State of the Earth's Cryosphere at the Beginning of the 21st Century: Glaciers, Global Snow Cover, Floating Ice, and Permafrost and Periglacial Environments—

GLACIERS

By RICHARD S. WILLIAMS, JR., *and* JANE G. FERRIGNO

With sections on GLACIERS OF THE SUBANTARCTIC ISLANDS
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ICE CORES, HIGH-MOUNTAIN GLACIERS, AND CLIMATE
By LONNIE G. THOMPSON

GLACIER MASS CHANGES AND THEIR EFFECT ON THE EARTH SYSTEM
By MARK B. DYURGEROV *and* MARK F. MEIER

GLOBAL LAND ICE MEASUREMENTS FROM SPACE (GLIMS)
By BRUCE H. RAUP *and* JEFFREY S. KARGEL

SATELLITE IMAGE ATLAS OF GLACIERS OF THE WORLD

Edited by RICHARD S. WILLIAMS, JR., *and* JANE G. FERRIGNO

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At the beginning of the 21st century, analyses of satellite images and other satellite data and ground observations showed that virtually all of the Earth's glaciers are retreating and losing mass. Ice cores from high-mountain glaciers provide more comprehensive information on climate change during shorter time intervals (centuries to millennia) than ice cores from the Antarctic and Greenland ice sheets (tens to hundreds of millennia). Change in mass balance of glaciers has many effects on the Earth System, especially the contribution of glacier meltwater to the rise in sea level. The Global Land Ice Measurements from Space (GLIMS) initiative focuses on scientific studies of changes in glaciers worldwide by using satellite images. The GLIMS initiative overlaps and continues the international research that this 11-chapter Satellite Image Atlas of Glaciers of the World project began in the late 1970s.

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STATE OF THE EARTH'S CRYOSPHERE AT THE BEGINNING OF THE 21ST CENTURY: GLACIERS, GLOBAL SNOW COVER, FLOATING ICE, AND PERMAFROST AND PERIGLACIAL ENVIRONMENTS

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Abstract

Part A-2, Glaciers, synthesizes information on glaciers in the 10 chapters (B–K) on geographic regions ("Satellite Image Atlas of Glaciers of the World," USGS Professional Paper 1386, 1988–2011). Part A-2 also includes additional information through 2009 about the state of glaciers in each of the 10 glacierized regions. Analyses of remotely sensed images from the Landsat multispectral scanner (MSS) and return beam vidicon (RBV) sensors acquired during the "baseline period" (1972–1981) provided the primary information to the authors of the 10 chapters. More recently published chapters incorporate satellite images from sensors on non-Landsat satellites, such as Advanced Spaceborne Thermal Emission and reflection Radiometer (ASTER) images in Chapter K ("Glaciers of Alaska," 1386–K). The baseline period was, in retrospect, a cooler interval of atmospheric temperature during the 20th century, when many glaciers worldwide advanced. A warmer interval of atmospheric temperature since the mid-1990s, which continues in 2012, has resulted in loss of glacier mass. Virtually all of the Earth's glaciers, from the smallest mountain glacier to the two large ice sheets in Greenland and Antarctica, are retreating and losing mass (except for surge-type glaciers and some tidewater glaciers). Eustatic (global) sea level is rising at the rate of about 3 mm a^{-1}, or perhaps by as much as 4 mm a^{-1}. About 50 percent of the rise has been caused by glacier meltwater and 50 percent by volumetric (steric) increase in the warming oceans. The eustatic rise in sea level is the principal glaciological hazard because of its high potential for displacing large human populations that currently occupy deltaic regions and because it will exact a high economic cost from nations that must protect coastlines against inundation and (or) must relocate and rebuild fixed infrastructures. An estimated 160,000 mountain glaciers were the primary contributors of meltwater into the oceans during the 20th century, especially shrinkage of Alaskan glaciers. During the 21st century, meltwater from receding mountain glaciers will continue. Two key unanswered questions are: (1) How will the only two remaining ice sheets on Earth—in Greenland and Antarctica—respond to continued warming of the atmosphere and oceans and changes in precipitation? (2) Will glacier meltwater from the Greenland ice sheet and (or) the Antarctic ice sheet become the primary driver of the eustatic rise in sea level as the 21st century proceeds?

A comprehensive review is presented of the state of the two largest glaciers on the Earth, the Greenland and Antarctic ice sheets, and the estimated 160,000 mountain glaciers and 70 largest ice caps (collectively referred to in this chapter as non-ice-sheet glaciers). Four independently authored sections are included in Part A-2: Glaciers of the Subantarctic Islands, which were not included in Chapter 1386–B, "Antarctica"; Ice Cores, High-Mountain Glaciers, and Climate, a review of the comprehensive history of climate variability during the past 10 millennia that high-mountain glaciers in the tropical and temperate latitudes north and south of the Equator preserved in their ice, which scientists extract as ice cores; Glacier Mass Changes and Their Effect on the Earth System, which addresses the source of glacier meltwater and delineates the scope of the loss of mass in glaciers since the 1980s; and Global Land Ice Measurements from Space (GLIMS), which combines analyses of satellite images and of non-image data from a variety of sensors orbiting the Earth with Geographic Information Systems (GIS) technologies in order to compile a comprehensive inventory of all the Earth's non-ice-sheet glaciers), a 21st century successor to the 20th century Satellite Image Atlas of Glaciers of the World.

Introduction

The concept of an Earth System (see pl. 1 and Supplemental Cryosphere Note 1), which is presented on p. A21–A34 in part A-1 of this chapter, serves as the context within which the cryosphere functions and within which this 11-chapter study of the Earth's cryosphere is best understood. The two primary components of the Earth System are the geosphere and the biosphere. The four subcomponents of the geosphere are: lithosphere (solid Earth), atmosphere (gaseous envelope), hydrosphere (liquid water), and cryosphere (frozen water).

Of the four elements that make up the Earth's cryosphere (fig. 1) (glaciers, snow cover, floating ice, and permafrost), glaciers are the most widely distributed geographically, although most of the area and volume of glacier ice are at high latitudes in the polar regions—in Greenland and Antarctica—in the two remaining ice sheets and in an estimated 70 large ice caps (tables 1, 2; Meier and Bahr, 1996). Glaciers are currently (2012) present on all continents except Australia and on many oceanic islands in both polar regions (fig. 2*A*). Meier and Bahr (1996) and Meier (1998a), basing their estimate on scaling analysis, determined that there are 70 large ice caps[1] and approximately 160,000 other non-ice-sheet glaciers on Earth. Glaciers range in area from 0.1 km^2 (mountain glaciers, such as fig. 2*B*) to the only two continent-covering ice sheets that remain on Earth (Meier, 1974), the Antarctic ice sheet, with an area of 13,586,400 km^2, and the Greenland ice sheet, with an area of 1,736,095 km^2 (table 1).

C. Simon L. Ommanney (written commun., 2009) suggests that atmospheric ice be included as a separate element in the Earth's cryosphere, but the four elements shown in figure 1 are the most relevant to any discussion of climate change on Earth. Forms of atmospheric ice that are deposited on the Earth's surface are implicitly included in the snow cover element. Snowflakes, sleet, hail, rime ice, and other forms of ice crystals are examples of ice that form in the atmosphere. Ice also enters the Earth System as cometary debris striking the Earth's atmosphere, but it does not remain in frozen state for long. The Visible Imaging System (VIS) NASA's Polar spacecraft (National Aeronautics and Space Administration [NASA], 1997) provides evidence that thousands of bodies of cometary ice bombard the Earth's upper atmosphere daily, introducing large amounts of water vapor into the upper atmosphere. C. Simon L. Ommanney (2009, written commun.) also suggests including "ground ice and subterranean ice (glaciares)" as elements of the Earth's cryosphere. However, according to the most recent (5th) edition of the Glossary of Geology (Neuendorf and others, 2005), "ground ice" is synonymous with "permafrost" (p. 285) and "subterranean ice" is synonymous with "ground ice" (p. 603). Therefore, both terms are included in "permafrost".

[1] Although Meier and Bahr (1996) refer to 70 ice caps worldwide, they are clearly referring only to the larger ones. Frank Paul (written commun., 2010) notes that there are an estimated 1,000+ ice caps on Earth. In Iceland, for example, the largest ice cap, Vatnajökull, has an area of 8,086 km^2; the smallest ice cap is 5 km^2 (Sigurðsson and others, in press).

Figure 1.—*Elements of the Earth's Cryosphere. Graphics design by James Tomberlin, U.S. Geological Survey.*

A

B

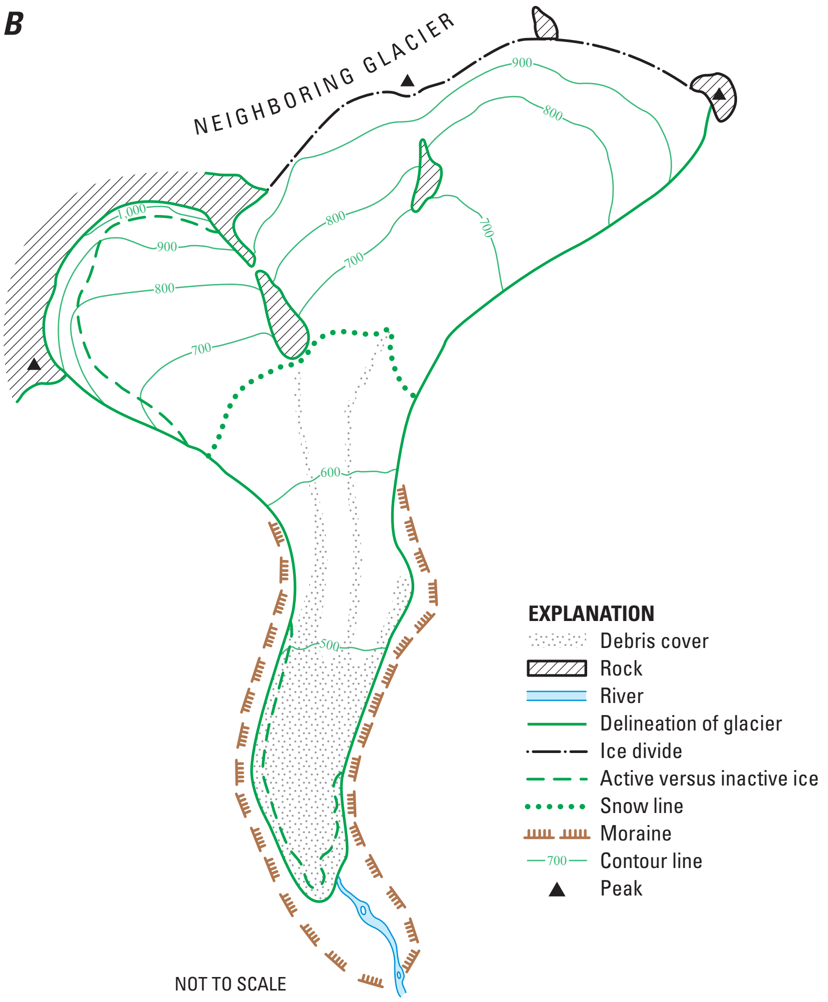


Figure 2.—**A**, *Geographic distribution of the principal glacierized regions (red) on Earth. Modified from Canadian Geographic (Shilts and others, 1998, figure on p. 52).* **B**, *Schematic diagram of a glacierized area, delineating features of a glacier and its landscape. Modified from Müller and others (1977, p. 4, fig. 1).*

Table 1.—*Areal extent of present-day glaciers on Earth*

[Area is listed in square kilometers (km²); ~, estimate; —, no information given]

Glacierized region	Source of data			
	Flint (1971); Sugden and John (1976)	Ohmura (2009)	U.S. Geological Survey Professional Paper 1386	Chapter or other source
	Area, estimated[1]	Area[2]	Area[1]	
South Polar region	**12,588,000**	[**12,307,000**]	**13,604,769**	—
Antarctic ice sheet[3]	—	12,300,000	13,586,400	Drewry and others (1982), table 2 in Part A–2 of 1386–A
Subantarctic islands	3,000	7,000	15,419	Table 8 in Part A–2 of 1386–A
South Shetland Islands (included in total of subantarctic islands)	—	—	(2,950)	Table 8 in Part A–2 of 1386–A
North Polar region	**2,137,274**	[**1,944,630**]	**2,040,397**	—
Greenland ice sheet	1,802,600	1,699,000	1,736,095	C; table 2 in Part A–2 of 1386–A
Ice caps and mountain glaciers peripheral to the Greenland ice sheet		49,000	48,599	C; table 2 in Part A–2 of 1386–A
Canadian Arctic islands	153,169	151,760	151,057	J
Iceland	12,173	**11,200**	11,048	D
Svalbard	58,016	**33,670**	36,591	E
Jan Mayen	117	—	113	E
Russian Arctic islands	**55,541**	[**56,860**]	**56,894**	F
Franz Josef Land	—	**13,760**	13,739.8	F
Severnaya Zemlya	—	**19,370**	18,325	F
Novaya Zemlya	—	**23,650**	24,413	F
Ostrov Ushakova (Ushakov Island)	—	Included in Severnaya Zemlya	325	F
Ostrova De-longa (DeLong Islands)	—	**80**	81	F
Ostrov Viktoriya (Victoria Island)	—	—	11	F
Russia (continental)[4]		**23,860**	[5]**22,044**	F
Qatorkŭhi Hisor/Hisor Tizmasi and Alayskiy Khrebet	—	—	2,335.8	F
Pamirs	—	—	7,493.4	F
Tien Shan	—	—	7,251.7	F
Main Caucasus	1,805	**1,390**	1,424.4	F
Zhongghar Alataū Zhostasy	—	—	1,000	F
Altay	—	—	906.5	F
Kamchatka	—	—	874.1	F
Koryakskoye Nagor'y (upland)	—	—	240.6	F
Khrobet Suntar-Khayata	—	—	201.6	F
Khrebet Cherskogo	—	—	156.2	F
Gory Byrranga	—	—	30.5	F
Sayany	—	—	30.3	F
Ural Mountains	28	**20**	28.7	F
Russia (continental)[4]		**23,860**	[5]**22,044**	F

Table 1.—*Areal extent of present-day glaciers on Earth—Continued*

[Area is listed in square kilometers (km^2); ~, estimate; —, no information given]

Glacierized region	Source of data: Flint (1971); Sugden and John (1976)	Source of data: Ohmura (2009)	Source of data: U.S. Geological Survey Professional Paper 1386	Chapter or other source
	Area, estimated[1]	Area[2]	Area[1]	
Russia (continental)[4]—Continued				
Orulgan	—	—	18.4	F
Saūyr	—	—	16.6	F
Kuznetskiy Alatau	—	—	6.8	F
Lesser Caucasus	—	—	3.8	F
Plato Putorana	—	—	2.5	F
Khibiny	—	—	0.1	F
North American continent	**76,880**	[124,340]	**125,232**	—
Alaska	51,476	74,720	74,600	K
Mainland Canada	24,880	49,046	50,041	J
Western United States	513	563	580	J
México	11	11	11.44	J
South America	**26,500**	[25,856]	**25,063**	I
Venezuela	—	**2.51**	2	I
Colombia	—	111	104	I
Ecuador	—	**110.8**	97	I
Perú	—	1,780	2,600	I
Bolivia	—	**509.5**	560	I
Chile and Argentina	—	23,342	21,700	I
Europe	—	6,129	**5,759**	E
Scandinavia	**3,810**	3,058 (includes Jan Mayen)	**2,909**	E
Norway	—	—	2,595	E
Sweden	—	—	314	E
Alps	**3,600**	**3,059.71**	**2,842**	E
France	—	—	350	E
Switzerland	—	—	1,342	E
Austria	—	—	542	E
Italy	—	—	608	E
Pyrenees	**33**	**11.43**	**8.11**	E
Asia	—	—	**122,590**	—
Middle East	—	4,000 (includes Turkey, Iran, and Afghanistan)	44	G
Turkey	—	—	24	G

Table 1.—*Areal extent of present-day glaciers on Earth—Continued*

[Area is listed in square kilometers (km^2); ~, estimate; —, no information given]

Glacierized region	Source of data: Flint (1971); Sugden and John (1976)	Source of data: Ohmura (2009)	Source of data: U.S. Geological Survey Professional Paper 1386	Chapter or other source
	Area, estimated[1]	Area[2]	Area[1]	
Iran	—	—	27	G; Moussavi and others (2009)
Himalaya	33,200	—	—	—
Kunlun chains	16,700	—	—	—
Karakoram and Ghujerab- Khunjerab ranges	16,000	—	—	—
Other	**49,121**	—	—	—
China	—	60,035	59,406	F
Mongolia	—	—	[6]659	A; Dashdeleg and others (1983)
India	—	40,000 (includes Pakistan)	16,755	F
Nepal	—	7,500 (includes Bhutan)	5,324	F
Afghanistan	—	—	~2,700	F
Pakistan	—	—	~15,000	F
Bhutan	—	—	1,317	F
Africa	12	**10.85**	**10**	G
Southwest Pacific region	**1,015**	[1,165]	**1,167**	H
New Zealand	1,000	**1,158**	1,159	H
Irian Jaya, Indonesia	15	7	7.5	H
Total area of glaciers on Earth	~14,898,320	[7]**14,550,618**	15,926,087	Chapters A–K and Drewry and others (1982)

[1]Values in bold font in columns 2 and 4 are total area in larger glacierized regions. In column 3 (Ohmura, 2009), values in bold font with brackets are regional summaries by the editors.

[2]Areas of non-ice-sheet glaciers are from table 1 (Ohmura, 2009, p. 145) and from table 3 (Ohmura, 2009, p. 148). Column 2 in Ohmura's table 1 (2009) presents areas as given in the 1988 World Glacier Inventory (WGI) that is archived in the database of the World Glacier Monitoring Service (Haeberli and others, 1989a); Ohmura's table 1 (2009) presents the latest data in the WGI. Nonbold font as given by WGMS (Haeberli and others, 1989a) represents the glacierized area in each geographic region. Areas of the Greenland ice sheet and ice caps and mountain glaciers peripheral to the Greenland ice sheet and to the Antarctic ice sheet are from table 3 (Ohmura, 2009, p. 148).

[3]Including ice shelves and ice rises.

[4]Including mountain glaciers in the Former Soviet Union; Republics of Georgia, Kazakhstan, Kyrgystan, and Tajikistan.

[5]Includes Ostrov Vrangelya (Wrangel Island) of 3.5 km^2. Ohmura (2009) shows no glaciers on Wrangel Island; Dowdeswell and others (2010) in Chapter F of this volume, note only snow patches and no glaciers.

[6]The total glacierized area of Mongolia is not known accurately, because systematic work, including field and satellite remote sensing measurements, has begun only recently (Ulrich Kamp, written commun., 7 and 29 September 2010; Brandon Krumwiede, written commun., 6 and 16 September 2010). The 659 km^2 is from Dashdeleg and others (1983) who primarily used topographic maps.

[7]Table 1 in Ohmura (2009) gives two totals of the area of non-ice-sheet glaciers on Earth as being 549,056 km^2; table 3 in Ohmura (2009, p. 148) gives the area as 554,180 km^2. Table 3 also gives the area for the Greenland ice sheet as 1,699,000 km^2 and gives the area for the Antarctic ice sheet as 12,300,000 km^2, the area being a total of 13,999,000 km^2 for both. But when Ohmura totals the area for non-ice-sheet glaciers and for the two ice sheets as 14,548,056 km^2, he is using the total non-ice-sheet glacier area from table 1 only. In the final total given in column 3, the editors have averaged the two discrepant total areas for non-ice-sheet glaciers in Ohmura's table 1 and table 3, then added the average of the two areas (551,618 km^2) to the total ice-sheet areas (13,999,000 km^2) as being Ohmura's (2009) total area of glaciers on Earth.

Response of Glaciers to Global Climate Change

All elements of the Earth's cryosphere are sensitive to changes in global climate, whether it be cooler or warmer intervals of atmospheric temperature or more or less precipitation in the form of snow. During the 1970s and 1980s, a cooler interval of atmospheric temperature prevailed. Then in 1991, Mount Pinatubo, Republic of the Philippines, erupted, depressing global temperature by about 1°C (Robock and Oppenheimer, 2003). Since the mid-1990s, the Earth's climate began a sustained interval of global warming. The Arctic region, in particular, has experienced a sustained warming interval, signaling, perhaps, a comparatively abrupt change in a long-term cooling (Kaufman and others, 2009). According to recent projections by glaciologists, the sea-ice cover in the Arctic Ocean will completely disappear by 2100 (Boé and others, 2009).

The area and volume of glaciers have been decreasing since the mid-1990s, with most non-ice-sheet glaciers losing mass as they thinned and their termini retreated (Zemp and others, 2008). Only some non-ice-sheet glaciers are advancing: surge-type glaciers and tidewater glaciers. Surge-type glaciers are "quasi-cyclical"; tidewater glaciers are "cyclical." Such glaciers fluctuate due to dynamic processes that are not directly related to climate nor do glaciologists fully understood these processes. Therefore, even though most glaciers that lose mass during a warmer atmospheric-temperature interval respond by thinning or by having their termini or margins recede, surge-type glaciers and tidewater glaciers in various glacierized regions are still advancing. Crichton (2004) used this fact in his novel, "State of Fear," by creating the fictional Snorrajökull (misspelled in Crichton, 2004, p. 43), a surge-type outlet glacier from the Vatnajökull ice cap, Iceland (no such glacier place-name has ever been used in Iceland) (Sigurðsson and Williams, 2008). Crichton probably modeled Snorrajökull after Brúarjökull or Dyngjujökull, both of which are surge-type outlet glaciers of the Vatnajökull ice cap (Sigurðsson and Williams, 2008).

Changes in the Earth's two remaining ice sheets—in Greenland and Antarctica (tables 1, 2)—are unquestionably taking place. However, the large area (and volume) of these complex ice sheets presents a tremendous challenge for glaciologists to study and to monitor. These ice sheets include hundreds of outlet glaciers, ice streams, and ice shelves that vary widely in observed changes in their area (Ferrigno and others, 2008) and in their volume (Steffen and others, 2008), as well as varying geographically both latitudinally (for example, Greenland ice sheet) and (or) longitudinally (for example, Antarctic ice sheet). Earth-orbiting satellites carrying various types of sensors are the only feasible means by which glaciologists can measure changes in area, surface elevation, and volume in the Greenland and Antarctic ice sheets. Landsat sensors can measure changes in area, laser altimeters such as the Ice, Cloud and land Elevation Satellite (ICESat) can measure changes in surface-elevation, and the tandem Gravity Recovery and Climate Experiment (GRACE) satellites can measure changes in volume.

Importance of Glaciers to Monitoring Global Environmental Change

Change in the global environment, especially changes in global and regional climates, includes changes in atmospheric temperature and amount of precipitation as snow. In terms of such changes, glaciers are key indicators of atmospheric warming (and cooling) regionally and globally. The non-ice-sheet

Table 2.—*Estimated present-day area and volume of non-ice-sheet glaciers and ice sheets on Earth*

[GIS, Greenland ice sheet; km^2, square kilometers; km^3, cubic kilometers; ~, estimate; —, not shown]

Type of glacier	Area (km^2)	Percent	Volume (km^3)	Percent	Primary source for data
Ice caps, ice fields, valley glaciers, and other types of mountain glaciers	603,592 [1](680,000)	3.79	~160,000 [1](180,000)	0.49 [1](0.55)	Chapter A, Part A–2, table 1, of this volume (Meier and Bahr, 1996)[1]
	[2][549,056 – 554,180]	[1](4.27)	[100,251 – [2]31,781]		[Ohmura, 2009][2]
Greenland ice sheet	1,736,095	10.90	2,600,000	7.91	Holtzscherer and Bauer (1954) and Chapter C of this volume: volume of Greenland ice sheet
Ice caps and mountain glaciers peripheral to the Greenland ice sheet (total included in ice caps and features listed above)	48,599	0.31	—	—	Weidick (1995): Greenland ice sheet area and area of ice caps and other mountain glaciers peripheral to the Greenland ice sheet in Chapter C of this volume
	—	—	[1](20,000)	[1](0.06)	Meier and Bahr (1996): Volume of ice caps and other mountain glaciers peripheral to the Greenland ice sheet[1]
Antarctic ice sheet	13,586,400	85.31	30,109,800	91.55	Drewry and others (1982)
East Antarctica	10,153,170	—	26,039,200	—	Drewry and others (1982)
West Antarctica	1,918,170	—	3,262,000	—	Drewry and others (1982)
Antarctic Peninsula	446,690	—	227,100	—	Drewry and others (1982)
Ross Ice Shelf	536,070	—	229,600	—	Drewry and others (1982)
Ronne-Filchner ice shelves	532,200	—	351,900	—	Drewry and others (1982)
Subantarctic Islands	[3][15,419]	[0.10]	—	—	Various sources[3]
Totals	15,926,087	100.00	32,889,800	100.00	

[1]Based on scaling methods, Meier and Bahr (1996) estimated the total number of large ice caps (70) and other non-ice-sheet glaciers (160,000); their estimated glacier area and volume are also based on scaling methods. Chapter A uses areal data for glaciers compiled from Chapters B–K and from other sources (see table 1, Part A–2 of 1386–A).

[2]Ohmura (2009) gives two totals for the area of the Earth's non-ice-sheet glaciers: table 1 (549,056 km^2) and table 3 (554,180 km^2). Ohmura (2009) also gives the estimated total volume of the Earth's non-ice-sheet glaciers in table 2 as 100,251 km^3 (from Chen and Ohmura, 1990) and as 131,781 km^3 (from Bahr and others, 1997).

[3]See table 8 in Part A–2 of 1386–A; area is included in total of non-ice-sheet glaciers.

glaciers, the non-tidewater glaciers, the non-surge-type glaciers—glaciers in the temperate zones, in particular—respond relatively quickly to changes in climate by losing mass and thereby exhibiting thinning and recession of termini during warmer intervals of atmospheric temperature (Sigurðsson and others, 2007). These same types of glaciers gain mass, thicken, and have advance of termini during cooler intervals. The latitudinal distribution of glaciers, from the Equator to the poles, makes them geographically important "global indicators" of climate change. Oerlemans (1994), analyzing the glacier records that the World Glacier Monitoring Service (WGMS) archived, found that the observed retreat of glaciers during the 20th century followed a linear trend that corresponded to a temperature increase of +0.66°C per century. Changes in glaciers, whether by thinning of the glacier or by recession of its terminus, are equally obvious to glaciologists and to lay persons (Perkins, 2003). No special instruments other than one's eyes are needed to convince scientists and non-scientists alike that changes in the Earth's glaciers are taking place. Scientists, however, require documentation in the form of ground and aerial photographs or from satellite images (see Hastenrath, 2008; Sigurðsson and Williams, 2008; Balog, 2009) to confirm what they observe. By contrast, carbon dioxide (CO_2)

is an invisible odorless, colorless greenhouse gas in the atmosphere that has increased to 390 ppm in 2009 since the Industrial Revolution began in the late 1700s, when it was 280 ppm.

Carbon dioxide that cannot be smelled or seen has increased by 39 percent during more than three centuries, but only scientists (and non-scientists curious enough to obtain proper instruments to acquire CO_2 data) can confirm the ongoing changes in the Earth's atmosphere. Volumetrically, glacier ice represents 2.15 percent of the Earth's water; 97.2 percent of the Earth's water is contained in the oceans and seas; 0.6 percent is held in groundwater; and 0.017 percent is in surface water (rivers, streams, lakes) (U.S. Geological Survey, 1976; Williams, 1986a, table 9.3). It is crucial to understand that less than 3 percent of the water on Earth is freshwater. Although glacier ice represents the second most important component of the global hydrologic cycle, it is the largest reservoir of freshwater . Glacier ice, therefore, represents water that is sequestered from the hydrologic cycle, so long as it remains frozen. The volume of water sequestered in glacier ice on land is, however, sufficient, to raise sea level by as much as 80 m if that ice melts (table 3) [*http://pubs.usgs.gov/fs/fs133-99/gl-vol.html*]. Many glaciologists consider that non-ice sheet glaciers, especially those in southeastern Alaska (Meier, 1984), are the main contributors of glacier meltwater to the observed rise in sea level (Arendt and others, 2002; Raper and Braithwaite, 2005; Meier and others, 2005, 2007; Bahr and others, 2009) (see also the discussion on Glacier Mass Changes and Their Effect on the Earth System, p. A192–A223, in part A-2 of this chapter). Because all the non-ice sheet glaciers could contribute a maximum of only about 0.45 m to rise in sea level (table 3; Meier and Bahr, 1996, table 1, p. 94), glacial meltwater from the Greenland ice sheet and (or) from the Antarctic ice sheet would have to be the source for major future rises in sea level from the melting of glacier ice (Intergovernmental Panel on Climate Change [IPCC], 2007a, b; Steffen and others, 2008). In addition, the increase in temperature of oceanic waters causes a volumetric (steric) expansion in the Earth's oceans, further contributing to the rise in sea level. During the 20th century, meltwater from mountain glaciers and from steric changes in the oceans each contributed about 50 percent to the observed rise in global sea level.

Ice cores represent a record of past climates and of the composition of the Earth's atmosphere. The record from an ice core from Dome "C," Antarctica, provides a continuous record for the last 800,000 years (EPICA Community Members, 2004). Depending on the latitudinal location of the glacier, ice cores can provide information on variations of El Niño in conjunction with the Southern Oscillation (tropical ice cores) (see Ice Cores, High-Mountain Glaciers, and Climate, by L.G. Thompson, p. A157–A171, in part A-2 of this chapter); ice cores can also provide information on concentrations of CO_2 and methane (CH_4), dustiness of the atmosphere, explosive volcanic activity from geochemical analysis of tephra layers, and $\delta^{18}O$ proxy temperature (the ratio of $^{18}O : {}^{16}O$ gives a measure of past temperature), among other constituents within the ice.

Changes in the area of glaciers alter the overall albedo of the Earth. Snow-covered or relatively clean glacier ice has a high albedo; it reflects incoming solar radiation back into space. Debris-covered glaciers and deglacierized terrain, whether vegetated or unvegetated, have a much lower albedo, and newly deglacierized terrain absorbs more solar energy, thus contributing to a warmer surface on the land.

Table 3.—*Estimated present-day volume of glaciers and of maximum potential rise in sea-level*

Glacierized region	Volume, estimated (cubic kilometers)	Percentage of volume of Earth's glaciers	Maximum potential rise in sea level (meters)[1]	Primary source for estimated volume
Ice caps, ice fields, valley, and other mountain glaciers	**180,000**	0.55	**0.45**	Meier and Bahr (1996)[2]
	[3][50,000]		[3][0.15 – 0.37]	Dyurgerov and Meier (2005)
	[3][100,251 – 131,781]		[3][0.25 – 0.31]	Ohmura (2004, 2009)
Greenland ice sheet (Inland Ice); local ice caps and glaciers	**2,600,000**	7.90	**6.50**	Holtzscherer and Bauer (1954)
	20,000	0.06	**0.05**	Meier and Bahr (1996)
	[3][2,900,000]		[3][7.3]	[Bamber and others (2001)]
Antarctic ice sheet	[2, 4]**30,109,800**	91.49	**73.44**	Drewry and others (1982)
	[3][24,700,000]		[3][56.6]	Lythe and others (2001)
East Antarctica	26,039,200		64.80	Drewry and others (1982)
West Antarctica	3,262,000		8.06	Drewry and others (1982)
Antarctic Peninsula	227,100		0.46	Drewry and others (1982)
Ross Ice Shelf	229,600		0.01	Drewry and others (1982)
Ronne-Filchner ice shelves	351,900		0.11	Drewry and others (1982)
Totals	**32,909,800**	100.00	**80.44**	

[1]Potential rise in sea level is defined as the maximum rise in sea level expected if all glacier ice in a specified glacierized geographic region were to melt. The potential rise is based on a density of 0.9 for glacier ice (Robin, 1967), an ocean area of 361,419,000 square kilometers (km^2) (National Geographic Society, 2005), and a volume of 400 cubic kilometers (km^3) of melted glacier ice to raise sea level 1 millimeter. The volume of glacier ice that is below sea level in the Greenland and the Antarctic ice sheets would have to be subtracted from the total volume of both ice sheets, however, to give a more accurate projection of maximum potential rise in sea level.

[2]By including the volume of ice caps and of other glaciers peripheral to the Antarctic and Greenland ice sheets, Dyurgerov and Meier (2006) increased the total volume of non-ice-sheet glaciers to 260,000 ±65,000 km^3. This chapter uses the estimates given by Meier and Bahr (1996).

[3]Ice volume (in cubic kilometers) and potential rise in sea level (in meters) from decrease in ice volume of non-ice-sheet glaciers, Greenland ice sheet (Inland Ice), and Antarctic ice sheet shown in brackets are from citations of scientific publications in table 1 of Part A1 (Intergovernmental Panel on Climate Change [IPCC], 2007a, p. 342). The calculations of ice volume by Ohmura (2004) and by Dyurgerov and Meier (2005) for the Earth's non-ice-sheet glaciers and their contribution to a potential rise in sea level exclude similar calculations for mountain glaciers and ice caps peripheral to the Greenland ice sheet and the Antarctic ice sheet. Ohmura's table 2 (2009) gives two estimates for the volume of the Earth's non-ice-sheet glaciers: 100,251 km^3 (from Chen and Ohmura, 1990) and 131,781 km^3 (from Bahr and others, 1997). Based on these discrepant totals for the volume of ice in non-ice-sheet glaciers, Ohmura (2009) gives a maximum potential rise in sea level of 0.25 to 0.31 meters if all the ice in the Earth's non-ice-sheet glaciers completely melted.

[4]The total volume of glacier ice in Antarctica is 30,109,800 km^3. To calculate the potential rise in sea level, only the grounded parts of the Antarctic ice sheet (including ice rises within the ice shelves) are used, for a total grounded-ice volume of 29,377,800 km^3. The total grounded-ice volume includes 25,921,700 km^3 for East Antarctica, 3,222,700 km^3 for West Antarctica, and 183,700 km^3 for the Antarctic Peninsula. The volume of ice rises on the Ross Ice Shelf is 5,100 km^3, and on the Ronne-Filchner ice shelves, the volume of ice rises is 44,600 km^3.

Other Important Scientific Aspects of Glaciers

In many arid and semi-arid regions, such as the Andes of South America and the glacierized interior mountains of Asia, meltwater from glaciers during the summer months provides the only source of water for drinking, crop irrigation, and other uses imperative to living organisms (Meier and Roots, 1982). As the glaciers disappear, so does the water supply for humans and other living organisms (Barnett and others, 2005; Francou and Coudrain, 2005; Bradley and others, 2006; Appenzeller, 2007; The Economist, 2007a; Vergara and others, 2007; Rosenthal, 2009). During the 20th century and during at least the initial decades of the 21st century, nations such as India will likely experience surplus water supplies, much greater than before because streamflow from melting glaciers adds to the overall volume of runoff. Eventually there will be a need for a readjustment to normal river discharge, when the glaciers are gone. In about 70 years, when the glaciers will be much diminished, especially those at lower elevations, nations in Asia and South America will likely find it difficult to reaccommodate to the normal river discharge, which had been considerably less before the melting of glaciers had begun.

Many countries depend on glacier meltwater to sustain or to augment the discharge of rivers into artificial reservoirs that provide water sufficient to generate hydroelectric power. Norway, Iceland, Switzerland, and British Columbia (Canada), for example, have institutions staffed by hydroglaciologists who monitor changes in the area and volume of their glaciers on an annual basis, because the information is needed to maintain the optimum generation of hydroelectric power. Under warmer climate conditions, glaciers lose mass (ice melts), thus making a greater contribution of glacier meltwater to glacierized hydrologic basins. Glaciers, therefore, can be considered an ephemeral landform that can completely disappear (melt away) from the landscape during intervals of warmer climate, like that which is occurring now (2012) in many glacierized regions, such as Iceland (Sigurðsson and Williams, 2008), and especially in tropical locations, such as Irian Jaya (Allison and Peterson, 1989), Indonesia, Africa (Young and Hastenrath, 1991), and South America (Chapters G, H, and I in this volume). Nesje and others (2008) discussed the disappearance of glaciers in Norway during warmer intervals in the Holocene Epoch, which became reestablished during cooler intervals.

Glaciers can also be the source of a number of hazards, including eustatic rise in sea level, increase in activity of surge-type glaciers and tidewater glaciers, jökulhlaups (both lacustrine and volcanic/geothermal), ice avalanches, debris avalanches, lahars, rockslides, and icebergs.

These hazards from glaciers will be discussed in Glacier Mass Changes and Their Effect on the Earth System by Mark B. Dyurgerov and Mark F. Meier, p. A192–A223, and in Glaciological Hazards: Global, Regional, and Local, p. A261–A274.

The Antarctic ice sheet is the Earth's largest repository of meteorites. The cold, dry climate of the polar plateau preserves all types of meteorites, including chondritic, achondritic (McSween and others, 1979), iron, stony-iron, shergottites (rock types from Mars; Peterson, 1997; Cassidy, 2003), lunar rock types (Eugster, 1989; Cassidy, 2003), and so on. Most of the meteorites now archived in the world's natural history museums have been collected from Antarctica. The Smithsonian Institution's National Museum of Natural History is the permanent archive for meteorites found in Antarctica under the U.S. Antarctic Search for Meteorites Program (ANSMET); Peterson (1997) noted that 8,000 meteorites had been collected during the three decades from 1976 to 1997 under the auspices of ANSMET. Until the Japanese discovery of an assemblage of meteorites in the Yamato (Queen Fabiola) Mountains of Antarctica in 1969 (Yoshida and others, 1971), about 15,000 meteorites had been found worldwide, including four from Antarctica (Williams and others, 1983). By 2012, scientists from many nations working in Antarctica had found more than 49,000 meteorites (R.P. Harvey, written commun., 11 July 2011), more than tripling the number of meteorites archived worldwide before 1969. Analysis of those meteorites has contributed to a highly fruitful increase in scientific knowledge about the origin of the Solar System and about the geochemistry of meteorites (Cassidy, 2003; Harvey, 2003). Blue-ice areas in Antarctica are the primary "collection" sites for meteorites, which can be collected annually because the sublimation of ice is occurring so rapidly (by as much as 1 m) in such areas that an annual "lag deposit" of meteorites is exposed at the surface. Landsat and other satellite images are used to identify and locate "blue ice" areas in Antarctica (Williams and others, 1983).

The first-time capability for monitoring and measuring changes in glacier area on a global basis by using Landsat images provided the impetus for undertaking the preparation of the 11-chapter U.S. Geological Survey Professional Paper 1386-A–K, Satellite Image Atlas of Glaciers of the World, a task that would eventually involve 113 glaciologists and other scientists contributing to the publication of the series. The use of satellite remote sensing technologies and geographic information systems (GIS) technologies will be discussed in Global Land Ice Measurements from Space (GLIMS) by Bruce H. Raup and Jeffrey S. Kargel, p. A247–A260, and in Monitoring Changes in Length, Area, and Mass Volume of Glaciers, p. A224–A246. Improved resolution in sensors such as the Geoscience Laser Altimeter System (GLAS) sensor on the ICESat satellite, has led to more accurate measurements of changes in the elevation of the Greenland and the Antarctic ice sheets. Data from the GRACE tandem satellites are providing accurate information about seasonal and interannual changes in mass of the Greenland and the Antarctic ice sheets.

Classification of Glaciers

Meier (1974) provided a definition of a glacier:

> A glacier may be defined as a large mass of perennial ice that originates on land by the recrystallization of snow or other forms of solid precipitation and that shows evidence of past or present flow. The definition is not precise, because exact limits for the terms large, perennial, and flow cannot be set. Except in terms of size, a small snow patch that persists for more than one season is hydrologically indistinguishable from a true glacier. One international group has recommended that all persisting snow and ice masses larger than 0.1 square kilometer (about 0.04 square mile) be counted as glaciers. Hence, in the absence of an agreed-upon upper size limit for glaciers, a body of ice as large as the Antarctic Ice Sheet (slightly smaller than the conterminous United States and Europe combined) could properly be considered a glacier.

Therefore, glaciers on Earth today (fig. 2*A*, p. A72) can range in size from 0.1 km^2 (mountain glacier) to 13.6×10^6 km^2 (Antarctic ice sheet). From the standpoint of satellite remote sensing under optimum conditions, the smallest glacier resolvable on a Landsat Multispectral Scanner (MSS) image (80-m picture element or pixel) is 0.1 km^2. C. Simon L. Ommanney (written commun., 2009) reports the recommendation by a Working Group meeting at the 2008 International Workshop on World Glacier Inventory, held in Lanzhou, China, that the minimum area of a valley glacier to be inventoried is

0.1 km^2 (Paul, Barry, and others, 2009). For a schematic diagram showing the delineation of a glacier and many of its features, see figure 2*B* (p. A72) from Müller and others (1977).

Geophysical Classifications

Glaciers can be described in terms of ice temperature and degree of surface melting. Ahlmann (1935) proposed three categories of "geophysical classification" described later by Paterson (1994): temperate glaciers (≥0°C throughout the ice), sub-polar glaciers, and high-polar glaciers (surface temperature is always less than 0°C). Glaciologists now use the term "polythermal" instead because most glaciers exhibit a variable temperature regime. On the basis of field work in Alaska, Canada, Washington State, and Greenland, Benson (1959, 1961, 1962) proposed that a glacier be divided into two areas: an accumulation area divided into three diagenetic facies based on the degree of surface melting and an ablation area (fig. 3). Between the ablation area and the accumulation area, Müller (1962) defined an equilibrium line at which the mass balance of the glacier is zero. Shumskiy (1964) also discussed the concept of firn facies on a glacier. Benson (Benson and Motyka, 1979) refined and updated his original idea (Benson, 1962) of diagenetic facies, renaming it "glacier facies." Williams and others (1991) applied Benson's concept to analysis of glacier facies on satellite images (fig. 4).

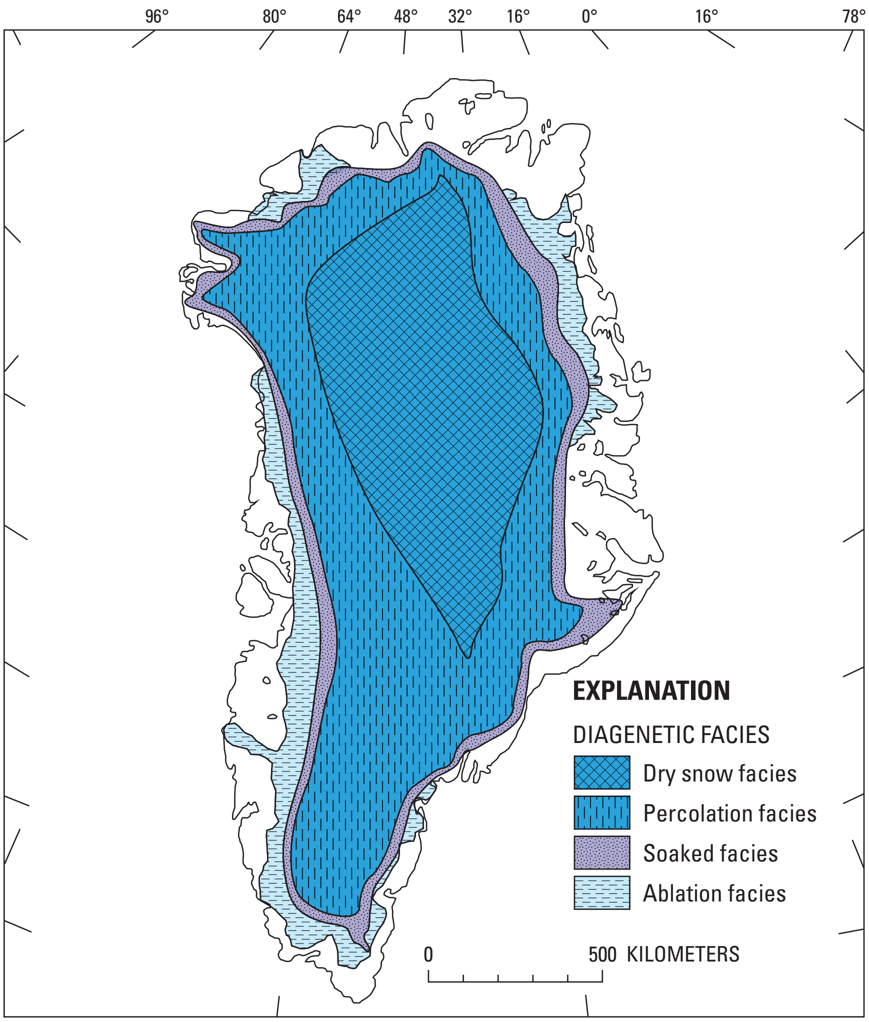


Figure 3.—*Diagenetic facies on the Greenland ice sheet. Modified from Benson (1961).*

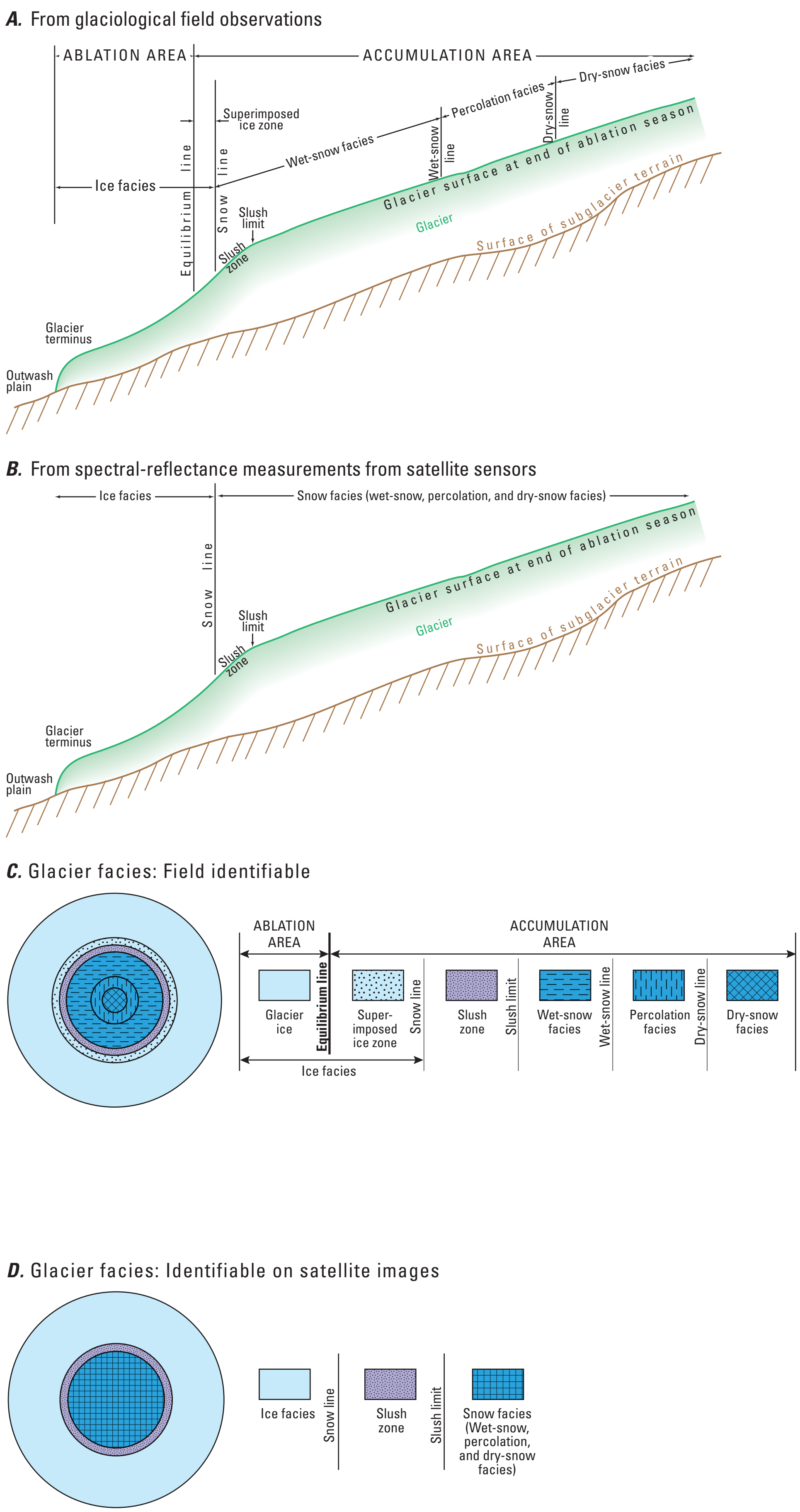


Figure 4.—*Cross-sections of a glacier showing glacier facies at the end of the balance year.* **A**, *From glaciological field observations.* **B**, *From spectral-reflectance measurements from satellite sensors. Modified from Williams and others (1991). Schematic diagrams of glacier facies.* **C**, *Field-identifiable facies.* **D**, *Facies identifiable on satellite images. Modified from Williams and Hall (1993).*

Morphological Classifications

Ahlmann (1940, p. 192) proposed a morphological classification of glaciers that categorized 11 different types into 3 broad groups: A, continental glaciers, ice caps, and highland glaciers; B, valley glaciers, transection glaciers, circus (sic, cirque) glaciers, wall-sided glaciers, and floating glacier tongues; and C, piedmont glaciers, foot glaciers, and shelf ice. In "Perennial Ice and Snow Masses," (United National Educational, Scientific, and Cultural Organization [UNESCO], 1970a) glaciers are classified in 8 primary categories: continental ice sheet, ice-field, ice cap, outlet glacier, valley glacier, mountain glacier, glacieret and snowfield, and ice shelf (10 categories if miscellaneous and rock glacier are included) (table 4). Armstrong and others (1973), in their "Illustrated Glossary of Snow and Ice," defined and illustrated 12 types of glaciers: ice sheet, inland ice sheet, ice cap, outlet glacier, piedmont glacier, ice piedmont, ice fringe, valley glacier, cirque glacier, ice stream, ice shelf, and glacier tongue.

