Sea Ice

Physics and Remote Sensing

Mohammed Shokr and Nirmal K. Sinha

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Geophysical Monograph 209

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Cover image: Photograph of sea ice in the North Water Polynya in the Arctic, viewed from the deck of the Canadian Coast Guard icebreaker *Pierre Radisson* in April 1998. Thin and deformed ice along with an open-water area are shown in the foreground and part of the Ellesmere Island (Canadian Arctic) appears in the background. The picture was taken by M. Shokr during the field program NOW'98.

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To the two most caring women in my life, my mother and my wife; my three sons who taught me more than I have ever taught them; and to all those who inspired me to finish this book knowing that they will not read it.

Mohammed Shokr

If I was successful as a scientist, it was only because of the unconditional support I received from my wife, Supti Sinha, and my three precious daughters, Priya, Roona, and Shoma, who often helped me in my cold laboratory. They were born while I was camping on an old floe near the north pole during long, long absences from home.

Nirmal Sinha

CONTENTS

PREFACE

During the winter months sea ice plays the most important role in the daily life of coastal communities of the cold regions of the Earth. As for the Arctic regions, especially Canada and Russia, the general trend in its retreat is leading to new opportunities for the extension of the navigational season and exploitation of natural resources. Growing interest in sea ice for socioeconomic purposes has also increased awareness of its role in affecting the global environment, the climate system, and most importantly the human aspects. The advent of remote sensing techniques from Earth observing satellites, as the major tool for obtaining global record of sea ice, is making it necessary to reexamine sea ice as a material with updated knowledge and approaches.

In the customary description of snow, fresh-, and seawater ice, often the basic sciences and measurement techniques used are assumed to be known yet not covered or explained in details. These are mostly related to the thermodynamic high-temperature state (being close to the melting point) of these materials, physics of solidification governing the development of microstructural features especially in sea ice, the impact of mobility and aging of sea ice on its geometrical and physical properties, and last but not least the birefringent properties of ice crystals that are used extensively for revealing internal substructures. By the same token, the customary description of the applications of remote sensing usually leaves out details of the electromagnetic wave interaction with ice and its snow cover, particularly the impact of the snow on the measured reflection, radiation, and radar scattering. This book is written with the hope of filling these gaps and to making a bridge between the physics and the remote sensing communities. This is spot-on whether the readers are working on sea ice in the two primary polar cryospheres or on snow/ice of the secondary cryospheres of Himalaya and other mountain ranges.

In many ways this is an entirely new book on natural floating ice in oceans with an emphasis on sea ice. The goal is to describe and explain the principles of physics and chemistry of sea ice and its space-borne observation using a suite of remote sensing imaging systems. The reader needs only to flip through the pages to note that the general appearance related to the format and illustrations are different from any technical book. The subjects are treated with a multidisciplinary and to some extent transdisciplinary approach. Underlining assumptions are that the reader/user will come with wide ranging engineering backgrounds — civil, mechanical, environmental, electrical/electronic, etc. or sciences — physics, chemistry, geography, oceanography, climatology, and even social sciences and anthropology related to human settlements. The style used, therefore, is to describe complex subjects in an understandable way so that the reader will not only gain a working knowledge of basic science related to the area of study, but also applications of ice information retrievals using remote sensing tools. It is also expected that the reader will cultivate an appreciation of the human aspects required to carry out the investigations and, of course, the real-life operational issues encountered in shipping, fishing, etc.

Design requirements for the construction of a few marine projects in ice-rich waters led the National Research Council (NRC) of Canada to start a program on sea ice engineering in the early 1970s. Among these projects were (a) the world's most northerly dock at Nanisivik mines in Strathcona Sound in Baffin Island, Canada, in 1975, (b) the world's first self-sustained ice breaking ore carrier, *MV Arctic*, to ship ores from Nanisivik to Europe, (c) man-made artificially thickened sea ice islands as drilling platforms for oil and gas exploration within the Canadian Archipelago, (d) offshore drilling activities in the Beaufort Sea and Hibernia off the coast of Newfoundland in the Atlantic Ocean, and (e) extension of navigational season in the eastern Arctic with the help of Side Looking Airborne Radar (SLAR) remote sensing. The second author of this book considers himself as lucky when opportunities emerged for him to get involved in all these Arctic R&D activities after he joined NRC in January, 1975, and also to get involved in sea ice preparation work for the Canadian Radarsat project since its beginning in 1981.

The advent of satellite remote sensing and its applications to sea ice in the 1970s fulfilled a wish by some of the pioneers of ice research in the 1950s. That was to develop instruments suited for recording the evolution and dynamics of sea ice in the polar environment. Space-borne sensors operating in the visible, infrared, and passive microwave emerged, but it wasn't until the introduction of the synthetic aperture radar (SAR) in 1978 onboard NASA's Seasat satellite when the sea ice community realized how revolutionary this tool would be for ice reconnaissance and parameter retrieval at fine resolutions. Work started in Canada in the 1980s to study the feasibility and then to develop the first Canadian SAR mission on board Radarsat-1 satellite. That opened many opportunities for individuals from different disciplines to join the interesting field of remote sensing of sea ice. The first author of this book also considers himself as lucky when opportunities emerged for him in Environment Canada, starting in 1988, to get involved in many projects to demonstrate the applications of imaging radar data in support of the Canadian sea ice monitoring program. Later he expanded his experience by developing applications using passive microwave and other remote sensing data categories. Opportunities also emerged for him to participate in field expeditions in the Arctic and the east coast of Canada and be in direct touch with the fascinating world of sea ice.

This book is intended to summarize experiences acquired mainly in Canada by researchers and operators to advance the sea ice research and operational monitoring programs. It is also intended to reach out to a variety of sea ice audiences interested in different aspects of ice related to physics, mechanics, remote sensing, operational monitoring, climatic impacts, etc. It is useful to briefly describe here the level of the book. Chapters 2–6 describe ice physics comprehensively to anyone with a high-school background in physics, chemistry, and mathematics. Chapters 7–10 are devoted to remote sensing by active and passive imaging sensors with an emphasis on microwave systems, currently used in many countries around the world for operational ice monitoring. Some topics may be considered as advanced. Chapter 11 describes the role of ice services provided by Environment Canada with a brief historical account of the development of Canada's interest in the Arctic and the ice service in Canada. The chapters are not free standing. Instead, links are established as much as possible between information about the different themes of the chapters.

M. Shokr and N. K. Sinha

ACKNOWLEDGMENTS AND RECOGNITIONS

When the first author started his collaboration with the Canadian Ice Service (CIS) in 1989 as a scientist employed by Environment Canada, the staff was very supportive and welcoming of his input in processing the remote sensing images of sea ice. More importantly, they granted him many opportunities to learn about sea ice and the operational ice monitoring program. The author would like to express his deepest appreciation to John Falkingham, Bruce Ramsay, Terry Mullane, Mike Manore, Dean Flett, Matt Arkett, and Dr. Roger De Abreu. Ken Asmus provided great help in fieldwork and laboratory experiments on sea ice. The first field program that the author joined to study Arctic sea ice was the Sea Ice Monitoring and Modelling Site (SIMMS) from 1990 to 1997. It was an inspiring experience that led the author to develop passion for sea ice and the cold region environment. The program was initiated by Dr. David Barber. The author would like to extend his appreciation also to Dr. Simon Prinsenberg of the Department of Fisheries and Ocean Canada for his support during field experiments in the Labrador Sea and to the team of the North Water polynya experiment for their support during field campaigns. Dr. Venkata Neralla and Roop Lalbeharry of Environment Canada have inspired the first author to complete the book. Thanks also go to Dr. Shawn Turner for his effort in reviewing parts of this book. Last but not least, the author would like to express his gratitude to Maria Latyszewsky, the chief librarian, and Lindsay Hall-Carvalho of the Environment Canada's Library in Toronto who facilitated many requests for material during the course of writing this book.

The second author would like to convey his sincere appreciation to innumerable persons who helped him in the field and to recognize the financial, technical, and strategic supports provided by several Canadian and international organizations since 1975. He is indebted to the captains and staff of several icebreakers of the Canadian Coast Guard, *North*, and the ice breaking ore carrier, *M.V Arctic* of Canarctic LTD.

George Hobson, director of Polar Continental Shelf Project (PCSP) of Energy Mines and Resources (now NRCan), from 1972 until 1988, provided moral, financial, and strategic support to the author for almost 10 years (1977–1986). This allowed the author to train an Inuit team who helped him carry out the year-round investigations on growth, structure, and engineering properties of sea ice at Eclipse Sound near Pond Inlet in Baffin Island. Special thanks are due to M. Komangapik, S. Koonark, and S. Koonoo for their efforts in collecting scientific data in the harsh climatic conditions of the Arctic and to D. Komangapic, S. Akoomalik, and J. Arnakallak for performing tedious measurements in the Pond Inlet field laboratory and tabulating the results and to I. Kilukishak and M. Komangapik for laboratory assistance in Mould Bay. The author wishes to express his sincere thanks to Hermann A.R. Steltnar and his wife Sophie Steltnar of Arctic Research Establishment (ARE) at Pond Inlet, for organizing and managing the field activities of the Inuit team, compiling the field data, and very importantly for providing the author with board and lodging during all of his trips to Pond Inlet.

For studies related to multiyear (MY) sea ice rubble field at the Hobson's Choice ice island, the author is obligated to Mike Scmidt, the PCSP camp manager, for providing all the logistical support and Tom Krochak's kitchen for supplying an endless line of food trays 24 h and for week after week. He would like to convey his sincere appreciations to Drs. Ed Stander and Paul Barrette, respectively, for their assistances in conducting the microstructural analysis of MY and Ward Hunt Shelf Ice.

Drs. Lorne Gold, Robert (Bob) Frederking, Robert Gagnon, Garry Timco, Mohammed Sayed, and Mary Williams of NRC provided encouragements, guidance, and useful critical comments for more than three decades. In fact, Bob Frederking introduced the author to the world of sea ice engineering. He led the author, literally by hand, to his first field experience at Strathcona Sound, Baffin Island, during the extremely cold polar environmental conditions of 24 h of dark "polar nights" of November–December 1975. Due to the unavailability of any remotely sensed images, the only view of sea ice the author had during this mission was whatever was visible under the illumination of headlights of trucks positioned on the newly constructed dock. The truck had to be turned to get different points of view. Ice cores were taken by suitably pointing the headlights to the spot for sampling.

Very special regards go to Dave Wright of the Division of Building Research (DBR) [renamed Institute for Research in Construction], NRC, who actually taught the author the refined techniques of "hot-plate melting" for making near perfect thin sections of "freshwater ice," fabricating a small field polariscope with built-in light source, and working with him during the first few, but crucial field trips in the High Arctic in 1976–1977. Ron Jerome of DBR-NRC provided technical assistance in almost all phases of laboratory and field investigations. This included the design of the large-field polariscope, 100 mm diameter "feather-weight" fiber-glass ice core auger, the NRCC borehole indentor (BHI), and the portable cold laboratory used at Pond Inlet, Resolute, and on board of icebreakers. The author is grateful to L. J. Mulvihill for fabricating the fiber-glass core auger and stainless-steel BHI to the extremely close tolerances required. He is indebted to Donald Martin and the staff of the National Capital Commission (NCC) of Ottawa for securing the entire Dow's Lake and providing technical assistance during the days of testing for the evaluation (and modifications) of these tools over several years, before they were taken to the Arctic.

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On the subject of extending the double-microtoming technique, originally developed for sea ice to investigate the microstructural aspects of snow morphology, the author is thankful to Dr. Pramod Satyawaliand Maj. Gen. Satya Sharma of Snow and Avalanche Study Establishment (SASE), India. He is also indebted to Dr. Alex Klein-Paste of the Norwegian University of Science and Technology (NTNU), Trondheim, and Armann Norheim of the Norwegian Civil Aviation (AVNR), Oslo, for refining the methods applying to snow on runways.

Vance Garbowski and Howard Massecar, helicopter pilot of Vancouver-based Quasar, provided field supports in Mould Bay and Baffin Bay. The author is indebted to Howard for rescuing and saving his life on 22 June 1984 in northern Baffin Bay during a tsunami storm that shattered the 1.5 m thick sea ice sheet and forced the author, David Lapp, and an Inuit polar-bear watcher to take shelter on a bergybit. Otherwise, the tales of this author with respect to the human aspects of Arctic research would not have been told in this book.

1
Introduction

1.1. BACKGROUND

Our world is divided into five regions according to the position of the Sun throughout the year: a tropical region around the equator, two temperate regions and two polar regions. On two equinoxes, March 21 and September 23, the Sun is directly overhead the equator and the Sun's rays reach both the North and the South Poles. On June 21 (the summer solstice), the Sun is directly overhead the Tropic of Cancer (about 23.5°N) in the northern temperate region, and on December 22 (the winter solstice) it is positioned directly overhead the Tropic of Capricorn (about 23.5°S) in the southern temperate region. In the two polar regions, mostly relevant to the material in this book, the Sun never sets in their summer and never rises in their winter. The Arctic region (or zone) containing the north polar region with latitudes greater than "about" 66.6°N and the Antarctic region (or zone) containing the south polar region having latitudes greater than "about" 66.6°S are the primary cryospheric regions of the world. Although the latitudes of the Arctic and the Antarctic circles depend on Earth's axial tilt, which fluctuates slightly with time (about 2° over a 40,000 year period), the variations in the boundaries of the polar region are very small and negligible. The secondary cryospheric regions are the Alps, Andes, Himalaya, Rockies, etc. Among the secondary cryospheric regions, the Himalayan belt covers and affects the largest effective area of human habitation.

Climate change has been affecting all the cryospheric regions of the world, and the effects can be directly observed and quantified using air- and space-borne remote sensing as well as land-based instruments. Remotely sensed images of the land- and ocean-based snow and ice information are paramount in understanding the state of health of Earth for sustainability of life. Other methods such as ice core analysis of ice caps and ice shelves are also used. After all, snow is the messenger of the sky and the environment.

Sea ice covers most of the oceanic surface of the primary cryospheric area of the global surface. The world of sea ice encompasses the polar region, particularly the Arctic basin and a belt around the continent of Antarctica. Out of the 71% of Earth's surface that is covered by ocean about 7%–15% is covered by sea ice at certain times (more in the winter and less in the summer). That is equivalent to 5%–10% of Earth's surface. Sea ice area in the Arctic varies between a minimum of about 4 million km^2 in September to a maximum of about 15 million km^2 in March. The corresponding figures for the Antarctic are 3 million and 18 million km^2 in February and September, respectively. Sea ice can develop very smooth or very rough surfaces. It can be soft or hard, a bare surface or snow covered, stagnant (fastened to the shoreline) or mobile pack ice, and stiff and silent or crushing with loud noise. It exhibits seasonal variations to which life in the polar regions is closely adapted. In the Arctic region, sea ice starts its growth in September/October and reaches its maximum in March when it covers the entire Arctic basin. This trend is reversed during the summer, and the ice extent reaches its minimum in September. In the Antarctic the annual fluctuations range between a minimum in February to a maximum in September when ice extends to latitudes between 55° and 65°S.

For a limited time during the summer months, certain areas of the polar waters in the Arctic zone are used extensively by ships (ice strengthened or escorted by icebreakers) where the floating bodies of new and old sea ice and icebergs can prove hazardous. The expected reduction of sea ice extent, the reduction of the navigationally hazardous old ice, and the increase in the duration of summer melting season will certainly increase marine

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activities in these areas. No doubt, the Arctic waters, particularly the legendary Northwest Passage that passes through the Canadian Arctic Archipelago and the Beaufort Sea, will be used more in the future for shipping goods between Asia, North America, and Europe.

Average sea ice covered area in either the Arctic or the Antarctic is about the same (about 15 million km^2) during the winter. However, because the mean thickness of sea ice is 3 and 1.5 m in the Arctic and the Antarctic, respectively, the maximum volume of the sea ice cover in the Arctic (about 0.045 million km^3) is nearly twice that of the Antarctic. In summer, ice extent shrinks significantly to about 50% of the winter coverage in the Arctic. Nearly 90% of the sea ice coverage disappears by the end of the summer in the Antarctic. Ice that melts completely during the summer is called "seasonal ice" or "annual ice." If the ice melts only partially, then the part that survives until the next winter and growth season is called "perennial ice." This can be second-year or multiyear ice depending on how many summers the ice has survived.

As a major component of the cryosphere, sea ice influences the global ocean and atmosphere in a profound manner. Its continuous interaction with the underlying oceans and the overlaying atmosphere leaves major impacts on weather, climate, and ocean current systems. Moreover, ice in one form or the other plays a significant role in the daily life of communities inhabiting the cold regions of Earth. Sea ice in particular influences the coastal areas in most of the circumpolar nations of the Northern Hemisphere. It affects to a lesser extent a few countries in the Southern Hemisphere. Of all the countries of the world, Canada has the longest coastline and has the largest reservoir of freshwater lakes and rivers with floating ice in them annually at least for half of the year. Except for Alaska, practically all the areas north of the 49°N in North America belong to Canada. While sea ice plays a major role in areas above 60° (north or south) it does not affect areas below that latitude except in the Hudson Bay, Labrador Sea, and the Gulf of St. Lawrence in Canada and to a relatively lesser extent in the Baltic Sea, Gulfs of Bothnia and Fin in Europe, the Sea of Okhotsk, north of Japan, and Bohai Bay in China. Above about 35°N in Eurasia and North America, most of the streams, rivers, and lakes (Black Sea, Sea of Azo, and Caspean Sea in Eurasia and the Great Lakes, to name a few among thousands, in North America) have some ice cover each winter. In fact, severity of winters in North America is often measured in terms of ice coverage of the five Great Lakes (Lake Superior, Lake Michigan, Lake Huron, Lake Erie, and Lake Ontario).

In spite of the fact that sea ice covers vast areas of sea surface of Earth, most of the people of the primary cryospheric regions of the world have not seen it or are even aware of it. That is because most people, even within the cold regions of Earth, live far from the areas affected by sea ice. Other than a few thousand multinational scientific observers and a few annual visitors, nobody lives in the south polar zone (beyond the Antarctic Circle). Only a few small communities of the Falkland Islands and Argentina consider the Antarctic region their home. On the other hand, beyond the Arctic Circle in circumpolar areas of Alaska, Canada, Norway, and Russia, perhaps a few million people live. This is incomparable to the nearly 1500 million people living in Afghanistan, Bangladesh, China, India, Nepal, Pakistan, and Tibet who are indirectly affected by the Himalayan cryosphere, but sea ice does not exists in that region.

It is not uncommon for people who live away from the circumpolar boundaries to be confused between sea ice and icebergs. Yet, general awareness about sea ice has been growing as public information about the decline of sea ice in the Arctic with its positive economic impacts and negative environmental impacts is spreading. This book, though not oriented to serve as a popular science document, provides scientific information with explanations that may hopefully expand the domain of interest in sea ice and attract a number of young scientists to pursue studies about its physical aspects as well as its detection using spaceborne remote sensing technologies.

The Arctic basin consists of primarily the Canadian and the Eurasian subbasins [for details on these two basins, see Chapter 3 in *Weeks*, 2010]. It is extremely difficult to obtain sea ice data in these areas because of the remote locations and extreme climate conditions in which ice exists. This situation also applies, perhaps to a lesser extent to ice-rich areas north of Russia because of year-round marine activities in Barents Sea, Kara Sea, Laptev Sea, and East Siberian Sea. Until the beginning of the twentieth century information about sea ice was mainly gathered and used by the local people who lived in the sub-Arctic regions. Later, increasing information was obtained from ship sighting and harbor icing records, but the purpose remained to assist the very limited marine operations. However, since the end of World War II in 1945 and the beginning of the Cold War, there had been a significant increase in human activity in both the polar regions and in particular the Arctic. Numerous weather stations equipped to gather scientific information and military bases with airports and radar lines were constructed in Canada, Alaska, and Greenland. Although some of the supplies for the construction and maintenance of these bases were transported by aircraft, ice-strengthened ships escorted by icebreakers were extensively used during the summer melt season. Submarines and buoys have also been used to gather data for sea ice in the Arctic Ocean.

Although Russia has a longer history of record on measurements of sea ice, dating back several centuries, even the information collected during the last century is not readily available to the rest of the world. However, Russia (actually the Soviet Union) played an important role in increasing the awareness of the Arctic outside Russia. Scientific interest in Arctic sea ice grew fast during the Cold War era as nuclear submarines of the United States and Soviet Union used the Arctic Ocean basin as a prime area to launch ballistic missiles. Funds were made available by the United States for scientists to launch field studies in the Arctic for the first time in the 1950s. In fact, many U.S.-based scientists together with their Canadian counterparts carried out their investigations on sea ice using facilities available in Canada. It is appropriate to mention here that the first English-language book on the physics of ice that covers significant sections on sea ice was written in Montreal, Canada, by *Pounder* [1965] and, incidentally, the first comprehensive English-language book on the physics of glaciers, the source of icebergs, was also written in Ottawa, Canada, by *Paterson* [1969].

The number and scientific scope of the field studies dedicated to sea ice measurements, relevant to spaceborne remote sensing and ground validation of remotely sensed images, peaked in the 1980s and 1990s. That was driven by the need for on-sight field observations and measurements of sea ice characteristics to validate interpretations of images obtained from satellites. Canadian participation was visible in the Arctic during that time. The field campaigns continued after that, though with less frequency and resources allocated to study ice physics. The focus of the work switched to monitor the decline of Arctic ice and its impact on biological and primarily economical aspects of life in the Arctic region.

The first major field program was the Arctic Ice Dynamics Experiment (AIDJEX), which was a collaboration between the United States, Canada, and Japan conducted in 1975–1976 [*Campbell et al.*, 1978; *Coon et al.*, 2007]. Other major experiments included the Marginal Ice Zone Experiment (MIZEX), conducted in 1983, 1984, and 1987 in the Greenland Sea [*Cavalieri et al.*, 1983; *Wadhams*, 1985]. This was followed by the Labrador Ice Margin Experiment (LIMEX), conducted twice in the winters of 1987 and 1989, in the Labrador Sea [*Drinkwater and Digby-Argus*, 1989; and a few papers in the special issue of *IEEE Transactions on Geoscience and Remote Sensing*, vol. 27 no. 5, September 1989]. A series of annual field experiments, called Sea Ice Monitoring and Modeling Site (SIMMS), were conducted in Resolute Bay and surrounding areas of the Canadian Arctic, between 1990 and 1997 [*LeDrew and Barber*, 1994]. The North Water Polynya (NOW'98) was conducted during April–June 1998 to characterize the sea ice coverage in the polynya [*Barber et al.*, 2001]. The Canadian Coast Guard (CCG) ice breaker *Pierre Radisson* was used as the working platform for this project (Figure 1.1). The Surface Heat

Figure 1.1 Canadian Coast Guard (CCG) ice breaking ship *Pierre Radisson* in the North Water Polynya in May 1998 (photo by M. Shokr).

Budget of the Arctic Ocean (SHEBA) experiment, a collaboration project between the United States and Canada, was carried out during the winter season of 1997–1998. During this period, the CCG icebreaker *Des Groseilliers* was frozen in the pack ice for one year and functioned as the base for scientific observations [*Perovich et al.*, 1999]. A wealth of sea ice information was obtained from these field experiments as well as many others. The long duration of the SHEBA experiment allowed the collection of a significant temporal record of ice conditions in relation to seasonal variation. The annual trips of the SIMMS program allowed for the study of interannual variability of ice cover at the same location and time of the year. However, there are limitations to the field measurements, such as the unavoidable point sampling approach to generate measurements that might be representative of the observations from the footprints of satellite data. Moreover, it is not easy to sample thin ice when it is not safe to walk on it. The photograph in Figure 1.2 shows the procedure of cutting a sample of thin ice (about 50 mm thick) using a gangway descending from an icebreaker while the operator is attached to a harness.

Interest in Arctic sea ice by the mid-1990s (after the end of the Cold War) shifted from being military-, security- or offshore-industry-driven to being environmentally driven. New concerns for the region is comprised of issues such as environmental conservation, including nuclear waste and other pollution issues, protecting the livelihood of the Arctic's inhabitants and species, and most importantly identifying sea ice as an indicator and result of climate change. As a result of sea ice being a strong indicator of climate change, sea ice monitoring in the polar regions has triggered an increase in funds to conduct more research at high latitudes in many countries, as well as those who operate in the Antarctica.

Figure 1.2 Sampling of 0.05 m thick sea ice in the Baffin Bay in May 1998 by the author (Shokr) using a gangway lowered from an icebreaker (photo by K. Asmus, Canadian Ice Service).

In the last decade or so, Arctic ice has gained extra attention by the international media and the general public as well as funding agencies because it is shrinking. Antarctic sea ice area, on the other hand, remains essentially the same. Moreover, the Arctic basin connects the Atlantic and Pacific Oceans. Therefore, if the ice diminishes or is replaced by thinner (navigable) ice, a marine navigation route (Northwest or Northeast Passage) may open. There will be a great positive economic impact of this possible scenario, especially for the countries that will have routes under their sovereignty, namely Canada and Russia.

Snow cover plays an important role in the thermodynamic evolution of sea ice. Although it accounts roughly for only 10% of snow/ice volume, some properties differ sharply from their equivalent properties of sea ice. Two familiar examples are the albedo and thermal conductivity. The albedo of dry snow is above 0.9, which is much higher than the albedo of bare first-year sea ice surface (about 0.52). Increased albedo allows less sunlight to penetrate the surface. Therefore, snow-covered ice (with thick and dry snow) receives less sunlight and so does the underlying water. More importantly, the thermal conductivity of the snow is one order of magnitude less than that of sea ice. This means that snow can insulate sea ice and slow down its growth. Therefore, the ice will generally be cooler and that causes a delay of melting in the spring in spite of reduced thickness.

