitor local air temperature and humidity using a model 207 Phys-Chem relative humidity-temperature sensor housed in a radiation shield approximately 1.5 meters above the ground. Spring flow was continuously recorded using an Omega FP-5300 flow meter, and spring outlet temperature was measured with a Campbell Scientific 107B temperature thermistor placed underwater. Spring temperatures were also monitored with Hobo dataloggers (Onsett Computer) placed into sealed Nalgene bottles. Thermistors leading from the dataloggers were placed into the spring water to record temperature every 30 min. Spot measurements were made in various locations using a Fluke model 51 digital thermometer (accuracy 0.05% + 0.3°C) with an external type J thermocouple.

Aerial and satellite imagery (ASTER/RADARSAT)

Satellite and aerial imagery provided local and regional information and was used in conjunction with available geological maps. Satellite imagery also provided a means to observe spring flow in the Expedition River outflow channel during the winter months. All aerial oblique images were obtained by using handheld 35mm cameras during fixed wing or helicopter flights over the various research sites. ASTER images of Axel Hieberg and

Ellesmere Island were obtained from the Land Processes Distributed Active Archive Center (LP DAAC). located at the U.S. Geological Survey's EROS Data Center ([illegible]) via the NASA EOS Data Gateway. ASTER images were processed with MultiSpec 3.0 (© 2003 Purdue Research Foundation). Processed RADARSAT images of the Expedition Fiord region were provided by the Canadian Centre for Remote Sensing, Ottawa.

GPS mapping

High-resolution (cm scale accuracy in both horizontal and vertical) GPS coordinates of spring outflows, channels, associated icings, frost mounds, sampling points and other features were obtained using a Trimble 4700 Total Station. Trimble's Survey Office software was used for post processing of data. A Garmin eTrex Vista was used to determine coordinates (accuracy <15m, 95% typical) of other springs, gypsum diapirs, pingos, and icings found during surveys conducted on Axel Heiberg and Ellesmere Islands.

Determination of permafrost depth

Regional permafrost depth was estimated by utilizing temperature data obtained from a nearby borehole at Mokka Fiord (60 km) that was reported by Taylor and Judge 1976.

Dissolved gases

Gas samples collected for this study were obtained at the base of Gypsum Hill in a pool type outlet known as "little black pond" (LBP) and were the primary source for samples in this study. The pool is approximately 1 meter in diameter and is filled with a black, course-grained sediment through which the spring flows. The 6°C water temperature and a flow rate of ~ 200 ml/s at this outflow have remained constant over the last 5 years. Gas bubbles emanating from the outlet provide an easy source for sampling the evolved gases. To obtain samples of the bubbles, 60 cc borosilicate serum vials were submerged in the spring and allowed to fill completely. Using a funnel to guide bubbles into inverted serum vials, bubbles were collected by displacing the water inside the vials. After approximately twice the volume of the vial had been purged the vials were stoppered and crimped while underwater and returned to the laboratory for analysis. In addition, samples of dissolved gas were obtained by filling serum vials underwater, inserting a stopper and crimping them while they remained underwater. *In situ* dissolved O_2 measurements were obtained in Phantom and Astro Lakes using a YSI 95 dissolved O_2 meter and with an Ocean Seven 316 multi-parameter probe (Idronaut, Italy). Water samples from Astro Lake were obtained with a Kemmerer bottle lowered into the water column. Once retrieved, the water was allowed to gently flow from the Kemmerer bottle via polyethylene tubing into 125 cc serum vials and subsequently sealed after overflowing at least two volumes of water.

Samples were analyzed for noble gases and tritium at the Lawrence Livermore National Lab (LLNL), Livermore, CA, following methods outlined by Clark et al. (1998). Measurement uncertainties are 2%. N_2/Ar determinations by LLNL used methods described by Kana *et al.* (1994), with a measurement uncertainty of 1%. Noble gases, N_2, O_2, CO_2, CH_4 and H_2S were also analyzed at NASA Ames Research Center using a SRS model RGA-200 quadrapole mass spectrometer and a Hewlett-Packard 5890–II GCMS. For gas analysis, samples were introduced into a vacuum system connected to the spectrometer through a leak valve. Air was used as the primary standard gas for concentration and ratio measurements, and care was taken to ensure that the ion current for the air standard was similar to that for the measurement gas. For water samples a headspace of 5 ml was first produced, using

He as described by Risgaard-Petersen and Rysgaard (1995) and Blicher-Mathiesen et al. (1998). For the measurement of He in the spring water the headspace was produced with air. For the helium analysis at the NASA Ames facility the error was 10%.

Water chemistry

pH and ORP were measured with an Orion 250Aplus pH/ORP meter. The sensor used for pH was the Orion 91-07 with a Ag/AgCl internal reference system, sealed reference and built-in thermistor for Automatic Temperature Compensation (ATC). The pH was measured after auto-calibration using 3 standard buffers (4.01, 7.0 and 10.01). pH range of the meter is -2.00 to 19.99 with a resolution of 0.01/0.1 pH units and a temp resolution of 0.1°C .

ORP was measured after calibrating with an Orion ASTM-approved ORP Standard (967961). The sensor was an Orion Epoxy Sure-Flow combination electrode (9678BN). The sample's ORP millivolts correlate back to the Normal Hydrogen Electrode (NHE) and ORP millivolts are displayed to 0.1mV resolution over the range of -1999.9 to +1999.9 mV.

Total alkalinity was measured with an Orion Total Alkalinity Test Kit using alkalinity reagent (700011). To a 100 ml sample, 10 ml of the Alkalinity Reagent was added, pH re-measured and the total alkalinity in ppm $CaCO_3$ was determined from the Orion pH conversion chart enclosed with the kit.

Routine water sampling was conducted at the springs, Phantom Lake, and Astro Lake. One liter of water was collected from depths of 5 m, 20 m, 100 m, and 200 m at Phantom Lake and 1 m, 15 m, 20 m, 25 m, 50 m, and 120 m at Astro Lake. Samples were collected at Colour Peak at the main spring outlet, a point half way down the slope (~50 m) and at the base. The water was filtered through GF/F glass fiber filters and the filters stored frozen in the dark for Chl *a* analysis. An amount of 250 ml of unfiltered water was collected for plankton enumeration and immediately fixed with Lugol's Iodine to a final concentration of 2.5%. In addition, 20 ml of water was collected in sterile glass vials for bacterial and viral enumeration and immediately fixed with formalin (final concentration 2.5 %). Furthermore, 500 ml of water was collected and filtered through GF/C glass fiber filters pre-combusted at 450 °C for 5 hours. Four 125 ml aliquots were stored in Nalgene bottles for dissolved organic carbon (DOC), dissolved inorganic carbon (DIC), anions and cations and trace elements, respectively. The DOC and trace element samples were fixed with nitric acid and the DIC samples with a few drops of the bacteriacide, benzalkonium chloride (ZEPHIRAN®). All samples were stored in the dark at 4 °C until analyzed. An Ocean Seven 316 multi-parameter probe (Idronaut, Italy) and a Yellow Springs Instrument (YSI) 6600 were employed to record profiles of lake depth, temperature, conductivity, salinity, pH, O_2 content, redox potential, Chl-*a* concentration, and turbidity.

In situ chlorophyll fluorescence analysis was utilized to non-destructively test for the presence of chlorophyll on the surface of the mineral precipitates at the springs. Chlorophyll fluorescence was determined using Pulse Amplitude Modulated (PAM) fluorometry (Schreiber, 1994, Campbell et al. 1998). A Walz Diving–PAM was used to make approximately 100 spot measurements at various locations at the springs to identify any sites where cyanobacteria might have been present.

Scanning Electron Microscopy

Samples of travertine and other mineralized surfaces that were collected for scanning electron microscopy (SEM) observations were placed in a 2.5% (vol/vol in PBS, pH 7) EM-grade gluteraldehyde solution to preserve cell morphology and structure. Upon return to the laboratory, the samples were dehydrated using an ethanol drying series and critical point drying, to further aid in the preservation of cell structure. Observations were made with a Hitachi cold field emission scanning electron microscope (S-4700) located at McGill University's Facility for Electron Microscopy Research.

Geochemistry, Petrographic analysis, and micropalentological examinations

Petrographic thin sections for optical microscopy and electron beam microprobe analysis were prepared from representative samples of travertine selected at various locations along the spring water flow path. Elemental compositions were measured by energy dispersive (EDS) and wavelength dispersive (WDS) spectrometers (using a JEOL 8900 X-ray microanalyzer at McGill University). In addition to petrographic analysis and micropalentological examinations, thin sections were sampled within and between laminations using an ultra-precision milling machine (MicroMill™ Sampler) to recover sample powders for chemical and isotopic analysis (^{13}C, ^{18}O). Analysis was performed at the Savannah River Environmental Laboratory (SREL) using a Finnigan MAT Delta Plus XL isotope ratio mass spectrometer. Bulk mineralogy was determined at NASA's Ames Research Center by X-ray diffraction (XRD). XRD patterns were obtained and mineralogical composition was determined by comparing sample diffraction patterns to known mineral standards.

Molecular phylogeny

Data on the microbial composition of the springs were obtained by applying a culture-independent approach. Inert substrates (pillows of fibra cells, Brunswick Scientific) were suspended in the stream of water exiting a spring for a period of two weeks. Samples of sediment and carbonate were also obtained for DNA extraction. Sample analysis was performed in the Pace Lab at UC Berkeley and at the Rainey Lab at Louisiana State University. DNA was extracted from the biomass buildup on the samples and used in PCR to amplify the 16S rRNA genes of the organisms using universal oligonucleotide primers. The PCR products were then cloned into the Invitrogen TA cloning vector, classed into clone types by RFLP and sequenced. Ribosomal -RNA sequences were then compared to other sequences of known organisms and comparative analysis was used to compute evolutionary relationships between the unknowns and the sequences of known organisms.

CHAPTER 2: WATER IN COLD, DRY ENVIRONMENTS: A REVIEW OF THE LITERATURE

2.1 Introduction

Within our solar system there are two worlds where landscapes are dominated by freezing temperatures. On one of these worlds, Earth, winter lasts four months of freezing darkness. Upon the other, Mars, winter has lasted three billion years or more. Despite the vast differences between the two planets today, Mars and Earth may have been much more similar early in their histories. If life developed on Mars as it did on early Earth it may have faced an untimely extinction as the cryopshere of Mars enveloped the entire planet, its atmosphere thinned and liquid water ceased to exist at the surface. To engage in a meaningful and successful search for evidence of life on Mars comparative studies of life located in the cold, dry environments of the Arctic and Antarctic are warranted. Mars may never have had global mean annual temperatures higher than what are normally experienced at Earth's polar regions today. The presence of liquid water in the Arctic and Antarctic regions, where mean annual temperatures are well below freezing, provides useful analogs of the processes that may have affected the evolutionary development of life and any subsequent burial and preservation of biomarkers or microfossils on Mars.

2.2 Water at the Earth's Polar Regions

Water is abundant at the Earth's polar regions, but in general it is not available as an exploitable resource by life because it is normally found as ice. The north and northwest arctic archipelago and much of the Antarctic are characterized by polar desert conditions with mean annual temperatures at or below -15 to -20°C (Maxwell 1982, McKay 1996). Periods of summer melt are short and the winters are long and cold. Most surface water is derived from the melting of local ice caps and glaciers during the short summers. The current glacial and periglacial environments have existed for tens of thousands to millions of years developing thick, continuous permafrost.

Permafrost describes a thermal condition of soil, bedrock, or other earth material with temperatures persisting below 0°C for more than two consecutive years (Woo 1986, IPA 1998). Nearly one fourth of Earth's land surface is underlain by permafrost. Landscape evolution is heavily influenced by the dynamic processes involved with the freezing and thawing of water in these regions and permafrost plays an important role in the occurrence, movement, storage and quality of surface water and groundwater (van Everdingen 1990). Cold climatic conditions are responsible for the formation of permafrost and large seasonal fluxes of heat in and out of the ground tend to push the permafrost to an equilibrium value determined by the local mean annual temperature. Permafrost thickness is therefore controlled by the balance between internal heat gain from the Earth's interior (the mean geothermal gradient for Earth being ~ 1°C/ 30-60 m) and the heat loss at the surface. As temperature fluctuations at the surface diffuse downward into the permafrost, their amplitudes diminish exponentially with depth. The depth at which there is no net change in temperature is known as the depth of zero annual amplitude and below this depth permafrost is unaffected by the surface thermal regime. While there are many factors that contribute to the amounts of heat that are added or removed from the ground such as vegetation, snow-cover, solar insolation etc., it is local climatic conditions (and particularly air-ground surface temperatures) set by the mean annual temperature that have the greatest effect upon the permafrost development.

In regions of continuous permafrost energy and water fluxes are strongly linked. Most scientific studies of northern hydrology have been limited to process studies including precipitation, snowmelt, evaporation, slope runoff and streamflow (Woo 1983). Mass balance and glacier runoff studies have also been carried out in the Circum-Arctic (Dowdeswell et al. 1997) including numerous studies at the White and Thompson glaciers next to the McGill Arctic Research Station (Cogley 1999, Ecclestone et al. 2000). Because permafrost acts as an aquiclude, most hydrologic activities are confined in or above the active layer (the zone just above the permafrost table subject to freezing and thawing (Woo 1986). As van Everdingen (1974) pointed out thirty years ago, there have been relatively few studies of subpermafrost groundwater in regions with thick continuous permafrost and this trend has held true to the present. Subsurface flow has been categorized into three basic classifications: suprapermafrost, intrapermafrost and subpermafrost groundwater (Williams and van Everdingen 1973). Suprapermafrost water, which flows above the permafrost and within the active layer, is the most common source for subsurface flow. Overland flow caused by saturation of the active layer is common in the High Arctic during the spring and early summer and has a major influence on landscape evolution. Intrapermafrost groundwater is found in zones of unfrozen ground or *taliks* within the permafrost. Taliks form as a result of the influence of heat sources such as lakes, rivers, perennial springs (Sloan and van Everdingen 1988) or high concentrations of salts such as saline springs or in some basal cryopegs. Subpermafrost water resides below the permafrost and in the high arctic its expression at the surface in the form of springs is rare. While thick permafrost is a significant barrier to infiltration, recharge to subpermafrost reservoirs is normally via open taliks linking the subsurface with a source of water such as lakes or basal melting of polythermal glaciers (Haldorsen and Heim 1999).

Perennial springs

Few examples of springs in high latitudes or regions of thick continuous permafrost exist. An example of deep groundwater flow from a confined aquifer in the McMurdo Dry Valleys of Antarctica is Don Juan Pond, which was discovered in 1962 (Harris and Cartwright 1981). Don Juan pond (DJP) is an intermittent brine pool located in the south fork of Wright Valley formed as result of groundwater discharge. It is small and shallow, being about 300m long and 100m wide and about 10cm deep. Its size varies from year to year depending on the amount of water it receives from groundwater and surface sources and the amount of evaporation that occurs in the summer. Extremes of groundwater input were calculated to be between 6-100 m^3 d^{-1}. It is chemically unique among surface waters in the Antarctic and is perhaps one of the more unique waters of the world. The brine consists of a concentrated solution of calcium chloride (>90%) and remains unfrozen all year, even when temperatures have dropped to –50°C near the eutectic point of the solution. The spring appears to originate from water that is in equilibrium with a sill of Ferrar Dolerite, a fractured basaltic rock constituting a confined aquifer (Harris and Cartwright 1981).

Several groups have reported the occurrence of springs in the high arctic. Spitbergen, the largest island in the Svalbard archipelago resides between 77° and 80°N where permafrost depths are estimated at 100-400 m. Roughly 60% of the island is covered in glaciers and a number of springs occur from Bockfjord in the northernmost part of the island to Sørkapp in the south (Banks et al. 1999, Lauritzen & Bottrell 1994). The Trollosen spring, a thermoglacial karst spring in South Spitsbergen has a reported discharge rate of about 18 m^3/s of a turbid, 4°C water during the summer. Its flow is thought to originate from moulins on a glacier about 5 km distant. Nearby springs discharge water that is 12-15°C and seem to have a separate hydrothermal source with a subsurface temperature in the deep aquifer of at least 30°C (Lauritzen and Bottrell 1994). The thermal springs at Bockfjord are associated with local volcanic sources, and geochemical evidence suggests that temperatures at depth reach between 130-180°C. They form travertine terraces with coatings of biofilms at the surface; however, very little work has been reported regarding the biological aspects of these springs.

Grasby et al (2003a, 2003b) have reported supraglacial sulfur springs located at 81°019'N, 81°359'W, in Borup Fiord Pass on northern Ellesmere Island in the Canadian High Arctic. Ten spring outlets were observed discharging from the surface of a 200 m thick glacier with active discharges estimated at 1 L/min at some locations and diffuse seeps elsewhere. The spring outflows are located about 500 m from the terminus of the glacier depositing native sulfur, gypsum, and calcite onto precipitate mounds on top of the ice surface. Grasby et al. (2003) measured outflow temperatures of 1-2°C and reported smelling H_2S in and around the outlets. The water chemistry of the springs is different from local melt water with higher pH and conductivity values than local melt. Light and electron microscopy and culture independent molecular methods provide evidence of a microbial community associated with the springs. Grasby et al. (2002b) also report the formation of a metastable carbonate known as vaterite, a rare hexagonal CaCO3 polymorph, that precipitates from the spring water as spheres 0.5m to 10m in diameter. They have not demonstrated that the springs are perennial and the low outflow temperatures and low solute concentration of the spring water implies that these are seasonal springs. Other glaciers in the region exhibit this same type of flow during the summer melt season and are common. However, what sets these supraglacial outflows apart from the others is the large amount of mineral precipitates that are carried up from beneath the glacier, indicating the water is flowing well below the surface and is interacting with the local geology underlying the glacier.

The only other known perennial springs at these high latitudes are those located on Axel Heiberg Island in the Canadian high arctic. Beschel (1964) first reported the occurrence of saline springs at the base of Gypsum Hill on the north side of Expedition Fiord describing their location, size, distribution, discharge rates and temperatures as well as providing an assessment of the basic chemistry. Although there were several authors that mentioned the springs in other manuscripts, they were not the focus of the papers (e.g. Schiff et al. 1991). Pollard et al. (1999) were the first to publish a detailed description of the hydrologic and geomorphic characteristics of the springs and provided additional information on a second set of springs 11km to the west at Colour Peak. Initial measurements of spring discharge, water chemistry, and descriptions of the mineral precipitates associated with the springs were among the data presented. They also provided data to show that the springs are perennial in nature and that the outflow temperatures and flow rates remain constant throughout the year. Omelon et al. (2001) reported the formation of ikaite ($CaCO3 \cdot 6H_2O$), a metastable carbonate mineral that develops at low temperature. Noting that the springs on Axel Heiberg Island flow all year with little variation in their temperature, Andersen et al. (2002) developed a thermal model to explain how the springs are able to maintain perennial flow. The ultimate source of this brine solution has been enigmatic for many years. Andersen et al. (2002) suggest that the source of the water is a combination from basal melting of the Muller ice sheet and water from glacially dammed Phantom and Astro Lakes located ~350 m above the outlet of the springs. Phantom and Astro lakes are sufficiently large to sustain through taliks and faulting beneath the lakes may also provide conduits into the subsurface. Andersen et al. (2002) suggest that water from the two lakes flows beneath the permafrost via the underlying evaporite strata and returns to the surface as a spring discharge flowing through the evaporite piercement structures. The resulting brine is warmed to the geothermal temperature within the underlying evaporite layer and loses heat to the surrounding permafrost as it flows upward to the Gypsum Hill and Colour Peak spring outlets. Heldmann et al. (2004) have investigated the large icing that results from the flow of the springs after freeze-up in the fall. The evolution of the icing occurs on an annual basis and can be divided into four distinct phases. The icing formation, icing melt, and icing washout processes are dependent upon ambient air temperatures. The icing deposit begins to form in late September once the spring water cools from its outlet temperature to the freezing point of -7°C (Pollard et al., 1999). The ice grows in lateral extent via channelized brine flow under the insulating ice cover. Mass balance calculations show that the total mass of the icing is consistent with the amount of total discharge from the spring outlets.

Perennial flow of water in the high arctic leads to the formation of a number of distinctive landforms patterned ground, palsas, pingos, seasonal frost mounds, icings and massive ground ice deposits. Frost mounds refer to a family of geomorphic features that have formed a result of or combination of (a): volumetric expansion of water during its phase change to ice, (b) the hydrostatic and hydraulic pressure of groundwater, and (c) the force of crystallization during freezing (Pollard 1988). They may be perennial or seasonal. Pingos and palsas are examples of perennial frost mounds and these can grow quite large over time (tens to hundreds of meters in diameter). Ground ice is a general term for all types of ice found in frozen ground ranging from pore ice to massive deposits of 10-20 m thick and several kilometers long (Pollard 2000, Permafrost Subcommittee, 1988). Ground ice can be divided into 2 broad categories: epigenetic and syngenetic. Epigenetic ground ice forms in situ as permafrost aggrades while syngenetic ground ice forms in combination with deposition. The latter includes various forms of buried ice – most notably buried glacier ice.

The freezing of water in the ground involves a complex set of process in which grain size, ground temperature, soil water content and chemistry, water transfer processes and rates determine the type and rate of ground ice formation. Although many types of ground ice are recognized, pore ice, wedge ice, segregated ice and buried ice are most significant in terms of volume and frequency of occurrence (Harry, 1989). Research by J.R. Mackay has established the technical basis for much our understanding about ground ice (Mackay, 1989; Mackay and Dallimore, 1992). One of the most widely used ground ice classifications (Mackay, 1972) identifies 10 types of ice and 3 primary sources of water, but excludes buried ice. Ground ice content sometimes exceeds the saturated moisture content of its host sediments, a phenomenon called excess ice. When permafrost containing excess ice thaws, the ground subsides proportional to the volume of excess ice; this is called thermokarst.

Permafrost containing massive ice is particularly thaw sensitive due to its high excess ice content.

Massive ice is defined as "a large mass of ground ice having a gravimetric ice content >250% "(Mackay, 1989). Ground ice is one of the most problematic features of permafrost and a major obstacle to development in arctic regions. Knowledge about the distribution, origin, and nature of massive ground ice and ice-rich sediments is necessary to assess potential impacts of thermokarst in response to natural and anthropogenic disturbance of permafrost. There is concern that global warming will not only cause a shift in the distribution of continuous and discontinuous permafrost but also widespread thermokarst. Ground ice studies also provide useful proxy information on arctic paleoclimates and paleogeomorphology.

2.3 Water on Mars

The arrival of Mariner 9 provided the first indications that the Martian surface had been modified by water. It unveiled a planet with deep canyons, large volcanoes, dune fields and what appeared to be fluvial features. The data collected by Mariner 9 set the stage for the Viking program, which placed two spacecraft in orbit around the planet and two landers on the surface. These four spacecraft provided continuous data for four years. The landers included measurements for biological activity, organic and inorganic chemistry, atmospheric composition, meteorological information, and surface imaging. The orbiters systematically mapped the planet at a resolution of 200 m/pixel and took a smaller number of images as high as 9 m/pixel. A good summary of the Viking results may be found in Carr (1981).

The Mariner, Viking, and most recently the Mars Global Surveyor, Odyssey, MER and Mars Express missions have provided visual and geochemical evidence that liquid water flowed on the surface of Mars at various times throughout geologic history (e.g., Carr, 1981). That Mars has had a long and complex evolutionary history is demonstrated by noting that the surface features present today could not have formed under contemporary conditions. Presently the planet appears cold and lifeless, however the overwhelming body of evidence suggests a much different past. This evidence implies that liquid water played a major role in the geomorphology of the planet. However, 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 as ground ice may exist beneath the surface. 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 lead 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. Liquid water can be maintained by the insulating properties of an ice cover (see e.g., Carr 1983, McKay et al, 1985, McKay and Davis 1991) or by geothermal activity (McKay et al. 1985), even when temperatures are below freezing.

Despite recent successful robotic missions to Mars, major questions regarding the history and distribution of water on Mars remain unanswered. How much water originally outgassed from the planet and how much remains on Mars today? Where did the water go? What were the relative roles of rainfall and snowfall and the subsequent release of water as groundwater? How warm (or cold) and wet was early Mars? Were there sudden shifts in climate or did this take place over billions of years? Does the Martian cryosphere retain a record of prebiotic chemistry or evidence of past life?

