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Introduction and background

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1.1 Aims and objectives of the book

The cryosphere can loosely be defined as all frozen water and soil on the surface of the Earth. This definition encompasses a diverse range of ice masses with a vast spectrum of spatial and temporal characteristics. It ranges from ephemeral river and lake ice to the quasi-permanent (on a millennial timescale) ice sheets of Antarctica and Greenland. Included in the definition is seasonal snow cover and permafrost. In compiling this book it was neither possible nor desirable to include all these different components. This is because the processes and interactions at play are as diverse as the components and, in some cases, unrelated. We have focussed here on two key components, which interact with each other and with the rest of the climate system in an inter-related way. They are land ice, in the form of ice sheets, caps and glaciers, and sea ice. Combined, these represent, at any one time, by far the largest component of ice on the planet, both by volume and area, yet respond to climate change over timescales ranging from seasons to millennia. Sea ice has been identified by the Intergovernmental Panel for Climate Change (IPCC) as a key indicator of short-term climate change, while land-based ice masses may have contributed as much as 50% of attributable sea-level rise during the twentieth century, and represent a large uncertainty in our predictions of a future rise (Houghton *et al.*, 2001). In the rest of this book the word cryosphere refers to these two components only.

The goal of this book is to provide, in a single volume, a comprehensive, up to date and timely review of our state of knowledge about the present-day mass balance of the cryosphere from observations and how it might change over the next millennium based on the latest modelling studies. The book is designed as a reference text covering all aspects of both the theory and practice of measuring and modelling the mass balance of land and sea ice. There are several excellent texts that cover the general physical principles of glaciology but none that deal, specifically, with the determination of the mass balance of either land or sea ice. In this respect, therefore, this is a unique contribution. Parts I and II cover the theoretical principles underpinning the methods used to observe and model mass balance, respectively. The subsequent parts present the state of knowledge of the present-day and predicted future

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mass balance of the cryosphere. These chapters are detailed reviews authored by leading scientists in their field. In all, 23 authors have contributed, but each chapter represents an integral part of the whole book rather than being a stand-alone contribution. This is not, therefore, a set of separate, unconnected contributions but an integrated, coherent treatise presenting (i) background material on the subject, (ii) our best guess as to the present and future state of health of land and sea ice, and (iii) how this information has been derived from both measurement and modelling.

In the 1990s a number of major advances have taken place in (i) our ability to monitor and estimate mass balance, and (ii) the sophistication and accuracy of numerical models of the various components of the cryosphere and more general Earth system models incorporating the cryosphere. Satellite, airborne and terrestrial programmes supported by the European Space Agency (ESA) and the National Aeronautics and Space Administration (NASA)[1] during the 1990s have, in particular, resulted in a quantum improvement in our knowledge of the mass balance of land and sea ice. New satellite programmes, dedicated to studies of the cryosphere, have been announced by both agencies, reflecting the recognition by governments and non-governmental organizations alike of the key role that the cryosphere plays in the Earth system and its vulnerability to changing climate.

1.2 Importance of the cryosphere in the Earth system

1.2.1 Sea level

The mass balance of land ice has a direct impact on sea level and is most likely contributing to sea-level rise, although the error bars on estimates of this contribution are as large as the signal (Church *et al.*, 2001). The Antarctic and Greenland ice sheets contain enough ice to raise global sea level by around 65 m and 6 m, respectively. Even a relatively small imbalance in these ice masses will have a significant effect on sea-level rise, which is currently believed to be between 1.5 and 2 mm per year. The uncertainty in the mass balance of Antarctica represents 1 mm, i.e. one-half, of the total signal (cf. Chapter 12). The level of uncertainty associated with the Antarctic ice sheet's future behaviour is such that we are not even certain of its sign (Houghton *et al.*, 2001). A better understanding of the dynamics of both ice sheets is therefore crucial to our ability to reduce the uncertainty in our predictions of future sea level. Smaller ice masses are, without question, presently contributing to sea-level rise at an ever increasing rate (currently at 0.41 mm per year, cf. Chapters 15 and 16) and represent one of the more sensitive elements of the global climate system, for reasons explained below.