Sea-ice-related information is rather scattered over vast areas of interdisciplinary fields (e.g., physics, chemistry, materials science, remote sensing, climate, oceanography, cryosphere, marine structure and operation, marine biology, and, not the least, civil, mechanical, and naval engineering, related to coastal and offshore engineering). Publications related to the physics and remote sensing are also scattered and not limited to English-language literature. When written in languages other than English, they obviously become not readily available. Some of the familiar books that address physics- and geophysics-related areas of ice include *Pounder* [1965], *Paterson* [1969], *Hobbs* [1974], *Untersteiner* [1968], *Wettlaufer et al.* [1999], *Thomas and Dieckmann* [2009], and Weeks [2010]. Weeks provided a brief, but very appropriate, historical background of the European and other explorers and their explorations of the Arctic and the Antarctic regions in addition to a comprehensive description of sea ice. Books that cover remote sensing of sea ice with some coverage on sea ice physics include *Hall and Martinec* [1985], *Haykin et al.* [1994], *Carsey* [1998], *Jefferies* [1998], *Jackson and Apel* [2004], *Sandven and Johannessen* [2006], *Reese* [2006], *Johannessen et al.* [2007], and *Comiso* [2010]. A few notable review papers on remote sensing of sea ice include *Sandven* [2008], *Breivik et al.* [2010], *Kwok* [2010], and *Heygster et al.* [2012].

1.2. HISTORICAL SYNOPSIS: CANADA AND THE ARCTIC

Except for very small areas of persistent ice-free conditions called polynyas (largest among them is the North Water Polynya) virtually all the northern oceanic areas of the second largest country in the world, Canada, freezes each winter. Historically, however, the Canadian north is divided loosely into High Arctic and Low Arctic regions. The High Arctic used to be the "No man's territory" until about 1958. The High Arctic of Canada is marked by a circle parallel to the latitude of about 70°N and spans between the longitudes of 75°W and 125°W. It is defined as the area consisting of a group of islands known as the Canadian Archipelago. These islands are separated from those of the Low Arctic by a wide channel of seawater, known as the Parry Channel, which includes Lancaster Sound in the east as shown in Figure 1.3. The width of this deep seawater channel varies between 40 and 100 km. It is a hostile environment for human activities because of the mobile pack ice in many areas. The vast area north of this widely varying channel was traveled by explorers from Europe, particularly Norway, only during the late nineteenth and early twentieth centuries. Naturally, the Inuit communities of the polar region of Canada concentrated their movements in the Low Arctic or the region south of the Parry Channel. The present-day settlements

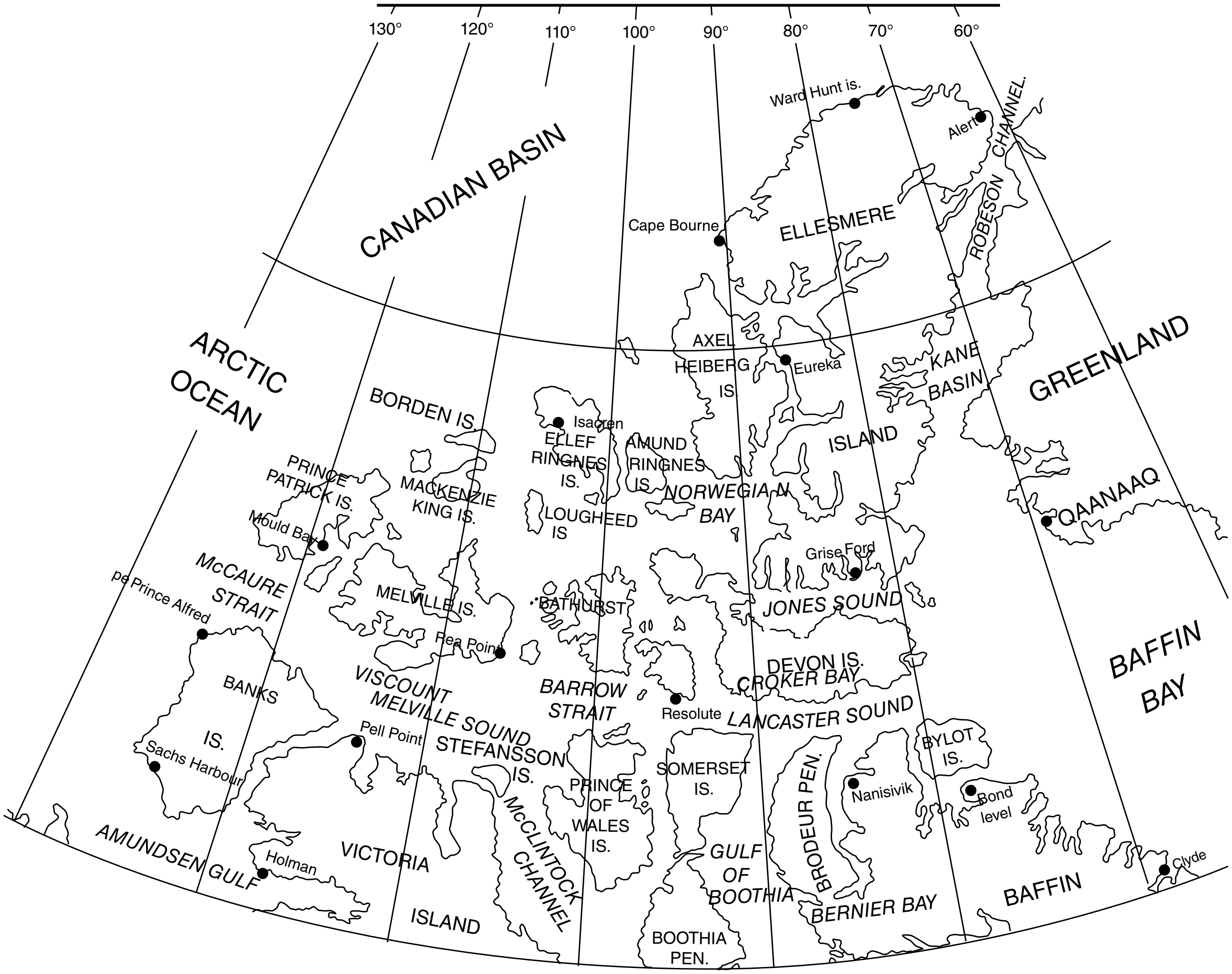

Figure 1.3 Map of the Canadian Arctic Archipelago showing locations of major measurement areas and the weather stations (operational and decommissioned) operated by Environment Canada.

of Resolute (74.72°N, 94.97°W) and Grise Fiord (76.42°N, 82.90°W), shown in Figure 1.3 (and as will be seen below), were established by the federal government of Canada in late 1950s to move Inuit communities to live in the High Arctic but did not succeed as planned. Even today, the Inuit communities of the newly formed territory of Nunavut like to live primarily in the Low Arctic. Nunavut joined the 10 provinces and 2 other territories of Canada on 1 April, 1999.

Historically speaking, the islands of the High Arctic remained isolated from human activities, except for the explorers and the adventurers, until the unrest in Europe spilled over to the north Atlantic. World War II changed the situation significantly. Interest in the Arctic as a strategic and possible economic region was boosted, but even then the climatological observations from the Canadian Arctic were scanty and inadequate for any meaningful analysis. Some geophysical and meteorological data were collected by the explorers, and some useful data could be extracted from records kept by expeditions that attempted to find the North west Passage. These observations were, however, inadequate for accurate climatological studies because information was collected on an opportunity basis over short periods. The records rarely extended over a period longer than a year. Moreover, the observations made at different localities were often made in different years. The results were not comparable because of the absence of continuity in the mode of data collection.

The advent of the Cold War shortly after the end of World War II, led to the creation of strong interests in Canada and particularly in the United States for the establishment of a network of Arctic stations. There was a general recognition that the weather in both Canada and the United States is dominated to a large extent by the Arctic air masses. Accurate forecasting in these two countries requires the detailed knowledge of the weather pattern of the polar region. Consequently, it was realized that observations from the Canadian Arctic would increase knowledge of the circulation of Earth's atmosphere and permit an extension of the period of reliability of weather forecasts. It was emphasized that year-round observation stations have to be established for carrying out regular observational programs directly linked with the investigation of Arctic meteorological problems.

On 12 February, 1946, the U.S. Congress approved the Arctic Project and eventually a suitable basis of cooperation between the governments of Canada and the United States was developed for the Canadian High Arctic. These two countries reached a working agreement on 27 February, 1947, for the establishment and operation of five weather stations in the Canadian High Arctic. This agreement was originally made for a period of 5 years. Consequently, five strategically placed permanent weather stations were built on islands within the Canadian Archipelago. These stations included Alert and Eureka on Ellesmere Island, Isachsen on Ellef Rignes Island, Mould Bay on Prince Patrick Island, and Resolute on Cornwallis Island (see locations in Figure 1.3). Following the first 5 year project, subsequent agreements affirmed that these five stations should continue to be operated jointly by Canada and the United States in accordance with the specifications agreed to at the Joint Arctic Weather Stations Conference, which was held annually. According to this agreement the Atmospheric Environment Services (AES) of Canada (now Meteorological Service of Canada, Environment Canada) provided all permanent installations and approximately half the staff, including an officer in charge who was responsible for the overall operation of the station.

The advancements in rocketry led to the dawn of the space race and made the world smaller and more easily accessible. A sudden thrust in the space race was imposed by the successful launching of the Soviet Union's satellite Sputnik in October, 1957, and directly forced and indirectly enhanced interests in the north polar zone. The launching of the first man-made satellite suddenly proved the world to be small and the polar regions within reach but, more importantly, levied a need for obtaining in-depth knowledge of the terrain in the north beyond the Arctic Circle—the High Arctic and the Canadian basin. In 1958, the United States lost the race for launching the first man-made satellite but succeeded in completing the much-publicized first underwater crossing of the Arctic Ocean using the submarine *Nautilus*. Although the Soviet Union (USSR) was prying the oceans with its fleet of nuclear submarines, no claim was made for any underice activities around the geographic North Pole by the USSR.

In a very direct manner, the underice activities and the space race forced Canada to increase the awareness of the Canadian Arctic Island and implement new measures to strengthen the Canadian sovereignty in all these islands of the High Arctic. This led to the establishment of two permanent human (Inuit) settlements, as mentioned earlier, Resolute at 74.72°N and Grise Fiord at 76.42°N. The United Nations Conference in 1958 on the Law of the Sea, which extended the resource and exploration rights of maritime nations on their continental shelves to a depth of 200 m, acted as the catalyst for Canada to undertake multidisciplinary scientific exploration of the north. Consequently, the Canadian government established the Polar Continental Shelf Project (PCSP) under the ministry of Energy Mines and Resources, or EMR (recently renamed as Natural Resources Canada or NRCan), in 1958. The most important aspects were the geophysical mapping of the High Arctic—areas of the land and the ocean (Figure 1.3) stretching from Alaska to Greenland and from the Arctic Circle to the North Pole. The important field of investigations included recording the magnetic and gravitational data required for the space program.

Nonetheless, the Canadian government decided not to invest heavily in building up its armed forces for the purpose of maintaining Canadian presence all over the Canadian High Arctic. Instead of sending armed forces personnel to the High Arctic, Canada always used a pool of scientists and experienced field workers. Since then the Canadian sovereignty of the High Arctic is essentially maintained by the labor of love of Canadian scientists (as well as their scientific collaborators from other countries). The birth of the new territory of Nunavut in 1999, ranging from mountains and fiords on the eastern shores of Baffin and Ellesmere Islands, through the lakes and tundra of the Barrens on the mainland, to the plateaus and cliffs of the Arctic coast, is changing the course of history of the Canadian Arctic.

Since 1959, the PCSP (Polar Shelf for short) with permanent base camps (shelters or shacks with rudimentary facilities) at Resolute on Cornwallis Island and Tuktoyaktuk in the Mackenzie Delta provided logistic support services to the scientists. This organization provides room and board (actually excellent nourishing food for the body and the soul) at the base camps, supplies land vehicles designed for all types of terrain, special field equipment, and responds to the Arctic field worker's greatest expense and concern: safe, efficient air transport (helicopters and aircrafts) and an excellent radio communications network. Civilian research scientists and engineers of Canada, of wide ranging disciplines, played and keep playing the most significant roles in advancing polar science and crucial roles in flying the nation's flag in the High Arctic, the Canadian basin, and the Arctic Ocean since 1959.

Table 1.1 Weather stations in the Canadian Arctic region (currently operational and decommissioned).

Weather Station	Latitude °N	Longitude °W	Start	End	Territory
Alert	50.583	126.933	1913	2006	Nunavut
Eureka	79.983	85.933	1947	Cont.	Nunavut
Isachsen (closed)	78.783	103.533	1948	1978	Nunavut
Grise Fiord	74.417	82.950	1973	1977	Nunavut
Mould Bay (closed)	76.233	119.333	1948	1997	NWT
Resolute	74.717	94.970	1947	Cont.	Nunavut
Nanisivik	72.983	84.617	1976	2011	Nunavut
Pond Inlet	72.689	77.969	1975	Cont.	Nunavut
Sachs Harbor	72.000	125.267	1955	Cont.	NWT
Holman	70.733	117.783	1941	1969	NWT
Clyde	70.486	68.517	1933	Cont.	Nunavut

The weather stations in the Arctic region (most of them are in the High Arctic) are listed in Table 1.1. These stations were established by the Canadian government and some were operated jointly with the U.S. military. Only five stations are currently operational: Eureka, Resolute Bay, Pond Inlet, Sachs Harbour, and Clyde. The locations of all stations are marked in Figure 1.3. When all stations were operational, they formed a network for providing weather services that were considerably significant by Arctic standards. A detailed historical account of development and expansion of meteorological facilities in the Arctic is given by *Smith* [2009].

1.3. FASCINATING NATURE OF SEA ICE

Most people living in cold countries, where snow and ice are part of the most familiar of natural phenomena, don't think much of scientific importance of these natural materials. We never realize that the solid state of water in all of its forms is actually a unique and the most fascinating natural crystalline material. Floating sea ice, in particular, is a very complex material. It features four readily noticeable and interesting characteristics. First, it is a composite material that encompasses three phases of matter: solid, liquid, and gas, depending upon temperature. Second, it exists in nature at temperatures very close to its melting point. In fact, the ice-water interface at the bottom of floating ice covers is always at the melting point. Third, it floats simply because it has lower density than the density of its melt (i.e., the liquid from which it solidifies). And snow deposits on floating ice sheets add to the complexities of the ice regime. Fourth and certainly the most important aspects of floating ice covers (both freshwater and sea ice) is the fact that they act like blankets and protect marine life in lakes, rivers, and oceans.

The first two characteristics make floating ice highly responsive to changes in atmospheric temperature, especially when it is thin. Its physical and radiometric properties, as well as the properties of a possible overlaid snow cover, change in order to maintain a state of thermal equilibrium between the ice and the atmosphere. The third characteristic is responsible for the floatation of ice on its melt and therefore moving in response to wind and oceanic current unless it is shore fast (called "land fast") or becomes "grounded" in relatively shallow waters. Land-fast ice, however, is subjected to tidal actions producing cracks, "ice hinges," and rubbles parallel to the shorelines. Sea ice is considered to be the fastest global-scale solid material moving upon Earth's surface. Given the complex nature of sea ice composition, its thermal state, and mobility, it is important to understand the processes involved in its formation and growth, particularly the desalination and deformation processes, as well as its decay. This should help to demystify the descriptions found in literature about ice, and sea ice in particular, as apparently peculiar, bewildering, confusing, puzzling, baffling, etc.

The heterogeneous and multiphase composition of sea ice arises because the salts and gases that dissolve in seawater cannot be incorporated into the lattice (polycrystalline) structure of sea ice. This structure is made up of pure ice crystals, leaving salts to be included within the interstices of the solid ice matrix in the form of liquid brine. Gases are also included in the form of gaseous bubbles. Other impurities such as microalgae, nonorganic deposits, and trace elements may also exist. A characteristic process that follows from this multiphase composition is the brine drainage (which takes other impurities with it) into the underlying ocean water. This process takes place at a rate that depends on the ice permeability and temperature. It continues throughout the lifetime of the ice cover, causing the bulk properties of the ice to be continuously changing.

Since ice exists in nature at temperatures of only a small fraction below its melting temperature, from the geophysical and materials science point of view it is considered to be a high-temperature material. High thermal state of ice in nature and related implications are discussed in details in Chapter 4. Even a very cold Arctic air temperature, T, of 233 K (or −40 °C) for pure ice is equivalent to a normalized (called homologous) temperature, T_h ($=T/T_m$) of $0.85T_m$ where T_m is the melting point of 273 K. This is only 15% below the melting point. Floating ice sheets rarely attain such low temperatures. Snow covers act like blankets and, as will be seen in Chapter 3 (section 3.1), keep the ice rather warm even in the middle of the winter. A realistic average temperature of a snow-covered floating ice sheet, say −10 °C (263.15 K) is only about 3.7% below the melting point. Unquestionably, this has to be considered as an extremely high temperature. This important aspect of natural ice is normally overlooked by the communities of ice scientists, engineers, or environmentalists. An important implication of this feature is the rapid and drastic changes of surface and bulk properties of ice cover in response to variations in meteorological conditions (temperature and precipitation). These variations often take place during a very short time frame (a few days or even hours) especially during the early formation and onset of melt periods. This is similar, albeit occurring at a much shorter scale, to long-term changes in the landscape of Earth's solid crust (existing at homologous temperatures, less than $0.15T_m$) in response to changes in the underlying viscous mantle. In both cases changes are caused by external or internal stimuli, but the landscape of Earth's crust evolves over tens of thousands of years because it exists at homologous temperatures far below those of ice.

Ice has lower density compared to its melt (fresh or saline water). Most liquids shrink as their temperature approaches the freezing point because the molecules tend to move slower. At the freezing temperature the molecules of the crystalline solids are usually tightly packed, exhibiting higher density. Water and ice are exceptions because of the polar nature of their molecule that develops hydrogen bonding between them. When water is cooled to near its solidification temperature, hydrogen bonding causes molecules to rearrange into lattice structure with "open gaps" within the lattice. This causes a decrease of water density below its maximum density of 1000 kg/m^3 at 4 °C. The density of pure ice is 917.6 kg/m^3 at 0 °C, and it increases slightly with decreasing temperature, reaching 934.0 kg/m^3 at −180 °C. Due to this unique feature, ice floats on water. This sustains marine life under the ice cover as seawater never freezes all the way to the ocean floor. Nevertheless, the floating ice becomes mobile under wind and ocean current forces and that causes continuous opening, closing, thickening, and deforming the ice pack. It represents hazards to marine operations.

One of the fascinations of ice, and sea ice in particular, that has attracted many researchers in the field of ice physics (including the authors of this book), is the microstructural feature. This is the study of ice "anatomy"; namely, its crystallographic structure that reveals much information about its multiphase components of solid ice, liquid (brine) inclusions, and gas bubbles. There is much to learn about ice from its microstructure photographs when they are produced in the best clarity and colorful quality. The microstructure reveals information about the water quality from which ice was developed, mechanical stresses to which ice was exposed to, oceanic and atmospheric conditions that prevailed during its growth, continuation or interruption of the growth, metamorphism of the snow overlaying the ice, and so on. Without microstructural information revealed by photographs of thin section of sea ice, one could only make an educated guess about the sea ice composition and growth history. Much of the information in the ice physics part of this book is supported with macro- and micrographs of ice and forensic type of investigations. These are photographs of very thin sections (less than 1 mm), with very smooth top and bottom surfaces parallel to each other, obtained by using a method called double-microtoming technique (DMT), presented in detail in Chapter 6. When a section with thickness in the range of about 0.5–0.7 mm is viewed and photographed using polarized light, different crystals exhibit different colors. The colorful micrographs reveal the hidden beauty of nature's art. The microstructural details provide also information related to complex structure-property relationships.

Out of admiration for its beauty, the authors decided to use terms such as "ice-rich" or "ice-covered" water in this book instead of the commonly used term such as "ice-infested water." To the people interested only in the economic aspects of making shortcuts from the Atlantic Ocean to the Pacific by moving through the Northwest Passage, sea ice has been a formidable obstacle and the water was considered to be 'infested' with ice. Ice-covered areas were considered, for example, by the British explorers in the nineteenth century as something that stood in their way and complaining about the ice forming too early or not breaking up. Sea ice, however, in one form or the other plays a significant role in the daily life of communities inhabiting cold regions of the world. The presence of ice in the seas is a welcoming sight for the people of the north, the Inuit.

Inuit communities in the Canadian Arctic and Greenland (*Kalaallit Nunaat* in Inuktitut), whose life is directly affected by sea ice, have developed many stories, myths, and profound knowledge about sea ice. Actually, the people from the Baffin Island migrated and settled in Greenland and share the same cultural aspects of the northern communities in Canada. The town of Qaanaaq (77.49°N, 69.38°W) is the northernmost permanent human settlement in the world. For them, like the others of the Low Arctic of Canada, the sea ice is home. The frozen seas have always been a very welcome feature and still are, as can be seen from Figure 1.4, even though the lifestyle has been changed significantly in the last 50 years after about 1960. The government of Canada acknowledged the people of the north by issuing a series of stamps on the Arctic as shown in Figure 1.5.

Figure 1.4 Using discarded sea ice samples from N. K. Sinha (after bending strength tests), the children of Qaanaaq, Greenland (Kalaallit Nunaat), posed proudly to show off one of their creations—an Inuit sculpture or *inukshuk* (photo of N. K. Sinha, from National Research Council Canada, *Sphere*, No. 4, 1994).

Prior to the onset of outside pressures in 1960, life in the north continued both on land and the ice, depending on the season. Inuit used to have temporary villages of igloos on the ice. Children were born on the ice. People would live and travel on the ice, albeit slowly, without the fast moving snowmobiles introduced in the 1970s. Consequently, the Inuit had developed some "age-based" terminologies for sea ice. While working on a project related to the interactions of ice with the newly constructed dock at Nanisivik in Borden Peninsula of Baffin Island [*Frederking and Sinha*,1977], the author (Sinha) came in contact with several persons from the nearby Inuit community of Arctic Bay (*Ikpiaqjuk*) and Pond Inlet (see Figure 1.3). They were working at the lead-zinc mine under construction. Although the author saw only the newly formed sea ice around the dock in Strathcona Sound during his first trip to Nanisivik in November–December, 1975, when it used to be dark all the time, he learnt a few simple terms of sea ice in Inuktitut (ee-nook-tee-tut). The word *Inuktitut* means "in the manner of the Inuit." Later in the spring of 1976 he could see the fully

Figure 1.5 First-day cover issued by Canada Post in 1995 featuring a series of stamps on the Arctic and a photograph by N. K. Sinha (National Archive Canada No. C-24520) of NRC borehole indentor system on sea ice in Resolute Bay ("Canada 95," The Collection of 1995 Stamps, published by Canada Post Corporation, pp. 40–41).

developed sea ice, or *siku*, and rubbles, or *ivuniit*, near the shoreline for the first time in his life. A few examples of sea ice names in Inuktitut are given below.

Qinu The darker frazil or grease ice, seen during the very beginning of winter, is known as *qinu*. No specific distinctions are made between frazil ice and grease ice. In general, the frazil refers to the stage when thin and tiny plates or needles of ice are suspended loosely in the water. Grease ice refers to the stage when the frazil crystals have coagulated sufficiently to form a soupy or oily layer on the surface.

Sikuaq As the ice grows thick enough to support a person to walk on it, the name is changed from *qinu* to *sikuaq*. Ice thickness is around 80 mm or more. This stage and up to a thickness of 0.30 m, is termed as "young" ice according to the World Meteorological Organization (WMO).

Tuvaq Ice in the open waters forms first around the shorelines. If it continues to stay attached with the land and continues to grow, then it is called, *tuvaq*, which is equivalent to land-fast ice according to WMO terminology.

Sinaaq One has to approach the edges of land-fast or *tuvaq* ice extremely carefully. The edges may crack and, therefore, always very unsafe to walk. The children will never be allowed to play and go close to the *sinaaq*.

Siku When the sea ice cover is sufficiently thick to support a group of people, it is called *siku*. This term is equivalent to the WMO term *First Year*, which is assigned to ice with thickness greater than 0.30 m.

Ivuniit During travel on the *siku*, or even going from the shore to level ice, often ridges or rubbles of ice are encountered. They are caused by winds and/or currents or tidal waves. These obstructions are named *ivuniit*.

Qavvaq Old, salt-free white or translucent sea ice is called as *qavvaq* or equivalent to *Multi Year* ice.

When an ice cover is thin, a small variation in thickness results in large variations on the load it can carry, especially for sea ice. Moreover, relatively larger variations in ice thickness occur during the early periods of winter when the ice cover is thin. Most drowning or break-through accidents occur when the ice is thin or not sufficiently thick to carry the load placed on the ice covers. In reality the introduction of snowmobiles has significantly increased range (and speed) of activities among the young Inuits, but also, unfortunately, it increased injuries and deaths, including ice-related drownings. Consequently, efforts in rescue operations have become very challenging [*Lifesaving Society*, 1998]. Anyway, as pointed out and illustrated earlier in Figure 1.2, it is not easy to sample thin sea ice, such as *qinu* or *sikuaq*, for metallurgical investigations. But one does not require an icebreaker either to collect samples of thin sea ice. The main problem is the fact that most ice scientists living in areas remote from the Arctic and the Antarctic sea ice are neither familiar with the bearing capacities of newly grown sea ice in nature nor are they equipped (mentally and physically) to go over thin ice. Above all, unless one lives in the areas of interest and is familiar to the formation of ice, year after years it is impossible to know the characteristics of the ice cover, particularly its thickness variability a few days after the freeze-up. Floating young sea ice cover, of thickness < 0.01 m, imposes severe physical limitations on field measurements and sampling. Prior to 1977, therefore, very little effort had been made to examine young (Y) sea ice as compared to numerous field investigations carried out on mature first-year (FY) sea ice. For that reason, most of the early studies on the initial growth of saline ice came from laboratory investigations.

Realizing the practical limitations—strategically, financially, as well as for the knowledge base—the author (Nirmal Sinha) decided to work closely with the local inhabitants of the Baffin Island. The goal was to tap and learn from Inuit knowledge about sea ice. No question, the Inuit elders were the experts on sea ice. Due to their hunting experience on sea ice in certain localities for many years and their long-term observations on ice covers, they have developed profound and rich understandings of the marine environment. Toward the end of 1976, an opportunity came to him in Pond Inlet to build long-term relationships with the local residents. Although the locals were the experts of ice, they never witnessed the inner beauty of ice crystals that can be seen in thin sections of ice under polarized light. Using his home-made portable polariscope, the author introduced the Inuit elders, the young, and the children to the wonderful world of colors of ice crystals. Almost instantaneously he developed loving relationships of the local people. In fact, they were bringing their ice samples, including pieces from an iceberg grounded in Eclipse Sound. The bonds resulted in conducting the long-term studies on annual sea ice in Eclipse Sound, described in *Sinha and Nakawo* [1981], *Nakawo and Sinha* [1981, 1984], and *Sinha* [1983a]. A deep knowledge on the growth and distribution of ice thickness in areas of oceans close to human settlements helps in assessing wide ranging characteristics of ice, its bearing capacity, and other engineering properties of annual ice elsewhere, including pack ice.