Contemporary Mars

At 1.52 AU from the Sun, the Martian surface receives about 50% of the sunlight that is incident on the surface of the Earth. Surface gravity is 0.38 of the Earth value. Lacking a significant magnetic field and a thick atmosphere, the surface of Mars is subject to higher levels of cosmic radiation, solar ultraviolet light and solar flare protons than the Earth. The contemporary atmosphere is composed predominantly of CO_2 (95.3%) with smaller percentages of N_2, Ar, O_2, H_2O, CO and noble gases (Kieffer *et al.* 1992). The Martian atmosphere is saturated throughout most of the year, however even when saturated the total amount of water is small with a global mean of around 10 precipitable microns (Carr 1996). The highest column abundance of water measured by the Viking orbiters above the northern polar cap during the summer was ~100 precipitable microns implicating the residual cap as the source of the water (Carr 1996). The atmospheric pressure measured by the Viking landers averaged about 8 mb but varied from 6-10 mb as CO_2 condensed to form the seasonal polar caps. The daily rotational period is almost identical to Earth, but because of its greater distance from the sun, the seasons are twice as long. The lack of a thick atmosphere on Mars and near absence of any greenhouse effect results in a mean surface temperature of 215° K (Meyer and McKay 1996).

One of the most striking features of Mars immediately noticeable is its asymmetry. The Martian surface is divided into two main components: the ancient cratered highlands of the southern hemisphere and the low-lying plains located mainly in the northern hemisphere. The major exceptions being the Hellas and Argyre impact basins, which extend 5 km below the datum and reside in the southern hemisphere. The cratered highlands cover two thirds of the planet's surface with average elevations 1-2 km above the datum while the northern lowlands average 1-2 km below the datum. Crater counts in the highlands indicate this region to be quite old with an age of about 3.8 Gyr (Carr 1996). In addition to the global dichotomy, there is evidence of widespread crustal bulging such as the large shield volcanoes that reside within the provinces of Tharsis and Elysium. Olympus Mons, the highest mountain in the solar system rises 27 km (about 3 times the height of Mt. Everest on Earth) above the Martian plains. To the east of the Tharsis bulge is a large crustal deformation called Valles Marineris that extends eastward for 4000 km until it merges with the chaotic terrain and large outflow channels south of the Chryse basin. Plate tectonics have been invoked by Sleep (1994) to explain the formation of the northern lowlands and recent magnetic striping patterns revealed by Mars Global Surveyor (MGS) have added confidence to this interpretation (Acuna *et al.* 1999). Mars, however, lacks the obvious tectonic features such as linear mountain chains, subduction zones etc. that are present on Earth.

The advance and retreat of the Martian polar caps had been observed by the end of the 18th century by Huygens, Cassini and Herschel. They had also determined that the axis of the planet was inclined like Earth's and deduced that Mars was subject to seasons. Coupled with the planets inclination, the elliptical orbit of Mars has a profound effect on the nature of the polar caps. Each fall a polar hood consisting of a CO_2 cloud covers the pole and CO_2 and water frost are deposited on the surface to a latitude of about 45°. In the spring, this hood dissipates, the frost sublimates back into the atmosphere and the residual cap is revealed. The summer temperatures in the northern polar region exceed the frost point for CO_2 causing it to sublime back into the atmosphere along with some water leaving only water ice at the north pole. At present, the summers are shorter and warmer in the southern hemisphere than in the north (southern spring coincides with perihelion). Summer temperatures at the south pole do not exceed 148° K and a perennial CO_2 frost covers the residual ice cap making it a year-round sink for atmospheric water. This seasonal cycle of CO_2 and the polar caps is reflected in the atmospheric pressure and when the caps are largest, the pressure is lowest (Squyres 1989).

Southern summer temperatures reach the threshold required to initiate global dust storms which in turn affects atmospheric temperatures, circulation patterns and the transport of water. It is presumed that dust from these storms is deposited onto the polar caps. The result of this cycle of freezing, ablation, and incorporation of dust has been the formation of distinctive sedimentary units known as the polar layered terrains. The layered terrains cover roughly the same area at each pole extending to about 80° latitude in the north and 70° latitude in the south. They consist of sequences of horizontal layers that are about 10 – 50 m thick. How much CO_2 is sequestered in the polar layered terrains, and the residual caps is still uncertain. Estimates by Kieffer and Zent (1992) suggest that the total amount in the residual caps is negligible and upon release would only increase atmospheric pressure by several mb. Jakosky *et al.* (1995) estimated that if CO_2 is stored within the layered polar deposits as clathrates or as solid CO_2, the deposits could contain as much as 0.85 bars. They also suggest that this could have been a repository for the CO_2 of an earlier, thicker atmosphere. Carr (1996) points out there is no supporting evidence for this and that the presence of clathrates or solid CO_2 in the polar deposits is little more than speculation at this point. There is a consensus that a significant climate record resides within the layered terrains and future *in situ* investigations and returned samples will be required to obtain this record. Zuber *et al.* (1998) using recently acquired MOLA data have estimated the amount of water residing at the Martian polar caps placing an upper limit for the current amount of water on the Martian surface at 3.2 to 4.7 x 10^6 km^3, or about 1.5 times the amount of ice covering

Greenland. If both caps are composed completely of water, the combined volumes are equivalent to a global layer of 22 to 30 m.

Evidence of Water on Mars

Geochemical evidence describing the history of water on Mars is limited, poorly constrained and not well understood (Carr 1992). However, some estimates of the total inventory have been made by modeling variables such as the D/H from the SNC meteorites and *in situ* measurements obtained by the Viking landers (Owen *et al.* 1988). These models suggest that the global inventory was either the equivalent of a few hundred meters of water at the surface at the end of the heavy bombardment, or, the equivalent of a few tens of meters of water. These and other attempts using alternative geochemical ratios have high levels of uncertainty. Geomorphological evidence for water at or near the surface of Mars is abundant and provides direct estimates based on observable features. There is a wide range of topographies but those that provide the best evidence for a fluvial origin can be grouped into two categories: outflow channels and valley networks. Carr (1991) estimated the amount of water required to form the geomorphic features to be several hundred meters. There are still many unanswered questions about the outflow channels and valley networks, especially regarding the relative roles of ice and liquid water, the processes required to initiate formation, and the climate regime in which these formative events took place.

Valley networks (Figure 2.1 a&b) consisting of narrow, quasi-dendritic drainage systems provide some of the best evidence for sustained flow of liquid water on the Martian surface (Baker *et al.* 1992). The valley networks have become the principal element of evidence for an ancient, warmer, denser atmosphere. Often these networks are incorrectly termed 'runoff channels' and can be distinguished from these because in general, the valleys lack bedforms that are direct indicators of fluid flow.

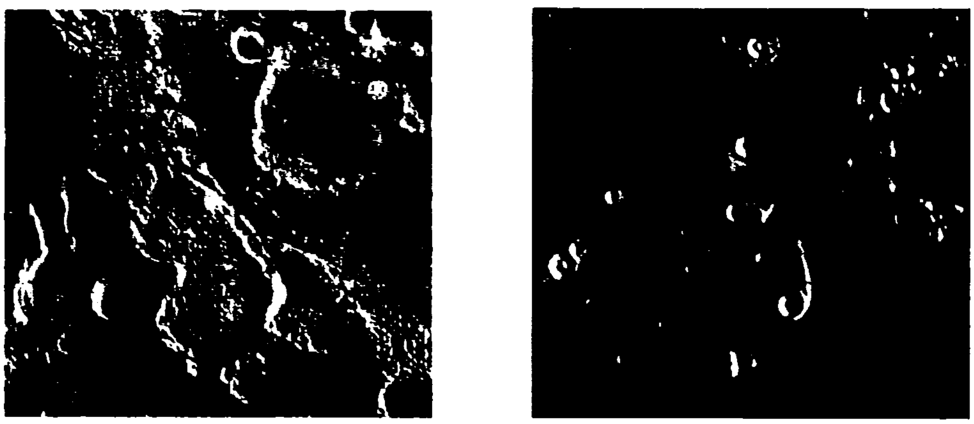
(a) **(b)**

Fig. 2.1 **(a)** Valley networks in the Warrego Valles region of Mars **(b)** Nanedi Vallis, an 800 km valley exhibiting clear evidence of past running water (NASA Photos).

The majority of the valley networks reside in Noachian terrains of the ancient cratered highlands concentrated between 65° S to 65° N latitude. Estimates by Baker *et al* (1992) place the formation age of most valley systems at 3.9 – 4.0 Gya. There are a number of examples of younger valley networks residing in Amazonian units. 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, theater-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 lava, wind etcetera, the role of groundwater sapping versus surface runoff, and climatological constraints. Erosion by running water is considered the primary mechanism for formation. Nanedi Vallis (Fig. 2.1b), 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 suggests that it was fed largely by groundwater rather than runoff.

Carr (1999) has postulated that the oldest valley systems formed as a result of precipitation and surface runoff during a warmer climatic period and towards the end of the heavy bombardment when temperatures could have been above freezing. A sharp decline in erosion rates at the end of the heavy bombardment indicates a significant climatic change took place at that time. Carr (1999) suggests that if this is true, the younger valley systems that formed near the end of the heavy bombardment may have formed under temperature regimes that were at or below the freezing point. Valley formation at this point would likely have been due to groundwater sapping. The valleys that formed in significantly younger units (i.e. Amazonian) could only occur where the thickening cryosphere was anomalously thin and heat flow was high such as the flanks of volcanoes.

Martian outflow channels (Figs. 2.2 a&b) are enormous features generally attributed to catastrophic flooding initiating from subsurface sources (Baker *et al.* 1992). They arise fully developed from spatially limited source regions and are most common along the boundary between the northern lowlands and the southern highlands. They can be over 200 km wide and over 2000 km long. Two general types occur: those in unconfined enclosures and those in confined closures.



(a) **(b)**

Fig. 2.2. (a) Outflow channels are believed to have formed by large catastrophic floods associated with the rapid drainage of ice-dammed underground reservoirs creating what has been termed 'chaotic terrain' (b) tear-drop shaped features found within the flood paths point downstream. (NASA Photos).

Unconfined outflow channels are most common around Chryse Planitia, an old highly eroded impact basin. The channels begin full-born in box canyons in the southern highlands, flowing northeast into Chryse Planitia. 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", a complex topography formed as the result of the removal of subsurface material and widespread collapse of topography. Typical bedforms include longitudinal grooves, teardrop-shaped islands 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.A. (Baker 1973, Baker and Milton 1974). On Earth approximately 18,000 years ago, Lake Missoula, 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 $\sim 10^7$ 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. These features include regional anastomosis (braiding), residual uplands streamlined by flow, flow constrictions and expansions, high width/depth ratios, low sinuosity, longitudinal grooves, inner channels and cataracts (Baker1973, Squyres 1989). The episodic catastrophic floods on Mars had discharge rates that 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 be 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, is testament to the magnitude of these voluminous discharge events. Current scientific consensus supports the notion that outflow channels result from the rapid release of large amounts of water across the Martian surface. Because of their enormous discharge volumes and concomitant latent heat energy, they could, even under current climatic conditions, flow for vast distances before being halted by freezing. 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.

The largest reservoir for water on Mars is probably in the subsurface as deep aquifers below the permafrost, and locked up in the permafrost as ground ice. Locating and sampling ground ice is a high priority activity for future missions to Mars in order to refine estimates of the total planetary water inventory and to locate resources for future human exploration efforts (Squyres 1989). Another important aspect of understanding the distribution and amounts of ground ice has to do with the search for evidence of chemical evolution or for past life. Surface soils analyzed at the Viking landing sites were found to be strongly oxidizing and these oxidizing agents may diffuse downward into the subsurface. Any organic compounds associated with prebiotic chemistry and the origin of life will be degraded or destroyed if it comes into contact with strong oxidants. At the Viking lading sites, volatile organics were not found in surface samples to the parts per billion level (for a recent review of this topic, see Benner 2000). The presence of ground ice would ameliorate this situation and aid in the preservation of early pre-biotic chemical compounds.

Clifford (1993) and Clifford and Hillel (1983) have calculated the loss of non-equilibrium subsurface ice from low latitudes. He found that his results depended on the pore structure of the regolith which is currently not well understood. Clifford's models of the Martian regolith, based on lunar analogues, suggest that it is porous and permeable to depths of at least 1-2 km. Ground ice stability will only occur if the atmospheric partial pressure of water exceeds the saturation vapor pressure for a given temperature. Under present conditions, ground ice is unstable at latitudes less than 30-40° because the mean annual temperatures are above the frost point temperature. Near surface ice in the low latitudes will sublime and diffuse into the atmosphere, the rates being dependent on the permeability of the overburden. Holding current climate conditions constant, Carr (1996) calculates that over the last 4 Gyr the dehydration front would have penetrated to a depth of several hundred meters at low latitudes and that any water that diffused into the atmosphere would then be trapped at the poles. In the higher latitudes, ground ice is stable to within a few meters of the surface and it is in these regions where the most definitive geomorphological evidence for ice-rich ground is found (Squyres 1988). Subsurface ice can alter the appearance of a planetary surface by a number of complex geomorphic processes and by observing the resulting features, the distribution of the ground ice can be inferred. Debris flows (particularly around impact craters) and terrain softening are perhaps the most common indicator of ground ice on Mars.

Rampart craters are distributed across all latitudes (Squyres *et al.* 1992) while terrain softening occurs only poleward from about 30°. Rampart craters form lobate debris aprons which have been interpreted to be a result of melting and fluidization of ground ice and sediments at the place of impact. The resulting mudflow moves out and away from the crater and then re-freezes in place. Kuzmin (1988) examined the size, depth and distribution of rampart craters and noted that in a given area, a certain critical size crater exists. Craters smaller than these sizes do not form fluidized ejecta. This is termed the onset diameter. Kuzmin et al. (1988) mapped the onset diameters as a function of location as a test of ground ice distribution predicted by thermodynamic models. There was a pronounced dependence on latitude for the distribution of onset diameters. At or near the equator the onset diameters were around 4 to 7 km while at higher latitudes they were much smaller (1-2 km). This provides an indication of the depth required to reach ground ice in the cryolithosphere. The results were in good agreement with the thermodynamic models. Terrain softening is a distinctive style of landform degradation that is observed primarily in the higher latitudes on Mars. It results in the viscous relaxation of topography, most notably crater rims. In the mid latitudes, this creep is seen over a wide depth range since temperatures are warmer. In high latitudes creep is restricted to the near surface active layer and is associated with warming during obliquity cycles. There are numerous other landforms on Mars such as thermokarst, patterned ground, rock glaciers, alases and pingos that are generally associated with periglacial environments and processes on Earth. Lucchitta (1981) and Squyres *et al.*(1992) have provided geomorphic evidence for many of these features using Viking images.

However some of the features (e.g., pingos) seen in the images are at the limits of resolution and definitive conclusions cannot be drawn.

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 result of Tharsis volcanism which triggered catastrophic flooding releasing water into the topographically lower northern plains. Gulick *et al.* (1997), assuming the Baker *et al* (1991) hypothesis was correct, explored the climatological consequences of having large pulses of CO_2 being released into the atmosphere coincident with the formation of an ocean. Their models suggested that episodic greenhouse forced climates initiated by the ocean would result in the production of fluvial valleys and glaciers. Sublimation of the ice-covered ocean could provide a means to return water to the Southern Highlands, temporarily closing the loop on the Martian hydrological cycle. These arguments have been very controversial particularly since few data are available to support arguments for or against the existence of large oceans on Mars. Interestingly, recent data from Mars Global Surveyor has added new information, but has not ended the debate. Head *et al.* (1999) have tested the original hypothesis of Parker (1989) by using high-resolution altimetric data from MOLA. They have reported their measurements to be in agreement with the observed shorelines made previously by Parker (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 the 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 supporting images 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. Given that the MGS camera has only imaged a very small per cent of the total 'ocean coastline' it may not be surprising that it has not found definitive evidence.

Geologically young, small-scale features resembling terrestrial water-carved gullies were observed by the Mars Global Surveyor (MGS) Mars Orbiter Camera (MOC) and first reported by Malin and Edgett 2000. The superposition of the gullies on geologically young surfaces such as dunes and polygons as well as the extreme scarcity of superposed impact craters indicate the relative youth of the gullies, suggesting that the gullies formed within the past few million years (Malin and Edgett, 2000). These features exhibit a characteristic morphology indicative of fluid-type erosion of the surficial material and liquid water has been suggested as a likely fluid. The gullies are reportedly found exclusively poleward of 30° latitude in both the northern and southern hemispheres. These regions correspond with the areas of ground ice stability on Mars (Clifford 1993) suggesting that gully formation may be intrinsically tied to the presence of subsurface ice in the martian polar desert environment.

Hydrothermal convection driven by magmatic activity or impact melt may have provided a mechanism for replenishing water in a martian aquifer. It is likely that springs 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 sub-Arctic 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 would form large icings and eventually would deposit the salts via freeze fractionation and/or evaporation. The Thermal Emission Spectrometer (TES) instrument on the Mars Global Surveyor (MGS) mission discovered an accumulation of crystalline hematite in a sedimentary rock formation that covers an area approximately 350 by 350-750 km in Sinus Meridiani and this is now the focus of exploration by the MER rover Opportunity. Initial results indicate that the hematite formed in a shallow, salty body of water (Kerr 2004). Given the nature of the site, it would not be unexpected to find other mineral precipitates as well and sites such as this will most assuredly be targeted for exploration as part of the search for evidence of past life.

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 grain 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 melt water entered into the lake to offset ablation of ice at the surface, the lakes could exist for several hundred million years or more. McKay and Davis (1991) employed calculations that explain the 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 and Gale may have provided an oasis for life from the Noachian-Hesparian boundary (2.5-3.8 Gyr BP) and lasting 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 represents a fraction of the total number of lakes in impact craters. With new high resolution images being returned by MGS, Mars Odyssey and now the MER surface rovers, the case for past standing water bodies on Mars is building although the case for a large lake within Gusev has not been demonstrated (Bell et al. 2004a, Squyres et al. 2004a). At the time of this writing, Spirit had found no clear evidence for lacustrine sedimentation, the dominant lithology being basalt, with the dominant geologic processes being impact events and eolian transport (Squyres et al. 2004a). NASA's Mars Exploration Rover Opportunity, which landed at Meridiani Planum, has found very compelling evidence for aqueous sedimentary and diagenetic processes that operated for long periods of time and over an area at least tens of thousands of square kilometers in size (Squyres et al. 2004b-c). The evidence includes cross-bedded sedimentary outcrops that are highly enriched with sulfur and sulfate salts (such as the ferric sulfate mineral jarosite [(K,Na,H3O,X+1)Fe3(SO4)2(OH)6]), and hematitic spherules (concretions) thought to have formed in an aqueous environment (Bell et al. 2004b, Squyres et al. 2004b-c).

Implications for the origin and Evolution of Life

The earliest evidence of life on Earth to date is preserved not as morphological fossils but as chemical signatures within sedimentary deposits found in the Isua Formation in Greenland. Characteristic isotopic shifts ($^{12}C/^{13}C$) associated with photoautotrophy have been measured within these sediments and provide an indicator for the existence of life as far back as 3.7 to 3.85 Gyr ago (Schidlowski *et al.* 1983; Mojziz *et al.* 1996). Awramik (1983) and Schopf and Packer (1987) reported definitive fossil evidence of life dating 3.5 Gyr ago within the Warrawoona Group formation in northwestern Australia. Pace (1997) conducted molecular phylogeny studies indicating that the last common ancestors on Earth were comprised of anaerobic hyperthermophiles. During the time that life had established itself on Earth, liquid water was present at or near the surface of Mars. The warmer and wetter period that seems to have occurred on Mars may have provided the requisite conditions conducive to the origin of life there. High rates of volcanism and abundant groundwater supplying chemically rich hydrothermal springs were probably common thus providing habitats similar to those found on early Earth (McKay 1991). If life evolved on early Mars it may have developed metabolic strategies similar to those used by the earliest life forms on Earth. Martian life would have had an ample supply of CO_2, S, and H_2 to be used in metabolic reactions as do many of the terrestrial chemolithoautotrophs. Hydrothermal sites on early Mars may also have been conducive to the preservation of microbial fossils in siliceous cherts or travertines associated with the mineralized waters of the springs.

Ice-covered lakes have also been proposed by Wharton *et al.* 1993, Wharton *et al.* 1995, and McKay and Davis. 1991 as suitable habitats that may have formed at various times early in Mars' history. 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 Gya. 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 be on Mars. A number of the large canyons found in Valles Marineris may have flooded with groundwater forming large ice-covered lakes. 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 exist for several hundred million years or more. If water was episodically present on Mars, it may have been possible that life survived to more recent times. Cabrol *et al.* (1998 and 1999) have argued that impact craters such as Gusev and Gale may have provided an oasis for life from the Noachian-Hesparian boundary and lasting for up to 2 Gya. Using high-resolution Viking images to located lacustrine features, they propose that Gale experienced a number of aqueous environments that transitioned from earlier warmer, wetter to cold and ice-covered waters. If this is the case, they suggest that the sedimentary record in these environments would be important to investigate for evidence of past life. Extant life on Mars may exist, but in all likelihood it will be difficult to locate. The surface, being dry, cold, and oxidizing, is probably not the place to look. However, the deep subsurface, at depths where geothermal heating is sufficient to maintain liquid water may be. A report by Stevens and McKinley (1995) of microbial communities in terrestrial subsurface basalts is especially exciting since they would represent a similar type of deep subsurface community on Earth.

At present, the most complling site for the search for evidence of past life may be the sediments located at Meridani Planum. If this region was a salty, acidic sea as suggested by Squyres et al. (2004c) then a record of life may be preseverved there.

PREFACE TO CHAPTER 3

Life is intrinsically linked to its physical environment and in regions where extreme conditions prevail this linkage becomes more pronounced. This chapter explores the thermal regime associated with the springs at Expedition Fiord, on Axel Heiberg Island. The existence of the springs in this region of thick, continuous permafrost is enigmatic at first glance. However, using observations of the spring flow rates and discharge temperatures in conjunction with the local geothermal gradient, a model can be constructed to describe how such springs can flow through such a thick and seemingly impenetrable barrier. A more thorough understanding of how water circulates under such strong climatic constraints will undoubtedly lead to a better appreciation for how microbial communities have adapted to a wide range of environmental conditions on Earth and possibly on Mars.

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CHAPTER 3: COLD SPRINGS IN PERMAFROST ON EARTH AND MARS

3.1 Introduction

Recent Mars Global Surveyor images indicate the presence of gully-like landforms that occur primarily at high latitudes providing evidence of recent fluvial activity (Malin and Edgett, 2000). These features appear to be geologically young enough that they would have formed under the present climatic conditions that include mean surface temperatures of -60°C and extensive permafrost. The absence of any association between these flow features and obvious geothermal heat sources (e.g. volcanic features) is problematic. Eutectic brines present in the shallow martian subsurface have been suggested as the likely fluid that formed these features (Hoffman et al. 2000). We believe cold perennial springs that flow through thick continuous permafrost on Earth may provide an analogue for these martian features.

On Earth, the presence of thick permafrost creates an effective aquitard preventing groundwater discharge and resulting in the separation of groundwater into sub-, intra-, and supra-permafrost systems (Williams and van Everdingen, 1973). However, there are two sets of perennial springs on Axel Heiberg Island that flow through 600 m of permafrost and are not associated with any volcanic heat sources. First reported by Beschel (1963) and more recently described in detail by Pollard *et al.* (1999), the springs are located at 79°26' N, 90°46' W (Figure 1). At nearly 80° N these are among the most pole-ward springs known and are currently the only known example of cold springs in thick permafrost on Earth.