In the Union of Soviet Socialist Republics (USSR) Glacier Inventory, Vinogradov (1966) proposed 33 different morphological types of glaciers. However, Vladimir Kotlyakov and his colleagues at the Institute of Geography in Moscow reduced the number of types to 13 (fig. 5*A*). United Nations Educational, Scientific, and Cultural Organization (UNESCO) (1970a) and Müller and others (1977) subsequently provided simple sketches of five basin types (fig. 5*B*) and five frontal characteristics (fig. 5*C*) of glaciers that glaciologists use worldwide in preparing glacier inventories. For their monumental work "World Atlas of Snow and Ice Resources" (Kotlyakov, 1997a), the Russian glaciologists expanded their initial classification scheme into three broad classes of glaciers and their complexes: ice sheet glaciation, reticular glaciation, and mountain glaciation (Kotlyakov, 1997a, v. 2, table 6, p. 146). Table 5 in this part (A-2) of Chapter A, modified from the legend for the maps in the Russian World Atlas, shows the glacier classes, morphological types, and morphological subtypes. Sugden and John (1976, p. 56) proposed a simple morphological classification of glaciers, based on the interaction of glacier ice with its topographic setting (table 6).

With the impetus provided from 1957 to 1958 by the International Geophysical Year (IGY) and 1965 to 1974 by the International Hydrological Decade (IHD), the international glaciological community expanded its collaborative work on monitoring the fluctuations of glaciers to include compiling and assembling data that would provide a world inventory of glaciers (United Nations Educational, Scientific, and Cultural Organization [UNESCO], 1970a). The efforts of the Russian and the Italian (Consiglio Nazionale delle Ricerche [CNR], 1959, 1961, 1962) inventories were motivated by the IGY. C. Simon L. Ommanney (written commun., 2009) quoted from the IGY report (International Union of Geodesy and Geophysics [IUGG], 1956), "The first Canadian glacier inventory was initiated in response to an IGY resolution, first proposed in 1955, by the Comité Spécial de l'Année Géophysique Internationale (CSAGI) International Union of Geodesy and Geophysics (IUGG) (1956), for countries to undertake a census of their glaciers. The specific recommendations for this inventory were published in 1959 (Annals of the IGY, 1959)." C. Simon L. Ommanney (written commun., 2009) recently reprised this history (Ommanney, 2009), also stating "The IHD initiative was, I believe, a recognition that not all countries had committed to the IGY inventory and that some standardization was desirable. Also that the emphasis had changed from glaciers as geophysical phenomena to [glaciers as] hydrological

Table 4.—*Classification and description of glaciers, according to the World Glacier Monitoring Service System*

[Examples of each type glacier are taken from chapters of the "Satellite Image Atlas of Glaciers of the World" (U.S. Geological Survey Professional Paper 1386); <, less than; >, greater than; km^2, square kilometers; NA, not applicable; —, not shown. Modified from United Nations Educational, Scientific, and Cultural Organization (1970a); Müller and others (1977); Müller (1978)]

Glacier classification	Description	Figure in Part A–2 of this chapter
Miscellaneous	All those not listed.	NA
Continental ice sheet	Ice mass that inundates areas of continental size, and its radial flow completely covers the landscape >50,000 km^2 (Armstrong and others, 1973) with the exception of nunataks. Composite of many outlet glaciers, ice streams, and, on the oceanic margin, floating ice shelves, as in Antarctica.	4
Ice field	Ice masses of insufficient thickness to completely bury the subsurface topography.	5
Ice cap	Dome-shaped ice mass with radial flow, which completely covers the landscape, except nunataks; area is <50,000 km^2 (Armstrong and others, 1973). Composite, usually, of many outlet glaciers.	6
Outlet glacier	Flows from an ice sheet, ice cap, or ice field, usually of valley-glacier or lobate form; the catchment area is generally not clearly delineated.	7
Valley glacier	Flows down a valley.	8
Mountain glacier	Cirque, niche, or crater type; includes ice aprons and groups of small units. Flow pattern is visible as an accumulation of material along the margin and (or) as a discharge of glacial rock flour, which signifies downslope flow of ice and active subglacial erosional processes.	9
Glacieret and snowfield	Glacierets are small ice masses of indefinite shape (situated in hollows, riverbeds, and on protected slopes) that have developed from snow drifting, avalanching, and (or) especially heavy snow accumulation in certain years; usually no marked flow pattern is visible, and therefore no clear distinction from a snowfield is possible. Exists for at least two consecutive summers.	—
Ice shelf	A floating ice sheet of considerable thickness attached to a coast and nourished by one or more outlet glaciers and (or) ice streams; has additional snow accumulation on its surface and (or) bottom freezing from sea water.	10
"Rock glacier"	A lobe-shaped mass of angular rock in a cirque or along a valley wall (sometimes with interstitial ice, firn, and snow) that moves slowly downslope. "Rock stream" or "block stream" is a better term. The description refers to a periglacial or mass-movement process, not to a glacier.	—

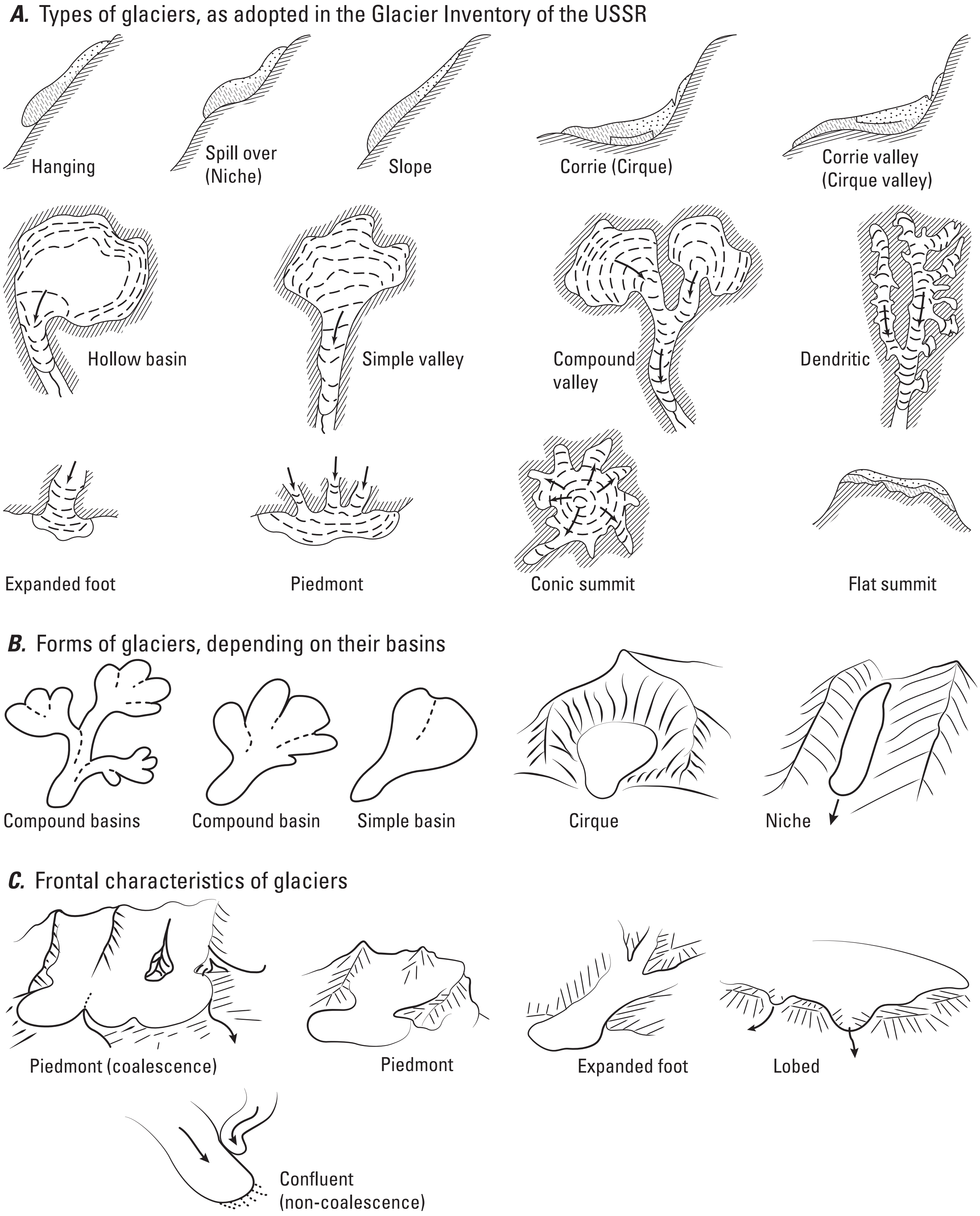


Figure 5.— **A**, *Types of glaciers, based on the glacier inventory of the U.S.S.R. Modified from Vinogradov (1966).* **B**, *Diagrams showing variation in forms of glaciers in different basins and,* **C**, *frontal characteristics of glaciers. Modified from Müller and others (1977, p. 15–17, figs. 3a–3e and figs. 4a–4e).*

Table 5.—*Types of glaciers and their complexes from the Russian World Atlas of Snow and Ice Resources*

[<, less than; >, greater than; km^2, square kilometers. From Kotlyakov (1997a, legend; 1997b, table 16, p. 146)]

Ice sheet glaciation
Continental glacier complex: ice sheets, ice caps, outlet glaciers and ice shelves
Large island glacial complexes (area >1,000 km^2): ice sheets and associated outlet glaciers
Island glacial complexes (area <1,000 km^2): ice caps and associated outlet glaciers
Outlet glaciers within the structure of island complexes
Ice shelves
Reticular glaciation
Glacial complexes of plateaux
Glacial complexes of valleys
Piedmont glaciers
Mountain glaciation
Glacial complexes of conic summits
Glacial complexes of high plateaux
Dendrite glaciers
Compound-valley glaciers
Valley glaciers
Hollow basin glaciers
Corrie-valley glaciers
Corrie glaciers
Hanging glaciers
Slope glaciers
Niche glaciers
Glaciers of the flat summits

Table 6.—*A simple morphological classification of glaciers*

[Modified from Sugden and John (1976, p. 56, table 4.1)]

Glaciers unconstrained by topography
Ice sheets (including ice domes)
Ice caps (including ice domes and lobate outlet glaciers)
Glaciers constrained by topography
Ice fields and outlet glaciers (constrained by subglacial valley)
Outlet glaciers from ice caps (constrained by subglacial valley)
Valley glaciers
Cirque glaciers
Other smaller mountain glaciers

[phenomena] with implications for the world's water supply." Now (2012), the emphasis has changed again to their relationship to climate and its changes with concomitant implications for rising sea level.

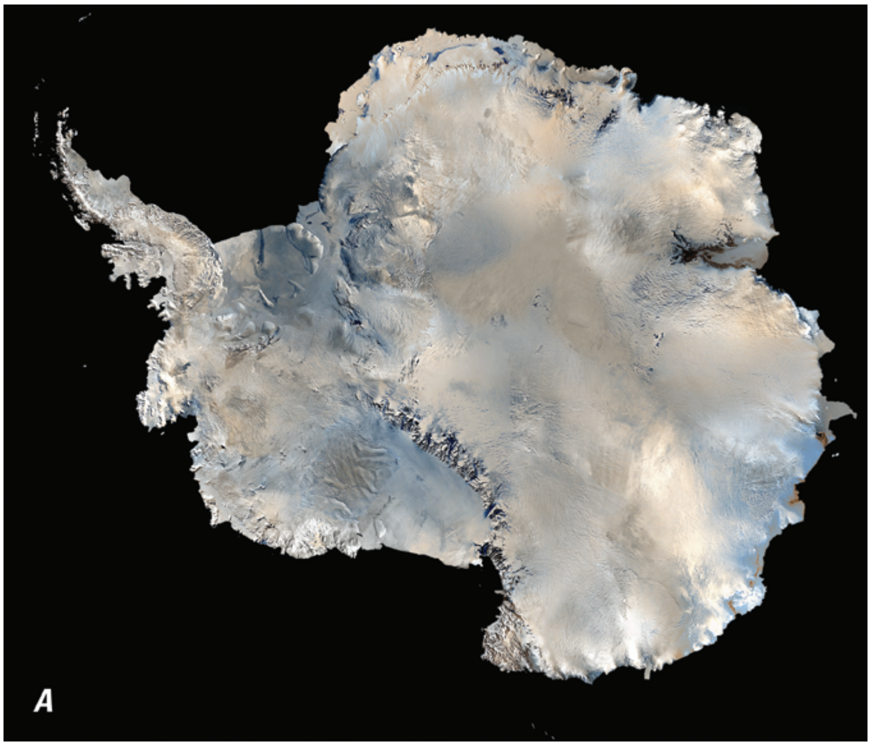
Fritz Müller, who, with C. Simon L. Ommanney and A.D. Stanley of Canada, had contributed pilot studies of glacier inventories to the UNESCO's "Perennial Ice and Snow Masses" booklet (United Nations Educational, Scientific, and Cultural Organization [UNESCO], 1970a), established a Temporary Technical Secretariat for a World Glacier Inventory (WGI) in the Department of Geography at the Swiss Federal Institute of Technology (ETH), Zürich, Switzerland. "Instructions for Compilation and Assemblage of Data for a World Glacier Inventory" were distributed to the international glaciological community (Müller and others, 1977; Müller, 1978). Because this ambitious international effort preceded the launch of an Earth-orbiting satellite, Earth Resources Technology Satellite-1 (ERTS-1), later renamed Landsat 1, the first medium-resolution civilian spacecraft capable of repeatedly imaging all of the Earth's glaciers that lay between the Equator and about latitudes 81° North and South, the WGI assumed that glaciologists would use conventional field mapping techniques and observations and analysis of vertical aerial photographs (single or stereographic pairs of photographs) for the source of data. For the classification (morphology) and description of glaciers that had been published in the three booklets that had been distributed (United Nations Educational, Scientific, and Cultural Organization [UNESCO], 1970a; Müller and others, 1977; and Müller, 1978), see table 6.

After the unexpected death of Müller in 1980, Karl Scherler became the Acting Director of the WGI Project. During his tenure, a supplemental UNESCO booklet was prepared in recognition of the increasing availability of images from satellites: "Guidelines for Preliminary Glacier Inventories" (Scherler, 1983). Of the four levels of comprehensiveness of data that the guidelines recommended, the first three were based solely on satellite images; only the fourth level required topographic maps in addition to satellite images. Scherler's tenure as Acting Director was followed by that of Wilfried Haeberli, who, as Director, eventually merged the WGI with the Swiss Permanent Service on the Fluctuation of Glaciers to create the current World Glacier Monitoring Service (WGMS).

The WGI (now WGMS) system for classifying glaciers (table 4) is used in the U.S. Geological Survey's 11-chapter "Satellite Image Atlas of Glaciers of the World" volume (U.S. Geological Survey Professional Paper 1386-A–K). Examples of seven types of glaciers are shown in this part's figures 6 through 12: continental ice sheet (fig. 6), ice field[2] (fig. 7), ice cap (fig. 8), outlet glacier (fig. 9), valley glacier (fig. 10), mountain glacier (fig. 11), and ice shelf (fig. 12). Each figure is drawn from satellite images included in the 10 geographic-region chapters of the volume.

The Global Land Ice Measurements from Space (GLIMS) Initiative, another ambitious international program to use satellite images from the ASTER sensor on the Terra spacecraft (formerly EOS-1), along with data from other satellite sensors, was initiated in 1995 by Hugh Kieffer of the U.S. Geological Survey (see section by B.H. Raup and J.S. Kargel on GLIMS, p. A247–A260, in part A-2 of this chapter). Because of the availability of satel-

[2] "Ice field" is the preferred form ("Glossary of Geology," by Neuendorf and others (2005, p. 318). The United Nations Educational, Scientific, and Cultural Organization (UNESCO) (1970a, p. 16) guide and Müller and others (1977, p. 13) used "ice-field". In Alaska, "icefield" is used in geographic place-names (for example, Harding Icefield). The editors' guide for publications of the International Glaciological Society uses "icefield", reserving "ice field" for an extensive area of pack ice.



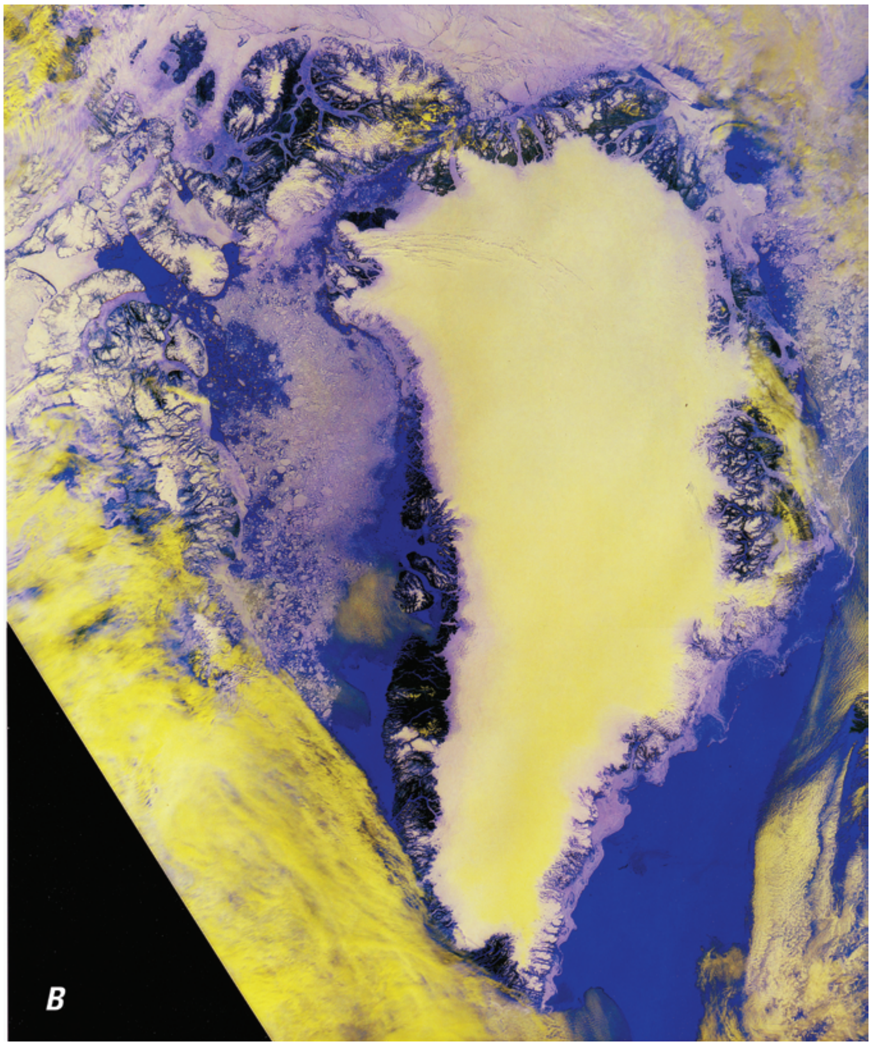


Figure 6.—<u>Continental ice sheet.</u> **A**, *NOAA AVHRR image mosaic of the Antarctic ice sheet (*http://terraweb.wr.usgs.gov/projects/Antarctica/AVHRR.html*) used as the image base for the USGS Satellite Image Map of Antarctica (I-2560) (Ferrigno and others, 1996, 2000).* **B**, *NOAA-H AVHRR image of the Greenland ice sheet (for an extended caption, see Chapter C of "Satellite Image Atlas of Glaciers of the World," p. C92, fig. 66).*

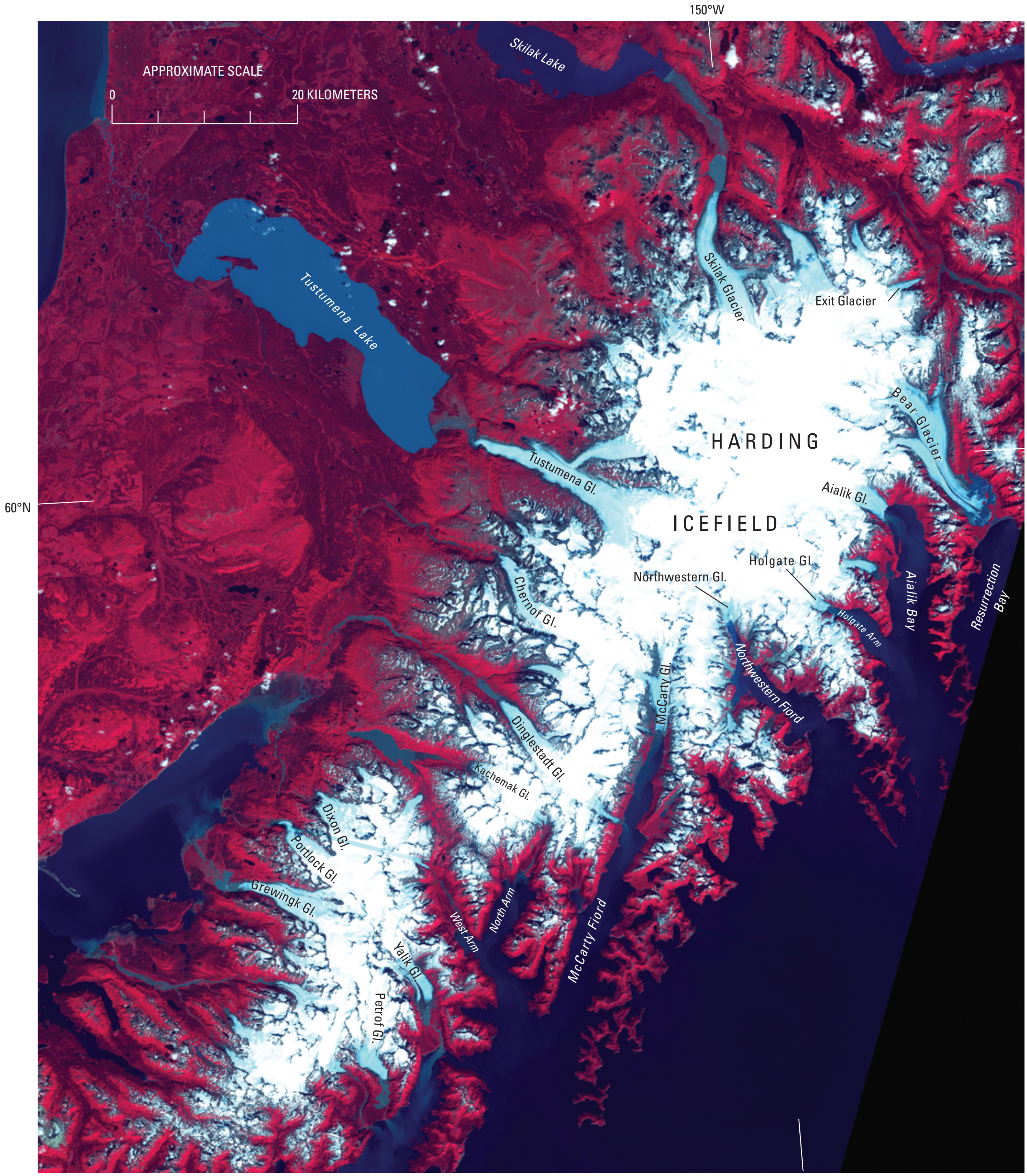


__Figure 7.__—*<u>Ice field.</u> Landsat 7 ETM+ image of the Harding Icefield, Alaska (for an extended caption see Chapter K of "Satellite Image Atlas of Glaciers of the World," p. K338, fig. 312).*

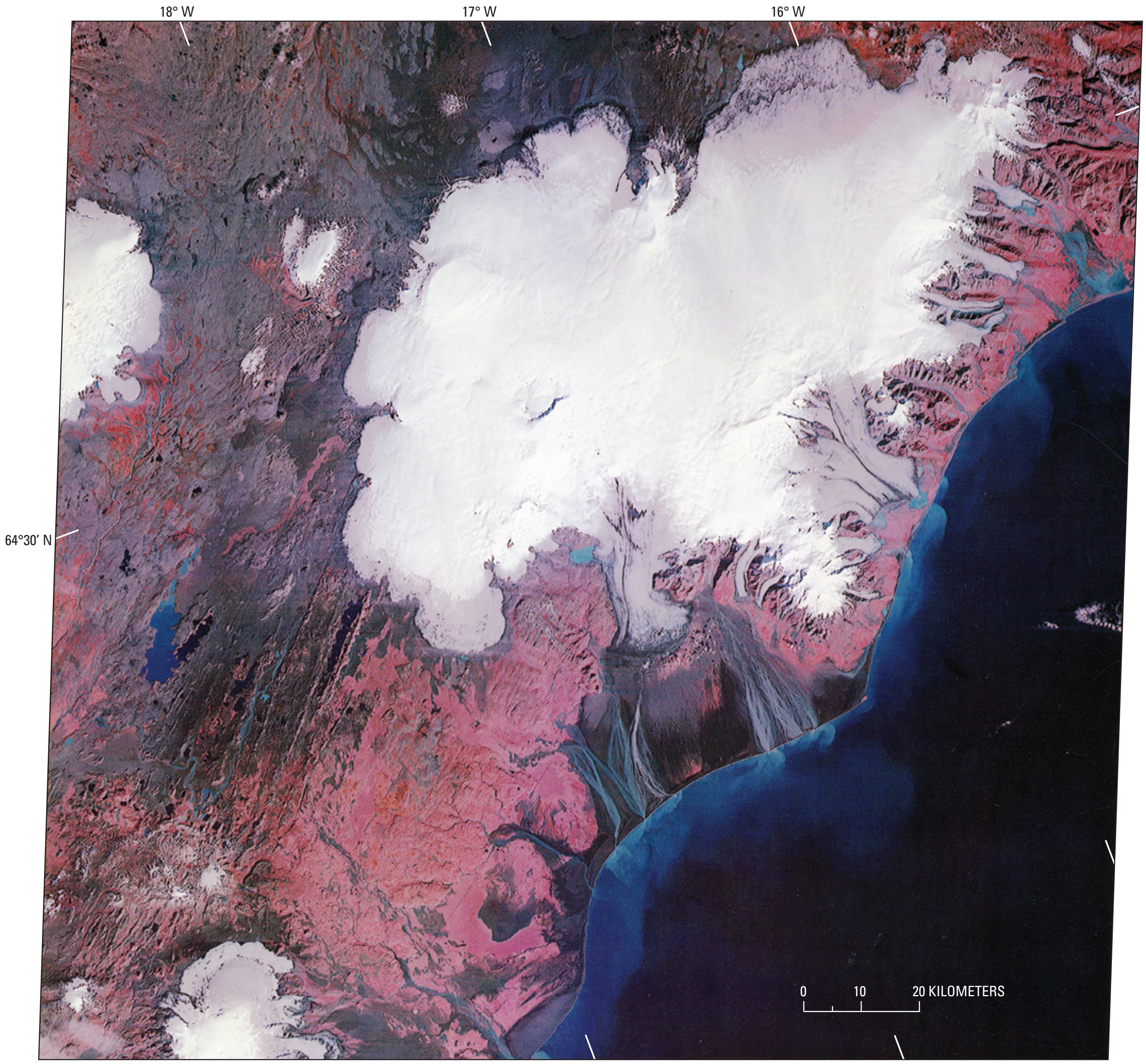


Figure 8.—*<u>Ice cap.</u> Landsat 1 MSS false-color composite image of the Vatnajökull ice cap, Iceland, on 22 September 1973. Image no. 1426-12070 was enhanced digitally by the USGS Earth Resources Observation System (EROS) Digital Image Enhancement System (EDIES). See Sigurðsson and Williams (2008, p. 197, fig. 188).*

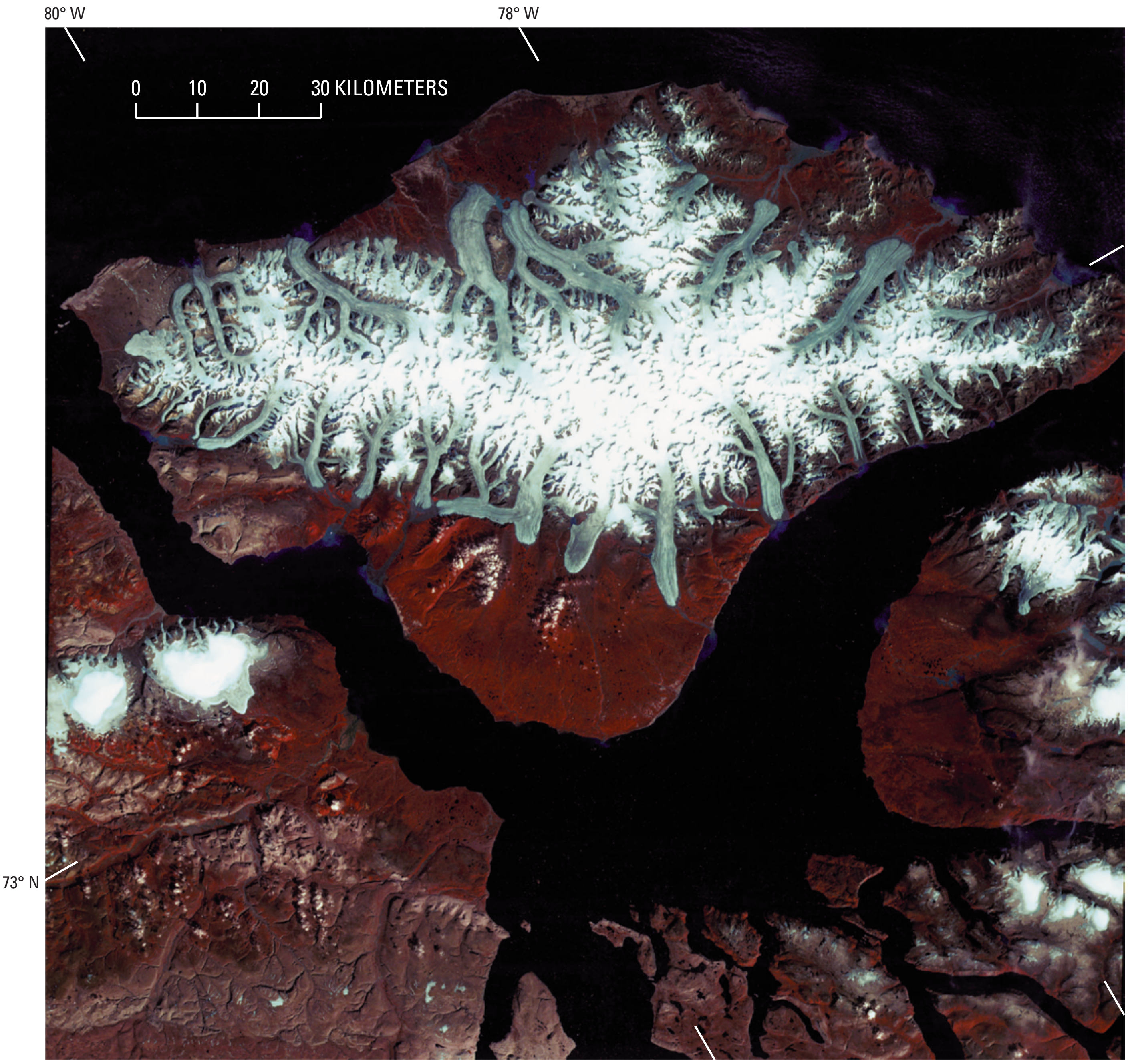


Figure 9.—*<u>Outlet glacier.</u> Landsat 2 MSS false-color composite image of outlet glaciers flowing from the ice field on Bylot Island, Nunavut, Canada (for an extended caption, see Chapter J of "Satellite Image Atlas of Glaciers of the World," its cover page and p. J173, fig. 3).*

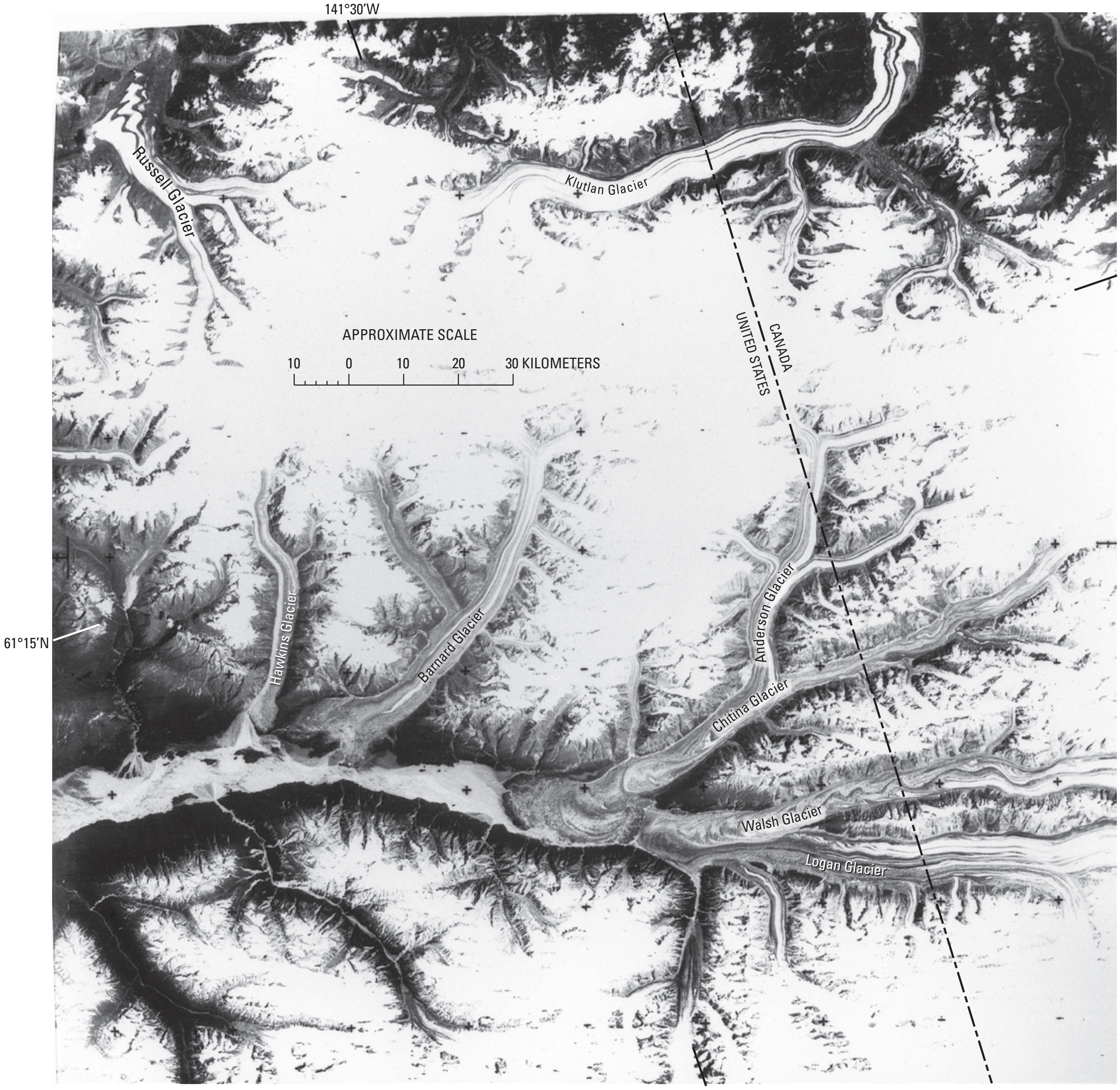


Figure 10.—Valley glacier. *Landsat 3 RBV image of valley glaciers, northwestern St. Elias Mountains, Alaska and Canada (for extended captions see Chapter K, and Chapter J of "Satellite Image Atlas of Glaciers of the World," p. K191, fig. 175; and p. J306, fig. 3, respectively).*

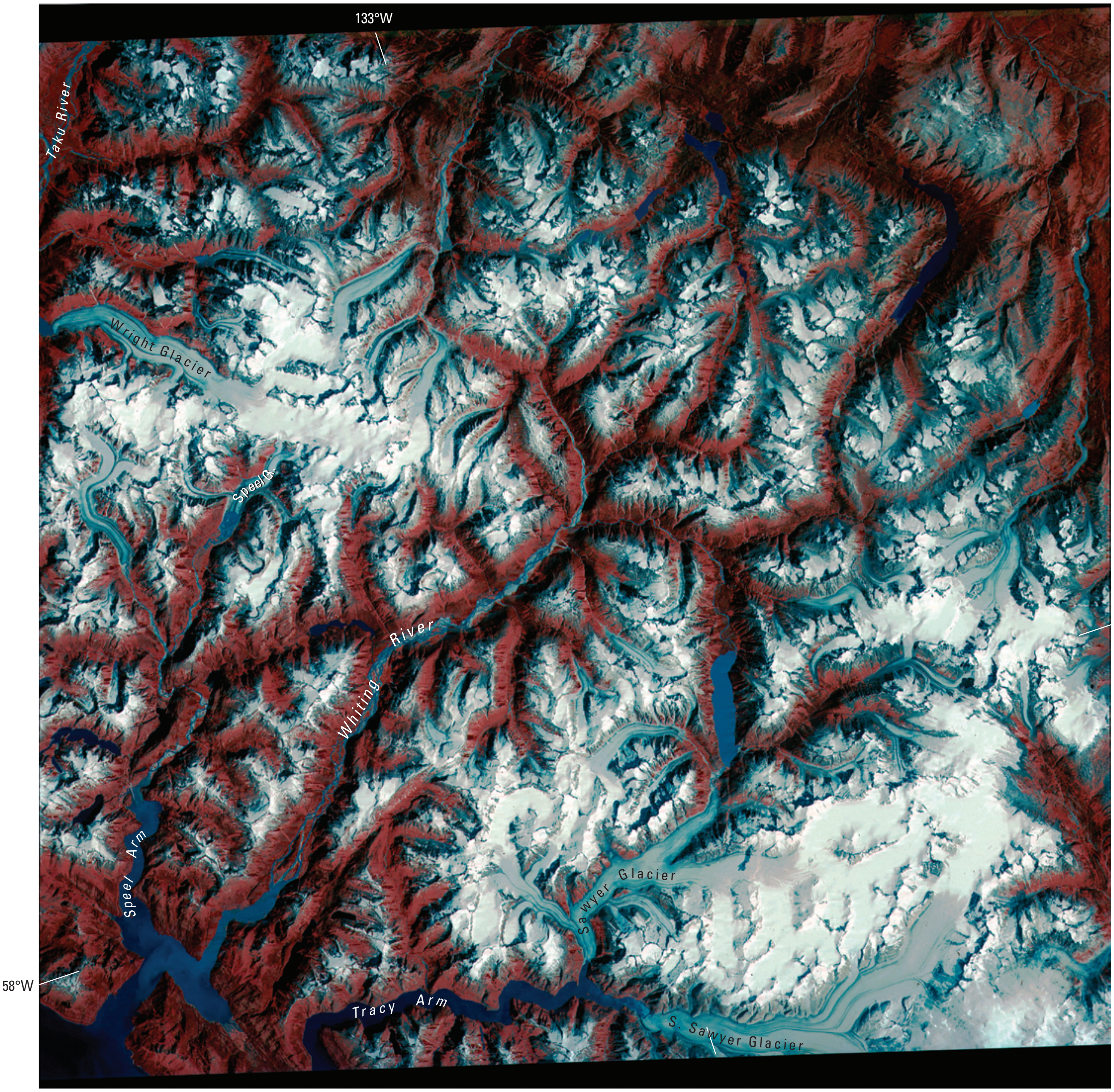


Figure 11.—*<u>Mountain glacier.</u> Landsat 5 TM image of cirque glaciers and valley glaciers in the Coast Mountains, Alaska (for an extended caption see Chapter K of "Satellite Image Atlas of Glaciers of the World," p. K103, fig. 84).*

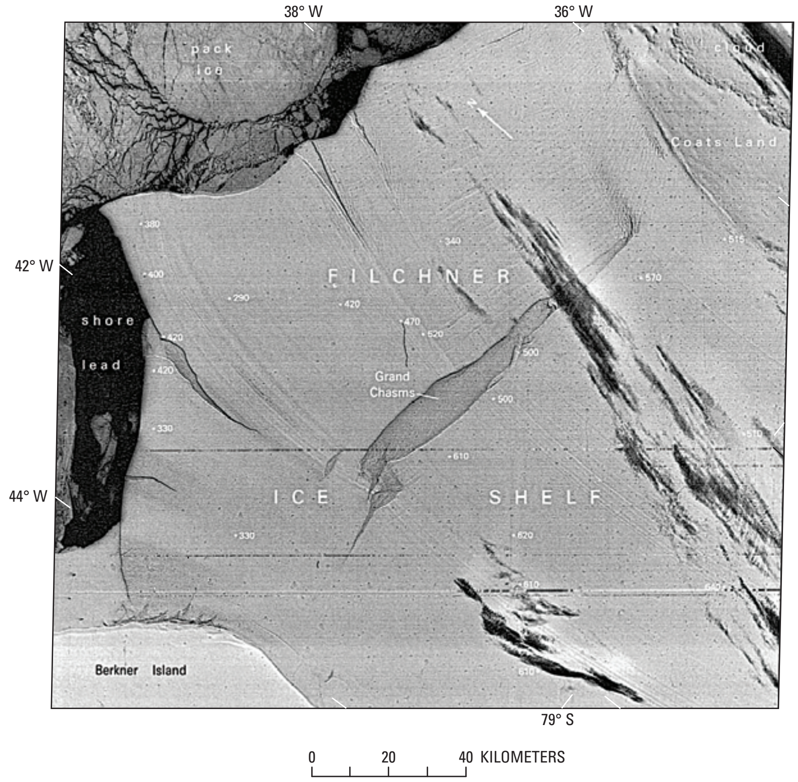


***Figure 12.**—<u>Ice shelf.</u> Landsat 1 MSS image of the Filchner Ice Shelf, Antarctica (for an extended caption see Chapter B of "Satellite Image Atlas of Glaciers of the World," p. B103, fig. 76).*

lite images with higher spatial resolution (from 15-m pixels with ASTER and Landsat 7 ETM+ to 0.60-m pixels with Quick Bird 2) and with the support of glaciologists associated with International Regional Centers of GLIMS, the GLIMS consortium prepared an Illustrated GLIMS Glacier Classification Manual (Rau and others, 2005). Their goal was to provide guidance in classifying glaciers for the GLIMS inventory of non-ice-sheet glaciers by expanding the WGI/WGMS glacier classification system (table 4).

Frank Paul (written commun., 2010) noted that the glaciological community has changed its emphasis from research on glacier hydrology to research on changes in glaciers caused by climate change. As a result of this he points out that a new distinction is now made between ice sheets on one hand and ice caps and glaciers on the other. The following paragraph by Frank Paul (written commun., 2010) provides justification for the shift in definition. The editors have made a few minor clarifications to the paragraph; the clarifications are shown in brackets. In addition, "ice cap" and "ice field" are shown as two words rather than as compound words in the original paragraph.