Microstructurally, sea ice is significantly different from freshwater river and lakes ice. It is also different from other floating ice types of land origin such as icebergs or land-attached ice shelves and floating ice islands. The latter forms have significantly larger volume than sea ice and therefore can easily be distinguished even if they are surrounded by sea ice. Icebergs and ice islands protrude a few meters above sea level. About 90% of all icebergs encountered in Canadian waters are calved from the glaciers of western Greenland. The annually observed

Figure 1.6 A tilted iceberg in Baffin Bay, June 1984, revealing underwater bulbous bow extension of the hull, like modern ships (photo by N. K. Sinha, unpublished).

icebergs in Baffin Bay and Labrador Sea ranges between 10,000 and 40,000. The Canadian Ice Service (CIS) produces charts of icebergs in the Canadian east coast water south of 60°N. The information is mainly based on visual observations from ships and aircrafts but also occasionally obtained from satellite imagery.

Icebergs are fascinating to watch. An example of a nontabular iceberg in northern Baffin Bay is shown in Figure 1.6. It can provide interesting lessons to naval architects regarding Mother Nature's efficient design of bulbous bows. A bulbous bow, in the language of naval architects, is an extension of the hull just below the waterline. It has many subtle shape variations, but it's basically a rounded front portion that flares out slightly as it blends into the traditional displacement hull construction. These forward protrusions are about twice as long as the width of the base. They would usually not extend forward past the top of the bow. This can be seen in Figure 1.6 also by imagining the iceberg to its original position before tilting. The basic principle of bulbous bow is to create a low-pressure zone to eliminate the bow wave and reduce drag. Hulls built with bulbous bow sections are common today for seagoing ships. Under certain conditions, depending on speed and sea state, the bulbous bow is very efficient at redirecting forces of hydrodynamic resistance and drag. Ice breaking ships do have a special shape of bulbous bow that is heavily reinforced to withstand the pressures induced by sea ice.

1.4. SEA ICE IN RESEARCH AND OPERATIONAL DISCIPLINES

Sea ice is an important component in a few study and application disciplines including navigation, physics, climatology, meteorology, oceanography, marine biology, and marine and offshore structural engineering. Each community has developed its interest in certain aspect(s) of sea ice, yet all of them draw from certain fundamental physical or geophysical properties of sea ice as explained below.

1.4.1. Sea Ice in Marine Navigation

Sea ice plays vital roles for naval transportation of goods and supplies during winter months in ice-rich waters of many northern countries. Some major areas of impacts are the Gulf of Bothnia for Sweden and Finland, Gulf of Finland, Northern Sea Route, Okhotst Sea for Russia, the Baltic Sea for Eurasia, the Gulf of Sr. Lawrence and Labrador Sea for Canada, and the Bohai Bay for China. Safe marine navigation in these water bodies requires timely information on ice accompanied with meteorological data. Technical products to map ice extent, types, strength, and surface features are produced regularly by a few ice centers in northern countries to serve marine navigation. Charts are generated based on interpretation of remotely sensed images. High-resolution charts cover local areas such as fjords, straits, and marginal ice zones. Medium-resolution charts cover areas that are used in determining ship routes. Regional forecast and regional ice charts provide ice extent, concentration, motion, and perhaps thickness.

In Canada, the Canadian Coast Guard (CCG) provides recommendations on safe navigation routes as well as escort services to ships transiting Canadian ice-covered waters. The CCG receives ice information from a few sources, of which the CIS of the federal department is the prime. Icebreaker superintendents have the updated conditions of the prevailing ice conditions and the anticipated trend of ice motion. Therefore, they are well positioned to provide reasoned advice regarding the best routes for the ships to pursue. Usually, the masters of the ship report to the CCG before entering an area of ice-covered water. This ensures a continuing monitor of the ship's position by CCG operators. If icebreaker support becomes necessary, it can be provided with minimum delay.

Records of sea ice concentration and thickness obtained from remote sensing data during the past two decades have confirmed a trend of Arctic ice cap thinning and shrinking. This may lead to opening a seaway connecting the Atlantic and the Pacific Oceans for marine navigation for a longer time or eventually for the entire year. The dream of many early European explorers, which occupied their adventures for more than 400 years, may soon be realized by the merit of climate change. The economic impact of this scenario is significant. Ice-free future sea routes will reduce the number of days of marine transportation between Europe and Asia and double the vessel fuel efficiency. More importantly, commercial marine transportation will have more access to natural resources in the Arctic. Two routes are expected

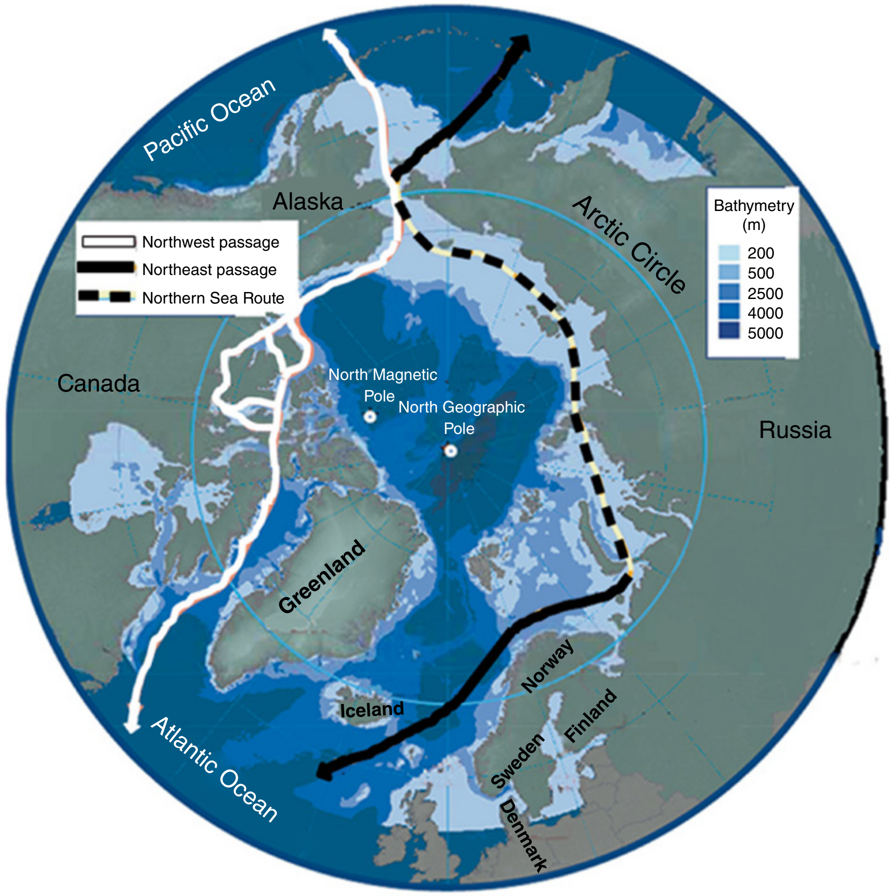

Figure 1.7 Expected ice-free Northeast and Northwest Passages that connect the Atlantic and Pacific Oceans. The outer boundary of the map is 50° latitude. The location of the north magnetic pole is as of 2008 (the average location moves in loops at around 40 km/year).

to open and link the Atlantic and Pacific Oceans: The Northwest Passage (NWP) and Northeast Passage (NEP). The latter encompasses the Russian Northern Sea Route (NSR). Figure 1.7 displays the two routes overlaid on a bathymetry map of the Arctic Ocean and surrounding seas. The NWP encompasses a few routes that pass through the Canadian Arctic Archipelago region. The Archipelago has always been a major impediment to fully opening the NWP. The NEP, on the other hand, is a set of marine routes that run along the coast of Siberia, making use of a few straits through the islands of the Russian Arctic. Sea ice will continue to be monitored in the future to spot any long-lasting opening of the routes.

1.4.2. Sea Ice in Physics

While large single crystals of pure ice can be found in low-temperate glaciers or can be prepared in cold laboratories, the usual icy objects found in nature or fabricated for human consumption consist of aggregate of many crystals. Geophysicists and materials scientists are interested in ice mainly because it is transparent or translucent and commonly occurs as a polycrystalline solid, similar to crystalline minerals that exist in nature, common metals, and alloys and ceramics. There is a vast literature on the physics and chemistry of water and ice. Probably the most comprehensive book on the early and fundamental studies on this topic was compiled by the physicist, *Dorsey* [1968] originally published in 1940 as part of the Monograph Series of the American Chemical Society. Since then, a number of books have been written on the physics and mechanics of ice. A comprehensive review of crystalline structure of lake and river ice is given in *Pounder* [1965], *Hobbs* [1974], and *Michel* [1978]. The information base for sea ice is very broad and rather scattered. It can be found, for example, in *Pounder* [1965], *Weeks and Ackley* [1982], *Weeks* [2010], and *Petrich and Eicken* [2009]. Journals devoted to ice include *Journal of Glaciology* published by the International Glaciological Society (IGS), *Cold Regions Science and Technology* (CRST), and to some extent by the *Journal of Geophysical Research* (JGR). Physics of ice in general and details on sea ice, especially of areas surrounding the continent of Antarctica and Okhotsk Sea, can be found in a number of articles in Japanese published by the Institute for Low Temperature Science, University of Hokkaido, Sapporo, Japan, since 1950s in the *Low Temperature Science Series*. It should be mentioned here that the first man-made snowflakes were also produced at this institute in Sapporo. On a regular basis, the International Symposium on Okhotsk Sea & Sea Ice (ISOSSI) is held in Mombetsu, Hokkaido, Japan. The Okhotsk Sea & Cold Ocean Research Association linked with the Sea Ice Research Laboratory, Hokkaido University, publishes the proceedings volumes of ISOSSI (articles in both English and Japanese).

The inner structures of any material that cannot be recognized by human eyes are known as microstructure. Microstructure-property (thermal, optical, mechanical, electrical, etc.) relationships of materials are very complex and depend on their texture and structure. The use of knowledge about ice microstructure for understanding and demystifying the behavior of several alloys at high temperatures was highlighted and pioneered in an early study by *Tabata and Ono* [1962]. No doubt, the extremely high thermal states, in case of sea ice, add even more complexities in the microstructure-property relationships. But then, gas turbine engine materials have also became highly complex over the years [*Sims et al.*, 1987].

As pointed out earlier, sea ice floating on its own melt exists at temperatures of about 5%, or less, than its melting point. This is significantly more than the maximum temperature allowed for the operation of man-made nickel-base directionally solidified (DS) columnar-grained (CG) superalloys, used in blades of modern gas turbine engines (jet or power-generating engines) and end casings of rocket engines. Nickel-based DS, CG superalloys were introduced in jet engines during the late 1960s [*Duhl*, 1987, in Chapter 7 of *Sims et. al.*, 1987]. Physicists, ceramicists, and high-temperature metallurgists, are therefore, interested in applying lessons learnt from studying structure-property relationship of ice directly to complex high-temperature, titanium- and nickel-base superalloys [*Sinha*, 2009a]. They can perform experiments under varying conditions of thermal and mechanical loading, in conjunction with microstructural analysis at the experimental (high) temperature, and model thermomechanical behavior of gas turbine engine materials, such as stress relaxation processes and strain rate sensitivity of yield strengths [*Sinha and Sinha*, 2011]. Moreover, the microstructure of sea ice exhibits impurity entrapment and grain- and subgrain-scale substructure very similar to high-temperature metallic alloys and ceramics. These features are key factors in determining many properties of sea ice, including thermal conductivity, microwave emissivity, dielectric constant, and mechanical strength.

Since ice has an accessible melting temperature, it could provide a model of how other materials will behave at very high temperatures near their melting points. Physicists have applied theories and experimental techniques that describe the behavior of snow and ice around their melting temperatures to high-temperature materials such as ceramics and advanced alloys used inside jet engines or gas turbine engines (examples include titanium-based and nickel-based superalloys). This has been substantiated by the fact that sea ice is a prime example of a natural material that exhibits impurity entrapment and grain-scale substructure very similar to binary alloys.

1.4.3. Sea Ice in Climatology

Sea ice plays a key role within the climate system and has long been thought to be a primary indicator of global warmimg. Climatologists are interested in sea ice because of a number of reasons, such as (1) its thermal and optical properties are important input to climate models, (2) its extent in the polar regions is a strong indicator of climate change, and (3) its strong influence in the high as well as midlatitude large-scale circulations of the atmosphere and ocean.

Thermal properties of sea ice determine the thermodynamic interaction between the cold atmosphere and the warm underlying ocean. This is demonstrated in two processes: the heat exchange between the two media and the heat of fusion released or acquired when ice freezes or melts, respectively. The thermal conductivity of sea ice is relatively low (2.25 W/m · K) compared to thermally conductive metals (205 W/m · K of aluminum or 401 W/m · K of copper). When covered with snow, the effective conductivity becomes even lower (thermal conductivity of snow is between 0.1 and 0.25 W/m · K, depending on its density and wetness). Therefore, the presence of the ice cover on the ocean surface reduces the ocean-atmosphere heat exchange even in the presence of thin ice [*Maykut*, 1978, 1982]. This reduces the moisture transfer from the ocean to the atmosphere, the momentum transfer from the atmosphere to the ocean, and the exchange of chemicals constituents between the two media.

The heat of fusion of water (defined as the heat required for transforming one gram of water at freezing temperature to ice) is one of the highest of all substances. When sea water freezes, it releases approximately 80 cal/g of heat to the atmosphere. This is the energy needed to establish the hydrogen bonding between the ice molecules. It is equivalent to energy required to raise the temperature of one gram of water from 0 to 80 °C. When ice melts it absorbs the same energy to produce water at the same temperature. This represents a huge amount of heat exchange between the ocean and the atmosphere that affects weather systems and eventually regional climate.

It should be taken into consideration in regional and global climate modeling.

The motion of the ice sheet may cause its divergence and therefore the appearance of openings within the pack ice in the form of cracks, leads or, polynya (for definition see Section 2.6.2). The winter flux of heat from these openings to the atmosphere can be two orders of magnitude larger than the heat flux through an adjacent thick ice cover [*Maykut*, 1982]. However, this sudden increase may not last for a long time as water in these openings usually freezes quickly. *Lüpkes et al.* [2008] found that even a few percent of open water or refrozen thin ice in the Arctic may increase the heat flux between the ocean and the atmosphere drastically and increase air temperatures within them by several degrees. In general, leads and polynya are considered to be important climatic features, and their distribution, along with the distribution of thin, ice is particularly important to the regional and perhaps global climate scenarios.

The primary optical property of sea ice that affects weather and climate systems is albedo (the ratio of reflected to incident optical radiation). Albedo from snow-covered ice (between 0.4 and 0.8) is significantly higher that albedo from seawater (≈0.06). This causes more sunlight to be reflected in the presence of ice and therefore the ocean will not be as warm as it would be in the absence of ice. That is how the ice-covered areas in the polar regions are climatologically cold. A possible decrease in the area of polar sea ice, caused by global warming, will cause more sunlight to be absorbed. This will trigger a positive feedback loop that amplifies the absorption and ice melting. In the past 30 years, air temperature in the Arctic has increased at twice the rate of the average increase around the globe. In relative terms, more increase in winter temperature has been observed compared to increase in summer temperature. Climatologists are therefore interested in information about the ice extent in the polar regions and the frequency and distribution of melt ponds at the surfaces.

Using daily ice concentration maps retrieved from satellite passive microwave a few operational centers track the ice extent in the Arctic region. Updated information is available, for example, from the Arctic Sea-Ice Monitor System maintained by the Japan Aerospace Exploration Agency (JAXA) in the IARC-JAXA information system (IJIS) located at the International Arctic Research Center (IARC) in Fairbanks, Alaska. The minimum ice extent usually occurs in mid-September. The lowest ice extent so far occurred on 5 September, 2012 (falling below 4 million km^2) while the second and third lowest occurred in 2007 and 2011, respectively. The ice extent in the Arctic has recently been reduced by 40%–50% of its average value in the 1980s. This has also been accompanied by reduction in ice thickness. A significant drop of about 0.6 m in multiyear (MY) ice thickness (and consequently overall ice thickness) is observed since 2005. The decrease in thickness has occurred at a rate of about 0.17 m per year [*Kwok and Sulsky*, 2010]. It should be mentioned that the Antarctic ice extent has increased slightly in the past three decades.

Early global climate models predicted that the polar regions would undergo the largest greenhouse warming [*Kattenberg et al.*, 1996]. However, current climate models underestimate the rate of ice retreat and thinning in the Arctic. The ice extent and thickness estimated from satellite remote sensing observations is four times faster than model calculations [*Stroeve et al.*, 2007]. Modelers indicate that this is mainly due to a lack of accurate representation of ice conditions in the models. For example, *Rampal et al.* [2011] argue that the underestimation is caused by the poor representation of the sea ice drift out of the Arctic Basin. In general, climate models still show deficits in reproducing the underlying processes for observed Arctic sea ice decline [Stroeve et al., 2007]. More utilization of ice information in climate models and further understanding of sea ice processes in relation to the climate system should ultimately improve regional and global climate modeling.

1.4.4. Sea Ice in Meteorology

In order to restore a temperature balance between the cold polar atmosphere and the ocean, the latter always tries to transport heat toward the poles. This process triggers what is known as atmospheric circulation, which is responsible for generating most weather systems over north-latitude areas. This includes the recurring low-pressure systems associated with storms and precipitation. Obviously, the presence or absence of sea ice in the polar regions affects these regional weather systems, directly or indirectly. The effect is forecasted using weather models. The models integrate a collection of physical processes into a consistent framework that can be used to forecast weather and project future climate conditions.

The traditional representation of sea ice information in the model is achieved by feeding ice information directly to the model at certain time steps. This approach was used in the early 1970s to initialize weather forecast models [*Stensrud*, 2007]. At a basic level, the required information at each grid cell should include ice concentration, but additional data such as ice thickness, surface roughness, and snow cover are also recommended. Other ice data that can be parameterized such as albedo, surface emissivity, and surface fluxes can be used as well. However, it should be noted that parameterization of surface fluxes is a challenging task because of the heterogeneity and the rapidly changing ice surface. Recent weather models incorporate a sea ice module (or a coupled ice-ocean model) to represent fundamental dynamic and thermodynamic processes.

It should be noted, however, that efforts to incorporate sea ice information in weather models are limited so far to identification of ice regardless of its type. However, thickness information (a proxy indicator of ice type) is also desirable even at coarse resolution.

One of the challenges in representing sea ice information in weather models is the spatial and temporal variation of the sea ice cover in terms of its physical parameters (e.g., temperature and emissivity) and surface geometry (leveled or deformed ice). While the surface temperature of the Arctic ice is fairly homogeneous in winter (can be as low as −40 °C), the open-water areas in leads and polynyas or open-water areas beyond the ice edge have a surface temperature at approximately −1.8 °C (the freezing point of seawater). This sharp contrast makes ice-covered oceanic regions among the largest spatially changing surface temperature on Earth. Accurate characterization of the spatial distribution of ice surface temperature is crucial to improving weather forecast, particularly during winter, in areas where leads become closely distributed within the ice cover. The same applies in summer when melt ponds with its substantially less albedo than surrounding ice also become closely distributed.

The importance of sea ice in meteorology and weather forecast can also be appreciated using the following similarity. Global ocean-atmosphere phenomena affect regions very far away from their origin. El Niño, for example, starts in the middle of the Pacific Ocean, yet it affects temperature and precipitation in North America in winter. Arctic sea ice, in turn, must have an effect on regions far from the Arctic. This is yet to be understood using regional weather models that focus on the Arctic, then upscale the results using a global model. A project has been undertaken in Environment Canada to develop a weather forecast system called Polar-GEM to improve the representation of sea ice at northern latitudes by coupling a detailed dynamical/thermodynamic sea ice model (GEM is Global Environmental Multiscale; the operational weather forecast model used by Environment Canada's Canadian Meteorological Centre). Particular emphasis is put on improving representation of surface processes such as using detailed dynamic-thermodynamic models coupled with ocean currents over the Arctic basin. Better representation of snow processes and air-sea interactions are also implemented using detailed snow models over sea ice.

A recent project has been undertaken by Environment Canada to develop a regional "sea ice analysis" system. It is primarily designed to satisfy the requirements for planning of marine transportation and other marine operations in ice-rich waters around North America. Additionally, it will satisfy the needs of regional numerical weather prediction models and regional sea ice model initialization. The term "sea ice analysis" means a forecast system using satellite- or ground-based observations incorporated in a data assimilation approach. The analysis is produced using the commonly known three-dimensional variational (3D-Var) data assimilation approach. In its first version, the system produces only ice concentration at approximately 5 km resolution using a 6 h persistence forecast from the previous analysis as the background state. The assimilated observations encompass measurements from a few satellite sensors. In addition, ice concentrations estimates from operational ice analysis of synthetic aperture radar (SAR) at the Canadian Ice Service as well as estimates from passive-microwave sensors are also assimilated. The system is described in *Buehner et al.* [2013]. Sea ice information required for weather forecast can be tracked at an appropriate scale using remote sensing observations, a sea ice model, or "sea ice analysis." Remote sensing is an important tool, though improvement of accuracy and spatial and temporal resolutions are still needed. Timely information is required, particularly in highly dynamic regimes such as ice edge where ice can drift up to 40 km in a day.

Growing scientific evidence has pointed to the influence of the current decline in Arctic sea ice (as explained later in this chapter) on atmospheric phenomena within and beyond the Arctic. *Budikova* [2009] found that changes in Arctic sea ice interact with processes related to atmospheric wind and temperature fields as well as to thermodynamic and radiative processes connected with water vapor, clouds, and aerosol feedbacks. *Cassano et al.* [2013] used climate model simulation to explore the effect of the increasing open water in the Arctic on Earth's atmosphere. They found that the increase in open water during the fall of 2007 (this season featured the third lowest Arctic sea ice extent on record) led to an increase in atmospheric temperature. The higher temperatures in the Arctic thus caused a decrease in the pole-to-equator temperature gradient, which in turn created a weaker jet stream and less storminess in the midlatitudes. To conclude this section, it is worth repeating that sea ice conditions, while of interest in their own right, are important information for improving weather forecast models, both regional and global.

It is worth noting that the term "meteorology" encompasses sea ice according to some definitions. For example, the CIS, a division of the Meteorological Service of Canada (MSC), is the leading authority for information about ice in Canada's navigable waters. Weather and ice services operate under the same federal Department of Environment (DOE) in Canada.

1.4.5. Sea Ice in Oceanography

During the winter months of the Northern Hemisphere, when the oceans in the Arctic region freeze, sea ice in the Antarctic region melt. A reversal in this process occurs

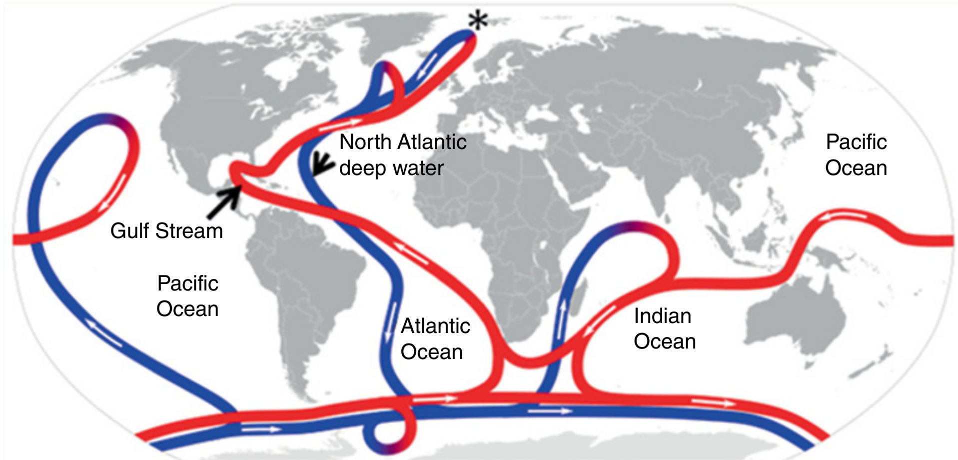

Figure 1.8 Major routes of the thermohaline circulation that circulates cold water (blue) and warm water (red) between the deep and surface ocean worldwide. The asterisk marks the location of Odden ice where warm surface water prevents the expansion of the ice cover. (For color detail, please see color plate section).

during the summer months. Freezing and thawing of the two primary cryosphere work like a seesaw of oceanic current—like a narrow board pivoted in the middle (the equatorial region in this case) so that as one goes up, the other goes down. Freezing and thawing of annual sea ice in the primary cryospheres affect the movement of ocean water in two ways: by rejection of salt to the underlying water during sea ice formation and growth and by melting large volumes of ice that has drifted away from the polar areas during the summer. The first process initiates what is known as thermohaline oceanic circulation while the second disturbs it. Thermohaline circulation (also known as the global ocean conveyor belt) is a large-scale pattern of seawater motion around the world driven by vertical density gradient. The adjective thermohaline is composed of two syllables: *thermo*, which refers to temperature, and *haline,* which refers to salt content. Changes in water density are associated with changes in its temperature but are caused mainly by the processes of salt rejection during the initial formation and growth of ice covers and throughout their lifetime (see Section 2.3.3 for details).

Figure 1.8 shows the major routes of the thermohaline circulation worldwide. Since the saltier water under the ice is heavier than the deep water, a density gradient is developed that causes water to sink at the location of ice formation and growth. This triggers the circulation in the oceans around world. The cold and dense polar water starts to move along the ocean bottom toward the equator while warm water from middepth to the surface level travels from the equator back toward the poles. Much of the world oceans' deep and bottom water is believed to be formed in polar latitudes as a result of ice formation and growth. Obviously, major changes in the amount of newly formed sea ice can disrupt normal ocean circulation [*Maykut*, 1978; *Carmack*, 1986]. Figure 1.8 also shows the location of formation of what is known as Odden sea ice in the Greenland Sea. Odden is a Norwegian word for headland. This is the crucial location for the initiation of the thermohaline circulation. The circulation takes the form of a large tongue that extends over 250,000 km^2 [*Comiso et al.*, 2001], and it can rapidly expand or shrink over a period of a few days. In some winters it persists for months, while failing to form in others.