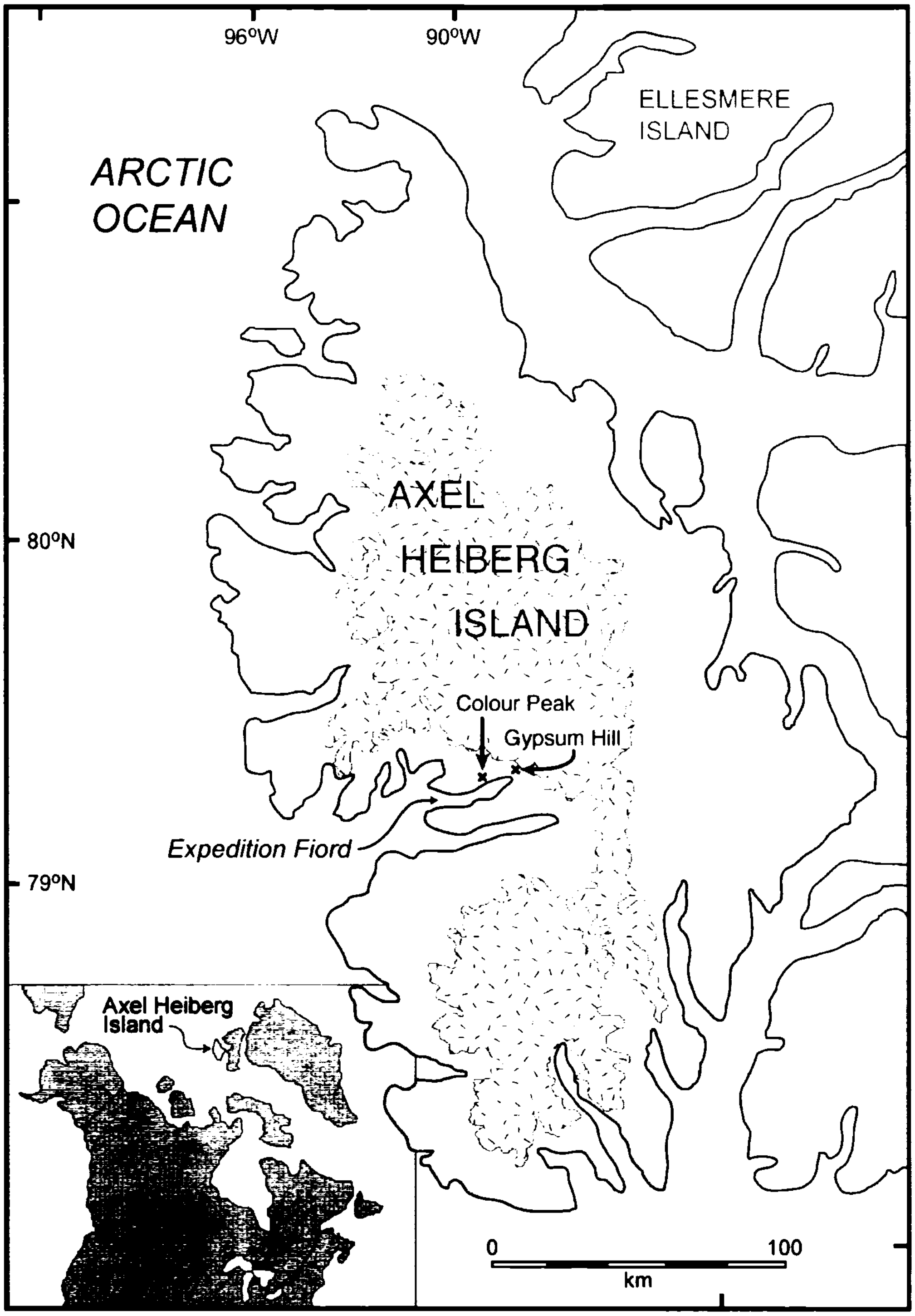


Figure 3.1. Location map of Expedition Fiord, Axel Heiberg Island, Nunavut.

Axel Heiberg Island is mostly bare ground with less than 35% covered by glaciers or ice caps and lies within the Sverdrup Basin Archipelago, a large intracratonic rift basin initiated during the late Paleozoic (Thorsteinsson and Tozier 1970). The local geology is complex and consists of folded and faulted sedimentary rocks ranging from Triassic to Tertiary in age. Upper Paleozoic evaporites locally intrude the overlying sedimentary clastic rocks in a series of piercement structures. Both groups of springs at Expedition Fiord occur in geologically similar settings 11 km apart and are associated with anhydrite piercement structures (Figure 2) (Hoen 1964).

Figure 3.2. Sub-glacial flow and Phantom Lake , a glacially dammed lake 397 m asl, are possible sources of the water flowing from the springs at Gypsum Hill and Colour Peak. The lake is dammed on one side by the Thompson Glacier and is situated next to an evaporite piercement structure. Faulting occurs beneath the lake through the piercement structure that may provide a conduit to the deep sub-surface as well as provide a connecting link between the other evaporite structures associated with the springs.

Thick, continuous permafrost reaching depths of nearly 600 m has been reported for this region (Taylor and Judge, 1976). The mean annual air temperature at the study site is approximately –15° C (Doran et al. 1996). The low temperature coupled with a potential evaporation that exceeds the low annual precipitation produces a region that is considered to be a polar desert. Perennial springs at both study sites discharge a brine with measured discharge temperatures that range from -2° to +6.5° C. At both sites, our results show that the main spring-flows maintain a 5-6° C mean discharge temperature throughout the year despite air temperatures that fall below -40° C during the winter months (Doran et al. 1996, Pollard et al. 1999). NBS thermometers housed in Stevenson Screens recorded a minimum temperature of –55° C during the 2000-2001 winter. Over the course of a year, the mean spring temperature at Colour Peak is 5.2° C, the maximum temperature is 5.5° C, the minimum temperature is 4.9° C, with the standard deviation being 0.2° C. Flow rates are also constant with little variation (Figure 3). The brines exhibit high concentrations of Na, Cl, Ca and SO_4 ions with total salt content > 10% (Beschel, 1963; Pollard et al. 1999). As pointed out by Fricker (1963) the high level of NaCl may have resulted from the contact between salts at depth and the emergent groundwater.

The Gypsum Hill site consists of approximately forty springs and seeps on the north side of Expedition River discharging along a band nearly 300 m long and 30 m wide and between 10-20 m ASL. The Colour Peak springs are located on the south-facing slope of Colour Peak at an approximate elevation of 30-40 m ASL. These springs are grouped into 3 distinct topographically controlled areas with 20 vents discharging directly into Expedition Fiord several hundred meters down slope. Unlike the springs at Gypsum Hill, the Colour Peak springs form distinct troughs, pipes, and sheet flows with well-developed travertine and other mineral precipitates (Pollard et al. 1999). Isotopic data suggest relatively recent but pre-bomb (prior to nuclear tests in the 1950's and 60's) meteoric water as the source for these springs (Pollard et al. 1999).

3.2 Thermal Model

The thermal and flow properties of the springs are remarkably constant given the extreme seasonal variations in air temperature (Figure 3.3).

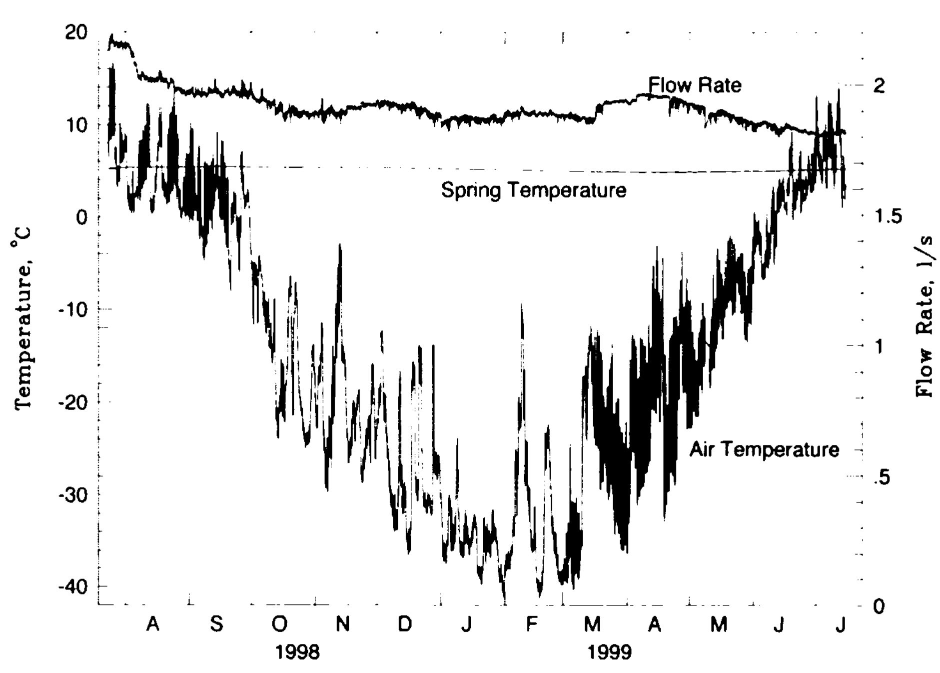


Figure 3.3 Spring discharge flow rate and outflow temperatures throughout the year

There is no evidence of recent magmatic geothermal activity in the area and we hypothesize that the heat content of the water in the springs derives from the geothermal temperature profile. We have developed a combined flow and thermal model of the subsurface flow similar to the approach of Demming et al. (1992) and Haldoresen and Heim (1999). We use the measured geothermal gradient of 37.3° C/km for a nearby borehole (60 km) that was reported by Taylor and Judge (1976). A schematic of the flow model is shown in Figures 3.4 & 3.5.

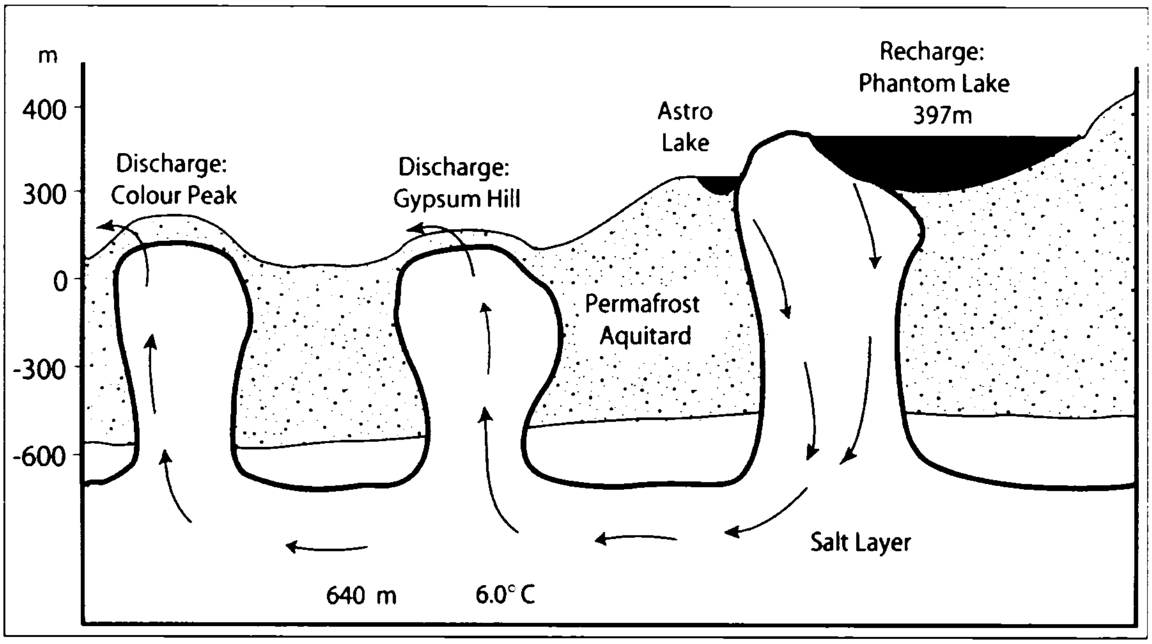


Figure 3.4. Schematic of spring system located at Expedition Fiord, Axel Heiberg (following the pattern of Haldorsen and Heim 1999).

The piercement structures and their associated evaporite units are the key to our hydrological model. The deep subsurface salt layer contains the main reservoir of water.

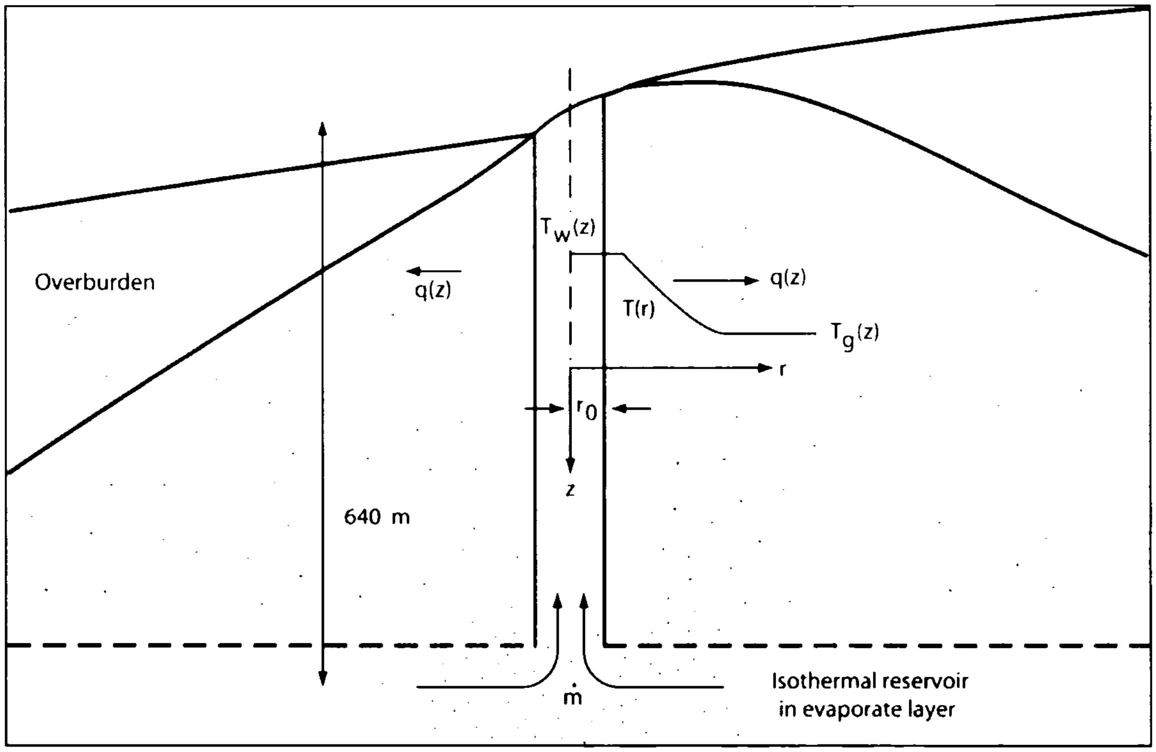


Figure 3.5 Schematic of the mathematical representation of the thermal balance of the upwelling spring water. Note that the r coordinate represents distance from the center of the cylinder while the z coordinate represents depth below the surface. T_w, T_r, and T_g are the temperature of the water in the cylinder, temperature at distance r from the cylinder, and the surrounding ground temperature (unaffected by the spring), respectively. The q(z) term indicates heat leaving the spring conduit.

Water enters and leaves this reservoir by way of the piercement structures that reach the surface. The major source of water entering the salt layer could be sub-glacial melt or water from Phantom Lake, which is situated approximately 397 m above the outlet of the springs. The basin of this glacially dammed, 200 deep, perennially ice-covered alpine lake is located within a structurally controlled valley along a fault line. The lake formed following the retreat of the Transit Glacier. Phantom Lake periodically drains a small portion of the total volume (1-2%) during summer months by overflowing a riegel located at the southern border of the lake and into a marginal channel that abuts one of the a piercement structures as mapped by

Hoen (1964). For both possible sources we suggest that water flows deep below the surface via the evaporite unit returning to the surface through the piercement structures associated with the springs. Below the surface, the water reaches the geothermal temperature and as it flows upward it loses heat to the surrounding permafrost.

If we consider time averaged conditions then the heat loss from the water to the ground as it moves upward at a depth z is given by

$$\dot{m}c\frac{dT_w(z)}{dz} = q(z) \qquad (1)$$

where q(z) is the rate of heat loss from the water to the ground as it moves up at a depth z, $T_w(z)$ is the temperature of the water in the spring as it flows upward, $\dot{m}$ is the mass flow rate of the springs in kg/s, and c is the specific heat of water in units of J/kg/°C.

To proceed further in this analysis we express q(z) as the heat loss from a uniform cylinder as shown in Figure 3.5. If we consider the radial coordinate, r, then the heat flowing out from the cylinder at depth z is given by

$$q(z) = -2\pi r\; k\frac{dT_r}{dr} \qquad (2)$$

where k is the thermal conductivity of the ground surrounding the cylinder which has an assumed azimuthly symmetric temperature profile T_r. At the boundary of the cylinder. $T_r = T_w$ and very far away from the cylinder $T_r = T_g$, where T_g is the geothermal temperature profile distant from the springs, $T_g(z) = T_o + \alpha z$, where T_o is the annual average surface temperature and α is the geothermal gradient.

We use these equations as the basis for a numerical model of the thermal balance of the spring water. Combining these equations and solving the resulting inhomogeneous differential equation gives the following expression for the mean temperature of the water in the springs

$$T_w(z) = T_o - \alpha\gamma e^{(z-z_b)/\gamma} + \alpha(z+\gamma) \qquad (3)$$

where $\gamma = c\dot{m}/\pi r k$ is a characteristic scale length for the spring flow and represents the distance that the flow moves before a significant loss of thermal energy by conduction occurs. Note that faster flow rates make γ larger while larger thermal conductivity coefficients make γ smaller. The values that must be specified to complete the calculation are the mean surface temperature, the geothermal gradient, and the depth (or temperature) of the salt layer. Given these parameters the mean exit temperature of the springs can be determined from Eq. 3 as a function of γ. This is shown in Figure 3.6 for Earth (Axel Heiberg) and Mars. For Axel Heiberg the mean temperature of the surface, the geothermal gradient, and the mean temperature of the underlying salt layer are −15°C, 37.3°C/km, and +6°C (640 meters depth), respectively. For Mars the corresponding values used are −60°C, 20°C/km, and 0°C. Note that the values used for Mars imply a 3 km thick permafrost and a source of water just below that permafrost (Clifford, 1993).

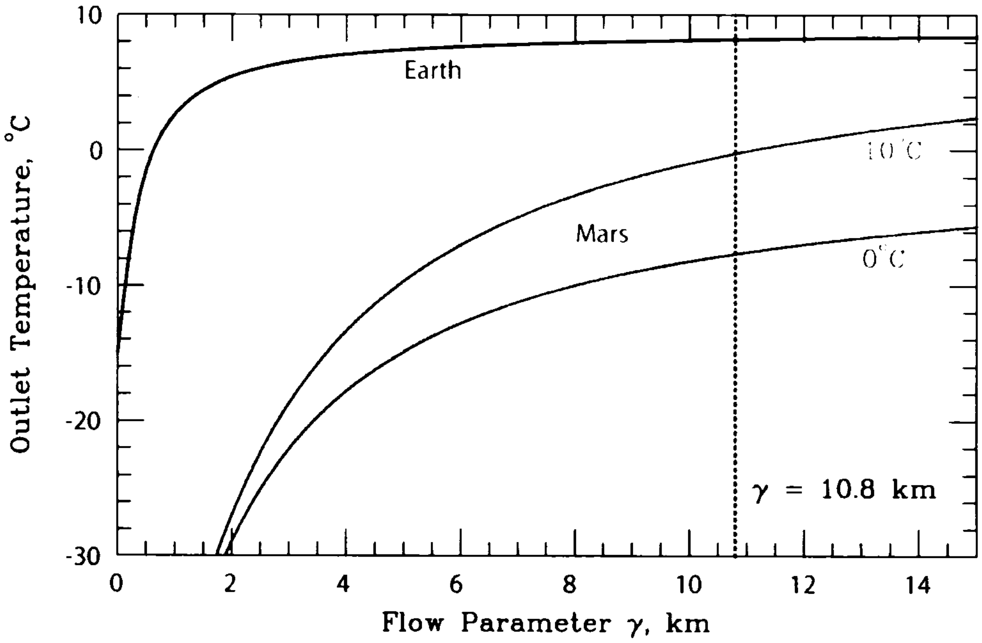


Figure 3.6. Outflow temperatures as a function of the flow parameter γFigure 3.6. Outflow temperatures as a function of the flow parameter γ where γ (= c $\dot{m}$ / Δ r k) is a characteristic scale length for the spring flow. The flow parameter γ represents the distance that a solution may travel through the permafrost before a significant loss of thermal energy by conduction occurs. Faster flow rates make γ larger while larger thermal conductivity coefficients make γ smaller. Boundary conditions for Earth are for the Axel Heiberg site: -15°C annual mean surface temperature, 37.3°C/km geothermal gradient, 640 m depth to source of springs in the subsurface salt layer (temperature = +6°C). For the springs at Axel Heiberg, the nominal value for γ=10.8 km, which implies an outlet temperature of ~5°C for Earth. Outlet temperatures for springs on Mars will vary depending on the boundary conditions. Here we show two curves for Mars with the boundary conditions: -60°C mean surface temperature, 20°C/km geothermal gradient and for water originating from just below the permafrost (3 km depth) at a temperature of 0° C and for water originating at 500 m below the permafrost (3.5 km depth) at a starting temperature of 10° C. Flows originating just below the permafrost have an outlet temperature of ~-5°C - brine solutions could be liquid at this temperature. Outflow temperatures above freezing are possible for flows originating from deeper locations.

The essential measurements to be compared to the model are the flow rate and the observed temperature of the discharge. Taylor (1991) determined the thermal conductivity (k =2.0 W m^{-1} K^{-1}) at the Gemini borehole and we use this value as a best estimate for the thermal conductivity at the spring sites located at Expedition fiord. Similarly, a value of $k = 2.0$ W m^{-1} K^{-1} is used for Mars using the estimates of Clifford (1993). The specific heat ($c = 3400$ J kg^{-1} K^{-1}, specific heat = 0.8) of the out-flowing was estimated from values published by Archer and Carter (2000) for NaCl solutions at low temperatures. For the mass flow rate reported by Pollard *et al.* (1999) of ~1 kg/s, and for a spring radius of ~0.5 cm, we thus obtain a nominal value of γ for the Arctic springs of 10.8 km. This in turn implies an outlet temperature of close to 5°C, which is in agreement with our observations. The large value of γ compared to the distance to the bottom of the spring justifies the neglect of seasonal variations in the analysis – the water is flowing quickly enough to be unaffected by local variations in the environment.

For Mars the model provides a quantitative way to understand how cold springs can penetrate thick permafrost. If the Martian springs were no bigger than the Arctic analogues then similar flow rates (γ = 10.8 km) would result in outlet temperatures between –5° C and –10° C, for solutions originating at 0° C just at the base of a 3 km thick permafrost. If the solution originated at a deeper, warmer, location then the exit temperature would be increased by a corresponding amount. To achieve exit temperatures above 0°C on Mars the solution would have to originate at a depth of 0.5 to 1 km below the permafrost.

The Arctic springs are maintained by recharge from glacial melt water produced during the summer months in the mountains above the springs. If this source were to end the flow would cease on timescales that we estimate to be a few hundred years based on the volume of the lake and the rate of discharge of the springs.

3.3 Springs on Mars

The spring features observed by Malin and Edgett (2000) could represent cold springs from distant glacial melting with the water flowing underground through salt layers as in the Arctic. Salts are known to be present on Mars and the existence of large accumulations of evaporite deposits is a logical possibility (Forsythe and Zimbeman, 1995). The springs could have formed during a recent period when due to obliquity changes glacial deposits were present in the southern highlands. Alternatively, the spring discharges may have resulted from relic groundwater stored in confined aquifers forced out by an expanding freezing front from a much earlier epoch (Mellon and Phillips 2001). In this case the Arctic model provides a mechanism for bringing the water to the surface but does not explain what prompted the release of the water after a long period of subsurface storage.

If the springs observed by Malin and Edgett (2000) are associated with flow through salt conduits as we observe in the Arctic, then we would expect these flows to leave behind large salt pans as the water froze and evaporated into the dry Martian air. Unfortunately, these saltpans may not be observable if there is a thin layer of airborne dust covering them.

As in the Arctic, cold springs emanating from thick permafrost could be sites for microbial life in an otherwise harsh environment. Equally important, the salts and other mineral deposits associated with these springs could preserve isotopic, molecular and morphological evidence of any life that may have been present.

3.4 Conclusions

We conclude that evidence of recent spring activity on Mars by Malin and Edgett (2000) can be understood as analogues to similar systems here on Earth which demonstrate that liquid water is capable of reaching the surface in regions of thick, continuous permafrost without strong volcanic heating sources. Mineral deposits associated with such springs in the Arctic contain biological signatures. They might on Mars as well.

3.5 Methods

The solution begins with the integration of $\dot{m}c[dT_w(z)/dz] = q(z)/r$ to give $q(z) = r_o\pi k[(T_w(z) - T_g(z)]$. Combining this with $T_g(z)=T_o+\alpha z$ gives $\dot{m}c[dT_w(z)/dz] = r_o\pi k[T_w(z) - T_o - \alpha z]$ which can be solved for the homogeneous term $T_w(z) = Ae^{(z/\gamma)}$ and the inhomogeneous term $T_w(z) = \alpha(z + \gamma)$. To determine the constant A we note that the temperature of the water in the salt layer at depth, z_b, is the same as the surrounding ground, i.e. $T_w(z_b) = T_g(z_b) = T_o + \alpha z_b$. Combining these gives our final result $T_w(z) = T_o - \alpha\gamma e^{(z-z_b)/\gamma} + \alpha(z+\gamma)$.

The seasonal dependence can be included explicitly using the same approximate solution for $q(z,t)$. In this case the ground temperature T_g is expressed as $T_g = T_o + \alpha z + A_o \sin(\omega t - z/D) e^{-z/D}$ where A_o represents the amplitude of the surface temperature fluctuation and the sin term accounts for this seasonal ground temperature oscillation, where ω is the annual frequency and D is the penetration depth of the annual thermal wave given by $D = (2k/\omega\rho C)^{1/2}$, ρ is the soil density, k the thermal conductivity, and C the specific heat per unit mass (Campbell 1997). Following this approach the solution yields an additional seasonally varying term which scales as $A_o D/2\gamma$. For our nominal values of γ (10.8 km) and D (3 m) this term is negligible, predicting a seasonal spring water temperature variation of 0.003°C.