> This section [by the Swiss glaciologist Frank Paul (written commun., 2009)] provides a short background on the different terminology now used for "perennial surface ice on land," [than what is used] in this book [sic, chapter], in the wider [glaciological] literature, and in global [glacier] monitoring programs. To a large extent, the preference for a specific term can be derived from its historic context so that there is not really a correct or false terminology. Today, the focus of many [glaciological] investigations has changed from hydrology to climate change detection and impact assessment. This [shift in emphasis] required also a revision of the terminology. When talking about perennial surface ice in climate related observing programs, a clear distinction has been made between ice sheets on the one hand and glaciers and ice caps on the

other hand [for example, Intergovernmental Panel on Climate Change (IPCC), 2007a; Eamer, 2007; Integrated Global Observing Strategy (IGOS), 2007]. This is also expressed in the selection of glaciers and ice caps as an essential climate variable (ECV) by Global Climate Observing System (GCOS), 2004. [see also discussion by Zemp and others (2008, p. 10) entitled "Perennial surface ice on land"]. In this context, the term 'glaciers' refers to several types of glaciers ([for example] mountain, valley, cirque, outlet, ice field) as given by the primary classification in (United Nations Educational, Scientific, and Cultural Organization (UNESCO), 1970a) while 'ice cap' refers to dome-shaped ice masses with radial flow. The currently existing two ice sheets of continental size (Greenland and Antarctica) are termed ice sheets to distinguish them from glaciers and ice caps. [Technically speaking, Greenland is the Earth's largest island; physical geographers do not classify it as a continent. Therefore, the Greenland ice sheet is not "continental" in size. However, the definition of "ice sheet" is a mass of ice that covers more than 50,000 km^2 (Armstrong and others, 1973). Ice caps cover less than 50,000 km^2.] This very simple and straight forward differentiation has been made to better distinguish both in terms of their climatic response (decades vs. millennia) and the volume of ice stored. For the data compilation of the World Glacier Inventory, the focus was on land ice as a water resource and a further differentiation [was] not required. Hence, from an historical perspective also, the two ice sheets were termed 'glaciers' (United Nations Educational, Scientific, and Cultural Organization (UNESCO), 1970a) and additional terminology was introduced later to distinguish the ice sheets from the 'other' glaciers. Prominent examples are terms like: 'glaciers other than ice sheets', 'small glaciers', 'mountain glaciers', or 'non-ice sheet glaciers' (as also used in this volume [sic, chapter]). In particular the 'small glaciers' terminology has created confusion as it was impossible to speak about differently sized 'non-ice sheet glaciers' as being large or small (their typical size range spans five orders of magnitude in km^2). The term 'mountain glaciers' is also confusing in this regard as this is also the name of a glacier type in the UNESCO primary classification. The other two terms provide a much clearer separation but still relate to the two ice sheets as being 'glaciers'. For the purpose of this book [sic, chapter] and to be consistent with the terminology used in previous volumes, the term 'non-ice sheet glaciers' is used here for '[mountain] glaciers and [ice fields] and ice caps'. However, it is recommended to use in future publications the terminology from the global observing programs, with ice sheets on the one hand, and glaciers and ice caps on the other hand (for example, Haeberli, 2004). This will also assist in a clearer communication of the science behind this part of the cryosphere to policy makers and the public.

Sensitivity of Glacier Morphology to Changes in Climate

Sugden and John (1976, p. 105) correctly point out that the morphology of glaciers "play[s] an important part in determining how they will respond to climate." An ice cap (fig. 8) is much more sensitive to climate warming than is a valley glacier (fig. 10), because an increase in altitude of the equilibrium line, which separates the accumulation area from the ablation area (fig. 4*A*), reduces the accumulation area on an ice cap more extensively (fig. 13, top) than on a valley glacier (fig. 13, bottom). The ice cap and the valley glacier will both thin, the margin of the ice cap will retreat, and the terminus of the valley glacier will retreat up valley. Sugden and John (1976) also conclude that, if the altitude of the equilibrium line rises above the uppermost surface of the ice cap, the glacier no longer has an accumulation area. In contrast, an accumulation area remains on a valley glacier so long as the altitude of the equilibrium line continues to intersect the glacier, albeit farther upglacier. According to C. Simon L. Ommanney (written commun., 2009), "The classic example of what is being discussed here is the Barnes Ice Cap on Baffin Island. It is a remnant of the last Ice Age and has remained only because the top of it is above the

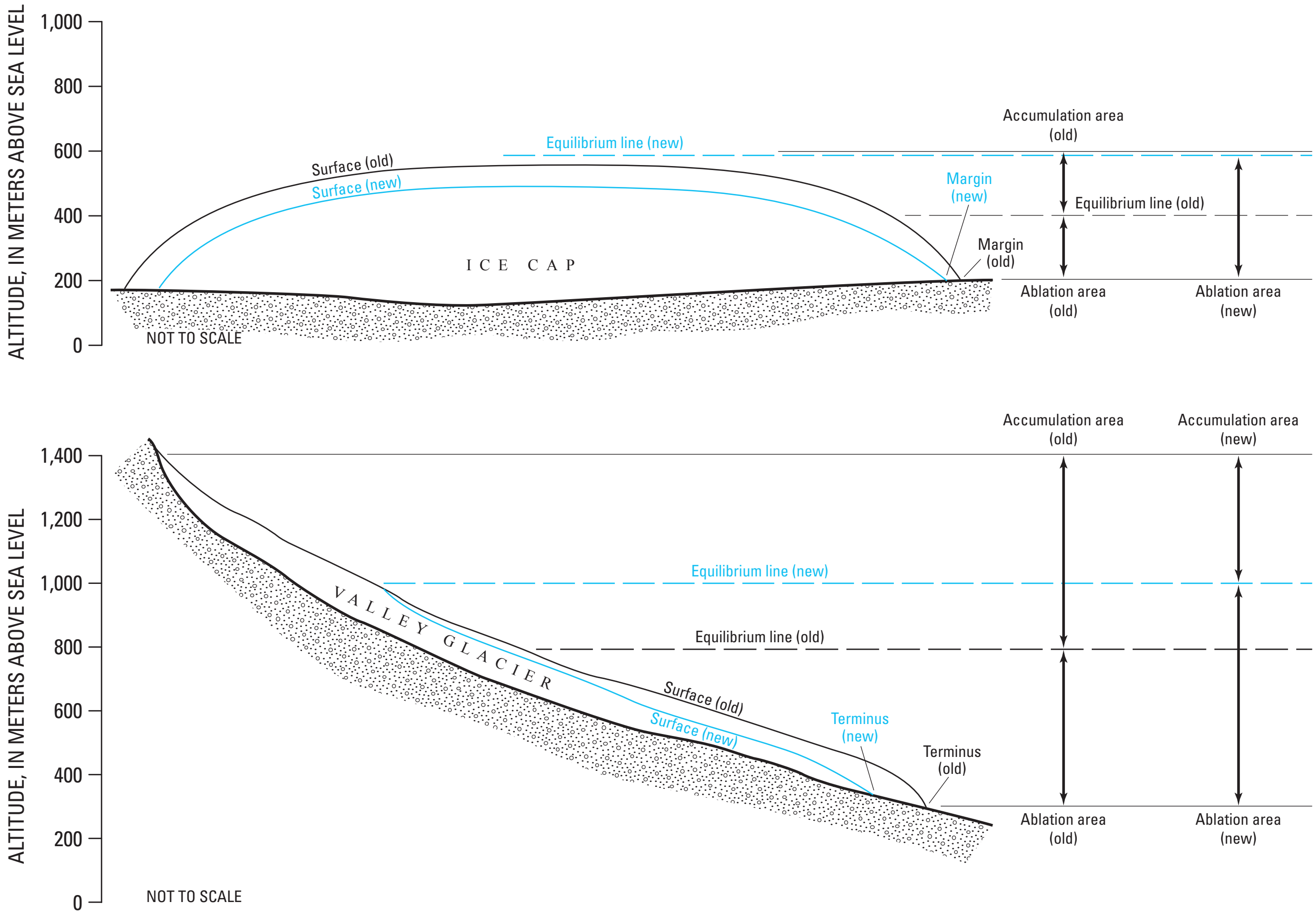


Figure 13.—*Sensitivity of ice caps and valley glaciers to changes in the equilibrium line altitude (ELA) in response to warmer climate as indicated in changes in areas of their respective accumulation and ablation areas. Modified from Sugden and John (1976, p. 105, fig. 63).*

freezing level. It doesn't really have an accumulation area in the traditional sense as its nourishment is through superimposed ice. With rising temperatures it will disappear completely and could then only be restored through the intervention of a new Ice Age. Baird (1952) wrote about this in the early 1950s long before we entered the current period of negative mass balances in the Arctic." (See also Gerald Holdsworth on the "Barnes Ice Cap: Geomorphology and Thermodynamics" in chapter 1386–J of this volume, p. J178–J184.) From field observations of glaciers and meteorological models, Oerlemans and Fortuin (1992) reported the sensitivity of non-ice-sheet glaciers (including small ice caps) to climate warming. They concluded that a temperature increase of 1°C would decrease the overall surface of the 12 glaciers they observed by 0.4 m a^{-1}. Oerlemans (2001, p. 115) also noted that, at about lat 60° N., glaciers in Alaska, Iceland, and Norway have a global mean sensitivity of 50 percent, a "relatively large sensitivity." Dyurgerov (2006, p. 133) concludes "Glacier sensitivity is the key parameter which links changes in climate, glacier volume and the eustatic component of sea-level rise." Ice caps (fig. 8) and ice sheets (fig. 6) are, therefore, especially vulnerable to changes in climate because of their "flattened dome" morphology. Oerlemans (2001, p. 125) reinforced this conclusion: "Glaciers that span a small altitudinal range will disappear more quickly than glaciers that have accumulation zones far above the present day equilibrium line. Flat ice caps, such as those found in Scandinavia, are the most vulnerable," because accumulation areas can enlarge during cooler climate conditions and shrink, or even disappear, during warmer climate conditions. Nesje and others (2008) noted that glaciers in Norway disappeared during the Holocene Epoch.

The morphology of ice caps and ice sheets informs us that these types of glaciers represent ephemeral landforms that add elevation to the landscape. The increase in altitude produces a longer duration of precipitation falling as snow because of the dry adiabatic lapse rate of -9.8°C/1,000 meters (wet adiabatic lapse rate of -5°C/1,000 meters) and greater amounts of precipitation because of the orographic effect of moist air masses forced to rise over regions of higher elevation (fig. 14). Melting glaciers, however, lower the topographic surface. Removal of the glacier ice is followed by isostatic adjustment (rebound) of the Earth's crust subsequent to melting of glacier ice; topographic uplift is therefore the consequence of the lowered elevation of the deglacierized terrain.

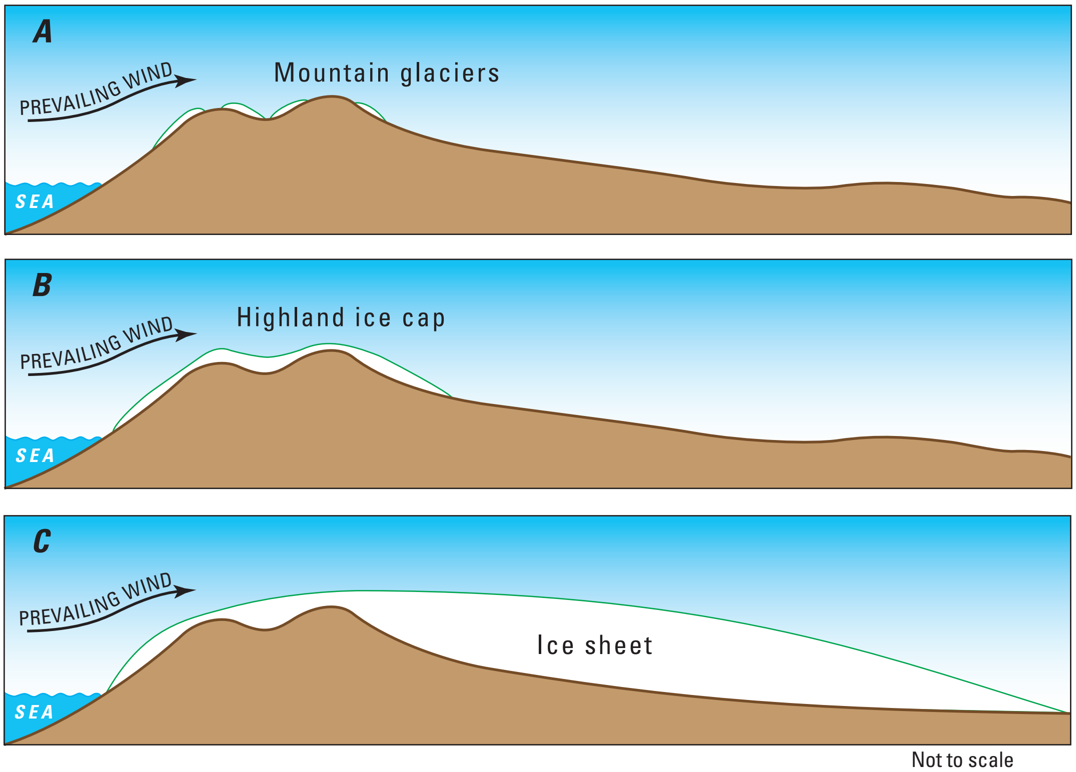


Figure 14.—*Evolution of glacierization of a mountainous area from accumulation of snow and ice during a period of climate cooling.* **A**, *The lowering of equilibrium line altitude (ELA) and concomitant increase in the accumulation area results in the formation of mountain glaciers;* **B**, *The increase in amount of precipitation from orographic uplift of moist maritime air masses leads to the formation of a highland ice cap;* **C**, *An additional expansion of the glacier cover produces an ice sheet. Modified from Flint (1957, p. 317, fig. 18-4; 1971, figure on p. 598).*

Glaciers on Planet Earth

Introduction

The Swiss pastor, Bernard Kuhn, wrote in 1787 that the large boulders that glacier ice transported proved that larger glaciers had existed in the past (Imbrie and Imbrie, 1979). In 1795 the Icelandic scientist Sveinn Pálsson completed the first comprehensive treatise on glaciers (Pálsson, 2004 [1795]). In his introduction, Pálsson (2004, p. 1) referred to the observations of glacierized mountains in Europe that other 18th-century European scientists had made (for example, Grűner, Saussure, Schrank, Walcher, deLuc, Fleischer, Strøm, and Vahl). C. Simon L. Ommanney (written commun., 2009) cited Muraltus (1669) as one of the earliest references to studies of European glaciers, and Ommanney included the monumental four-volume work by Saussure (1779–1796). During the 19th century, geologists and geographers in Switzerland, England, France, Germany, and Italy began the study of glaciers that eventually became the subdiscipline of glaciology (Carozzi, 1984). C. Simon L. Ommanney (written commun., 2009) also cited nine additional studies of glaciers in Switzerland by 19th century Swiss, French, English, and German scientists to provide a fuller overview of the historical development of what eventually would be called glaciology: Agassiz (1840), Charpentier (1841), Martins (1846), Tyndall and Huxley (1857), Ramsay (1860), Pfaff (1876), Schweizerischen Akademie der Naturwissenschaften (SANW) (1881–2002), Trotter (1885), and Bonney (1893). It has always been a mystery that the Swiss Louis Agassiz, who later emigrated to the United States and is considered to be the first scientist to envision a past age of temperate landscapes covered with glacier ice, thousands of meters thick, never traveled to Iceland to directly observe firsthand, as Pálsson did in the 1790s, the large ice caps and associated outlet glaciers that cover about 10 percent of the land area of Iceland. C. Simon L. Ommanney (written commun., 2009) provided references to published works by English, American, Italian, and French scientists, who studied glaciers in other glacierized regions of the Earth: Forbes (1844), Tyndall (1859), Blake (1867), Baretti (1880), Science (1885), Freshfield (1888), Topham (1889), Bonaparte (1890), Geikie (1881), and de Déchy (1892).

For more than 100 years (since 1894; Haeberli and Wallén, 1992), Swiss scientists, with the support of the government of Switzerland and of international scientific organizations, such as the United Nations Scientific and Cultural Organization (UNESCO) and the United Nations Environment Programme (UNEP), provided international leadership in the study of glaciers, working with the International Commission of Snow and Ice (ICSI). Initially (see history of ICSI and its predecessors by Radok, 1997), research was focused on local and regional scales. Later, under the leadership of Fritz Müller, research was expanded to all of the Earth's glaciers with the establishment of the Permanent Service on the Fluctuation of Glaciers (PSFG) in 1967 and by the creation of the Temporary Technical Secretariat (TTS) for the World Glacier Inventory in 1976 in Zürich, Switzerland (Müller and Scherler, 1980). Two international "symposia" have been held since the World Glacier Inventory project was established, to review progress on compilation of the inventory and to publish scientific papers resulting from the "symposia." The first, the International Workshop on the World Glacier Inventory, was sponsored by the International Association of the Hydrological Sciences and held in Riederalp, Switzerland, in September 1978; the proceedings volume was edited by Müller and Scherler (1980). The second "symposium" was sponsored by the International Glaciological Society and held in Lanzhou, China; the proceedings volume, World Glacier Inventory, was edited by Braithwaite and others (2009), and published in the Annals of Glaciology series.

The Permanent Service on the Fluctuation of Glaciers (PSFG) and TTS were combined in 1986 when the World Glacier Monitoring Service (WGMS) was established (Haeberli, 1998) under the leadership of Wilfried Haeberli (Haeberli and Wallén, 1992). Eight volumes of "Fluctuations of Glaciers," each containing worldwide data covering 5-year intervals, have been published: the first four by PSFG (v. 1 (1959–1965), v. 2 (1965–1970), v. 3 (1970–1975), and v. 4 (1975–1980)) and the last four, including the latest in the series, by WGMS (v. 5 (1980–1985), v. 6 (1985–1990), v. 7 (1990–1995), and v. 8 (1995–2000), (Haeberli and others, 2008)).

Leadership provided by Swiss scientists and the Swiss government to the global glaciological community in the late 20th century was in the operation of an international center (WGMS) for the compilation of a global inventory of glaciers and in the archiving and distribution of data on glaciological parameters. British scientists have also provided international leadership in glaciology for more than a century according to C. Simon L. Ommanney (written commun., 2009); British scientists provided leadership

> […] not only through the early work of U.K. citizens on glaciers around the world, but also through the creation of the British Glaciological Society [founded in 1945, renamed the International Glaciological Society in 1962], which put the subject on a professional/scientific footing and through the Journal of Glaciology [volume 1, number 1, first published in 1947] and later through meetings it convened, and on which it published in the Annals of Glaciology [volume 1 published in 1980] that pointed the way ahead. There is also the role of the International Commission on Snow and Ice. Although the decision to create the Commission Internationale des Glaciers [1894], the forerunner of ICSI, was made in Zürich, it was not by the Swiss specifically, but by the International Geological Congress. Although the continental Europeans F.-A. Forel and Prince Roland Bonaparte were involved, it was Captain Marshall Hall of Witshire, U.K., who had worked with Forel [Forel, 1894], who was the inspiration and made the formal proposal. Although Forel was the first President, it then went in succession to Austria, Germany, the USA, and Austria. Details are in Uwe Radok's history of ICSI [Radok, 1997]. The history of the first 50 years of the IGS has been written by Peter Wood [Wood, 1986].

For more than a century, therefore, glaciers have been the focus of research by scientists, especially the measurement of fluctuations in area (for more than 100 years), in mass balance (for more than 50 years), and in the changes in the position of termini (for more than 100 years) during changes in climate and in the erosional and depositional landforms these changes leave on the landscape (Sugden and John, 1976; Hambrey and Alean, 2004). Glaciers have also been a subject for photographers, who often combine scientific documentation with the recording of aesthetic beauty. Shaler and Davis (1881) were the first scientists to use ground photographs to document glaciers and glacier morphology. Oblique aerial photographs of glaciers by Austin Post (Post and LaChapelle, 2000), Bradford Washburn, and Robert F. Krimmel famously combine scientific documentation and artistic qualities; the Glaciers of Alaska Chapter (1386–K) of this volume contains many of their photographs. Pfeffer (2007) used photographs to sequentially document the retreat of the Columbia Glacier, a tidewater glacier in Alaska. Oblique aerial photographs can be combined with satellite images and ground photographs can be used to produce a comprehensive "preliminary inventory"of glaciers in a specific region and, with maps, can document precise geographic locations of glaciers, such as in Iceland (Sigurðsson and Williams, 2008). The chapters for this volume (1386-B–K) generally use the same combination of images, often supported with direct ground measurements, to document change in glaciers over time. Glaciers have also attracted the attention of geophysicists who study the processes that govern the flow of glacier ice (Nye, 1952, 1959; Glen, 1958, 1974; Scheidegger, 1970; Lliboutry, 1971; Paterson, 1994; Hooke, 2005), including surge-type glaciers (Nye, 1958; Kamb and others, 1985).

Total Area of the Earth Covered by Glaciers

To accurately determine the total area of the Earth covered by glaciers has been a goal of geologists and geographers for more than a century. Three major problems have prevented achieving that goal: availability of accurate data; difficulty of discriminating between glacier ice and snow pack, especially when snow pack overlies a glacier margin or when glacier ice is blanketed by supraglacier debris; and having only a restricted period of time (for example, less than a decade) during which to determine a glacier's area because regional and global atmospheric warming or cooling can result in a rapid reduction or increase in total area, especially of non-ice-sheet glaciers. Until remotely sensed data from satellites became available, thereby helping to solve the first and third problems, the primary source of data for glaciologists was map coverage, if it existed. The second problem, difficulty of discriminating glacier ice from snowpack, affected the accuracy of these maps, which were compiled by plane-table or by aerial photogrammetric mapping techniques.

The compilations of the Earth's glacierized area by Hess (1933), Thorarinsson (1940), Flint (1957, 1971), Thiel (1962), Barry (1985), Haeberli and others (1989a), and Ohmura (2009) (table 7) are primarily based on conventional cartographic data. Ohmura (2009) used data archived at both WGMS and at the NSIDC, which include glacier inventory data derived from analysis of Advanced Spaceborne Thermal Emission and reflec-

Table 7.—*Compilation of total area of Earth's glacierized regions, from various sources 1933–2009*

[Area is in square kilometers]

Source	Area
Hess (1933)	15,363,060
Thorarinsson (1940)	15,098,775
Thiel (1962)	14,270,000
Flint (1971), Sugden and John (1976)	14,898,320
Barry (1985)	14,440,000
Haeberli and others (1989a)	15,861,776
Williams and Hall (1993)	15,812,780
Ohmura (2009)	14,550,618
Williams and Ferrigno (table 1, Part A–2 of this chapter)	15,926,087

tion Radiometer (ASTER) images by glaciologists working with the Global Land Ice Measurements from Space (GLIMS) initiative. Dyurgerov and Carter (2004) provided data on glacier area in the pan-Arctic, not including the Greenland ice sheet. The accurate regional compilations by the WGMS (Zürich) are primarily based on available maps, supplemented by aerial photographs and satellite images (Haeberli, Müller, and others, 1989). The WGMS also focuses on recording fluctuations of non-ice-sheet glaciers, emphasizing the quantitative measurement of changes in terminus and in mass balance. The most recent WGMS summary is found in Global Glacier Changes: Facts and Figures (Zemp and others, 2008). In July 1972 medium-resolution Landsat images of most of the glacierized regions of Earth became available, forming the basis for the multinational 11-chapter Satellite Image Atlas of Glaciers of the World (U.S. Geological Survey Professional Paper 1386-A–K), a dataset of worldwide glacier area from Landsat MSS images and from ancillary data for the baseline period 1972–1981, an interval of time during which the global climate was cooler than it had been before as well as cooler than it had been after that period.

Completing the 10 regional chapters of this volume (1386-B–K) enabled us to compile the areal extent of glacier ice in each of the 10 glacierized regions and for the entire Earth (table 1). Our conclusion is that 15,926,087 km^2 of the surface of the Earth is covered by glaciers. The total glacier area is not a "snapshot" at a single point in time; rather we have compiled it from the areas of glacierized regions given in each of the 10 chapters, supplemented by data from other reliable sources. For example, the area (and volume) of ice in the Antarctic ice sheet is from the research by Drewry and others (1982) and Drewry (1983). Table 1 also provides the data on regional glacier areas that Flint (1971) compiled (also cited by Sugden and John (1976)) with our compilation from the 10 chapters of this volume and our ancillary data. Table 1 also includes Ohmura's (2009) compilation of glacier data from the 10 chapters so that readers may compare Ohmura's (2009) compilation with that of Flint (1971) (also cited by Sugden and John (1976)) with our compilations. Comparing the regional subtotals of Flint (1971) (also cited by Sugden and John (1976)) with those compiled by Ohmura (2009) and with those in this chapter reveals differences in glacier area. More recent data on glacier area in some regions complicate the issue and will be discussed in the context of changes resulting from more than a decade of global warming of the atmosphere, land, and oceans.

South Polar Region

Antarctic Ice Sheet (Chapter B of this volume; Swithinbank, 1988)

Antarctica has the highest average elevation (2,440 m) of all of the Earth's continents, and it is the coldest continent. The coldest temperature ever recorded on the Earth's surface was at Vostok Station on 21 July 1983 at 2,439m; it was -89.6°C (Australian Geographic, 1994). Drewry and others (1982) and Drewry (1983) of the British Antarctic Survey compiled the most accurate information to date on the area (13,586,400 km^2; table 1) and volume (30,109,800 km^3; tables 2, 3) of the Antarctic ice sheet from their airborne radio-echosounding surveys. Lythe and others (2001) published a new subglacial topographic model of Antarctia that allows more accurate calculation of the thickness of the Antarcic ice sheet.

Only about 2 percent of Antarctica is ice free (Drewry and others, 1982; Drewry, 1983); the other 98 percent is covered by an ice sheet that flows into the Southern Ocean through fringing ice shelves, ice streams, and outlet glaciers. When the glacier ice reaches the ocean, it becomes afloat at its grounding line, forming an ice front. If it stays grounded, it forms an ice wall (Swithinbank, 1988). Glaciologists divide the Antarctic ice sheet into two parts, separated by the Transantarctic Mountains. The part of the ice sheet in East Antarctica is the larger of the two, with an area of 10,135,170 km^2 and a volume of 3,262,000 km^3 according to Drewry and others (1982) and Drewry (1983); it is approximately five times larger and has eight times more volume than the part of the ice sheet in West Antarctica (table 2). Large ice shelves, into which flow numerous outlet glaciers and ice streams, are in Antarctica; one large ice shelf is located in West Antarctica, the Ross Ice Shelf (Ferrigno and others, 2007), and two large ice shelves are located in East Antarctica: the Ronne-Filchner Ice Shelves (Ferrigno and others, 2005) and the Amery Ice Shelf (Brooks and others, 1983; Foley and others, in press). The two largest glaciers in the world flow into two of the ice shelves. The 400-km-long Byrd Glacier, an outlet glacier, flows into the Ross Ice Shelf (Ferrigno and others, 2007) through the Transantarctic Mountains; it drains an area of more than 1,000,000 km^2 (Swithinbank, 1988, p. B1). The approximately 1,000-km-long Lambert Glacier drains into the complex Amery Ice Shelf system (Foley and others, in press).

Until the launch of the ERTS-1 (Landsat-1) Earth-orbiting satellite on 23 July 1972, cartographers were able to map Antarctica only by relying on ground-based surveying methods and on aerial photogrammetric techniques, which themselves relied on image-identifiable features on the ground to provide precise geographic location. Conventional ground surveys had previously provided such positioning, and they continued to provide needed ground control for aerial photogrammetry until replaced by accurate positioning made possible by global positioning satellites (GPS). The database of Landsat MSS, of return beam vidicon (RBV), and of thematic mapper (TM) images of Antarctica north of about lat 81° S. gradually increased to the point of being extremely useful for regional glaciological studies and cartographic projects. Williams and Ferrigno (1988a, b) evaluated and compiled tables of optimum Landsat 1, 2, and 3 MSS and Landsat 2 RBV images of Antarctica acquired between 1972 and 1982. Of the 10,000 images we examined directly in Goddard Space Flight Center archives, 45 percent of the 2,514 nominal scene centers of images were cloudfree or had <10 percent cloud cover. Landsat imaging sensors and subsequent imaging sensors (for example, RADARSAT) provided glaciologists with data to compile accurate planimetric and topographic maps of Antarctica and to document, in a time-lapse manner, changes occurring in ice-sheet elevation and ice-sheet area, with particular focus on changes in the floating (ice fronts) and grounded (ice wall) cryospheric margin of Antarctica (see, for example, Ferrigno and others, 2008). Successive satellite images have shown that ice fronts, especially the margins of ice shelves, have exhibited the greatest changes in area between acquisitions of satellite images; changes in position of ice walls appear to be very slow.

Several maps of Antarctica have been compiled by mosaicking satellite images. The first satellite image map of Antarctica to be compiled was the NOAA Advanced Very High Resolution Radiometer (AVHRR) image mosaic map (U.S. Geological Survey, 1991; Ferrigno and others, 1996, 2000) and a RADARSAT image mosaic map (U.S. Geological Survey, 2003). The National

Geographic Society (2001) used the RADARSAT image mosaic map as the base for its latest map of Antarctica. From a cartographic perspective, the 2003 RADARSAT map is the most accurate small-scale map ever compiled of Antarctica because the images used in the compilation were acquired during a short period: 19 September to 14 October 1997, almost a "snapshot." Robert Bindschadler, a glaciologist at NASA's Goddard Space Flight Center and leader of the West Antarctic Ice Shelf Initiative (WAIS), directed a project to obtain Landsat Enhanced Thematic Mapper (ETM+) images of Antarctica north of about lat 81° S. The project produced a new image dataset with excellent geometric accuracy. The dataset, which contains about 1,100 mosaicked images acquired between 1999 and 2001, is known as the Landsat Image Mosaic of Antarctica (LIMA). Glaciologists and cartographers use LIMA to map changes in the cryospheric coast of Antarctica [*http://lima.usgs.gov*].

In collaboration with Scott Polar Research Institute (SPRI) (Cambridge, England, U.K.), the U.S. Geological Survey (USGS) led an international collabortive effort to define the cryospheric coast of Antarctica. In this project, "Coastal-Change and Glaciological Maps of Antarctica," Charles Swithinbank of SPRI used Landsat MSS and TM images to map changes in the dynamic coast of Antarctica (Williams and Ferrigno, 2005) [*http://pubs.usgs.gov/fs/2005/3055/*]. Later maps in the USGS I-2600 map series also used Landsat ETM+, RADARSAT, Corona, and MODerate-resolution Imaging Spectroradiometer (MODIS) images to document changes.

Of the 24 maps initially planned by the international collaboration, 9 have been published in two formats, printed maps with analysis of changes in an "accompanying pamphlet," and online ArcGIS datasets. Three additional maps (Amery Ice Shelf [I-2600-Q] and Drygalski Ice Tongue/Oates Coast [I-2600-J, -K] (Foley and others, in press) are in production by the USGS. The third map of the entire cryospheric coast of Antartica will be published online only, with changes in the cryospheric coast of Antarctica available online in digital (ArcGIS) format (Ferrigno and others, in press). The 9 conventional maps and accompanying pamphlets (Swithinbank and others, 1997, 2003a,b, and 2004; Ferrigno and others, 2005, 2006, 2007, 2008, 2009, 2010) include three maps (I-2600-A, -B, and -C) of the Antarctic Peninsula showing changes in its cryospheric coast compiled from satellite images, historic and modern maps, oblique and vertical aerial photographs, and field survey datasets; the three maps (Ferrigno and others, 2006, 2008, 2009) resulted from a long-term collaboration between the USGS's Glacier Studies Project (GSP) and the British Antarctic Survey's (BAS) Mapping and Geographic Information Centre (MAGIC) (Ferrigno and others, 2002) [*http://pubs.usgs.gov/fs/fs17-02/*], with the Landsat TM image mosaic map of the Antarctic Peninsula provided by the German Institut für Angewandte Geodäsie (IfAG) of the Bundesamt für Kartographie und Geodäsie.

The Antarctic Peninsula was chosen as the focus for the USGS, BAS, and IfAG (BKG) collaboration because the region has experienced significant climate warming (+3°C) during the past six decades. As a result, some of its ice shelves have disappeared and others are experiencing a rapid retreat of their ice fronts (Rau and others, 2004). Ice shelves, one of the seven types of glaciers (fig. 12, tables 4–6), appear to be very sensitive to global climate warming from increases in atmospheric temperatures and in the temperatures of ocean currents. Alley and others (2008) have developed a simple law in terms of which ice-shelf calving (spalling off of ice from the near-vertical terminus of an ice shelf) forms icebergs.

Glaciers of the Subantarctic Islands

By RICHARD S. WILLIAMS, JR.

When Swithinbank (1988) wrote the first chapter (Antarctica, 1386–B) of this volume, no usable Landsat images during the "baseline period" (1971–1982) of most of the subantarctic islands were available. Persistent cloud cover and snow-covered terrain prevented acquiring such images, as did the fact that acquiring the images required onboard recording capability because satellite receiving stations were absent in the Southern Hemisphere, and that need had to compete with other regions that required images. Even today (2009), persistent cloud cover prevents many Earth-orbiting satellites from acquiring images of these remote islands. Because snow obscures the landscape, including the margins of glaciers, seasonal snow cover can further reduce the usability of satellite images. Therefore, information on the glaciers of the subantarctic islands was compiled from various sources for inclusion in part A-2 (Glaciers) of this chapter in order to supply the missing information.

According to Mercer (1967, p. 243), "The Sub-Antarctic islands may be defined as those lying between the southern [Hemisphere] limit of trees and the Antarctic continental shelf." The subantarctic islands range in size from 0.4 km^2 (Scott Island) to 6,675 km^2 (Île Kerguelen). The islands are either individual or part of an archipelago; only glacierized subantarctic islands are included in table 8, even if the glacierized area is small compared to the island's total area.

After 1982, subsequent to the Landsat image "baseline period," images of the subantarctic islands were acquired by various satellites, such as Satellite Pour l'Observation de la Terre (SPOT) and Envisat, among others. Figure 15 is an Aqua MODIS color image of Île Kerguelen showing the Cook Glacier, a 500-km^2 ice cap with 12 named outlet glaciers, visible on the western side of the island. Figure 16 shows a Terra MODIS image of snow-covered South Georgia Island. Figure 17, from Mercer (1967), clearly shows the glacierized area of both islands.

Because there are so few usable satellite images of the subantarctic islands showing minimum snow cover and without cloud cover, various ancillary sources of published and web-accessible information were evaluated to determine the area of each island and its glacierized extent (table 8).

Although it was difficult to determine the accuracy of much of the source material, the information judged most reliable from these sources formed the basis for determining the glacierized area of each subantarctic island or island archipelago. Two scientific papers on the area of glaciers on Heard Island (Ruddell, 2006; Thost and Truffer, 2008) are considered to be reliable sources. Sources analyzed include the following: Mercer (1967), Flint (1971), Sugden and John (1976), Anonymous (1978), LeMasurier and Thomson (1990), Alberts (1995), Rubin (1996), Ruddell (2006), and Thost and Truffer (2008).

The American Geographical Society's "Southern Hemisphere Glacier Atlas" by Mercer (1967) contains general descriptions of the geography and geology of all the subantarctic islands and two map plates. Map 18, "Antarctica and Sub-Antarctic," includes inset maps of the subantarctic islands: Bouvetøya (with two glacier place-names), South Sandwich Islands (with 11 named islands in the archipelago), and Heard Island (with 8 named glaciers from

Table 8.—*Glacier area of the glacierized subantarctic islands and archipelagos*

[Area is in square kilometers; —, not shown; ~, estimate; bold font indicates total for Balleny Islands and for the entire subantarctic islands and archipelagos]

Island or archipelago	Area of island or archipelago	Glacierized area	Latitude[1]	Longitude[1]
Balleny Islands	**505**	**480**	66°55´S.	163°20´E.
Buckle Island	[2]101	96	66°50´S.	163°12´E.
Sturge Island	[2]207	[3]197	67°28´S.	164°38´E.
Young Island	[2]197	[3]187	66°25´S.	162°24´E.
Bouvetøya	[3]54	[3]50	54°26´S.	3°24´E.
Heard Island	[3]390	[3]257	53°06´S.	73°30´E.
Île Kerguelen (Grand Terre)	—	~400	48°58´S.	68°72´E.
Îles Kerguelen	[3]7,215	—	49°73´S.	70°58´E.
Peter Island	[3]158	[3]150	68°47´S.	90°35´W.
Scott Island	[3]0.4	~0.3	67°40´S.	179°92´W.
South Georgia Island/ South Georgia Group	[3]3,755	[3]2,225	53°50´S. – 55°00´S.	35°50´W. – 38°67´W.
South Sandwich Islands	[3]310	[3]240	57°45´S.	26°30´W.
South Orkney Islands	[3]622	[3]530	60°35´S.	45°30´W.
South Shetland Islands	[3]3,688	[3]2,950	62°00´S.	58°00´W.
Adelaide Island	[2]3,888	[2]3,888	67°15´S.	68°30´W.
Anvers Island	~2,800	~2,500	64°33´S.	63°25´W.
Brabant Island	[2]1,369	[2]1,369	64°15´S.	62°20´W.
Hoseason Island	[2]47	[2]47	63°44´S.	61°41´W.
Trinity Island	[2]233	[2]233	63°45´S.	60°44´W.
Total glacier area	—	**15,419**	—	—

[1]All geographic coordinates, except those of Îsles Kerguelen, Scott Island, and South Georgia Island, are from Alberts (1995); geographic coordinates of Îsles Kerguelen, Scott Island, and South Georgia Island are from Rubin (1996).

[2]From Alberts (1995).

[3]From Rubin (1996).

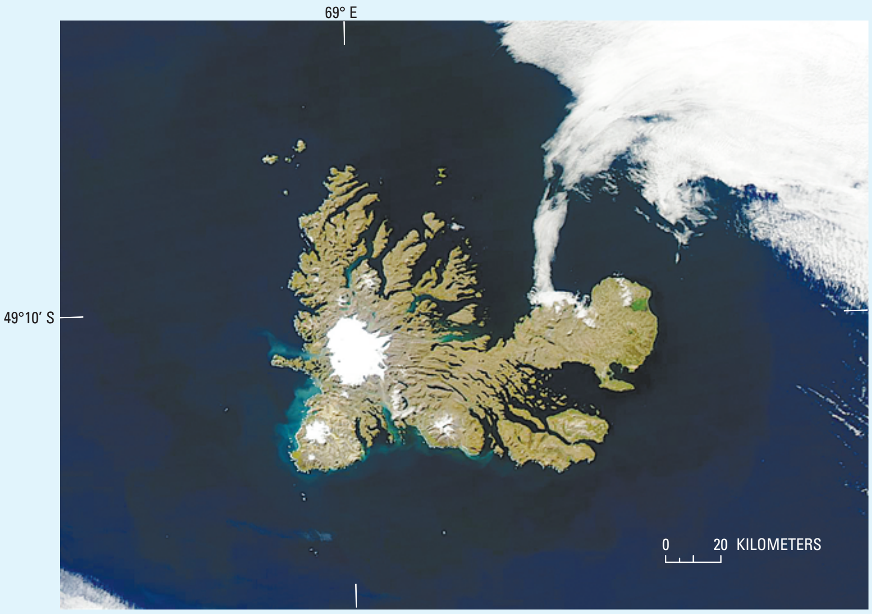


Figure 15.*—MODerate-resolution Imaging Spectroradiometer (MODIS) image of Île Kerguelen (Îles Kerguelen) on 4 March 2004. Cook Glacier (ice cap) and its outlet glaciers (12 named) are visible, as are glaciers (5 named) on three peninsulas: Presqu' île de la Société de Géographie (north of Cook Glacier), Presqu' île Rallier du Baty (southwest of Cook Glacier), and Presqu' île Gallieni (southeast of Cook Glacier). NASA Aqua MODIS image.* [http://visibleearth.nasa.gov; *KerguelenIslands; File-Kerguelen. A2004064.0945.250m.jpg]*

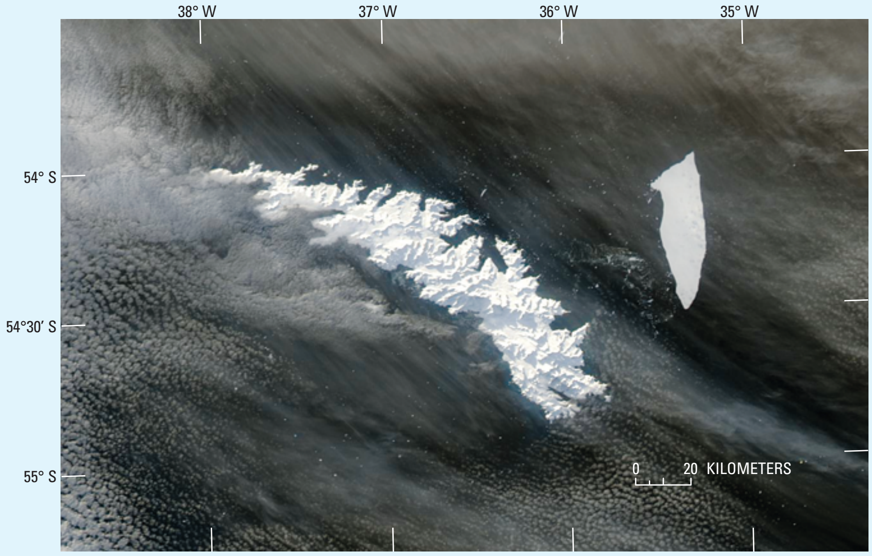


Figure 16.*—MODerate-resolution Imaging Spectroradiometer (MODIS) image of South Georgia Island on 15 July 2004. The island is completely covered with snow, masking the 43 percent of the island not covered with glaciers (see fig. 17). The A-38B tabular iceberg and associated smaller icebergs can be seen east of the island. The "parent" A-38 iceberg (140 km long, 40 km wide; Ferrigno and others, 2005) calved from the Ronne Ice Shelf, East-Antarctica in early October 1998, before breaking into two tabular icebergs (A-38A and A-38B) on 22 October 1998 (Ferrigno and others, 2005). During the 6-year interval between the initial calving event and the date of this MODIS image, the A-38B iceberg has been reduced from its original dimension to about 60 km long, 15 km wide. MODIS image from the Terra satellite of the National Aeronautics and Space Administration. NASA Terra MODIS image at* http://veimages.gsfc.nasa.gov/6835/SouthGeorgia.A2004197.1635.250m.jpg.

the ice field). Map 18.2 (fig. 17) includes maps of the "Glaciers of Kerguelen" (including Cook Glacier, an ice cap with 12 named glaciers, and three glacierized peninsulas with 5 named glaciers), and the "Glaciers of South Georgia" (with 15 named glaciers). "Geographic Names of the Antarctic," by Alberts (1995, 2d ed.), now supplemented by a periodically updated web-accessible database at *http://geonames.usgs.gov/antarctic/index.html*, provides brief descriptions of each island and archipelago, except for Île Kerguelen. Berthier and others (2009) discuss changes in the glaciers on the Kerguelen Islands during 1963 through 2006. The Polar Regions Atlas of the U.S. Central Intelligence Agency (Central Intelligence Agency, 1978) was also consulted, but it lacked specificity. The most informative source turned out to be the Lonely Planet booklet on Antarctica (Rubin, 1996), which provides excellent narrative descriptions of each of the subantarctic islands. Drawing from a number of uncited sources, Rubin (1996) provides area data for most of the islands as well as the percentage of each island that is glacierized.

All the islands and archipelagos are volcanic in origin, and some have active volcanoes. LeMasurier and Thomson (1990) is the best source for information on the volcanic geology and geologic history of the islands. The second edition of the Smithsonian Institution's "Volcanoes of the World" (Simkin and

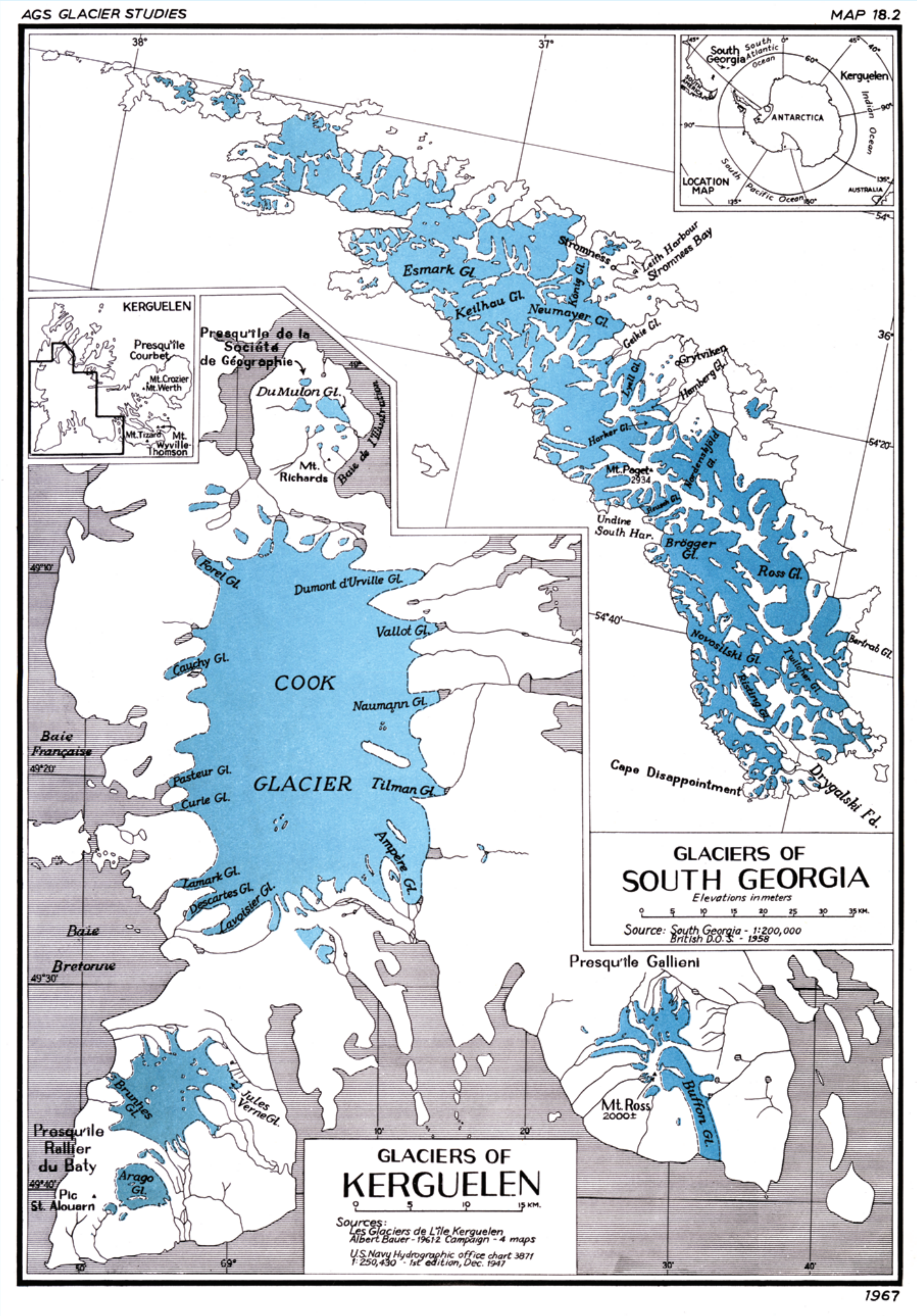


__Figure 17.__—Maps of the Glaciers of Kerguelen and of the Glaciers of South Georgia showing glaciers on the two subantarctic islands. Map plate from the American Geographical Society Glacier Studies (Mercer (1967, p. 325, Map 18.2©)).Used with permission of the American Geographical Society.