As shown in Figure 1.8, the circulation is closed in the Greenland Sea as the warm water moving from the south near the ocean surface level reaches that high-latitude area and starts to cool. The cooler water becomes dense and therefore sinks. The circulation continues afterward. The cycle may be slowed down if significantly less salty water (or freshwater) is provided. In this case, the sinking of seawater will be activated only by the colder and denser water at the surface. The cycle may be interrupted if much less sea ice is formed or the atmospheric temperature is maintained at the warm side so that warmer water coming from the south will not be cooled enough to make it denser. This anomaly is called the "great salinity anomaly." This phenomenon was discovered during the late 1960s and early 1970s in the Nordic Seas (this includes the Greenland, the Norwegian and the Iceland Seas). It was later observed in the 1980s and 1990s. A notable study that addresses this phenomenon is presented in *Häkkinen* [1999]. The question that would be raised at this point is: What would give rise to the major impulse of freshwater that causes this anomaly? As mentioned before, excessive rain, snow, or river runoffs are possible answers. Yet, excessive melting of glaciers and icebergs is a much more serious cause. The disappearance of sea ice may contribute to this anomaly as well.

Normally, under ocean currents and wind, Arctic sea ice moves southward from the Arctic Basin to the Greenland Sea through Fram Strait (located between Greenland in the west and Svalbard in the east). As it continues its journey south and when the melt season approaches, the sea ice starts to melt. Since sea ice is much less salty than the surrounding seawater, the melt decreases the salinity of the seawater near the surface, albeit at another location from its initial formation. Excessive melting of sea ice (or icebergs) may lead to a great salinity anomaly. Therefore, any major change in Arctic winds that move a considerably larger volume of sea ice through the Fram Strait into the North Atlantic or any significant increase in the temperature of the Nordic Seas are possible scenarios that may trigger this anomaly.

1.4.6. Sea Ice in Marine Biology

Sea ice provides habitat, shelter, breeding, feeding, nursery, and hunting grounds for a variety of species, ranging from microorganisms (alga and bacteria) to birds and marine mammals (polar bear, walrus, and a few species of whales and ice-associated seals). Sea ice also supports the habitats of many different fish as it maintains the seawater at a warm enough temperature for their survival. A retreat of sea ice will decrease the available platforms that birds and mammals use to rest on and from which to hunt. There have been concerns that this may lead to significant loss in the population of those species. A dedicated chapter on sea ice as a critical habitat for polar marine mammals and birds in presented in the book by *Thomas and Dieckmann* [2009]. The survival of the animal populations is related to the presence of areas of ice-free waters within sea ice cover. These areas provide migration routes and sources of abundant underice food reserve for animals. A notable example of such areas is polynyas. These are areas within an extensive ice cover that contain open water and thin ice even in the middle of winter when atmospheric temperatures are very low.

Animals that depend on sea ice for their survival prefer different habitats within the ice cover. For example, bearded seals and walruses prefer areas of thin or broken ice cover in relatively shallow water because their main food source is benthic invertebrates. Walruses and belugas live along leads within pack ice. Ringed seals prefer fast ice because of its stability for the successful rearing of their pups and the sufficient snow cover for construction of birth lairs. They usually congregate along the edges of the ice and they use holes in the ice to breathe (Figure 1.9). Polar bears also use this ice type since their main food is ringed and bearded seals. Traditionally, Inuit hunters also use the breathing holes for locating and getting their kills for food as well as for the furs. Harp seals live and give birth to their pups on heavy ice floes within ice-edge zones (Figure 1.10). In winter, Arctic foxes venture onto the sea ice to scavenge for remains of seals killed by polar bears.

As for the Antarctic, Weddell seals spend most of their time beneath fast ice while coming onto the sea ice surface only to rest and have their pups. Leopard and Crabeater seals are species of deep pack ice. However, since Leopard seals are predators and cannot move easily on sea ice, the Crabeater seals often rest on rougher ice where they are safe from predation. Emperor Penguins, found only in the Antarctic, prefer the fast-ice zones. In general, marine biologists confirmed that animals prefer ridged ice (compared to level ice) as an attractive habitat. From this discussion it is obvious that the ice information needed to identify the suitable habitat for each species involves some key surface features, not physical properties or crystalline structure of the bulk ice. Remote sensing observations that can be used to identify such features (e.g., fast ice, ridges, leads, ice edge, etc.) are presented in Chapter 9.

At the bottom of the marine food chain many microorganisms find their habitats within the volume of the sea ice. A notable review on this subject is presented in *Horner* [1985]. When seawater freezes, the salts are entrapped within and between the ice crystals in the form of brine, filling a network of "pockets" and "channels" (section 2.3.1). Brine volume fraction can represent up to 30% of the ice volume (section 3.4). The diameter of the brine pockets is usually less than 1 mm (Section 4.5.2). Brine provides the necessary nutrients to host extensive algal, phytoplankton (microalgae), and bacterial communities. Ice algae contribute considerably to the total

Figure 1.9 Seal breathing hole in first-year fast ice in Lancaster Sound, Canadian Arctic (photo by M. Shokr, unpublished).

(a) (b)

Figure 1.10 Calving season for seals on sea ice in the Labrador Sea: (a) mothers and (b) a puppy (photos of N. K. Sinha, March 1989, unpublished).

primary production in the Arctic (25%) and in the Antarctic (20%) [*Legendre et al.*, 1992]. About 200 species are found in Arctic sea ice, the most common among them is diatoms. These microorganisms sustain the secondary production and that, in turn, supports life of all marine animals including fish, birds, seals, bears, and many others, all the way to the killer whales.

In addition to nutrients, the survival of these organisms hinges upon the availability of light that penetrates the ice sheet to reach the depth where sufficient nutrients are available for organisms to grow. This is known as photosynthetically active radiation (PAR). While sea ice absorbs most of the incident light and the snow cover reflects most of it, the PAR at 1 m depth of an ice sheet represents only 1%–5% of the visible spectrum that penetrates the sheet. Incorporation of microorganisms from the water column into the sea ice may occur during the ice formation. However, high concentrations of microalgae have been observed during the springtime within the interstices of the lower margin of sea ice floes. In early spring the Sun starts to shine after the end of long polar nights, and the PAR reaching the bottom of the ice sheet (can be 2 m thick) would be sufficient for the alga growth. Microalgae tend to concentrate in the bottom ice layers because it is more favorable microhabitat than the surface layers. This is due to less stressful temperature and higher saline environment. Nevertheless, microalgae are also often found in a thin layer of seawater immediately under the water-ice interface. For at least 1–3 months, ice algal blooms enhance and extend biological production in polar waters. Depending largely on climatic and environmental variability, biomass accumulation of sea ice algal populations eventually depends upon the duration of their growth season. More information on biomass accumulation at the bottom of the ice is presented in section 4.5.4.

1.4.7. Sea Ice and Offshore Structures

For offshore structures—both floating and fixed, including near shore structures such as docks and ports, and vessels in ice-rich waters—the influence of sea ice is probably the most significant factors to be considered by the designers and structural engineers. Structural frameworks and their mass must be able to withstand the local and overall forces exerted by moving ice. The structures must be protected from encroachments of ice using ice information managements (with the help of remote sensing). The movements of the ice could be continuous or intermittent. The ice could be first year, second year, or multiyear, flat floes, or rafted and ridged, and may consist of bergybits or icebergs. Development of new technologies for exploration for undiscovered oil and gas in the Arctic, construction of environmentally friendly production drilling platforms, transportation of gas and oil from such remote parts of the globe, and avoidance of oil spills and cleanup in case of accidents are some of the huge challenges to the energy industry with high-technology demands on structures, vessels, and pipelines.

During ice-structure interactions, there is only one choice. Ice must fail, not the structure. Figure 1.11 depicts a notch in a second-year (SY) sea ice floe, in northern Baffin Bay, made during a dedicated ramming event by the ice breaking, bulk carrier, *MV Arctic* in June 1984. During this planned ice breaking expedition, old ice floes were instrumented and ice characteristics were recorded before conducting the tests. The CCG icebreaker, *Louis*

Figure 1.11 Notch in second-year sea ice floe made by ice breaking bulk carrier, *MV Arctic*, with CCGS icebreaker, *Louis S. St-Laurent* in the background, Baffin Bay North, June 1984 (photo by N. K. Sinha, unpublished).

S. St-Laurent, also participated in this expedition. Soon after these ice breaking trials, *Louis S. St-Laurent* joined USCGC *Polar Sea* to become the first North American surface vessels to reach the North Pole on 22 August 1994. *Louis S. St-Laurent* was launched in June 1966 and joined the *SS Manhattan* expedition in the Northwest Passage during the 2 weeks, 8–22 September 1969, together with two other icebreakers USCGC *Northwind* and *Staten Island*. *Louis S. St-Laurent* is classed as a heavy Arctic icebreaker and is the largest icebreaker and flagship of the CCG.

There are a number of excellent books on ice-structure interactions and ice engineering, e.g., *Cammaert and Muggeridge* [1988], *Sanderson* [1988], and *Jones et al.* [1991]. Up to date information in the areas of ice-structure interactions and related field can be found in the journal *Cold Regions Science and Technology* (CRST). Moreover, the conferences of the Offshore Mechanics and Arctic Engineering (OMAE) of the American Society of Mechanical Engineering (ASME) and the Port and Oceans under Arctic Conditions (POAC) provide current status of developments in ice engineering. These conferences also provide platforms for disseminating research results, new developments, and novel concepts in the multidisciplinary field of polar science and technology.

1.4.8. Sea Ice for Search and Rescue and Transportation

During the long winters in many cold regions of Earth, there are more possibilities for finding relatively flat lake or sea ice surfaces than land-based areas for aircraft operations with little or no preparations. For search and rescue (SAR) operations and temporary winter air transportation, frozen lakes and fjords provide ideal locations for landings and takeoff. Usually first-year sea ice sheets in fjords are relatively free from ridges and rubble fields. This is particularly the situations where there are mountains or hilly areas on both sides and thereby protected from severe wind effects. For the Arctic and the Antarctic, sea ice in refrozen leads within the pack ice also provides ready-made sites for emergency operations for SAR, especially with short takeoff and landing (STOL) aircraft like Canada's famous *Twin Otters*. Once landed using the light STOL aircraft, measurements on the ice conditions can be performed and evaluated for operations of larger airplanes [*Sinha et al.*, 1996]. Figure 1.12 illustrates a set of plots for weight versus ice thickness, as functions of flexural strength of ice, recommended for operations of aircraft.

The bearing capacity of an ice sheet can be affected more by ice quality than by ice thickness. Safe estimates of strength values can be made by experienced ice specialists through observations of the type, quality, temperature, and uniformity of the ice, which may be supplemented by field measurements of ice bending strength described in details in Appendix A of *Sinha et al.* [1996]. These estimates can then provide the basis for decisions concerning use of the airstrip for unlimited movements, or allowing loads in excess of the maximum recommended for limited use. It is emphasized that an engineering analysis, including a detailed survey and investigations of the ice cover, should be made by a qualified ice specialists to approve a runway for an unlimited number of landings.

For practice, simulated emergency landing by an SAR aircraft of the Canadian Armed Forces were made in many areas of the High Arctic. An example of landing made on a refrozen lead in first-year sea ice is shown in Figure 1.13.

Based on the charts in Figure 1.12, history was made by a Canadian airline, First Air, by landing a fully loaded (64,640 kg) Boeing B727 jet aircraft on first-year sea ice in Frederick Hyde Fjord in northern Greenland at about 83°11′ N, 29°50′ W [*Pole*, 1995]. The landing location was 29°50′ W, but at that high latitude, the longitudes cover significantly less distances. The longitude of the 30 km long fjord was between 28°W and 32°W. Also, it was the most northerly point on sea ice cover that a B727 jet aircraft had ever landed and operated on a commercial basis. In all, 16 landings were made in 7 days from 19 May to 25 May, 1994 [*Sinha* ,1995].

The ice strip was 100 m wide and 2.5 km long. The ice thickness varied from 2.24 m at one end to 2.41 m at the other end of the runway. The ice was thus extremely uniform, and about 50% of the surface was absolutely snow free. Actually, even the snow-covered areas had very little snow. Consequently, from the air the entire fjord was

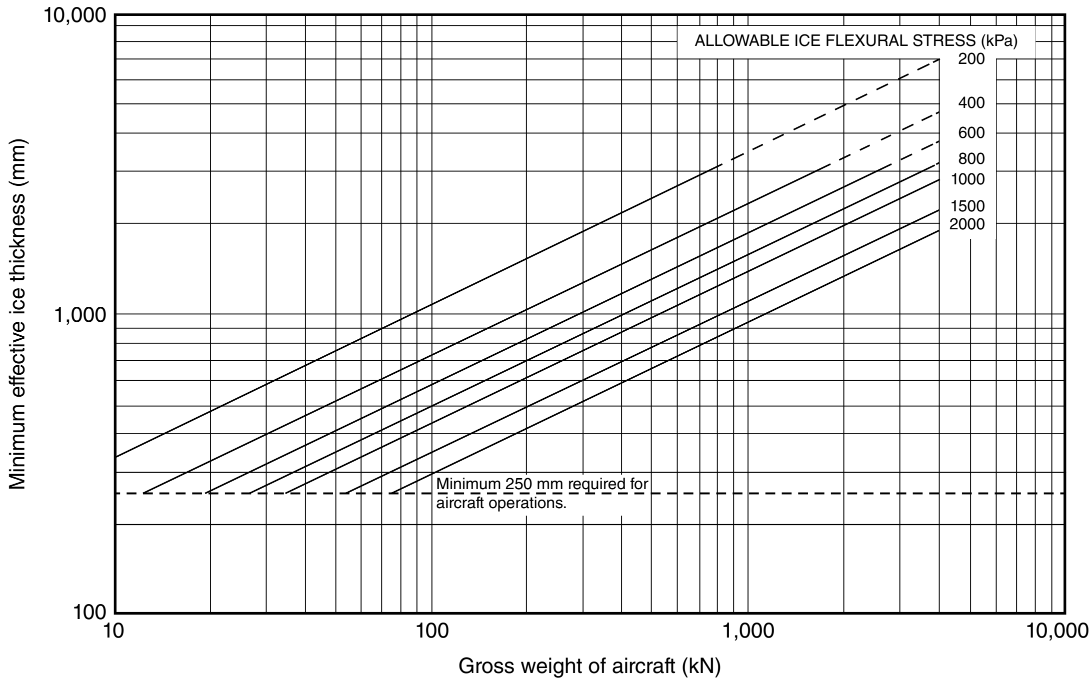

Figure 1.12 Minimum thickness recommended for landing aircraft on sea ice as functions of flexural strength (from Fig. 1 in *Sinha et al.*, 1996].

Figure 1.13 Search and rescue (SAR) Twin Otter of the Canadian Armed Forces on a snow covered refrozen lead in FY sea ice near Beechey Island, Nunavut, Canada (photo by N. K. Sinha, unpublished).

Figure 1.14 Boeing 727 on first-year sea ice in Frederick Hyde Fjord, Greenland, in May 1994; the foreground shows snow-free surface (photo by N. K. Sinha, unpublished).

turquois in color. An example of unloading the aircraft is shown in Figure 1.14, we which also shows clearly the snow-free surface of the ice.

Frozen seas also provide vital links for surface transportation between the communities and hunting opportunities in the Arctic and subarctic areas of many northern countries, especially Canada. Use of ice covers for transportation, however, is not limited only to the waters covered with sea ice. Ice surfaces of frozen lakes and rivers are routinely used for making temporary runways, winter roads for transportation of goods, and recreational trails and winter sports throughout Canada. Increase in tourism in the Arctic is also opening new opportunities for use of sea ice covers and the land. Snowmobiles replaced the dog teams almost 40 years ago, but nothing could replace the supreme yet elegant design of the traditional Inuit sledges. Only difference is that the sledges are pulled by the snowmobiles. Dog teams and rides on sea ice have become the sources of tourist attractions (Figure 1.15).

Knowledge of sea ice and its conditions year after year in their areas of activities are critical to the survival of Inuit culture and language. With the advent in microwave

(a)

(b)

Figure 1.15 (a) Dog team at Pond Inlet with Bylot Island in the background in 1978 by N. K. Sinha and (b) dogs simply ignoring author's commands at Qaanaaq in 1994 (photo by the master of the dog team).

Figure 1.16 Qaanaaq child proudly proving practical use of a core hole and a sense of sharing as the author was making notes on the ice core for structural details (photographed by N. K. Sinha, March 1994).

communications systems, television signals reached the people of the north during the middle of the 1970s. Saturdays became the days of the "Hockey night of Canada" popularized by the Canadian Broadcasting Corporation (CBC). In the early days of his research, based at Pond Inlet, the author (Sinha) found it impossible to get anyone to assist him on the ice during the days of hockey nights. In fact, they would invite him to spend time with them, share their food, hear their stories, and learn the intricacies of handling pucks on the ice. Frankly, until then the author did not pay much attention to chasing the pucks on ice with sticks; he was a soccer player. Little did he know in those days that he would end up spending lots of time on ice arenas for optimization of structure, texture, and temperatures suitable for speed skating versus figure skating or hockey. Although lifestyle has changed significantly in the last 50 years, day-to-day life among the Inuit communities are still intertwined with sea ice. For most of the year Inuit life is still tied closely to ice on the ocean. It is extremely important, therefore, for sea ice scientists and engineers, to work together with the local people of the north, and share information and experience. A child in Qaanaaq showed the author a symbolic gesture of working together in Figure 1.16, and the children of that place collectively provided another example of sharing in Figure 1.4.

1.5. SEA ICE AND REMOTE SENSING

Due to the remote locations and the extreme climate conditions where sea ice exits, the only tool that can be used to study this ice is by space-borne remote sensing. This potential was realized more than 40 years ago. In the

Preface of the notable book on Antarctic sea ice [*Jeffries*, 1998], the editor refers to a few statements made by Lyn Lewis and Willy Weeks in a report published in 1971 by the Scientific Committee on Antarctic Research. One statement underlines the thought that prevailed up to the 1950s that Antarctic sea ice is probably not any different than Arctic ice, so why bother studying it? But realizing that this may not be true to some extent, the authors of the report added: "It is clear that future work will depend critically on the logistics available to allow surface observations beyond the fast ice edge at all seasons of the year. Of almost equal importance will be the development of instruments and recording equipment suited for use in the polar environment" (*Lewis and Weeks*, 1971). The identification of that need was made one year before the launch of the first U.S. space-borne microwave instrument for Earth observations; the Electrically Scanning Microwave Radiometer (ESMR). The foresight of the authors was true. ESMR was a milestone instrument for sea ice observations in the polar regions, following the success of previous satellites that used optical sensors. A brief synopsis of satellite remote sensing for sea ice applications is presented in section 7.2.

Satellite data of sea ice represent one of the longest Earth observation records from space. Since the early 1970s many different satellite Earth observation sensors were developed and used effectively in routine surveillance of sea ice. Sensors operate in different bands of the electromagnetic spectrum: optical, infrared, and microwave (passive and active) (see section 7. 1). Information from these bands complement each other. Different sensors provide information at different scales. The basic premise of sea ice parameter retrieval from remote sensing data is the significant difference between physical and radiometric properties of sea ice and open water; namely physical temperature, reflectivity, and microwave emissivity. Optical sensors discriminate between sea ice and open-water based on their contrast in albedo. Thermal infrared sensors use the difference in the physical temperatures (ice surface is usually colder than water surface). Passive microwave sensors use the difference between ice and water microwave emissivity. Radar sensors use the difference between backscatter from the ice and the water surface, although they both occupy a wide range of values and they overlap.

Many sea ice parameter retrieval algorithms have been developed using observations from different sensors. Examples include surface temperature, retrieved from thermal infrared sensors; ice concentration and extent from passive microwave; ice types from passive and active microwave; and surface features, ice drift, and deformation from radar imagery data. The two most commonly used categories of sensors are active microwave (radar), which produces imagery data at fine resolutions of a few tens of meters, and passive microwave, which produces images at a coarse resolution of a few kilometer or tens of kilometer. Radar imagery generates information at tactical scale (a few tens or hundreds of kilometers of an imaged scene) primarily to support ship navigation though it has been used also to produce synoptic views of ice motion and deformation in the polar regions (section 10.7). Passive microwave generates information at a synoptic scale (a few thousands of kilometers of an imaged scene), although it has been used to produce ice concentration maps at medium-resolution scales of a few kilometers.

Some of these algorithms have been used to generate useful records of ice concentration, extend, surface temperature, thickness (with some limitations), motion, and deformation. The information is used to support marine operational tasks as well as climate-related studies. However, the use of remote sensing data in operational sea ice monitoring programs still depends on visual analysis of satellite imagery data. The reason is the strict requirement of the operational analysis for robustness. Algorithms may not be reliable under all possible ice conditions, but the operational environment requires nothing less than reliability. Visual analysis of the data is certainly subjective, but it incorporates many factors that cannot be incorporated in a quantitative algorithm (e.g., climatic information, recent history of the ice field, records of meteorological data, in addition to other heuristic rules used by expert ice analysts). The approach of visual analysis of satellite images to retrieve operational sea ice information will probably continue for many years until a coincident multisensor data system is developed to acquire multichannel optical, thermal, and microwave data simultaneously.

In addition to the valuable information offered by remote sensing about sea ice, another unnoticeable positive development has been brought by this technology to the sea ice community. That is the engagement of many researchers and operators who come from the remote sensing systems and applications in the sea ice properties and physical processes. Many researchers who spend their career in developing hardware instruments or software methodologies for remote sensing data collection and analysis find themselves, at one point or another, involved in testing or validating their ideas using sea ice. Some of those people have later developed interest in sea ice as a phenomenon and pursued serious research work in this field. This happened to scientists who came from the fields of electrical or civil engineering, mathematics, earth sciences, physics or geography.

1.6. ABOUT THE BOOK AND ITS ORGANIZATION

This book combines information on two aspects of sea ice: physics and remote sensing. Research communities from the two disciplines do not usually interact; in general, the active members may not be aware of the details about each other's realm of interest or work. When geophysicists and materials scientists embarked on intensive studies of sea ice in the 1950s, they soon realized the need for tools to monitor ice conditions at larger scales. The scales of observations are certainly different, but microstructural features seen at the grain- or subgrain-scale level are also repeated at the floe-scale and floe-fragments levels. Macroscopic images often look very similar to space-borne images. The intergranular activities like shearing between the grains and subgrains are often reflected in the interfloe behaviors like ridging, rafting, and rubble formation. Similarly, intragranular processes like dislocations pileups, cell formations, or intragranular cracking activities appear to be similar to many deformation patterns seen at the large scales. The tools, airborne and the space-borne remote sensing, eventually became available by the mid-1970s. On the other hand, when the remote sensing researchers started their work to retrieve sea ice parameters from the observations, they too realized the importance of knowledge more on the physics of sea ice in order to support the interpretation of the data and the retrieval of ice information. Today, it is common to see basic information on ice physics cited in the literature of remote sensing, but it is not common to see the opposite. This book furnishes an opportunity for the two research communities to meet and perhaps learn a bit more about each other's work.

Knowledge about sea ice physics is relatively well established, while knowledge about remote sensing of sea ice is still in developing mode. The remote sensing information, however, has an edge because considerable amount of information has become widely available on the Internet. This availability does not apply to the information on ice physics. Unsurprisingly, at this time of information explosion, if the required information is not readily available on the Internet, the common notion is that it does not exist! This book is an attempt to restore a balance between the two subjects in a single presentation, although the vast areas of engineering physics of ice remained untouched here. It provides only the basic physics of sea ice for the ice remote sensing researchers to develop better physical insight into the subject of their study. It also provides a reasonable scope of applications of ice remote sensing for the ice physicists and geophysicist to comprehend the potential and limitations of the remote sensing applications.

The argument about the importance of a satisfactory level of knowledge about sea ice physics for remote sensing research applies laterally to research in other ice-related disciplines such as engineering, climatology, and glaciology. However, judging from several recent publications, it seems that researchers in these fields tend to remain confined within the level of knowledge about ice physics that they keep circulating since 1970s and 1980s. One of the reasons for this tendency is the fact that those researchers usually come from different scientific backgrounds. For example, ice-related engineering problems are dealt with by engineers trained mainly in the broad fields of civil, mechanical, or aerospace engineering to whom the ice is a "nuisance." Glacier ice and to some extent sea ice is studied mostly by the geographers trained in the vast aspects of geography but not necessarily physics. None of these communities, except for the physicists, treat snow and ice as extremely high temperature materials.

Remote sensing of sea ice has been conducted by researchers from a wide range of scientific background that encompasses computer sciences, environmental sciences, geography, and electrical and civil engineering. One goal of this book is to bring out some progress made in the field of ice physics that could expand the scope of knowledge of the researchers and operators in the different aspects of sea ice and hopefully influence their work.

The book is intended to reach out to a variety of sea ice audiences who study different aspects of the ice phenomenon related to physics, remote sensing, mechanical behavior, climatic impacts, operational monitoring, etc. For many readers, it is not crucial to gain deep understanding of ice as a material. For that reason the authors have tried to present the material in a simplified and gradual manner whenever possible so the reader can gain the amount of information that suits his/her purpose. To facilitate this presentation approach, definitions are introduced constantly with cross referencing, treatments of both physics and remote sensing material start from first principles, many illustrations are used, derivation of commonly used equations is presented, many references are provided, etc. In short the book is intended to be of educational value.

The book places more focus on sea ice research and applications achieved in Canada. A few Canadian institutions have been involved in different aspects of sea ice physics and remote sensing research and applications. The National Research Council of Canada undertook research into ice physics and mechanics to support shipping and engineering operations in ice-rich waters. The Canadian Ice Service of Environment Canada (the federal department of the environment) provides the most timely information about ice in Canada's navigable waters. Canada Centre for Remote Sensing (the government of Canada's center of excellence for remote sensing and geodesy) has conducted extensive research and science programs to support development of remote sensing applications of sea ice. The Canadian Space Agency made the data from the Canadian satellite Radarsat suitable and available for sea ice applications. A few academic institutes have expanded their research programs on several geophysical and climatic aspects of sea ice and its remote sensing applications. Among them is the University of Manitoba's Centre for Earth Observation Science, University of Calgary Department of Geography, and University of Waterloo Department of Geography and Environmental Management. Research and operational work from these and other Canadian academic institutes is quoted frequently in the book.