PREFACE TO CHAPTER 4

On Axel Heiberg Island in the Canadian High Arctic there are two sets of perennial saline springs that flow through 600 m of permafrost. In the previous chapter it was shown that the springs are able to maintain flow through the permafrost without the need for a volcanic heating source. This chapter links the flow of the water through the permafrost with the origin of the water itself as determined by measurements of the dissolved gases. In this chapter we report on measurements of the gas composition of bubbles that are released from these springs. The gas is composed primarily of N_2 with relative concentrations of Ar, Kr, and Xe almost identical to air. No O_2 is detectable and Ne is 60% of air values. Helium is present at about 0.5% relative to N_2 and low levels of CO_2 and CH_4 are present. We believe that ~50% of this gas originates from the direct release of air by nearby alpine glaciers and local ice sheets into groundwater that infiltrates sub-ice sedimentary deposits. Below the surface O_2 is consumed by the oxidation of organic matter and other redox reactions and He, derived from subsurface radioactivity, accumulates. The elevated levels of He are consistent with a long residence time of the water below the surface while the low Ne level reflects the solubility in ice for this gas.

Submitted to JGR Planets, 2004.

CHAPTER 4: COMPOSITION AND SOURCE OF GAS BUBBLING FROM A SALINE HIGH ARCTIC SPRING

4.1 Introduction

Current global climate models indicate that global warming will be most acute in polar regions and that the Arctic will be particularly vulnerable (Moritz et al. 2002). Predicting and understanding the effects of climate change in these high latitude ecosystems will be increasingly important as we try to understand the impact of any future climate change across the planet. The trapping and isolation of gas bubbles in glacial ice has provided a source of ancient atmospheric gases that records information regarding past climatic conditions. The processes by which air is trapped and bubbles are formed by the sintering of dry firn are complex and highly dependent on local climate. If surface melting of the glacier occurs during summer the resulting meltwater formation and refreezing will alter the gas composition (Stauffer et al. 1985). While there is a vast supply of glacially trapped gas within the polar ice caps, logistically these repositories are difficult and expensive to utilize. Gas derived from glaciers and ice caps emanating from springs may provide an alternative source. Two sets of perennial springs are located at Expedition Fiord at 79°26′ N on Axel Heiberg Island in the Canadian Arctic Archipelago (Figure 4.1). Perennial springs are rare in high latitudes and presently the only reported springs such as these are those found at Spitzbergen (Haldorsen and Heim 1999) and at Axel Heiberg (Pollard et al. 1999). The Axel Heiberg springs emerge ~50m asl from the base of gypsum-anhydrite diapirs that formed as a result of the Eurekan orgeny during the Tertiary (Pollard et al. 1999). This region is heavily glaciated with thick continuous

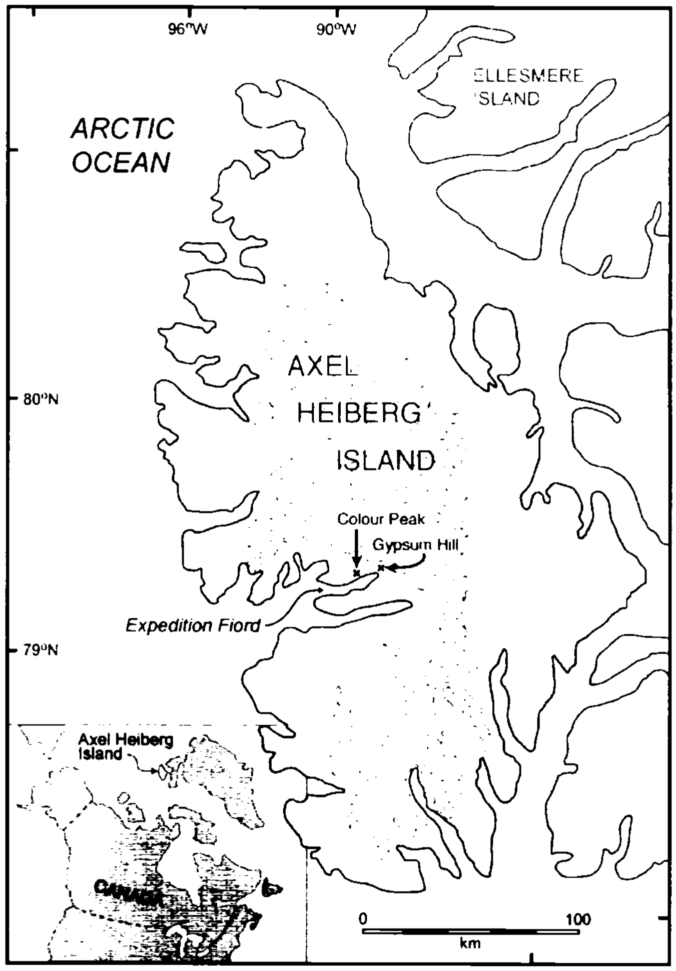


Figure 4.1 Location map of Expedition Fiord, Axel Heiberg Island, Nunavut Canada.

permafrost reaching depths of 600 m or more and a mean annual temperature of –15°C (Doran et al. 1996). In this study we analyze the gas bubbling from the spring outflows and present a model that explains the origin and evolution of the constituent gases.

The presence of thick permafrost in the Arctic Archipelago makes if difficult for groundwater to reach the surface or for surface waters to recharge deep aquifers (Woo et al. 2000). While the formative processes that resulted in the springs at Axel Heiberg are not well understood, Haldorsen and Heim (1999) have suggested that the springs located at Spitzbergen may have formed during the end of the last glaciation when sea level was higher and the present day spring outlets were submerged. A similar situation may have occurred at the Expedition Fiord springs during a time when the marine limit was higher. Andersen et al. (2002) developed a combined flow and thermal model of the subsurface flow associated with the springs at Expedition Fiord using the locally measured geothermal gradient. Their data show the thermal and flow properties of the springs remain remarkably constant given the harsh climate and extreme seasonal variations in air temperature and they noted that the thermal properties of the springs are derived from the local geothermal gradient. In addition, they postulated that the groundwater is derived from local glacially dammed alpine lakes and/or subglacial discharge infiltrating the sedimentary units associated with the numerous piercement structures prevalent in the area. Phantom Lake and Astro Lake were considered to be the likely sources due to their large volumes, proximity to the diapirs and multi-year ice-covers (possibly perennial). As pointed out by Boulton et al. (1994) and Haldorsen and Heim (1999), subglacial melt may provide a means for the deep circulation of groundwater beneath ice sheets and glaciers. Local to the springs at Axel Heiberg is the White Glacier, a multithermal alpine glacier with permafrost at the lobes and temperate bottom temperatures occurring above the equilibrium line (Blatter 1987). The geology of the Expedition Fiord region is characterized by complex folding and faulting of sedimentary rocks and given that much of the sedimentary units mapped appear to extend beneath the glaciers and ice sheets it is plausible that there is recharge via the porous subglacial strata.

4.2 Methods

Gas samples collected for this study were obtained at the base of Gypsum Hill where approximately 40 springs and seeps extend over an area of 900 m^2. One pool type outlet known as "little black pond" (LBP) was the primary source for samples in this study. The pool is approximately 1 meter in diameter and is filled with a black course-grained sediment through which the spring flows. The 6°C water temperature and a flow rate of ~ 200 ml/sec at this outflow have remained constant over the last 5 years. Gas bubbles emanating from the outlet provide an easy source for sampling the evolved gases. To obtain samples of the bubbles, 60 cc borosilicate serum vials were submerged in the spring. Using a funnel to guide bubbles into the serum vials, bubbles were collected by displacing the water inside the vials. After approximately twice the volume of the vial had been purged the vials were stoppered and crimped while underwater and returned to the laboratory for analysis. *In situ* dissolved O_2 measurements were obtained in Phantom and Astro Lakes using a YSI 95 dissolved O_2 meter and with an Ocean Seven 316 multi-parameter probe (Idronaut, Italy). Water samples from Astro Lake were obtained with a Kemmerer bottle lowered into the water column. Once retrieved, the water was allowed to gently flow from the Kemmerer bottle via polyethylene tubing into 125cc serum vials and subsequently sealed after overflowing at least two volumes of water.

Samples were analyzed for noble gases and tritium at the LLNL following methods outlined by Clark et al. (1998). Measurement uncertainties are 2%. N_2/Ar measurements by LLNL followed methods described by Kana et al. (1994) with a measurement uncertainty of 1%. Noble gases, N_2, O_2, CO_2, CH_4 and H_2S were analyzed at NASA Ames Research Center using a SRS model RGA-200 quadrapole mass spectrometer and a HP. For gas analysis the sample was introduced into the vacuum system connected to the spectrometer through a leak valve. Air was used as the primary standard gas for concentration and ratio measurements and care was taken to ensure that the ion current for the air standard was similar to that for the measurement gas. For water samples a headspace of 5 ml was first produced, using He as described by Risgaard-Petersen and Rysgaard (1995) and Blicher-Mathiesen et al. (1998). For the measurement of He in the spring water the headspace was produced with air. Error is: 10%. There are several sources of error in the models: For the solubility of neon in ice we have assumed that the solubility is simply equal to the solubility of neon in water. However alternative measurements suggest it may be 0.9 the solubility in water (Hood et al. 1998, Top et al. 1987). For the equilibrium between gas and water at LBP we have assumed that the bubbles at LBP are in equilibrium with the brine. This is plausible given the loose sandy mixture at the bottom of LBP which provides surface area for interactions that should help establish equilibrium. N_2 and Ar measurements in the water from LBP are consistent with this equilibrium at the 20% level, but more precise measurements are necessary to show the exact degree of equilibrium. If equilibrium is not reached then there will be an addition effect on the gas ratios due to the rates of diffusion associated with each gas.

One might expect that heavier gases such as Xe would diffuse more slowly. For the salting out effect we have assumed that the reduction in solubility due to salt is independent of gas species. For glacial air we have assumed that the gas in the ice bubbles is identical to air. This will however, depend on the nature of the glacial surface and climate. If there is a melting period in the summer then this could alter the composition of gases trapped in the firn ice.

4.3 Results and Discussion

Gas concentrations in the bubbles relative to N_2 are presented in Table 4.1. The gas in the bubbles is composed primarily of N_2 with relative concentrations of Ar, Kr, and Xe almost identical to air. O_2 is absent and ^4He is enriched by a factor of about 100. Neon is depleted and CO_2 and CH_4 are present at low levels. Measurements shown in Table 4.1 suggest that the N_2/Ar ratio in the bubbles is similar to air and the N_2/Ar ratio in the brine is as expected for equilibrium between the bubbles and the liquid.

The ratios of noble gases measured from the springs are somewhat puzzling given that the relative concentrations of Ar, Kr and Xe are identical to air. The high ^4He content is most likely a result of the radioactive decay of uranium and thorium and is consistent with a long residence time for the groundwater on the order of 10,000 years (Castro et al. 2000). Oxygen loss does not appear to be solely due to carbon consumption or a higher level of CO_2 would be produced which is not observed. It likely that the O_2 is consumed by the oxidation of reduced species such as sulfide that subsequently form sulfates and remain in solution.

Table 4.1. Summary of data for dry gas from LBP bubbles and gas dissolved in the water.

Quantity	Springs	Air	Dissolved in 0° C Water	Note
N_2	98.9±1 %	78%		1,2
O_2/N_2	<0.001	0.27	0.56	1
Ar/N_2 in gas	0.0112±0.002	0.0120		1,2 (±0.0001)
Ar/N_2 in water	0.025±0.006		0.027	1
$^{36}Ar/^{40}Ar$	$3.1\pm0.3\times10^{-3}$	3.38×10^{-3}	3.38×10^{-3}	1
Ne/Ar	$1.21\pm0.01\times10^{-3}$	1.95×10^{-3}	0.47×10^{-3}	2
Kr/Ar	$1.23\pm0.01\times10^{-4}$	1.22×10^{-4}	2.42×10^{-4}	2
Xe/Ar	$9.47\pm0.09\times10^{-5}$	9.31×10^{-5}	38.4×10^{-5}	2
He	540±6 ppm	5.2 ppm		1,2,3
CO_2	350±15 ppm	350 ppm		3
CH_4	0.26±0.02%			3
^3H	<2.1±10% T.U.			4

1. NASA ARC MIMS analysis
2. LLNL analyses
3. NASA ARC GCMS
4. Data from Pollard *et al.* 1999

Clearly the gas in the springs is derived from air. There are two possible sources for this gas. The gas could be derived from air dissolved in surface meltwater or lake water that then flows into the subsurface via the salt layers. Examples of possible sources of lake water are Phantom Lake and Astro Lake. Water from these lakes enters the subsurface via the piercement structures described by Pollard et al. (1999) as shown in Figure 4.2a. Both lakes have water temperatures of ~0°C and *in situ* measurements of the dissolved O_2 in Phantom and Astro show that both lakes are well ventilated with dissolved gases throughout the water column in air equilibrium. N_2

and Ar measurements in Astro confirm that these gases are also in air equilibrium. Astro Lake is in direct contact with gypsum strata and the water column below 20m has salinity values at or near saturation with gypsum (0.2 ppt).

An alternate source is bubbles trapped in glacial ice that is discharged into the subsurface during basal melting. If the basal meltwater is isolated from the atmosphere the composition of the gas will be maintained. Basal melting of the glacier would provide a means for the gases in the glacial bubbles to be forced into solution in the meltwater. If the meltwater was not isolated from the atmosphere it would come into air equilibrium (S=1) and would be indistinguishable from the lake water input discussed above. Thus for the glacial bubbles to contribute to the gas ratios in the springs the basal melt region of the glacier must be isolated from the atmosphere and the glacial melt must find its way into the subsurface without being exposed to the atmosphere. Presumably this occurs at the base of the glaciers that are overriding the regional piercement structures (Figure 4.2b). The high pressure at the base of the glacier would ensure that the glacial gas goes completely into solution in the melting ice. To keep the available gas in solution (preventing bubble formation) the pressure would have to exceed 2 atm equivalent to 20 m of ice.

These two sources of gas have distinctly different relative ratios of the permanent gases. The ratios in air and the ratios in dissolved water are shown in Table 4.1. For example the ratio of N_2/Ar in air (37) is about twice the ratio in air dissolved in water that is in equilibrium with the atmosphere (84).

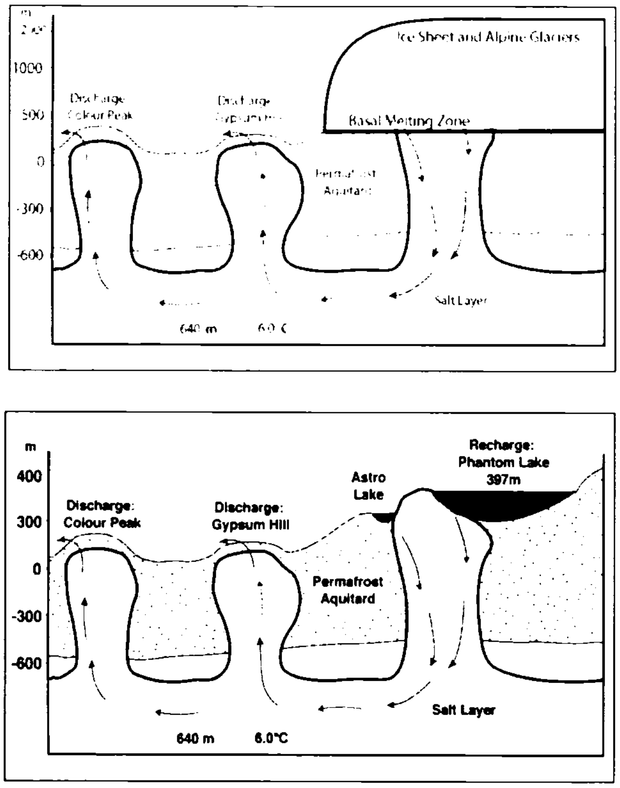


Figure 4.2a & 4.2b. Diagram of system showing glacial air trapping, flow through the glacier, subsurface melting, entering the diapir system, flowing out in the springs. Glacial gas trapped in the firn zone of a glacier contains all the air gases except neon. Neon is dissolved in the surrounding ice. Thus the ratio of Ne to Ar is determined by the volume fraction of bubbles in the glacial ice. If meltwater mixes with the glacial ice at the base of the glacier the total gas content of each gas will depend on the glacial bubbles and the dissolved gases in the meltwater. In this analysis we treat the solubility of neon in ice as equal to the solubility in water.

As the water from either of these sources descends into the subsurface via the piercement structures, dissolution of halite occurs and salt concentrations reach levels 3-5x seawater. The effects of this increase in dissolved salts on gas solubility are shown in Figure 3a. The increasing salt concentration results in a linear decrease in gas solubility and an overall increase in the level of gas saturation. For the salinity value of 100 ppt at LBP the solubility of gases is about half of its value for fresh water. Thus even if the O_2 is removed, the N_2 in water originating from the lakes will have a sufficient increase in saturation to form bubbles when the water emerges at the surface. For example, for lake water the saturation of the emerging water will be approximately 1.5 (= 0.78 x 2) and bubbles will form containing 1/3 of the available gas. For glacial gas the saturation could be higher depending on the fraction of bubbles in the ice.

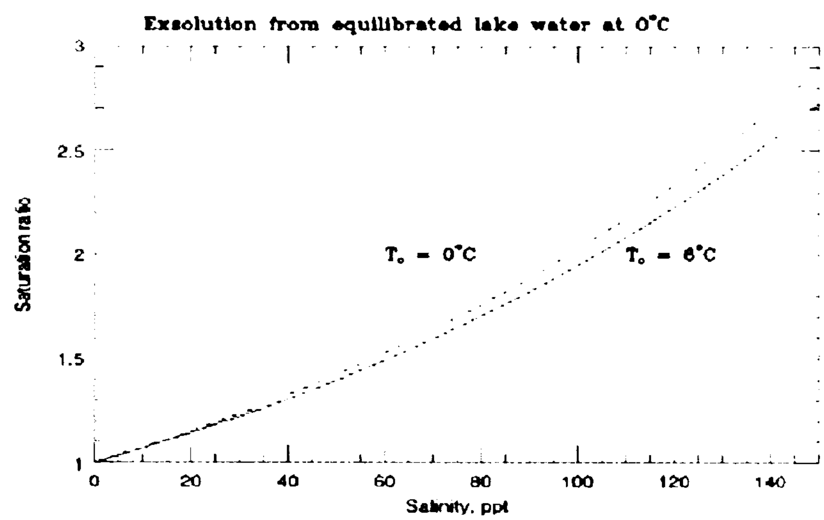


Figure 4.3a. Increased saturation due the presence of salt based on the formulae (eq. 27 & 28) in Benson and Krause (1976). Note that the effect of salt on solubility is treated as an independent of gas species. Curves for 0°C and 6°C are shown. There is a slight reduction in the salting out effect with increased temperature.

As bubbles form in LBP gases will come out of solution and reach equilibrium between the partial pressure of each gas in the bubble and the partial pressure of the dissolved gas. Because of the differing solubility of each gas the equilibrium values will in general be different.

We have computed the expected concentration of gases in LBP bubbles for a mixture of glacial ice and lake water. There are two free parameters in our model: the amount of bubbles in the glacial ice and the relative fraction of glacial ice to lake water. The glacial bubbles are assumed to be exactly air composition with the exception of Ne as discussed below. With these parameters chosen we can compute the total gas content of the water flowing from LBP. Using the Henry's law coefficients (Benson and Krause 1976) corrected for the salinity of 100 ppt we can determine the partitioning between bubble and dissolved spring water for each gas.

The depletion in Ne is used to determine the bubble content of the glacial ice. Ne and He are the only air gases that are soluble in ice (although He is also soluble, the He we observe is primarily derived from subsurface radiogenic sources). Hood et al. (1998) show that the solubility of Ne in ice causes it to partition differently from other noble gases in air/water/ice systems. As discussed in Hood et al. (1998) the solubility of Ne in ice is not well determined experimentally but is approximately equal to the solubility of Ne in water. This provides a natural mechanism for explaining the depletion of Ne in the gas bubbles. If the gas originates as glacial air then all the gases would be trapped in the bubble except Ne which would be dissolved in the surrounding ice. (See schematic in Figure 4.2b). Thus the Ne concentration relative to other gases can be used to determine the volumetric fraction of bubbles in the glacial ice. We find that the depletion of Ne to be 0.6 times the other noble gases implies a bubble volume in the glacial ice of 13.5 cm^3 per kg of glacial ice.

Figure 4.3b shows that, as expected, when there is mostly glacial ice the fractions of Xe and Kr are too low and N_2 is too high. This is because the glacial gas is derived directly from air and the ratios of these gases are air-like. However when these gases form bubbles in the spring they partition between the bubble and the spring water in way that reflects their solubility in water. The most soluble gases (Xe and Kr) are therefore depleted in the bubbles with respect to the less soluble gases like N_2. Thus air alone coming in with the glacial water will not give air-like ratios in the bubbles at the springs.

In contrast when there is mostly aerated meltwater the fractions of Xe and Kr and too high and N_2 is too low. In this case the infiltrating water is enriched in the most soluble gases compared to the less soluble gases. Thus as the increased salinity forces gases out of solution there is more Xe and Kr available than N_2. The net result is an enhancement of these gases.

Only for values near equal mass mixing do all ratios come close (but never all identical) to air values. The data points (open squares) correspond to values of the mass mixing ratio that go from 0.48 (Xe) to 0.59 (Kr). A least squares fit gives a best-fit value at 0.52 and a root-mean-square error to the measurements of 2.5%. Thus we find that a meltwater and glacial ice-mixing model can fit the data and results in reasonable values for the required bubble concentrations.

It is interesting to compare the supersaturations in the noble gases in the spring water with other natural systems that show high dissolved gases. Ocean and lake water typically have noble gas concentrations that approach 1-2% above saturation – two orders of magnitude smaller than our results.

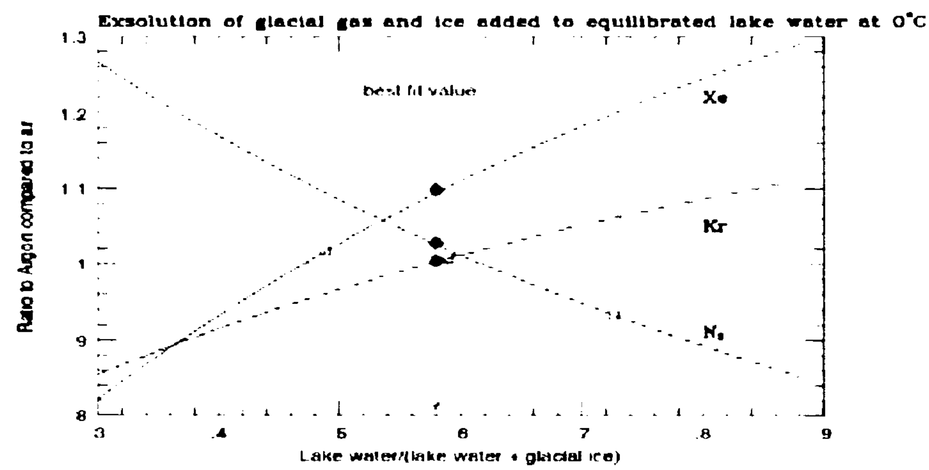


Figure 4.3b. shows the predicted gas ratios as a function of the fraction of the amount of aerated lake water (S=1, T=0°C) in the combined lake water and glacial ice mix. Gas is carried into the system both by the glacial bubbles and by the dissolved gases in the lakewater. Neon is carried only by the ice and the water. The size of the bubbles in the glacial ice determines the neon ratio. For the results shown in Figure 4.3b the bubble size is constrained to give the observed depletion in neon of 0.62. As expected when there is mostly glacial ice Xe and Kr are too low and N2 is too high. In contrast when there is mostly aerated meltwater X and Kr and too high and N2 is too low. Only for values near equal mass mixing do all ratios come close (but never all identical) to air values. The data points (open squares) correspond to values of the mass mixing ratio that go from 0.48 (Xe) to 0.59 (Kr). A least squares fit gives a best fit value at 0.52 and a root-mean-square error to the measurements of 2.5%.