Siebert, 1994) is the best recent source; it is updated monthly by the "Bulletin of the Global Volcanism Network," published by the Smithsonian National Museum of Natural History. Simkin and Siebert (1994, p. 160–161) discuss historic eruptions on Buckle Island of the Balleny Islands archipelago and on six islands in the South Sandwich Islands archipelago; Deception Island, a partially submerged volcanic caldera in the latter archipelago, is second only to Mount Erebus on Ross Island, with regard to frequency of historic volcanic activity. The glacier-covered active volcanoes on some of the subantarctic islands are of scientific interest because the interaction of volcanic activity with glaciers can cause glacier outburst floods (jökulhlaups) (see section on Jökulhlaups in Glaciological Hazards: Global, Regional, and Local, on p. A266–A273, of this chapter).

Prior to the present chapter's compilation, Flint (1971) and Sugden and John (1976) estimated the area of glaciers of the glacierized subantarctic islands as being 3,000 km^2; Ohmura (2009) estimated the area as 7,000 km^2. Table 8, compiled from the various sources indicated, shows the best current (2012) estimate to be 15,419 km^2. The area of subantarctic islands includes many of the islands to the west and northwest of the Antarctic Peninsula (Ferrigno and others, 2006, 2008, 2009) that are not connected to the peninsula. Drewry and others (1982, p. 88) stated that their compilation of glacier area for Antarctica does not include non-conterminous islands (for example, islands not connected by ice shelves to the mainland). Drewry and others (1982, p. 88) did not include the South Shetland Islands in their compilation because they were not "attached" to the continent. The data in table 8 are provisional totals, subject to revision when specific criteria are met: the images are georeferenced; the images are usable (minimum snow cover, cloud free); and the spatial resolution of the images is that of Landsat ETM+ panchromatic images (15-m picture element (pixel)) or higher. Images from SPOT (10-m pixels), Ikonos (1-m pixels), or Quick Bird 2 (0.62 pixels) meet the spatial resolution criteria, but until they become available for each glacierized island, glacier areas cannot be measured accurately.

The remoteness of the subantarctic islands has limited field observations of changes in these glaciers; persistent cloud cover and snow cover have reduced the opportunities for acquiring usable satellite images. However, Zemp and others (2008, p. 46–47), using WGMS data, summarized what is known about South Georgia Island: during the 20th century, glaciers receded from their more advanced positions at the end of the 19th century, fluctuating until the 1980s. Most glaciers have been in retreat since the 1980s. Except for advances of some glaciers during the 1960s, glacier recession has occurred on Heard Island during the latter half of the 20th century. Thost and Truffer (2008) report that the glacier area was reduced about 29 percent from 1947 to 2003. Donoghue and Allison (2008) have also reported on retreating glaciers in the subantarctic.

The area covered by the Antarctic ice sheet (13,586,400 km^2) represents nearly 85 percent of the total area of glacier ice on Earth, and its volume (30,109,800 km^3) represents nearly 91.5 percent of the total volume of glacier ice on Earth (table 3). Because more than 70 m of the maximum potential rise in sea level is sequestered in the ice sheet, any accelerated reduction in volume of ice in the Antarctic ice sheet would have global ramifications. By contrast, accelerated melting of the Earth's non-ice-sheet glaciers would produce approximately 0.5 m of rise in sea level, even if all were totally melted (table 3).

In the late 1970s, Mercer (1978) set forth a hypothesis that an increase in CO_2 in the Earth's atmosphere would lead to climate warming, which, in turn could cause a decrease in the volume of glacier ice in the region covered by the Ross Ice Shelf and the Ronne-Filchner Ice Shelves of West Antarctica. If the ice shelves and glacier ice in the divide between them completely disappeared, an ocean channel would form between the Transantarctic Mountains on the east and the group of subglacier volcanic mountains on the west. If the ice shelves, which "buttress" and "impede" the flow of outlet glaciers from East Antarctica and ice streams from the west were removed, the drawdown of glacier ice above sea level would result in a rise of global sea level of about 6 m. Bamber and others (2009) concluded that a rapid collapse of the West Antarctic part of the Antarctic ice sheet would raise global sea level 3.3 m; they also supported Mercer's 1978 hypothesis that the West Antarctic part of the Antarctic ice sheet is inherently unstable. Mitrovica and others (2009) discussed the regional impact of the global rise in sea level resulting from collapse of the West Antarctic part of the Antarctic ice sheet. The Fourth Assessment Report of the Intergovernmental Panel on Climate Change (IPCC) (2007a) estimated a sea-level rise of ~5 m from collapse of the West Antarctic part of the Antarctic ice sheet. Mitrovica and others (2009) showed that the rise in sea level would not be uniform globally; coastal locations in eastern North America, for example, would be ~30 percent higher than the projected 5 m.

In 1994, the West Antarctic Ice Sheet (WAIS) Initiative [*http://neptune.gsfc.nasa.gov/wais/*] was begun by NASA and the National Science Foundation as a multidisciplinary study to address the concern of rapid climate change and future sea level, using satellite technology and field-work techniques, for example, by using the Global Positioning System (GPS) of satellites to make precise geodetic measurements of horizontal and vertical change. The first WAIS Workshop was held in 1996.

In the early-to-mid 1990s, the USGS initiated a follow-on study, "Coastal-Change and Glaciological Maps of Antarctica," after the "Antarctica" chapter, 1386–B (Swithinbank, 1988), in this volume was completed. The primary emphasis of the study was on publishing coastal-change maps of West Antarctica [*http://pubs.usgs.gov/fs/2005/3055/*]. Satellite images could measure changes in the fringing ice shelves of the cryospheric coast[3] of Antarctica, especially those on the Antarctic Peninsula. Cook and others (2005) documented the successive retreat of the outlet glaciers and the ice shelves on either side of the Antarctic Peninsula. The British Antarctic Survey has observed a significant increase in temperature during the past 50 years, with a +3°C increase measured on the western side at its stations on the peninsula (British Antarctic Survey, 2007). Several ice shelves have completely disappeared during the past two decades, including the Jones,

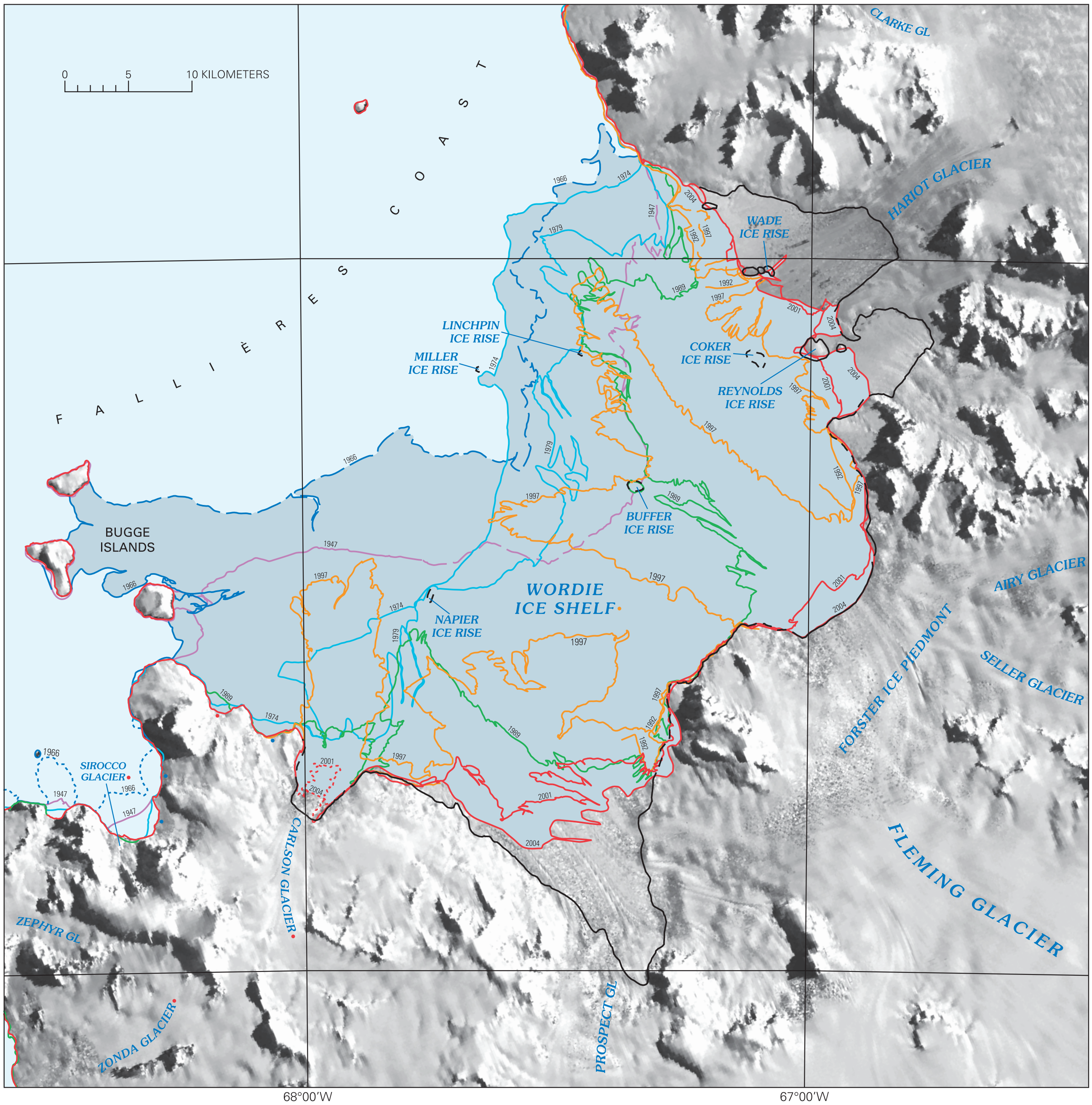
[3] Cryospheric coast refers to coasts (especially in Antarctica and Greenland) where outlet glaciers and ice streams flow directly into the ocean, ice shelves, or fjords. The terminus of such "tidewater" glaciers may be grounded (ice wall) or floating (ice front). In either case, ice is lost primarily by calving into the ocean.

Larsen A, and Müller Ice Shelves and Wordie Ice Shelf (fig. 18) (Ferrigno and others, 2006, 2008, 2009; Brown, 2008). The Larsen B part of the Larsen Ice Shelf is also beginning to break up (Brown, 2008; Ferrigno and others, 2008; National Snow and Ice Data Center [NSIDC], 2008). The Wilkins Ice Shelf on Alexander Island is also breaking up (Ferrigno and others, 2009).

Significant changes have been occurring in the fringing ice shelves of Antarctica, especially those on the Antarctic Peninsula. Of the 244 ice fronts of glaciers measured on the peninsula and associated islands during a period of 60 years (from about 1940 to about 2000), 87 percent have retreated, and the pattern of the retreating ice fronts shows clearly that the retreat is affecting glaciers located farther south (Cook and others, 2005; Ferrigno and others, 2006, 2008, 2009). The Wilkins Ice Shelf, for example, lies mostly south of lat 70° S. (Brown, 2008; Ferrigno and others, 2009) and is breaking up quickly (National Snow and Ice Data Center [NSIDC], 2008). Also, Mercer's hypothesis of accelerated retreat of outlet glaciers and ice streams following the removal of ice shelves has been confirmed on the Antarctic Peninsula (De Angelis and Skvarca, 2003; Rignot, Cassasa, and others, 2004; Rignot and others, 2005; Pritchard and Vaughan, 2007). The outlet glaciers that the ice shelves had previously buttressed have now become tidewater glaciers. Alley, Fahnestock, and Joughin (2008) addressed the impact of loss of ice shelves in Antarctica and the resulting acceleration of flow of ice streams and outlet glaciers into the ocean. Oppenheimer and Alley (2004) had previously summarized the importance of the loss of the West Antarctic part of the Antarctic ice sheet in terms of future climate policy.

Although the use of satellite remote sensing techniques, such as data from the Geoscience Laser Altimeter System (GLAS) on the Ice, Cloud, and land Elevation Satellite (ICESat) and the Gravity Recovery And Climate Experiment (GRACE), is beginning to provide the needed data that will accurately measure changes in the overall mass balance of the Antarctic ice sheet (Steffen and others, 2000), neither the West Antarctic nor the East Antarctic part of the ice sheet has yet been accurately measured. Future sensors on Earth-orbiting satellites that measure surface elevation, such as NASA's ICESat-II or the European Space Agency Cryosat-II, in addition to mascon data from GRACE are needed to acquire the long-term datasets. Rignot and others (2008) used interferometric synthetic-aperture radar (InSAR) data acquired by four satellites (two European, one Canadian, and one Japanese) from 1992 to 2006, to determine the total mass flux of glacier ice entering the Southern Ocean around 85 percent of the cryospheric coastline of Antarctica. The mass flux from larger drainage basins in the interior was calculated for the period 1980 to 2004 from a regional atmospheric climate model. Rignot and others (2008, p. 106) concluded:

> In East Antarctica, small glacier losses in Wilkes Land and glacier gains at the mouths of the Filchner and Ross ice shelves combine to a near-zero loss of 4±61 Gt yr^{-1}. In West Antarctica, widespread losses along the Bellinghausen and Amundsen seas increased the ice sheet loss by 59% in 10 years to reach 132±60 Gt yr^{-1} in 2006. In the Peninsula, losses increased by 140% to reach 60±46 Gt yr^{-1} in 2006. Losses are concentrated along narrow channels occupied by outlet glaciers and are caused by ongoing and past glacier acceleration. Changes in glacier flow therefore have a significant, if not dominant[,] impact on ice sheet mass balance.



EXPLANATION

Ice Wall or Ice Front* Reliability: Good	Poor	Floating Glacier Tongue*	
red solid	red dashed	red dotted	2000s
orange solid	orange dashed		1990s
green solid			1980s
light blue solid	light blue dashed		1970s
dark blue solid	dark blue dashed	dark blue dotted	1960s
purple solid	purple dashed		1950s and 1940s

*Selected ice fronts, iceberg tongues, and floating glacier tongues show date of observation

- Grounding line
- Former grounding line
- Location of fluctuating unnamed glacier or ice front
- Named and unnamed glaciers and (or) ice fronts that have fluctuated an average of more than 200 meters per year (m a^{-1}) during some time interval since the 1940s
  - Retreat
  - Advance
  - Retreat and advance
- **Former Larsen Ice Shelf**—Farthest extent observed since the 1940s

Figure 18.—*Wordie Ice Shelf map inset, part of "Coastal-change and Glaciological Map of the Larsen Ice Shelf area, Antarctica: 1940–2005." Selected ice-front positions between 1947 and 2004 are shown, as well as named and unnamed glacial features. See Ferrigno and others (2008) for full map, map explanation, and accompanying pamphlet with tables. Named glacial features and numbered unnamed glacial features are listed in pamphlet and in tables 3 and 4, respectively. Average annual change of the ice front, calculated for the time intervals between years when measurements were made, is detailed in table 5A.*

North Polar Region

Greenland Ice Sheet (Chapter C of this volume; Weidick, 1995)

The Greenland ice sheet is the second largest glacier on Earth. Its area is 1,736,095 km^2, and the volume of the inland ice is estimated to be 2,600,000 km^3 (tables 1–3). The Greenland ice sheet has a very different geographic setting than that of the Antarctic ice sheet, even though both are surrounded by water; coastal ocean currents influence the cryospheric coasts of both ice sheets. In the case of the Greenland ice sheet, warmer ocean currents penetrating fjords have resulted in calving of tidewater glaciers, acceleration of flow, and retreat of termini, as is the case with Jakobshavn Isbræ in western Greenland (Howat and others, 2007). The continent of Antarctica sits astride the South Pole, has the highest mean elevation of any continent, is surrounded by the Southern Ocean and by the Circum-Antarctic Current, and is distant from other continents. Greenland, on the other hand, lies in the North Atlantic Ocean and is influenced by the northward flowing Gulf Stream to its south, which then turns east towards Iceland before reaching Greenland. Intense low-pressure systems track along the Gulf Stream along a frontal boundary of cooler (to the west) and warmer (to the east) maritime air masses. In addition, seasonal warm air masses from the North American continent can reach the southwestern coast of Greenland. Therefore, the Greenland ice sheet is more vulnerable to melting under a warming of the global climate than is the Antarctic ice sheet. In fact, during the last interglacial, about 130,000 years B.P., most of the Greenland ice sheet may have melted (Koerner, 1989). Small ice sheets and ice caps (a smaller version of an ice sheet, being less than 50,000 km^2 in area; see table 4) and associated outlet glaciers, however, probably persisted along the coast of the subglacial mountain range that extends north to south on the eastern part of Greenland. Letréguilly and others (1991) and Fabre and others (1995) have carried out modeling studies of Greenland to determine changes in area and volume of the Greenland ice sheet under warmer climate conditions; Otto-Bliesner and others (2006) also carried out simulations of a warmer Arctic during the last interglacial. Figure 19, modified from Fabre and others (1995), shows the remnants of the Greenland ice sheet under two different modeled temperature regimens, the first an increase of 3°C, the second at an increase of 5°C. In the 5°C model, only six ice caps, five on the eastern part of Greenland, remain. Letréguilly and others (1991) stated that no glacier ice would remain in Greenland at a temperature "increase" of 6°C. The model of the two ice sheets at an increase of 3°C (fig. 19) indicates that the greatest changes in the Greenland ice sheet will first occur in southern Greenland and that, even with a +3°C increase in temperature, the Greenland ice sheet (two ice sheets in this model) is still significant in area and volume. Greenland is important climatologically, in part because the "Inland Ice" (Greenland ice sheet), a regional center of cooling, sequesters a large volume of ice in the global hydrologic cycle, and it produces large numbers of icebergs (Weidick, 1995). Of the annual accumulation, it is estimated that 35 to 50 percent is discharged from the margin and from tidewater outlet glaciers as calf ice and icebergs (Weidick, 1995). To maintain an annual mass balance in equilibrium, the other 50 to 65 percent of the ablation is through discharge of meltwater subglacially, englacially, and supraglacially. The Arctic Climate Impact Assessment (ACIA) (2005, p. 205) report summarized the Intergovernmental Panel on Climate Change (IPCC) (2001) estimates

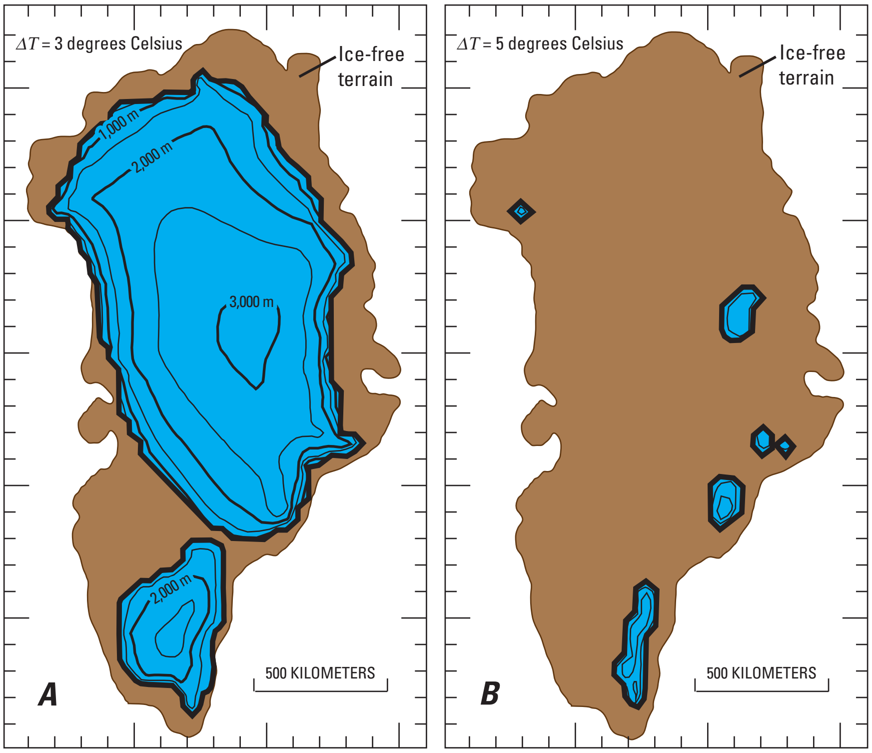


Figure 19.**—Maps of the Greenland ice sheet modeled for two warmer steady-state climates.* **A, +3°C, and* **B***, +5°C. Snow accumulation was allowed to remain within the range of present-day distribution. Modified from Fabre and others (1995, p. 5, fig. 3).*

of the mass balance of the Greenland accumulation as being 520±26 $km^3\ a^{-1}$. The estimated ablation is 564 $km^3\ a^{-1}$, of which 329±32 $km^3\ a^{-1}$ is from runoff, and 235±33 $km^3\ a^{-1}$ is from icebergs. The net loss of mass is therefore about 44 $km^3\ a^{-1}$.

As recently as the 1990s, the Greenland ice sheet became a focus of study by many glaciologists for two reasons: (1) global circulation (or climate) models projected an enhanced warming of the Arctic under global climate warming and (2) Earth-orbiting satellites with specially designed sensors to accurately measure changes in area, surface temperatures, and elevation of the Earth's surface became available or would soon become available. Also, during the 1990s, R.H. Thomas and W.B. Krabill at NASA had begun repeat surveys of the Greenland ice sheet using the geodetic airborne laser altimeter (GALA) (Thomas, 2001; Krabill and others, 2002). Bamber and others (2001) published information about new ice thickness and subglacial topography for the Greenland ice sheet. Using modeling, Oerlemans (1991) had shown the sensitivity of the mass balance of the Greenland ice sheet. Braithwaite and Olesen (1993) examined the sensitivity to climate change of the rate of ablation on the margin of the Greenland ice sheet. Ohmura and others (1996) and Gregory and others (2004) addressed changes in the mass balance of the Greenland ice sheet that a rise in annual average temperature of >3°C caused. They concluded that, under such conditions, the Greenland ice sheet would melt. Krabill and others (1999, 2000, 2004) analyzed the sequential sets of GALA data and found that the higher elevation parts of the Greenland ice sheet were in relative balance but that the lower elevation parts were thinning. Analyzing 40 years of data, Paterson and Reeh (2001) determined that the ice sheet in northwest Greenland had thinned. From satellite altimeter data acquired from 1992 to 2003, Johannessen and others (2005) found that the interior of the Greenland ice sheet had thickened because of snow accumulation. Abdalati and others (2001) discussed changes in outlet glaciers and thinning of the margin of the Greenland ice sheet. Thomas and others (2000, 2001) found thickening at higher elevations but noted that the thickening varied geographically, with thinning occurring in the southeast and south (Krabill and others, 1999). Zwally and others (2005) came to the same conclusion from analyzing satellite radar altimeter data. Hanna and others (2005) analyzed the relationship of discharge of meltwater to the mass balance of the Greenland ice sheet. Velicogna and Wahr (2006a) reported on an acceleration of loss of mass in spring 2004. Using data from the GRACE satellites, Luthcke and others (2006) were able to quantify mass loss in parts of the Greenland ice sheet.

During the beginning of the 21st century Zwally and others (2002), Rignot and Kanagaratnam (2006), and Howat and others (2007) reported changes in the flow of outlet glaciers from analysis of satellite radar/interferometry; they also suggested a relationship between the observed increase in volume of meltwater and the increase in the velocity of outlet glaciers. Thomas and others (2009), however, consider the warming of deeper ocean waters penetrating beneath outlet glaciers with deep beds to be a more important process in changing Greenland's outlet glaciers than is an increase in surface melting on the ice sheet and on associated outlet glaciers. Warming of ocean waters is considered to be a factor in the breakup of ice shelves on the Antarctic Peninsula, so ocean warming may be common to the reduction in glacier ice in both polar ice sheets. Using data from active and passive microwave sensors on Earth-orbiting satellites, Abdalati and Steffen (2001), Steffen and others (2004), Hall and others (2004), and Fettweis and others (2006) showed changes in patterns of seasonal melt (onset, termination, and length of melt). The greatest areal extent of surface melt was found to have occurred in 2002 (fig. 20) (Arctic Climate Impact Assessment [ACIA], 2004, 2005).

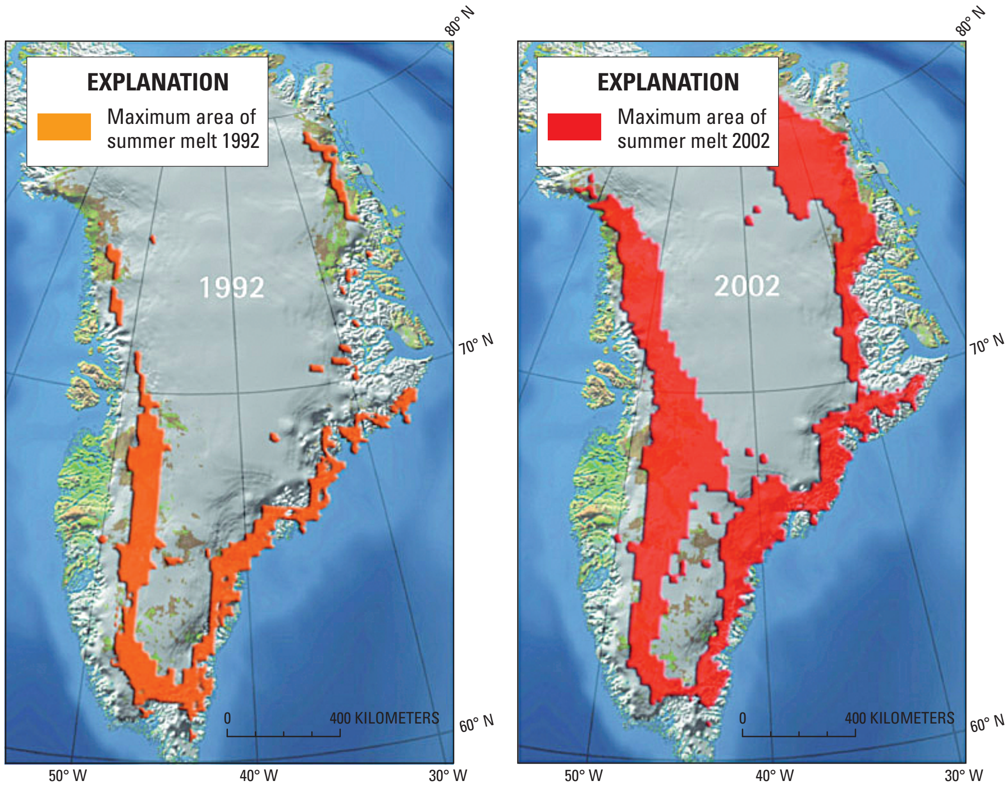


Figure 20.*—Maximum area of summer melt on the surface of the Greenland ice sheet in 1992 and as of 2002. Modified from K. Steffan in Arctic Climate Impact Assessment (ACIA) (2004, p. 40; 2005, p. 205, fig. 618).*

However, Mernild and others (2009) and Mote (2007) reported a record area of surface melt and runoff on the Greenland ice sheet in 2007; Tedesco and others (2008) reported extreme melt in the northern part of Greenland during the 2008 melt season. Hall and others (2004, 2006) analyzed MODIS surface-temperature data for the Greenland ice sheet during the melt seasons of 2000 to 2005, and they linked surface temperature and the seasonal melt pattern on the ice sheet (using MODIS data from 2000 to 2006) to mass loss (using GRACE data from 2004 to 2006) (Hall and others, 2008) (fig. 21). Hall and others (2009) also examined surface and near-surface melt on the Greenland ice sheet using both MODIS LST and QuikSCAT data for the 2007 melt season; Quick Scatterometer (QuikSCAT) data recorded an 11 percent greater area of melt than MODIS land surface temperature (LST) data because QuikSCAT detected both surface and subsurface melt. Analysis of GRACE data substantiated the acceleration in melting of the Greenland ice sheet (Chen and others, 2006).

<u>Glaciers in Greenland, Independent of the Greenland Ice Sheet (Chapter C of this volume; Weidick, 1995)</u>

Within mainland Greenland and its numerous islands, but peripheral to the ice sheet, are a number of local ice caps and other mountain glaciers. For these glaciers Weidick (1995) calculated an area of 48,599 km^2 (table 1), and Meier and Bahr (1996) estimated a volume of 20,000 km^3 (table 2). Dyurgerov and Meier (2006), in their analysis of the area and volume of the Earth's glaciers, estimated an area of 70,000 km^2 and a volume of 90,000 km^3 for glaciers in Greenland that are independent of the ice sheet.

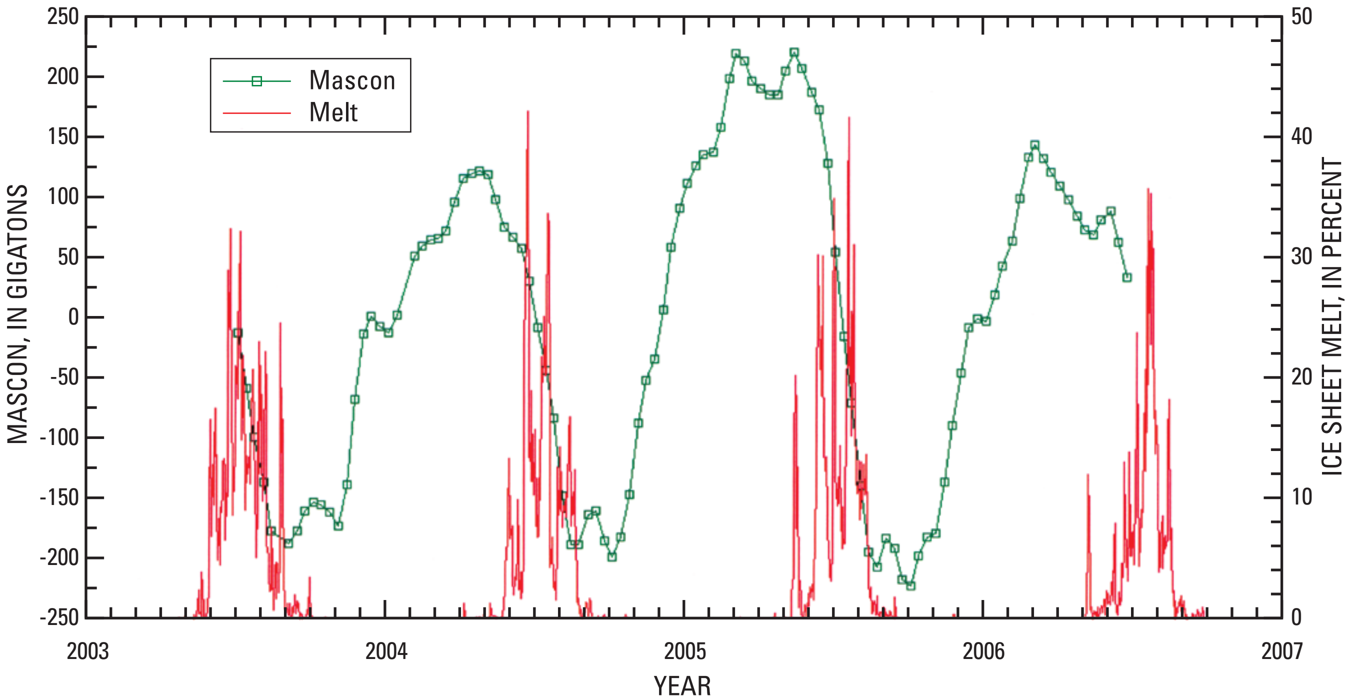


Figure 21.—*Percentage of ice-sheet melt derived from MODerate-resolution Imaging Spectroradiometer (MODIS) images and data on MASs CONcentration (mascon) derived from the Gravity Recovery and Climate Experiment (GRACE) for the entire Greenland ice sheet, for the 3-year trend July 2003 through July 2006. Signals from Earth and ocean tide and atmospheric mass have been removed. Modified from Hall and others (2008, p. 10, fig. 11).*

Canadian Arctic Islands (Chapter J, Glaciers of North America of this volume; Williams and Ferrigno, 2002a)

The total area of ice caps and other glaciers on islands in the Canadian Arctic is 151,057 km^2 (table 1). In a discussion of ice caps and other glaciers of the Canadian Arctic Islands in Chapter J (J-1, Glaciers of the Arctic Islands of the Glaciers of Canada; Andrews, 2002; Jeffries, 2002; Koerner, 2002) of this volume, very little change in the glaciers was noted. The Arctic Climate Impact Assessment (ACIA) (2005, p. 204), which noted no trend apparent in the mid-1980s, reinforced the 2002 conclusion. According to measurements since the mid-1980s, however, mass-balance values have become more negative, with summer mass balances increasingly so. Arctic Climate Impact Assessment (ACIA) (2005, p. 204) noted, "This indicates that, as in other parts of the Arctic, summer temperature drives variations in the annual mass balance." Similar conclusions were reached by Dowdeswell and others (1997) and by Zemp and others (2008). Dowdeswell and others (1997) note that during the past 80 years the Meighen Ice Cap has lost about 13 m in the elevation of its ice surface.

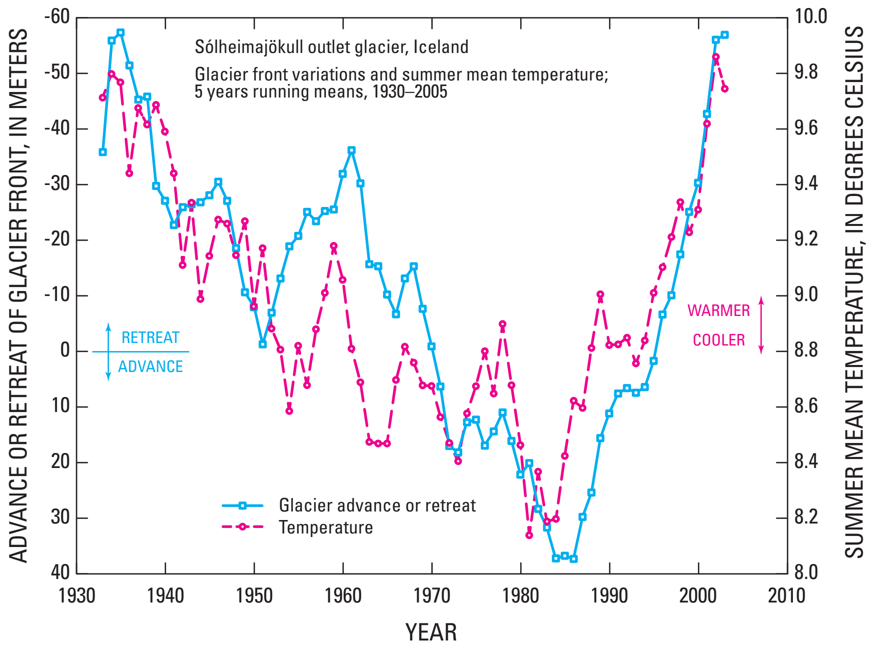
Glaciers of Iceland (Chapter D, Glaciers of Iceland, of this volume)

The glaciers of Iceland, with an area of 11,048 km^2 in 2000 (table 1), have been the subject of scientific study since the end of the 18th century, when Sveinn Pálsson completed a comprehensive treatise on Iceland's glaciers (Pálsson, 2004 [1795]). A systematic program to monitor annually the fluctuations of Iceland's glaciers was begun in 1930 (Sigurðsson and others, 2007); at the present time about 50 sites at the termini of various glaciers are monitored (Arctic Climate Impact assessment [ACIA], 2005, p. 206). Sigurðsson (1998) summarized the fluctuations of glacier termini from 1930 to 1995. Sigurðsson and others (2007) analyzed the data to determine how the termini of Iceland's non-surge-type glaciers have responded to variations in 20th century climate, especially to warmer summers, since 1930.

Iceland's glaciers generally advanced during the Little Ice Age, reaching their greatest areal extent about 1890 (Sigurðsson, 2005). Glaciers generally retreated from 1890 to 1930, retreated more rapidly from 1930 to 1960, slowed from 1960 to 1970, then started to advance in the 1970s to mid-1980s, when the advances peaked, and then slowed from the mid-1980s to the mid-1990s (Sigurðsson and others, 2007). Most glaciers began to retreat in the early 1990s; all were retreating by 2000; the fluctuation of glacier termini can be correlated with variations in climate during the 20th century (Sigurðsson and others, 2007). Data on summer temperature at the meteorological station at Stykkishólmur, northwestern Iceland, along with data on variations in the terminus of Sólheimajökull, an outlet glacier of the Mýrdalsjökull ice cap in south-central Iceland, are presented in figures 22*A* and 22*B* (Sigurðsson, 2006). Sigurðsson and others (in press) provide an overview of Iceland's glaciers.

All of Iceland's named glaciers, historic and modern (269), were documented by Sigurðsson and Williams (2008). These are the glaciers for which the areas were measured in 2000. Ice caps and associated outlet glaciers constitute 98 percent of the area; mountain glaciers constitute the remaining 2 percent.

A



B

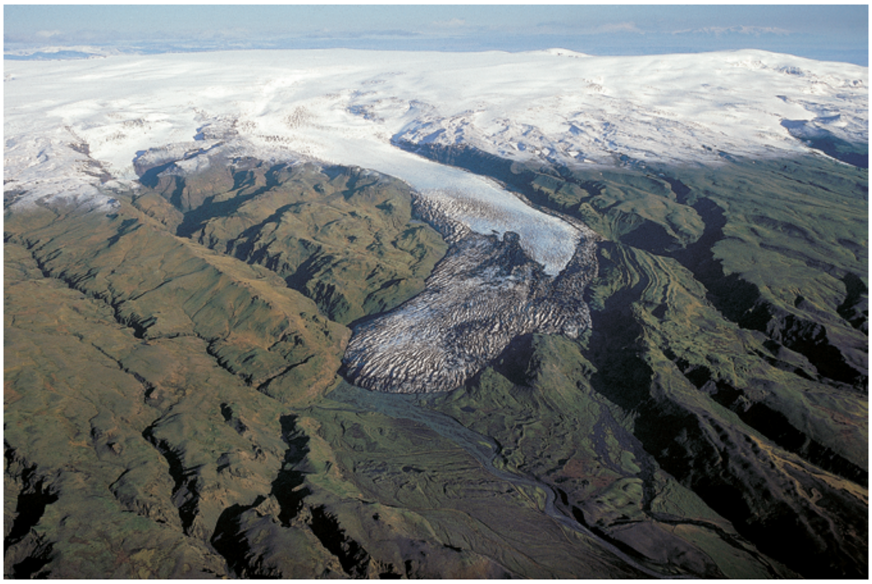


Figure 22.—**A**, *Annual variations of the terminus of Sólheimajökull outlet glacier (blue), southern Iceland, and the mean summer (May–September) temperature (red) at the Stykkishólmur meteorological station, northwestern Iceland, 5-year running mean, 1930–2005. Modified from Sigurðsson (2006, p. 89, fig. 2). The vertical scale is adjusted to facilitate comparison.* **B**, *Oblique aerial photograph of Sólheimajökull on 30 October 1985 by Oddur Sigurðsson, Icelandic Meteorological Office. (From the "Geographic Names of Iceland's Glaciers: Historic and Modern" (Sigurðsson and Williams, 2008, cover page and p. 181, fig. 171)).*

Glaciers of Svalbard, Norway (Chapter E, Glaciers of Europe, of this volume; Williams and Ferrigno, 1993)

The area of glaciers on the islands of the Svalbard archipelago is 36,591 km^2 (Liestøl, 1993). On table 1, this area is 37 percent less than the area (58,016 km^2) estimated by Flint (1971) and cited by Sugden and John (1976). The discrepancy does not represent a reduction in area because of global climate warming; rather, the explanation lies in the fact that Liestøl (1993) used modern maps of Svalbard, Norway, that he had compiled when he calculated the glacier areas. Earlier maps were all quite inaccurate, especially when map projections were used that were not appropriate for the latitude of Svalbard.

According to Arctic Climate Impact Assessment (ACIA) (2005, p. 206), "No significant changes in mass balance [of Svalbard glaciers] have been observed during the past 30 years." Dowdeswell (1995) reports a slightly negative or near-zero net mass balance for the glaciers of Svalbard; he also forecasts a negative mass balance of 0.5 m a^{-1} for a 1°C increase in temperature. Dowdeswell and others (1997) show a 35-m loss of ice on the Austre Brøggerbreen ice cap, northwestern Spitsbergen, during the past 80 years, so climate warming during the 20th century may have reduced the volume of ice in Svalbard.

Glaciers of Jan Mayen, Norway (Chapter E, Glaciers of Europe, of this volume; Williams and Ferrigno, 2003)

Jan Mayen has a glacierized area of 113 km^2 (table 1) on the ice cap and its 20 outlet glaciers on Beerenberg, an active stratovolcano on Nord-Jan (Orheim, 1993). The maximum postglacial advance was at the end of the Little Ice Age (about 1850 on Jan Mayen). Several advances and retreats of outlet glaciers occurred during the 20th century. Orheim (1993, p. El53) stated, "The glaciers on Jan Mayen are especially sensitive to change in climate." No recent research is available to determine whether the overall warming in the Arctic region is affecting the mass balance of the outlet glaciers on Jan Mayen. Lubick (2006, p. 30) notes that during the past century, Arctic and subarctic regions have warmed by 2° or 3°C, two or three times the amount that has occurred elsewhere on the planet (except, of course, for the Antarctic Peninsula, which has experienced the same magnitude of warming).

Glaciers of the Russian Arctic Islands (Chapter F, Glaciers of Asia, of this volume; Williams and Ferrigno, 2010)

The seven major Arctic islands of Russia have a total glacierized area of 56,894 km^2 (Dowdeswell and others, 2010, table 1). Most (56,477 km^2 or 99+ percent) of the glacierized area is on three island archipelagoes: Franz Josef Land (13,739 km^2), Severnaya Zemlya (18,325 km^2), and Novaya Zemlya (24, 413 km^2). To the east, Ostrov Ushakova (325 km^2), Ostrova De-longa (81 km^2), Ostrov Viktoriya (11 km^2), and Ostrov Vrangelya (3.5 km^2) (included in the Continental Russia total in table 1) have a total glacierized area of 420.5 km^2. The four islands to the east are not well known glaciologically.

During the second half of the 20th century, the termini of glaciers on the three island archipelagoes retreated (Kotlyakov, 2006). This is essentially the same conclusion reached in Arctic Climate Impact Assessment (ACIA) (2005, p. 206–207): a negative mass balance and retreat of glaciers on Noyava Zemlya; a reduction of 210 km^2 in area of glaciers on Franz Josef Land from 1953 to 1993; and a probable negative mass balance of glaciers on Severnaya Zemlya. Negative mass balances for glaciers on all three island archipelagoes were reported in Dowdeswell and others (1997).

North American Continent Region

Glaciers of Alaska (Chapter K, of this volume; Molnia, 2008)

Glaciers cover 74,600 km^2 of Alaska (table 1) in 11 mountain ranges, one large island (Kodiak Island), one island archipelago (Alexander Archipelago), and one island chain (Aleutian Islands) (Molnia, 2008). The discrepancy between this figure and the lesser figure estimated by Flint (1971) and Sugden and John (1976) (table 1) is accounted for by the 1:250,000-scale maps of Alaska available to them at the time. The USGS later published 1:63,360- and 1:62,500-scale maps of Alaska, but Alaska still stands alone (as of 2012) among all U.S. states and territories in not having modern (1:25,000-scale) topographic map coverage. Maps at that scale would provide much more detail of Alaska's terrain, including its glaciers.

Since the end of the Little Ice Age (from the early 18th to the late 19th century in Alaska), glaciers at lower elevations have been thinning and termini of glaciers have been retreating, except for about 12 tidewater glaciers; at higher elevations, glaciers are thickening or in equilibrium (Molnia, 2008). Arendt and others (2002), extrapolating from their geodetic airborne laser altimeter (GALA) (also known as Light Detection And Ranging, or LiDAR) surveys of 67 Alaskan glaciers, found a rate of glacier thinning of -0.52 m a^{-1} between the mid-1950s (topographic maps) and the GALA surveys; a follow-up GALA survey of 28 glaciers in 2000 and 2001 found that the rate of thinning had more than tripled to -1.8 m a^{-1}. Dowdeswell and others (1997) reported that glaciers in northern Alaska also exhibited more negative trends in mass balance during the 1990s, compared with previous decades.

Glaciers of Mainland Canada (Chapter J, Glaciers of North America, of this volume; Williams and Ferrigno, 2002a)

Glaciers cover an area of 50,041 km^2 in western and northwestern Canada and in Labrador (table 1). Mainland Canada contains about 25 percent of Canada's glacierized area (Ommaney, 2002a, b, c; Williams and Ferrigno, 2002b), and the Arctic Islands contain 75 percent. Some glaciers along the border between Canada and the United States (Alaska) are shared (Clarke and Holdsworth, 2002a, b). For those glaciers in western Canada where mass-balance measurements have been carried out for several decades, there has been a loss of ice. Place and Peyto Glaciers, for example, now have strongly negative mass balances (Zemp and others, 2008), and both have shown significant retreat of their termini from 1965 to 2005.

Glaciers of the Western United States (Chapter J, Glaciers of North America, of this volume, Williams and Ferrigno, 2002a)

In the western United States, glaciers cover 580 km^2 (Krimmel, 2002) (table 1), about 10 percent more area than reported by Flint (1971) and Sugden and John (1976) report. The discrepancy can be explained by the current (2009) availability of more accurate maps to measure the areas of glaciers in Washington, Oregon, California, Montana, Wyoming, Colorado, Idaho, Utah, and Nevada. Fountain and others (2007) have been studying glaciers in the American West by analyzing data from maps and from remote sensing, using GIS technology. Williams (2009) summarized the recent work on 8,303 glaciers (and snow patches) in the western United States by Andrew

Fountain and his colleagues. Although most glaciers have lost area, their study shows considerable geographic variability: higher elevation glaciers have lost less area, while 52 glaciers lost 66 percent of their area in Montana and California. In Glacier National Park, Montana, all the glaciers have retreated since the end of the Little Ice Age in the late 19th century (Key and others, 2002). Barnett and others (2005) and Kerr (2007) reported on long-term shrinkage of mountain snowpack in the western United States; modeling projections forecast a significant shrinkage of the mountain snowpack in California and Oregon (Kerr, 2007).