Materials related to physics of ice are presented in Chapters 2–6 while remote sensing materials are presented in Chapters 7–10. Chapter 2 addresses the initial formation and growth of sea ice with its two phases of lateral and vertical growth. It then presents models of ice growth to estimate its rate under different atmospheric, oceanic, and snow effects. This is followed by discussions on the processes of brine entrapment as the freezing proceeds at the ice-water interface and the subsequent brine drainage from the ice mass during the first year of its life. Forms of ice deformation, decay, and aging are also introduced with geometric characterization of some forms. The chapter concludes with a section on sea ice classes and regimes. This includes a brief account of commonly known regimes; namely polynyas, pancake ice, marginal ice zone, and forms of floating ice of land origin.

Chapter 3 presents data on basic physical properties of sea ice and derivation of simple models to determine elected parameters such as volume fractions of sea ice constituents and the dielectric constant of the two-phase sea ice composition. Typical values of key physical parameters are given. The chapter addresses three basic physical properties of sea ice (salinity, density, and temperature) as well as relevant thermal properties (conductivity and heat of fusion). The dielectric constant is presented in details because of its relevance to remote sensing.

Chapter 4 presents detailed information on polycrystalline structures of freshwater lake and river ice and sea ice. It starts with general structural features of ice and relates the information to other material at high temperature. A number of basic terms and definitions related to the polycrystalline aspects of ice that are used in this book are introduced. At the core of this chapter is the crystallographic classification of natural sea ice. This is the key information for the interpretation of texture of the ice, from which information about ice growth conditions can be inferred. Many examples of age-based sea ice types and their characteristic features are presented by photographs of thin sections. Information that can be retrieved from these data is presented in the last section (e.g., geometric characteristics of crystals, brine pockets, and air bubbles).

Chapter 5 is about two major field experimental programs, conducted in the 1980s in the western Arctic of Canada. The first is the Mould Bay experiment that marked the beginning of the field programs in support for the development of the Canadian Radarsat project. The study of sea ice from first-year to second-year type and the continuation of the growth of first-year ice under the second-year ice are two unique achievements from this program. The second was a program conducted on an ice island that broke away in 1982 from the East Ward Hunt ice shelf in Ellesmere Island, Canada. Only limited information from both of these experiments has been published. Thus this book is providing an opportunity to disseminate technical as well as human aspects of conducting long-term experiments on sea ice in the Arctic.

Chapter 6 is dedicated to the methods developed for revealing microstructural characteristics of sea ice and performing forensic type of investigations. Following an introduction to polarized light, polarizing sheets, large field-of-view polariscopes, and the birefringent properties of ice, it moves on to present detailed techniques for preparing thin sections of ice and snow. Special emphasis is given to the double-microtoming technique (DMT) that avoids the use of any warm-to-touch glass plates or surface melting—hence the best choice for saline ice. The procedures for examining and photographing thin sections under polarized and scattered light, or their combinations, are then presented. The chapter is concluded with descriptions of thermal and chemical etching, and the dual process of etching/replicating (DPER) technique in conjunction with scanning electron microscopy (SEM) for examinations of subgrain boundaries and substructures involving line defects or dislocations. These are not possible with the traditional polarized lights methods. Moreover, these procedures can be performed in the field and were actually used by the authors in the Arctic (using tents and make-shift shelters), and many examples from these field experiments are presented in this book. It is stressed that microstructural analysis should be performed in the field when the ambient conditions are cold (preferably below −15 °C) and immediately after recovering the samples from the sea.

Chapter 7 addresses a few concepts of remote sensing relevant to sea ice applications and particularly its parameter retrieval. The breadth of the material in this chapter is designed to appeal to researchers and users of remote sensing data who want to develop quick acquaintance with scientific issues that are outside their domain of experiences. After a historical synopsis of satellite remote sensing for sea ice, the chapter includes sections on electromagnetic wave properties and processes. Principles of optical, thermal, and microwave remote sensing are introduced in separate sections. More focus is placed on the imaging radar sensing since this is the prime data source for operational sea ice monitoring.

The chapter provides also a brief account of radiative processes in the atmosphere, ocean, and snow on sea ice. These three media contribute to the satellite observations of sea ice. The material in this chapter provides sufficient background to facilitate comprehending the information presented in the next chapters.

Chapter 8 presents data sets on radiative properties and satellite observations of sea ice, its snow cover, and surrounding open water. These include radar backscatter, microwave brightness temperature, visible and near-infrared reflectance and albedo, emissivity, and penetration depth in the microwave spectral region. Data are arranged according to the commonly used age-based ice types and for snow under different physical conditions (wetness, density, and grain size). Limited information is provided to explain the overlap between measurements from different ice types or snow cover. Most of the data are obtained from several satellite observations but airborne and ground measurements are also used. No attempt has been made to explain discrepancies (if they exist) between data sets of the same parameter from different sources. The data can serve as benchmarks to guide the process of retrieval of ice parameters from remote sensing data.

Methods of retrieval of surface features of sea ice are presented in Chapter 9. The features include surface deformation (ridges, rubble ice, and brash ice between floes), leads within the ice cover, forms (stages) of surface melt, and frost flowers cover on thin ice. For each feature the most suitable sensors are identified and the most common methods of retrieval are described. Geometrical characterization of ice ridging is included. The characterization of leads in terms of their geometries and the wind conditions that generate them is presented. Procedures for determining the onset of surface melting and its advanced phases using optical, passive microwave, and imaging radar data are also described. The advantage of combining observations to retrieve surface features is demonstrated. The chapter concludes with a presentation on frost flowers on thin ice surface in relation to the conditions of their formation and their effect on modulating the received radar backscatter.

Chapter 10 presents methods for the retrieval of seven sea ice parameters: ice types, concentration, extent, thickness, surface temperature, snow depth, and ice displacement and velocity. Whenever possible, methods for each parameter are grouped according to the type of observation: optical, thermal infrared, passive microwave, and radar. Ice type is an important operational parameter, and it can be retrieved from all categories of remote sensing observations. Special focus is placed on retrieval of ice concentration because this has been the most successful retrievable parameter from remote sensing observations. Detailed descriptions of four methods that represent four different retrieval approaches are included. As for the rest of the seven parameters, the most commonly used retrieval methods are presented. Once again, more coverage is given to the ice thickness retrieval. Some critical issues that affect the accuracy of retrieval of all parameters, especially ice concentration, are presented. The information in this chapter aims at understanding the potential and limitations of the ice parameter retrieval algorithms.

Chapter 11 presents a brief historical account showing the development of interest in Arctic sea ice in Canada and a synopsis of the Canadian ice monitoring program, which is currently operated by the Environment Canada (the federal Department of the Environment). The historical narration focuses on the turning point when sea ice became a service offered by the federal government after it had been just an issue that had to be "attended to" for a few centuries. The material on operational ice monitoring program and the examples of operational ice charts products from CIS serve as supporting material to a few parts throughout the book. In general, this chapter adds another human dimension to the sea ice material in this book.

2

Ice Physics and Physical Processes

Development of a sea ice cover at any previously ice-free location encompasses a number of phases: (1) nucleation of ice crystals and initial formation, including consolidation that depends on the prevailing weather and atmospheric conditions, (2) vertical growth or "ice congelation," (3) motion and deformation depending on sea and wind states, (4) melt and decay, and (5) aging. Wind, current, and wave-induced motion and associated strain (hence stress) stimulate deformation in different forms. Cracking, rafting, raised edges, and piling are common forms in thin floating ice with thickness less than 0.3 m, known as young ice. Formation of ridges and rubble fields are common in thicker ice. Grounded and/or land-fast ice (i.e., fastened to the shore) are obviously less affected by these elements. Shore fast ice is, however, subjected to the effects of tides, forming cracks, hinges, and rubbles near the shorelines.

This chapter focuses on the first two phases with its major processes of initial ice formation, lateral and vertical growth, entrapment of inclusions, and the continuous desalination processes. It also includes a brief presentation on the synopsis of ice motion, deformation, and decay. Aging of ice is addressed in Chapter 5.

Formation and growth of ice entails three major stages. The first is the initial formation when minute ice crystals are nucleated in the seawater and grow sufficiently to recognizable sizes and shapes in the form of needles or tiny discs, called frazil ice. The frazil crystals are usually suspended in water and are often herded by wind action. At this stage of ice formation, the water has a slushy consistency. The ice particles are loose and isolated and have not yet frozen to the point of consolidation. The second stage is marked by lateral growth of the spicular crystals to form small rounded discs or flat patches, and the crystals have coagulated to form a soupy layer on the surface, known as grease ice. These two stages are well known to the Inuit as *qinu* (in Inuktitut) as described in Chapter 1. In the third stage, ice congelation starts when it grows vertically (i.e., thickens). These three stages are addressed in section 2.1. on initial ice formation and section 2.2. on ice growth.

One of the key processes in sea ice formation and growth is the rejection of salts originally dissolved in the seawater. This takes place mainly at the ice-water interface as water solidifies, adding to the volume of pure ice at the interface and growth of thickness of the ice sheet. During the initial stages of solidification, the rejection of salts is associated with the nucleation of crystals of pure water molecules. The crystal lattice has no room to accommodate any salt molecules. As these crystals grow in size and numbers, the entire surface is covered with solids with arms or dendrites at the ice-water interface (section 2.3.1.). This convoluted geometry, in turn, causes some of the rejected salts to be entrapped into the gaps between the ice discs or plates and eventually become enclosed within the volume between the plates as brine. This is what has become known as "brine pockets." Brine pockets could be interconnected or merge to form brine channels during the early growth period. At this stage, the brine continues to drain rapidly to the underlying seawater (with also some expulsion to the surface depending on the ambient conditions). The desalination processes continue throughout the lifetime of a sea ice cover. These processes are addressed in details in section 2.3.3.

Since ice is one of a very few crystalline substances for which the solid phase is less dense than its melt, it always floats on the water surface. Unless this floating ice becomes grounded to the shore land in the form of land-fast or simply fast ice, it may undergo a complex motion at different scales, leading to several forms of deformation of the ice cover. Mobility and deformation of sea ice are presented in section 2.4., while ice decay is addressed briefly in section 2.5. Ice classes and regimes are

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presented in section 2.6. Ice classes address the traditional age-based ice classification system [developed by the World Meteorological Organization (WMO)], which is used in the operational sea ice monitoring programs. "Ice regime," on the other hand, includes a wide ranging phenomena defined by their origin or formation processes such as pancake ice, congealed ice, fast ice, marginal ice zone, polynyas, and floating ice bodies of land origin.

2.1. INITIAL ICE FORMATION

Nucleation of ice crystals in freshwater lakes, rivers, or oceans with saline water occurs when the water surface cools down to the freezing temperature. As described earlier, the onset of freezing is seawater characterized by the formation of crystals known as frazil ice; a collection of loose, randomly oriented needle-shaped ice crystals in the sea or in freshwater. Since the formation of ice could originate in fresh or saline water, a few relevant properties of both types are discussed first.

2.1.1. Relevant Seawater Properties

Freshwater lakes and rivers contain very little dissolved impurities (less than 0.5%) and their effects on the growth of ice can be neglected for all practical purposes. That is not the case for sea water. The two properties of seawater that are directly related to sea ice formation are its salinity and density. The isotopic composition of water in case of freshwater ice may not be considered as important, but it may play some role in the case of seawater and ice formed from seawater. It is relevant because snow cover on sea ice may have a slightly different isotopic composition than the seawater. Consequently, this property may be used in distinguishing snow ice, formed from solidification of snow saturated with meltwater, from frazil ice formed from the seawater (section 4.3.3). The isotopic composition of glacier ice cores at decadal resolution are routinely used to infer climate record because the land-based ice bodies developed from snow depositions in the past [*Von Grafenstein et al.,* 1999]. A brief account of the molecular composition of water, including its isotopic composition, is introduced followed by another introduction to the two major properties of water that affect the ice formation and growth, namely salinity and density.

2.1.1.1. Molecular Composition of Water

The atomic and the molecular structure of water in all its phases (liquid and solid) have intrigued scientists for many centuries. Although the chemical formula of water is very simple, its atomic structure, depending on temperature and mechanical or electrical forces, is extremely complicated and difficult to study. The stoichiometric composition of water has been investigated by numerous investigators over a very long period of time and has been documented thoroughly in many reports and books. It was known for a long time that if two atoms of hydrogen are combined with one atom of oxygen under the same temperature and atmospheric pressure, then a molecule of water vapor forms. According to Avogadro's hypothesis, this leads to the basic molecular structure of pure water consisting of two atoms of hydrogen and one atom of oxygen, or simply H_2O. The ratio of the combining volumes of oxygen (O_2) and hydrogen (H_2) at 0 °C and a pressure of 0.76 mm Hg is $O_2/H_2 = 1/2.00288$. Pure water, however, may be divided into several types depending on isotopes of oxygen and hydrogen.

Hydrogen has three isotopes: ^1H (known as H), ^2H (known as deuterium, D) and ^3H (known as tritium), which is radioactive and its half-life is 2.5 years. Oxygen, on the other hand, has six isotopes: ^{14}O, ^{15}O, ^{16}O, ^{17}O, ^{18}O, and ^{19}O (oxygen ^{16}O is simply known as O). The three isotopes of ^{16}O, ^{17}O, and ^{18}O are stable, whereas ^{14}O, ^{15}O, and ^{19}O are radioactive but short-lived. The combination of oxygen and hydrogen atoms leads to six stable isotopes of water formed from stable atoms of hydrogen and oxygen.

The isotopic content of water in oceans, lakes, and rivers varies depending upon its origin. It contains about 99.73% of $^1H_2{}^{16}O$, 0.2% of $^1H_2{}^{18}O$, 0.04% of $^1H_2{}^{17}O$, and 0.03% of $^1H^2H^{16}O$ [*Shatenshtein et al.,* 1960]. The last one is simply termed HDO. However, if both the hydrogen atoms in this compound are deuterium (^2H), i.e., the molecule is $^2H_2{}^{16}O$ or D_2O, then the compound is known as heavy water. Based on the above percentages, ordinary water, which is generally referred to as H_2O, is mainly composed of $^1H_2{}^{16}O$.

The isotopic water composition has an impact on evaporation and consequently on the composition of snowflakes in the upper atmosphere. At a given temperature, the evaporation of the two major isotopes of water, $^1H_2$ ^{16}O and 0.2% of $^1H_2$ ^{18}O, depends on their masses. Naturally, the heavier water evaporates at a slower rate than the lighter one. As the temperature rises, the proportionality of these two varieties of water increases. Water containing the heavy oxygen ^{18}O evaporates more under warmer atmospheric temperature, making higher percentage of the relevant water isotope in the falling snow. This allows for the retrieval of paleoclimatic information to understand the evolution of the cryosphere. Although this information is lost when snow falls on land, rivers, lakes or oceans it is maintained when snow is deposited on areas that could sustain it for long periods. Such areas include ice caps and sheets, which are formed from the massive accumulation of snow on flat plateaus such as those found in several islands in Canada (i.e., Ellesmere Island, Penny Ice Cap in Baffin Island), Greenland, and Antarctica.

A chemical analysis of deep ice cores extracted from sections of ice caps on top of flat bedrock, where ice had settled with no side movement, provides insight into the climatic conditions of the period related to the analyzed section of the core. This can be achieved by measuring $\delta(^{18}O)$, which is expressed as the fractional difference between the ratio $^{16}O/^{18}O$ in the sample and the ratio in "standard mean ocean water" (SMOW) measured in percent. In general, the values of $\delta(^{18}O)$ in ice cap snow are negative. Since ^{18}O is heavier than ^{16}O, then water molecules containing the former evaporate proportionately less than the latter. Nevertheless, under warmer atmospheric temperatures, more molecules containing ^{18}O evaporate. This means that higher negative values of $\delta(^{18}O)$ found in the precipitated snow are likely associated with a warmer climate although there are also a number of other effects that can alter this parameter [*Hobbs,* 1974]. Since each depth segment of an ice cap core has originated in the past, chemical analysis of ice cores at different depths can then be used to identify relatively warmer or colder periods of snow deposition.

Because snow has a lower content of ^{18}O than ^{16}O compared to seawater, granular ice, formed from slush of snow saturated with meltwater, can be chemically distinguished from frazil ice formed from slush containing seawater and frazil crystals. These two types of ice may have similar appearances and may not be identified unless forensic type of microstructural analysis are carried out on both horizontal and vertical sections. *Jefferies et al.* [1994] used both crystalline structure and oxygen isotopic composition analysis of sea ice cores from the Ross Sea in the Antarctic to determine the amount of snow that contributed to the development of sea ice. In fact, for first-year sea ice in Okhotsk Sea, *Toyota et al.* [2007] found measurable differences in the values of $\delta(^{18}O)$ for snow, frazil, and columnar ice.

2.1.1.2. Seawater Salinity

Seawater holds the same isotopic water composition as freshwater, yet it contains a considerable percentage of dissolved salts and gases. Water salinity is the most relevant property for sea ice formation, composition, and growth. It is usually measured as the ratio of the weight of salts (in grams) dissolved in one 1000 g (1 k) of seawater. Hence, it is usually presented in parts per thousands (ppt or ‰). Alternatively, oceanographers define ocean salinity in terms of the practical salinity unit (PSU), which is the conductivity ratio of a seawater sample to the conductivity of a solution of potassium chloride (i.e., the standard solution of measuring electrical conductivity). The first measure is adopted in this book.

Seawater salinity is 35‰ on average in most marine areas, though slightly higher values (36‰) are observed in some regions in the Atlantic Ocean and Indian Ocean and slightly less values (34‰) are observed in the polar region. Within the Canadian Archipelago, water salinities are often found to be in the range of 30‰. The brackish water is usually found in estuaries, inland seas, or lakes. Salinity is affected by the relative amount of precipitation and evaporation as well as the mixing with freshwater in the vicinity of river mouths. If a large river is emptying its water into a sea, the local seawater salinity is significantly less than the typical value of 35‰. Variation of salinity is one criterion for water classification. It is also revealed in different mechanisms of sea ice formation as well as structure and physical properties of ice. Freshwater implies salinities less than 0.5‰, while brackish water bodies have salinities more than that of freshwater but less than the usual salinity of seawater. Thus brackish water is defined as water having salinities in a wide range, between 0.5‰ and 29‰ According to this nomenclature, saline water can be defined as water with salinity ranging from 29‰ to 50‰. Brine is a term that describes water with salinity higher than 50‰.

The primary salt dissolved in seawater is sodium chloride (NaCl), but other salts exist such as sodium sulfate ($Na_2SO_4 \cdot 10H_2O$), magnesium sulfate ($MgSO_4$), and magnesium chloride ($MgCl_2 \cdot 12H_2O$ and $MgCl_2 \cdot 8H_2O$). The ionic proportion of chloride, sodium, sulfate, and magnesium in these compounds (regardless of the water salinity) is close to 55.03%, 30.59%, 7.68%, and 3.68%, respectively. The rest of the ionic proportion, namely 3.02%, is composed of calcium (Ca), potassium (K), bromide (Br), cobalt (CO), and other elements in negligible amounts.

Solutes in seawater are rejected during the process of solidification. Most are rejected to the water at the ice-water interface. Some are entrapped as inclusions within the ice mass. Brine inclusions are known as brine pockets. They are located, as will be shown later, along boundaries and subboundaries of ice crystals. The entrapped brine naturally remains in sea ice in thermal equilibrium with the surrounding ice. The temperature of the ice determines the salinity of the liquid in the brine pockets. With the decrease in temperature, different solutes in the brine start to precipitate at different temperatures. This precipitation process makes the ionic concentration of the liquid as temperature dependent, as can be seen in Table 2.1.

Table 2.1 Major salts in sea ice and their precipitation temperatures.

Salt Name	Composition	Precipitating Temp. (°C)
Calcium carbonate	$CaCO_3 \cdot 6H_2O$	−2.20
Sodium Sulfate	$Na_2SO_4 \cdot 10H_2O$	−8.20
Magnesium chloride	$MgCl_2 \cdot 8H_2O$	−18.0
Sodium chloride	$NaCl \cdot 2H_2O$	−22.9
Magnesium chloride	$MgCl_2 \cdot 12H_2O$	−36.8
Calcium chloride	$CaCl_2 \cdot 6H_2O$	−55.0

Adapted from *Weeks and Ackley* [1982].

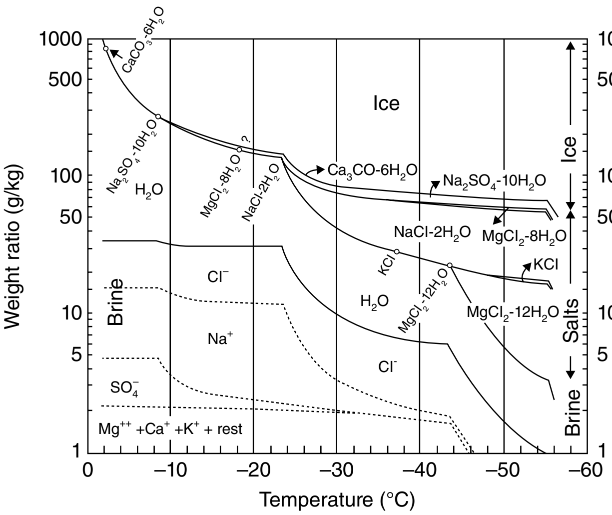

Figure 2.1 Sea ice phase diagram [*Assur,* 1958].

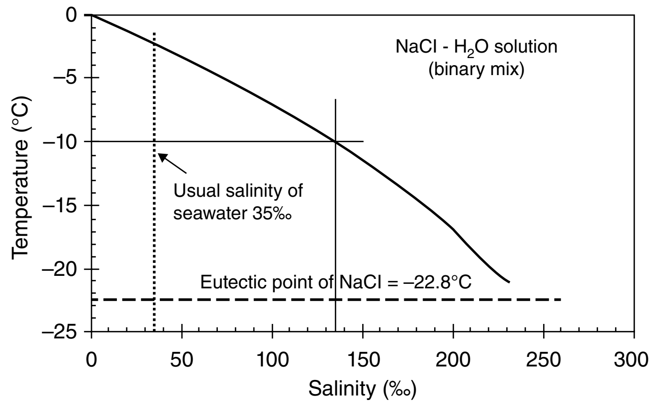

Figure 2.2 Phase diagram of binary mixture consisting of NaCl and H_2O.

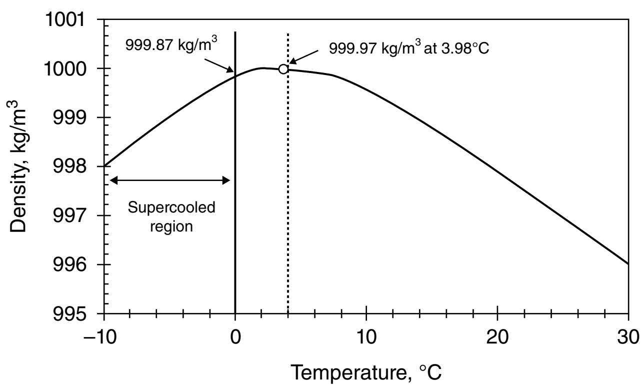

Figure 2.3 Temperature dependence of density of pure water; Note the maximum density at 4 °C.

From the viewpoint of materials science, natural sea ice is a very complex material. In order to understand the physics of sea ice, one must have an appreciation of the so-called sea ice phase diagram. A phase diagram illustrates the amounts and composition of the phases (ice, brine, and precipitated solid salts) that exist at different temperatures as a given volume of seawater freezes and its temperature decreases. The reader must consult *Weeks* [2010] for a general understanding of this topic. Only a very brief description will be given here. A comprehensive phase diagram for "standard" sea ice was developed by *Assur* [1958] and is presented in Figure 2.1. It shows the weight ratio of each component: pure ice, salts, and liquid brine at any temperature when the three phases are in equilibrium. Temperatures for precipitation of different salts are also shown. For all practical purposes, a phase diagram for the binary mixture of sodium chloride and water is sufficient to represent the salt contents in sea ice. Only that part of the phase diagram is presented in Figure 2.2. If brine exists in sea ice at −10°C, for example, some water in the brine mixture has to freeze in order to bring the mixture to its equilibrium concentration point, i.e., to a salinity of 135‰ as shown in Figure 2.2. As the temperature of the mixture decreases further, more water molecules continue to solidify until the temperature reaches −22.8°C, which is the eutectic temperature of NaCl. At the eutectic temperature all sodium chloride precipitates with no liquid left in the binary mixture.

2.1.1.3. Sea Ice Density

Another property of "pure" water, which is quite relevant to ice formation and growth processes, is its density. Much like any other material, water is subjected to thermal expansion as its temperature rises or contraction with the decrease in temperature. However, the behaviour of pure water is very interesting very close to its solidification point in the temperature range of 4 °C–0 °C. Lowering the temperature below 4 °C causes the density not to increase, as expected, but to decrease as shown in Figure 2.3. The maximum density (at 4 °C) is 999.972 kg/m^{-3}. Another property of pure water, though not unique to water, is the fact that it can be supercooled far below the normal freezing point without solidification. As the water is supercooled below the normal freezing point, the density continues to decrease with the decrease in temperature, as shown in the figure. This behavior affects the freezing mechanism of pure water as will be seen later in the next section.

Dissolved salts in water depress the temperature of the maximum density, below that of pure water (about 4 °C) as well as the freezing point. For seawater, however, both the temperature of maximum density and the freezing point decrease almost linearly as the water salinity increases, as illustrated in Figure 2.4. At a critical salinity of 24.69‰, the two temperatures are equal to −1.32°C. For salinities higher than the critical value, the freezing temperature is higher than the temperature of maximum density. At salinity of 35‰, the two temperatures are −1.88 and −3.5°C, respectively.