1.2.2 Ice–ocean–atmosphere feedbacks

Land-based ice interacts with the global climate system in a number ways and over a range of different timescales. The principal interaction (shared with sea ice) is due to very high

[1] The crucial contributions by ESA and NASA to the study of the cryosphere is reflected in their joint sponsorship of this book.

reflectance characteristics. The albedo of clean snow ranges from 80 to 97%, while that of clean ice varies between 34 and 51% (Paterson, 1994). This is many times greater than the albedos of other naturally occurring surfaces, such as water (∼1%), forest (10 to 25%) and bare soil (5 to 20%). This contrast gives rise to many important feedbacks between the cryosphere, the atmosphere and the underlying ocean or land surface, which can locally exacerbate the effects of global climate change on very short timescales. Increasing snow cover reduces the amount of solar radiation absorbed, producing a cooling effect that leads to increased snow cover. The converse is also true: increased temperatures reduce snow cover, which leads to more solar radiation being absorbed and a further increase in temperature. This is usually called the ice-albedo feedback mechanism and is one of the most important interactions that the cryosphere has with the rest of the Earth system. It is a particularly important factor for sea ice as this is such a dynamic component of the cryosphere, fluctuating in extent by about a factor of 5 between summer and winter in the Southern Ocean (Chapter 8). The albedo of sea-water is about 1% and the effect of sea-ice cover is, therefore, to reduce the amount of solar radiation absorbed at the surface by as much as 95%, which could amount to around 100 W/m^2. By comparison, the estimated radiative forcing effect of the increase in atmospheric CO_2 from 1750 to 2000 is 1.46 W/m^2 (Houghton *et al.*, 2001).

In addition to the albedo effect, land-based ice masses often have a dramatic effect on regional climates. At synoptic scales, the topographic blocking of the large ice sheets has an important influence on atmospheric circulation. At intermediate scales, ice masses often have their own distinctive regional climates, for instance ice masses are normally associated with strong katabatic winds. At local scales, the presence of a glacier in a valley will alter the micro-climate dramatically. In fact, valley glaciers are extremely sensitive indicators of climate change because they cannot buffer the effects of an atmospheric warming by increasing long-wave emission once their surfaces have reached melting point.

It should also be noted that, in addition to being active components of the climate system, ice sheets and glaciers are efficient recorders of past climate change. Information obtained from ice cores has played an important role in characterizing past changes on timescales varying from decades to ice ages (for example, the Vostok ice core from East Antarctica extends back 420 000 years; Petit *et al.*, 1999). Sub-polar and alpine glaciers are often an important source of water for both hydroelectric schemes (in Norway and Iceland, for example) and irrigation/human consumption (such as in the Himalayas and Karakoram). In many areas they are a key resource for the biggest growth industry on the planet: tourism.

Both the ice sheets and sea ice play an important role in ocean circulation, and, in particular, in the formation of deep water that forms part of the ocean conveyor belt (Stossel, Yang and Kim, 2002). Rapid and dramatic changes in climate have been identified from ice-core data during the last glacial in the northern hemisphere (Stocker, 2000). These fluctuations (in particular Heinrich events and Dansgaard–Oeschger oscillations) involve major changes in a few decades, and the former are probably associated with the shutdown of the thermohaline circulation by massive iceberg discharge events from the Laurentide ice sheet (which covered North America). Ice is clearly implicated in rapid climate change during glacial times, and it has been suggested that the Greenland ice sheet could, under conditions of global warming, influence the thermohaline circulation of the North Atlantic