Concentrations of dissolved gases reach the levels reported here in two other instances: excess air in ground water and perennially ice-covered lakes. Dissolved gas concentrations due to the excess air component in ground water as high as 200% are reported (Aeschbach-Hertig et al. 2000). Excess air is incorporated into the ground water by the physical trapping of air bubbles within the pores spaces of soils. All air gases are incorporated without fractionation by this mechanism. Thus, Neon, which is only slightly soluble is the component of choice for tracking excess air in groundwater. The supersaturation of gases in the perennially ice-covered lakes of the Antarctic dry valleys arises by the exsolution of gas from dissolved water during the freezing process (Wharton et al. 1987, Craig et al. 1992, Andersen et al. 1998, Hood et al. 1998). In this case the enhancement of gas is not simply air but is fractionated by the relative solubility in water (Craig et al. 1992, Andersen et al. 1998), the rate and completion of freezing (Craig et al. 1992) and the solubility in ice (Hood et al. 1998). The level of superstaturation in the emergent groundwater reported here represents a new mechanism for explaining extreme supersaturations; in this case the excess gas originates from bubbles in the glacial ice. Except for Neon this is similar to the physical incorporation of bubbles in groundwater via normal infiltration. In addition salinity affects the level of saturation --- but does not fractionate the gases. The use of the excess air component as a means to understand conditions present during infiltration and the dependence of noble gas studies upon this more complete knowledge has been demonstrated and may eventually provide paleoclimatic proxy data as important as other noble-gas derived data (Aeschbach-Hertig et al. 2000).

4.4 Conclusions

The gases in the Axel Heiberg spring outflows are derived from the dissolution of air contained in glacial ice bubbles (~50%) and from meltwater containing dissolved gas concentrations that are in atmospheric solubility equilibrium

(~50%). This suggests that the base of the surrounding ice sheets and glaciers are well anchored. If melting rates change, the ratio of gas derived from gas bubbles to dissolved gas in air solubility equilibrium would shift accordingly. Surface melts in the firn zone during summer (surface temperatures above 0°C) would cause a fractionation in the glacial gases due to the relative solubilities of the gases (Stauffer et al. 1985). That this is not consistent with our data suggests that the firn summer temperatures were not above zero. Our results show that the neon is depleted in glacial air while Ar, Kr, and Xe are not. We suggest that this is due to the depletion of Ne from the bubbles in the firn ice. The neon depletion is consistent with the known solubility of neon in ice. These results suggest that a greater degree of caution must be exercised when using excess Ne as a tracer of glacial input (Hamme and Emerson 2002, Hohmann et al. 2002). ^{4}Helium is highly enriched relative to air indicating a significant portion of the groundwater is very old (>10,000 yrs) although the springs themselves may be relatively young. Within the groundwater there is a reduction in solubility of the gases due to a high salinity concentration that promotes degassing and bubble formation at the spring outflows. Groundwater from glacially derived waters may provide a significant source of firn ice noble gases for study. We show that mixing of glacially derived excess air with dissolved gas in air solubility equilibrium complicates climatic interpretations. More detailed analysis of isotopic data (particularly Xe) may provide greater insight with regard to climatic change.

PREFACE TO CHAPTER 5

Microbial ecosystems exist in extreme environments on Earth and understanding their environmental properties (physical, chemical and biological) will enhance the liklihood of detecting evidence of life (past or present) on Mars. Chapter three presented a combined flow and thermal model of the springs located at Expedition Fiord to demonstrate the mechanism by which the springs are able to maintain perennial flow. Chapter four focused upon the origin and age of the spring water by considering the dissolved gases in the groundwater. This study presents a detailed analysis of the chemical and physical components of the spring ecosystem and provides the first data regarding the biological components. Using culture independent methods, phylogenetic analysis of environtaxa provides insight into the phylogenetic and physiological diversity of the organisms present in the spring water. Petrographic analysis and micropalentological examinations of mineral precipitates were used to investigate possible biomineralization and preservation.

CHAPTER 5: PERENNIAL SPRINGS IN THE CANADIAN HIGH ARCTIC: ANALOGUES OF MARTIAN HYDROTHERMAL SYSTEMS

5.1 Introduction

Water is present throughout the universe occurring in the gaseous phase within the interstellar medium and as low temperature, amorphous ice accreted upon interstellar dust grains (Committee on Planetary Biology and Chemical Evolution 1990). Liquid water appears to be much less common and to date the only place liquid water has been observed is on our own planet, Earth. There is strong evidence that liquid water exists elsewhere in our solar system beneath the thick ice-cover of several jovian moons and perhaps beneath the thick permafrost of Mars (McCord *et al.*1998, McCord *et al.*2001, Baker 2001). Water, with its unique chemical and thermodynamic properties provides a medium within which life has originated, evolved and thrived on Earth – but only in its liquid form. Ongoing efforts to discover, catalog, and understand life on Earth have yielded results that provide insight not only about the history of life on Earth but also about the range of habitats and extreme environments where life may be found beyond Earth. Studies of life on Earth residing in extreme environments are helping to define the search for life elsewhere in the universe. Characteristics of extreme terrestrial environments may include physical/chemical properties such as desiccation, high salinity, high radiation, low light, high pressure, high and low temperatures, extreme pH, and high or low oxygen tension. During the last twenty years investigations of terrestrial ecosystems around the planet have yielded an astonishing number of newly recognized microorganisms living in habitats in almost every conceivable place (Rothschild and

Mancinelli 2001). Despite enormous metabolic diversity and other clever adaptations used by the organisms that inhabit these niches, one fundamental requirement they all have in common is the necessity of liquid water which is required by cellular machinery to carry out metabolic and reproductive functions. Thus, the search for life is by default a search for liquid water.

A long-term, multinational commitment to the exploration of Mars is now underway (Squyres *et al.*2004, Kerr 2004). Currently there are three spacecraft in orbit and two rovers operating on the surface of Mars with the primary goal of understanding the history and presence of water on Mars. While these and previous missions have provided a wealth of information, future missions designed to detect life, past or present, will rely upon the lessons learned from examples of terrestrial microbial life, to design, build and test the methods and tools that will be needed on Mars.

Earth's polar regions provide relevant analogues to ecosystems which may have existed on an earlier, more clement planet Mars (Andersen *et al.*1992, McKay 1993). In regions of thick, continuous permafrost, the behavior of water is qualitatively different, its presence occurring primarily in the mineral phase as ice. Melting occurs principally as a result of seasonal and geothermal warming, and perennial liquid water environments are uncommon. While there is ample evidence that liquid water once existed at or near the surface of Mars, current climate models suggest that mean annual temperatures on early Mars remained below 0°C and permafrost has dominated the planet throughout its history (Segura *et al.*2002). Liquid water, being the non-negotiable prerequisite for the origin and evolution of living systems would have been required on Mars for life to evolve there. These aqueous environments would have been constrained by similar physical and chemical processes unique to polar settings on Earth. The Canadian High Arctic, particularly Axel Heiberg and Ellesmere Islands offer a number of analogous features found in association with the low temperature climate regime and the resulting thick, continuous permafrost.

Two sets of perennial springs are located near the McGill University Arctic Research Station next to Colour Lake at Expedition Fiord on Axel Heiberg Island (Figures 5.1 – 5.6). The springs at the base of Gypsum Hill were first reported by Beschel (1963) and until recently, there have been few data published on them since his initial description. Beschel's publication provided information detailing the location, discharge temperatures, flow rates and major ions of the spring water. Later studies by Pollard (1991) discussed the development of seasonal frost mounds and icings that form as result of the spring discharge. Curiously, despite the multitude of glaciological, hydrological and limnological studies that took place in this area (e.g. Adams, 1987, Allan *et al.*, 1987; English *et al.*, 1991; Schiff *et al.*, 1991, Boike *et al.*, 1993; Fishback, 1995), there were no further studies of these springs or reports of the springs at Colour Peak 11km to the west. The first detailed examinations of the springs located at Gypsum Hill and Colour Peak were conducted by Pollard *et al.*(1998, 1999). They demonstrated that the springs were perennial, described the two sites in the context of regional geology and physiography, provided additional geochemical results and speculated about the origin and age of the groundwater. Omelon (1999) and Omelon *et al.*(2001) examined the inorganic geochemical components and described the nature of the mineral precipitation occurring at the springs. Their field observations and modeling efforts suggest that the mineral ikaite ($CaCO_3$_$6H_2O$) a metastable form of calcium carbonate develops during the winter and early spring. Andersen *et al.*(2002) developed a combined flow and thermal model of the springs to demonstrate the mechanism by which the springs are able to maintain perennial flow. The origin of the groundwater and the age of the spring water were discussed in detail by Andersen *et al.* (2004 submitted JGR Planets) by evaluating the gases dissolved in the groundwater. They found that the spring water appears to be originating from sub-glacial flow from regional glaciers and ice caps mixed with water from local alpine lakes. The age of the water has yet to be determined however, Andersen *et al.*(2002) placed constraints using ^{4}He and tritium data. Tritium in the water is quite low (<0.8TU, Pollard *et al.* 1999) and helium-4 is supersaturated at values approximately 500 times atmospheric levels. This indicates that the water is pre-bomb (>50 years), and given the accumulation of helium-4 in the groundwater, possibly much older - up to 10,000 years. Andersen *et al.* (2004) point out that this is an upper limit and may be revised significantly with the use of better indicators such as krypton-81 (e.g., Sturchio *et al.* 2003).

5.2 Location and setting

At nearly 80° N, these are among the most poleward springs known, and are at roughly the same latitude as the thermal springs located at Spitsbergen (Lauritzen and Bottrell, 1994). The region is heavily glaciated with 30-35% of Axel Heiberg Island being covered by either ice-sheets or glaciers. Axel Heiberg Island lies within the Sverdrup Basin Archipelago, a large intracratonic rift basin initiated during the late

Paleozoic (Trettin, 1991). The local geology is complex and consists of folded and faulted sedimentary rocks ranging from Triassic to Tertiary in age. Upper Paleozoic evaporites locally intrude the overlying sedimentary/clastic rocks in a series of piercement structures. Both groups of springs at Expedition Fiord occur in geologically similar settings 11 km apart and are associated with anhydrite piercement dome structures (Hoen, 1964). Thick continuous permafrost 400-600 m thick has been reported for this region (Taylor and Judge, 1976).

The mean annual air temperature at the study sites is approximately –15° C (Doran *et al.*). The low temperature, coupled with a potential evaporation that exceeds the annual precipitation produces a region that is considered a polar desert. The perennial springs are comprised of two groups of springs, one at Gypsum Hill and the second at Colour Peak. The outflow waters from the springs have a year round ambient temperature range of -4°C to 7°C. At both sites, published results show that the main spring-flows maintain a 6.5° C discharge temperature and constant flow rate throughout the year, despite air temperatures that fall well below -40°C during the winter months (Pollard *et al.*, 1999; Andersen *et al.* 2002).

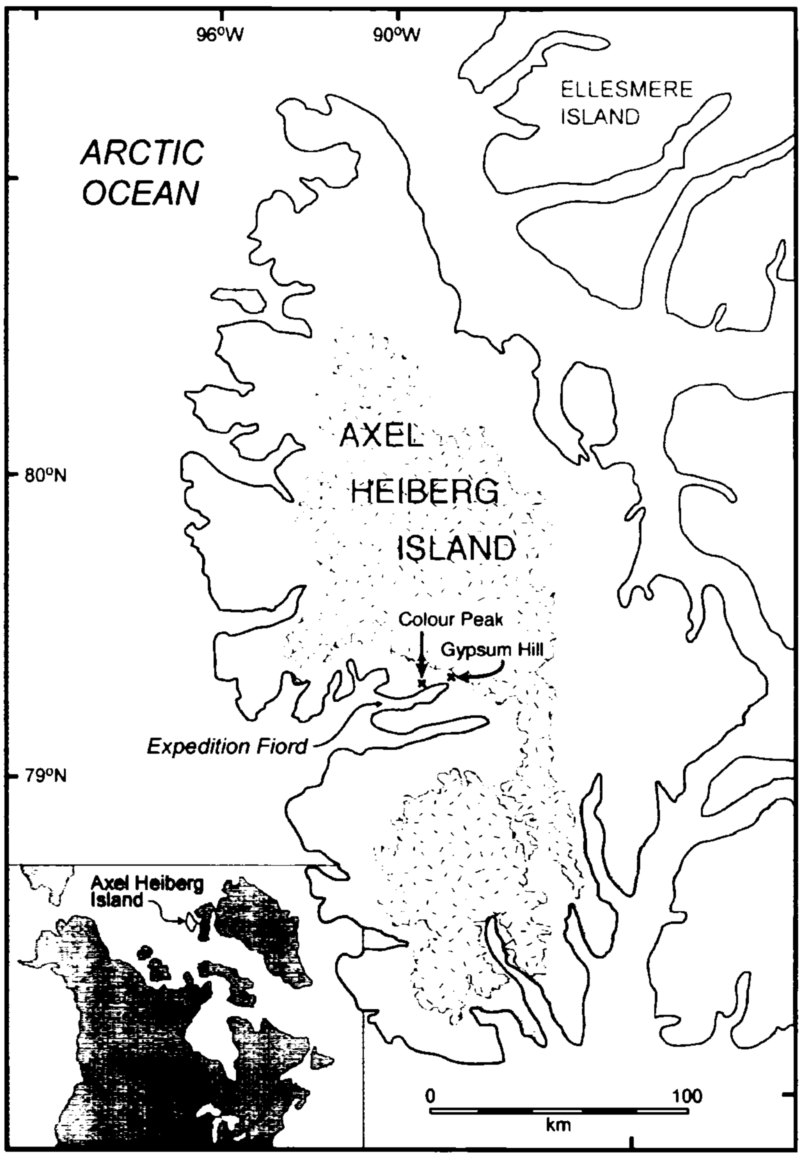


Figure 5.1 Location map of Expedition Fiord, Axel Heiberg Island, Nunavut Canada

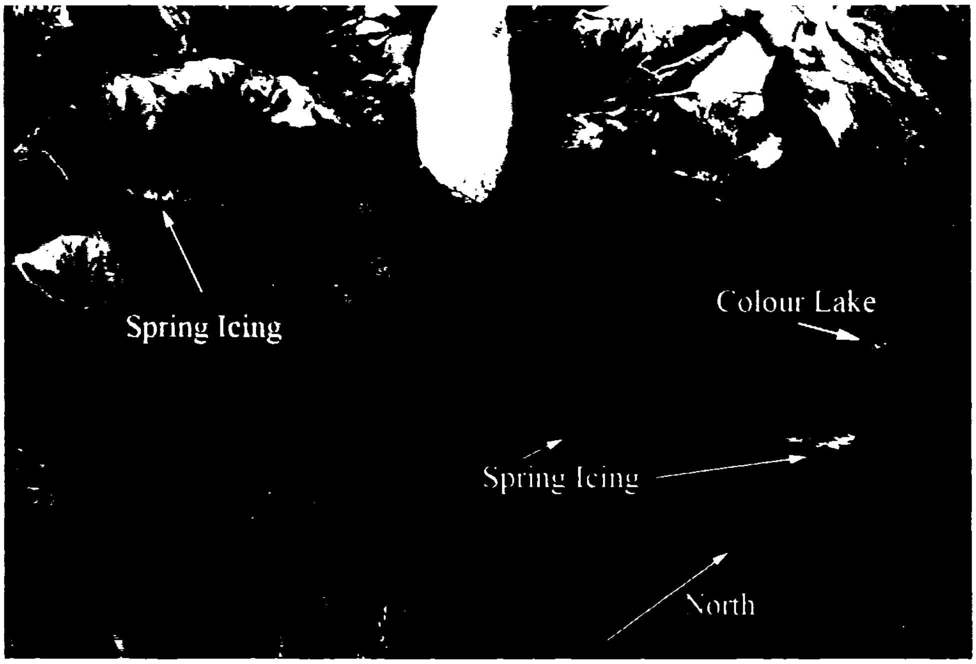


Figure 5.2 Expedition Fiord Study area. Residual icings resulting from spring flow are clearly visible in this ASTER image acquired on July 15 2002. Also, note the ice-cover on Colour Lake (above right) and Expedition Fiord (left).

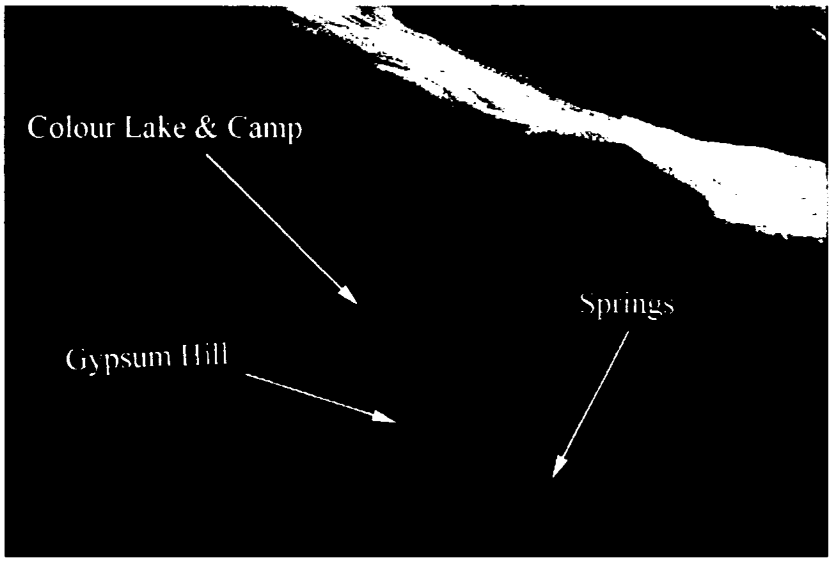


Figure 5.3 McGill University's Arctic Research Station
is located at Colour Lake. The perennial springs first described by Beschel (1963) are found at the base of Gypsum Hill on the north side of the Expedition River.

The outflow waters are anoxic, moderately saline (~9% NaCl), contain ~10 ppm methane, ~4000 ppm sulfate, and H_2S is present. Spring flow is derived from subpermafrost ground water rising to the surface along a "through talik" within the thick permafrost. Evidence of microbial activity is present in the form of biofilms on channel and sediment surfaces and long, white filaments in the outflow channels. The presence of hydrogen sulfide is also indicative of sulfate reduction. A summary of the currently published chemical data is presented in Table 5.1.

Table 5.1. Chemical composition of the spring waters at Gypsum Hill and Colour Peak spring, local precipitation and Colour Lake (From Pollard *et al.*, 1999).

	T (°C)	pH	Na mg/l	K mg/l	Ca mg/l	Mg mg/l	Cl mg/l	SO_4 mg/l	Alkalinity meq/l	Cond. (mS/cm)
Gypsum Hill										
Pollard et al.1999	6.0	6.74	34600	59.9	2300	118	38100	3700	756	105.0
Pollard (1991a)	5.8	6.50	28024	62.6	2115	115	44079	3763	-	100.5
Beschel (1963)	-	-	26600	55.0	2007	190	42225	3980	-	-
Colour Peak										
Pollard *et al.* 1999	6.2	6.78	52900	186	3628	316	68600	2300	532	170.0
Precipitation										
Pollard (1991a)	0.5	6.80	0.13	0.04	1.35	0.05	0.37	1.82	-	0.007
Colour Lake										
Schiff *et al.*(1991)	-	3.74	1.84	5.28	63.4	17.0	1.06	286	-	0.55

The Gypsum Hill site consists of approximately forty springs and seeps on the northeast side of Expedition River discharging along a band nearly 300 m long and 30 m wide and between 10-20 m ASL (Figures 5.3, 5.4). There are outflows occurring in Expedition River but thick ice and snow in the winter and high stream flows during the summer have prevented the discovery of their locations. The Colour Peak springs are located on the south-facing slope of Colour Peak at an approximate elevation of 30-40 m ASL (Figure 5.5). These springs are grouped into 3 distinct topographically controlled areas with 20 vents discharging directly into Expedition Fiord several hundred meters down slope. Unlike the springs at Gypsum Hill, the Colour Peak springs form distinct troughs, pipes, and sheet flows with well-developed travertine and other mineral precipitates (Figure 5.6). The source for the springs is most likely derived from subglacial melt from nearby ice caps and polythermal glaciers (Andersen *et al.* 2002), and D/O isotope data indictates meteoric water (Pollard *et al.*, 1999).

Figure 5.4 "Little Black Pond" a perennial seep at the base of Gypsum Hill. The seep is approximately 1m in diameter. The outflow temperature is a constant 6°C.

Figure 5.5 Perennial springs located at Colour PeakFigure 5.5 Perennial springs located at Colour Peak, Expedition Fiord. In this image remnants of residual ice can be seen throughout the site, however by mid-July most ice has melted or evaporated away leaving icing pastes and salts behind.

Figure 5.6 Spring channel at Colour Peak. The channels become lined with mineral precipitates forming protective troughs. Sediments adjacent to the spring flow are subject to much higher rates of weathering.

5.3 Materials and Methods

This study used a combination of field observations and laboratory experiments with the principal aim of developing a broad understanding of this unique ecosystem and making a preliminary identification of the major microorganisms present. Field observations and measurements were made during both early spring when temperatures were well below freezing and during the height of summer when temperatures were, on average, above freezing. Water and representative samples of mineral precipitates were obtained at the springs and returned to the laboratory for additional detailed analyses.

Meteorological measurements:

Climatic conditions at Expedition Fiord were monitored during this project at two principal sites. Meteorological data were collected at Colour Lake with a Campbell Scientific CR10 datalogger emplaced in 1992 by Doran *et al.*(1996). The station at Colour Lake initiates measurements every 30 seconds and records averages every 30 minutes. A model 05013 RM Young wind monitor (wind speed and direction) and LiCor LI200SZ (solar flux) are affixed to the mast ~2.5 m above ground, and a model 207 Phys-Chem relative humidity-temperature sensor housed in a radiation shield is attached to the mast ~2.2 m above the ground. Ground temperature is measured with two Campbell Scientific 107B temperature thermistors buried at depths of 2 cm and 34 cm. Data from the station were collected twice a year (spring, summer) providing a continuous record for the years 1998-2002. A second

Campbell Scientific CR10 datalogger was placed next to the largest spring outflow at Colour Peak to monitor local air temperature and humidity using a model 207 Phys-Chem relative humidity-temperature sensor housed in a radiation shield approximately 1.5 meters above the ground. Spring flow was continuously recorded using an Omega FP-5300 flow meter and spring outlet temperature was measured with a Campbell Scientific 107B temperature thermistor placed underwater. Spring temperatures were also monitored with Hobo dataloggers (Onsett Computer) placed into sealed Nalgene bottles. Thermistors leading from the dataloggers were placed into the spring water to record temperature every 30 min. Spot measurements were made in various locations using a Fluke model 51 digital thermometer (accuracy 0.05% + 0.3°C) with an external type J thermocouple.

GPS mapping

High-resolution (cm scale accuracy in both horizontal and vertical) GPS coordinates of spring outflows, channels, associated icings, frost mounds, sampling points and other features were obtained using a Trimble 4700 Total Station. Trimble's Survey Office software was used for post processing of data. A Garmin eTrex Vista was used to determine coordinates (accuracy <15m, 95% typical) of other springs, gypsum diapirs, pingos, and icings found during surveys conducted on Axel Heiberg and Ellesmere Islands.

Water chemistry

Routine water sampling was conducted at the springs. Samples were collected at Colour Peak at the main spring outlet, a point half way down the slope (~50 m) and at the base. Water samples were collected in acid washed or precleaned (Fisher iChem) 125 and 250 ml polyethylene bottles. Each bottle was rinsed three times with sample water at the time of collection. Water was filtered through GF/F glass fiber filters and the filters stored frozen in the dark for Chl *a* analysis. 20 ml of water was collected in sterile glass vials for bacterial and viral enumeration and immediately fixed with formalin (final concentration 2.5 %). Furthermore, 500 ml of water was collected and filtered through GF/C glass fiber filters pre-combusted at 450 °C for 5 hours. Four 125 ml aliquots were stored in Nalgene bottles for dissolved organic carbon (DOC), dissolved inorganic carbon (DIC), anions and cations and trace elements, respectively. The DOC and trace element samples were fixed with nitric acid and the DIC samples with a few drops of the bacteriacide, benzalkonium chloride (ZEPHIRAN®). All samples were stored in the dark at 4 °C until analyzed.

Oxidation and reduction potential (ORP or EH) and pH were measured with an Orion 250Aplus pH/ORP meter. The sensor used for pH was the Orion 91-07 with a Ag/AgCl internal reference system, sealed reference and built-in thermistor for Automatic Temperature Compensation (ATC). The pH was measured after auto-calibration using 3 standard buffers (4.01, 7.0 and 10.01). pH range of the meter is -2.00 to 19.99 with a resolution of 0.01/0.1 pH units and a temp resolution of 0.1°C .

EH was measured after calibrating with an Orion ASTM-approved ORP Standard (967961). The sensor used was an Orion Epoxy Sure-Flow combination electrode (9678BN). The sample's ORP millivolts were standardized to the Standard or Normal Hydrogen Electrode (NHE) and the instrument's ORP millivolts are displayed to 0.1mV resolution over the range of -1999.9 to +1999.9 mV.

Total alkalinity was measured with an Orion Total Alkalinity Test Kit using alkalinity reagent (700011). To a 100 ml sample, 10 ml of the Alkalinity Reagent was added, pH re-measured and the total alkalinity in ppm $CaCO_3$ was determined from the Orion pH conversion chart enclosed with the kit.