Glaciers of México (Chapter J, Glaciers of North America, of this volume, Williams and Ferrigno, 2002a)

México's three highest mountains, all volcanoes, have a total glacierized area of 11.44 km^2 (White, 2002) (table 1). Volcán Popocatépetl (5,465 m), an active volcano, has erupted at least 25 times since 1345; some of the recent eruptive activity has affected its glaciers. By the early 21st century, glaciers on all three volcanoes had been reduced in area. Julio-Miranda and others (2001) reported a 72 percent decrease in area of glaciers on Volcán Popocatépetl from 1958 to 2000. It is likely that the three-summit Volcán Iztaccíhuatl (5,230 m), a dormant stratovolcano, and Volcán Popocatépetl will no longer have glaciers by the end of the second decade of the 21st century. Volcán Pico de Orizaba (Volcán Citlattépetl) (5,610 m), historically active, has the largest glacierized area—an ice cap and associated firn field, Gran Glacier Norte, that has seven outlet glaciers (White, 2002); because of its higher elevation, this volcano will retain its glaciers after glaciers disappear from the other two volcanoes.

South American Continent Region

Glaciers of Venezuela, Colombia, Ecuador, Perú, Bolivia, Chile, and Argentina (Chapter I, Glaciers of South America, of this volume; Williams and Ferrigno, 1998)

Glaciers of varying size extend from north to south along the crests of the Andes and associated cordilleras, with outlet glaciers from ice fields and ice caps extending to sea level only in the southern part of the Andes in Chile and Argentina (Lliboutry, 1998a, b): Northern Patagonian Ice Field, Southern Patagonian Ice Field, and Cordillera Darwin in Tierra del Fuego. Venezuela has 2 km^2 of glacier area; Columbia has 104 km^2; Ecuador has 97 km^2; Perú has 2,600 km^2; Bolivia has 560 km^2; and Chile/Argentina have 21,700 km^2. The total glacierized area of South America is 25,063 km^2 (Williams and Ferrigno, 1998 (table 1).

Most of the glacierized area of South America, about 87 percent, is in the three large ice fields in southern South America: the Northern (4,200 km^2) and Southern (13,000 km^2) Patagonian Ice Fields of Patagonia; the ice field of Cordillera Darwin (2,300 km^2), in Tierra del Fuego, Chile; and other smaller glaciers (fig. 23).

All the low-latitude glaciers in the Andes of Venezuela (Schubert, 1998), Colombia (Hoyos-Patiño, 1998), Ecuador (Jordan and Hastenrath, 1998), Perú (Arnao, 1998), and Bolivia (Jordan, 1998) retreated during the late 20th century and early 21st century (Zemp and others, 2007; Vuille and others, 2008). As with glaciers in other high-mountain regions of the Earth, shrinkage

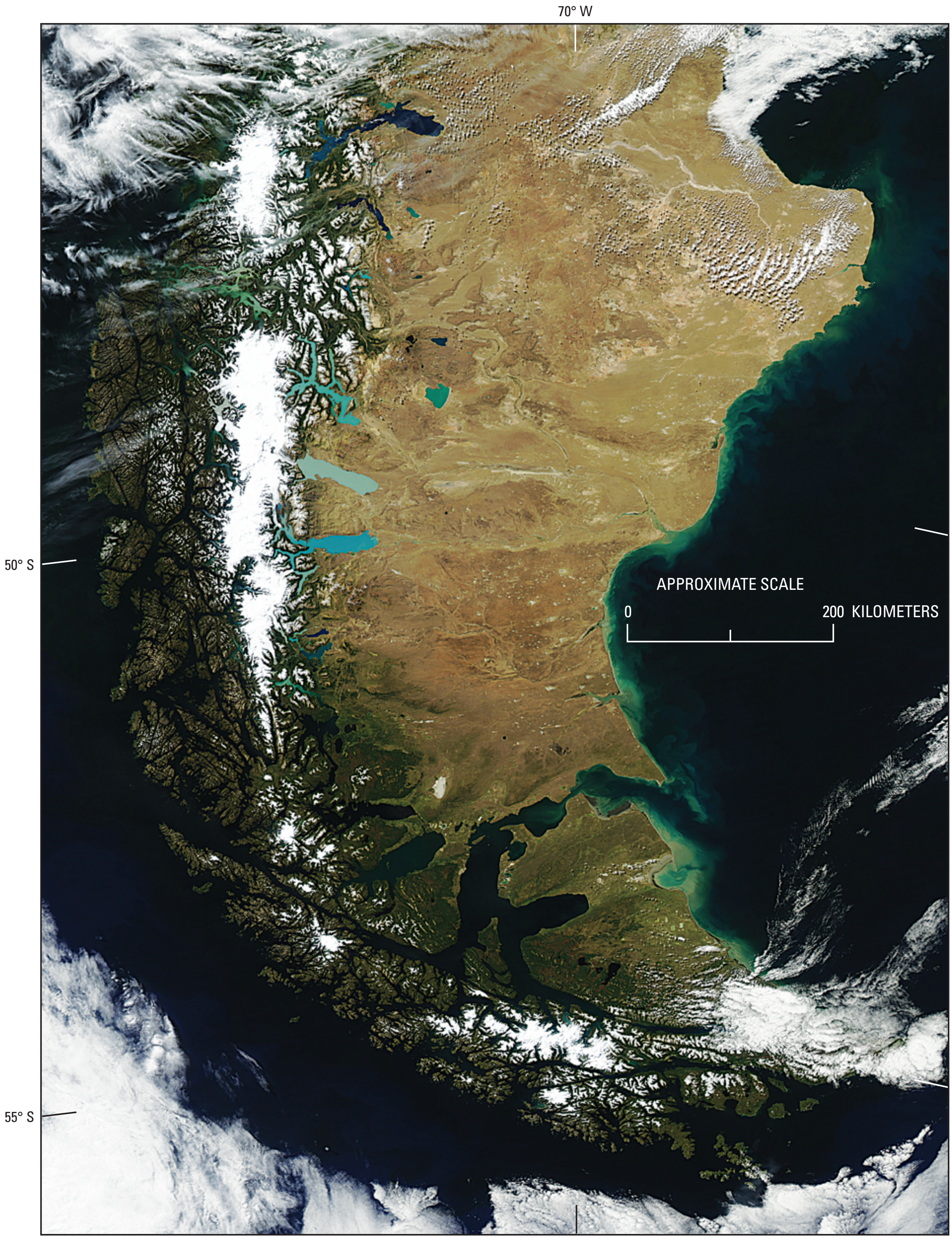


Figure 23.—*MODerate-resolution Imaging Spectroradiometer (MODIS) image on 2 October 2005 of the Northern Patagonian Ice Field, the Southern Patagonian Ice Field, the ice field of the Cordillera Darwin, outlet glaciers from each ice field, and other glaciers. Some of the glacially scoured "Finger Lakes" (south to north, Lago Argentino, Lago Viedma, and Lago San Martín) in the center of the image are "robin's-egg-blue color", the result of glacial rock-flour sediment entering the lakes. MODIS image from the Aqua satellite; Goddard Space Flight Center, National Aeronautics and Space Administration.*

of glaciers and, ultimately, complete loss of glacier ice will markedly reduce the amount of water in glacierized hydrologic basins. For many high-mountain communities downstream, such as in Bolivia, glacial meltwater is the primary souce of water for drinking and irrigation (Rosenthal, 2009). From 1850 through 2002, the glacier on Nevado de Santa Isabel, Colombia, reduced in area 87 percent (Zemp and others, 2007). Hastenrath (2008) published photographic documentation of glacier recession in the Cordillera Blanca of Perú. The study by Hastenrath and Ames (1995) of Yanamarey Glacier in the Cordillera Blanca, Perú, showed 1.55 km retreat of its terminus, 0.9 km^2 decrease in area, and 0.01 km^3 shrinkage in volume from 1948 through 1988. Qori Kalis, an outlet glacier from the *Quelccaya Ice Cap*, Perú, which has been the subject of more than 30 years of study by Lonnie Thompson, will probably retreat into the ice cap (disappear) by 2012 (Sever, 2007); Thompson (2007, written commun.) stated that Qori Kalis is now retreating at 205 m a^{-1}, more than 40 times more rapidly than in 1978 (fig. 24). Thompson, an ice-core glaciologist, took the iconic photograph of the annual ice layers exposed in the margin of the *Quelccaya Ice Cap* in 1977; a second photograph taken from the same location of the camera as in 1977 was taken in 2002 (fig. 24). Four additional photographs of Qori Kalis taken from the same camera position (different location from the 1977 and 2002 photographs) show the receding terminus in 1978, 2002, 2005, and 2008 (fig. 24).

Chilean, Argentinian, Japanese, U.S., and other scientists (Casassa and others, 2002) have conducted extensive research on the Northern and Southern Patagonian Ice Fields, remnants of an extensive ice sheet that extended well to the east during glacial intervals (Lliboutry, 1998b), and the largest glaciers in South America (fig. 23). Casassa and others (1997, p. 106) stated that "Most glaciers in Patagonia are retreating rapidly." Skvarca and DeAngelis (2002) analyzed Landsat TM images from 1986, 1997, and 2001 of five outlet glaciers from the Southern Patagonian Ice Field, all of which calve ice into a fjord, (1) or into a lake (4). Four of the calving tidewater/lacustrine glaciers showed significant changes in their termini during the 20th century: Glacier Brüggen (Glaciar Pío XI) showed advances and retreats while Glaciar O'Higgins, Glaciar Viedma, and Glaciar Upsala showed only recession. The relatively stable Glaciar Perito Moreno had small advances and retreats. The glaciologists' analysis of 34 outlet glaciers, the termini for which are on land, shows that all but one had retreated from 1986 to 2001. Zemp and others (2008) note that the lower parts (near the margins) of the Southern Patgonian Ice Field were thinning at a rate of 30 m a^{-1} (from Rignot and others, 2003).

Aniya and Wakao (1997) and Aniya (2001), who studied changes of 21 outlet glaciers in the Northern Patagonian Ice Field by comparing data from 1944 to 1945 (maps) and from 1995, 1998, and 1999 (hand-held oblique aerial photographs), report a general recession of the outlet glaciers. The only exception is a strong advance and retreat of Glaciar San Rafael, a tidewater glacier. Zemp and others (2008) note that the Northern Patagonian Ice Field had lost 140 km^2, 3.4 percent of its area from 1942 to 2001.

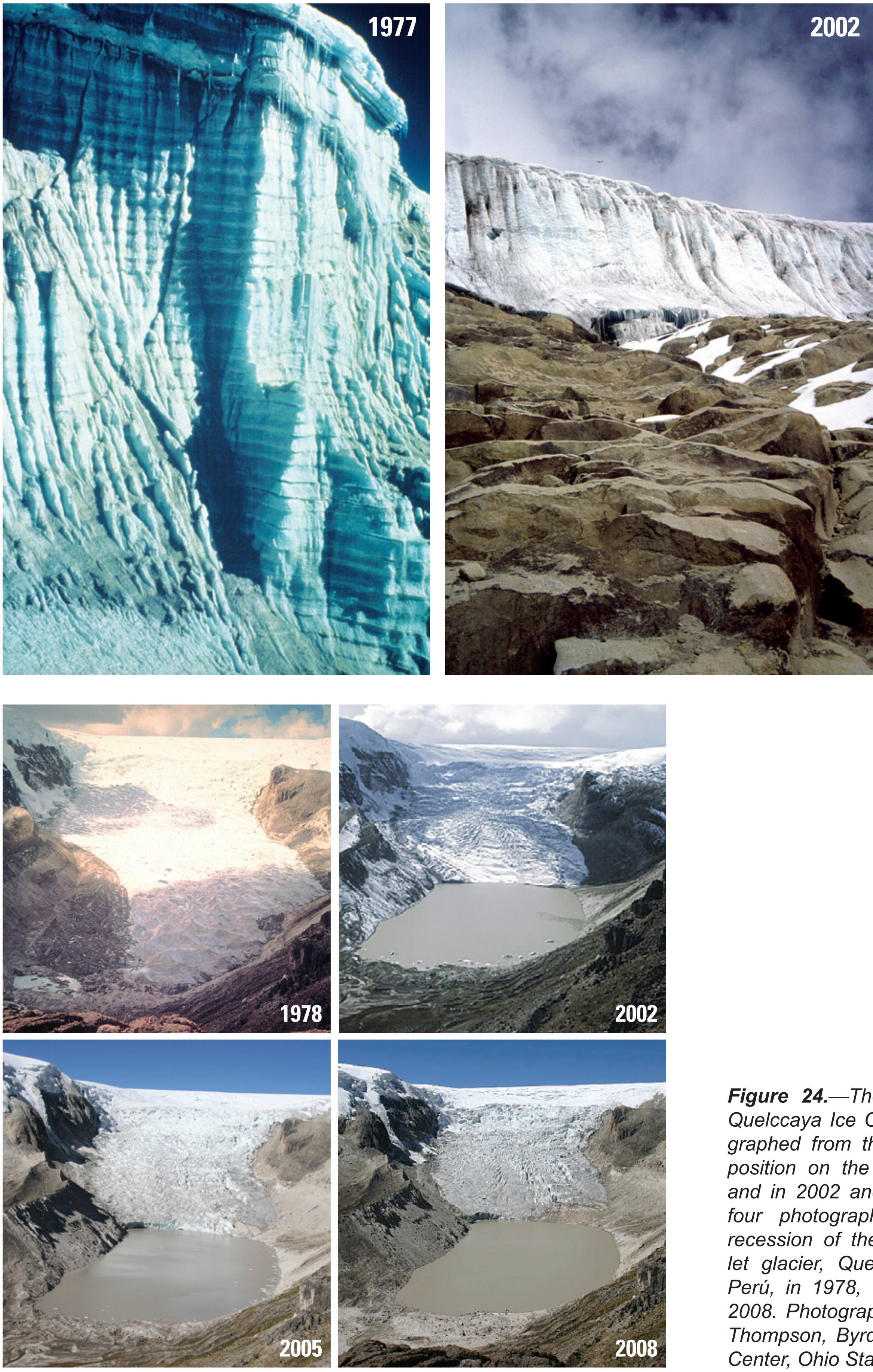


***Figure 24.**—The margin of the Quelccaya Ice Cap, Perú, photographed from the same camera position on the ground in 1977 and in 2002 and a sequence of four photographs showing the recession of the Qori Kalis outlet glacier, Quelccaya Ice Cap, Perú, in 1978, 2002, 2005, and 2008. Photographs by Lonnie G. Thompson, Byrd Polar Research Center, Ohio State University.*

European Region

Glaciers of Norway, Sweden, France, Switzerland, Austria, Italy, and Spain (Chapter E, Glaciers of Europe, of this volume; Williams and Ferrigno, 1993)

The glaciers of Norden (Scandinavia) (Norway, Østrem and Haakensen, 1993; Sweden, Schytt, 1993) and the glaciers of the Alps (Austria, Rott, 1993; France, Reynaud, 1993; Italy, Barbero and Zanon, 1993; and Switzerland, Scherler, 1993) are the best mapped and most closely monitored in the world, during more than 100 years for Austria, France, Italy, and Switzerland, for example. The monitoring is sustained because of the importance of forecasting runoff from wholly or partially glacierized hydrologic basins so that full reservoir capacity (optimum hydraulic head) is maintained in order to maximize the generation of electricity from hydropower, especially in Norway and Switzerland. Strong glaciological research institutions support the generation of electricity by hydropower in Canada (BC Hydro), Norway (Norges Vassdrags og Energiverk (NVE) (Norwegian Water Resources and Energy Directorate)), and Iceland. Iceland established a new glaciological unit in Veðurstofa Íslands (Icelandic Meteorological Office) in 2009, having transferred the Vatnamælingar (Hydrological Service) from Orkustofnun (National Energy Authority) and merged it with another glaciological unit in Veðurstofa Íslands.

Among the Swiss scientists who pioneered the measurement of glacier fluctuations during the past century are the late Fritz Müller and Wilfried Haeberli. In particular, Müller and Haeberli were responsible for establishing the Permanent Service on the Fluctuation of Glaciers and the World Glacier Inventory, now merged into the World Glacier Monitoring Service in Zürich, a testament to their "vision" of having glaciologists from all nations contribute to a global database of glacier changes, including collecting and archiving periodic measurements of the fluctuation of glacier termini and changes in mass balance.

When the areal extent of the glacierized areas of Scandinavia, the Alps, and the Pyrenees (Serrat and Ventura, 1993) were compiled in 2009 (table 1) and compared to the areas estimated by Flint (1971) and Sugden and John (1976), all these regions show significant reduction. Except perhaps for the Pyrenees, excellent maps existed when Flint (1971) prepared his table, and therefore the reductions in area for Scandinavia and the Alps in 2009 are scientifically valid. For Scandinavia, the areal extent of glaciers has been reduced from 3,810 km^2 to 2,909 km^2, a reduction of 24 percent; for the Alps, a reduction of 3,600 km^2 to 2,842 km^2, 21 percent; and for the Pyrenees, a reduction from 33 km^2 to 8.11 km^2, a loss of 75 percent.

In Norway and Sweden, glaciers reached their maximum extent at the end of the Little Ice Age in the 18th century, with fluctuations throughout the 19th century (Zemp and others, 2008). According to Zemp and others (2008), most Scandinavian glaciers retreated throughout the 20th century; the general recession was interrupted by advances ca. 1910 and ca. 1930, in the late 1970s, and in ca. 1990. With an increase in snowfall in the 1960s, maritime glaciers in Norway had positive mass balances. Since the beginning of the 21st century, however, all monitored glaciers in Norway had negative mass balances (Andreassen and others, 2005; Zemp and others, 2008). Nesje and others (2008) reviewed the history of fluctuations of mountain glaciers in Norway, their present state, and what is likely to happen in the future.

For the Alps as a whole, there has been a general trend of glacier recession for 150 years, interrupted by re-advances in the 1890s, 1920s, and 1970s to 1980s (Zemp and others, 2008). Paul and others (2004, p. 12) report, "Analyses of multispectral satellite data indicate accelerated glacier decline around the globe since the 1980s." In the Alps, rapid distintegration of Alpine glaciers has been reported by Haeberli (1994), Paul and others (2004), and Zemp and others (2006). Paul and others (2004, p. 12) report an 18 percent reduction in area of 930 Alpine glaciers from 1985 to 1999, a rate seven times higher than the decadal mean from 1850 through 1973; the volume loss from 1973 to 1999 is 25 km^3. Haeberli and Hoelzle (1995) report that the total volume of glaciers in the Alps was about 130 km^3 in the mid-1970s. The estimated loss of glacier volume is 19 percent for the final 25 years of the 20th century. Paul and others (2004) report that the volume loss of 25 km^3 was accompanied by a loss of 675 km^2 in area of alpine glaciers during the same period. If the loss of glacier ice continues at the same rate, no glaciers will exist at elevations less than 2,000 m by 2050 or at elevations less than 2,500 m by 2100. Paul, Kääb, and Haeberli (2007) report on continuing changes in the glaciers of the Alps from their analysis of satellite images.

The Pasterze Glacier in the eastern Austrian Alps has been studied for more than 100 years, including using satellite data in addition to ground data during many of these years. Landsat and other satellite data continue to show decreases in areal extent, even showing measurable changes from year to year (Hall and others, 1992; Bayr and others, 1994; Hall, Bayr, and others, 2003; Bayr and Hall, 2009). The glacier has also lost much of its accumulation area, as determined from Landsat data (Bayr and others, 1994). Because supraglacial debris covers a large part of the western part of the glacier terminus, it is difficult to measure the actual extent of the glacier tongue, even in the field. However, K. Bayr used GPS data and compared measurements with satellite-derived measurements to derive the full extent of the glacier terminus (Bayr and Hall, 2009).

In the Pyrenees between Spain and France, Chueca and others (2007, p. 547) report, "The last two decades of the 20th century and the beginning of the 21st century have been characterized by rates of glacial shrinkage as high as those observed around 1860–1900, immediately after the maximum of the LIA [Little Ice Age]. . . ." Chueca and others (2007), analyzing changes in area and volume of glaciers in the Maladeta massif from 1981 to 2005, report that the area of glaciers had been reduced by 35.7 percent during the 25-year period, a period characterized by an increase in maximum temperatures and decrease in snowfall throughout the region.

Southwest Asia Region

Glaciers of Turkey and Iran (Chapter G, Glaciers of the Middle East and Africa, of this volume; Williams and Ferrigno, 1991b)

In Southwest Asia, small glacierized areas (24 km^2) exist in the mountainous regions of Turkey, mostly in the southeastern Taurus Mountains (Kurter, 1991) and in three mountainous regions of Iran (20 km^2) (Ferrigno, 1991) (table 1). Moussavi and others (2009) produced a new inventory of glaciers in Iran based on analysis of aerial and satellite imagery of five glacierized regions; they report a glacierized area of 27.12 km^2. A report by Yavaşlı and Ölgen (2008) documented retreat of glaciers on Mount Suphan [Süphan], Turkey, from analysis of sequential satellite images. The latitudinal location of their glaciers and warming of the global climate suggest that the glaciers of Turkey and Iran will continue to be reduced in area and volume.

Central Asian Region

Glaciers of Russia and Independent States of the Former Soviet Union (Chapter F, Glaciers of Asia, of this volume; Williams and Ferrigno, 2010)

Kotlyakov and others (2010) reviewed glaciers in Russia, including the Russian Arctic islands, and the independent states that were part of the former Soviet Union (see earlier discussion of the Russian Arctic islands (p. A119). The total areal extent of the glacierization of continental Russia is 22,044 km^2 (table 1); this total includes glaciers in 20 mountain ranges in Russia, the 3.5 km^2 of glaciers on the Arctic island Ostrov Vrangelya, and glaciers in the Republics of Georgia, Kazakhstan, Kyrgystan, and Tajikistan in the former Soviet Union. Kotlyakov (2006) estimated the total volume of the glacierized area to be 1,570 km^3, and he reported the present state of glaciers in seven geographic regions to be summarized as follows (Kotlyakov, 2006, p. 397–402). Ural Mountains: Glacierization in the Urals was much diminished by the close of the 20th century, with additional degradation during the 1990s. According to Zemp and others (2007), some glaciers have completely melted away. Caucasus: The total area of glacierization decreased by 794 km^2 (36 percent) from 1900 to 1970; although the loss in glacier area (240 km^2) has continued from 1970 to 2000, it has slowed during these decades. Pamirs-Alai: Recession of glaciers continued from 1972 to 1990, although the loss in area diminished from 1990 to 2001. North Tien Shan: From the late 1950s to early 1970s, the glaciers were generally in equilibrium; for the 15 years following the early 1970s, mass balances were negative, and a marked recession of glaciers occurred; since the mid-1980s, mass balances have been more stable. Altay: From 1952 to 1998, glacier area in Central Altay was reduced by 15 percent; for the entire region of Altay, the areal extent of glacierization has been reduced 7.1 percent. Suntar-Khayata Mountains and Khrebet Cherskogo (Chersky Range): Negative mass balance of glaciers predominated during the period from 1960 to 2000. Kamchatka: Glaciers on the Kamchatky peninsula are affected by climate variability and by active volcanism on glacier-capped volcanoes; on the Kronotsky peninsula, most glaciers are retreating and have negative mass balances.

Glaciers of China (Chapter F, Glaciers of Asia, of this volume; Williams and Ferrigno, 2010)

The glacierized area of China is 59,406 km^2 (Shi Yafeng and others, 2008; Shi Yafeng and others, 2009) (table 1). The overall loss of glacier area since maximum areal extent in the 17th century is estimated to be 20 percent; since the 1960s, there has been an additional 6 percent loss (Zemp and others, 2007). In western China, including glaciers around the Tibetan Plateau, glaciers in this area show an accelerating loss of area and mass from the 1970s through the 1990s (Ding and others, 2006).

Glaciers of India (Chapter F, Glaciers of Asia, of this volume; Wiliams and Ferrigno, 2010)

The glaciers of India cover 16,755 km^2 (Vohra and others, 2010) (table 1). Analyzing sequential satellite images from 1972 to 2000, Prasad and Singh (2007) report that the areal extent of glaciers and snow cover is shrinking in the western Himalaya. Raina (2009) published a Discussion Paper on "Himalayan Glaciers" that summarized historical and modern knowledge, of glacial studies, glacial [glacier] retreat, and climate change; his report, issued by the Indian Ministry of Environment and Forests, contains many historical and modern photographs of the termini of selected glaciers. Raina (2009) is dubious about any correlation between climate change and variations in glaciers of the Himalaya, concluding that changes in the termini of the Gangotri and Siachen Glaciers are in response to changes in climate many millennia ago. Other scientists, such as Richard Armstrong of the University of Colorado, note that little or no change was taking place at elevations higher than 5,400 m but that changes in glaciers were occurring at lower elevations; his observation is similar to those documented for glaciers in Alaska (Molnia, 2008). Molnia (2008, p. K2) stated "In spite of significant changes at lower elevations, not every Alaskan glacier is thinning and retreating. In several ranges, no changes were noted in glaciers situated at higher elevations."

Glaciers of Mongolia (Summary from Kamp and others, 2011, written commun.)[4]

Only a few studies on glacier mapping exist for Mongolia, and systematic mapping has begun only recently. The Digital Chart of the World (DCW) outlines only a few glaciers in Mongolia; the World Glacier Inventory (WGI) includes very few glaciers without providing information on when the data were collected. The World Glacier Monitoring Service (WGMS) lists one Mongolian glacier. The international consortium Global Land Ice Measurements from Space (GLIMS) only recently added a specific Regional Center for Mongolia (cooperation between the University of Montana and the National University of Mongolia) to its list of regional observation centers.

Thus, knowledge of the number of all glaciers in the two glacierized areas (Mongolian Altai and Khangai Mountains of Mongolia) remains uncertain. Glaciers can be found between lat 46°25' and 50°50' N. and between long 87°40' and 100°50' E. at 2,750 to 4,374 m above sea level (a.s.l.) (Davaa and Basandorj, 2005). In general, spatial distribution is sporadic and decreases from northwest to southeast. Dashdeleg and others (1983) put the total number of glaciers at 262 covering a total area of 659 km^2 (table 1). The glaciers are responsible for supplying 10 percent of the water resources within Mongolia (Myagmarjav and Davaa, 1999) and store approximately 63 km^3 of freshwater per year (Davaa and Basandorj, 2005). Baast (1999) estimated that during the 40 years from the 1960s until the end of the 1990s the area of glaciers in Mongolia decreased by 6 percent.

[4] Editors' note: The Glaciers of Mongolia review was not included in the printed version of Chapter F, Glaciers of Asia, of this volume because of lack of data at the time. Five scientists who had begun work on a review of Mongolia's glaciers provided the information given in this entry. The authors and their affiliations of this summary are as follows: Ulrich Kamp, Brandon Krumwiede, and Kevin McManigal, Department of Geography, University of Montana, Missoula, MT 59812, U.S.A.; Michael Walther, MOLARE Research Center for Climate and Landscape Studies, National University of Mongolia, Ulaanbaatar, Mongolia; and Avirmed Dashtseren, Geographical Institute, Mongolian Academy of Sciences, Ulaanbaatar, Mongolia. See also footnote 6 in table 1 (p. A75) of this part.

Glaciers of Nepal, Afghanistan, and Pakistan (Chapter F, Glaciers of Asia, of this volume; Williams and Ferrigno, 2010)

Higuchi and others (2010) report the area of glaciers in Nepal as being 7,929 km^2, based on analysis of Landsat images by Qin Dahe (1999). Mool, Bajracharya, and Jonsi (2001) also analyzed Landsat images and reported 5,324 km^2, the value given in table 1. The estimated area of glaciers in Afghanistan is approximately 2,700 km^2 (Shroder and Bishop, 2010a); the glacierized area in Pakistan covers 15,000 km^2 (Shroder and Bishop (2010b) (table 1). Information on the state of glaciers in Afghanistan and Pakistan has been limited because of inadequate maps and limited field observations owing to civil unrest. Because glaciers in adjacent mountains are losing mass and area and their termini are retreating, it is likely that glaciers in all three countries are similarly affected (Zemp and others, 2007). Schmidt and Nüsser (2009) used historic maps and remotely sensed images (Corona, ASTER, Landsat, and QuickBird) of the Raikot Glacier, northern Pakistan, from 1934 to 2007, to document a retreat of about 200 m of the glacier terminus. Schmidt and Nüsser (2009) concluded that the retreat had been interrupted by advances during the 1950s and 1980s and that thinning of the Raikot Glacier was not significant from 1934 to 2007.

Glaciers of Bhutan (Chapter F, Glaciers of Asia, of this volume; Williams and Ferrigno, 2010)

The area of glaciers in Bhutan is 1,317 km^2 (Iwata, 2010) (table 1). From 1963 to 1993, the glacier area was reduced by 8 percent (Zemp and others, 2007; Zemp and others, 2008).

Africa Region

Glaciers of Kenya, Tanzania, Uganda, and Zaïre (Chapter G, Glaciers of the Middle East and Africa, of this volume; Williams and Ferrigno, 1991)

The area of glaciers in Africa was 10 km^2 according to Young and Hastenrath (1991) (table 1) on two volcanoes—Mount Kenya, Kenya, and Kilimanjaro, Tanzania—and in the Ruwenzori, Uganda and Zaïre. Hastenrath (2008) has published photodocumentation of the continuing recession of glaciers in all three geographic locations. According to Zemp and others (2007), about 50 percent of the glaciers in these locations have melted away. According to Zemp and others (2008), glaciers in the Ruwenzori have been losing area and their termini have been retreating since the late 19th century, and some glaciers have completely melted away; on Kilimanjaro, the glaciers have diminished in area from about 20 km^2 in 1880 to about 2.5 km^2 in 2003. Hastenrath and Greischar (1997) also documented the recession of glaciers on Kilimanjaro from 1912 to 1989 (see fig. 39): on Mount Kenya, 8 out of 18 glaciers melted away during the 20th century (Hastenrath, 2005; Zemp and others, 2005), and Lewis Glacier on Kilimanjaro decreased in volume from about 7.7 km^3 in 1978 to about 0.3 km^3 in 2004 (Hastenrath and Polzin, 2004; Zemp and others, 2008). Lewis Glacier has retreated by more than 800 m during the years 1893 through 2004 (Zemp and others, 2007). Perkins (2009c) summarized recent work by Lonnie G. Thompson and others that projects a complete loss of glacier ice on Kilimanjaro by 2022.

South Pacific Region

Glaciers of New Zealand (Chapter H, Glaciers of Irian Jaya, Indonesia, and New Zealand, of this volume; Williams and Ferrigno, 1989)

The area of glaciers in New Zealand is 1,159 km^2 (Chinn, 1989) (table 1). Six glaciers are located on Mount Ruapehu, North Island; most glaciers are along the crest of the Southern Alps, South Island. Chinn (1996) and Zemp and others (2007) report that New Zealand has lost between 25 percent and 50 percent of its glacier area since the end of the Little Ice Age at the end of the 18th century. After the mid-1980s some glaciers increased in volume and advanced. Since the beginning of the 21st century, the number of glaciers that were retreating increased; from 1977 to 2005 about 11 percent of the volume of glacier ice was lost (Zemp and others, 2007).

Glaciers of Irian Jaya, Indonesia (Chapter H, Glaciers of Irian Jaya, Indonesia, and New Zealand, in this volume; Williams and Ferrigno, 1989)

The area of glaciers in Irian Jaya, Indonesia, is 7.5 km^2 (Allison and Peterson, 1989), 50 percent less than Flint (1971) and Sugden and John (1976) reported (table 1). The discrepancy may be genuine or it may be an artifact of the inaccurate maps available to Flint (1971). According to Zemp and others (2008), nevertheless, all the glaciers in the mountains of Irian Jaya, Puncak Mandala, Ngga Pilimsit, Puncak Jaya, and Puncak Trikora have undergone extensive retreat since their maximum extent at the end of the Little Ice Age in the mid-19th century. Glaciers on Puncak Jaya, which has the highest elevation (5,030 m), covered about 20 km^2 in 1850, but by 2002 they were reduced to <3 km^2, primarily by melting of the large Meren Glacier (Zemp and others, 2008). Zemp and others (2008) also report that all glaciers, except those on Puncak Jaya, had melted away. Hastenrath (2008) provided photo-documentation of the recession of glaciers in Irian Jaya. Prentice and Brackett (2002) reported a decrease in glacier area in Puncak Jaya from 12 km^2 to 2 km^2 by 2001. Using IKONOS satellite images acquired on 8 June 2000 and 11 June 2002, Klein and Kincaid (2006) calculated that glaciers on Puncak Jaya lost 7.48 percent of their areal extent during the intervening years, from 2.326 km^2 to 2.152 km^2. Prentice and others (2003) reported on the recession of Irian Jaya's glaciers from 1936 to 2000; by the end of the 20th century, their mass balance was negative and the altitude of equilibrium lines have increased, with some variations during that 60-year interval of time.

Conclusions

By the beginning of the 21st century, glaciologists from the glacierized regions of the world and their colleagues from other nations have access to more than 100 years of ground-based measurements of fluctuations of glacier termini in many regions and to more than 50 years of modern topographic maps of many glacierized regions. In addition, glaciologists have compiled more than 20 years of mass-balance measurements for about 40 glaciers, and more than 100 glaciers had 1 to 5 years of such measurements (Dyurgerov and Meier, 1997a). Also, more than 35 years of medium-resolution satellite images of the Earth's glacierized regions are now (2009) available to glaciologists.

The World Glacier Monitoring Service (WGMS), located in Zürich, Switzerland, is the most important global archive for long-term records of the fluctuations of glaciers (since 1894) and of measurements of changes in mass balance of glaciers (since the late 1940s). By 2008, WGMS had archived about 36,000 measurements of termini fluctuations for about 1,800 glaciers and about 3,400 mass-balance measurements of 230 glaciers from the Earth's glacierized regions. A World Glacier Inventory effort in the 1970s yielded a detailed inventory of more than 100,000 glaciers, data that WGMS archives (Zemp and others, 2008, p. 9), as does NSIDC.

Most of the data in the WGMS were derived from direct field observations and measurements, from measurements on maps (compiled from stereophotogrammetric methods), or directly from vertical aerial photographs, either from individual photographs or from stereophotographs. Since 1972, however, glaciologists have been using data obtained from sensors on Earth-orbiting satellites to measure fluctuations of termini of glaciers, changes in area of glaciers, and changes in volume of glaciers. The Satellite Image Atlas of Glaciers of the World (11-chapter USGS Professional Paper 1386-A–K) is primarily based on analyses of Landsat images. The GLIMS initiative primarily uses the ASTER sensor on the Terra satellite sent into orbit in 1999. Global Land Ice Measurements from Space (GLIMS), by Bruce H. Raup and Jeffrey S. Kargel, describes the GLIMS initiative in detail (p. A247–A260). Measurements of changes in glaciers, by both conventional and remote sensing methods, are also presented (see p. A224–A246).

By the early part of the 21st century, and with the benefit of hindsight from more than 100 years of glacier measurements by conventional means and more than 35 years of measurements derived from an increasingly sophisticated and technologically advanced suite of imaging and other sensors on Earth-orbiting satellites, it was obvious that glaciers—from the smallest mountain glacier to the Greenland and Antarctic ice sheets—are especially important indicators of changes in the Earth's climate, regionally and globally. As Wilfried Haeberli, Director, WGMS, points out, "Their shrinkage and, in many cases, even complete disappearance leaves no doubt about the fact that the climate is changing at a global scale and at a fast if not accelerating rate" (Zemp and others, 2008, p. 8). Since the mid-1980s and especially since the mid-1990s, virtually all of the Earth's non-ice sheet glaciers (mountain glaciers, ice caps and ice fields and associated outlet glaciers) have experienced thinning and retreat of their termini and margins (fig. 25); the only exceptions have been surge-type glaciers, tidewater glaciers, and some temperate and high-latitude glaciers at higher elevations.

Glaciologists, such as Stefan Hastenrath, in his photographs of the recession of tropical glaciers at or near the Equator (Hastenrath, 2008), and Lonnie G. Thompson, in his photographs of the rapid recession of glaciers in the Andes (fig. 24) and at other tropical latitudes, document the accelerated thinning of the retreat of the termini and the disappearance (complete melting away) that low-latitude glaciers have been experiencing. The gifted landscape photographer Jim Balog, Extreme Ice Survey, which the National Geographic Society supports, has been using ground and time-lapse photography/imagraphy to document "unusual evidence of melting glaciers in Alaska, British Columbia, the Rocky Mountains, Bolivia, Greenland, Iceland, and the French and Swiss Alps" (Balog, 2009, p. 13).

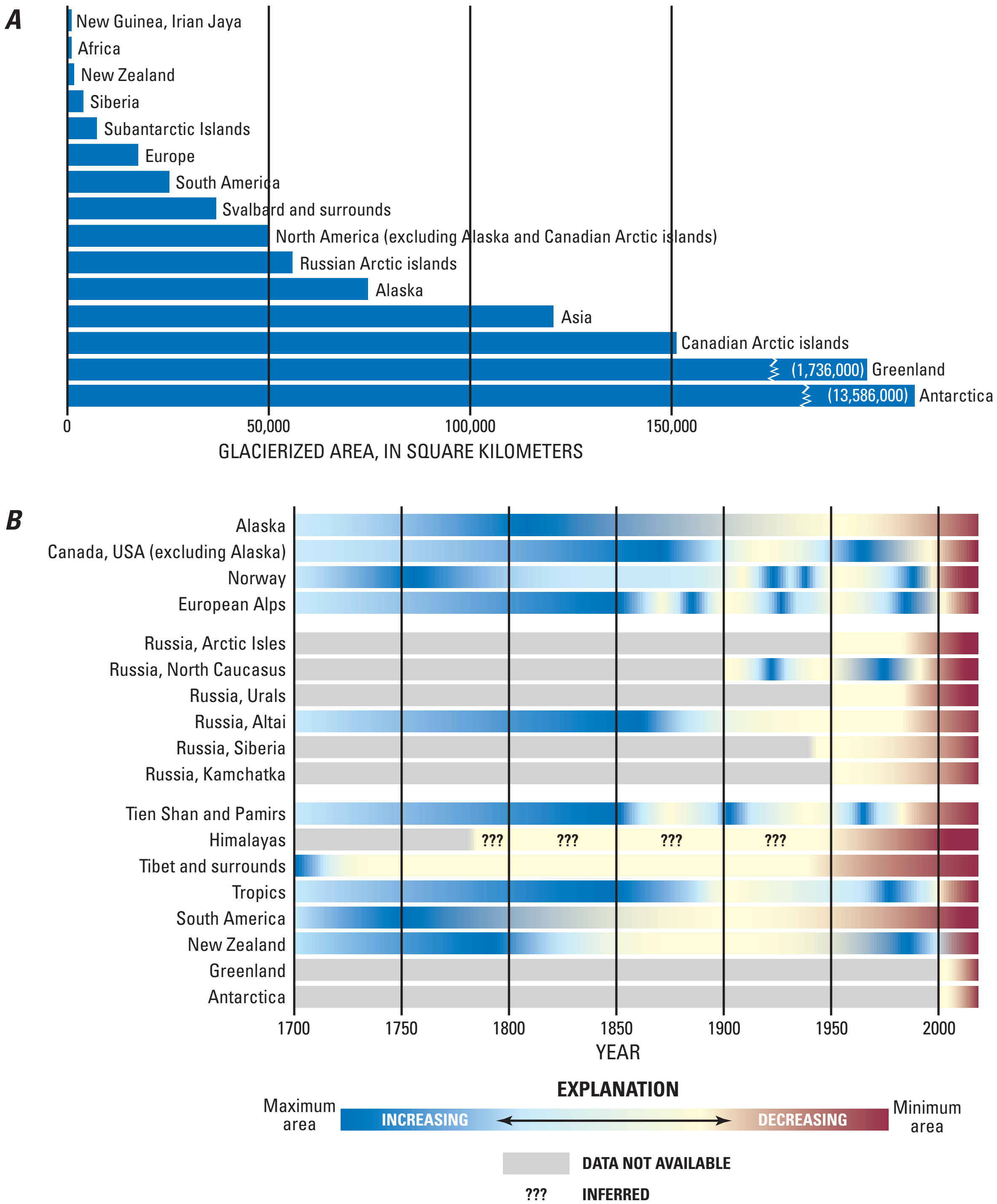


Figure 25.—**A**, *Extent of glaciers in the glacierized regions of the world, from the least glacierized (Irian Jaya, Indonesia, 7.5 km²) to the most glacierized (Canadian Arctic islands; 151,057 km²). The Antarctic and Greenland ice sheets were excluded from the original diagram; both ice sheets are added to this modified figure. At the same scale on which the bar graph for Svalbard reaches about 3 cm, a bar graph for the Antarctic ice sheet would have to be 10.45 m long and for the Greenland ice sheet 1.35 m long (see table 1). The area of the Antarctic ice sheet contains 85.32 percent of the Earth's glacierized area, and the area of the Greenland sheet contains 10.90 percent of Earth's glacierized area; both ice sheets together contain 96.22 percent of the Earth's glacierized area (see table 2). Only 3.78 percent of the area of the Earth's glaciers are non-ice-sheet glaciers. Modified from Zemp and others (2007, p. 132–133, fig. 6B.11a).* **B**, *Color-coded bar graphs showing decreasing glacierized areas in the main geographic regions. Changes in the area of the Antarctic ice sheet are seen primarily as retreat of ice shelves on the Antarctic Peninsula. Changes in the area of the Greenland ice sheet are seen as increased area of summer melt (water) on the surface of the ice sheet. Modified from Zemp and others (2007, p. 132–133, fig. 6B.11b).*

The loss of all the Earth's non-ice sheet glaciers would cause a maximum of about a 0.45 m rise in global (eustatic) sea level (table 3), only about 50 percent more than the total estimated 0.3 m rise in sea level from glacial meltwater and the steric expansion of volume of the oceans during the 20th century. There are signs that the two ice sheets, the Greenland ice sheet and Antarctic ice sheet, are being affected by the warming of the Earth's climate. Total melting of the Greenland ice sheet would cause a 6.5-m rise in global sea level (7.3 m according to Intergovernmental Panel on Climate Change [IPCC], 2007a). Total melting of the Antarctic ice sheet would add 73.4 m to global sea level (56.6 m according to the Intergovernmental Panel on Climate Change [IPCC], 2007a). Loss of land ice (ice above sea level) from either or both of these ice sheets would therefore be the source of significant increases in future sea level (table 3) even though regional differences would occur in the rise of sea level globally (Mitrovica and others, 2009). The Greenland ice sheet is undergoing longer periods and more extensive areas of seasonal melting (Hall and others, 2009) that vary from year to year (fig. 20), and the long-term trend is toward longer duration and an increase in melt area. Increasingly, a possible acceleration in loss of ice from the Greenland ice sheet is attracting the attention of glaciologists (Lubick, 2006; Velicogna, 2009) and decisionmakers (Nature, 2008).

In Antarctica, the most visible sign of change is dynamic thinning of its cryospheric coast (Pritchard and others, 2009) and rapid retreat in its fringing ice shelves, especially the complete loss of some of the ice shelves including Larsen A Ice Shelf, Wordie Ice Shelf (fig. 18), and Wilkins Ice Shelf that are on the Antarctic Peninsula (Ferrigno and others, 2006, 2008, 2009). The retreat of ice shelves on the Antarctic Peninsula has been the most obvious impact of climatic warming in the region, and the change is affecting ice shelves farther south along the peninsula and adjacent regions. Successive satellite images are documenting the breakup and retreat of the Wilkins Ice Shelf (Ferrigno and others, 2009).

The latest Fourth Report of Intergovernmental Panel on Climate Change (2007a, b) did not include much depth or detail concerning changes to the Greenland or the Antarctic ice sheets because of the limited scientific literature on measurements and modeling when the report was written. For example, the Intergovernmental Panel on Climate Change [IPCC] (2007b, p. 17) report stated, "Contraction of the Greenland Ice Sheet is projected to continue to contribute to sea level rise after 2100." Vaughn and Arthern (2007) addressed the difficulty in forecasting future changes in the Greenland and the Antarctic ice sheets; they said, however, that it was extremely important to reduce the current uncertainty. Long (2009), referring to changes in the Greenland ice sheet, said that improved knowledge of both contemporary and past changes is needed.

Two reports were prepared for the 2009 United Nations Climate Change Conference (Copenhagen Summit) held in Copenhagen, Denmark, on 7–18 December 2009, in order that attendees, science journalists, and the general public be updated on the latest climate science since the publication of the Fourth Assessment Report of the Intergovernmental Panel on Climate Change (IPCC) (2007a, b). "The Copenhagen Diagnosis, 2009: Updating the World on the Latest Climate Science" included several sections written by glaciologists on changes in the cryosphere (Allison and others, 2009). The United

Nations Environment Programme published several chapters in its "Climate Change Science Compendium, 2009," including chapter 2, "Earth's Ice," on the current state of ice globally (McMullen and Jabbour, 2009).