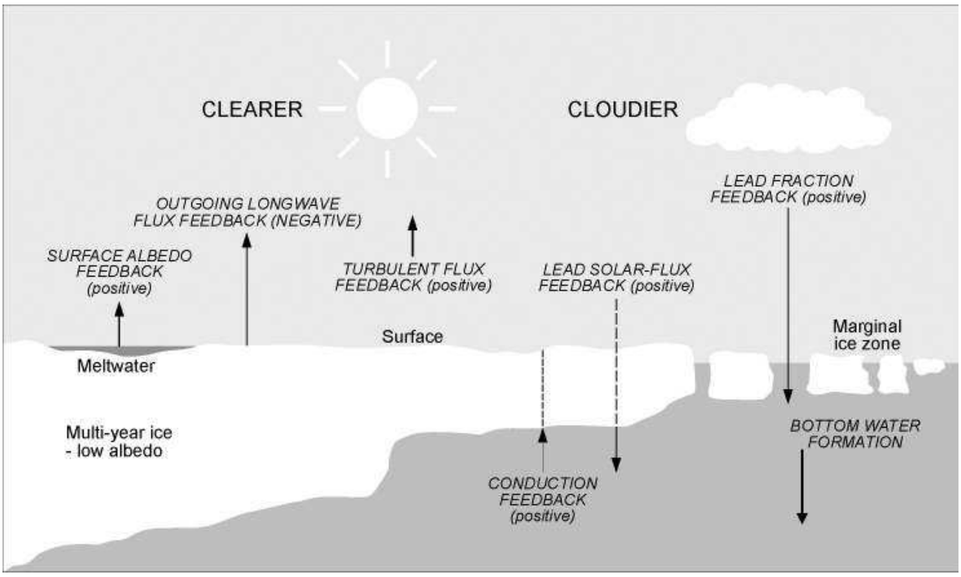

Figure 1.1. Schematic diagram illustrating the key interactions of sea ice with the rest of the climate system. (Courtesy E. Hanna, University of Plymouth.)

in the future. During the present-day, iceberg discharge and bottom melting from floating ice shelves fringing most of the Antarctic ice sheet contribute to the fresh-water budget and mixing in the Southern Ocean.

Sea ice is the most dynamic and variable component of the cryosphere examined in this book and, partly as a consequence, it is strongly coupled to the rest of the climate system. Figure 1.1 is a schematic diagram illustrating some of the key feedbacks that take place, in particular close to the marginal ice zone. Sea ice acts as a thermal blanket, as well as greatly reducing the exchange of moisture and CO_2 between the ocean and the atmosphere. Leads within the pack ice are areas of intense energy and moisture exchange, due to the large contrast between surface oceanic and atmospheric temperatures. Intense cooling of the surface waters take place in leads and polynyas, producing dense, cold water that forms North Atlantic deep water in the northern hemisphere and Antarctic bottom water (ABW) in the south. The Weddell Sea Polynya, for example, is believed to play a key role in ABW formation (Goosse and Fichefet, 2001). A statistically significant reduction in Arctic sea-ice extent and thickness has been observed since about 1970 (see Chapter 8), and this could, if it continues, result in substantial, but largely uncharted, changes to the climate of the North Atlantic through the interactions and feedbacks described above.

1.3 Timescales of variability

This book discusses the mass balance of components of the cryosphere ranging in size from the Antarctic ice sheet, covering an area of 1.3×10^7 km^2, to valley glaciers a few kilometres in length. We also discuss trends in sea-ice cover, which, as mentioned, varies markedly on a seasonal basis. It is important, therefore, to discuss briefly the issue of timescales of variability and to place current fluctuations in context with respect to past variations.