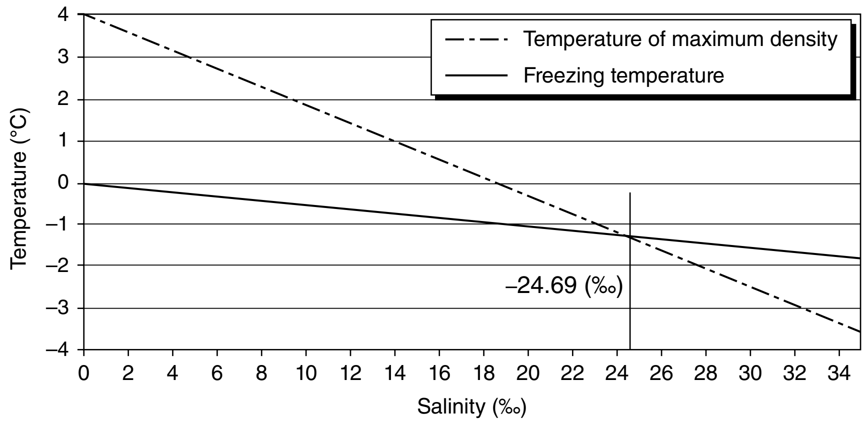

Figure 2.4 Dependence of the temperature of maximum density and the freezing point on water salinity; the two curves intersect at the salinity of 24.69‰.

It is appropriate to mentioned here that the density of seawater at the surface of the ocean varies between 1020 and 1029 kg/m^3 depending on the salinity of the water. Although the density of pure ice is about 917 kg/m^3, the density of newly formed sea ice is only marginally less than 1000 kg/m^3 due to its high brine content. Nonetheless, it is lighter than the sea water and floats on the surface. Of course brine in the ice drains very quickly and makes the ice even lighter.

Air is also dissolved in seawater. Through the constant stirring of the sea surface by wind and waves, gases are transferred from the atmosphere to the water. The three common gases that make up 99.93% of the atmosphere are nitrogen (78.08%), oxygen (20.95%), and argon (0.93%). The sum of their total percentages in the dissolved air in seawater is 98.5% (62.6% nitrogen, 34.3, oxygen, and 1.6% argon) [*Pilson,* 1998]. The solubility of oxygen in water is relatively higher than nitrogen. This results in a lower nitrogen content in water and hence ice. Carbon dioxide represents only 0.035% of atmospheric gases, but it constitutes a relatively larger percentage (1.4%) of the total air dissolved in seawater. The marine organisms may be a factor that contributes to the higher CO_2 percentage. As the temperature or the salinity of seawater increases, the amount of gas that ocean water can dissolve decreases slightly. When seawater freezes, air is segregated and entrapped in the form of pockets along with brine as the pockets at the intercrystalline boundaries.

2.1.2. Seawater Freezing Mechanism

As mentioned above, the water salinity is the controlling factor that governs the temperature dependence of both the freezing point and the density of the water. Consequently, the freezing mechanism differs between fresh, brackish, or seawater, though they all freeze when the water surface is cooled down to or below its freezing temperature. Additionally, freezing requires the presence of a nucleus around which ice crystals can form. In case of natural water bodies nuclei could be dust particles, snowflakes, frozen water droplets or any type of impurities deposited at the upper surface of the water. In the absence of such nuclei, water can remain in liquid phase at temperatures well below its freezing point. For pure water, this phenomenon is called supercooling. For impure water, such as seawater, brackish, or even freshwater with low salinity, it is known as constitutional or compositional supercoiling, as will be clarified later in section 2.3.1. Although by definition, pure water freezes at 0°C under normal atmospheric pressure, it can be supercooled under that pressure down to about −42°C. Considering the large amount of dissolved impurities in seawater compared to freshwater in lakes and rivers, the amount of supercooling in seawater is probably a few hundredths to tenths of a degree below the freezing point [*Weeks and Ackley,* 1982].

Ice formation in freshwater entails the following processes. As the water surface cools, the cooler layer at the top becomes denser and therefore sinks, allowing warmer water to rise. This vertical convection continues until the surface temperature reaches the critical temperature of maximum density, about 4.0°C. At this temperature the convection essentially stops. Further cooling of the surface layer makes it less dense and therefore remains at the surface. The surface then responds faster to any further drop in atmospheric temperature until it reaches the freezing temperature of 0°C. At this point ice crystals start to form around appropriate nuclei. It should also be pointed out here that freshwater (depending on its purity and the absence of nucleating agents) can be supercooled well below that temperature if it is not mechanically disturbed. The temperature of the water under the newly formed ice away from the interface remains isothermal at 4.0°C. It gradually cools as a result of the heat exchange at the ice-water interface and the freezing continues. The density of clear freshwater lake ice is slightly less than about 917 kg/m^3, or close to that of a single crystal of pure ice. This is about 10% less than the density 999.8 kg/m^3 for pure water at 0°C.

Seawater, as a solution, undergoes a markedly different freezing process from that of freshwater, particularly when the salinity exceeds the critical value of 24.69‰. It can be seen in Figure 2.3 that for salinities below this critical value the temperature of the maximum density is higher than the freezing temperature of the seawater solute. As the surface of the seawater cools, the denser cooler layer also sinks and a vertical convection current is created similar to the case of freshwater. The difference, however, is that this current does not stop before the surface reaches the freezing point. The freezing point of water with salinity 35‰ is −1.8°C, which is higher than −3.5°C, the critical temperature, of maximum density for the solution. This means that the entire depth of the seawater body must be cooled to the freezing temperature before any formation of ice. But this does not happen in nature. The convection current is usually limited to the upper 50–200 m depth, called the mixing layer of the ocean, as can be visualized from Figure 2.5.

Many areas of the ocean are stratified with increasing salinity and hence become denser toward the bottom. When the colder and denser upper area sinks, the increased density of that layer may not be as high as that of the lower layer, known as a pycnocline zone, which features a rapid increase of density with depth due to continuous change in its temperature or salinity. This stratification imposes a lower limit to the vertical convection in the mixing layer. This limit is about 50 m in the Arctic Ocean, which has slightly less salinity than the Atlantic Ocean. Any further cooling of the surface below the freezing temperature of the isothermal mixing layer will cause surface freezing.

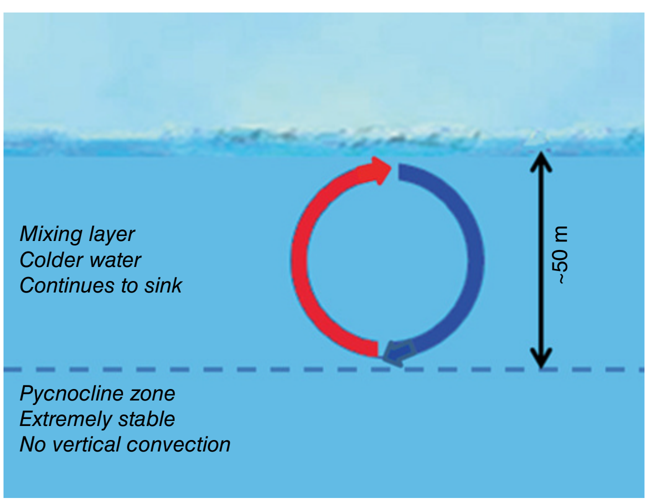

Figure 2.5 Illustration of the lower limit for convection current in seawater appropriate for the Arctic.

Differences in the freezing processes in freshwater and seawater can then be summarized as follows. In the case of freshwater, the convection current stops when the surface temperature reaches about 4°C (the value corresponding to maximum density). Therefore, there is a time lag between the end of the convection current and the actual start of freezing when the surface temperature reaches the freezing point. This does not happen in the case of seawater as the convection current stops at the moment when the surface temperature reaches the freezing point, depending on water salinity. Consequently in order for ice to form in seawater, more heat has to be removed compared to that required for freshwater. In other words, sea ice formation is relatively slower than freshwater ice. As a rule of thumb, sea ice forms faster under any of the following three conditions: (1) in areas of low salinity such as those near the mouths of rivers, (2) in areas where there is no currents such as inlets, bays, and straits, and (3) in areas of shallow waters near coasts or over shoals or banks where only a small body of water must be cooled.

2.1.3. Initial Ice Crystals and Frazil Ice

Attaining the freezing point is not sufficient for water to freeze because the formation of ice crystals requires nuclei to start solidification. In primary cryospheric regions when the ambient air temperature is below 0°C, small ice nuclei are almost always present in the atmosphere. These ice particles are deposited continuously on seawater surfaces to initiate nucleation of crystals at the surface. If the water surface is very calm and the air temperature decreases slowly to develop only small temperature gradient, the surface water may be supercooled and crystals may be nucleated. Once nucleated, the pure single crystal of ice takes a spherical shape because the surface-to-volume ratio is a minimum for a sphere. The spherical shape, however, is not sustained due to the anisotropic growth rates in different planes of the hexagonal lattice of ice crystals. The crystal starts growing parallel to its basal plane with *c* axis normal to this plane (definitions are given in section 4.1.3). It takes the form of a discoid because the growth in the basal plane is preferred in hexagonal crystals like ice and will be described in detail in section 4.2.1. This process of initial growth following the nucleation processes takes place irrespective of the salinity of the water. Under calm conditions, without any mechanical or thermal disturbance, each discoid floats at the surface. Each of the tiny disc float with their flat surface parallel to the water surface or horizontal plane. Since the major surfaces of these discs are parallel to the basal plane, the *c* axis of these crystals tends to be oriented in the vertical plane. Each of the floating discs, therefore, is oriented favorably for growth in the horizontal plane. If the initial ice crystals are nucleated around snowflakes, during snow flurry activities in the air, then those crystals with their flat surfaces parallel to the surface tend to grow faster than the others. Of course, this is not the case when snowfall occurs and covers the water surface. In those situations, the initial growth processes are complex and flakes with their flat surfaces normal to the water surface are in favorable conditions for growth (for details on this topic, see section 4.3.3.4).

Knowledge of the early stages of nucleation of crystals at the snow-free surface of saline water and the initial growth comes from studies conducted in the laboratories [*Weeks and Assur,* 1963; *Weeks and Lofgren,* 1967; *Cox and Weeks,* 1975]. Under calm conditions, due to the favorable orientation mentioned above, ice discs continue to grow laterally up to 2–3 mm depending on the salinity of the water [*Weeks and Ackley,* 1982]. The freezing front becomes wavy as the disc size increases. Initially circular boundaries assume tree-like or dendritic structures by developing "arms" around their peripherals as shown schematically in Figure 2.6. The term dendritic comes from dendrology—the study of trees. This is true for both freshwater and seawater ice but due to different mechanisms. In freshwater ice, with low concentration of solutes, formation of dendrites is primarily a thermally induced phenomenon. The temperature gradient around the crystal is asymmetric due to complications of heat flow. This is mainly due to the instability at the growing front induced by increase in size of the discs. A small perturbation at the ice-water interface ends up in even more supercooled liquid so the interface becomes more unstable. In saline water, on the other hand, initiation of dendritic arms is a salinity-related phenomenon. As a discoid continues to grow it pushes the salt-rich solute to its boundaries. Salt concentration starts to build up around the disc with varying distribution leading to varying degrees of supercooled water. This causes instabilities in the freezing such that any further crystal growth takes place anisotropically (though still in the horizontal plane). The growth direction follows the path of supercooled water, i.e., the direction of least resistance for growth. Details on this aspect of growth are provided later in the section 2.3.2.

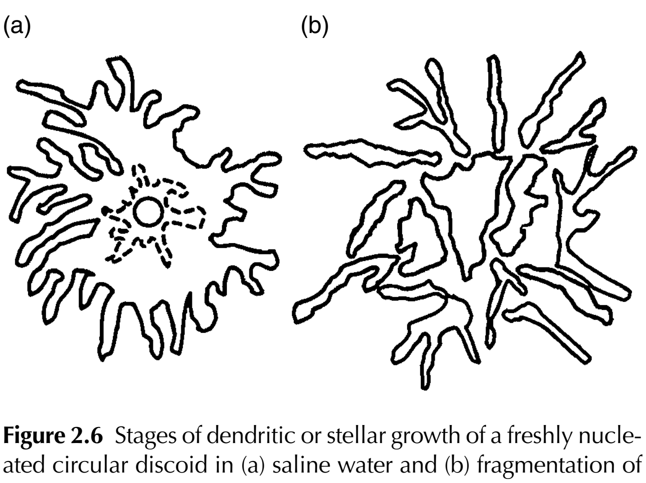

Figure 2.6 Stages of dendritic or stellar growth of a freshly nucleated circular discoid in (a) saline water and (b) fragmentation of the stellar crystal.

The initial discoidal crystal shape develops into dendritic or stellar form with very fragile arms. The average diameter of these crystals may be around 2.5 mm [*Weeks,* 1959]. The change from discoidal to stellar crystal shape is accelerated under rapid cooling when the solute is rejected at a rate fast enough to cause significant variation of the salt distribution in the vicinity of the crystal in the plane parallel to the major plane of the discoid. Moreover, the arms of the stellar crystals may grow thicker because they are surrounded by supercooled water and may not be able to sustain their own weight and consequently break.

As the growth continues, the fragile arms of stellar crystals, therefore, start to break off and form needle-shaped crystals as illustrated in Figure 2.6b. This is triggered mainly by the wave actions that tilt and bend the delicate discs to fracture into segments. Actually this type of fragmentation of thin ice discs occur in both freshwater and seawater. Early observations [*Suzuki,* 1955] showed that a fewer number of needles are observed in case of freshwater ice. These needles are confined to the thin supercooled layer at the surface [*Hallet,* 1960]. It is well known that frazil crystals are not necessarily limited to thin layers at the water surface. Turbulent wave action also causes inclined discs and needles to sink into the thick supercooled mixed layer.

The needle-shaped ice crystals may be geometrically oriented or randomly oriented and usually carried away by the ocean surface currents. The ocean wave may tend to compress those needles and orient them with their long axis parallel to the vertical plane. The spatial distributions of the discoid, the broken discoid, and the needle-shaped crystals are usually sporadic and loose. The frazil needles are typically less than 1 mm in thickness and a few millimeters to a few tens of a millimeter in length. The aggregation is known as frazil or grease ice [*Weeks and Ackley,* 1982]. Reviews on frazil ice in rivers and ocean can be found in *Osterkamp* [1978] and *Martin* [1981].

When frazil crystals cluster together over a large area of a water surface, they dampen surface motion and give the ocean surface a greasy or oily appearance. It is a soupy layer on the ocean surface that does not reflect much light, hence gives the surface a dark matt appearance. It is known as *qinu* in Inuktitut, the language of the indigenous Inuit people in the Canadian Arctic and Greenland as described in Chapter 1. In general, this is known as "grease ice," though it is not actually solid ice. Ocean currents cause frazil crystals to herd into streaks or pile up downwind against the edges of floating ice floes to depths that can be on the order of 1 m [*Martin,* 1981; *Sinha,* 1986]. Figure 2.7 provides an example of frazil ice

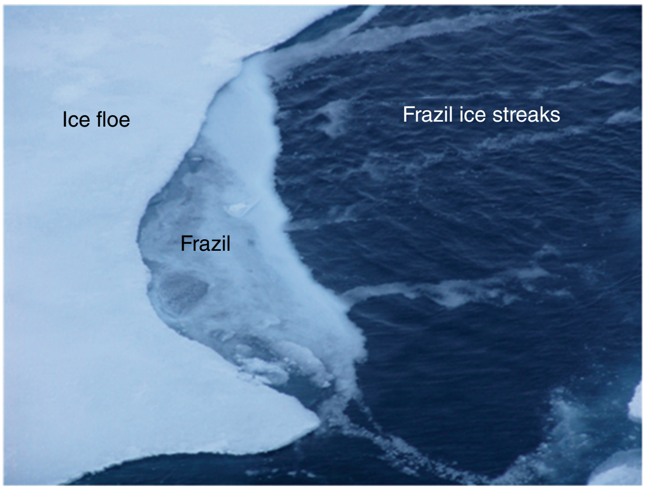

Figure 2.7 Frazil ice blowing into bands and accumulated next to an ice floe (adapted from the collection of Dr. Pablo Clemente-Colon, National Ice Center).

blowing into bands and accumulating against the edge of an ice floe.

Due to the elongated shape, frazil needles may not create enough buoyancy force to keep them floating and can be carried by the turbulent currents to areas below the water or ice surfaces. It is, therefore, not uncommon for frazil crystals to exist as suspended elements in deeper waters or solid masses below ice sheets. When frazil crystals form a layer at the ice-water interface at the bottom of ice sheets, the crystalline growth habit of the ice cover, such as columnar structure, is interrupted. However, as will be seen in Chapter 4 dealing with microstructural aspects of natural ice sheets, crystals of frazil ice may also act as seeds for new growth of columnar grained ice.

During the early part of the growth season, frazil crystals in fjords and channels can also be herded and pushed by the wind toward the shores, which eventually consolidate to form thick layers of vertically oriented crystals. Microstructure and strength properties of this type of frazil ice have been examined in detail by *Sinha* [1986], and a brief description of the structural aspects of this type of frazil ice is given in section 4.3.3.2.

Before closing this section, it should be mentioned that the initial ice cover can also develop from snow deposition on open-water surface and eventual freezing of water-saturated snow. Relatively thin ice covers can also be thickened by the solidification of flooded snow overlaying the existing ice cover. This commonly occurs when a snow cover becomes heavy enough to depress the ice surface below its freeboard. The influx of seawater through the permeable snow saturates the snow mass, which subsequently freezes into what is known as granular or snow ice. Microstructural aspects of granular, snow ice will be presented in section 4.3.3.1.

2.2. ICE GROWTH

Understanding the processes of ice growth, particularly during the early seasons of freezing and growth, is important both from the climatic and remote sensing viewpoints. Ice grows mostly thermodynamically due to the colder temperature of the atmosphere with respect to the temperature of the seawater. It can also grow, to a much lesser extent, mechanically as a result of its mobility. Mobility of ice sheets enhances the possibility of cracking the sheets or breaking the edges of ice floes on collision and piling up of the broken ice blocks.

The rate of thermodynamic growth of sea ice depends mainly on three factors that can be measured: air temperature, ice thickness, and snow cover. Other factors include the solar radiation, wind conditions, and the density and albedo of snow. Some of these factors are often difficult to quantify. The apparent or macroscopic and crystallographic form of ice at any growth stage particularly during the early growth is determined mainly by the oceanic conditions, whether calm or turbulent. Wind and ocean current are also active factors because they determine the mobility of floating ice. One of the most important processes that occur during sea ice growth is the brine rejection to the underlying seawater and the brine entrapment within the ice mass. After the initial formation of ice discoid with its dendritic boundaries as described earlier, the ice growth commences laterally as the minute ice discoid grow sidewise or frazil crystals herd horizontally. The growth will then proceed vertically in the direction of maximum heat flow.

2.2.1. Lateral Ice Growth

The general characteristics of thin ice covers depend on the state of the ocean surface, namely whether it is quiescent or turbulent. For relatively calm atmospheric and oceanic conditions, ice discoids and frazil crystals or frazil streaks that are formed during the onset of freezing continue to grow sideway until they touch each other and cover the entire surface area of the water. Under a quiescent water surface, frazil crystals can also be herded to form streaks of ice. The ice particles eventually consolidate if cold temperature persists and the water surface continues to be calm, leading to the formation of nila. This is a continuous but flexible sheet and translucent in the beginning. Initially, it has a dark appearance but it takes on a gray appearance as the sheet continues to thicken (Figure 2.8). Nilas may grow up to the thickness range of 10–20 mm if the water surface remains calm, but this rarely happens in open seas. In nature, even under relatively calm weather conditions, there could be oceanic currents and low-amplitude waves generated by convections or very light breeze. Nilas are usually broken into large pieces (a few meters to tens of meters wide) due to wind effect and/or oceanic conditions. While floating and moving, the fractured pieces may slide over each other to form what is known as surface rafting (Figure 2.9). This phenomenon is discussed in more details in section 2.4., but it suffices here to mention that rafting is a characteristic of nilas. In fact, rafting is a useful surface feature to identify nilas in remotely sensed images.

The second scenario of lateral ice growth is observed under wind and turbulent ocean surface conditions. This is commonly seen in open seas where the water surface is usually more turbulent than lakes. When the ocean surface is rough at the time of initial ice formation, turbulence will not allow consolidation of the herded frazil crystals into nilas. Instead, it causes frazil to undergo cyclic compression following the wave action. If compressed enough, herded crystals may bond with each other due to the freezing of water between them. Eventually they may form discs that may start as slush before the solidification when water between crystals continues to freeze. It is, therefore, common to see the formation of small and thin pancakes during the very beginning of freezing as shown in Figure 2.10. Actually, these small features can also become an integral part of the nilas leading to different shades of grayness, as can be noticed in Figure 2.8.

The disc-shaped features are called pancakes simply because they are round or oval in shape. The diameter of these discs range from a few fractions of a meter to several meters. Often their edges are raised as results of rotation and collision against each other by the ocean-induced motion. Figures 2.10, 2.11, and 2.12 illustrate different stages of the growth of pancake ice. It can be seen that the larger pancakes are formed by the consolidation of smaller ones. Large pancakes may also break due to wave actions. An example of cracked and separated fragments of a large pancake is marked as (a) in Figures 2.12.

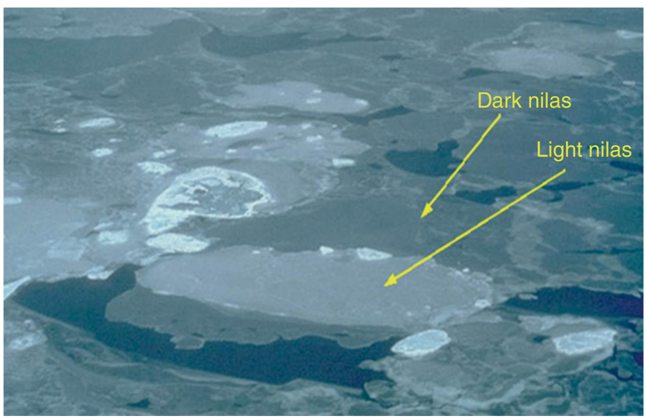

Figure 2.8 Dark and light nilas in Hudson Strait, Canada, during mid-November, photograph taken from altitude of about 700 m (Courtesy: Canadian Ice Service).

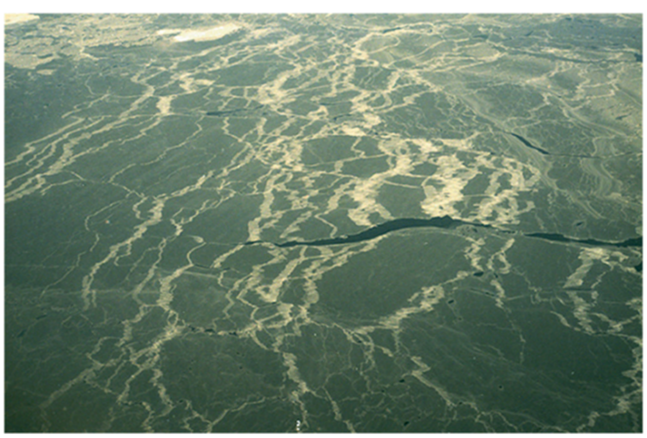

Figure 2.9 Nilas with rafting and cracking in Nares Strait, north of Baffin Bay, in April 1994 (photo by N. K. Sinha, unpublished).

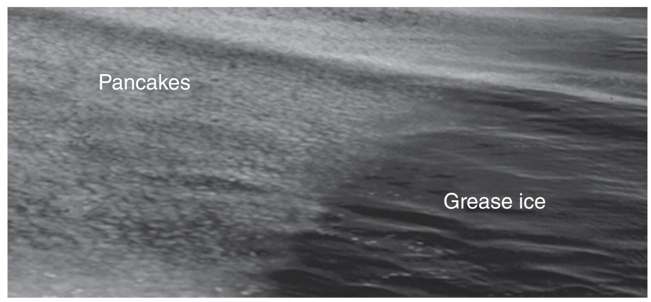

Figure 2.10 Grease ice and small pancakes with diameters in the range of 100–200 mm in Davis Strait in early November 1982; note the differential damping of the ocean waves in the two ice regimes (photo by N. K. Sinha, unpublished).

Figure 2.11 Freely floating and partially consolidated pancakes, with diameters of 0.5–1.0 m, damping the ocean waves in Davis Strait (photo by N. K. Sinha, unpublished).

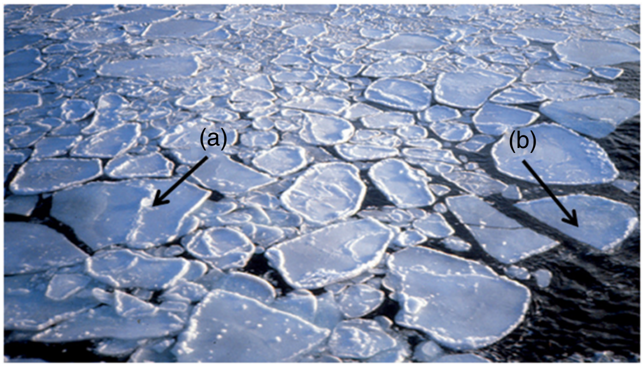

Figure 2.12 Pancakes with diameters up to 5 m in Davis Strait in November 1982; note (a) the intrapancake ridge and (b) the fractured pancake (photo by N. K. Sinha, unpublished).

Fragmented sections of pancakes, such as (b), may also be pushed against each other and get fused by forming intrapancake ridges as pointed out by (a) in Figures 2.12. If conglomeration continues, large thin sheets are formed laterally.

2.2.2. Vertical Ice Growth (Congelation Ice)

Following the formation of nilas either by completion of freezing of agglomerate of needles, snow particles, and small discoids at the surface or by consolidation of pancakes, the ice sheet has no room to grow horizontally. The ice skin at the surface can only grow vertically down along the direction of the maximum heat flow from the underlying water to the atmosphere. This process is known in the ice community as congelation. In metallurgy, this type of unidirectional growth is known as directional solidification (DS). In fact, turbine blades used in the hottest sections of gas turbine engines (e.g., jet engine) are exclusively made of nickel-based superalloys using the DS process [*Sims et al.,* 1987] and were shown to behave mechanically like DS columnar-grained ice [*Sinha,* 2009b].