In situ chlorophyll fluorescence analysis was utilized to non-destructively test for the presence of chlorophyll on the surface of the mineral precipitates at the springs. Chlorophyll fluorescence was determined using Pulse Amplitude Modulated (PAM) fluorometry (Schreiber, 1994, Campbell *et al.* 1998). A Walz Diving–PAM was used to make approximately 100 spot measurements at various locations at the springs to identify any sites colonized by cyanobacteria.

Molecular phylogeny

Data on the microbial composition of the springs were obtained by applying a culture-independent approach. Inert substrates (pillows of fibra cells, Brunswick Scientific) were suspended in the stream of water exiting a spring for a period of two weeks. Samples of sediment and carbonate were also obtained for DNA extraction.

Sample analysis was performed in the Pace Lab at UC Berkeley and at the Rainey Lab at Louisiana State University. DNA was extracted from the biomass buildup on the samples and used in PCR to amplify the 16S rRNA genes of the organisms using universal oligonucleotide primers. The PCR products were then cloned into the Invitrogen TA cloning vector, classed into clone types by RFLP and sequenced. Ribosomal -RNA sequences were then compared to other sequences of known organisms and comparative analysis was used to compute evolutionary relationships between the unknowns and the sequences of known organisms.

Scanning Electron Microscopy

Samples of travertine and other mineralized surfaces that were collected for scanning electron microscopy (SEM) observations were placed in a 2.5% (vol/vol in PBS, pH 7) EM-grade gluteraldehyde solution to preserve cell morphology and structure. Upon return to the laboratory, the samples were dehydrated using an ethanol drying series and critical point drying, to further aid in the preservation of cell structure. Observations were made with a Hitachi cold field emission scanning electron microscope (S-4700) located at McGill University's Facility for Electron Microscopy Research.

Geochemistry, Petrographic analysis, and micropalentological examinations

Petrographic thin sections for optical microscopy and electron beam microprobe analysis were prepared from representative samples of travertine selected at various locations along the spring water flow path. Elemental compositions were measured by energy dispersive (EDS) and wavelength dispersive (WDS) spectrometers (using SEM and microprobe facilities at McGill University). In addition to petrographic analysis and micropalentological examinations, thin sections were sampled within and between laminations using an ultra-precision milling machine (MicroMill™ Sampler) to recover sample powders for chemical and isotopic analysis (^{13}C, ^{18}O). Analysis was performed at the Savannah River Environmental Laboratory (SREL) using a Finnigan MAT Delta Plus XL isotope ratio mass spectrometer. Bulk mineralogy was determined at NASA's Ames Research Center by powdered X-ray diffraction (XRD). A representative sample of the mineral precipitate from Colour Peak was ground to a grain size of 5-10 microns using a pre-cleaned mortar and pestle. XRD patterns were obtained and mineralogical composition was determined by comparing sample diffraction patterns to known mineral standards.

5.4 Results and Discussion

Climate, permafrost and locations of interest

The north and northwest arctic archipelago is characterized by polar desert conditions displaying very cold, dry winters and cool summers with maximum precipitation occurring in July (Maxwell 1982). Intermittent climate records for the Expedition Fiord locale are available for the last 30 years with a more complete record for the last 12 years. A long-term meteorological station is also located at Eureka (79°59'N, 85°56'W, 10 m ASL) on Ellesmere Island approximately 120 km NE of Expedition Fiord.

Long-term averages for Eureka are characterized by a mean annual air temperature of -19.7°C and mean monthly temperatures for January and July of -36.1°C and +5.4°C, respectively (extreme winter minima of -55°C are not uncommon). Inland, July air temperatures on Ellesmere Island frequently reach 20°C and nival streams rapidly recede after mid-June and are often dry by mid-July. Precipitation at Eureka is lower than measured at Expedition Fiord due to a rain shadow effect caused by the mountains on Axel Heiberg Island (Edlund and Alt 1989). Early climatological research at Expedition Fiord reported a mean annual accumulation of 371 mm (water equivalent) on the Müller Ice Cap (Müller 1963).

Table 5.2 Monthly Mean Average Temperature1998-2002

	1998	1999	2000	2001	2002
January	-31.96	-32.6	-31.16	-32.42	-33.4
February	-34.38	-27.86	-23.83	-33.76	-35.24
March	-28.48	-17.89	-15.94	-30.02	-31.02
April	-16.65	-4.68	-4.88	-20.8	-25.07
May	-5.47	5.17	6.82	-10.13	-6.26
June	5.39	9.55	6.64	2.3	1.98
July	7.36	0.29	-2.64	5.23	4.42
August	2.86	-9.68	-13.88	3.34	4.08
September	0.1	-25.34	-23.89	-5.92	-2.61
October	-15.27	-31.69	-25.7	-20.3	-10.07
November	-23.55	-35.03	-24.48	-23.53	-26.93
December	-33.33	-35.58	-22.1	-26.28	-24.08
Mean	-14.45	-17.11	-14.59	-16.02	-15.35

Table 5.3 Thawing index

	1998	1999	2000	2001	2002
January	0	0	0	0	0
February	0	0	0	0	0
March	0	0	0	0	0
April	1.07	10	17.65	0	0
May	7.79	160.68	211.4	2.7	4.65
June	162.13	286.55	207.19	92.62	83.51
July	228.66	57.35	20.62	163.09	135.61
August	91.36	0.29	0	114.09	128.17
September	36.51	0	0	7.89	13.34
October	0	0	0	0	8.96
November	0	0	0	0	0
December	0	0	0	0	0
Total	527.52	514.87	456.86	380.39	374.24

Table 5.4 Freezing index

	1998	1999	2000	2001	2002
January	-990.7	-1010.57	-965.81	-1004.93	-1035.31
February	-962.77	-780.15	-690.93	-945.34	-986.78
March	-882.79	-554.66	-494.18	-930.57	-961.77
April	-500.69	-150.44	-164.06	-624.09	-752.15
May	-176.43	-0.46	-0.07	-316.71	-198.68
June	-0.53	0	-8.12	-23.67	-25.12
July	-0.36	-48.27	-102.42	-1.03	-1.19
August	-2.8	-300.44	-430.37	-10.69	-1.72
September	-33.6	-760.32	-716.74	-185.55	-91.62
October	-473.34	-982.51	-796.72	-629.34	-321.25
November	-706.5	-1050.9	-734.33	-705.98	-807.98
December	-1033.28	-1102.86	-685.04	-814.7	-746.61
Total	-5763.79	-6741.58	-5788.79	-6192.6	-5930.18

Year-round data from an automatic weather station at Colour Lake (elevation 64 m ASL) collected since 1992 indicates a mean annual temperature of –15° C (Doran *et al.* 1996). Data collected during this study are presented in Tables 5.2-5.4 and indicate a similar trend, with the mean annual temperature being -15.5°C and approximately 451 thawing degree-days in the summer. The year 1998 was an unusually warm year with a mean annual temperature of -14.45°C however this was followed by a much cooler year with 1999 averaging -17.11°C for the year. The strong seasonal pattern typical of polar climates is readily observed in the plots of multiyear temperature and light data (Figures 5.7-5.8). A temperature data logger (Hobo, Onsett Computer) was placed into the spring outflow at Colour Peak approximately 75m downslope from one of the main outlets. It recorded temperature data in the stream for two continuous seasons (Figure 5.9) between July of 1999 and July 2001. The south-facing stream reached temperatures nearing 10°C during the summer months. One interesting observation of the data set is the occurrence of sudden drops in temperature over the duration of the measurements. As seen in figure 5.9, there are about 15 such sudden temperature drops. To better understand why these events were occurring the events were plotted on an hourly scale and compared to local met data from Colour Lake. One such event occurred on 11 Nov 1999. The temperature data from the logger for this day can be seen in Figure 5.10. The water temperature suddenly drops from several degrees above 0°C to nearly –10°C. The relative humidity (Figure 5.11) shows a sharp upward trend as does the local air temperature. The implication from this data is that the depression in temperature is due to snowfall that has accumulated enough to bridge the stream (Figure 5.6) in and around the outflow channel just above the thermistor. A portion of this snow then collapses into the stream causing a short-lived temperature drop due to freezing point depression as the snow crystals mix with the brine. Other *in situ* measurements taken during the formation of the spring icing in early April (with air temperatures ~-30°C) demonstrated that as the outlflow brine mixed with snow and formed a liquid/slush

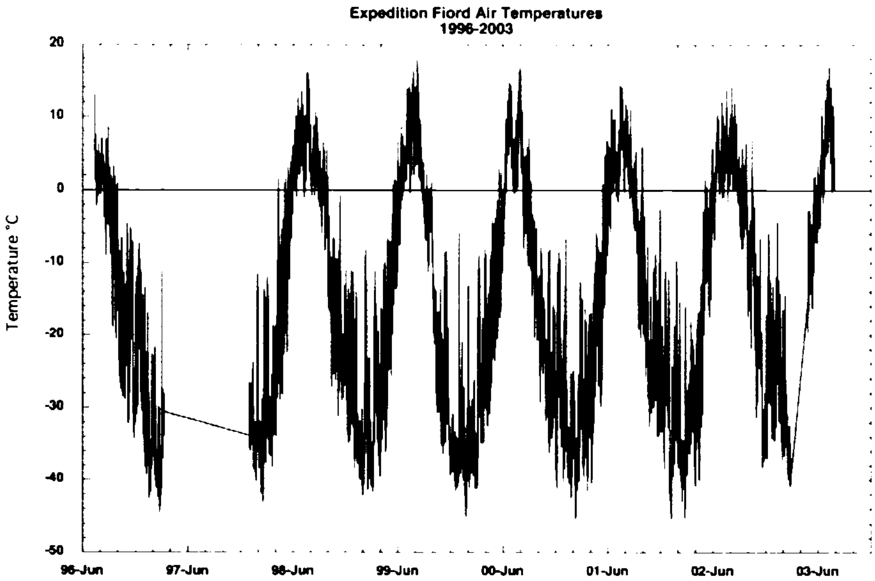
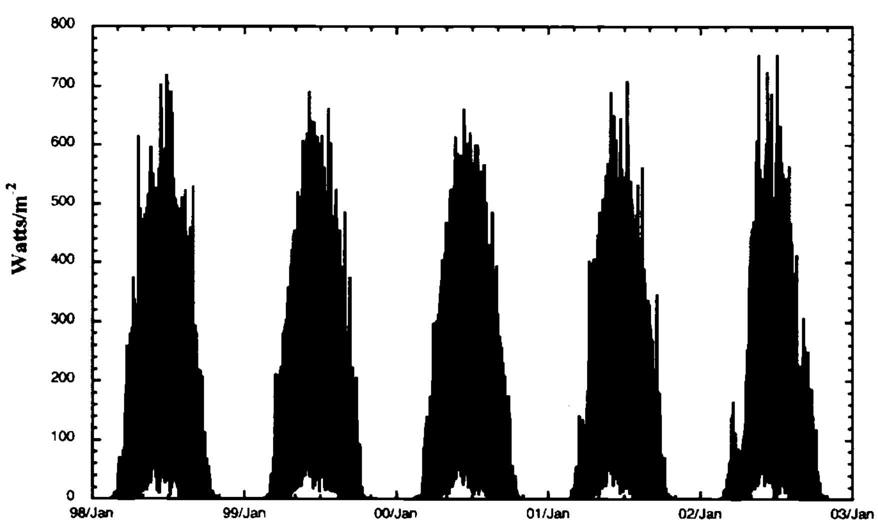


Figure 5.7 Temperature data recorded at Colour Lake between 1996 and 2003. Battery failure on two occasions caused the two gaps in the data set.



5.8 Solar flux observed at Colour Lake between 1998 and 2003demonstrates the strong seasonal pattern resulting from the high latitude.

mix, temperatures dropped to –10°C. As the snow melts and is cleared out of the channel, the temperature returns to its nominal value. Examination of other sudden temperature drops shows a similar pattern of a sudden decrease in water temperature with a concomitant rise in air temperature and humidity. Accurate measurements of snowfall in remote settings is technically challenging due to high power consumption of the sensor. These dropouts may provide a simple qualitative method to monitor local snowfall events that would otherwise not be measured. These sudden temperature fluctuations may also have an impact on the structure and function of any microbial consortia established in the outflow channel.

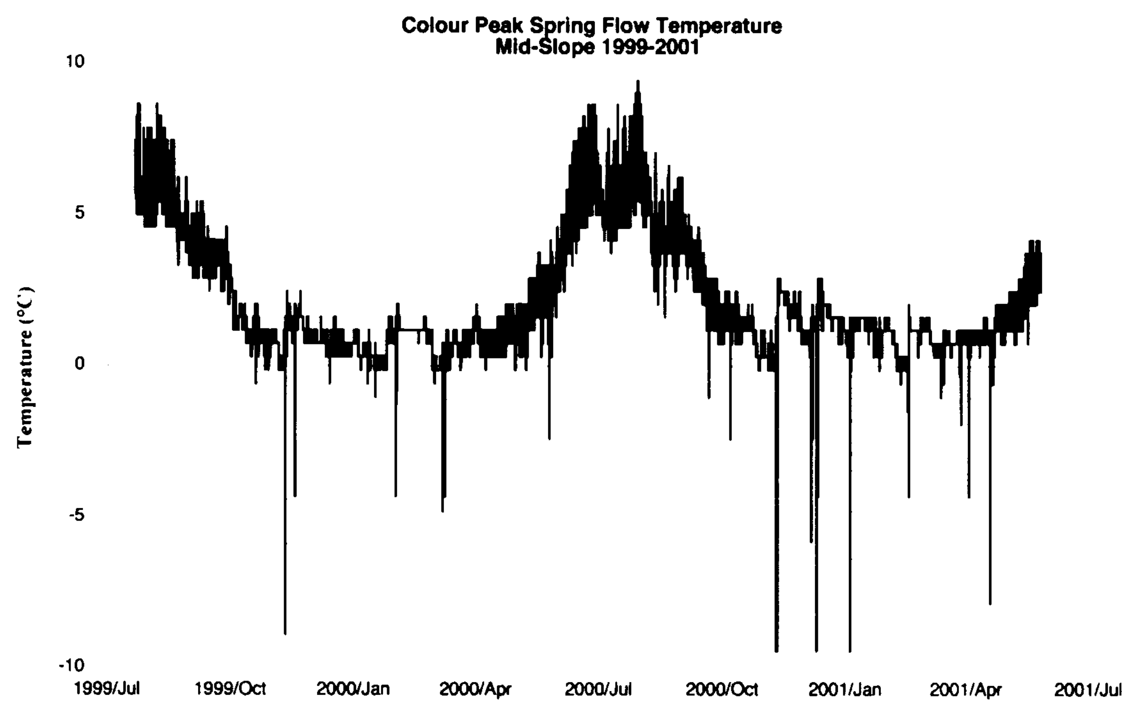


Figure 5.9 Spring-flow temperatures were measured over a period of two years at a site approximately 75m below the outlet at Colour Peak. Despite the mean annual temperature of -15°C, the spring flow remains above 0°C most of the time. During the summer months the water warms considerably, reaching nearly 10°C. The temperature drops are most likely the result of snow that has built up during precipitation events and then collapsing into the briny spring water, temporarily lowering the water temperature.

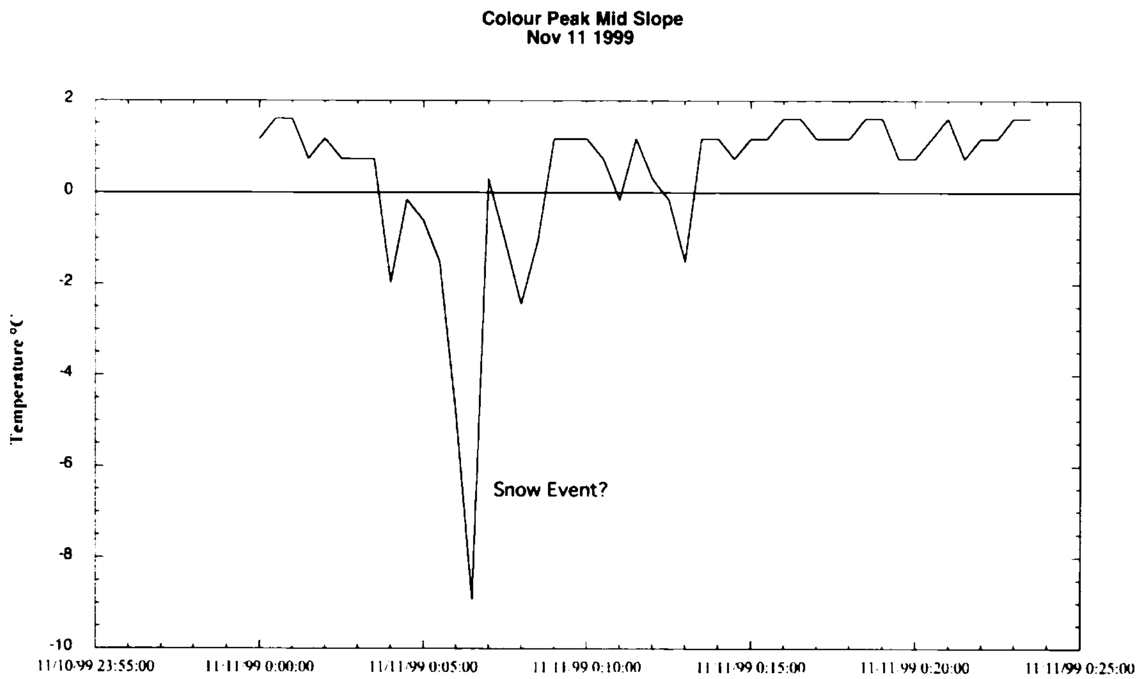


Figure 5.10 Plot of outflow temperature vs. time shown on an hourly basis for the day of 11 Nov 1999. A short-lived temperature drop occurred on 11 Nov 1999 just before 7am. By 3pm, the temperature had recovered to its 'normal' value.

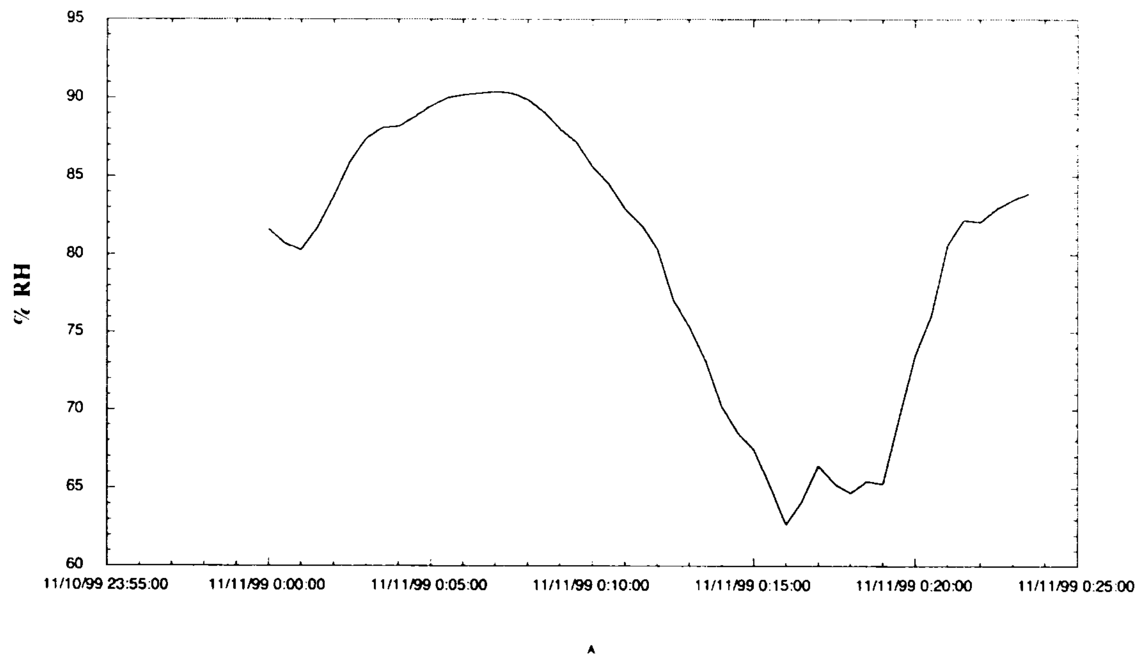


Figure 5.11 Relative Humidity vs. time for 11 Nov 1999 at the Colour Lake met station. At the time the water temperature is decreasing the RH is peaking at 90%. Air temperature for this time also rose (not shown). Sudden increases in air temperature and RH are typical during storms that may bring snow.

The regional climate regime controls the formation and depth of penetration of permafrost. Permafrost depth has not been measured at Expedition Fiord, however, a permafrost thickness > 400 m was documented in an exploration well on the east side of Axel Heiberg Island roughly 60 km from Expedition Fiord. Other exploration wells in the area reveal that permafrost is generally between 400-600 m thick. Permafrost features include extensive polygonal ice wedge development in unconsolidated fluvial and colluvial deposits at lower elevations. Wet areas develop tussock microtopography while poorly sorted circles and stripes characterize midslopes. A pingo exists in outwash gravels near Middle Fiord (Figure 5.12) and various frost and icing blisters and localized icings and icing mounds develop in the Expedition River floodplain (Figure 5.13). The pingo located at Middle Fiord flows during the summer months but it is has not been observed during the winter or early spring to determine if the flow is perennial. A second smaller pingo is located just to the north but discharge was not noted at the time it was observed. Massive ground ice is widespread in the Eureka Sound Lowlands (Pollard 199) but is limited to ice-cored moraines at Expedition Fiord.

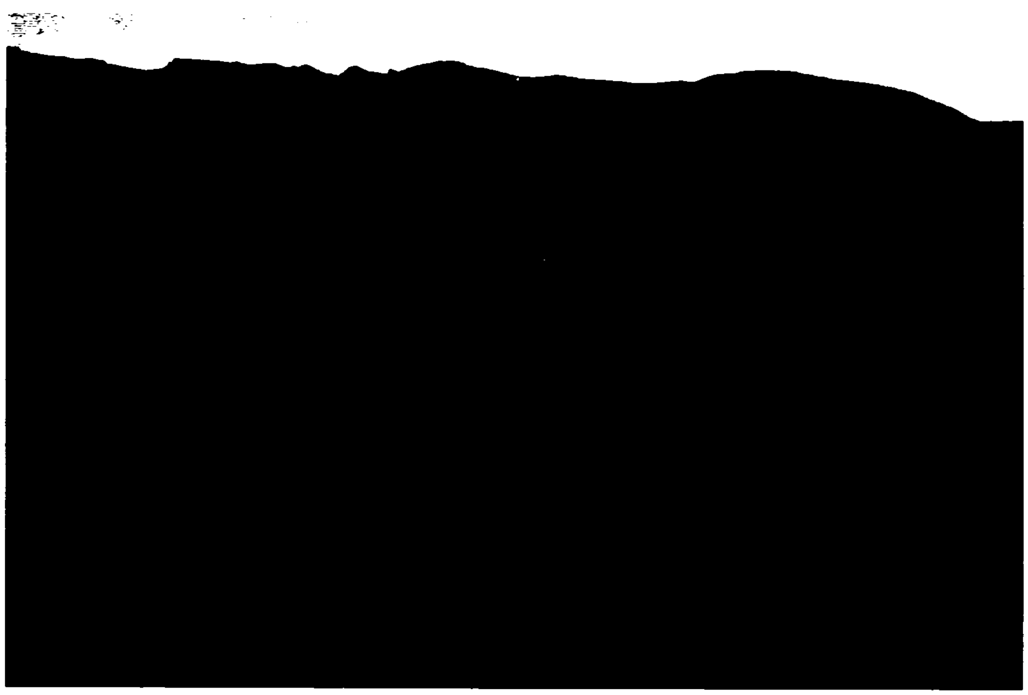


Figure 5.12 Pingo located at Middle Fiord, west-central Axel Heiberg Island. Note the discharge of water emerging from the middle of the pingo and cascading down its flanks. The pingo is approximately 30m in height and 100m in diameter.

Figure 5.13 Seasonal Frost mound formed as a result of spring activity at the base of Gypsum Hill.

Surveys for other sites of interest

Aerial surveys were conducted during this project in part to discover new sites of groundwater discharge. Table 5.5 lists locations of known spring discharges or of periglacial features that are associated with liquid water at or near the surface in this region of thick permafrost. These surveys successfully added three new spring sites, and three new icings that will require further study. The spring at Lost Hammer diapir is significant as it's similar in nature to the springs at Colour Peak, Gypsum Hill and Stolz diapir. The site is characterized by a large mound of salt 3m in height and 3m in diameter that forms 'fumarole-like' structure. A saltpan extends about a half a kilometer to the west. Water is present as a pool in the central portion of the structure, and it appears that during the winter months, as temperatures decrease, the water level builds up and flows out and over the top. During July 2004, water flowing beneath and through the side of the structure was saline with a temperature of -3.5 °C. Hydrogen sulfide was present in and around the discharge site. It is likely that the spring flows throughout the year as there is a large amount of salt built up and small terraces and rimpools are found on the outside edge of the structure. Without continual replacement, this salt structure would quickly erode away by way of aeolian or fluvial processes.