The present-day ice sheets of Greenland and Antarctica cover a total area of approximately 1.36×10^7 km^2. This is about half of the land area covered by ice sheets during their maximum extent some 21 000 years ago (at the last glacial maximum, LGM). Both present-day ice sheets were significantly larger and extended to the edge of the continental shelf. In addition, ice sheets covered much of the northern hemisphere continents north of 45° (the Laurentide, Innuitian and Cordilleran ice sheets in North America, and the Scandinavian and Barents Sea ice sheets in Eurasia). A concomitant increase in sea-ice extent also existed during this period. Any assessment of the present-day cryosphere should be made, therefore, in the context of the global deglaciation that took place at the end of the last Ice Age. This deglaciation started at approximately 14 500 years before present (BP) and was largely completed by 11 500 years BP, although it should be noted that these dates vary significantly on a regional basis (for instance, the Cordilleran ice sheet did not reach its maximum until 4000 years after the LGM; Clague and James, 2002). The transition from the glacial world to the present-day, inter-glacial one was not a gradual process. In particular, it was punctuated by a major re-advance of ice masses during a cool period known as the Younger Dryas (12 000 years BP). More recently, during the Holocene epoch (the last 10 000 years), the world has seen several periods during which global temperatures have changed dramatically and ice masses have temporarily re-advanced or retreated. Examples of the former are the climatic optimum or hypsothermal between 5000 and 6000 years BP (when global temperatures were roughly 1 °C warmer than present) and a secondary optimum at 1000 AD, which saw Viking expeditions to Iceland, Labrador and Greenland, and the establishment of farming in Greenland. A period of relative ice advance separated these two optima, and the second one was terminated by the Little Ice Age, centred around 1700 AD and lasting for approximately 400 years. Our assessment of the present-day mass balance of the cryosphere must therefore been seen against a background of natural variability on a range of timescales varying from the major glacial–inter-glacial cycles to sporadic, centennial events such as the Little Ice Age or even shorter climate fluctuations such as the quasi-decadal North Atlantic oscillation.

Antarctica has response times of the order of 10 000 years or more. It is, therefore, amongst the slowest components of the climate system (along with the deep ocean, see Table 1.1). This timescale is principally associated with the very slow flow of ice within an ice sheet, where velocities are typically of the order of 1 to 10 m per year away from the faster flowing zones such as ice streams (where velocities are of the order of 1 km per year but which only occupy a small fraction of the present-day ice sheets). This has the important implication that Antarctica (and possibly Greenland) is still responding to climate changes associated with the Earth's emergence from the last Ice Age some 8000 years ago (see Chapter 13 for more details). Since the ice sheet was significantly larger at the LGM, this implies that some of the observed recent sea-level rise may be attributable to this long timescale, non-anthropogenic cause. Furthermore, a component of the present-day mass balance of the ice sheets may not be related to climate change during the Holocene.

Table 1.1. *Estimated timescales from various components of the climate system to reach equilibrium (i.e. their response times).*

Component	Response time
Free atmosphere	days
Atmospheric boundary layer	hours
Oceanic mixed layer	months to years
Deep ocean	centuries
Sea ice	days to centuries
Snow and surface ice	hours
Lakes and rivers	days
Soil and vegetation	days to centuries
Glaciers	decades to centuries
Ice sheets	millennia
Mantle's isostatic response	millennia

Source: McGuffie and Henderson-Sellers (1997).

For land ice, the dynamic response time, t, is proportional to the size of the ice mass (Paterson, 1994) and can be approximated by the relationship

$$t \approx H/a_0, \tag{1.1}$$

where H is the maximum ice thickness and a_0 is the ablation rate. If typical values for Greenland are used in equation (1.1), we find that it has a present-day response time of about 3000 years, while for a valley glacier t can be of the order of few hundred years. Land-ice masses are integrators of the climate. In contrast, sea-ice extent is directly related to the immediate climate and, as a consequence, may be one of the earliest indicators of recent, and possibly anthropogenic, climate change. In the chapters that follow, discussions of mass balance are implicitly linked to the time constants for a response, and it is important, therefore, that the reader keeps this in mind and is aware of the implications for the interpretation of a short-term record from satellite or *in situ* observations. Modelling studies of variability also implicitly incorporate the timescale for various types of response. The emphasis in this book is on variability in mass balance from decadal to millennial timescales. We do not deal with glacial–inter-glacial variations, except where they may be influencing the present-day state, as discussed above.