Glacial geologists and glaciologists have traditionally considered that the time frame in which ice sheets can lose significant volumes of ice is in terms of 10,000 years to 100,000 years (Sugden and John, 1976, p. 103). Truffer and Fahnestock (2007) disagree with that conventional view, arguing that glaciologists must reevaluate how quickly ice sheets can decrease their mass. Fairbanks (1989) shows that the loss of ice in the Laurentide, Scandinavian, and Patagonia ice sheets during the latter part of the last glacial was extremely rapid, with several rapid rises in global sea level ("meltwater pulses") (Fairbanks, 1990): for example, 24 m in 1,000 years as the ice melted (fig. 26). Fairbanks was writing about "temperate" ice sheets. The glacier ice in the southern part

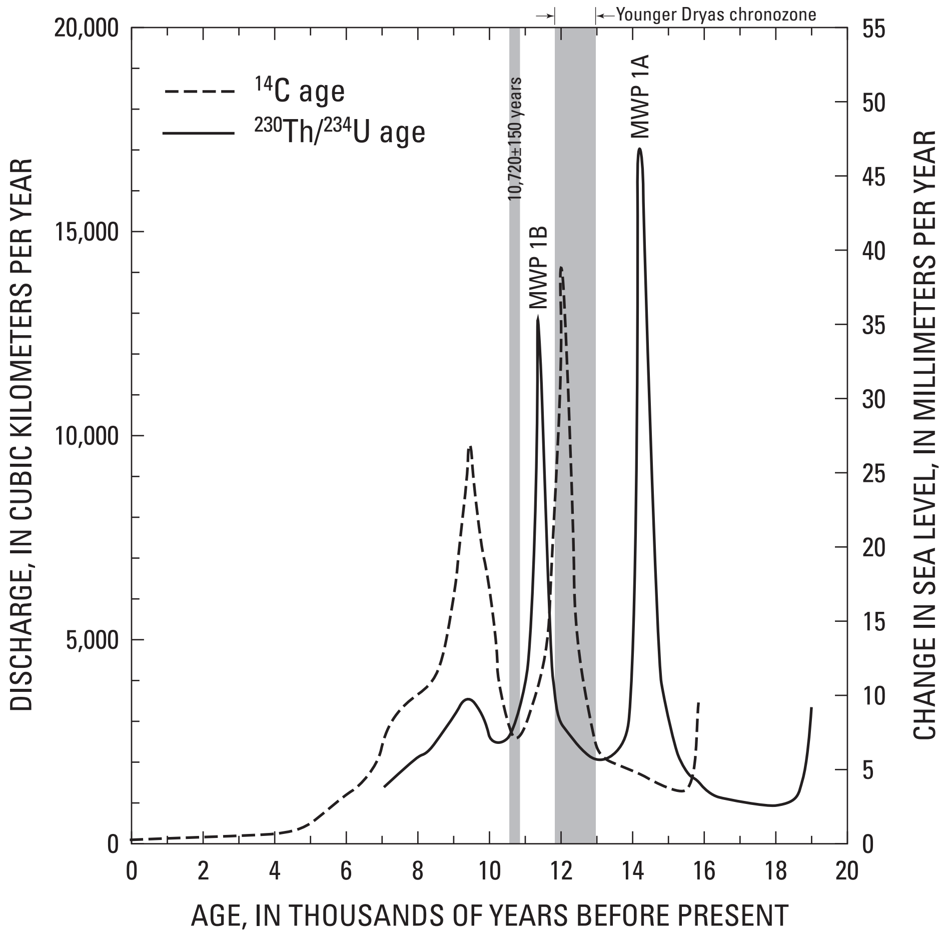


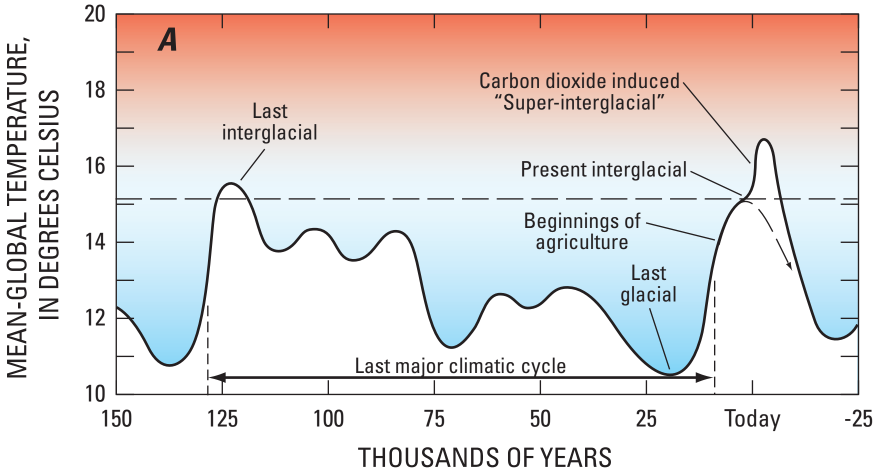
Figure 26.*—Annual discharge rate of glacial meltwater emanating from rapidly melting ice sheets at the end of the Pleistocene Epoch and near the beginning of the Holocene Epoch, plotted against the annual rate of rise in eustatic sea level. The ^{230}Th/^{234}U-dated sea-level curve is shown with a solid line. The ^{14}C-dated sea-level curve is shown by a dashed line; it is not corrected for changes in production of ^{14}C. The meltwater pulse IB coincides with the 1,787 m δ^{18}O anomaly in the ice core from the Dye 3 site in the Greenland ice sheet (10,720±150 years layer-counting calendar age) based on the ^{230}Th/^{234}U calendar chronology. The Younger Dryas chronozone is shown by a shaded pattern (see fig. 35). Modified from Fairbanks (1990, p. 943, fig. 3).*

of the Greenland ice sheet can also be considered to be "temperate" because the surface temperature of the southern margin is near 0°C during the summer (Hall and others, 2006).

Jonathan Overpeck, quoted by Lubick (2006, p. 30), stated "the Arctic may have reached a point where it will tip into a super interglacial state, a long warm period unlike anything experienced in the past 1 million years" (figs. 27*A* and 27*B*). Geologists often say that "the present is the key to the past," meaning that the analyses of processes taking place in the Earth System today (2012) can be used to describe how past geologic events occurred. In the context of the presence of glaciers on Earth today, however, glaciologists should perhaps say that "the past is the key to the future." An ephemeral ice sheet in Antarctica first appeared about 35 Ma (million years ago) in the late Eocene Epoch of the Cenozoic Era (figs. 28, 29). According to Katz and others (2008), the Earth System transitioned from a "greenhouse" to an "icehouse" climate in the late Eocene Epoch. Up to that point in the Cenozoic, the Earth had no ice sheets and, because of elevated concentrations of CO_2 in the atmosphere, probably had only non-ice-sheet mountain glaciers at high elevations. Ward (2009) points out that a CO_2 concentration of 1,000 ppm would probably produce an Earth with no glacier ice. The last time CO_2 was that voluminous was during the early Eocene (Climatic Optimum; in ~50 to 53 Ma) and the late Paleocene (Thermal Maximum; about 55 Ma) (fig. 29). The 2009 CO_2 concentration was 390 ppm, rising at 2 ppm each year, and may be accelerating. It is possible that a CO_2 concentration of 1,000 ppm could be reached in less than a century. A glacier-free Earth could result soon after 2100 (Ward, 2009).

It is likely that the potentially rapid loss of glacier ice from the Greenland ice sheet and from the Antarctic ice sheet during the next 100 years and beyond has enormous implications for the economic and societal impact of a eustatic rise in sea level far more rapid than most glaciologists envision. Ferrigno and others (2009) document the rapid reduction in area of the Wilkins Ice Shelf, southwest of the Antarctic Peninsula, and inferred that warming of coastal waters is causing rapid changes in the cryospheric coast farther south than the Wordie Ice Shelf on the west side of the Antarctic Peninsula (Ferrigno and others, 2008) and in the Larsen Ice Shelf on the east side of the Antarctic Peninsula (Ferrigno and others, 2006, 2008). Using GRACE data, Velicogna (2009) reports an acceleration in the loss of glacier ice from the Greenland and the Antarctic ice sheets. Pritchard and others (2009, p. 971) write that "dynamic thinning of glaciers now reaches all latitudes in Greenland, has intensified on key Antarctic grounding lines, has endured for decades after ice-shelf collapse, penetrates far into the interior of each ice sheet and is spreading as ice shelves thin by ocean-driven melt."

The current and rising concentration of CO_2 in the Earth's atmosphere is unmatched during the past 800,000 years. Hansen (2004, 2007a) clearly addresses the issue of rise in sea level as a "global warming time bomb," adding that "scientific reticence" (Hansen, 2007b) is partly to blame for not "emphasizing" the potential for a rapid rise in sea level as a major environmental issue. As will be presented later (p. A261–A264), the rise in global sea level can be considered a global glaciological hazard because of inundation of coastal lowlands and deltas and low-lying islands and displacement of human populations in such regions (pl. 1, lower right-hand map).



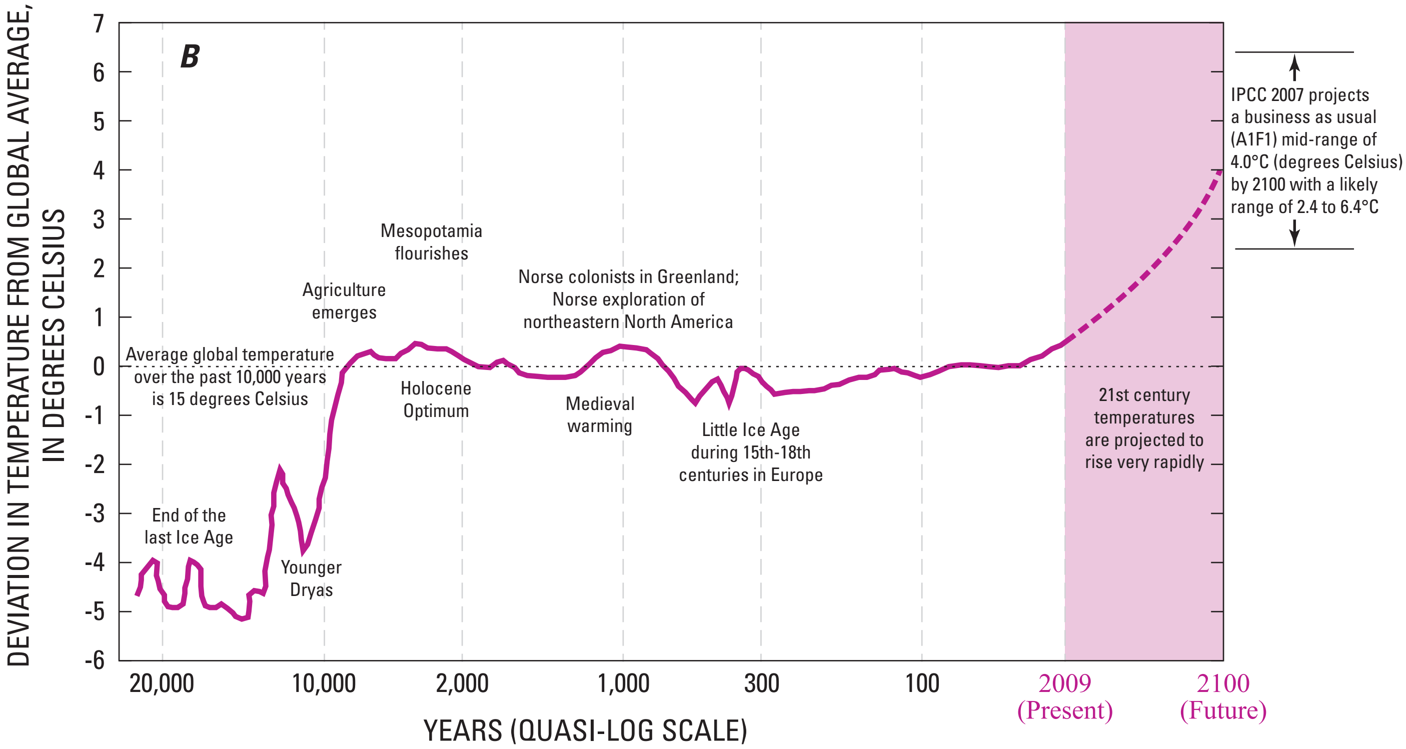


***Figure 27.*—A**, *Climate "forecast" for the next 25,000 years. A CO_2-induced "super-interglacial" at the end of the current interglacial may produce a much warmer climate, thereby delaying the natural onset of the next glacial. Modified from Imbrie and Imbrie (1979, p. 186, fig. 48).* **B**, *Past, current, and projected global temperature from about 20,000 years before the present to 2100 C.E. Modified from data published by the World Health Organization, the World Meteorological Organization, and the United Nations Environment Programme in 2003 (McMichael and others, 2003) and data published by the International Panel on Climate Change (IPCC) (2007a).*

History of Glacier Ice and Climate Change in the Earth System

Introduction

All planetary systems in the universe have a so-called "zone of habitability" around a single or binary star in which water can exist in its liquid phase (Ward, 2009). The distance and width of the "habitable zone" depend on the size of the star (Cowen, 2008). In the Solar System, the Earth, at about 150×10^6 km from the Sun, lies beyond the "habitable zone," the inner zonal boundary of which lies at 100°C, the outer zonal boundary at 0°C. Earth would be the "Ice Planet," not the "Water Planet," without its atmosphere.

Earth's atmosphere is constituted primarily by two principal non-greenhouse gases: nitrogen (N_2) at 78 percent and oxygen (O_2) at 21 percent. The concentration of greenhouse gases in the Earth's atmosphere—such as water vapor (H_2O), carbon dioxide (CO_2), and methane (CH_4)—constitutes only about 1 percent of the atmosphere, but these gases elevate the Earth's mean surface temperature to about 15°C. Carbon dioxide constitutes only about 0.0003 percent (390 ppm in 2009), but its natural variability over time can produce significant changes in the Earth's mean surface temperature; it either lowers the temperature (less CO_2) or raises it (more CO_2).

The concentration of greenhouse gases in the Earth's atmosphere permits the presence of water on ~71 percent of the Earth's surface. Under what conditions, geologists ask, and how often has ice formed on Earth during its 4.6 billion year (Ga) history? In other words, when have one or more elements of the cryosphere (glaciers, snow cover, sea ice, and permafrost) been present on Earth? Four key factors contribute to the answer: (1) the concentration of CO_2 in the Earth's atmosphere (Royer and others, 2004); (2) the latitudinal position (Nield, 2007) and topographic elevation of the Earth's continents (Ruddiman and Kutzbach, 1991); (3) the volume and global geometry of cool and warm oceanic currents (Broecker, 1991); and (4) the astronomical cycles of predictable variations in the shape of the Earth's orbit around the Sun. These latter variations are (a) eccentricity—from near-circular to elliptical, a 100,000-year cycle; (b) obliquity—tilt of the Earth's axis of rotation with respect to the ecliptic plane, its obliquity, which varies between 21.8° and 24.4°, a 41,000-year cycle; and (c) precession—Earth's closeness to the Sun in the Northern and Southern Hemispheres—of the equinoxes, a 23,000-year cycle) (Imbrie and Imbrie, 1979; Mayewski and White, 2002).

In the geologic past, continents could receive seasonal *snow cover* at high elevations and in polar locations, and prolonged periods of cold temperatures could cause *permafrost* to form. Seasonal *sea ice* was likely to form in polar regions, poleward of the Arctic and the Antarctic circles of latitude, where no solar insolation is received during six months each year, unless warm oceanic currents keep polar oceans at temperatures greater than freezing. However, none of these three (of the four) elements of the cryosphere is likely to leave any evidence of colder paleoenvironments in the Earth's geologic record. Among the three, only evidence of past permafrost conditions can be studied in that record; even so, the likelihood of preservation is low.

But glaciers, the fourth element of the cryosphere, leave abundant evidence of their occurrence in the Earth's deep past. That record includes changes in eustatic sea level caused by variation in the volume of glacier ice on land and by deposition of tills or tillites (indurated tills) (Hambrey and Harland, 1981) and other glacial deposits, such as varves and glacial deltas. Glaciers also erode the landscape—for example, leaving striations on rocks and excavating fjords.

There is geologic evidence for at least three extended intervals of continental glaciation during the Proterozoic Eon (Cryogenian Period of the Neoproterozoic Era), the Paleozoic Era (Carboniferous-Permian periods) and the Cenozoic Era (Imbrie and Imbrie, 1979; Martini, 1997). Each of these continental glaciations represents an "Ice Age," but the geologic evidence for pre-Cenozoic ice ages is fragmentary, increasingly so as one goes further back in time (Crowell, 1982). Although the sections to follow discuss the "Snowball Earth" hypothesis (Proterozoic Eon), the Permian Ice Age, and the Cenozoic Ice Age, greatest emphasis is given to the Cenozoic at the end of the Eocene and beginning of the Oligocene, 33.7 Ma. At that time, the Earth "changed from a greenhouse world (which was generally ice-free) to an icehouse world (which was heavily glaciated [sic, glacierized])" (Pekar, 2008, p. 603). Since the late Paleocene Thermal Maximum (about 52 Ma) and a subsequent warm interval during the Eocene Climatic Optimum (about 50 Ma), Earth has undergone a decrease in global temperature (with a few intervals of warming) and the formation of ice sheets (Zachos and others, 2001). During the Pleistocene and Holocene, Earth has also undergone numerous cycles of glacials (glacial intervals) and interglacials (intervals between glacials) in what is broadly referred to as "the" Ice Age.

Divisions and Subdivisions of Geologic Time

Geologists divide the Earth's 4.6 Ga (billion years = giga-annum or 10^9 years) history into four Eons, from oldest to youngest (Orndoff and others, 2007) (fig. 28): the Hadean (>4 Ga), the Archaean (~4 Ga to 2.5 Ga), the Proterozoic (2.5 Ga to 542 Ma), and the Phanerozoic (542 Ma to the present). The Archean is further divided into four Eras, and the Proterozoic is divided into three Eras (Neoproterozoic, Mesoproterozoic, Paleoproterozoic). The Phanerozoic is also divided into three Eras: Paleozoic (542 Ma to 251 Ma), Mesozoic (251 Ma to 65.5 Ma), and Cenozoic (65.5 Ma to present) (fig. 28). Each Era is further divided into Periods; and the Periods of the three Paleozoic, Mesozoic, and Cenozoic Eras are further subdivided into Epochs. The Ediacaran Period (630 Ma to 542 Ma), the youngest period of the Neoproterozic Era, and the Periods and Epochs of the Phanerozoic, are defined by the fossils found within the sedimentary rocks of each.

Geologists attribute the general decrease in temperature of the Earth during the Cenozoic Era (the most recent 65.5 Ma) (fig. 29) to a decrease in the concentration of CO_2 in the Earth's atmosphere (Ward, 2009). The present discussion emphasizes the onset of glaciation during this interval of time. The Cenozoic Era is conventionally subdivided into two periods: the Tertiary Period (from 65.5 Ma to 1.8 Ma) and the Quaternary Period (from 1.8 Ma to the present). The Quaternary Period is divided into two epochs: the Pleistocene Epoch (1.8 Ma to 11,500 years ago) and the Holocene Epoch (11,500 years ago to the present) (fig. 28).

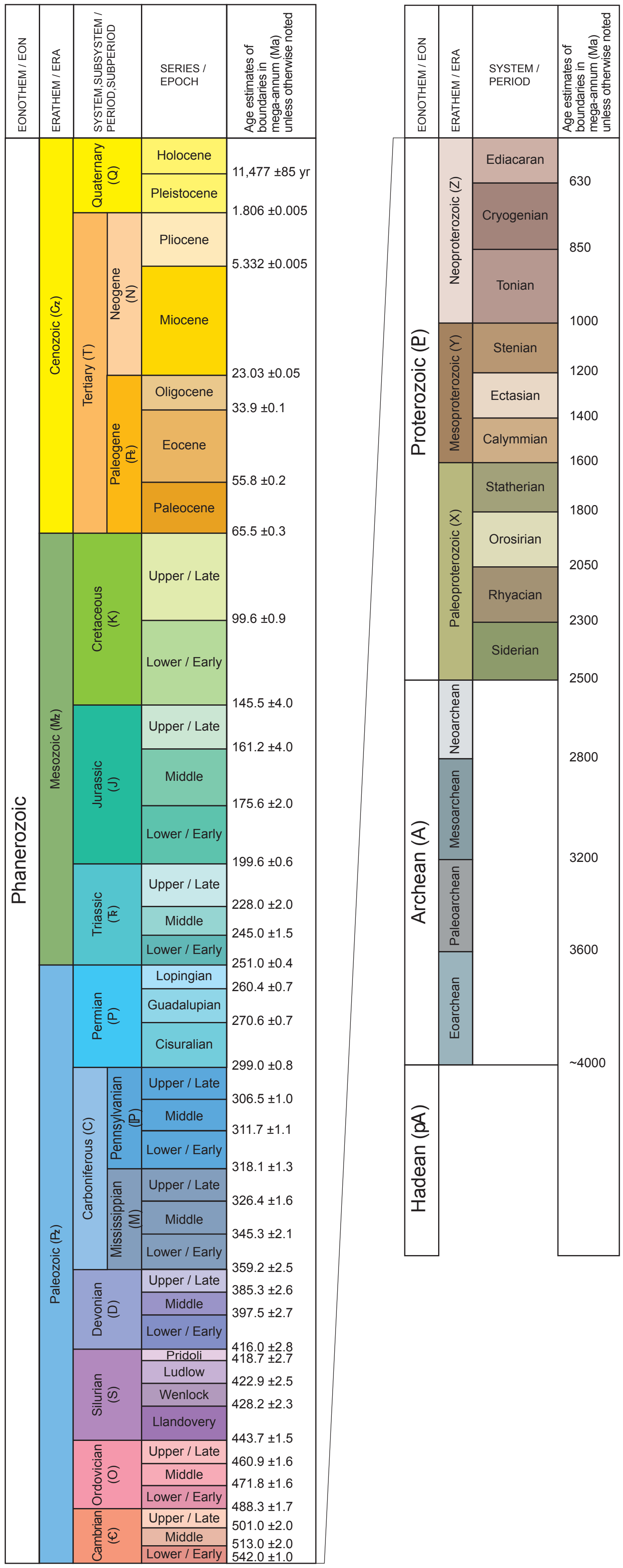


Figure 28.—*Geologic time scale from the beginning of the Hadean Eon (4+ Ga) to the present, showing subdivisions into eons, eras, periods, and epochs. Age of start and ending of each interval is modified from Orndorff and others (2007).*

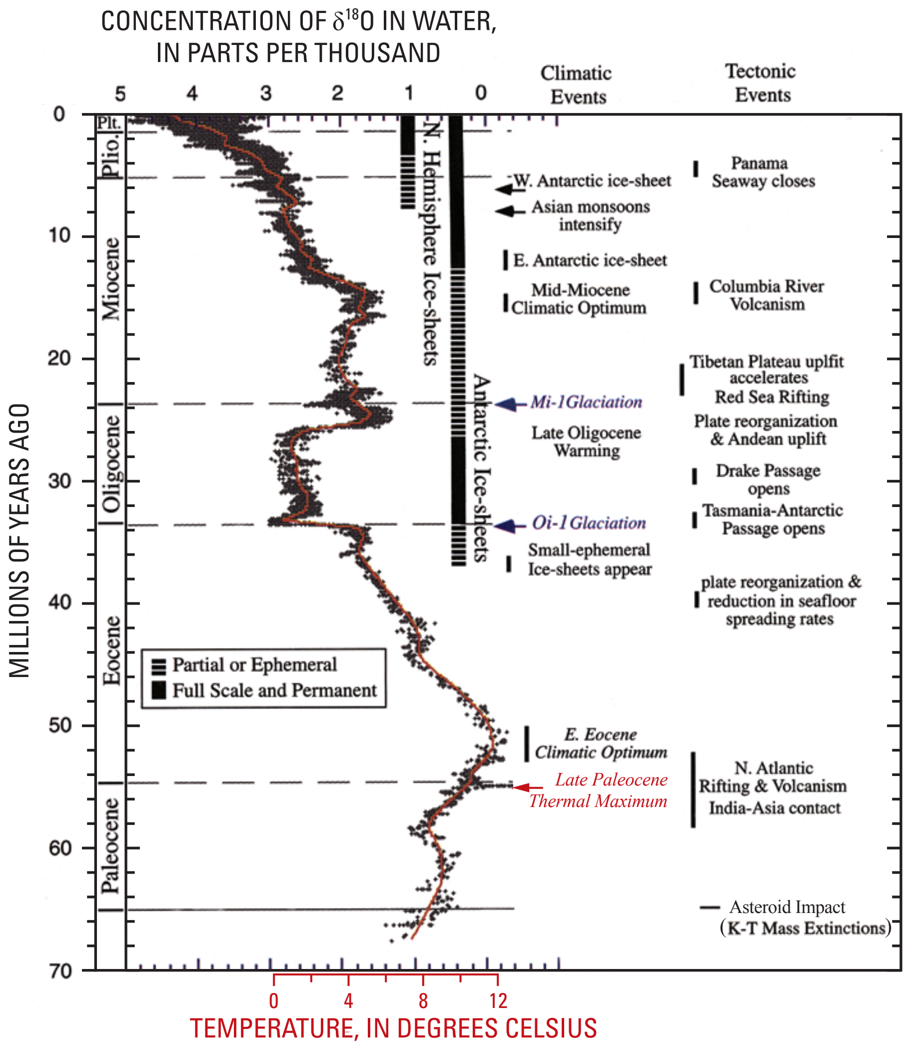


Figure 29.—*Proxy temperatures [$\delta^{18}O$ (‰)], climatic events, and tectonic events during the Cenozoic Era. A long-term trend of decreasing global temperatures is apparent, along with a change from global "greenhouse" climate to global "icehouse" climate, resulting in increased glaciation, starting in the late Eocene. Modified from Zachos and others (2001, p. 688, fig. 2).*

Some geologists, such as Gould (1991), argue that the Holocene Epoch should be deleted from the geologic time scale because the multiple cycles of glacials and interglacials during the Pleistocene did not end with the last major phase of melting of the continental ice sheets in North America, Eurasia, and South America: the Earth is still in an ice age. Other geologists (Zalasiewicz and others, 2008) argue that the Holocene is only the latest of many late Quaternary interglacials and the only one conferred with the status of an Epoch. They further express a need to define a new epoch (Anthropocene) to mark the point at which humans began to have a measurable impact on the Earth System. Interestingly, significant areas of agriculture were first established 11,500 ago, the beginning of the Holocene (fig. 26). Modifications of the Earth's surface for agriculture caused significant changes in natural land cover (for example, deforestation) requiring engineering of the landscape (for example, by constructing dams and redirecting streams and channels to irrigate crops) (Diamond, 1997; Ruddiman, 2005). Perhaps a more comprehensive solution would be to delete the Holocene Epoch as a stratigraphic subdivision, make the Pleistocene Epoch synonymous with the Quaternary Period (1.8 Ma to the present), and add the Anthropocene Epoch as synonymous with the time interval of measurable human impact on one or more components, elements, processes, or cycles that comprise the Earth System. Atmospheric chemist Paul Crutzen made this suggestion in 2000 (Johnson, 2009).

Neoproterozoic Glaciation ("Snowball Earth")

Evidence of glaciation on Earth first appeared in the Cryogenian Period (850 Ma to 630 Ma) of the Neoproterozoic Era. Kopp and others (2005) suggested an even earlier glaciation. Kennedy and others (1998) postulated two or four continental glaciations during the Neoproterozoic. The precise number of glaciations remains speculative (Crowell, 1982). Hoffman and others (1998) postulated a theory that Lubick (2002) deems controversial: that the mean annual temperature at the Earth's surface dropped low enough to cause a large accumulation of glacier ice on Rodinia continents clustered along the Equator (fig. 30). This temperature would be low enough to freeze the Earth's oceans (Kirschvink, 1992; Hoffman and Schrag, 2000). Hoffman and Schrag (2000) coined the term "Snowball Earth" and proposed four stages: Snowball Earth Prologue (stage 1); Snowball Earth at its Coldest, with mean surface temperatures of -50°C (stage 2); Snowball Earth as it Thaws (stage 3), and; Hothouse Aftermath (stage 4). Various hypotheses have been proposed to account for such glaciation. Donnadieu and others (2004) proposed continental break-up. Kopp and others (2005) and Rasmussen and others (2008) proposed that evolution of eukaryotes (organisms whose cells contain a nucleus) and of cyanobacteria (bacteria that obtain their energy through photosynthesis), which together initiate oxygenetic photosynthesis, can account for these glaciations. Other scientists suggest that, although continental glaciation covered all land areas, some oceans remained free of ice (Allen and Etienne, 2008). Kopp and others (2005) stated that an earlier "Snowball Earth" occurred during the Paleoproterozoic Era, ~2.3–2.2 Ga, which caused a global glaciation following initiation of oxygenetic photosynthesis. Walker (2003) proposed that "Snowball Earth" provided the necessary environmental conditions for the inception of primitive organisms.

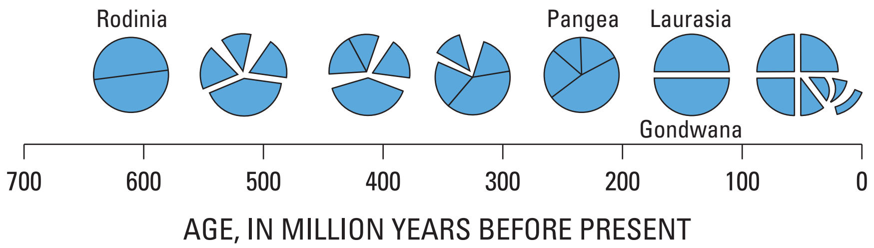


***Figure 30.**—Schematic diagram of continental spreading from the Neoproterozoic Era to the present time (Cenozoic Era). The coalesced Rodinia supercontinent (Neoproterozoic Era) was followed by the breakup of the Rodinia supercontinent during the middle Paleozoic Era, the recoalescence into the Pangea supercontinent (middle Paleozoic Era), and the subsequent breakup of the Pangea supercontinent into Gondwana in the southern Hemisphere and Laurasia in the Northern Hemisphere during the late Paleozoic Era. Subsequent fragmentation and continental drift produced the current configuration of the continents. Modified from Servais and others (2009, p. 5, fig. 11B).*

Late Paleozoic Glaciation (Carboniferous-Permian Periods)

Crowell (1982) describes continental glaciation during the Late Ordovician and Early Silurian Periods and even during the Middle Devonian. The best documented widespread continental glaciation, however, with glacial deposits and other evidence for glaciation on all of the continents that made up the Gondwana supercontinent (fig. 30), occurred during a 50-M.y. interval during the Late Carboniferous and Early Permian. Visser (1997) points out that the Madagascar and African-Arabian-Peninsula components of the Gondwana supercontinent covered 35 percent of the surface land area. Rocks in Africa provide the best evidence of widespread Late Paleozoic continental glaciation. Evidence of glaciation during the Late Carboniferous–Early Permian interval is also found in parts of South America and Antarctica (Visser, 1997; Ziegler and others, 1997).

About 300 Ma, the Earth's continents moved together to create a supercontinent called "Pangea" (fig. 30). The supercontinent extended from the North Pole to the South Pole, thereby providing highland areas on which ice sheets formed, and they radiated outward onto the accreted continental landmasses described above. The Pangean supercontinent began to break apart soon after formation, so that at the end of the warm Mesozoic Era, about 90 Ma, today's continental configuration began to develop (fig. 30), although the continents were much closer to each other. Scientists have found evidence of glaciations even during the Mesozoic, however. During the Late Cretaceous Period, a continental ice sheet covered an area estimated to have been 6 million km^2 (Bornemann and others, 2008).

Cenozoic Era Glaciation

By the early part of the Cenozoic Era, the drifting continental fragments of Pangea had moved away from each other and formed the Atlantic Ocean. During this movement, some fragments collided with one another. For example, Africa collided with Europe and eventually formed the Alps. India collided with Asia, which elevated the Tibetan Plateau and the Himalaya. In contrast, Antarctica split from Australia, which rafted to the northeast, whereas Antarctica stayed centered at the South Pole. Subducting oceanic plates of the eastern Pacific Ocean descended beneath and uplifted continental plates to the east. North-south-oriented mountain ranges bordering North America (Coast Ranges) and South America (Andes) formed. This produced extensive volcanism along the Pacific margin of North and South America.

The global climate of the early Cenozoic Era was warmer than that of the late Mesozoic. A short-lived Thermal Maximum peaked in the Late Paleocene (about 54 Ma) and a longer Climatic Optimum developed in the early Eocene (53 Ma–50 Ma) (Zachos and others, 2008) (fig. 29). According to Sluijs and others (2007), the late Paleocene Thermal Maximum was caused by warming of ocean-bottom water. Methane clathrates in sedimentary deposits of the sea floor may have disassociated and released large amounts of methane (CH_4) into the Earth's atmosphere. Methane is about 20 times more effective than CO_2 as a greenhouse gas in warming the Earth's surface.

At the beginning of the Eocene (55.8 Ma), the Earth had a "greenhouse" climate. As the Eocene progressed, mean global temperatures cooled, slowly transforming the Earth to an "icehouse" climate (McGowran, 1990; Prothero and others, 2003; Katz and others, 2008; Koeberl and Montanari, 2009). By the beginning of the Oligocene (33.7 Ma), the Earth was at the beginning stages of an "icehouse" climate, the result of a decrease in CO_2 in the Earth's atmosphere (Tripati and others, 2005). Recent research suggests that sea ice in the Arctic Ocean first appeared about 47.5 Ma (Perkins, 2009b). The downward temperature trend was not linear, however, but was interrupted by rebounds to somewhat warmer temperatures. Besides atmospheric and oceanic effects, climatic events appear to have been related to major tectonic events during the Cenozoic (fig. 29). Zachos and others (2001, p. 686) concluded from analyses of deep-sea sediment cores that during the 65 Myr of the Cenozoic Era, "Earth's climate has undergone a significant and complex evolution… This evolution includes gradual trends of warming and cooling driven by tectonic processes on time scales of 10^5 to 10^7 years, rhythmic or periodic cycles driven by orbital processes with 10^4- to 10^6-year cyclicity, and rare rapid aberrant shifts and extreme climate transients with durations of 10^3 to 10^5 years."

As the Cenozoic progressed, continents continued to move toward their present configuration, periods of extensive volcanism occurred, and the Andes and the Tibetan Plateau were elevated. The Antarctic continent was isolated as the Southern Ocean formed around it. Cyclical perturbations of Earth's orbit also affected global climate: eccentricity (100 Ka-year cycle), obliquity (41 Ka-year cycle), and precession (23 Ka-year cycle). The concentration of atmospheric greenhouse gases, primarily CO_2 (Pekar, 2008) but also CH_4, began to change. The combined effects of all these processes led to renewed continental glaciations. Glaciation had been absent since the end of the Permian Period, a hiatus of about 230 Myr.

According to Zachos and others (2001), the first significant glaciation began in Antarctica about 34 Ma during the late Eocene (Haug and others, 2004) (fig. 29). Barrett (2003) states that a decrease in CO_2 in the Earth's atmosphere brought about this change. Late Oligocene warming, about 25 Ma, reduced the glaciation of Antarctica, a reduction that continued through the middle Miocene Climatic Optimum, about 15 Ma. During the Miocene, about 11.5 Ma, the Antarctic ice sheet expanded (Holbourn and others, 2005) and became permanent, first in East Antarctica (11 Ma) and later in West Antarctica (about 7 Ma). Glaciation in the Northern Hemisphere began by 8 Ma, in the late Miocene. In the middle Pliocene Epoch, 3.3 to 3.0 Ma, global temperatures rose significantly to 2° to 3°C warmer than at present (2009) (Robinson and others, 2008). Even during the mid-Pliocene warming, however, the ice sheet in Antarctica persisted, as it had since the middle Miocene Epoch (about 14 Ma) (Kennett and Hodell, 1995).

During the late Pliocene, the Panamanian seaway between North and South America closed, thereby uniting the two continents and altering oceanic circulation between the eastern Pacific and North Atlantic Oceans. The closing of this "Panama Gap" about 3 Ma (Schmittner and others, 2004), coupled with the continued isolation of Antarctica and the development of the Antarctic circumpolar current, strengthened the Great Ocean Conveyor Belt (Broecker, 1991) (fig. 31). This group of connected global oceanic currents transported heat from tropical regions to polar regions and generated cold saline currents

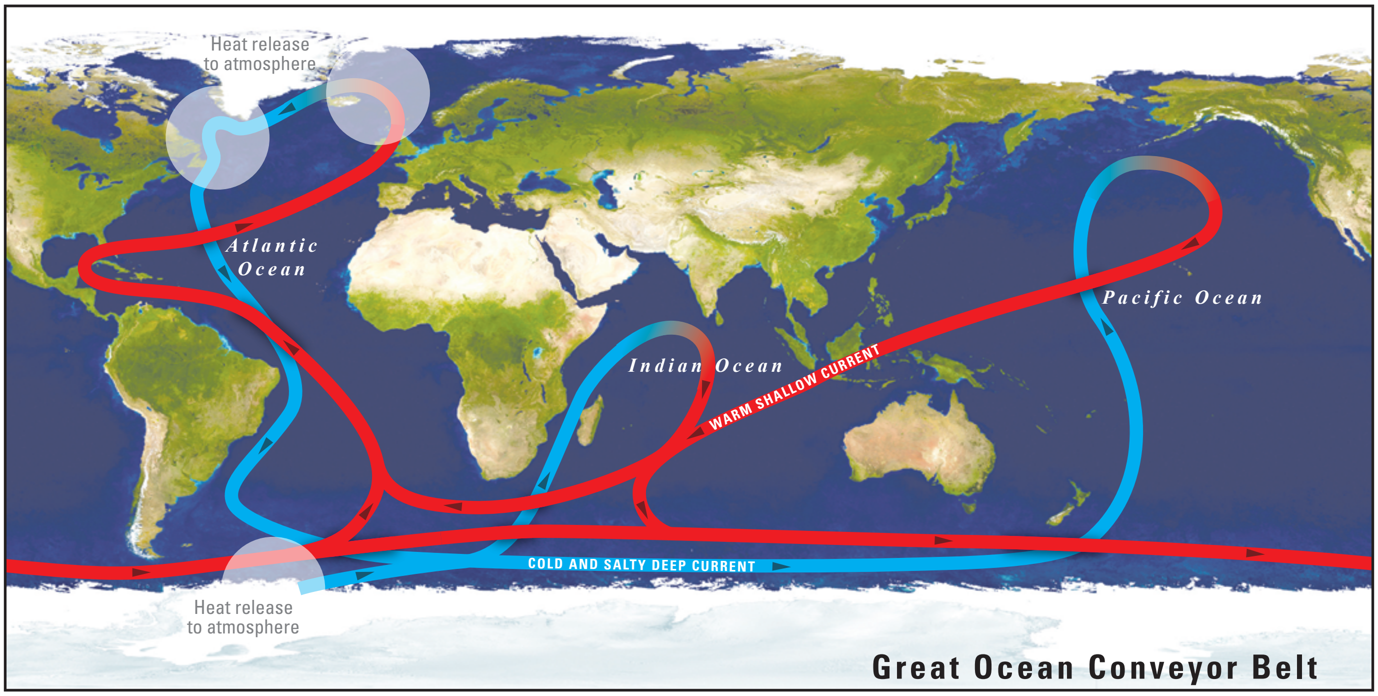


Figure 31.—The great ocean conveyor belt of global ocean currents as described in Broecker (1991). Winds drive warm salty ocean currents in the global pattern of atmospheric circulation. Note that the flow of warm currents is relatively unimpeded in the Pacific Ocean. In the Atlantic Ocean, however, México and Central America block the westward flow, forcing the current northward (Gulf Stream). The cold currents in the polar regions are denser than warm equatorial waters, and therefore they sink, forming cold deep water. Atlantic deep water forms near Greenland, travels to Antarctica, adds cold salty water from Antarctica, and then continues into the Pacific Ocean. Modified from Intergovernmental Panel on Climate Change (IPCC) (1996, p. 271, fig. 2); after Broecker (1991).

that flowed equatorward from both polar regions (fig. 31). The closing of the Panamanian seaway between North and South America produced a continental geometric configuration that initiated the Gulf Stream, and it set the stage for Pleistocene glaciation, including the formation of the Greenland ice sheet (Schmittner and others, 2004) 2.7 Ma (Haug and others, 2004). According to Lunt and others (2008), a reduction in atmospheric CO_2 was also a requisite for the Greenland ice sheet to form.

By the late Pliocene, the global geographic distribution of the continents was similar to that of today. In Antarctica, the continent with the highest average elevation (about 2,500 m), the Antarctic ice sheet was centered on the South Pole, which produced cold, saline ocean bottom water. Continental land masses with mountain ranges were concentrated mainly in the Northern Hemisphere, except for the Andes in South America. The Great Ocean Conveyor Belt dominated global oceanic circulation. Reduction in the concentration of CO_2 in the atmosphere (DeConto and others, 2008) and tectonic uplift of extensive highland regions, such as the Tibetan Plateau (Ruddiman and Kutzbach, 1991), the Himalaya, and the western ranges of North and South America, disrupted global atmospheric circulation. The combination of all these factors set the stage for repetitive 100,000-year glacial cycles, which lasted about 90,000 years, and short interglacials (about 10,000 years each), in which orbital eccentricity, obliquity, and precession began to dominate global climate.

During the late Pliocene and through the Quaternary (Pleistocene and Holocene), Ice Age evidence of multiple glacial-interglacial cycles were recorded in sedimentary beds deposited on the sea floor. Until 1 Ma, most glacial-interglacial cycles lasted 41,000 years. For the past 1 Myr, however, 100,000-year cycles have dominated (Paillard, 2006) and those of the past 800,000 years are recorded in ice cores from Antarctica (Jouzel and others, 2007). These paleoclimatic cycles show a slow onset of continental glaciation in North America, South America, and Eurasia and a strengthening of glaciation in Antarctica and Greenland, followed by rapid deglaciation of continental ice sheets in North America, South America, and Eurasia (fig. 32). Changes in the Earth's orbital eccentricity appear to have caused the 100,000-year cycles, but in the view of Huybers and Wunsch (2005, p. 491), "the timing of Late Pleistocene glacial terminations [can be statistically proved] purely in terms of obliquity forcing." Broecker and Denton (1990) and Kawamura and others (2007) showed that increased summer insolation in the Northern Hemisphere caused the rapid deglaciation (fig. 33). Although scientists now have a clear record of glacial-interglacial cycles and of the slow growth followed by rapid decay of continental ice sheets during the Ice Age, a complete understanding of causes is still to be obtained (Raymo and Huybers, 2008; Huybers, 2009). For example, Siegenthaler and others (2005) report a stable carbon cycle—climate relationship during the late Pleistocene. The ice sheets in Antarctica and Greenland persisted during the interglacials, although the Greenland ice sheet probably melted during the last interglacial, while retaining only smaller ice masses along its mountainous divide (Koerner, 1989; fig. 19, p. A114).

Even though the last 12,000 years is called an "Epoch" (Holocene Epoch), independent of the Pleistocene Epoch, the Earth still remains in the grip of the 2.7-Myr-long late Cenozoic glacial (Ice Age) which began in the late Pliocene. The Holocene, the latest interglacial in a series of glacial-interglacials, is the only interglacial that has been given the status of an "Epoch." As can be seen in Chapters B–K of this volume, glaciers are still present on all continents

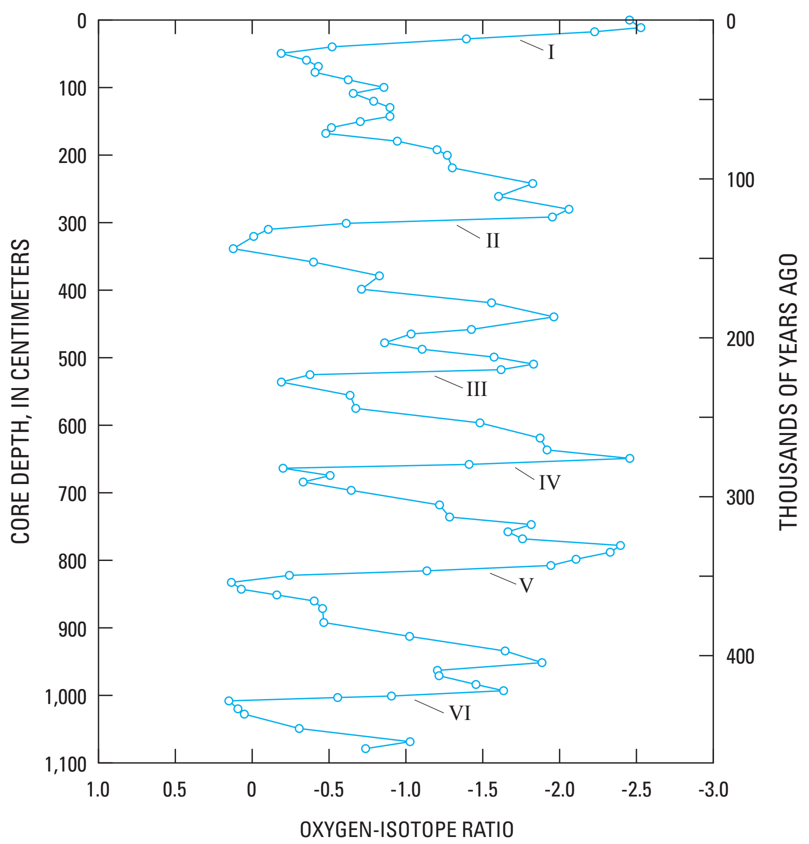


__Figure 32.__—Variations in the oxygen-isotope ratio of a marine sediment core showing the 100,000-year cycles of glacials and interglacials. The curves record the slow onset of glaciation and six intervals of rapid deglaciation at the end ("termination") of each cycle. Modified from Imbrie and Imbrie (1979, p. 157, fig. 38).

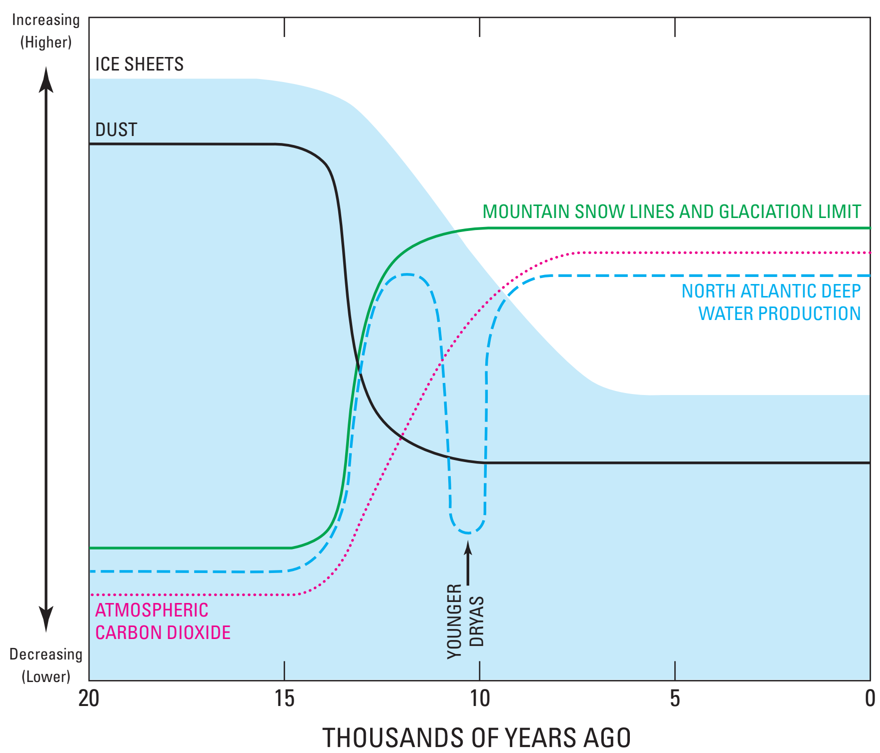


__Figure 33.__—Schematic diagram showing some of the changes in the Earth System during the transition between the end of the Pleistocene Epoch and the beginning of the Holocene Epoch. These changes decreased dust in the Earth's atmosphere, whereas CO_2 concentration and production of deep water in the North Atlantic Ocean increased. Elevation of the snow line (and therefore the glaciation limit) in mountain ranges increased, signifying the retreat of mountain glaciers, of ice caps, and of snowfields and the melting of continental ice sheets in North America, Eurasia, and South America. A sudden short interval of cooling is indicated by the Younger Dryas. Modified from Broecker and Denton (1990).

except Australia (fig. 2). Present-day Earth has an active cryosphere. Glaciers (two ice sheets and an estimated 160,000 other glaciers, including 70 large ice caps (Meier and Bahr, 1996)) cover about 11 percent of the Earth's surface. Seasonal snow cover; seasonal expansion and contraction of the area of sea ice in the polar regions; seasonal lake and river ice; and permafrost in polar regions and high-mountain areas are prevalent. Planet Earth, therefore, currently remains in a partial glacial, even though it is reduced in scope from a full glacial.