Nilas and pancake ice are usually composed of frazil or granular ice crystals. Congelation ice starts to grow under this layer and very similar to what is commonly seen in lakes and rivers. The top layers become part of the transition layer to contrast with the congelation layer at the bottom. Structurally, these two types have distinguishable fabric. The differences lie in the solidification process involved. Granular or frazil ice in the transition layer develops after the interparticle liquid freezes as the heat continues to flow from the bottom to the top of the ice sheet. The columnar grains formed during congelation are significantly larger than the grains of frazil or granular ice. They are shaped like a pencil with their length along the vertical direction. The grain shape, orientation, boundaries, subboundaries, and the geometry of the entrapped inclusions (liquid brine and gases) are affected by the history of ice growth. The growth is related to the meteorological conditions (including oceanic) that prevailed at each growth stage.

There are innumerable examples that can be cited to confirm that congelation sea ice in the form of columnar ice is commonly found in relatively narrow channels, fjords, and bays, and in fast ice along the shores of open seas in the Arctic: for example, *Weeks and Lee* [1958] at Hopedale, Labrador; *Weeks and Hamilton* [1962] at Point Barrow, Alaska; *Pounder* [1965] for Button Bay off Hudson Bay, Resolute Bay, etc.; *Peyton,* [1963] at Barrow, Alaska; *Nakawo and Sinha* [1981] in Eclipse Sound, Baffin Island; *Weeks and Ackley* [1982] in Alaskan coast; *Shokr and Sinha* [1994] in Resolute Bay; Nunavut, and *Timco and Frederking* [1996] in Canadian Beaufort Sea. It is also found in undeformed second-year ice [*Bjerklund et al.*, 1985; *Sinha,* 1985*b*] as well as in multiyear ice [*Shokr and Sinha,* 1994; *Sinha,* 1991]. *Gow et al.* [1987] collected cores from ice floes in pack ice in Farm Strait and found that they typically contain 75% congelation versus 25% frazil ice. *Eicken et al.* [1995] confirmed this result in the Eurasian Sea of the Arctic basin and *Meese* [1989] confirmed it in the Beaufort Sea. Naturally, congelation ice is not expected in the pressure ridges or rubble fields containing damaged and crushed material, except within the fragments or blocks of ice inside the ridges or rubbles [*Sinha,* 1991]. *Meese* [1989] found 85% frazil ice in thin ice (0.44 m) in a lead in the Beaufort Sea and between 30% and 70% in areas near a pressure ridge. *Weeks* [2010] mentions that exact percentages of frazil versus congelation ice separated on a regional basis do not appear to be available. In general, in the Arctic, the wind and ocean wave conditions after the initial ice formation usually favor the formation of congealed ice.

There are marked differences in the conditions for growth of sea ice in the Arctic and the Antarctic. Practically all the area of sea ice formation in the Antarctic is a seasonal or temperate, equivalent to sub-Arctic zone. *Treshnikov* [1966] estimated that about 75% of the Antarctic ice melts during the summer, compared to about 25% in the Arctic. Older sea ice survives essentially in Weddel Sea and some areas of the Ross and Bellinshausen Seas. Moreover, the vast perimeter of the annual ice around the continent of Antarctica is actually within the South Temperate Zone (STZ), not really in the south polar region (between 66.6°S and North pole), and can be considered primarily as the marginal ice zone (MIZ). This MIZ in the STZ is subjected to strong winds of the Southern Ocean and the velocity shear across the Antarctic polar front [*Wadhams,* 1986]. The persistent stormy conditions associated with high winds cause the water surface to be almost always turbulent. Naturally, the Antarctic sea ice is expected to feature less congelation ice and more frazil ice. *Gow et al.* [1987] measured the ratio of frazil to congealed ice in the Weddell Sea during the austral summer of 1980 and found that the average values were 57% and 43%, respectively. However, they reported significant variations in these ratios from floe to floe. For example, they found floes containing between 6% and 90% congelation ice and between 3% and 100% frazil ice.

Lange and Eicken [1991] reported that sea ice in floes of all ages in the Weddell Sea in the Antarctic is dominated by granular ice of frazil origin. However, in a field study in the western area of the Ross Sea, *Jefferies and Adolph* [1997] found that thermodynamic thickening of the ice in the inner pack ice was dominated by congelation ice growth. They observed that 65% of the ice in the inner pack (up to 400 km from the coast) was congelation ice with a mean thickness of 200 mm, while 22% of ice in the outer pack (>800 km from the coast) was frazil ice with a mean thickness of 120 mm. They attributed the preponderance of congelation ice in the inner pack ice due to a less stormy environment, which is necessary for significant congelation ice growth. *Worby et al.* [1998] suggested that at least 39% of the ice volume in the East Antarctic is columnar, 47% frazil, 13% snow ice and 1% other types. As part of the Japanese Antarctic Climate Research (ACR) program, a 2-year study of atmosphere/sea ice/ocean interaction processes off Queen Maud Land and Enderby Land, *Kawamura et al.* [1995] performed extensive investigations of sea ice from 1990 to 1992. They examined the structure and texture of sea ice, along with measurements on vertical distribution of oxygen isotope concentration, $\delta(^{18}O)$ (presented earlier in section 2.1.1.), at 16 stations in Lützow-Holm Bay. The ice thickness varied in the range of about 2–3.5 m. Significant variation in the amount of granular and mixed columnar/granular ice was noticed. The thickness of this layer varied a great deal, between 0.5 and 2.5 m, but all the cores exhibited columnar structure at the bottom. For more and up-to-date information on Antarctic sea ice, the reader must consult the National Institute of Polar Research in Tokyo and the Low Temperature Science Laboratory of Hokkaido University in Supporro, Japan.

2.2.3. Superimposed Ice

Superimposed ice is formed on existing ice surfaces as a result of one of the following three processes: freezing of rain on existing ice surface, refreezing of ice surface or snow melt, and freezing of water-logged snow. The second scenario occurs when the ice or snow surface melts due to warm atmospheric temperature during winter, then refreezes when cold atmospheric temperature resumes. The third scenario occurs when the load of snow cover is sufficient to depress the growing ice sheet below its freeboard so that the water floods the surface and later refreezes. In both cases the freezing may proceed either from top down or from bottom up depending on the temperature difference between the ice surface and the atmosphere. From the viewpoint of the crystallographic classification of natural ice, superimposed ice is usually categorized as snow or granular ice. Incorporation of snow into the sea ice reduces the δ (^{18}O). Therefore, it is possible to differentiate between snow ice and other structurally different ice types (of seawater but not snow mixture origin) using a threshold on the δ (^{18}O) value.

Superimposed ice is not common in the Arctic, though it is observed in subarctic areas such as the Labrador Sea, but the ice covers are significantly thinner than those in other areas. However, this type of ice is frequently observed in the Antarctic. In the southern hemisphere, the annual sea ice develops primarily in the South Temperate Zone (STZ), as mentioned earlier, around the coastal areas surrounding the continent of Antarctica, except for the areas covered by Ross Sea, Amundsen Sea, Bellingshausen Sea and Weddel Sea, all within the Western Antarctic (and remarkably just west of the International Date Line and/or longitude of 0°). Since the coastal line of this continent follows essentially the Antarctic Circle with a latitude of about 66.6°S, the sea ice regime is confined within a narrow belt between about 60°S and 66.6°S, except, of course, the western Antarctic. The area, north of the Antarctic Circle but beyond the coastline of the continent in the western Antarctic, is occupied largely not by sea ice but by a collar of freshwater ice in the form of shelf ice. There are a number of huge ice shelves, for example, Ross Ice Shelf, Ronne Ice Shelf, and Thwaites Ice Toungue. Thus, strictly speaking, the Antarctic sea ice regime is not within the south polar region. Since it is found in STZ, this should be named as STZ ice, 'not polar ice'. The STZ ice is naturally expected to be thinner due to the warmer atmospheric temperature of the STZ. Additionally, heavier snowfall in the STZ (primarily drifted from the land to the ocean surfaces) compared to that in the Arctic also dampens the growth of sea ice. No wonder, the Antarctic sea ice is thinner than its counterpart in the Arctic. Here, of course, we mean the annual sea ice as the "undeformed" first-year (FY) ice grown in oceans without the adverse effects of storms and severe wind. This speculation is confirmed by *Weeks* [2010] who concluded that the thickness of undeformed ice in the Antarctic rarely exceeds 1 m, which is remarkably thinner than 1.5–2.5 m typically measured in the Arctic.

Only a few field studies have been conducted in the past to identify the superimposed ice in the Antarctic. *Worby et al.* [1998] observed that more than 50% of thin ice surfaces in the eastern Antarctic were flooded. The surface flooding is caused by a number of processes that include surface deformation and ice breakup, wave penetration in ice, snow loading, or upward rejection of brine from the ice subsurface layer [*Perovich and Richter-Menge,* 1994]. In the western Antarctic, namely in the Ross and Amundsen Seas, *Jefferies et al.* [1994] found that superimposed snow ice varied from 13% to 43%, whereas frazil ice averaged between 25% and 55%. They combined textural analysis with measurements on the stable isotope $\delta(^{18}O)$ to discriminate between snow ice and frazil ice, similar to the 1990 to 1992 studies of *Kawamura et al.* [1995] in Lutzow-Holm Bay.

The application of the differential isotope, $\delta(^{18}O)$, method is certainly unique and very powerful. However, the technique for using such isotopic procedures requires accessibilities with laboratories equipped with rather sophisticated and expensive equipment, and highly trained operators. Moreover, only a limited number of samples can be analyzed because of high expenses involved. An alternate powerful and cheap method that can be performed in field laboratories (e.g., inside tents) is to use the double-microtomed thin sectioning technique in conjunction with measurements utilizing thermal etching, chemical etching, and replicating in addition to polarized light. Replicas can be made in the field and can also be readily examined with optical microscopes (see Chapter 4, specifically section 4.4.1). If desired, the replicas can be examined later with a scanning electron microscope (SEM). Since the microstructures of granular snow ice and frazil ice are significantly different from each other, the two types of ice can be identified readily. A minimum of two sections (i.e., horizontal and vertical) are required for proper identification. These techniques are described in detail in Chapter 6.

2.2.4. Thermodynamic Ice Growth

Ice grows in thickness mostly thermodynamically, although mechanically induced growth also occurs in rough seas. Thermodynamic growth entails increasing thickness in response to the negative energy budget between the ocean and the ice sheet. The ocean is the source of the heat that is transferred to the atmosphere through the ice. Mechanical growth entails piling up of broken ice along the edges of ice floes due to several processes such as rafting, ridging, and rubble pile-up. Larger thickness, of course, is typically generated by dynamic processes. However, as will be seen in section 5.1.4, thermodynamic ice growth can indeed occur up to a depth of 5.1 m of continuous columnar-grained ice, with four distinct interfaces indicating five growth seasons. This was observed by *Sinha* [1986] in a multiyear floe. *Eicken et al.* [1995], inclined to show that thermodynamic ice growth leads to maximum thickness of about 3.5 m for multiyear ice. Thermodynamic growth rate is mainly controlled by three factors: (1) the severity and duration of cold air temperatures, (2) snow accumulation on the surface, and (3) ice thickness. Other factors include proxy for long-wave and latent heat fluxes, ocean heat flux, and short-wave flux in spring and summer.

Extremely rapid freezing can occur in a "lead" if an ice sheet fractures and separates when the ambient air temperature is very low. The Canadian Ice Service (CIS) confirmed (through personal communications) that under a steady air temperature of −25 °C, a thin skin of ice will thicken quickly to reach 100 mm during the first 24 h. This was actually verified by experimentally simulating a lead in Mould Bay, as described in Section 5.1.3. However, as the ice grows, snow is also accumulated on top of the ice cover, and the growth rate decreases with increase in ice thickness. It may take a few weeks to reach the stage of mature first-year ice with thickness greater than about 1.2 m [*Canadian Coast Guard,* 1999]. At warmer temperatures below but closer to the freezing point (in Labrador Sea, Bohai Bay and Ohkhost Sea for examples), it would take 3–4 weeks for ice to grow up to the stage of thin first-year ice (0.3 m thick).

Growth and salinity profile of annual sea ice, in conjunction with microstructural investigations, have been studied extensively for several years in Eclipse Sound near Pond Inlet, Baffin Island, Canada [*Sinha and Nakawo,* 1981, for 1977–1979; *Nakawo and Sinha,*1981, for 1977–1978]. They reported rates between 5 mm/day (i.e., 0.021 cm/h) and 15 mm/day (0.0625 cm/h). Another extensive series of long-term study program, from 1981 to 1985, was carried out in Mould Bay, Prince Patrick Island, Canada. Growth of ice was recorded at several stations across the 7.8 km wide bay. The details of this project will be described in section 5.1. As an example, a growth rate of 20 mm/day was noted during the first 10 days of growth, without any snow cover, in September, 1981. However, for the first 200 days, after the beginning of freezing, during the 1981–1982 winter season, the average growth rate varied from 7.5 mm/day to 9.5 mm/day, depending on the location across the width of the bay. The differences in the growth rate were caused by the differences in the accumulated snow depth.

Melnikov [1995] conducted in situ measurements in the western Weddell Sea in 1992 during the U.S.-Russian Ice Station Weddell 1 Expedition in the Antarctic and found much higher growth rates of 3.8 mm/h for ice up to 90 mm thick (19–20 May), 1.3 mm/h for thicker ice up to 280 mm thick over the next 8 days, and only 0.3 mm/h during 81 days of observations on ice between 0.42 m and 0.97 m thick (18 March to 7 June). The growth rate decreases as ice thickness increases until ice reaches thermodynamic equilibrium with the atmosphere. The ice growth then stops. This equilibrium occurs when the thickness reaches about 3 m in the case of Arctic ice and 1–2 m in the case of Antarctic ice [*Whitman,* 2011].

2.2.4.1. Modeling Ice Growth

Sea ice growth rate and modeling is important from oceanic, remote sensing, and marine navigation viewpoints. Its oceanic impact is demonstrated in the dependence of the salt rejection from the skeletal layer at the bottom of the ice sheet to the underlying saline water layer. As the salinity of the layer and therefore its density increases, it sinks deeper into the ocean, starting a vertical convection current. As a rule of thumb, the faster the rate of sea ice growth the more salt is entrapped within the ice. Considerable salt is usually entrapped in the subsurface layer of thin ice and that affects the radiometric emission and scattering from the surface. As ice grows, the salinity varies with time at any point, but it eventually stabilizes at what is known as "stable salinity." This term is defined and described in details in Section 3.2. It suffices here to mention that *Nakawo and Sinha* [1981] related stable salinity of a section of ice at a given depth with its corresponding growth rate. Finally, ice growth rate is important for marine navigation because it facilitates at least an approximate estimation of ice thickness distributions in areas of limited ice dynamics.

Many studies have been conducted and a few methods have been developed to estimate ice growth rate using direct field measurements and/or simple models with input from meteorological and climatic data. Early studies on ice thickness prediction include *Lebedev* [1938], *Zubov* [1938], *Tabata* [1958], and *Billelo* [1961]. A quick but rough empirical equation to estimate ice thickness (h) in centimeters from thermodynamic growth was presented by *Lebedev* [1938]:

$$h = 1.32(\mathrm{FDD})^{0.58} \tag{2.1}$$

where FDD is the accumulated freezing degree-days (or AFFD used some times), defined as the sum of the average daily subfreezing degrees (below 0°C for freshwater and about −1.8°C for sea water) for a specific period. For example, if the total number of days with air temperature below freezing is 3 and the average daily temperature is −2.8°C, −3.8°C and −5.8°C, then for sea water with freezing point of −1.8°C, the FDD becomes the sum of the daily temperatures after subtracting the freezing temperature; i.e., 1+2+4=7°C. This method is used in the operational sea ice monitoring environment to roughly estimate ice thickness for the early growth period. Equation 2.1, however, does not take into consideration any environmental factors such as ice motion, snow accumulation, ice thickness, surface radiation budget, and surface physical conditions.

Under calm ocean conditions and stable weather, the growth rate of thin ice can be determined using a simple heat flux equation:

$$g = \frac{Q^*}{L\rho} \tag{2.2}$$

where Q^* is heat flux from ocean to the atmosphere, L is the latent heat of fusion for water, and ρ is the density of ice. Under thermal equilibrium Q^* can be determined using the following equation:

$$Q^* = (T_s - T_w)/k \tag{2.3}$$

where k is the thermal conductivity of ice, T_s and T_w are the ice and water surface temperatures, respectively. A more accurate estimate of Q^* can be achieved by estimating its four components:

$$Q^* = Q_E(T_s) + Q_H(T_s) + Q_{LW}(T_s^4) + Q_{SW} \tag{2.4}$$

where Q_E Q_H, Q_{LW}, and Q_{SW} are latent, sensible, long-wave and short-wave fluxes, respectively.

The growth rate of ice can be inferred using a simple model of heat balance at the ice-water interface and energy balance at the upper surface. The latter is more important during the early growth period. Figure 2.13 is an idealized schematic diagram showing a section of floating sea ice of thickness h_i with snow cover of depth h_s. Ice is formed, primarily under two heat transferring influences: (1) the oceanic heat flux F_w from the warm ocean to the colder ice interface and (2) the conductive heat flux F_c through the bulk of the ice and snow, created by the difference between the colder air temperature and the warmer ocean temperature. These two influences are balanced by the quantity of heat released during freezing, which is determined as the multiplication of the latent heat of water for freezing L_i, the ice density ρ_i, and the rate of increase in ice thickness.

The energy balance at the ice-water interface determines the rate of ice growth. Following the convention of positive fluxes in the upward direction, this can be formulated as:

$$F_w - F_c + \rho_i L_i \left(\frac{dh_i}{dt}\right) = 0 \tag{2.5}$$

The ice thickness can be determined by integrating this equation provided that expressions for F_c and F_w (as a function of time) are provided. The continuity of heat flux throughout the ice and snow layers and into the atmosphere (as illustrated in Figure 2.13) leads to the following net conductive heat flux equation:

$$F_c = -(T_a - T_w) \Big/ \left(\frac{1}{k} + \frac{h_i}{k_i} + \frac{h_s}{k_s}\right) \tag{2.6}$$

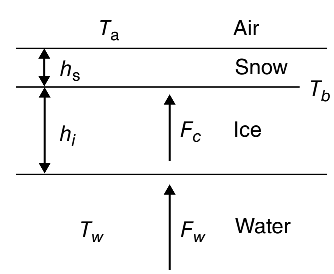

Figure 2.13 Idealized heat fluxes in a snow-covered floating ice sheet.

where T_a and T_w are the air and water temperature, respectively, k is the effective heat transfer coefficient between ice surface and atmosphere, and k_i and k_s are the thermal conductivity of ice and snow, respectively. Substituting this equation into the (one before it) yields

$$\frac{dh_i}{dt}\rho_i L_i = F_w - (T_a - T_w) \Big/ \left(\frac{1}{k} + \frac{h_i}{k_i} + \frac{h_s}{k_s}\right) \quad (2.7)$$

By assuming a constant heat transfer coefficient k, the integration of the above equation produces an expression for the ice thickness:

$$h_i^2 + \frac{2k_i}{k}\left(1 + \frac{k_i h_s}{k_s h_i}\right)^{-1} h_i = \frac{2k_i}{\rho_i L_i}\left(1 + \frac{k_i}{k_s} r_s\right)^{-1} \left[\int (T_w - T_a)\,dt - \int F_w\,dt\right] \quad (2.8)$$

This equation can be used if time series of snow depth, air temperature, and heat flux from the ocean are provided. A simplified form of the above equation can be written under the following assumptions: (1) the ice surface is snow free ($h_s = 0$), (2) the oceanic heat flux F_w is zero, and (3) the atmospheric temperature T_a is stable and equal to the ice surface temperature, therefore $k \to \infty$:

$$h_i^2 = \frac{2k_i (T_w - T_a)}{\rho_i L_i} t \quad (2.9)$$

Equation (2.9) can also be derived for predicting growth in terms of accumulated freezing degree-days, following *Sinha and Nakawo* [1981], using the thermal balance principle, which dictates that heat fluxes through ice and snow layers must be equal, and they are both equal to the heat released during ice freezing. Assuming, once again, that there is no heat flux through the ice-water interface, then the increase of ice thickness Δh_i in time Δt can be written as

$$L_i \rho_i \,\Delta h_i = k_i \left(\frac{T_w - T_b}{h_i}\right)\Delta t = \left(\frac{T_b - T_a}{h_s}\right)\Delta t \quad (2.10)$$

where T_b is the temperature of the snow base or the snow-ice interface (Figure 2.13). The second equality in this equation gives

$$T_b = (k_i h_s T_w + k_s h_i T_a) / (k_s h_i + k_i h_s) \quad (2.11)$$

Substituting T_b in equation (2.10),

$$\Delta h_i = \frac{k_i k_s}{L_i \rho_i}\left(\frac{T_w - T_a}{k_s h_i + k_i h_s}\right)\Delta t \quad (2.12)$$

If $(T_a)_N$ is the mean air temperature of the Nth day from the freezing-up date, then equation (2.12) gives the ice growth on that day:

$$(\Delta h_i)_N = \frac{k_i k_s}{L_i \rho_i}\left[\frac{T_w - (T_a)_N}{k_s (h_i)_{N-1} + k_i (h_s)_N}\right] \quad (2.13)$$

where $(h_s)_N$ is the average snow thickness on the day under consideration (N) and $(h_i)_{N-1}$ is the ice thickness at the end of the previous day.

The total thickness of ice over a number of days D can then be written as the summation of ice growth per day:

$$\sum_{N=1}^{N=D} (\Delta h_i)_N = \frac{k_i k_s}{L_i \rho_i} \sum_{N=1}^{N=D}\left[\frac{T_w - (T_a)_N}{k_s (h_i)_{N-1} + k_i (h_s)_N}\right] \quad (2.14)$$

Equation (2.14) can be rearranged to give

$$\sum_{N=1}^{D}\left[T_w - (T_a)_N\right] = \sum_{N=1}^{D} \frac{k_i k_s}{L_i \rho_i}\left[(k_s (h_i)_{N-1} + k_i (h_s)_N (\Delta h_i)\right]_N \quad (2.15)$$

Growth of ice in Eclipse Sound near Pond Inlet, Baffin Island, Canada, during the two winter seasons 1977–1978 and 1978–1979 was determined from equation (2.15) [*Sinha and Nakawo,* 1981]. Results are shown in Figure 2.14 as a function of accumulated degree-days of freezing. Dates given at the top of the figure indicate the time of the season. Calculations were based on a constant snow thickness of 11.4 cm and appropriate ice and snow conductivities. The equation produces reasonable results in general but underestimates ice thickness during the early part of the season and overestimates it toward the end. The use of measured snow improved the agreement with observations during the early growth period, but not during the late season. This could be attributed to increasing solar radiation.

Equation (2.14) can be rewritten in an integral form to relate the growth of ice in terms of accumulated degree-days of freezing:

$$\int_1^D (T_w - T_a)\,dt = \int_1^D \frac{L_i \rho_i}{k_i k_s}(k_i h_s + k_s h_i)\,dh_i \quad (2.16)$$

For constant snow thickness, the above equation further reduces to

$$\int_1^D (T_w - T_a)\,dt = \frac{L_i \rho_i}{2k_i}(h_i)^2 + \frac{L_i \rho_i h_s}{k_s} h_i \quad (2.17)$$

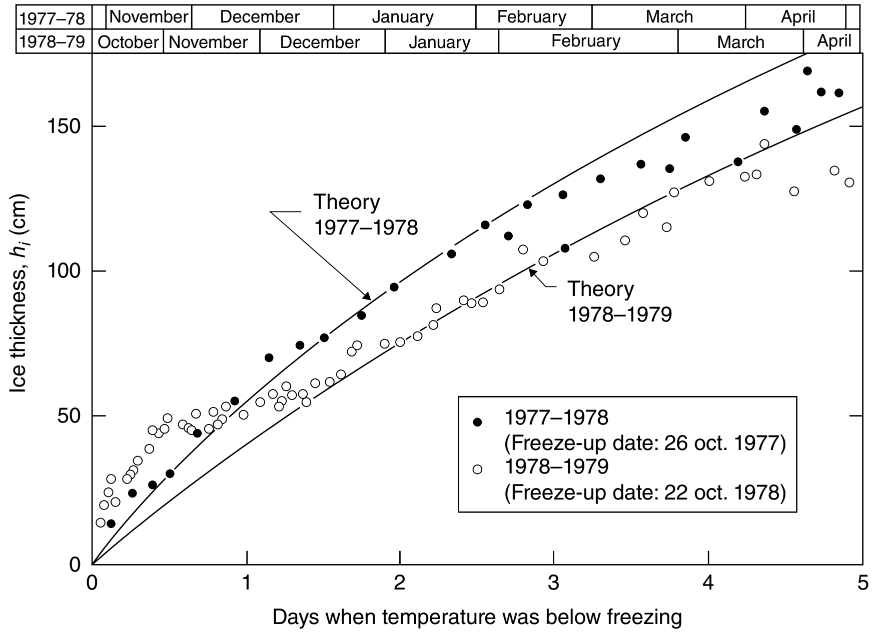

Figure 2.14 Growth of ice in Eclipse Sound measured during winters of 1977–1978 and 1978–1979, compared to calculations. Dates given at the top of the figure indicate the time of the season [*Sinha and Nakawo,* 1981].

In the absence of any snow cover, it can further be reduced to

$$h_i^2 = \frac{2k_i}{L_i \rho_i} \int_1^D (T_w - T_a)\, dt \qquad (2.18)$$

This is the same equation as equation (2.9). The right-hand side represents the accumulated freezing degree-days.

Equation (2.17) was used to calculate ice thickness versus the number of freezing degree-days for different values of snow thickness and ($T_w - T_a$). The results are presented in Figure 2.15. The following values were used in the equation: ρ_i = 917 kg/m^3, L_i = 334.9 × 10^3 J/kg, k_i = 2.0 and k_s = 0.25 J/m · K · s. The point for day 1 in each graph is generated using the thermal conductivity of seawater, 0.6 J/m · K · s, instead of ice. The plots clearly show the blanketing effect of the presence of snow or higher air temperatures. Note the pronounced sensitivity of the temporal growth of ice thickness to the accumulation of the first 100 mm of snow.

The simple model of ice growth described above does not take into consideration the complex effects of (1) uncertainty of thermal conductivity of snow cover, (2) ocean heat flux, and (3) surface ablation. These factors are addressed briefly in the following section.