1.4 Geographical context

The largest ice mass on the planet (by a factor of 10) is the Antarctic ice sheet (see Chapter 12). It contains around 80% of the world's fresh-water supply and covers an area in excess of 1.3×10^7 km^2. As mentioned, the dynamic response time of an ice mass is proportional to its size, and as a consequence it has, in general, the longest response time of any of the

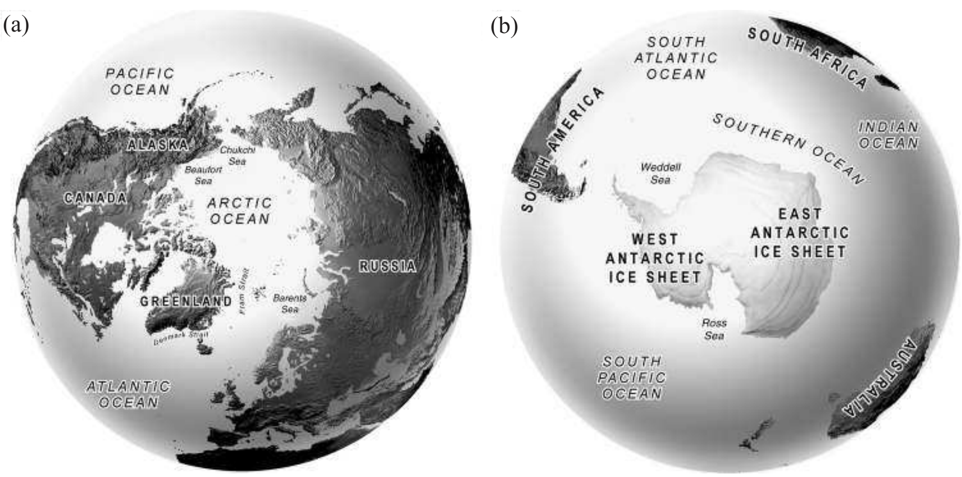

Figure 1.2. Polar stereographic map projections of (a) the Arctic Ocean and surrounding continents and (b) Antarctica and the Southern Ocean.

ice masses discussed in this book. In contrast, Antarctic sea ice covers as much as 1.9×10^7 km^2 of the Southern Ocean during the austral winter, but has a strong seasonal cycle of growth and decay with a minimum extent of less than 4×10^6 km^2. Geographically, the Arctic has a very different setting (Figure 1.2).

The Arctic Ocean is almost completely land-locked with relatively narrow channels such as the Bering and Fram Straits providing for ice or water transport into or out of the area. By contrast, the Southern Ocean surrounds a continent and is responsible for the largest current on the planet: the Antarctic Circumpolar Current. These differences are fundamental to the climate of the regions and the behaviour of both the land and sea ice. The continental character of East Antarctica results in it being one of the driest deserts on the planet with annual precipitation as low as 7 cm per year in parts of the interior. Precipitation over Greenland is around 20–100 cm per year. The Greenland ice sheet is around a factor of 10 smaller in area and volume compared with its southern hemisphere counterpart, but it has an extensive ablation area (unlike its southern hemisphere counterpart) which is extremely sensitive to changing climate. The other islands and continental areas surrounding the Arctic Ocean are also glaciated to a larger or greater extent. Almost 60% of the Svalbard archipelago is, for example, glaciated. These smaller ice masses of the Arctic represent about half the total global volume of land ice not contained in the ice sheets. The other half is distributed amongst glaciers, mainly throughout the northern hemisphere, including the European Alps, the Himalayas, Karakoram and other areas in central Asia. There is relatively little land ice in the southern hemisphere outside of Antarctica. The exception to this is the Patagonian ice field that stretches for some 350 km along the spine of South America from a latitude of around 48–51.5°S and is the largest ice cap outside of the polar regions. There are in excess of 160 000 glaciers on the planet, ranging in length from kilometres to hundreds of kilometres. The cryosphere is clearly extremely heterogeneous in its distribution, size, response time and interaction with the rest of the planet. To observe and model such a heterogeneous constituent of the Earth system presents a daunting prospect, and in the following chapters we present the 'state of the art' and our best attempt at tackling this challenge.

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