Latest Cenozoic Glaciation

Geologists traditionally used the terms Ice Age and Pleistocene Epoch interchangeably, because field work during the late 19th century and first half of the 20th century in Europe and North America found geological and geomorphological evidence in the landscape that described "four" glacials and three interglacials. They used "relative dating" by mapping the position of terminal moraines, erratic boulders, and other glacial geologic evidence. Geologists used "relative" positions of deposits ("younger" on top, "older" underneath; law of superposition) because, until radiometric-dating techniques were developed, it was usually not possible to determine absolute dates of deposits and strata. To support archaeological research in the American Southwest in the early 20th century, Andrew F. Douglas developed tree-ring chronologies (dendrochronolgy). The Laboratory of Tree-Ring Research at the University of Arizona contains many thousands of cross-correlated tree-ring records. Those records enable paleoclimatologists and geoarchaeologists to accurately date past climate changes on the basis of variations in width and density of rings (caused by temperature and precipitation) for trees growing in temperate latitudes. This allows annual resolution to pre-Columbian time (several hundred years) and, in a few cases, several thousand years.

The capability to date older glacial events did not develop until 1949 when Willard F. Libby, 1960 Nobel Prize-winning physical chemist, developed the carbon-14 (^{14}C)-dating technique. He demonstrated that the amount of ^{14}C isotopes in terrestrial and marine fossil organisms could be measured precisely. The half-life of the ^{14}C isotope is 5,730 years; so if only one-quarter of the amount of ^{14}C in a once-live organism remains, it died 11,460 years ago. Modern Accelerator Mass Spectrometer (AMS) ^{14}C dating can measure very small amounts of residual ^{14}C isotopes, thereby extending ^{14}C dating back 40 Ka (and somewhat longer in some instances) to the late Pleistocene. During the second half of the 20th century, many other radiometric dating techniques (for example, ^{238}U, ^{230}Th) became available. Also, isotopes of oxygen preserved in marine fossils permit scientists to determine past sea-surface temperatures.

In 1956, the first glacier ice cores were obtained from the Greenland ice sheet (Langway, 2008). During the subsequent 50 years, an ice-core record of the last 130,000 years has been obtained from Greenland. Ice cores from Antarctica currently span the last 800,000 years. Numerous ice cores also have been obtained from high-mountain glaciers in temperate and tropical regions. Some of these records extend back 20,000 years, but they are more usually only a few thousand years (see p. A155–A171). Concentrations of CO_2 and of NH_4^+ in these cores allow estimates of Pleistocene atmospheric and sea-surface paleotemperatures (paleoenvironmental parameters).

In 1964, the Deep Sea Drilling Program (DSDP) was started by the Joint Oceanographic Institution for Deep Earth sampling (JOIDES) with funding from the National Science Foundation (NSF). This drilling program was directed at obtaining sediment cores from the sea floor. In 1984, the DSDP was renamed the Ocean Drilling Program (ODP) and focused on those areas of thickest sediments from which lengthy geological records could be obtained. Scientists were able to correlate the ^{18}O isotope content in these cores ($\delta^{18}O$ water) with water temperature, because many marine organisms, such as foraminifers, incorporate the ^{18}O isotope into their calcite ($CaCO_3$) shells. As a consequence, geologists could use the calcite $\delta^{18}O$ to determine past ocean-water temperature (paleotemperature).

Emiliani (1961) constructed a graph showing variation in oceanic paleotemperatures, which could be organized into Marine Isotope Stages (MIS). Subsequent research showed paleotemperature variation during most of the Plio-Pleistocene Ice Age. The geologist Nicholas Shackleton (1969) determined that the $^{18}O/^{16}O$ oxygen ratios could also be used to show that the concentration of the heavier oxygen isotope (^{18}O) in the ocean was higher during glacials than during interglacials. Marine Isotope Stage 1 (MIS 1) indicates the Holocene Epoch; the last glacial spanned MIS 2, 3, and 4; MIS 5 indicates the last interglacial. Shackleton (1969) correlated MIS 5e (the oldest substage of MIS 5) with the Eemian interglacial (the Sangamon interglacial in North America), which glacial geologists first described from moraines and other glacial deposits on land. MIS 5e is dated at about 125,000 years ago. During MIS 5e, ^{18}O content was lower in the ocean, and sea level was about 6 m higher than today, probably the result of extensive melting of the Greenland ice sheet. The ^{18}O content in the ocean also provided a means for determining the volume of glacier ice on land during a glacial.

As with many advances in science, access to new data (such as from ice cores) and the invention of new devices (such as the Hubble Space Telescope) and new analytical tools (radiometric dating) permitted scientists to make significant advances in knowledge. The confluence of new data from ice cores and marine sediment cores with the many new analytical techniques that are being developed has ignited a revolution in new knowledge about climate variability during the Plio-Pleistocene and Holocene Epochs.

Stokstad (2001) summarized six primary methods currently used to analyze paleoclimates: tree rings, pollen, geomorphology, ice cores, corals, and marine sediments. Table 9 shows the time interval to which each technique can be applied, the scientific discipline(s) involved, the nature of the information derived, and the relevance of the derived information to paleoclimate change. Key findings relevant to paleoclimate variability, and changes in the Earth System during the Ice Age (especially sea-level change, and other observed changes underway at present), will be summarized in the discussion that follows.

Quaternary geologists, glaciologists, and paleoclimatologists now have an ever increasing understanding of paleoclimate and paleoenvironmental changes, including variations in: volume of glaciers on land, sea level, CO_2 and CH_4 in the atmosphere, and paleotemperature. Of particular interest to glaciologists and paleoclimatologists is the previous interglacial, the last glacial, and the current interglacial, especially cyclical changes in the volume of ice on land (reductions in the volume of ice in the Antarctic and Greenland ice sheets during interglacials) and changes in sea level in response to waxing and waning of the volume of glacier ice on Earth (Donn and others, 1962).

Earth's orbital variation, variations in atmospheric greenhouse gases, changes in North Atlantic oceanic circulation, and decrease in atmospheric dust appear to produce long glacials (about 100,000 years) and short interglacials (about 10,000 years). Figure 33 schematically shows the changes that Broecker and Denton (1990) discussed as the last glacial came to a close and the latest interglacial began.

Analyses of ice cores from Greenland and Antarctica and of marine sediment cores show that gradual changes in paleoclimates and paleoenvironments are atypical for both glacials or interglacials. Instead, glacials and interglacials are characterized by abrupt changes (Kerr, 1993; National Research Council, 2002; Mayewski and White, 2002; Mayewski and others, 2004; Climate Change Science Program, 2008).

Ice cores withdrawn from Dome C by the European Project for Ice Coring in Antaractica (EPICA) span the last 800,000 years and thus provide a continuous climate record through Marine Isotope Stage (MIS) 20.2 (Jouzel and others, 2007). The Antarctic Geological Drilling Program (ANDRILL) has recovered a core of sedimentary and volcanic rocks from the McMurdo Ice Shelf that records 13 Myr of paleoclimate history in Antarctica (Naish and others, 2007). Analysis of the sedimentary sequence in the core shows 60 cycles of advance and retreat (glacial and interglacial) of the margin of the Antarctic ice sheet. This extraordinary record extends from the Miocene

Table 9.—*Analytical techniques used to determine past climates and other past environmental changes in the Earth System, from various data sites and from greatest to least distant range of time*

[Ma, million years ago; ^{14}C, carbon 14. Modified from Stokstad (2001)]

Analytical technique	Geoscience discipline	Range of time for which the analytical technique allows dating	Types of sites studied	Climate or other environmental information obtained
Geomorphological	Geomorphology, Quaternary Geology, Geochronology	Neoproterozoic Era—Cryogenian Period, about 850 Ma to end of Little Ice Age	Continental and insular land masses; sites (few) on submerged continental or insular platforms; landforms	Glacial moraines and other glacial deposits. Glacial erosion features, such as wave-cut marine terraces and prior shorelines
Marine Sedimentary	Sedimentology, Geochemistry, Paleoclimatology	Lower Jurassic Period of Mesozoic Era (about 180 Ma) to present Plate Tectonics limits dating the oldest sea floor to about 200 Ma	Global ocean bottom sediments	Fossils (^{14}C-dating): 40,000 years maximum. Isotopes: temperature, salinity, ice volume, carbon dioxide in atmosphere
Glacier Ice Cores	Glaciology, Geochemistry, Paleoclimatology	From Pleistocene Epoch of Quaternary Period (800,000 years) to present	Ice sheets, ice caps, high-mountain glaciers on all continents except Australia and oceanic islands	Seasonal to decadal volume of ice sheets; carbon dioxide and ammonia in atmosphere; atmospheric dust; proxy surface temperature
Pollen and Spores	Palynology, Paleobotany	Miocene Epoch of Teritary Period (about 6 Ma) to present	Local to regional; terrestrial deposits (land and lake)	^{14}C-dating: 40,000 years maximum. Tephra layers
Tree Rings (Dendro-chronological)	Dendrochronology Botany	Generally limited to Holocene Epoch of Quaternary Period (about 11,000 years)	Trees that grow in temperate climates	Annual resolution of rings: variations in temperature and precipitation
Coral Reefs	Marine Biology, Marine Geology	Continuous record for about 400 years to present (Late Pleistocene Epoch and Holocene Epoch of Quaternary Period). Fossil corals provide short records back to last interglacial (about 130,000 years ago)	Tropical oceans; eastern margins of continents; islands	Sea surface temperature; ^{14}C-dating; higher and lower sea levels in the past

to the 21st century. The upper part of the core shows that the most recent 800,000 years had a cold polar paleoclimate (Naish and others, 2007). During the middle Pliocene, about 3.5 Ma, the Antarctic ice sheet may have been markedly reduced, with global temperatures 3° to 4°C warmer than at present because of an increase in atmospheric CO_2 (Kennett and Hodell, 1995; Schiermeier, 2008).

Iceland also preserves a record of changes in Ice Age paleoclimate, including stratigraphic evidence for multiple cycles of glacial and interglacial intervals. Fourteen ice caps (not including two contiguous ice caps) and mountain glaciers (Sigurðsson and Williams, 2008) at present cover about 10.7 percent of Iceland's total area (11,048 km^2 of 103,000 km^2). Einarsson and Albertsson (1988) studied the interbedding of volcanic and marine and glacial sediments and soils with plant species and marine fauna of Iceland and concluded that about 3 Ma, Earth's climate suddenly became cooler. Dwarf willows (*Salix phylicifolla*), birch (*Betula pubescens*), and two other tree species, similar to modern flora but uncommon to Iceland, almost totally replaced deciduous trees and conifers. Icelandic ice caps formed and ice sheets waxed and waned at least 12 times during glacials (Einarsson and Albertsson, 1988). More recent research found 20 glaciations recorded in Iceland during the Ice Age. The Tjörnes Peninsula contains evidence of 14 glacials (Wohlfarth and others, 2008). After about 15,000 years, the Icelandic ice sheet began to shrink, but it expanded during the Younger Dryas cool interval. By 9,000 years ago, the ice sheet had melted. Only ice caps and mountain glaciers remained, and Iceland's modern landscape developed (Wohlfarth and others, 2008).

During the last interglacial, about 130,000 years ago, temperatures in Greenland were about 5°C warmer than at this stage in the present interglacial. This warmth reduced the Greenland ice sheet by 50 percent, and it separated it into two smaller ice sheets (see fig. 19). This loss of ice volume raised sea level about 3.4 m (Wohlfarth and others, 2008).

In North America, the Laurentide ice sheet also melted rapidly. Only two ice sheets remained in northeastern Canada about 8,000 years ago (deVernal and others, 1997). Some glaciologists consider this rapid Laurentide melting to be a bellwether of future changes in the Greenland ice sheet (Carlson and others, 2008). Data from the base of Greenland ice cores indicate that extensive melting of the Greenland ice sheet occurred at the end of the last interglacial (Koerner, 1989). Otto-Bliesner and others (2006) used models to assess effects of a warmer paleoclimate in the Arctic during the last interglacial on glaciers in the region. Their models show a considerable melting of polar glaciers, contributing to a projected 2.2- to 3.4-m rise in sea level. Schrag and Alley (2004) present evidence that Earth's atmospheric CO_2 concentration is likely to double or triple by the end of the 21st century. Even a doubling could elevate global temperatures by 4°C, which might cause a sudden reduction in the volume of ice in Greenland and Antarctica.

Ice cores from the NGRIP drilling site (Greenland ice sheet) and cores from North Atlantic sediment provide a record of paleotemperature variations on the Greenland ice sheet during the last glacial and fluctuations of the Laurentide ice sheet as it melted. Deglaciation of the Laurentide ice sheet (60–28 Ka) was characterized by six distinctive ice-stream surges (called "Heinrich events," H1–H6) (fig. 34) from the continental margin into the North Atlantic. The ice streams discharged debris-laden icebergs, which melted and deposited the debris onto the sea floor (Ahn and Brook, 2008).

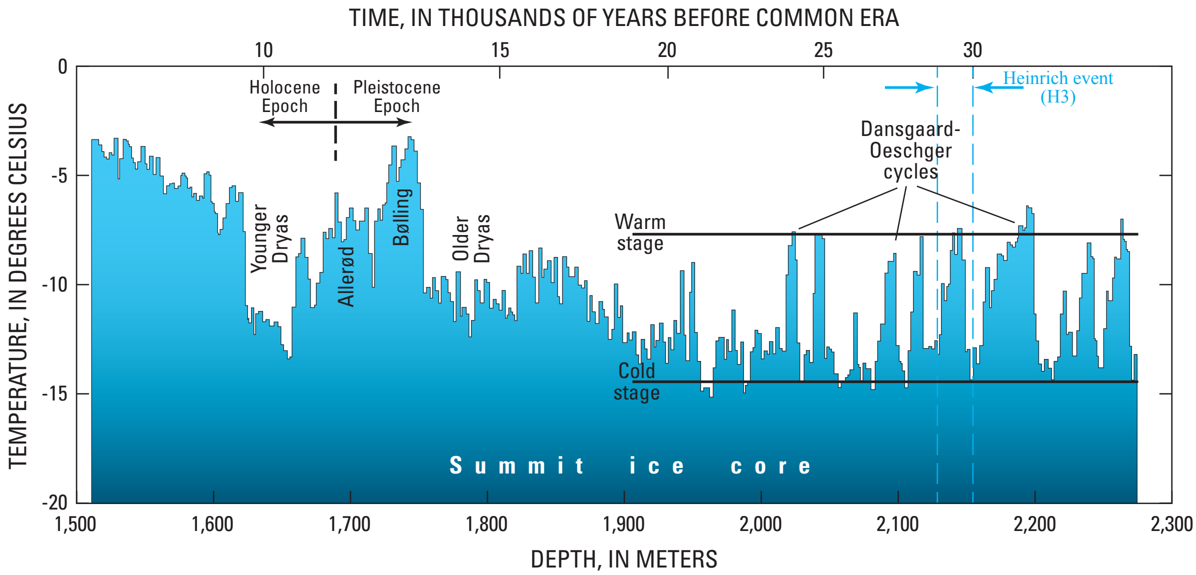


Figure 34.*—Temperature variations in the Summit ice core (Greenland ice sheet) during the last 35 kyr of the late Pleistocene and Holocene Epochs. The late Pleistocene was characterized by cold temperatures, interrupted by about 7 short Dansgaard-Oescher warm intervals. One Heinrich event (H3) is shown when an ice stream on the eastern margin of the Laurentide ice sheet surged. At the end of the Pleistocene, longer warm intervals occurred, including Bølling and Allerød. A final cold period, the Younger Dryas, occurred at the end of the Pleistocene near the beginning of the Holocene. The Holocene is a warm interval with less extreme variations in temperature after the Younger Dryas (see fig. 35). Modified from Ahn and Brook (2008, figure on p. 84). The original source is the Greenland Ice-Core Project (see also Kerr (1993, figure on p. 891)).*

The NGRIP ice core recorded 25 warming events during the last glacial ("Dansgaard-Oeschger events") (fig. 34) (Andersen and others, 2004; Jouzel and others, 2007).

In transition between the last glacial and the current interglacial, a warm period was followed by a colder period (Older Dryas), that is, two warm periods (Bølling and Allerød) were separated by a brief colder interval (fig. 34). The beginning of the Holocene was characterized by an extended cold interval (Younger Dryas), as cold as preceding glacials, followed by warming that extends to the present but with fewer variations in temperature (figs. 34 and 35).

Glaciologists and paleoclimatologists have yet to explain why the thousand-year-long Younger Dryas occurred. Although highly speculative, one hypothesis suggests that an asteroid impact may have altered the temperature (Kerr, 2009). Another short-lived, abrupt cooling event occurred about 8,200 years ago (figs. 28, 29). Clarke and others (2003) ascribe this cooler interval to disruption of circulation in the North Atlantic Ocean caused by a jökulhlaup from Lake Agassiz into Hudson Bay and the Labrador Basin as the Laurentide ice sheet melted (see discussion of Pleistocene jökulhlaups, p. A268–A269). Aeolian dust in the troposphere, volcanic tephra and aerosols (for example, sulphuric acid, H_2SO_4) in the stratosphere, and dust and aerosols in the atmosphere from impacting comets or asteroids decrease the amount of solar insolation reaching the Earth's surface, thus cooling the planet. Major

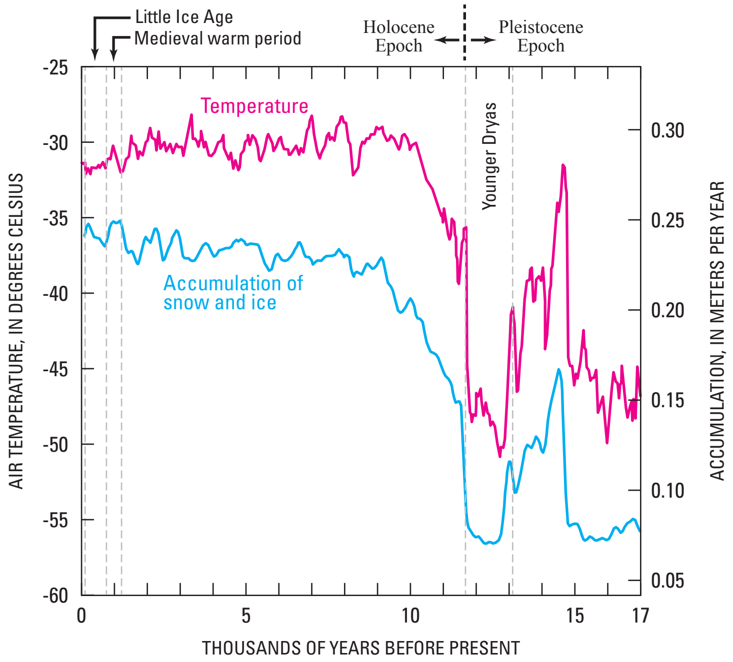


***Figure 35.**—Temperature variations during the late Pleistocene Epoch and near the beginning of the Holocene Epoch, determined as proxy temperatures from ice cores extracted from the central part of the Greenland ice sheet. Note the 1,000-year Younger Dryas cold interval (about 10 ka to 11 ka), the short Medieval Warm Interval (about 1100–1300 C.E.), and the Little Ice Age (about 1400–1880 C.E.). Modified from Alley (2000, p. 9, fig. 12).*

explosive volcanism has cooled the Earth's atmosphere and caused changes in tree growth (Briffa and others, 1998). Growth changes can last for one year following a volcanic event, and sometimes for several years, as in the case of the eruption of Komaga-Take in July 1640 and of Tambora in April 1815. The eruption of Tambora, in particular, caused the longest period of cooler climate recorded during historic time (D'Arrigo and others, 2009). The eruption of Mount Pinatubo in June 1991 cooled Earth for one or two years (Robock and Oppenheimer, 2003). The effusive volcanic eruption from the Laki fissure in south-central Iceland (June 1783 to February 1784) cooled Europe, Asia, and North America (The Economist, 2007b).

Extraterrestrial dust below or above the ecliptic plane, in which most of the planets in the Solar System move in their orbits around the Sun, has been suggested as an alternative hypothesis to explain the 100,000-year cycle of glacials and interglacials during the Ice Age (Monastersky, 1997). Muller and MacDonald (1997) hypothesized that the Earth moves below or above the ecliptic in a 100,000-year cycle, encountering interplanetary dust that reduces insolation of Earth's surface. However, most glaciologists support the Milankovich Theory that the three variables in Earth's orbit account for the current 100,000 year glacial-interglacial cycles (Imbrie and Imbrie, 1979).

Holocene Epoch

During the Holocene Epoch, the Younger Dryas cold interval occured (Fairbanks, 1990); the Laurentide ice sheet in North America melted (about 5,000 years ago); the Scandinavian ice sheet melted (about 7,000 years ago); and the Patagonian ice sheet was reduced to two separate ice fields along the Andes between Chile and Argentina. The margins of the ice sheets in Antarctica and Greenland fluctuated through advances and retreats, as did margins and outlet glaciers of ice caps and ice fields and the termini of mountain glaciers (Denton and Porter, 1970; Sugden and John, 1976). Figure 36 schematically represents global fluctuations of glaciers during most of the Holocene, when climate was quite variable (Mayewski and others, 2004). Schaefer and others (2009) emphasized that the timing of advance and retreat of mountain glaciers during the Holocene in the Northern and Southern Hemispheres remains an unsolved problem and reinforces the need to better understand "the complexity of climate dynamics" (Balco, 2009).

Little Ice Age

As the Industrial Revolution was beginning in the late 1700s, human activities had not yet begun to significantly alter the Earth's atmosphere. As noted by Ruddiman (2005), however, humans had already begun the process of altering the Earth's climate 8,000 years ago. The expansion of agriculture into previously forested areas changed the natural land cover, for example. The CO_2 concentration in the atmosphere was about 280 ppm, nearly the maximum to be expected during an interglacial, as EPICA Community Members (2004) determined from analyses of ice cores. The advance of glaciers worldwide during the Industrial Revolution cannot, therefore, be attributed to the slow accumulation of greenhouse gases (such as CO_2 and NH_4^+, and so on) in contrast to the 20th and 21st centuries but must rather be the result of other cooling processes. This is another indication that climate change is the result of multiple forcings, some predictable, some unpredictable, and some not even expected (Overpeck, 1996; National Research Council, 2002; Climate Change Science Program, 2008).

After a Medieval Warm Interval (about 1100 to 1300 C.E.) (fig. 35), global cooling (1400s to the late 1800s or nearly 500 years) caused a readvance of glaciers worldwide. François Matthes (1939), a geologist and topographer with the USGS coined the term "Little Ice Age" for this advance. Grove (1988) comprehensively documented the effect of lower Little Ice Age temperatures on various geographic regions, including the advance of glaciers in Iceland, Eurasia, Greenland, North America, and South America. Grove (1988) also discussed possible causes for this extended interval of cooling, including changes in the Earth's orbital parameters, reduction in solar insolation, reduction in concentration of atmospheric CO_2, and increase in explosive volcanism.

Some scientists have hypothesized that variations in the output of energy from the Sun can also change climate. The number of sunspots during the Sun's 11-year magnetic cycle can vary from a few at solar minimum (Livingston and Penn, 2009) to more than 100 at solar maximum. During the period 1645 to 1715, there was a prolonged absence of sunspots (the so-called "Maunder Minimum"); this minimum spans the coldest years of the Little Ice Age (Hammond, 1976). Recent research refutes reduced insolation as a cause of the Little Ice Age (as well as the parallel hypothesis that an increase in solar energy explains climate warming during the late-20th century). The variability in solar energy is inadequate to produce the requisite insolation changes (Duffy and others, 2009).

It is obvious from the history of climate variability during the latter part of the Holocene that warmer intervals, such as the Medieval Warm Period, and colder intervals, such as the Little Ice Age, are caused by natural fluctuations in the Earth's climate system (Jones and others, 2001). At the end of the Little Ice Age, in the late 1800s, the termini of mountain glaciers and outlet glaciers from ice caps and ice fields generally deposited terminal moraines farthest from

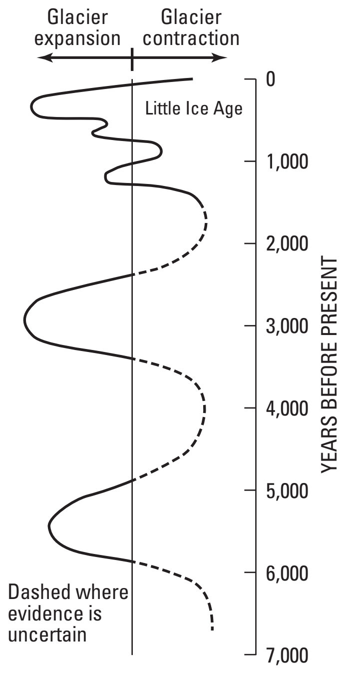


Figure 36.—*Schematic diagram of glacier fluctuations worldwide during the last 7,000 years of the Holocene Epoch. Modified from Sugden and John (1976, p. 124, figure 6.17).*

subsequent, recessional terminal moraines and the present (2012) positions of termini. The only exceptions are the submarine terminal moraines that tidewater glaciers deposit and the morainic debris deposited by surge-type glaciers.

From the late 19th century throughout the 20th century and into the beginning of the 21st century, most such glacier termini have retreated from their positions at the end of the Little Ice Age, even though there have been a few intervals of cooler climate in which non-ice-sheet glaciers either stopped receding or readvanced. As concentrations of greenhouse gases in the Earth's atmosphere have continued to increase, specific changes have become more evident. Good examples include increase in global temperatures during the 1990s and the early 21st century, the warmer summertime temperatures in temperate regions, and the rapid recession of the termini of non-ice-sheet glaciers globally. Changes in the two remaining ice sheets are also becoming more evident. Ice shelves on the Antarctic Peninsula (part of the Antarctic ice sheet) are retreating (Ferrigno and others, 2006, 2008, 2009a), glacier ice on land is thinning, and summertime melting (area affected and duration of melt) on the Greenland ice sheet is becoming more extensive.

From the perspective of Ice Age glacial and interglacial cycles, the Earth is approaching the end of the normal 10,000-year duration of an interglacial. Global climate is expected to continue its natural variability of warmer or cooler than normal intervals, but superimposed on this is an anthropogenic component of warming of the Earth's atmosphere, land areas, and oceans. This component is the result of increased concentration of atmospheric greenhouse gases, which arises from the collective activities of 6.7 billion human beings (as of 2009). This ongoing global warming is evident in the rapid changes occurring in all components of the Earth's cryosphere: glaciers, snow cover, sea ice and lake and river ice, and permafrost and periglacial environments. It is especially evident to geologists and glaciologists who have long studied and directly witnessed the shrinkage in volume of the glaciers, the retreating termini of glaciers (mountain and outlet), and the retreating margins of ice caps and ice sheets.

Imbrie and Imbrie (1979) suggested that the increasing volume of greenhouse gases may create a "super-interglacial" (fig. 27) (see also figure 1 in Haeberli and Wallén (1992, p. 6)). A "super-interglacial" inplies a warmer climate than existed at the end of the last interglacial (Eemian; MIS5), when a substantial part of the Greenland ice sheet melted, producing a rise of several meters in global sea level. The rate of melt of the ice sheets has enormous implications for the rate at which sea level rises during the 21st century and beyond, especially the economic, social, and psychological costs associated with the displacement of human beings when the infrastructure in low-lying islands and coastal regions is inundated (see Hansen, 2004, 2007a; Alley, 2007; and Pennisi and others, 2007).

There is an increasing consensus among Earth scientists that a warmer global climate will cause significant changes in the Earth's cryosphere, including loss of glacier ice, reduction in seasonal extent of sea ice, especially in the Arctic (Carroll, 2008; Boé and others, 2009), reduction in permafrost, and less seasonal snow cover. A number of scientific summaries about expected changes in the cryosphere were presented to the United Nations Climate Change Conference that was convened in Copenhagen, Denmark, 7–18 December 2009. Summaries written by Allison and others (2009), Raloff (2009), and United Nations Environment Programme (2009) are particulary valuable and echo the changes noted above.

Ice Cores and Climate

Introduction

Ice cores obtained primarily from ice sheets and ice caps provide distinct records of past concentrations of CO_2 and CH_4 in the atmosphere. Evidence of past volcanic activity from layers of tephra (airborne particulates deposited from explosive volcanic eruptions), layers of aeolian dust, records of atmospheric temperature that provide information on the variability of past climate, and other chemical and isotopic "markers" of major explosive volcanic events are indicators of natural events that may have caused cooling of the Earth's climate (Oeschger and Langway, 1989). Ice cores from the Antarctic and Greenland ice sheets provide the longest continuous records, about 800,000 years and 130,000 years, respectively. It is likely that the planned drilling and retrieval of an ice core from Dome A in Antarctica will extend the record back to more than 1 Ma (White, 2004; Jones, 2007). Langway (2008), in his comprehensive documentation of the history of the earliest ice cores obtained from the Greenland and Antarctic ice sheets, essentially covers the period from 1956 to 1981 (Langway and Weertman, 2005). During that time span, ice cores were recovered from five sites in the Greenland ice sheet, including three Greenland Ice Sheet Program (GISP) sites, and two sites in the Antarctic ice sheet (Langway, 2008, p. 4, fig. 4). Langway (2008, p. 47) lists ice cores obtained from four sites in Greenland ice sheet: Greenland Ice Sheet Project Program (1971–1981) (Dansgaard and others, 1982); Greenland Ice Sheet Program 2 (GISP-2); Greenland Ice Core Project (GRIP) (Pinklin and others, 1993); and North Greenland Ice Core Project (NGRIP). During 1989 to 2007, ice cores were also obtained from four sites in the Antarctic ice sheet: Vostock, Dome Fuji, Dome C (Brook, 2005), and Kohnen Station. The first meeting of the International Partnership in Ice Coring Science (IPICS) to present and discuss the findings was held in Washington, D.C., in 2004. The second meeting was held in Brussels, Belgium (Brook and Wolff, 2006), to continue the research.

Another important aspect of the study of gases, chemicals, isotopes, and particulates in ice cores—and the dating of layers—is the correlation of events recorded in ice cores with cores from marine and terrestrial environments. Large-magnitude explosive volcanic events eject tephra (particulates) and aerosols into the stratosphere; subsequently, particulates reenter the troposphere and are eventually deposited on land, in the oceans, and on glaciers, where they form a precise datable (time-restricted) layer. Ejecta from smaller explosive volcanic events are confined to the troposphere. The tephra is subsequently deposited downwind from the source in a layer of decreasing distal thickness. In Iceland, for example, if Hekla erupts explosively when wind direction is from the south, southeast, or east, the resultant tephra layer is deposited in Iceland (on land and ice caps), in the Denmark Strait (marine sediments), and in Greenland (ice sheet). A datable record of a specific volcanic eruption in Iceland is thereby preserved concurrently over a wide geographic area. Tephrochronologists can date successive tephra layers in such deposits, thus establishing a series of correlative layers in three different environments.

Another example of correlating terrestrial deposits with ice cores involves a 400,000± year record of paleotemperature (deuterium, a naturally occurring stable isotope of hydrogen) in the Vostok (Antarctica) ice core and a radiometrically dated paleotemperature ($\delta^{18}O$, a naturally occurring stable isotope of oxygen) record in a calcite core taken from Devils Hole, Nevada (Landwehr and Winograd, 2001). Making the standard assumption that major climatic events are synchronous in the two paleotemperature records, Landwehr and Winograd (2001) were able to extrapolate the time scale from Devils Hole to Vostok. The Devils Hole chronology also permitted accurate dating of other chemical properties in the Vostok ice core (CO_2, CH_4, nitrous oxide (Spahni and others, 2005), and aerosols).

Ice cores from the Antarctic and the Greenland ice sheets also contribute to the glacial-interglacial record of paleoclimates and paleoenvironments (Charles and others, 1994). The Antarctic and Greenland ice cores are especially important because they provide the longest records of Pleistocene paleoclimates. Ice cores from temperate and tropical glaciers (primarily ice caps and icefields) provide a shorter record, a maximum of about 20,000 years (Bowen, 2005). But geologists can still decipher seasonal atmospheric paleocirculation and other paleoclimatic processes and related paleoenvironmental information from these lower-latitude glaciers (Cecil and others, 2004, 2010; Naish, 2007).

Ice Cores, High-Mountain Glaciers, and Climate

By LONNIE G. THOMPSON[5]

Introduction

The near-global retreat of high-mountain glaciers and the shrinkage of ice fields and ice caps and retreat of their associated outlet glaciers are perhaps the most visible evidence for 20th-century climate change and the recent increase in the globally averaged near-surface temperatures of the Earth's surface. The loss of high-mountain glaciers—including ice fields, ice caps, and other glaciers—not only diminishes regional water supplies but also destroys the long, scientifically valuable, and often detailed climatic histories that are unavailable from any other terrestrial or marine record. The question arises as to the significance of 20th-century retreat of glaciers within a much longer perspective. In a few cases, the paleoclimate histories contained within ice cores from these shrinking ice fields and ice caps can provide this critical temporal context. The ongoing global-scale rapid retreat of mountain glaciers threatens freshwater supplies in many of the world's most populous regions while at the same time contributing to the rise in global sea level.

Innovations in lightweight drilling technology have enabled paleoclimatic research on ice cores to expand from the polar regions (for example, ice cores from the Greenland and Antarctic ice sheets) to ice fields and ice caps in many of the highest mountains on Earth. During the last few decades, much effort has been focused on retrieving ice cores from glaciers in subpolar regions, such as western Canada and eastern Alaska; from the mid-latitudes, such as the Rocky Mountains and the Alps; and from tropical mountains in Africa, South America, and China. Unlike climate records from polar ice cores, those from lower latitude high-mountain glaciers—including ice caps, ice fields, and other types of glaciers—present information necessary for studying climatic processes in areas where human activities are concentrated, especially in the tropics and subtropics, where 70 percent of the world's population lives. During the past 100 years, changes in global and regional-scale climates and environmental changes to which human beings are vulnerable have accelerated. These changes will increasingly affect us in the current century and in future centuries. The following overview of the archives of these past changes as held in high-mountain glaciers on millennial to decadal time scales includes a review of the recent, global-scale retreat of these high-mountain glaciers under present climate conditions. The significance of this retreat for the longer term perspective, which can be inferred only from paleoclimate records contained in ice cores, is also discussed.

Geographic Locations of Mountains Where Glacier Ice Cores Have Been Obtained

The sites where many of the high-altitude glacier ice cores have been retrieved are shown in figure 37 and listed in table 10. Among the earliest efforts to retrieve climate records from high-mountain glaciers were research

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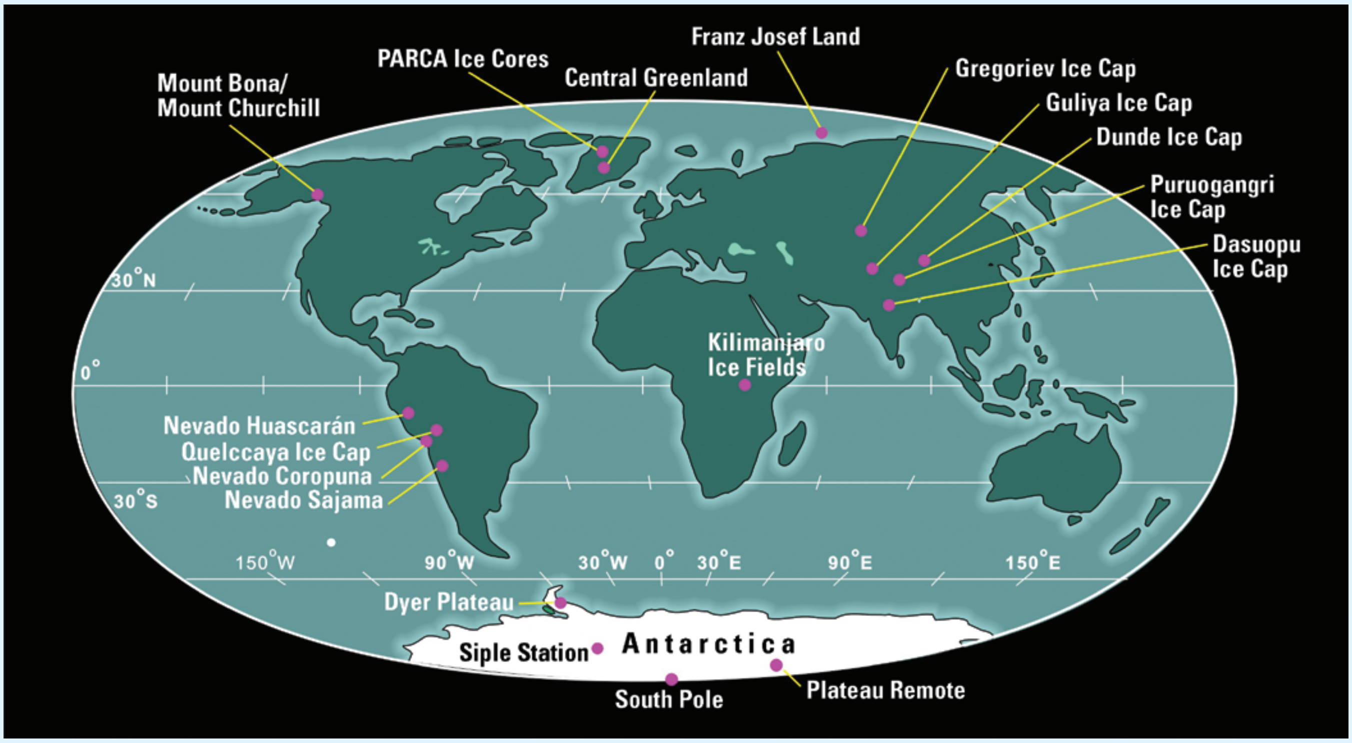


Figure 37.—*Geographic locations of sites where ice cores have been obtained by the ice-core paleoclimate research group at the Byrd Polar Research Center, The Ohio State University. PARCA, Program for Arctic Regional Climate Assessment.*

programs from 1974 to 1979, which involved surface sampling on the *Quelccaya Ice Cap* (5,670 m above sea level [asl]; lat 13°56' S., long 70°50' W.) in southern Perú. (See section on *Quelccaya Ice Cap* by Stefan Hastenrath in Chapter 1386–I, "Glaciers of South America," p. I64–I65, of this series (Morales-Arnao, 1998)). The results from this preliminary research by the Institute of Polar Studies (now the Byrd Polar Research Center, or BPRC) at The Ohio State University paved the way for the first tropical high-altitude, deep-drilling program on the *Quelccaya Ice Cap* in 1983, which yielded a 1,500-year climate record. Meanwhile, in western Canada in 1980, a 103-m ice core was drilled on Mount Logan (5,340 m asl; lat 60°35' N., long 140°35' W.) by Canada's National Hydrology Research Laboratory. The record from this core extends back to A.D. 1736. The ice caps on both mountains have been redrilled recently, Mount Logan in 2002 and *Quelccaya Ice Cap* in 2003. During the intervening decades high-mountain ice-core research has expanded significantly throughout the world, with programs successfully completed on the Tibetan Plateau, the Himalaya, the Cordillera Blanca of northern Perú, Bolivia, East Africa, the Swiss and Italian Alps, Alaska, and the Pacific Northwest of the United States. In 2002, the BPRC paleoclimate ice-core group succeeded in retrieving the deepest ice cores ever recorded (460 m) from the col between Mount Bona (5,008 m asl) and Mount Churchill (4,770 m asl) in the St. Elias Mountains of southeastern Alaska. (See descriptions of glaciers in this mountain range in Chapter 1386–K, "Glaciers of Alaska," of this series (Molnia, 2008).

Table 10.—*A selection of sites of high-mountain glaciers from which research organizations have obtained ice cores since 1980*

[NA, time series for this ice core is not yet available; place-name in italics indicates that it is informal usage (unapproved)]

Abbreviations for the organizations:

BPRC, Byrd Polar Research Center (previously called the Institute of Polar Studies), The Ohio State University (USA)
NHRC, National Hydrology Research Centre (Canada)
NGP, National Glaciology Programme (Canada)
AINA, Arctic Institute of North America, University of Calgary (Canada)
NIPR, National Institute of Polar Research (Japan)
IQCS, Institute for Quaternary and Climate Studies (now called the CCI, Climate Change Institute), University of Maine (USA)
USGS, U. S. Geological Survey (USA)
PSI, Paul Scherrer Institut (Switzerland)
IRD, Institut de Recherche pour le Développement (France)
LGGE, Laboratoire de Glaciologie et Géophysique de l'Environnement (France)
IGRAS, Institute of Geography, Russian Academy of Science (Russia)
IFU, Institut für Umweltphysik (Germany)
UNIBE, Universität Bern (Switzerland)
LIICRE, Key Laboratory of Ice Core and Cold Regions Environment (China)
UNH, Institute for the Study of Earth, Oceans, and Space, University of New Hampshire (USA)

Name of glacierized mountain or glacier, and state or country	Location (latitude and longitude)	Elevation (meters above sea level)	Year drilled	Leading organization	Length of core (meters)	Length of record (years)
Mount Bona/Mount Churchill, Alaska	61°24´N. 141°42´W.	4,420	2002	BPRC	460	NA
Mount Logan, Canada	60°35´N. 140°35´W.	5,340	1980 2002	NHRC, NGP, AINA, NIPR, IQCS, UNH	103 190 220 345	225 NA NA NA
Eclipse Dome, Canada	60°51´N. 139°47´W.	3,017	1996	UNH	160	NA
Fremont Glaciers, Wyoming	43°09N. 109°37´W.	4,100	1991 1998	USGS	160 50, 160	275 NA
Gora Belukha, Kazakhstan/Russia	49°48´N. 86°34´E.	4,062	2001	PSI, UNIBE	140	200+
Fiescherhorner, Switzerland	46°32´N. 8°02´E.	3,880	1988 2002	PSI, UNIBE	30 150	42
Col du Dôme, France	45°50´N. 6°50´E.	4,250	1994	LGGE, IFU	139	75
Mont Blanc, France	45°45´N. 6°50´E.	4,807	1994	LGGE	140	200+
Djantugan, Russia	43°12´N. 42°46´E.	3,600	1983	IGRAS	93	57
Gregoriev, Kyrgyzstan	41°58´N. 77°55´E.	4,660	1991 2001 2003	IGRAS, BPRC IGRAS IGRAS	20, 16 21.5 22	53 NA NA
Dunde ice cap, China	38°06´N. 96°24´E.	5,325	1987	BPRC, LIICRE	183	10,000+
Malan ice cap, China	35°50´N. 90°40´E.	6,056	1999	LIICRE	102	112+
Guliya ice cap, China	35°17´N. 81°29´E.	6,200	1992	BPRC, LIICRE	302	110,000+
Purugangri ice cap, China	33°55´N. 89°05´W.	6,000	2000	BPRC, LIICRE	208	NA
Dasuopu glacier, China	28°23´N. 85°43´E.	7,200	1996	BPRC, LIICRE	162	1,000+
Qomolangma, Nepal/China	27°59´N. 86°55´E.	6,500	1998	UNH, LIICRE	80	154
Kilimanjaro ice fields, Tanzania	3°04´S. 37°21´E.	5,895	2000	BPRC	52	11,700
Nevado Huascarán, Perú	9°06´S. 77°36´W.	6,050	1993	BPRC	166	19,000
Quelccaya ice cap, Perú	13°56´S. 70°50´W.	5,670	1983 2003	BPRC	155, 164 168, 129	1,500 1,800
Nevado Illimani, Bolivia	16°37´S. 67°47´W.	6,350	1999	IRD, PSI	137, 139	18,000
Nevado Sajama, Bolivia	18°06´S. 68°53´W.	6,540	1996	BPRC	132	20,000

Climatic and Environmental Information from High-Mountain Ice Cores

The records within the Earth's high-mountain ice caps and other glaciers contain a wealth of data that contribute to a spectrum of critical scientific and societal issues. These data have enabled glaciologists to reconstruct high-resolution climate histories; explore the oscillatory nature of the climate system; estimate the timing, duration, and severity of abrupt climate events; and estimate the relative magnitude of 20th-century global climate change and its impact on the Earth's cryosphere, among other achievements. Many of the measurements made on polar ice cores (for example, from the Greenland and Antarctic ice sheets) and the resulting information are also relevant to ice cores from high-mountain glaciers. Researchers can utilize an ever-expanding ice-core database of multiple kinds of proxy information (for example, stable isotopes, insoluble dust and tephra, major- and minor-ion chemistry, precipitation reconstructions) that spans the globe in terms of geographic coverage and that is of the highest possible temporal resolution. Numerous parameters can be measured that yield information on regional histories of variations in temperature, precipitation, moisture source, aridity, vegetation changes, volcanic activity, and anthropogenic input (table 11). Many of these physical and chemical constituents yield signals in the ice of wet and dry or cold and warm seasons that allow the variations to be analyzed backward in time, much as does tree-ring analysis (dendroclimatology).

Isotopic ratios of oxygen ($\delta^{18}O$) and of deuterium (δD) in water are among the most widespread and important of the measurements made on ice cores. The information that $\delta^{18}O$ and δD provide is based on the fractionation of the oxygen and hydrogen atoms into their light and heavy isotopes (^{16}O and ^{18}O, and ^{1}H and ^{2}H or deuterium (D)) and on the higher vapor pressure of $H_2{}^{16}O$ over $HD^{16}O$ and over $H_2{}^{18}O$. Early work on these stable isotopes (as opposed

Table 11.—*Principal sources of paleoclimatic information from ice cores*

[A variety of environmental information can be obtained from high-altitude ice cores. In the low latitudes, pollen and entrapped microorganisms have been measured; insects and organic material are sources for accelerator mass spectrometer (AMS) carbon-14 dating (after Bradley, 1999). Abbreviations: δD, delta deuterium; $\delta^{18}O$, delta oxygen-18; ^{10}Be, beryllium-10; SO_4, sulfate; CO_2, carbon dioxide; CH_4, methane; N_2O, nitrous oxide; ECM, electrical conductivity measurement]

Parameter	Analysis
Paleotemperatures:	
Summer	Melt layers
Annual? Days with snowfall?	δD, $\delta^{18}O$
Humidity	Deuterium excess (d)
Paleoaccumulation (net)	Seasonal signals, ^{10}Be
Volcanic activity	Conductivity, nonsea salt, SO_4
Tropospheric turbidity	ECM, microparticle content, trace elements
Wind speed	Particle size, concentration
Atmospheric composition: long-term and anthropogenic changes	CO_2, CH_4, N_2O content, glaciochemistry
Atmospheric circulation	Glaciochemistry (major ions)
Solar activity	^{10}Be

to unstable, or radioactive, isotopes) in polar ice cores resulted in transfer functions between the isotopic ratios and the water temperature at the moisture source and the atmospheric temperature as the water vapor condenses into precipitation (Dansgaard, 1961).