A saline discharge at Junction diapir is located on the southern edge of the diapiric structure flows into a stream that empties into a larger stream on the valley floor. A residual icing has not been seen associated with the discharge indicating that flow does not persist through the winter. The origin of the water for both Lost Hammer and Junction diapirs may originate from the Stacie Ice Cap which is located several km to the south of each feature. No other large bodies of water are present near the outflows.

Sub-glacial discharge has been observed at an alpine glacier near the north eastern end of Skaere Fiord. The flow was discovered by twin-otter pilots flying over the glacier during the early spring when air temperatures were still well below freezing. Initial reconnaissance during the summer found an icing that was approximately 0.5km in diameter with red-tinted fine clays covering the surface. The discharge was flowing from the central area of the icing at a rate of ~250 ml/sec, the water tasting strongly of iron, but not saline. This feature may similar to the spring reported by Grasby *et al.* (2003a, 20003b) located in Borup Fiord Pass on northern Ellesmere Island. At the Borup Fiord site, sulfur, gypsum and calcite pastes are present on the glacier and icings. At the Skaere Fiord site it appears that iron is a major constituent of the icing paste however chemical analyses have yet to be made. Whether or not either spring is perennial is still to be determined.

Table 5.5 Location index for known springs, icings and pingos in the Canadian High Arctic.

Feature	Latitude	Longitude
Gypsum Hill Springs	N79°22.192'	W090°43.939'
Colour Peak Springs	N79°22.866'	W091°16.270'
Stolz Diapir Springs[2]	N79°05.343'	W87°02.228'
Junction Diapir Spring[1]	N79°03.736'	W090°41.893'
Lost Hammer Spring[1]	N79°04.285'	W090°12.755'
Skaere Fiord Glacial Outflow[1]	N78°56.702'	W088° 17.666'
Borup Fiord Glacial Outflow [2]	N81°01.660'	W081°38.484'
Middle Fiord Pingo 1	N 79°43.653'	W94°13.794'
Middle Fiord Pingo 2	N79°47.154'	W94°09.343'
Icing 1 North West Axel[1]	N 80°22.749'	W93°57.584'
Icing 2 West Axel[1]	N79°58.966'	W93°57.539'
Icing 3 Strand Fiord[1]	N79°17.517'	W090°06.314'

[1]located during this study
[2]located via previous literature

Gypsum Hill and Colour Peak Water Chemistry

Results for the major anions, cations and trace elements for two outlets at Colour Peak and Gypsum hill are reported in table 5.6 and 5.7. These data are very similar to previously reported results (see Table 5.1) with little change noted in the major ions since Beschel's first observations or those of Pollard et al. (1999). The water can be characterized as a near neutral, anoxic sodium chloride brine with high levels of calcium and sulfate. The brine may be considered as formation water produced from reservoirs in the Sverdrup Basin derived from evaporated seawater. Total phosphates, nitrates, nitrites and dissolved organic carbon and dissolved iron were all found to be below detectable limits. Oxidation-reduction potential measurements, which provide a measure of the electromotive force of a water, are reported in tables 5.6, 5.8 and 5.9. The Eh of the main outlets of Colour Peak and Gypsum Hill are shown in table 5.6. Both show the outflowing water to be highly reduced. At Colour Peak, additional measurements were made downslope within the stream channel. It can be seen that the redox values do not change significantly even after having flowed 75m. At approximately 30 and 50 m the Eh drops again suggesting that additional seeps are contributing reduced groundwater to the flow. Stable isotopes for carbon and oxygen were obtained from mill filings following precision drilling of a polished thin section. Tracks between seven different laminae were sampled. Little variation was seen between samples. The mean value for $\delta^{13}C$ was –8.93 while the mean value for $\delta^{18}0$ was –18.88. This may reflect temperature stability of the water in which the carbonates precipitate. Temperature measurements taken at various discharge points are fairly variable (table 5.11) ranging from a low of -4.0°C to just over 8°C.

Table 5.6 Major ion chemistry of the spring water collected at Colour Peak and Gypsum Hill outlets. *ORP values are reported as mV normalized to a hydrogen electrode (NHE).

	Colour Peak spring	Gypsum Hill spring
Temperature (°C)	6.4	6.6
pH	6.82	7.66
ORP (mV, NHE)*	-122.4	-97.1
Conductivity (mS/cm)	170	105
Density (g/cm^3)	1.114	1.056
Ca^{2+} (all in mol/kg)	0.0905	0.0574
Mg^{2+}	0.0130	0.0049
Na^+	2.30	1.5043
K^+	0.0048	0.0015
Cl^-	1.9324	1.0732
SO_4^{2-}	0.0240	0.0385
PO_4^{3-}	LDL	LDL
NO_3^-	0.001	LDL
DIC (mmol/l)	0.278	0.389
DOC (mg/l)	LDL	LDL
Alkalinity	0.0184	0.0155

Table 5.7 Elemental analysis of water collected from Colour Peak (CP), Gypsum Hill (GH)and Expedition Fiord sea water. All results reported as ppm.

Element	CP Top	CP Mid	CP Bottom	GH-LBP	Exp. Fiord
Sr 88	87.7	88.4	86.7	39.9	6.9
Mn 55	0.111	0.102	0.089	MDL	MDL
Cr 52	MDL	MDL	EQL	EQL	MDL
Fe 54	MDL	MDL	MDL	MDL	MDL
Na 23	67189.3	63116.8	60061.4	28461.7	9909.256
Mg 24	389.9	361.1	343.9	106.9	1175.13
Mg 25	381.9	358.3	340.7	105.7	1148.5
K 39	282.1	272.5	266.3	58.1	367.9
Ca 44	4083.8	4034.6	3968.0	2155.5	EQL
Rb 85	0.895	0.905	0.89	EQL	EQL
Cs 133	0.146	0.148	0.153	MDL	MDL
Sb 123	MDL	MDL	MDL	MDL	MDL
Mo 98	MDL	MDL	MDL	MDL	MDL
Cd 114	MDL	MDL	MDL	MDL	MDL
Ag 109	MDL	MDL	MDL	MDL	MDL
Se 82	MDL	MDL	MDL	MDL	MDL
As 75	MDL	MDL	MDL	MDL	MDL
Ba 137	EQL	EQL	MDL	MDL	MDL
Pb 208	0.027	EQL	0.033	MDL	0.038
U 238	MDL	MDL	MDL	MDL	MDL
Al 27	MDL	MDL	MDL	MDL	MDL
Co 59	MDL	MDL	MDL	MDL	MDL
Ni 60	EQL	EQL	MDL	MDL	MDL
Cu 65	MDL	MDL	MDL	MDL	MDL
Zn 66	EQL	EQL	MDL	EQL	MDL

Table 5.8 pH, ORP and outlet temperature at Gypsum Hill and Lost Hammer spring

	pH	ORP	Temp°C
GH 1	7.66	-97.10	6.4
GH-2	7.06	-58.10	0.2
GH-3	7.35	-81.00	6
GH-4	7.37	-152.40	5.8
Lost Hammer	****	60.00	-3.5

Table 5.9 pH and ORP measurements starting at the outlet and moving down slope at Colour Peak spring, 4 July 2004.

	Outlet	10m	15m	30m	50m	75m
pH	6.8	7.14	7.64	8.4	8.61	8.39
ORP (mv)	-122.4	-86.4	-91	-112.5	-121.7	-69.2

Table 5.10 ^{13}C and ^{18}O measurements taken from material milled from polished thin section. Little variation is seen between the layers.

	$\delta^{13}C$ vs. V-PDB	$\delta^{18}0$ vs. V-PDB
CP #1	-8.42	-18.34
CP #2	-9.31	-19.84
CP #3	-8.98	-19.18
CP #4	-8.97	-17.31
CP #5	-8.98	-18.93
CP #6	-9.04	-19.55
CP #7	-8.84	-18.99

Table 5.11 Temperature and pH for outflows at Colour Peak and Gypsum Hill

Site No.	Colour	Peak	Gypsum	Hill
	Temp °C	pH	Temp °C	pH
1	-1.4	6.4	1.4	7.7
2	-4.0	6.7	-1.1	7.8
3	1.3	6.9	3.3	7.9
4	7.1	6.3	6.4	7.4
5	-2.8	6.6	2.4	7.4
6	4	7.5	3.6	7.5
7	8	7.7	5	7.5
8	8.2	7.6	5.3	7.3
9	3.2	6.7	5.8	7.4
10	5.4	6.6	5.6	7.3
11	2.5	6.6	5.1	7.9
12	2.5	6.62	0.3	7.4
13	5.4	6.6		
14	9.5	6.6		
15	3.4	6.72		
16	2.4	7.21		

Characterization of the microbial community

The purpose of this phase of the study was to provide an initial description of the microbial community associated with the springs. The culture-independent, or molecular approach, to prokaryotic diversity assessment has become prominent in recent years. This involves isolating DNA or RNA from an environmental sample, or the microorganisms concentrated from within a given sample, amplifying and cloning the individual 16S rRNA genes which originate from individual cells or groups of cells, and determining the degree of diversity and the phylogenetic position of the individual sequences. These sequences can be placed in a taxonomic context and since they are known to originate from the environmental sample under investigation, the source organisms of the sequences are referred to as "environtaxa" (Rainey and Ward-Rainey, 2000).

This culture-independent approach has provided insight into the extent of microbial diversity, and exposed the inability of culturing approaches to recover the full range of diversity present in a wide variety of environments (Pace 1997). The ability to detect novel taxa in this way is arguably the greatest strength of the molecular approach. According to Hugenholtz et al., (1998), the domain Bacteria can be divided into 36 phylogenetic divisions, 13 of which are composed exclusively of so-far uncultured taxa. Based on the phylogenetic position of some environtaxa it is possible to predict the physiological type of the organism from which the sequence originated and thus design an enrichment or isolation protocol to target the culturing of that organism or closely related species. It has become clear that a complete picture of the prokaryotic diversity present in a given environmental sample is most likely to be obtained through the use of a biphasic approach, employing both culturing and culture-independent techniques however for the purpose of this study, only the culture independent method was used. However data obtained thus far will aid with the planning and development of future more focused studies of the spring ecosystem.

The sites for sampling for the microbiological studies of these polar springs consist of two types: outflow spring water, and spring water with associated material in pools, channels and soils in runoff areas. This study focused solely upon the main outflow at Colour Peak.

Outflow Spring Water:

The outflow spring water represents an interesting and extreme environment at these spring sites because of its extreme physical/chemical parameters (i.e. origin, cold temperature, moderate salinity, anoxic, sulfate saturation). Microorganisms able to survive and be physiologically active in these spring waters may possess the following characteristics; (a) cold-adapted, either psychrophilic/psychrotrophic, (b) moderately halophilic, (c) anaerobic or facultatively anaerobic (c) possibly methanogenic, sulfate-reducing, sulfur oxidizing, because of the presence of CH_4, SO_4, and S^o, in the spring water.

Preliminary data on the microbial composition of the spring water has been obtained by applying the culture-independent approach. Inert substrates (pillows of Fibracells (Brunswick Scientific)) were suspended in the stream of water exiting the spring for a 2 week period. DNA was extracted from the biomass buildup on the substrates and used in PCR to amplify the 16S rRNA genes of the organisms present. Subsequent cloning and sequencing and phylogenetic analysis of 55 environtaxa provides the first insight into the phylogenetic and physiological diversity of the organisms present in the spring water. These data are presented in Tables 5.11 and 5.12. The majority (76%) of the fifty-five environtaxa showed high sequence similarity to *Thiomicrospira* species (sulfur-oxidizing organism). Other sequences show high similarity to sulfate-reducing members of the delta (*Desulfocapsa* sp.) and epsilon (*Sulfurospirillum* sp.) proteobacterium groups. A single sequence was found to have 99% sequence similarity to species of the genus *Haloanaerobium* a group of low G+C Gram positive, anaerobic halophiles (Rainey et al., 1995).

Table 5.12 Environtaxa present in the spring water at Colour Peak.

97% commonality with *Desulfocapsa* thiozymogenes, a sulfate reducer
97% commonality with *Thiobacillus* hydrothermalis, a sulfur/sulfide oxidizer
96% commonality with *Thiomicrospira* sp., a sulfur/sulfide oxidizer
A very close match with *Shewanella* sp., a Fe(III) reducer
98% commonality with *Haloanaerobium acetoethylicum*
91% commonality with an unidentified gamma proteobacterium

Spring water and associated material in pools, channels and soils in run off areas:

As the outflow water leaves the springs and enters pools, channels, and surrounding soils, the physical/chemical parameters change considerably and will have a significant effect on the microbial communities. Unlike the outflow water, the physical/chemical parameters encountered in these habitats become very heterogeneous. Anaerobic, microaerophilic, and aerobic conditions, as well as moderately saline to hypersaline conditions are present. The one parameter that appears to remain relatively constant is the cool temperature of the water although the surrounding air/soil/sediment temperature would vary considerably, 10°C to –40-50°C. In these areas the diversity should be more extensive than in the direct outflow of the spring, and a wide variety of prokaryotes (anaerobic, microaerophilic, and aerobic; halotolerant and halophilic, chemolithoautotrophic and heterotrophic) may be present. The prokaryotes should be more diverse in the run off channels and surrounding areas than in the outflow spring water due to the presence of organic substrates eg. plant leaves that blow in from the surrounding areas.

The preliminary results described above clearly demonstrate that the culture-independent approach can be applied to this environmental situation and that the results obtained do correlate with the geophysical conditions found.

Mineralogy and geochemistry

Mineral precipitates occur in association with spring activity at both Gypsum Hill and Colour Peak ranging from simple salt crusts to complex channels, terraced mounds, and cascade structures. In some areas the travertine develops distinct forms such as elevated channels and tunnels, compared to other locations where mineral precipitates are thin and poorly developed. Precipitate wall thickness ranges from several millimetres to several centimetres. In less steep, slower flow areas, terraced mounds or barrages are commonly observed at large breaks in slope where the water fans outwards and water depths decrease. Rimstone pools ranging in size from 5-15 cm in lateral dimensions and 2-5 cm deep are common to these areas (Figure 5.14). Low temperature, near neutral pH discharges and the associated alteration minerals are of interest particularly since they may have been more common on early Mars and may have provided a site for the preservation of biological information.

Figure 5.14 Metastable carbonate crystals up to 0.5 cm in length form rimstone pools in areas of low, non turbulent flow at Colour Peak. Clays, organic detritous and other sediments fill in the depressed area in the central area. These large, white carbonate crystals, tentatively identified as the mineral ikaite by Omelon *et al.*(2001) have only been observed to form during the colder months when temperatures are well below freezing. Scale is 10 cm.

The powdered x-ray diffraction of the samples collected from the spring channels showed the material to be calcite (Figure 5.15) however trace constituents would not be seen using this method. Polished thin sections were produced from samples collected within the spring channels at Colour Peak (Figures 5.16 and 5.17). These samples have been found to contain highly laminated, birefringent, fine grained sparite crystals producing layers 50 –100 μm thick. The layers are tightly joined with a narrow micritic area between the bands of crystals (Figures 5.18-5.21). Figure 5.22 shows a magnified view of the micritic region using light microscopy under oil immersion at 1000x. At this magnification small granules are seen populating the region between the large crystals. WDS microprobe (Figures 5.23a-c) and EDS analysis (Figure 5.24) both indicate the presence of iron and sulfur in this region. however the levels are not what might be expected for what appears to be a large population of pyrite grains. Given that the water is rich in sulfide and is moving to the surface through marine clay that contains iron, it is not unreasonable to expect FeS_2 to precipitate out inorganically or perhaps with the additional catalyst of microbial activity. Nevertheless, additional analyses will be required to understand this fully.

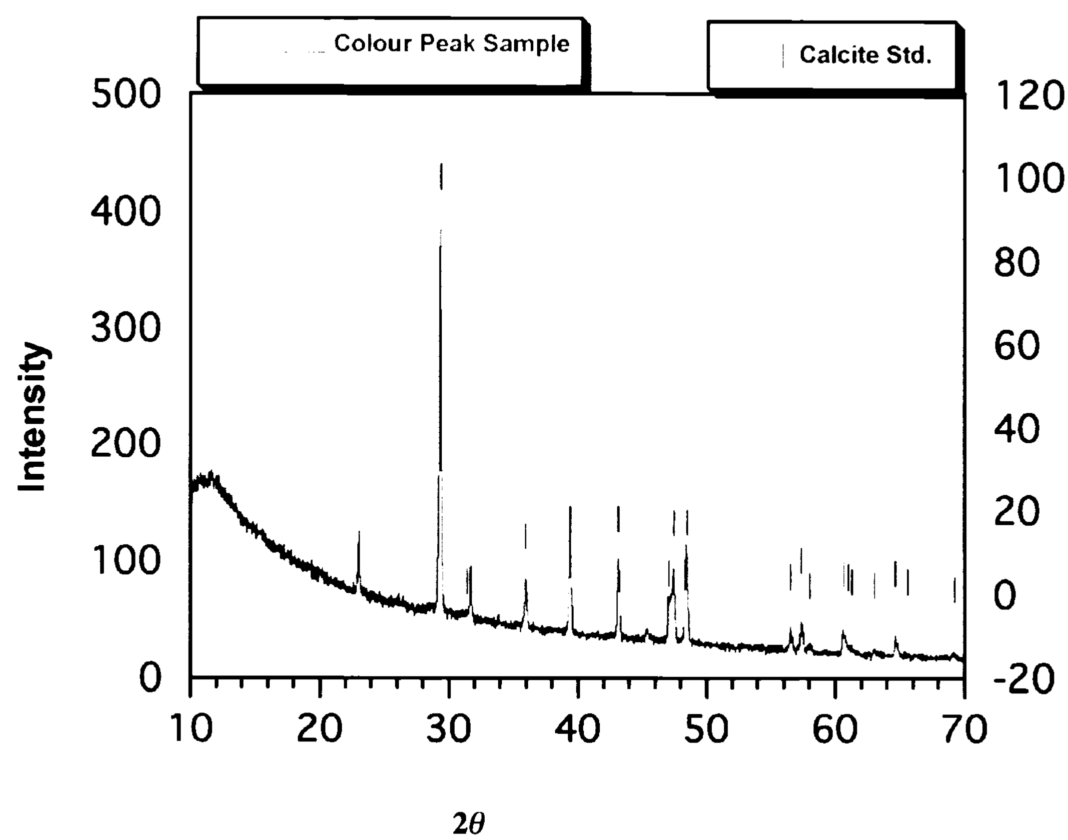


Figure 5.15 X-ray diffraction pattern of mineral precipitate obtained from outflow channel at Colour Peak indicates the material is calcite.

Figure 5.16 The carbonate deposits exhibit a smooth, botryoidal texture.

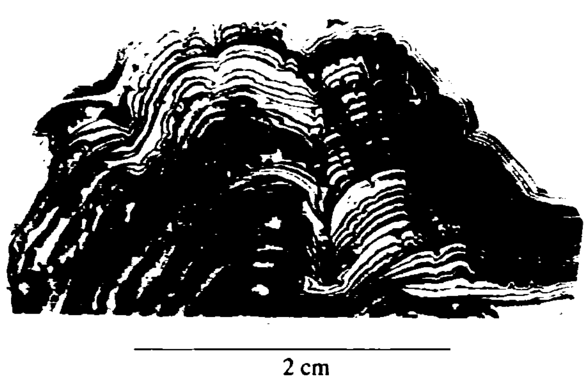


Figure 5.17 Polished thin section prepared from the carbonate material seen in Figure 5.16 that was collected at Colour Peak.

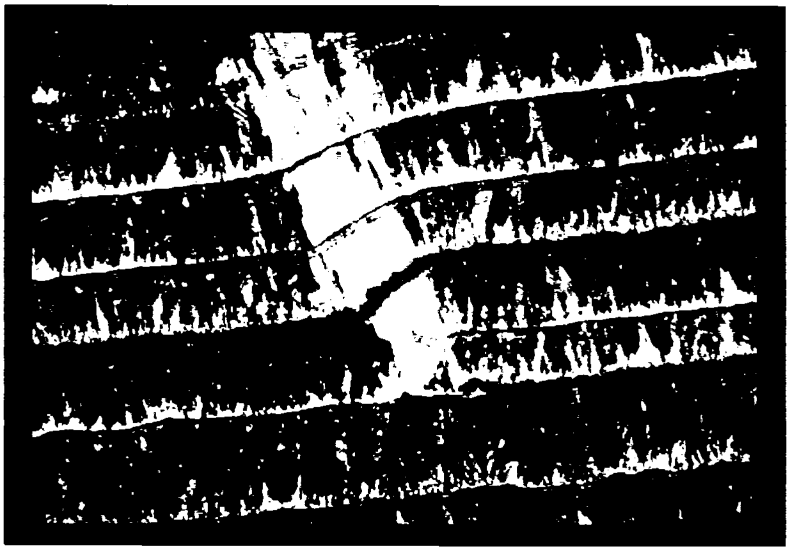


Figure 5.18 Photomicrograph (magnification 200x) of polished thin section taken from the channel travertine at Colour Peak spring.

Figure 5.19 SEM image of stromatolitic carbonate from Colour Peak. The surface of the sample is located at the upper right corner. Note the stacks of large carbonate crystals. Scale bar is 50 μm.

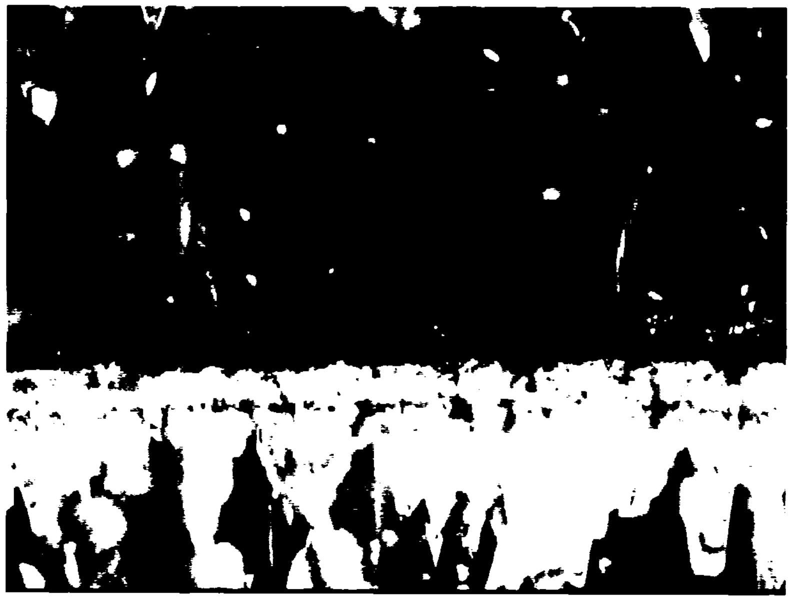
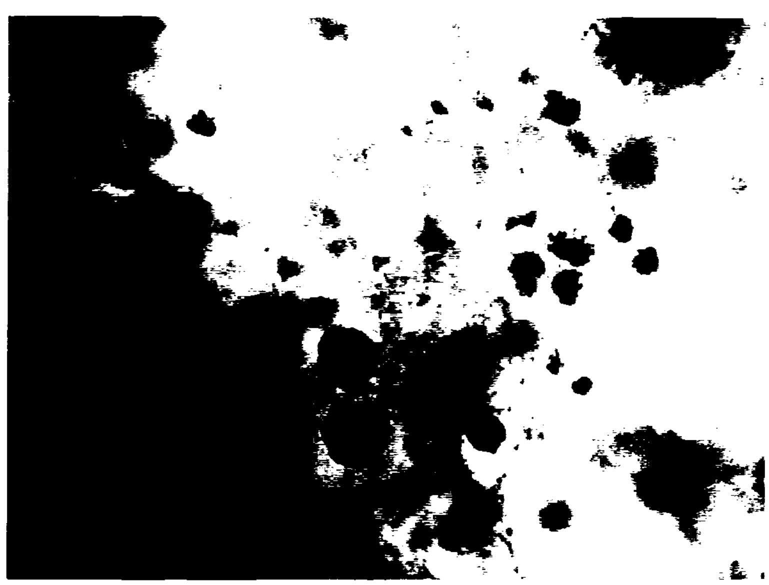
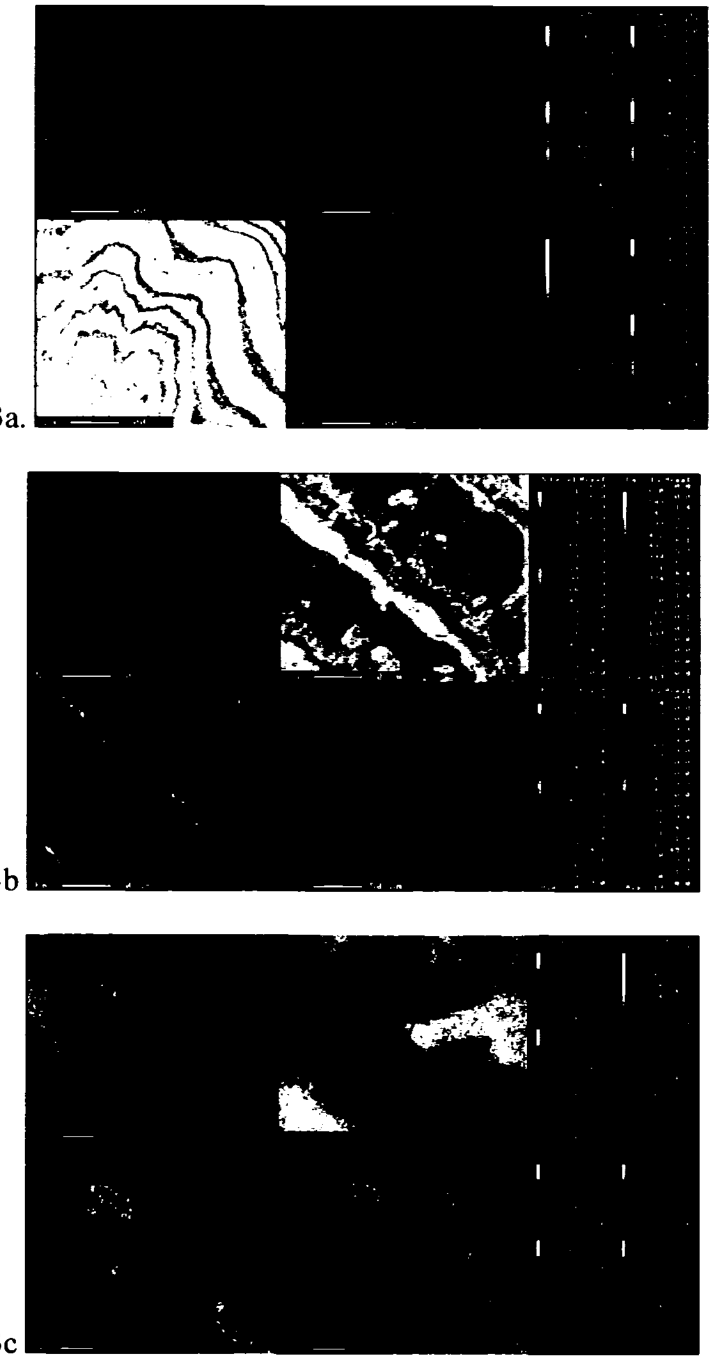


Figure 5.20 Photomicrograph (magnification 450x) of polished thin section prepared from a sample of carbonate from Colour Peak. The crystals are highly bifringent and tightly ordered. The micritic area resides between the large calcite crystals.