2.2.4.2. Effect of Snow Cover

Snow cover on sea ice impedes ice growth. It acts as a blanket that affects the thermal diffusivity and optical transmissivity between the underlying ice sheet and the atmosphere. The high albedo of dry snow causes most of the received solar radiation to be reflected, thereby affecting the surface energy balance [*Maykut,* 1986]. More importantly, the poor thermal conductivity of the snow slows down the heat flux from the ice-water interface to the atmosphere through the ice volume. Depending on its wetness, the thermal conductivity of snow varies between 0.1 and 0.4 W/m · K while that of sea ice is roughly 2 W/m · K [*Schwerdtfeger,* 1963; *Ono,* 1967; *Mellor,* 1977; *Massom et al.,* 2001]. In general, the thermal conductivity of snow is one order of magnitude less than that of sea ice. This means that 50 mm of snow deposited on sea ice with thickness of 0.50 m will reduce the ice growth rate by half. The effectiveness of snow as an insulator depends also on its compactness (i.e., age). Newly fallen snow is soft and fluffy, so it is an excellent insulator because of its high air content. The compacted snow, on the other hand, is a relatively poor insulator. Table 2.2 describes various snow conditions in terms of the sea ice thickness that would have the same insulating effect. It is obvious that a few millimetres of even loosely packed snow on top of sea ice will slow down the ice growth rate significantly.

Except for a few field studies, such as the weekly measurements through the entire ice growth seasons of 1977–78 and 1978–1979 in Eclipse Sound [*Sinha and Nakawo,* 1981; *Nakawo and Sinha,* 1981], presented in Chapter 3, and the studies in Mould Bay in 1981–1985 described in Chapter 5 (section 5.1), the physical properties and effect of snow over on sea ice have not received as much attention in the literature as snow over land. The field

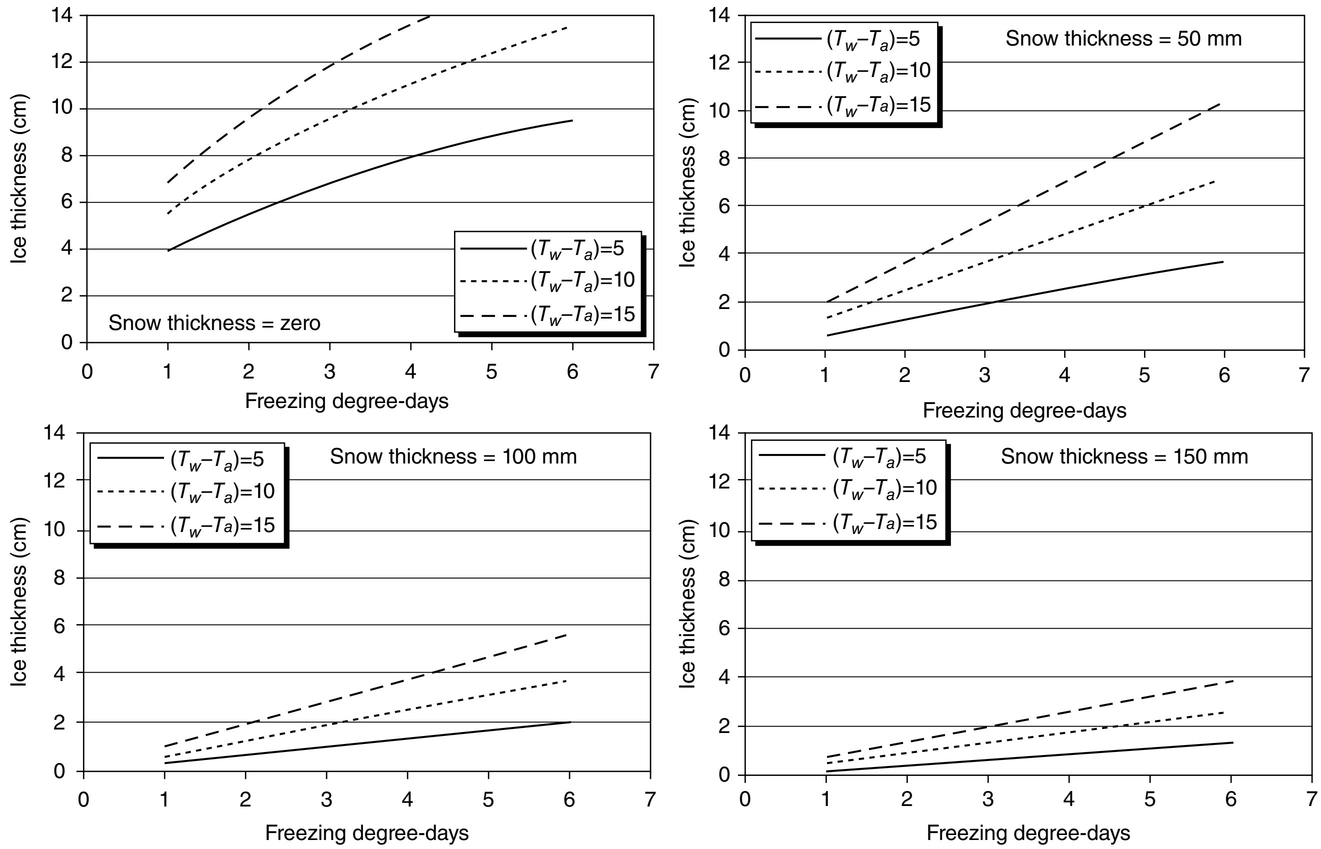

Figure 2.15 Dependence of ice thickness on freezing degree-days for different snow thickness and air temperature, calculated by using equation (2.17).

Table 2.2 Insulating effect of 50 mm of different snow types in terms of their equivalent insulating effect of sea ice thickness.

0.05 m of Various Snow Types	Equivalent to an Ice Thickness of:
Fresh soft snow	2.64–3.80 m
Semisettled snow	1.75–1.93 m
Settled snow cover	0.61–0.97 m
Old snow	0.41–0.61 m
Hard packed snow	0.24–0.31 m
Extremely cold snow	0.19–0.24 m

investigations on the characteristics of snow from the beginning to the end of winter growth season on sea ice at Eclipse Sound were carried out, primarily, for modeling thermodynamic growth of ice and growth rate dependence of brine inclusion and microstructural features. Mould Bay studies were performed to document the effect of snow cover on the temporal as well as spatial variability of ice thickness across the 7.8 km wide bay. Relatively recent studies that address snow depth and distribution on sea ice, with no data on ice thickness and distribution, include *Barber et al.* [1994], *Barber and Nghiem* [1999], *Haas et al.* [2008], *Nicolaus et al.* [2009], and *Iacozza and Barber* [2010]. As a general rule, snow depth increases on ice that has been formed earlier (since more time allows more snow to accumulate) or deformed later (e.g., on rough ice including ridged and ice or rubble fields). Temporal evolution of snow on ice is controlled by the timing of snowfall and wind speed. In the Arctic region, most of the snowfall takes place in the fall and spring [*Sturm et al.,* 2002]. Strong wind of at least 7.7 m/s is required to redistribute the snow across the ice surface [*Li and Pomeroy,* 1997].

Measurements of snow depth over sea ice are usually conducted using a point sampling approach. This approach was adopted in most field studies. A few early studies include the AIDJEX in the Beaufort Sea [*Cox and Weeks,* 1974*; Martin,* 1979], the Eclipse Sound experiments in 1977–1979 [*Sinha and Nakawo*, 1981; *Nakawo and Sinha*, 1981], the Mould Bay experiment [*Holt and Digby,* 1985; see also Section 5.1], and the MIZEX'84 in Fram Strait [*Tucker et al.,* 1987]. It was also used in a series of experiments on Arctic sea ice, known as Sea Ice Monitoring and Modeling Site (SIMMS), conducted between 1990 and 1997 in Resolute Passage, Canadian eastern Arctic [*Barber et al.,* 1992]. Details on temporal and spatial distribution of snow and ice across the 7.2 km wide fjord in Mould Bay, Canadian western Arctic during 1981–1982 are given in Chapter 5.

Snow depth varies significantly between different regions and at different times. Moreover, snow depth has been reported to be greater on multiyear ice (MYI) than firstyear ice (FYI). For example, in the MIZEX'84 experiment snow depth on FYI was found to be 80 mm on average and never exceeded 0.20 m, while it averaged 0.40 m on MYI [*Tucker et al.,* 1987]. A similar observation is made in *Shokr and Barber* [1994] from the measurements during the first experiment in the SIMMS program (between 15 May and 8 June, 1990) when they found that the average snow depth was 20.5 and 36.2 cm on FY ice and MY ice, respectively, with higher standard deviation from MY ice. This is due to the undulating topography of the MY ice surface. *Tucker et al.* [1987] related the thicker snow cover on MY ice to the ice thickness. They suggested that, for thinner ice types such as FY ice, more heat is conducted from the ocean to the ice surface and therefore snow would be more susceptible to loss by sublimation. In the AIDJEX experiment, snow depth on MYI averaged 10 cm on hummocks and 30 cm on depressions (melt ponds). *Warren et al.* [1999] analyzed data from several Canadian weather stations listed in Table 1.1 that operated for 37 years from 1954 to 1991 in the Arctic to explore the temporal evolution of snow on ice. They concluded that the ice remains mostly snow-free during August but snow starts to accumulate rapidly in September and October. It continues to accumulate though moderately in November and very slowly in December and January. It accumulates moderately again from February to May. The Chukchi Sea region shows a steadier accumulation throughout the autumn, winter, and spring. The average snow depth on the MY ice reaches a maximum of 34 cm in May. The average snow density increases with time throughout the snow accumulation season, averaging 300 kg/m^3, with little geographical variation.

Snow cover on Antarctic sea ice is usually deeper than that on ice in the Arctic. *Jefferies et al.* [1994] observed more than a meter of snow in some regions in the Antarctic. In a study to compare differences between snow cover over FY and MY ice in the Weddell Sea, Antarctic, *Nicolaus et al.* [2009] measured snow properties and thickness on level ice. They found that snow on MY ice was thicker, colder, denser, and more layered than on FY ice. Snow metamorphism, however, was similar between the two ice types because it depended more on surface heat fluxes and less on underground properties.

The point sampling measurements of snow thickness can be reliable only for examining spatial distribution of snow over a small area and for short periods. The alternative is to use (1) physical models to estimate the evolution of the snow distribution or (2) remote sensing observations to retrieve snow depth and other properties. Data from using the second approach are presented in section 10.6. A model to estimate the evolution of the snow distribution, called the snow model has been developed and used to simulate evolution of snow over land using input from meteorological and vegetation cover data [*Liston and Elder,* 2006]. *Iacozza and Barber* [2010] applied this model to sea ice and compared results against in situ measurements of snow on fast ice in Tuktoyuktuk, Nortwest Territory (NWT), Canada. They found a significant disagreement between observed distribution and model output, which was attributed to meteorological data being incomplete or inaccurate.

2.2.4.3. Effect of Oceanic Heat Flux

The effect of the upwelling oceanic heat flux (F_w) from the deeper water to the ice-water interface is neglected in equations (2.9) and (2.18). Without this flux, the ice will grow as long as the atmospheric temperature continues to be below the freezing point of the seawater. However, in the presence of this flux, the ice growth may be hindered. This amount of heat is usually small in most regions of the Arctic (around a few W/m^2) as indicated by *Steele and Flato* [2000], so it can indeed be neglected in the heat balance equation. In the Antarctic, where the upwelling convective heat flow from the deep ocean can reach several tens of W/m^2 [*McPhee et al.,* 1998], the oceanic heat flux cannot be neglected. That partly explains the slower ice growth in the Antarctic where ice has been found to be generally thinner compared to the Arctic.

In extreme cases oceanic heat flux may impede or prevent ice formation, even in areas of cold atmospheric temperatures. It also causes melting of the ice bottom during summer or at least slowed down ice growth later in the fall [*Yu et al.,* 2004]. But perhaps the most striking phenomenon caused by oceanic heat flux is what is known as the "sensible heat polynya." Here, the upwelling warm water is so strong that it causes the ice to be significantly thinner or impossible to form because the water surface temperature keeps rising to offset the colder atmospheric temperature. Using a one-dimensional thermodynamic model of sea ice, *Maykut and Untersteiner* [1971] found that ice thickness is sensitive to even small changes in the oceanic heat flux.

2.2.4.4. Effect of Surface Ablation

While ice thickness is expected to increase as it grows at the bottom, the total thickness may, actually, decrease slightly due to surface ablation. This is triggered by sublimation, a process of solid to vapour transformation (directly related to the high-temperature state of ice, being close to the melting point) or surface melt or both. As for sublimation, the air temperature, humidity, and wind velocity are the driving factors. Surface melting, on the other hand, is caused by a rise in atmospheric temperature or the increase in incoming solar radiation. The solar radiation factor can be neglected in the case of polar ice due to the near absence of sunlight during most of the winter season and the low elevation of the sun during the rest of the growth period. Otherwise, surface melt ablation must be taken into consideration. This is determined by surface temperature, which is a strong function of the radiative flux. For a surface at steady temperature, conservation of energy requires that the net radiative flux Q_n must be balanced with the heat fluxes plus the heat consumed by surface melt:

$$Q_n - F_c = F_s + F_e + L_f(dm/dt) \qquad (2.19)$$

Once again, F_c is the conductive heat flux through the bulk of ice and snow. The last term in the right-hand side represents the energy consumed by surface melt, where L_f is the latent heat of fusion (the energy absorbed when a substance melts or released when it freezes), and m is the mass of water formed per unit surface area as a result of melt; F_s and F_e are the sensible and latent heat, respectively, given by the following expressions:

$$F_s = \rho_a C_s C_p U(T_s - T_a) \qquad (2.20)$$

$$F_e = \rho_a C_e L_v U(q_a - fq_0) \qquad (2.21)$$

where ρ_a is the air density, C_s is the sensible heat transfer coefficient, C_e is the evaporation coefficient, C_p is the specific heat at constant pressure, L_v is the latent heat of vaporization (in joules per kilogram), U is the wind speed at a reference height (10 m), T_a and T_s are the air and surface temperatures, q_a and q_0 are the specific humidity at 10 m above the surface and at the surface, respectively, and f is the relative humidity. Both F_c and m depend on T_s. If equations (2.20) and (2.21) are substituted into equation (2.19), then T_s can be determined following a method suggested in *Maykut* [1978]. The net flux in equation (2.19) can be decomposed into its radiation components:

$$Q_n = (1-\alpha)F_r - I_0 + F_L \Downarrow + F_L \Uparrow \qquad (2.22)$$

where α is the ice surface albedo (the ratio of reflected to the incident solar short-wave energy), F_r is the incoming short-wave radiation, I_0 is the short-wave flux penetrating the ice surface, and $F_L \Downarrow$ and $F_L \Uparrow$ are the incoming and emitted long-wave radiation, respectively. In the above equations, the flux direction toward the surface is considered to be positive. *Cox and Weeks* [1988] present expressions for L_v as a function of air temperature, the emitted long-wave radiation in terms of air temperature, the difference in specific humidity $(q_a - q_0)$ (which is a function of both air and surface temperatures), I_0 (which is a function of the reflected short-wave $[(1-\alpha)F_r]$, and α (which is a function of ice thickness). They also present typical values of ρ_a, C_s, C_e, C_p, and f. *Maykut* [1978] gives daily values of air temperature, incoming short-wave and long-wave radiation in the Arctic based on a polynomial smoothing technique developed by *Maykut and Untersteiner* [1971].

Edgar and Ackley [1981] used the above formulation to determine the meteorological variables responsible for the difference between the ice surface ablation seasons in the Arctic and Antarctic. They found that Antarctic ice rarely exhibits surface ablation (i.e., melt pond) as compared to the Arctic ice. They determined the reason for this was the low relative humidity associated with the relatively dry winds off the Antarctic continent as well as the smaller effective radiation parameters in the Antarctic.

2.3. INCLUSIONS IN ICE

Water is an excellent solvent for many substances in the air and the atmosphere and this feature is specially required for the survival and growth of most living organisms. Dissolved salts and gases exist at a much higher quantity in seawater (inorganic salts constitute about 34‰–35‰ by weight) than in lake or river water (around 1‰). However, when water solidifies to ice, the ice lattice retains practically nothing. Except for very minute quantities of certain elements and compounds, all the dissolved materials in the water are rejected on solidification when the amorphous state of water changes to a crystalline structure.

At ordinary atmospheric pressures and temperature (say, higher than −70 °C), crystalline structure of ice is close-packed hexagonal (Figure 4.4 in Chapter 4). The oxygen-oxygen lattice of the hexagonal crystal practically does not allow any atoms other than oxygen (O) and hydrogen (H). There are a few exceptions, but their solubility is extremely low. These are a few common acids, such as HCl and HF, ammonia (NH_3), a few alkalis (KOH, NaOH) and their derivatives such as NH_4F (NH_3 + HF). However, brine and gas inclusions are trapped as pockets in between the ice crystals.

For this reason, sea ice may be considered as a rather complex porous material consisting of pure ice crystals and inclusions that can exist in three different phases depending on its age, thermal state, and history. The inclusions could be liquid in the form of brine pockets, gas in the form of bubbles either isolated (separate) or inside the brine pockets, and solid in the form of precipitated salt crystals. For practical purposes, sea ice can be considered as a two-phase (binary) material comprising of pure ice and one type of dominant inclusion. The dominant inclusion could be brine pockets with or without air bubbles trapped inside them or air pockets. In case of new and up to a year old ice, first-year (FY) ice, the dominant inclusions exist in the form of brine pockets

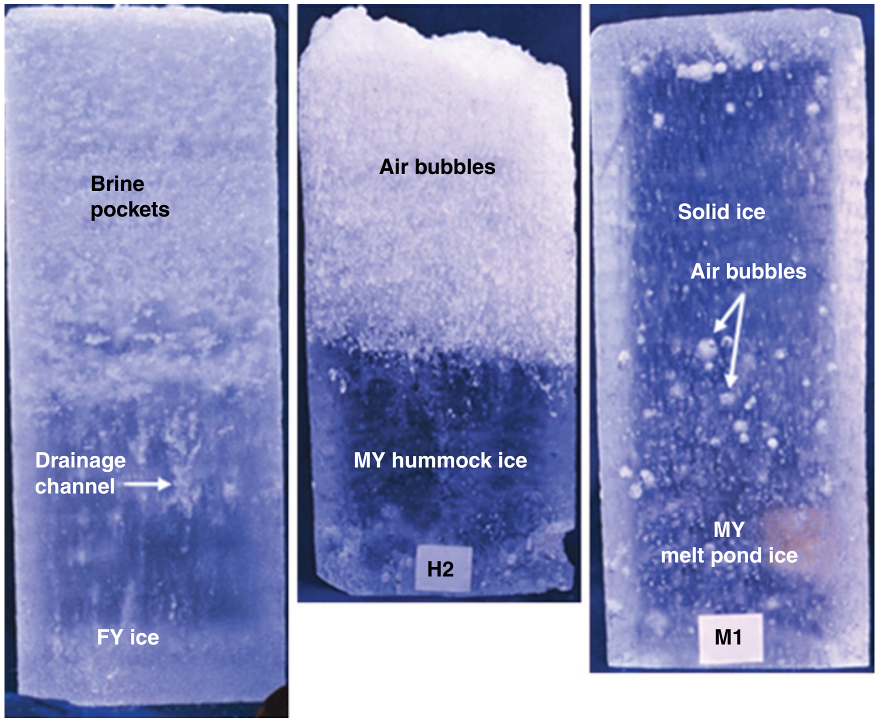

Figure 2.16 Three 100 mm diameter cores: FY ice, MY ice (hummock), and MY ice (melt pond); note brine drainage channels in the FY ice, porous top of layer of the MY hummock, and scattered air bubbles in the melt pond (photographed by M. Shokr).

with or without solid precipitates, for temperatures higher than about −40 °C, and particularly higher than about −23 °C, as can be seen in the phase diagram, Figure 2.1. This binary concept is used in modeling the thermal and electrical properties of sea ice to be described in section 3.6.2. It should be noted here that gas inclusions in the form of very tiny air bubbles are commonly seen inside the brine pockets in FY ice. While brine continues to drain throughout the life time of FY ice, gas inclusions remain trapped in the ice.

First-year and old ice covers can be distinguished visually in the field based on the general topography of the surface. The FY ice floes are usually flat, unless there are ridges, which can be distinguished very clearly. On the other hand, the undulating surface is the main discerning feature of old, MY ice floes. These two types of ice can be identified clearly by sampling ice cores or ice blocks. The main discerning feature is the inclusion type (brine pockets or air bubbles), which can be readily seen. The geometric characteristics of the inclusions can be ascertained readily by external examination of an ice core. In fact, this is the only quick method for discriminating level ice and melt pond ice in case of MY floes.

Figure 2.16 shows the top sections of three 100 mm diameter cores extracted from a flat FY ice floe and from a hummock and a melt pond in an MY floe in Parry Channel near Resolute in May 1993. The cores were photographed after cutting each core vertically into two halves along its center. The photograph of the FY ice shows the intensive brine inclusions that gives the core a "foggy" appearance. It shows also a brine channel with its converging tributaries. The white top of the old hummock ice exhibits the highly porous subsurface layer that actually triggers the high radar backscatter values, which distinguishes MY from other types in radar remote sensing imagery (section 8.1.1). Note the sharp transition between the bubble-rich area at the top of the core and the clear area below. Melt pond ice, on the other hand, features scattered air bubbles that are much less in number and size. These inclusions may originate in superimposed ice discussed earlier in section 2.2.3. The latter probably forms from the freezing of snow-clogged water at the surface of the melt pond. This layer is usually followed by relatively clear ice, which represents either the original MY or newly frozen ice from freshwater accumulated in a melt pond with no snow nucleation. In general, air bubbles in MY ice are much larger than brine pockets in FY ice, with their major axis in the order of a few millimeter compared to a millimeter or less in length of brine pockets [*Bjerklund et al.,* 1985; *Perovich and Gow,* 1996].

The landmass in the Arctic is full of numerous freshwater lakes. Bubbles in the lake ice are formed when the growing ice surface pushes the dissolved air to the growing front. In columnar-grained ice, the air is also pushed to some extent to the longitudinal boundaries between the grains. At a certain stage the air particles are numerous enough to form bubbles, which are generally trapped within the grain boundaries of the columnar grains. They are usually long because the grains are oriented with their long axis in the vertical or growth direction. Air bubbles are readily visible in directionally solidified (DS) columnar-grained freshwater ice as well as translucent glacier ice. Figure 2.17 shows air inclusions in the top 0.28 m of Lake Resolute in May 1993 (before onset of melting). Rounded air bubbles of diameters ranging from fractions of a millimeter to a few millimeters are visible along with elongated bubbles of a few millimeters in length. The shape and distribution of air bubbles in freshwater ice depend on the directional growth of the crystals and their type, namely whether they are S1 or S2 crystal type (see definitions in Table 4.1).

The S1 type of ice covers develop in freshwater bodies under calm conditions without any snowing activities before and during the start of the surface freezing. Following the nucleations at the surface level, the crystals in S1 type of ice grow rapidly in the horizontal plane until there is no room for further growth. The size of the crystals or the grains depends on the number of available sources of nucleation at the surface and air temperature adjacent to the surface of the water. During the processes of the initial surface growth, the dissolved air is segregated and pushed away from the nucleation point or center of the grains toward the boundaries of the crystals. Naturally, the air pockets develop at the boundaries of the crystals, commonly described as "grain boundaries." The size of the air pockets depends on the size of the gains and hence the rapidity of surface growth. Once the surface is covered fully, the crystals have no choice, but to grow in the vertical direction like columns. As the columns grow downward, the dissolved air is continuously pushed toward the long vertically oriented grain boundaries. This growth conditions leads to relatively clear (almost transparent) ice with vertically oriented long air bubbles.

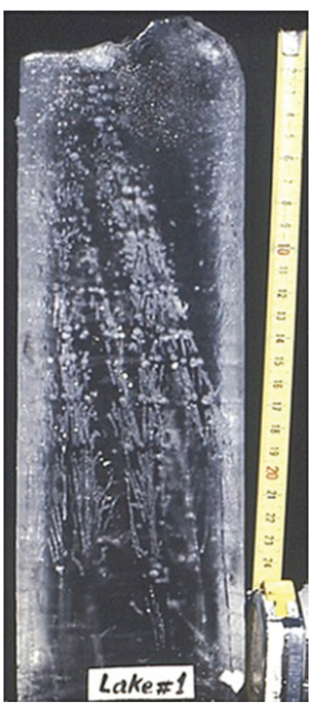

Figure 2.17 Photograph of a 100 mm diameter core of S1 ice, with elongated air bubbles, from Lake Resolute, Nunavut, Canada, obtained in May 1993 (photo by M. Shokr, unpublished).

Ice of glacier origin (see section 2.6.5) is also rich in bubbles. Cracks and crevices are also common phenomena in glaciers. They are often filled with meltwater or rain that eventually solidifies, leaving behind large gaps of air bubbles. The cracks also heal with time due primarily to the high thermal state of ice leading to sublimation and solidification in the narrow spaces inside the cracks. This leads to formation of rows of bubbles, another mechanism that contributes to the bubble-rich medium of glacier ice. Examples of such cracks in iceberg ice in the Labrador Sea off Newfoundland can be seen in *Barrette and Jordaan* [2002]. Healing of different cracks lead to different families of bubbles, and they are usually visible even by the naked eye when exposed surfaces of floating icebergs are examined. Close examinations of ice sampled from glaciers and/or icebergs reveal the details of the families of healed cracks. A method for examining detailed microstructure and texture of glacier ice, using solid-state thin sectioning combined with observations by cross-polarized, parallel-polarized, and scattered light techniques are described in section 4.3.3.7.

Figure 2.18a shows a photograph of a small ice island, with exposed fractured surface, spotted in Wellington Channel near Resolute in the Canadian central Arctic in May 1993. Note the huge freeboard above the surrounding FY sea ice. The layered structure of the shelf ice can also be clearly seen in the exposed fractured surface. This ice island was identified to be a part of one of the larger islands calved from the Ward Hunt Ice Shelf (WHIS) at the northern tip of Ellesmere Island, Canada. The leading edge of WHIS also fractured into several pieces in 1982 and the largest one, named Hobson's Choice, was used as a floating scientific base for geophysical research for many years, as will be seen in section 5.2. WHIS is still the largest ice shelf in the Northern Hemisphere. More than 50 years ago, it used to be a strip about 250 km long