Figure 5.21 Micrite layer between the calcite crystals. Scale bar is 10 μm.

Figure 5.22 Photomicrographs (above @ 500x and below @1000x) of the micrite layer observed within the polished thin section. The 'dogtooth' crystal is about 5 μm across. Black grains smaller than a micron in size are likely iron or sulfur granules (or perhaps FeS_2 particles) deposited either as an inorganic precipitate or as a result of microbial metabolic activity. Additional EDS mapping is required to confirm the elemental composition.





5. 23a.

5.23b

5.23c

Figure 5.23 a-c High resolution WDS microprobe mapping reveals the distribution of S, Ca, Fe, and Si at the boundary between the very pure, highly organized calcite crystals.

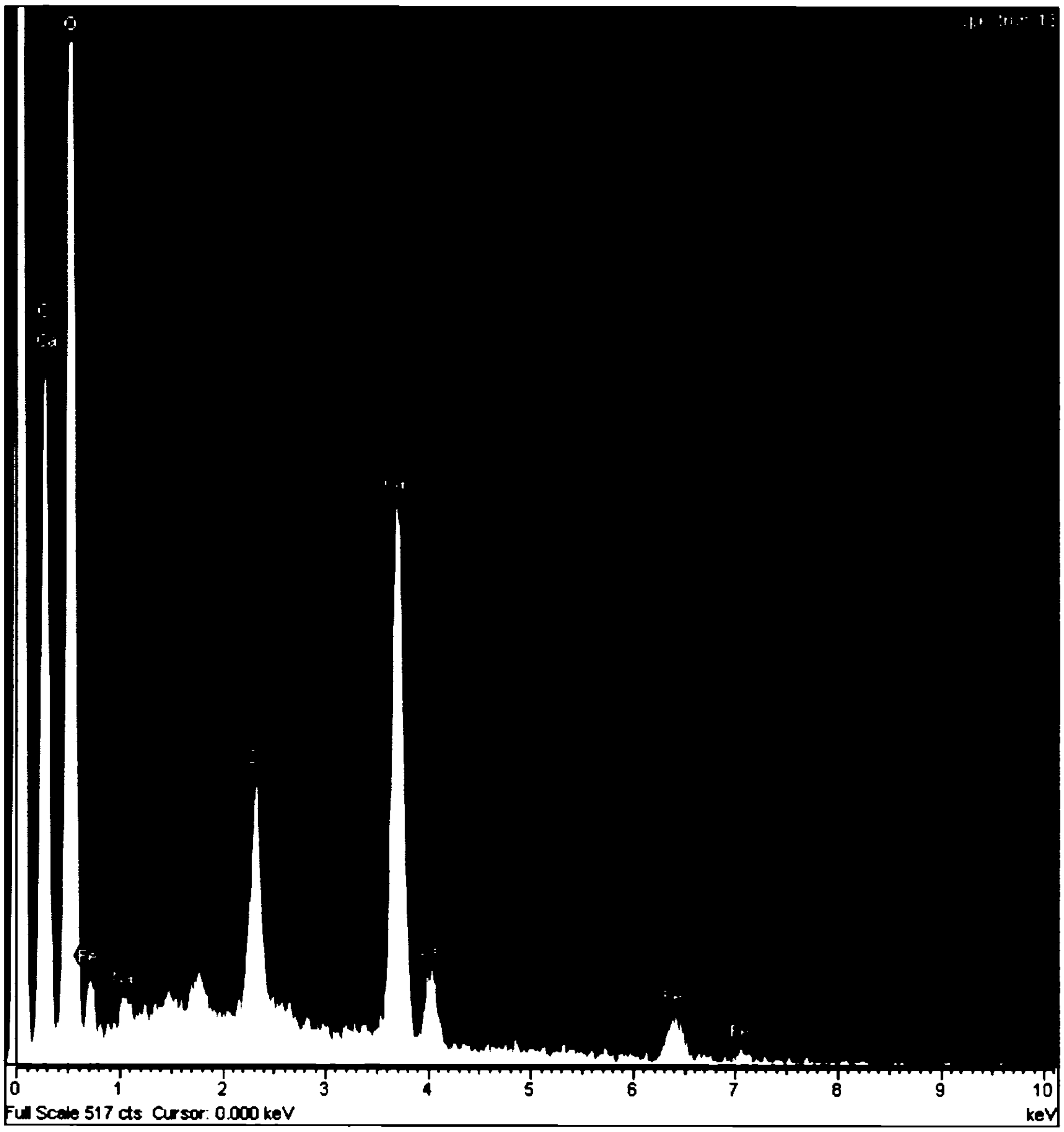


Figure 5.24 SEM EDS analysis of micritic region. Sulfur and iron are present; however the iron peaks are somewhat diminutive relative to what might be expected if pyrite is present. Additional sample analysis, including EDS mapping across the micritic boundary, will be required to determine with certainty what the granules visible under light microscopy are composed of.

Figure 5.25 Close-up of metastable calcite crystals forming in rimstone pool seen in figure 5-14.

Figure 5.26 The rimstone pools, originally filled with the large white carbonate crystals undergo change to the more compact, dark setting seen above. The botryoidal shape seen in the flow channels becomes pronounced as crystals form on top of and around topographic highs and lows. During the summer organic material such as leaves (note the leaf in the foreground) may be incorporated into the carbonate as well.

Large (up to 0.5 cm in length) metastable calcium carbonate crystals form rim pools in quiescent low-flow areas of the springs (Figure 5.25). These crystals have only been observed in early April when local air temperatures are well below freezing. Omelon *et al.* 1999 argues that these crystals are a low temperature form of calcite known as ikaite. The initial appearance of the crystals is white with little to no debris embedded within the matrix. Over the course of several weeks the surface of the crystals takes on a bluish-green tint ranging from a bright green to eventually a dull, dark green to nearly black. Using light microscopy, the staining is visible as a green band bleeding away from the micritic area between layers of calcite crystals. It is not clear what causes the color change. Microscopic observations of fresh samples revealed no prokaryotes such as cyanobacteria colonizing the surface and subsequent *in situ* surveys for chlorophyll using a PAM fluorometer revealed that chlorophyll was completely absent over a several hundred square meter area of travertine. Additional extractions of calcite and water for chlorophyll were also negative. The only chlorophyll bearing cells to be found were bits and pieces of detritus that had blown onto the surface of the carbonate but these were relatively few and far between.

The springs erupt through a layer of marine clay and one possibility is pigment staining from the local clay that eventually comes into contact with the carbonates. Another interesting possibility that will require further examination is that the green tint imparted to the surface of the carbonate is a result of the formation of green rusts. Green rusts are mixed Fe(II)/Fe(III) hydroxides belonging to the sjögrenite-pyroaurite class of minerals (Hansen 1989; Genin *et al.* 1996) that are found in many suboxic environments. Green rust along with a suite of other very fine-grained minerals such as ferrihydrite, lepidocrocite and goethite may form as water with Fe(II) moves from a reducing environment into an oxidized setting. Green rust, built of brucite-type layers of Fe(II)Fe(III)–hydroxide, has anions such as sulfate, chloride, carbonate and water filling the interlayers. It is thought that the formation of green rusts may be mediated by the reduction of Fe(III) oxyhydroxides by dissimilatory iron-reducing bacteria (DIRB) (Frederickson *et al.* 1998). Understanding mineral transformation reactions initiated by a consortia of DIRB will provide a better understanding of the formative stages that underlie the development of the cool water carbonates present at Colour Peak. Blast results from DNA extracted from the inert substrates suspended in the spring flow show the presence of organisms with a 94% commonality to *Shewanella putrefaciens* & to *Geospirillum barnesii*. Both of these bacteria are known to be metabolically diverse and have been shown to be capable of mediating FeII/III transformations.

Green rusts are difficult to study because they oxidize shortly after exposure to air, and indeed this is our observation at the spring site. Upon removal from the water the green tinted crystals quickly oxidize and the color changes to the more familiar orange/brown hue. *In situ* mössbauer or RAMAN spectroscopy may prove to the method of choice for its identification (or elimination as an explanation) at the arctic spring site. Future work will also necessitate the identification of organisms that may associated with microbially mediated redox processes involved.

5.5 Conclusions

The springs located at Expedition Fiord (Colour Peak and Gypsum Hill springs) provide a unique opportunity to study a microbial community in an extreme polar environment. The ranges in temperature, pH, redox, nutrient availability and the large seasonal variations in light have undoubtedly shaped the structure and function of the ecosystem as well as impacting the biological record left in the sediments. This study has presented details regarding the diversity of organisms and the physical and chemical environment in which they live. The planet Mars may have once hosted numerous springs in a physical setting not to dissimilar to the high arctic. Studies of terrestrial microbial ecosystems in regions of thick, continuous permafrost will undoubtedly aid in the future search for evidence of life in the permafrost of Mars.

CHAPTER 6: CONCLUSIONS

The primary purpose of this study was to conduct a multidisciplinary research effort to examine the geomorphic, physical-chemical, and biological aspects of perennial springs and associated mineral precipitates located near Expedition Fiord, Axel Heiberg Island, Nunavut. Three guiding hypotheses were developed to focus the work: (1) the thermal regime of the perennial springs is controlled by the local geothermal gradient, not local magmatic hotspots; (2) the source of the water for the springs is local meteoric water derived from either sub-glacial melt, local alpine lakes, or a combination of both; (3) prokaryotic communities associated with the springs are diverse, contribute to the formation of the mineral deposits and are preserved within the sedimentary record. A secondary goal was to apply the results from this work to conceptual models of environmental conditions that may have been present on early Mars with the express purpose to further future exploration of that planet.

Accomplishments were numerous and the fundamental questions that were posed by the project were satisfactorily answered. A combined flow and thermal model of the subsurface flow associated with the springs at Expedition Fiord using the locally measured geothermal gradient was developed to explain how such springs sustain themselves in regions of thick permafrost on Earth and possibly on Mars. It is interesting to now note that additional insight regarding the role of climate and the initiation of the springs has resulted from the original work on the thermal regime. It has been puzzling how the springs actually were initiated. The flow model helps us to understand how the springs are maintained in essentially a steady state. It appears that the formation of a through talik is largely controlled by two important contstraints: the local mean annual temperature and the eutectic point of the solution moving upward towards the surface. For the brines at Expedition Fiord the eutectic point is around –25°C. The ideal NaCl solution is –21.2° C however additional salts depress this temperature somewhat. As long as the local mean annual temperature remains warmer than the eutectic point of the salts, the subsurface water will have a way to stay liquid and dissolve its way upward through the permafrost. If the temperature falls below this, springs would not form unless there was some other hydrothermal activity to provide the necessary heat.

Measurements of the dissolved gases in the spring outflows have constrained the origin and age of the spring water as well. The ratio's of the gases and the depletion of neon suggests that the water is derived from a mixture of glacial ice melting at the base of the glaciers and lake water or other surface runoff that is in air equilibrium. Future work will still be required to ascertain with any certainty the age of the groundwater. Results from the gas analysis indicate the water is probably pre-bomb (more than 50 years old) and possibly older given the elevated levels of helium (~500 times the atmospheric level). This age estimate is poorly constrained and will require refinement. New techniques using onsite gas stripping as a means to enrich krypton-81 are now available and recent developments in Atom Trap Trace Analysis (ATTA) are making this measurement practical (Sturchio et al. 2003).

The springs are now seen as more than just an outflow of salt water onto the permafrost. A robust community of microorganisms inhabits the water and upon the surfaces over which the anoxic brine flows. Initial samples that were extracted for DNA have shown that the system is host to a large number of functional groups and most likely novel classes of organisms both psychrophilic and psycrotrophic. A surprising find was the near total lack of cyanobacteria on the surfaces of the carbonate precipitates. It is still not clear why they are absent, but the combination of low redox, high salinity and high mineralization rates may limit their ability to grow there. Whether or not the microorganisms present are involved in the mineralization that is taking place at the springs is not clear. The lack of microorganisms such as cyanobacteria results in a mineral precipitate that is largely devoid of direct evidence of microbial activity. Again, it could be that small size and high sedimentation rates prevent the attachment and eventual entombment of large numbers of microorganisms within the mineral matrix. Comprehensive studies of the micritic region of the travertine may provide the answers. The small (sub micron) particles of iron and sulfur deposited along crystal boundaries may provide the best chemical evidence of life's interaction with the mineralization process. Detailed isotopic analysis may indicate that inorganic processes dominate the system or that the presence of microorganisms affects the distribution and chemical composition within this boundary layer between the large, structured calcite crystals. This has important implications for future efforts for the detection of life on Mars. In the absence of indisputable microfossils, indirect evidence will be required to decipher the sedimentary records on Mars. Unlike other hot springs (such as Yellowstone) that leave very distinct biomarkers within the carbonates, the springs at

Expedition Fiord tell us that the search for evidence of life in the sedimentary rocks of Mars may not be so easy.

Recent (July 2004) reconnaissance flights over Axel Heiberg and Ellesmere Islands have revealed another example of a perennial spring at Lost Hammer Diapir. Future flights in the Arctic Archipelago and perhaps the use of satellite imagery will undoubtedly reveal other examples. However we can be fairly certain that the term 'rare' will be a common trait among them all. This research will continue. The role microorganisms play, if any, in mineralization has yet to be discovered. A systematic and detailed study of the microbial ecosystem present at the springs needs to be undertaken. In addition to continued identification and cataloging of the microbes, their physiological types and what role they may play in biomineralization, the physical setting at the spring sites requires furthur study. Geophysical measurements of the subsructure and uplift rates of the diapirs would provide data that would help resolve the formative processes involved with the initiation of the springs.

This project is the first in depth study of the perennial springs located at Expedition Fiord. The multidisciplinary approach and the backdrop of searching for life on another planet drew together a multitude of other students and scientists. This study has provided the intellectual framework for a science program that included other graduate and undergraduate theses, for example C. Omelon (MSc) and J. Heldmann (Ph.D.) and the publications arising from these theses (six manuscripts). Results of this work have also appeared in two book chapters that use the ideas and data presented for background material about life in extreme environments and the search for life on Mars (McKay et al. 2004, Doran et al. 2004). This study represents the first and possibly only study to date that systematically analyzed high arctic perennial springs as a biophysical analog of Mars. This work will help refine the search for life beyond the Earth and perhaps one day the remnants of a similar set of springs and its biota will be discovered on Mars.

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APPENDIX

Blast Results for Colour Peak artificial substrate

clone names	RFLP abundance	division	closest blast hits, decreasing scores	notes
Colour Peak 2, 5, 6, 7, 8, 10, 11, 18, 22, 25, 28, 36, 37, 38, 39, 40, 41, 42, 43, 47, 48, 49, 53, 54, 55, 59, 60, 61, 62, 64, 65, 70, 71, 73, 74, 76, 78, 82, 87, 90, 92, 94	42/55 = 76%	gamma proteobacteria	97% to *Thiomicrospira* sp. Art-3	sulfur-oxidizing, isolated froma salt spring Artern Thuringia
			97% to *Thiomicrospira* sp. JB-A2	sulfur-oxidizing, isolated from an intertidal mud flat
			96% to *Thiomicrospira* sp. L-12	isolated from Galapagos rift hydrothermal vent
			96% to *Thiomicrospira* sp. Ch-1	isolated from the continental shelf near Concepcion, Chile
			95% to *Thiomicrospira* sp. Strain L12	sequence from gill symbiont of Thyasira flexuosa
			95% to *Thiomicrospira crunogena*	from Mid-Atlantic ridge hydrothermal vent sample
			94% to *Thiomicrospira* sp. Strain Milos T-1	sulfur-oxidizing, from shallow hydrothermal vent in the Aegean Sea
Colour Peak 1	1/55 = 2%	gamma proteobacteria	94% to unid bacterium BPCO3	hydrocarbon seep sediment
			91% to *Methylobacter* sp. BB5.1	methane and trichloroethylene oxidation, estuarine methanotroph
			92% to *Solemya reidi* endosymbiont	sulfur-oxidizing chemoautotrophic endosymbiont
			91% to *Escarpia spicata* endosymbiont	sulfur-oxidizing chemoautotrophic endosymbiont
			90% to *Anodontia philliiana* endosymbiont	sulfur-oxidizing chemoautotrophic endosymbiont
			90% to *Pseudomonas iners*	
			91% to *Methylomicrobium* sp. NI	oxidation of trichloroethylene and dimethyl sulfide
			94% to *Shewanella putrefaciens*	halotolerant, metabolically diverse
Colour Peak 63	1/55 = 2%	gamma proteobacteria	98% to unid bacterium JTB148	deep sea sediment, Japan Trench
			96% to unid bacterium Sva0091	cold marine sediments, Norway
			95% to unid bacterium BPC022	hydrocarbon seep sediment
			95% to unid bacterium Sva0854	cold marine sediments, Norway
			92% to *Ridgeia piscesae* endosymbiont	sulfur-oxidizing endosymbiont

Colour Peak 86	1/55 = 2%	gamma proteobacteria	95% to unid bacterium BPC036	hydrocarbon seep sediment
			91% to *Methylobacter* sp. BB5.1	methane and trichloroethylene oxidation, estuarine methanotroph
			91% to *Escarpia spicata* endosymbiont	tubeworm endosymbiont
			91% to *Anodontia philiipiana* endosymbiont	sulfur-oxidizing endosymbiont
			90% to *Pseudomonas iners*	
Colour Peak 79	1/55 = 2%	gamma proteobacterium	90% to unnamed purple bacterium Thd2	anaerobic oxidizer of FeII by photosynthesis, AEM 60(12) p. 4517
			90% to *Methylococcus capsulatus*	thermophilic acetate-oxidizing and sulfur-reducing organism, distinct lineage in proteo?
			90% to *Halomonas desiderata*	alkaliphilic, halotolerant denitrifying bacterium isolated from municipal sewage
			90% to *Alkalispirilum mobilis*	non-phototrophic member of Ectothiorhodospira family, salt-tolerant
			90% to *Thiocystis gelatinosa*	
			90% to *Chromohalobacter marismortui*	moderately halophilic bacterium
			90% to *Rhabdochromatium marinum*	purple sulfur bacterium from a salt marsh mat
Colour Peak 13	1/55 = 2%	distinct group in proteobacteria with *Thiobac. hydrothermalis* and *Thiobac. halophilus*	98% to *Thiobac. hydrothermalis*	mesophilic obligately chemolithoautotroph, from deep-sea hydrothermal vent in Fiji Basin
			97% to *Thiobac. halophilus*	Arch Micr 166(6)p 394
			92% to *Thiobacillus* sp.	autotrophic sulfide oxidizer (to sulfur), from sulfidic waste reactor
			90% to *Ectothiorhodospira* sp. "Bogoria Red"	bacteriochlorophyll b containing, unpublished
			89% to *Allochromatium vinosum*	
Colour Peak 81	1/55 = 2%	delta proteobacterium	97% to *Desulfocapsa thiozymogenes*	disproportionation of inorganic compounds, sulfate-reduction, Arch Micro 166 p. 184
			97% to *Desulfocapsa* sp. 113	sulfate-reducer, from meromictic Lake Cadagno, Switzerland
			97% to *Desulfocapsa* sp. 348	sulfate-reducer, from meromictic Lake Cadagno, Switzerland
			97% to *Desulfocapsa* sp. 282	sulfate-reducer, from meromictic Lake Cadagno, Switzerland
			97% to *Desulfocapsa* sp. 386	sulfate-reducer, from meromictic Lake Cadagno, Switzerland
			97% to *Desulfocapsa* sp. 167	sulfate-reducer, from meromictic Lake Cadagno, Switzerland
			95% to *Desulfocapsa sulfoexigens*	sulfur and thiosulfate disproportionation, isolated from marine surface sediment

Colour Peak 20, 31, 35, 56	4/55 = 7%	epsilon proteobacteria	95% to *Sulfurospirillum arcachonense*	microaerophilic sulfur-reducing bacterium, IJSB 47(4) p. 1212
			95% to *Campylobacter* sp.	(unpublished)
			95% to *Dehalospirillum multivorans*	tetrachloroethene-utilizing strictly anaerobic bacterium
			94% to *Geospirillum arsenophilus*	(unpublished)
			95% to unid bacterium ACE-24	Antarctic maritime lake and fjord benthic zones
			94% to *Sulfurospirillum deleyianum*	microaerophilic sulfur-reducing bacterium, IJSB 47(4) p. 1212
			94% to *Geospirillum barnesii*	dissimilatory FeIII-reducing bacterium
Colour Peak 51	1/55 = 2%	Gram Positive-*Bacillus/Clostridium*	99% to *Haloanaerobium praevalens*	strictly anaerobic halophile isolated from a can of Swedish salted herrings
			98% to *Haloanaerobium saccharolyticum*	fermentative halophilic anaerobe, Anaerobe 1 p. 185
			98% to *Haloanaerobium fermentum*	strictly anaerobic fermentative halophile isolated from fermented puffer fish ovaries
			98% to *Haloanaerobium acetoethylicum*	anaerobic halophilic fermentor
Colour Peak 66	1/55 = 2%	gamma or epsilon proteobacterium	96% to unid SRang1.25	bacterial community in Sulfur River, Parker Cave, KY
			96% to unid SRang1.23	bacterial community in Sulfur River, Parker Cave, KY
			95% to unid bacterium PVB_OTU_2 clone 15	microbial mat in a hydrothermal vent system, Hawaii
			95% to unid bacterium PVB_OTU_2 clone 7	microbial mat in a hydrothermal vent system, Hawaii
			95% to *Thiomicrospira denitrificans*	deep-sea hydrothermal vent sample
			95% to unid bacterium PVB_OTU_3 clone55	microbial mat in a hydrothermal vent system, Hawaii
			94% to unid bacterium Phenol-1	sulfidogenic 2-bromophenol-dehalogenating and phenol-degrading consortium
Colour Peak 93	1/55 = 2%	unaffiliated to any known groups	88% to unid bacterium ACE-29	Antarctic maritime lake and fjord benthic zones
			88% to unid bacterium BD2-16	deep-sea sediments
			88% to unid bacterium vadinBA30	anaerobic digestor clone from wine distillery waste
			88% to *Pirellula* sp.	planctomycete