The use of $\delta^{18}O$ and δD as temperature proxies for polar ice is now widely accepted; however, it is still a source of controversy for lower latitude ice cores. Some geoscientists who have studied the problem believe that $\delta^{18}O$ is a function of the amount of precipitation rather than being a temperature proxy at lower latitudes (Rozanski and others, 1993; Kang and others, 2000; Qin Dahe and others, 2000; Baker and others, 2001; Tian and others, 2001). However, real-time comparisons of $\delta^{18}O$ measured on precipitation on the northern Tibetan Plateau and air temperature reveal a very close relation between the two (Yao and others, 1996). Correlations between ice-core records from the Himalaya and temperature records for the Northern Hemisphere show that the dominant factor controlling mean $\delta^{18}O$ values in snowfall on longer than annual time scales must be temperature, not precipitation (Thompson and others, 2000; Davis and Thompson, 2004; Thompson and Davis, 2005; Thompson and others, 2006). On seasonal to annual time scales, temperature and precipitation both influence the local $\delta^{18}O$ signal (Vuille and others, 2002, 2003, 2005).

The annual net balance[6] on an ice cap can be constructed by measuring the thickness of ice between seasonal variations in one or more parameters that indicate warm or cold seasons (such as $\delta^{18}O$) or that characterize wet or dry seasons (such as aerosol concentrations). Because ice is viscous, it tends to flow not only vertically but also horizontally, resulting in thinning of the annual layer with depth. To correct for this deformation so that the original thickness of an annual layer at the time of its deposition in the past can be reconstructed, vertical-strain models are used that take into account the changing densities with depth, the thickness of the glacier, and the rate of thinning (Bolzan, 1985; Reeh, 1988; Meese and others, 1994).

Aerosols in the atmosphere are either deposited on high-mountain ice fields, ice caps, and other glaciers as nuclei of snow (wet deposition) or carried by turbulent air currents to high altitudes (dry deposition). Either way, these insoluble-mineral dust particles and soluble salts (chlorides, for example) record variations in environmental conditions such as regional aridity. The concentration and size distribution of insoluble dust are also helpful for qualitative reconstructions of wind speeds. Evidence of volcanic eruptions is found in sulfate concentrations and (or) the presence of microscopic tephra particles in the ice cores. If these volcanic layers are geochemically or petrologically identifiable (for example, as the 1815 eruption of Tambora or the 1883 eruption of Krakatoa, both in the East Indies), they can serve as valuable reference horizons to calibrate the time scale within and between ice cores (tephrochronology). Biological aerosols, such as pollen grains (Liu and others, 1998) and nitrates that may have been injected into the atmosphere by vegetation upwind of a glacier (Thompson and others, 1995; Thompson, 2000), have been useful for reconstructing past climate and environmental changes that caused changes in regional flora.

[6] "The *balance* or *mass balance* b at any time is the algebraic sum of the accumulation and ablation … The mass balance at the end of the balance year is the *net balance* b_n for the year. It can be subdivided into a *winter balance* b_w which is positive, and a *summer balance* b_s which is negative." (Paterson, 1994, p. 28).

The record of human activity also is available from high-mountain ice cores, although this type of research lags behind that on analysis of ice cores from the polar regions. Analysis of heavy metals in high-mountain glaciers is relatively new, but what is available from Mont Blanc (Monte Bianco) in the Savoy Alps, on the border of France and Italy, provides information about increasing industrial production and other activities associated with expanding populations and urbanization (van de Velde and others, 1999). Measurements of carbon dioxide, methane, and other atmospheric gasses trapped in ice bubbles in low-latitude ice cores are not as extensive as they are from polar ice cores; however, the available measurements show that so-called greenhouse-gas concentrations correlate with the temperature proxy $\delta^{18}O$ (Yao, Duan, and others, 2002; Yao, Thompson, and others, 2002). The information from these studies complements other proxy records that define the Earth's climate history, the ultimate yardstick by which the significance of present and projected anthropogenic effects are being and will continue to be assessed.

The Significance of Climate Records from High-Mountain Glaciers

When ice-core records from high-mountain glaciers are combined with high-resolution proxy records such as those from tree rings, lacustrine and marine cores, corals, and other sources, geologists are afforded an unprecedented examination of the Earth's climatic history. Such records can extend over several centuries or even millennia. The longest records have revealed the nature of climate variability since the Last Glacial Maximum (LGM), 18 to 20 ka (thousand years ago), and even longer ago (Thompson and others, 2006). The more recent parts of the ice-core records—which have annual and, in some cases, seasonal resolution—can yield high-resolution temporal variations in the occurrence and intensity of coupled ocean-atmosphere phenomena of global significance, such as El Niño[7] and monsoons; the latter are most strongly expressed in the tropics and subtropics. Such information is particularly valuable because long-term meteorological observations from these regions are scarce; most records are of short duration (decadal at most, as opposed to a century or more in the United States and Europe) and in many cases intermittent.

Three records from the Andes (Nevado Huascarán, northern Perú; and Sajama and Illimani, Bolivia) and one from the western Tibetan Plateau (Guliya ice cap) extend to or past the end of the last glacial stage (of the Pleistocene Epoch), along with other climate proxy records (for example, Guilderson and others, 1994; Stute and others, 1995). These records confirm that the LGM was much colder in the tropics and subtropics than previously believed (Thompson and others, 1995, 1997, 1998; Ramirez and others, 2003). Although this period was consistently colder, it was not consistently drier throughout the lower latitudes than it was in the polar regions. For example, the effective moisture along the axis of the Andes during the end of the last glacial stage was variable, being much drier in the north than in the Altiplano region in the central part of the range. Meanwhile, in western China, the southwest Indian monsoon system was much weaker during the last glacial stage than during the Holocene Epoch. However, the region of the Tibetan Plateau in which the Guliya ice cap is located also received (and receives) mois-

[7] [Editors' note: El Niño is defined as a phenomenon in a region of the equatorial Pacific Ocean characterized by a sea surface temperature that is ≥0.5°C warmer than normal, averaged over three consecutive months; this increase affects the global climate system.]

ture generated from the cyclonic activity carried over Eurasia by the prevailing wintertime westerlies. Not only were lake levels in the western Kunlun Shan higher than in tropical lakes during the LGM (Li Shijie and Shi Yafeng, 1992) but the dust concentrations in the Guliya ice-core record were consistent with those of the early Holocene, when the summer Asian monsoons became stronger, suggesting that higher precipitation and soil-moisture levels inhibited local sources of aerosols during this cold period (Davis, 2002).

Ice cores from the Andes can also contribute to our understanding of past environmental and climatic conditions of the Amazon Basin. The extent of the Amazon rainforest during the LGM is controversial, but the nitrate concentration record from the Nevado Huascarán ice core sheds some light on this debate (Colinvaux and others, 2000). Pollen studies from the Amazon Basin suggest that the areal extent of the rainforest has not changed much between the LGM and the Holocene. Proponents of the "refugia" theory (for example, Clapperton, 1993) assert, however, that the cold, dry climate in the tropics caused a major retreat of the rainforest flora (and associated fauna) into small, geographically isolated areas, leaving most of the Amazon Basin covered by grasslands. In the Nevado Huascarán ice core, the nitrate-concentration profile is similar to (although lagging behind) the $\delta^{18}O$ record throughout most of the ice core, and the very low concentrations of nitrate that are concurrent with very depleted $\delta^{18}O$ suggest that biological activity upwind of the Cordillera Blanca of Perú was impeded by the cold and dry climate ~19 ka. Limited evidence for tropical ecosystems suggests that rainforest and forest soils may be a major source of active species of atmospheric nitrogen such as NO and NH_3, which are precursors for NO_3^- and NH_4^+, which occur as aerosols. The low NO_3^- concentrations in Late Glacial Stage ice may imply that the forest cover was significantly reduced in response to dry conditions (Thompson and others, 1995).

Most of the deep ice cores from the low latitudes extend through at least the Holocene Epoch and show spatial variations in climate, even between records from the same region. For example, the Holocene $\delta^{18}O$ profiles from Nevado Huascarán, Perú, and from Nevado Illimani, Bolivia, which are similar in showing nearly identical early Holocene isotopic enrichment, differ from the profile from Nevado Sajama, Bolivia, which shows relatively low isotopic variability throughout the last 10,000 years. Although Nevado Sajama and Nevado Illimani are geographically close, they are on opposite sides of the Andes, with Nevado Illimani to the east in the Cordillera Real (Jordan, 1998; see especially fig. 17, p. I99). Like Nevado Huascarán far to the north, Nevado Illimani has long received most of its precipitation from the northeast after the water that makes up the precipitation has cycled through the Amazon Basin. Nevado Sajama, which is on the high, dry Altiplano, is more subject to Pacific influences, local hydrological processes, and sublimation.

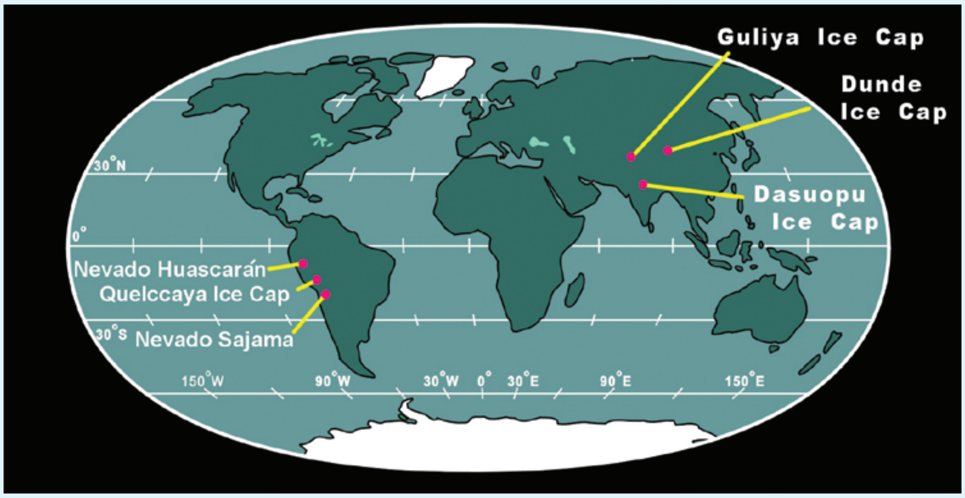
Holocene ice-core records from high-mountain glaciers around the world show evidence of major climatic disruptions (droughts and abrupt cold events) during an epoch that geologists previously believed to have been stable. Beginning between 4.2 and 4.5 ka and lasting several hundred years, major dust events are observed in ice cores from Nevado Huascarán and from Kilimanjaro ice fields (in Africa) beginning between 4.2 ka and 4.5 ka (Thompson, 2000; Thompson and others, 2002, respectively), and lasting several hundred years. The timing and character of the dust spike are similar to those seen in a marine sediment-core record from the Gulf of Oman (Cullen and others, 2000) and in a speleotherm $\delta^{13}C$ record from a cave in Israel (Bar-Matthews and others, 1999). This dry period is also documented in several other proxy climate records throughout Asia and North Africa (Dalfes and others, 1997). Two other periods of abrupt, intense climate change in East Africa are observed in the ice cores from the Kilimanjaro ice fields at ~8.3 ka and 5.2 ka (Thompson and others, 2002). The latter event is associated with a sharp decrease in $\delta^{18}O$, indicative of a dramatic but short-term cooling.

More recently, a historically documented drought in India in the 1790s, which was associated with monsoon failures and a succession of severe El Niños, was recorded in the records of insoluble and soluble aerosol concentrations in the ice core from the Dasuopu glacier, China (Thompson and others, 2000). The ice core from the Dasuopu glacier recorded another Asian monsoon failure in the late 1870s (Lamb, 1982) in the dust-flux record, which is a calculation that incorporates both the dust concentration and the annual net balance. The dust concentration on the Dasuopu glacier is also linked to the magnitude of the Southern Oscillation[8] and the phase of the Pacific Decadal Oscillation (Davis, 2002), thus indicating a linkage between these two tropical processes. However, recent research on Tibetan Plateau ice cores drilled north of lat 32° N. shows that their climate records are influenced not only by the South Asian Monsoon and other tropical coupled atmospheric-oceanic processes such as the El Niño–Southern Oscillation (ENSO) (Davis and others, 2005) but also by atmospheric-pressure variations such as those seen in the North Atlantic Oscillation (NAO) (Wang and others, 2003; Davis and Thompson, 2004). Thus, the records from ice cores of high-resolution isotopes, chemistry, dust, and accumulation that were retrieved from across the Tibetan Plateau help us to accurately reconstruct the spatial and temporal variability of climate in this region.

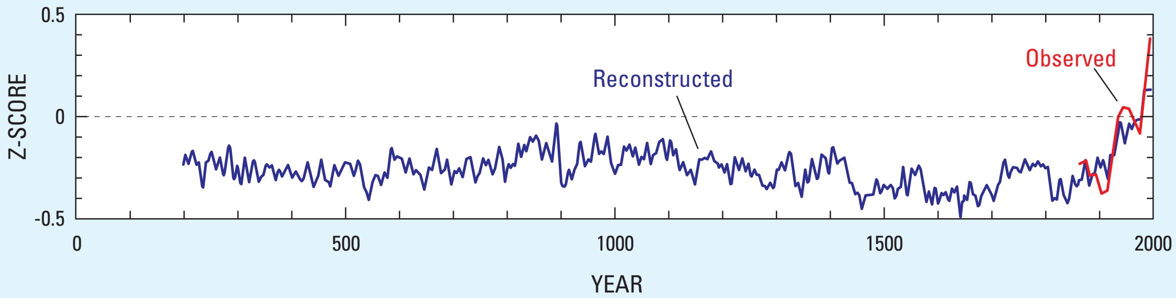
Ice-core proxy records document local, regional, and larger scale environmental fluctuations. There is little purpose in trying to reconstruct the history of global climate change from a single ice core, especially at high resolution on short time scales. However, certain parameters, such as $\delta^{18}O$, clearly record large-scale regional variability in sea-surface temperatures (SSTs), whereas other parameters, such as aerosols, may be more sensitive to local and regional conditions. Although the high-mountain ice-core records that extend back through the last two millennia differ regionally from one another and from the polar records, many of them document common climatic variations on hemispheric and even global scales. For example, composites of the decadally averaged $\delta^{18}O$ profiles of three South American ice cores (Nevado Huascarán, *Quelccaya Ice Cap*, and Nevado Sajama) and three Tibetan Plateau ice cores (Dunde and Guliya Ice Caps and Dasuopu Glacier) show different interhemispheric trends during the last millennium (Thompson and others, 2003) (fig. 38). For example, the Little Ice Age, a cold event between the 15th and 19th centuries that is noted in many northern European climate records (Grove, 1988), is more evident in the South American ice-core composite than that from the Tibetan Plateau. The Medieval Warming, a period before the Little Ice Age that appears in the Greenland ice-core records, is also more evident in the Andes time series. However, both of the composites show isotopic enrichment (indicating warming) beginning in the late 19th century and accelerating through the 20th century.

[8] [Editors' note: The Southern Oscillation describes changes in atmospheric air pressure above the Pacific Ocean.]

A. Locations of ice cores used in the ice core composite record



B. Reconstructed and observed Northern Hemisphere temperature record



C. Composite $\delta^{18}O$ record from Andes Mountains and Tibetan Plateau ice cores

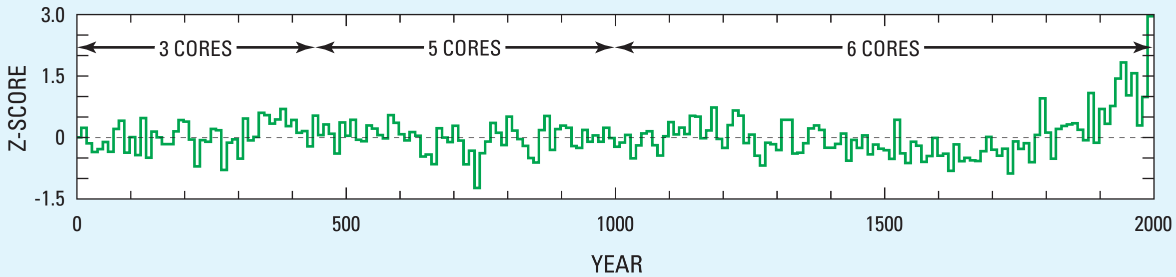


Figure 38.—**A**, *Geographic location of ice cores used in the ice-core composite record.* **B**, *Northern Hemisphere temperature records. The ice-core reference period for B is 1961–1990. The blue line is a reconstruction of temperature from Jones and Mann (2004); the short red segment of the line depicts the meteorological observations compiled by Jones and Moberg (2003).* **C**, *Composite of decadal averages of the isotope $\delta^{18}O$ from ice cores from the Andes Mountains and Tibetan Plateau during the past two millennia. The Z score is the standard deviation from its respective mean (Mosley-Thompson and others, 2006). The ice-core reference period for C is the 2,000 years preceding the 21st century.*

When all six of the profiles from these high-mountain glaciers are combined, the resulting profile is similar to the Northern Hemisphere temperature records of Mann and others (1998) and Jones and others (1998), which cover the last 1,000 years (Thompson and others, 2003) (fig. 38). Not only do these comparisons argue that temperature plays the most important role in the composition of oxygen isotopic ratios in glacier ice but they also demonstrate that the abrupt warming beginning in the late 19th century through the 20th century (and continuing into the 21st century) transcends regional influences, unlike earlier climatic variations. Indeed, on a global basis, the 20th century was the warmest period in the last 1,000 years, a period of time that also encompasses the interval of the Medieval Warming.

To capture a larger temporal time scale, the composite record of decadal $\delta^{18}O$ averages from these six ice cores has been extended back another 1,000 years. The composite consists of all the ice cores for the last 1,000 years, five for the last 1,550 years, and three for the last 2,000 years (fig. 38). Because 70 to 80 percent of the snowfall over both the Tibetan Plateau and Andes arrives during the Northern Hemisphere summer and winter, respectively, combining the six isotopic time series results in a more representative history from high elevations situated in low latitudes. The combined Tibetan Plateau and Andes records of decadally averaged $\delta^{18}O$ are shown in figure 38, along with the 2,000-year Northern Hemisphere multiproxy climate history, coupled with meteorological observations since 1860 (Jones and Moberg, 2003; Jones and Mann, 2004). These records place the dominant signal, that of the 20th century warming, within a longer-term perspective. This global-scale, high-elevation warming is significant because, although seasonal and annual temperature variations are small in the tropics, the high-mountain ice fields, ice caps, and other glaciers in these regions experienced significant and rapid loss of mass in the past two decades (for example, accelerated thinning and retreat of termini and margins) and this rapid loss is currently continuing. (See also table 3.1 (p. 19) and fig. 3.1 (p. 20) in "The Distribution of Glaciers in the Tropics" (Kaser and Osmaston, 2002).) The changes in high-mountain glaciers are discussed more fully below.

Post-1950 Climate Warming and Its Effects on High-Mountain Glaciers

Meteorological data from around the world suggest that the Earth's globally averaged temperature has increased +0.6°C since 1950. The El Niño year of 1998 saw the highest globally averaged temperatures on record, whereas 2002, a non-El Niño year, was the second warmest. The year 2003 was followed by 2004, a La Niña[9] year, and yet 2005 matched or exceeded the record temperature of 1998. The warming during the 20th century, which has accelerated during its last two decades, is recorded in high-mountain glaciers both within the ice-core records and by the rapid thinning and retreat of many of the ice fields, ice caps, and other glaciers. This glacier retreat is observed in almost all regions of the world:

- The Caucasus and other Eurasian mountain ranges in the mid-latitudes (Mikhalenko, 1997; Kotlyakov and others, 2010);

[9] [Editors' note: La Niña is defined as a phenomenon in a region of the equatorial Pacific Ocean characterized by a sea surface temperature that is ≥0.5° C colder than normal, averaged for three consecutive months.]

- Central Europe (Williams and Ferrigno, 1993);
- Western North America (Haeberli and others, 1999; Huggel and others, 2002; Williams and Ferrigno, 2002; Hoelzle and others, 2003; Meier and others, 2003); Alaska (Molnia, 2008);
- Iceland (Sigurðsson, 1998, 2005, 2006; Sigurðsson and Williams, 1998; Sigurðsson, and others, 2007);
- Tibetan Plateau (Shi Yafeng and others, 2010) ;
- Tropics (Allison and others, 1989; Thompson and others, 1993, 2000; Qin Dahe and others, 2000);
- Andes (Williams and Ferrigno, 1998);
- Irian Jaya, Indonesia (Allison and Peterson, 1989; Prentice and others, 2003);
- New Zealand (Chinn, 1989, 1996);
- Great Rift Valley Region of East Africa (Young and Hastenrath, 1991).

The case of Kilimanjaro is worth remaking. In 1912, the many ice fields on Kilimanjaro covered an area of 12.1 km^2, but in 2000 the ice covered only 2.6 km^2 (fig. 39). If the current rate of retreat and thinning continues, the perennial glacier ice on this mountain is likely to disappear by 2020 (Thompson and others, 2002). When the 20th century shrinking of Kilimanjaro's ice fields is placed within a 11.7-ka interval, the perspective of the ice-core records shows that the current wasting of these ice fields is unprecedented during the Holocene Epoch, even within the context of the evidence of intense and abrupt climatic disruptions in the past. For example, the ice-core evidence reveals that a prolonged drought occurred ~4.5 to 4.0 ka, coincident with the greatest drought historically recorded for tropical Africa. Kilimanjaro's Northern Ice Field (NIF), nevertheless, survived this drying event. In the visible stratigraphy of the longest NIF core, no major melt features are visible throughout the core's length except in the topmost meter. Comparing the chemical and physical properties preserved in the NIF with those in the nearby rapidly shrinking Furtwängler Glacier, which is saturated with water (see map (figs. 7, 8) and Landsat images (figs. 6, 7, 9) in Young and Hastenrath, 1991, p. G58–G63), confirms that conditions similar to those today (2009) have not existed in East Africa during the past 11 millennia (Thompson and others, 2002).

The ongoing retreat of glaciers in the Andes is now well documented (Ames, 1998; Williams and Ferrigno, 1998; Thompson and others, 2003; Thompson, 2004) and can be placed within a longer time perspective. The retreat of Qori Kalis, the largest outlet glacier from the *Quelccaya Ice Cap*, has been studied by terrestrial photogrammetry since 1978; it has the best documented recent history of any tropical glacier (figs. 24 and 40). The rate of its retreat from 1983 to 1991 (14 m a^{-1}) was almost three times the rate of its retreat from 1963 to 1983 (5 m a^{-1}); during the 2000 and 2001 balance year, its rate of retreat reached 205 m a^{-1}. The retreat stopped during two periods (1991–1993 and 2002–2003), but the glacier resumed its retreat in subsequent years. Figure 40 is a 10-map sequence that documents the rapid and acceler-

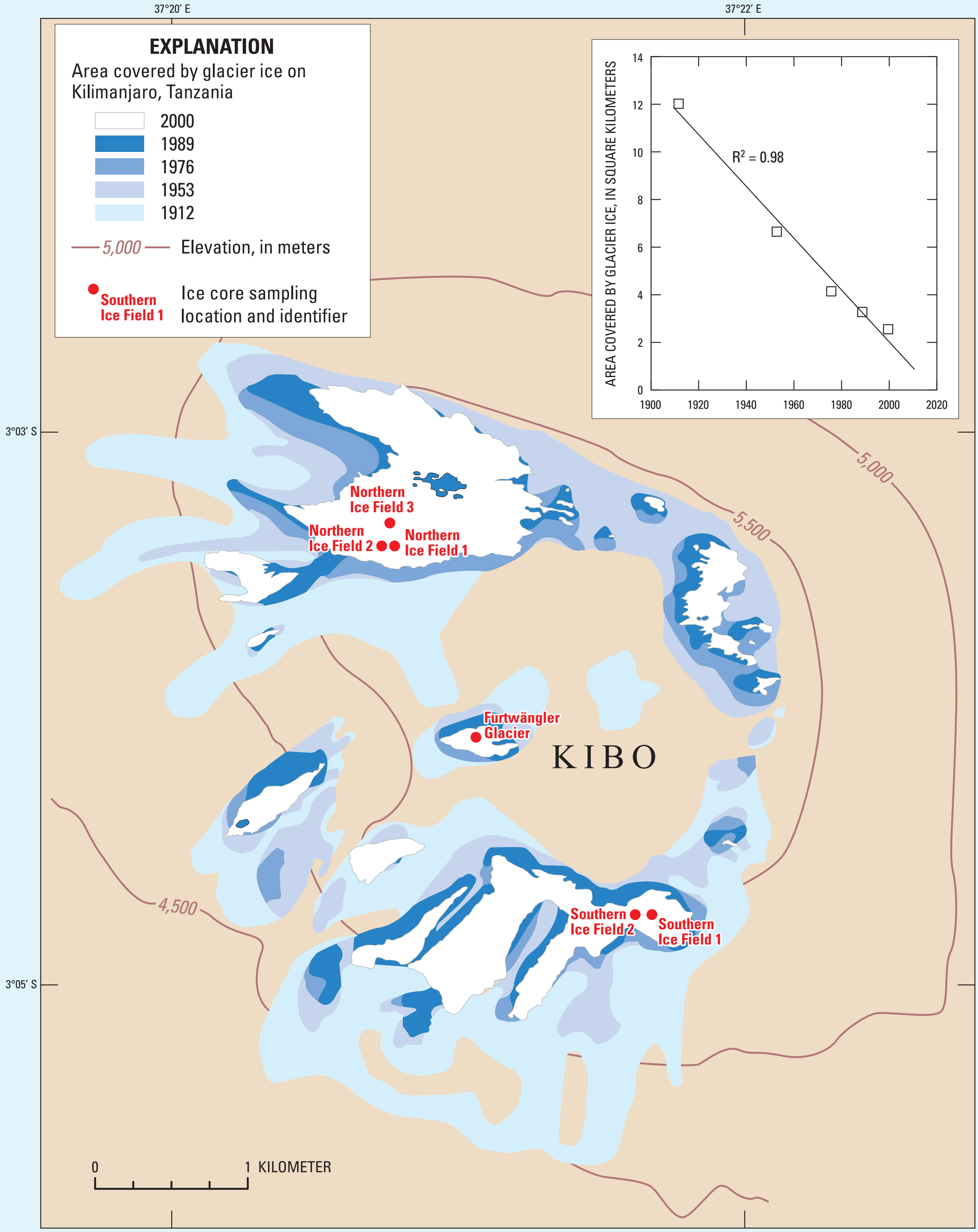


Figure 39.—*Outlines of the Kilimanjaro Ice Fields in 1912, 1953, 1976, 1989, and 2000. The inset graph shows a least-squares regression line plot of the 5 discrete years of area calculations and the nearly linear (R^2, the coefficient of determination, is 0.98) decrease in ice area from 1912 to 2000. The red dots on the map indicate locations of ice-core sites, three from the Northern Ice Field, one from the Furtwängler Glacier, and two from the Southern Ice Field. See also Young and Hastenrath (1991) and Hastenrath and Greischar (1997).*

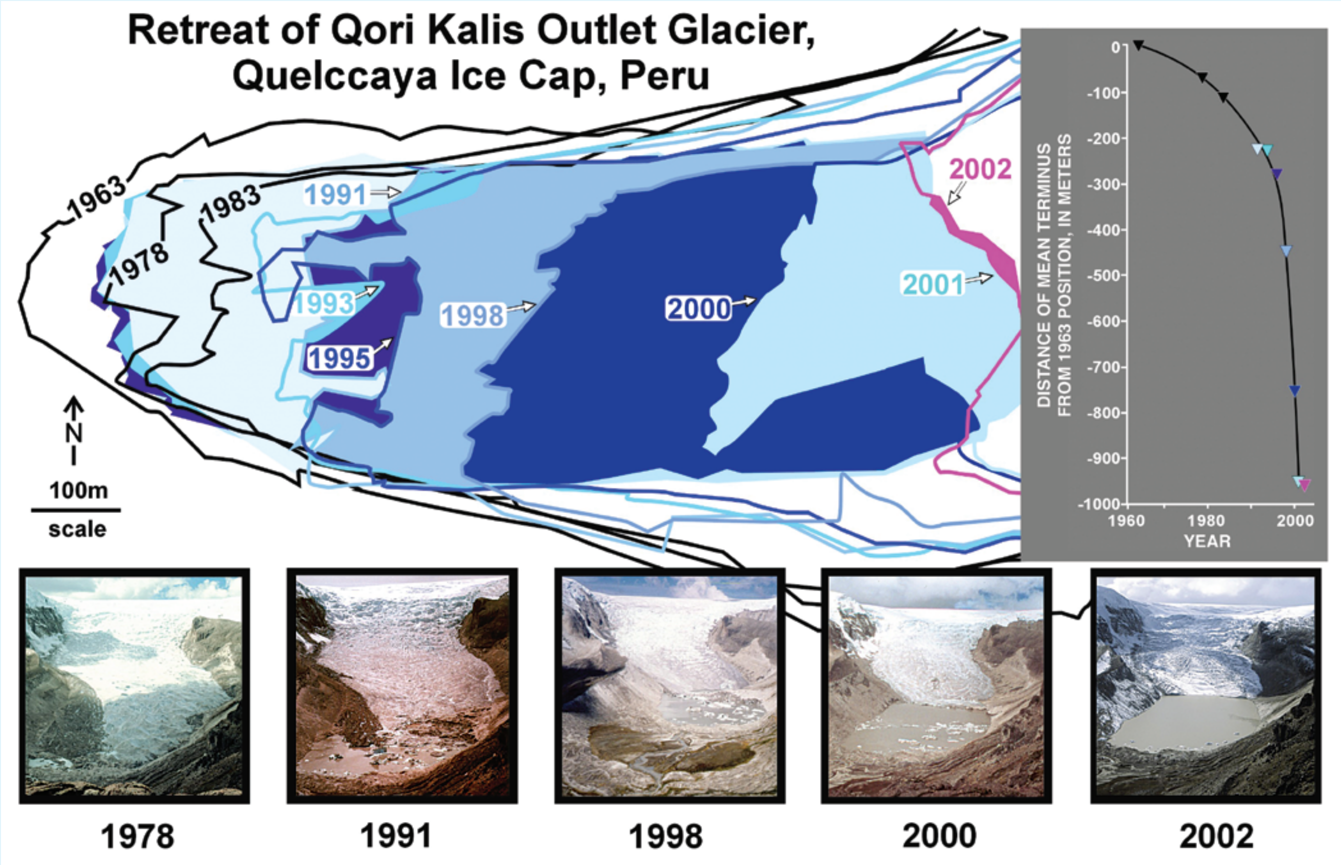


***Figure 40.**—Photographic and cartographic history of the retreat of Qori Kalis, an outlet glacier from the Quelccaya Ice Cap, Perú, 1963–2002. Accompanying graph shows retreat of terminus during that period of time.*

ating wastage, in which the Qori Kalis outlet glacier terminus is receding about 10 times faster during the period from 1987 to 2002, than it receded during the initial measurement period from 1963 to 1978.

The shrinkage that is affecting the Qori Kalis outlet glacier is actually occurring around most of the margin of the *Quelccaya Ice Cap* (fig. 24), and the rapid melting of the ice is now exposing plant material that was buried between 5.2 and 4.8 ka (accelerator mass spectrometer ^{14}C dated), a time during the middle of the Holocene when glaciers throughout the Andes began expanding (Mark and others, 2002; Thompson, 2004). These localized data, coupled with the ~18,000-year δ^{18}O history from Nevado Huascarán (Thompson and others, 1995), suggest that warmer and drier conditions prevailed in the early Holocene (~11 to 6 ka) and that less extensive glacierization of high mountains accompanied the drought. Around 5 ka, cooler, wetter conditions prevailed, initiating the expansion of high mountain glaciers. Multiproxy studies of sediment from glacial lakes throughout the central region of the Andes corroborate an overall aridity through the late glacial stage and the early Holocene. After 4.8 ka, wetter conditions and glacier expansion followed, with the last 2,300 years being the wettest of the Holocene. Conditions were not always stable on centennial to millennial time scales, however, and considerable variability in climates is evident on the scale of decades to centuries (Abbott and others, 2003).

Future Priorities

Ice cores from high-mountain glaciers are unique and valuable archives of climate information because they record variations in atmospheric chemistry and conditions. Ice cores from the tropical and subtropical latitudes are especially valuable, because these latitudes comprise more than 50 percent of the Earth's surface area where the major energy that drives the global climate system is absorbed and where most of the world's population is concentrated. Since 1982, the phenomenon that combines El Niño and the Southern Oscillation has captured worldwide attention as populations and governments have come to realize the wide geographic extent of the phenomena's widespread and often devastating effects on the weather in many regions. As we begin to understand how this coupled atmospheric-oceanic process works, we also see its linkages with other important systems such as the Asia/African monsoons. Because both of these tropical systems influence precipitation and temperature over large regions, their effects are also recorded in the chemistry (and in the net accumulation) of snow that falls on high-mountain glaciers. Seasonal and annual resolution of chemical and physical parameters in ice-core records from the Andes have allowed glaciologists to reconstruct the variability of the ENSO phenomenon during several hundred years (Thompson and others, 1984, 1992; Henderson, 1996; Henderson and others, 1999). Because the effects of El Niño and La Niña events vary geographically, ice-core records are needed from the entire length of the Andes—from the Sierra Nevada de Mérida, Colombia, to the Cordillera Darwin of southern Patagonia—in order to more fully resolve both the temporal and spatial variability of the ENSO phenomenon, including their frequency and intensity, along with temperature variations that occurred long before human documentation.

The variability of the South Asian monsoon is also of vital importance for the large percentage of the world's population that lives in the affected areas. Cores from the Tibetan Plateau have yielded variability in the annual- to millennial-scale histories of monsoons across this large region, thereby affording us information on the interaction between the monsoon system and the prevailing westerlies that can be traced back (and linked by teleconnection) to atmospheric processes in the Atlantic Ocean. Recent research has suggested that the current warming trends may be affecting the strength of the Asia monsoon system (for example, Anderson and others, 2002) and that information about past climate on the Tibetan Plateau and its association with monsoonal strength can be gained from additional analyses of ice cores and from related glacier studies in the Himalaya.

Some of the most dramatic evidence for the warming of the global climate that is underway today (2012) comes from the drastic recession of both total area and total volume of high-mountain glaciers. This rapid retreat causes concern for two reasons.

First, these glaciers are regional "water towers," and their loss threatens water resources necessary for producing hydroelectric power, for crop irrigation, and for municipal water supplies in many nations. The high-mountain ice fields, ice caps, and other glaciers constitute a "savings account" that is drawn upon during dry periods to provide water supplies to human populations downstream. The accelerated melting of high-mountain glaciers is rapidly depleting that "savings account," an "account" that was built up for hundreds to thousands of years but that current climatic conditions are not replenishing. Figure 41 illustrates, the current thinning and areal shrinkage of almost all the Earth's high-mountain glaciers. The land between latitudes 30° N. and 30° S., which constitutes 50 percent of the global surface area, is home to 70 percent of the world's population and to 80 percent of the world's growth in population (births). However, only 20 percent of global agricultural production takes place in these climatically sensitive regions.

The second concern that the disappearance of these high-mountain glaciers brings about is that they contain paleoclimatic histories that are unattainable from any other source. As the glaciers melt, the records of variability of the change in the environment and in the climate preserved within them are lost forever. Only these scientifically valuable records can assure that we accurately determine how climate has changed in the past in many regions and that we carefully assess the probable magnitude and rate of future changes.

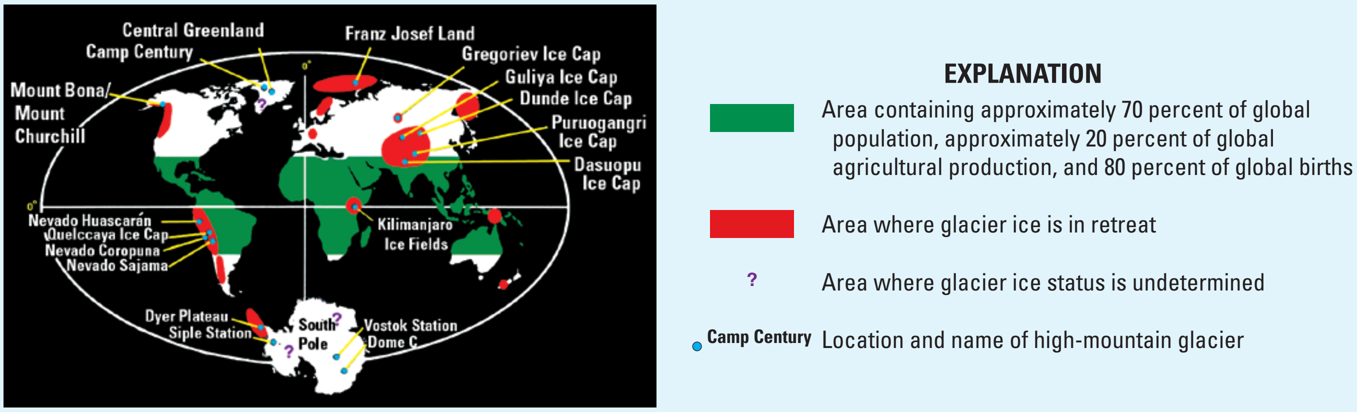


Figure 41.—*Late 20th-century status (retreat or advance) of selected high-mountain glaciers of the Earth's cryosphere, the location of ice-core sites, and the contemporary location of important human activities.*

Sea-Level Variability and Volume of Glacier Ice on Land

Introduction

A common "thread" addressed throughout Part A-2 (Glaciers) is the relationship between changes in sea level and the volume of glacier ice on land. The change in global (eustatic) sea level during the numerous glacial and interglacials of the last 2.7-Myr-long Pliocene-Pleistocene Ice Age is reviewed (Barron and Thompson, 1990; Matthews, 1990; Raymo and others, 2006), but the last 1 myr, when 100,000-year cycles of glacial and interglacials dominated the Earth's climate system, are especially emphasized.

Many scientists have examined the ~125-m rise in eustatic sea level since the Last Glacial Maximum (LGM) about 18,000 years ago (Fairbanks, 1989; Pirazzoli, 1996; Clark and Mix, 2002; Day and others, 2007). The acme of the LGM varied geographically with respect to maximum volume of glacier ice sequestered in ice sheets, ice caps, and mountain glaciers. For the Northern Hemisphere, the LGM ranges from 26.5 ka to 19 ka to 20 ka; for the part of the Antarctic ice sheet in West Antarctica it ranges from 15 ka to 14 ka (Clark and others, 2009). About 7 ka sea level had more or less stabilized to its present position (Stanley and Warne, 1997; Day and others, 2007).

Definition of Sea Level

"Sea level" refers to "the height of the sea surface" (Neuendorf and others, 2005, p. 581) and is a synonym for "mean sea level" (mean level of the sea between mean high tide and mean low tide). Mean sea level is used as a datum for nautical charts and is defined as "the arithmetic mean of hourly heights observed [on tide gauges] over some specified time, usually 19 years" (Neuendorf and others, 2005, p. 400). Coastal charts published by the National Ocean Survey (NOS) of the National Oceanic and Atmospheric Administration use mean low water as the chart datum; USGS topographic maps of coastal areas use mean high water as the datum.

Geologists refer to two different types of sea level—relative and eustatic. Relative sea level refers to the height of mean sea level relative to a specific geographic feature or other defined datum on a coast or estuary. On a geological scale, changes in relative sea level (positive or negative) result from natural processes, such as postglacial rebound of the Earth's surface within or adjacent to a former ice sheet (Mitrovica and Vermeersen, 2002a, b), tectonic subsidence or uplift, eustatic changes, and compaction of sediments on a delta. Human activities, such as damming rivers, withdrawing subsurface fluids (groundwater, oil, gas), and diverting river channels, can also change relative sea level.

"Eustatic sea level" pertains to global changes of ocean volume. Eustatic changes dominant in the past few million years were caused by addition or removal of water from continental ice caps [sic, ice sheets]" (Neuendorf and others, 2005, p. 220). Steric changes in ocean volume caused by variation in sea-water temperature also create eustatic changes (Roemmich, 1990). Historically, mean sea level in coastal regions has been measured by tide gauges. Mean sea level in the open ocean is now measured by geodetic sensors on Earth-orbiting satellites (Sahagian and Zerbini, 2002; Cazenave and others, 2009; Wdowinski and Eriksson, 2009). Using modern Global Positioning

System of satellites (Wells and others, 1986), geologists can accurately measure latitude, longitude, and elevation above the local geoid at tide-gauge stations. An integrated system to measure long-term changes in global sea level is needed (Eden, 1990; Peltier and Tushingham, 1991).

Variation in Volume of the Earth's Oceans

Sean C. Solomon (written commun, 1989) compiled table 12, which shows that variation in ocean volume is the integral of 16 elements and processes in the Earth System: cryospheric, atmospheric, hydrospheric, lithospheric, and climatic. Those 16 processes can change the ocean volume on different scales of historical and geologic time. Bowen (1978, p. 158), however, listed the five primary processes that control sea level over time: "(1) long term tectonic changes, (2) glacial isostasy, (3) hydro-isostasy, (4) geoidal changes, (5) glacio-eustatic movements in sea level due to alternating glaciation and deglaciation."

With the current configuration of the Earth's continents ("large islands") and small islands, the volume of glacier ice currently (2012) on land sustains a sea level that is in dynamic equilibrium between the 29.1 percent of non-glacierized land above sea level (that is, Earth's terrestrial surface) and the 70.9 percent of non-glacierized land below sea level (that is, sea floor) (National Geographic Society, 2005, p. 136; middle right graphic on pl. 1). Some of the land area covered by the Greenland and Antarctic ice sheets is below sea level owing to an isostatic depression of the Earth's crust caused by the weight of the glacier ice. The disappearance (melting) of the two ice sheets would result in isostatic uplift of that depressed crust. See Alpha and Ford's (1989) isometric maps of Antarctica from three perspectives: with present-day ice sheet, with no ice sheet, and after isostatic rebound, the so-called "glacial rebound," a process still ongoing in northeastern North America and in the Nordic countries (Norden) (Schneider, 1997). To accurately determine the maximum potential sea-level rise (table 3), one would need to subtract the volume of glacier

Table 12.—*Ocean-volume variations—An integral of many effects that must be measured and understood*

[Modified from viewgraph presented by Sean C. Solomon, Department of Earth, Atmospheric, and Planetary Sciences, Massachusetts Institute of Technology, at the NASA Solid Earth Science Workshop, Coolfont, West Virginia, July 1989]

[Editors' note: Glaciers and ocean temperature are shown in bold font because meltwater from glaciers and warming of the ocean are the principal processes causing the present-day rise in global sea level]

Glaciers	Continental freeboard
Ocean temperature	Erosion
Postglacial rebound	Historic record
Sea level	Seismic stratigraphy
Rotation of the Earth	Global climate warming
Wobble excitation	Plate motion rate
Wind stress	Magnetism
Sedimentation rate	Global change

ice below sea level beneath both ice sheets from the total volume of ice given for each ice sheet. No such correction was made for table 3 because the total volume of ice below sea level in both ice sheets is not well enough known.

During the Last Glacial Maximum (Wisconsinan), about 20,000 ka, the volume of glacier ice on land was about three times greater than today (Flint, 1971; van den Heuvel and Buurman, 1974). As a consequence, eustatic sea level was 125 m lower than today (Fairbanks, 1989; Clark and others, 2001). About 8 percent more land was above sea level, so 37.1 percent of the Earth's surface was terrestrial and 62.9 percent was covered by oceans (Williams and Hall, 1993) (refer to upper right graphic on pl. 1). The bottom right graphic on plate 1 shows the current continental configuration with no glacier ice existing on land. On a glacier-free Earth, sea level would rise about 80 m, which would significantly reduce the area of land above sea level (to about 25 percent). The oceans would expand to cover about 75 percent of the Earth's surface. Plate 1 (lower right graphic) shows the substantial inundation: Florida; the Atlantic Coastal Plain; Mississippi River delta and embayment in the United States; Amazon and other river basins in South America; northern and central Russia; south-central Australia; and, most importantly, the heavily populated, food-producing deltaic areas in China, Pakistan, India, and Bangladesh. The last time that the Earth was essentially free of glacier ice was ~45 Ma during the Eocene, when excessive atmospheric CO_2 (≥1,000 ppm, 2.5 times higher than at present (2012) and 3.5 times higher than at the beginning of the "Industrial Era") raised the mean atmospheric temperature +6°C above current mean temperature (Zachos and others, 2001; Kennedy and Hanson, 2006). Kennedy and Hanson (2006) also estimate that sea level was ~50 m higher than present.

History of Sea-Level Change During Phanerozoic Time

Servais and others (2009) (fig. 42) show the variation in eustatic sea level during the past 542 myr (Phanerozoic Eon or Eonothem). Eustatic sea levels were relatively high during most of the Early Paleozoic Era (100 to 200 m above present (2012) sea level; Haq and Schutter, 2008), and they were declining in the Late Paleozoic during an interval of continental glaciation when the Pangean Supercontinent was intact (at about 230 Ma) (fig. 30). Eustatic sea level rose during most of the Mesozoic Era, when Pangea had separated into two continents (at about 38 Ma) (fig. 30). After these two continents fragmented, sea level began a fluctuating decline throughout the Cenzoic Era

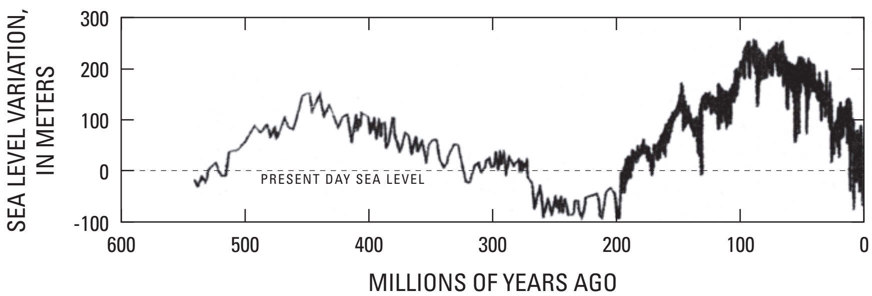


Figure 42.—*Variation in global (eustatic) sea level during the Phanerozoic Eon. Modified from Servais and others (2009, p. 5, fig. 1 (c)).*

(Browning and others, 2008) (fig. 28). This deline accelerated when ephemeral ice sheets began to form in the late Eocene Period (fig. 28). Figure 43 also shows the increase in amplitude of eustatic fluctuations during the Ice Age.

The best sea-level data came from the latter part of the Ice Age, when the 100,000-yr cycles of glacial-interglacial dominated (fig. 44), especially information about changes in ice volume and sea level during the last glacial (Lambeck and Chappell, 2001). Fluctuation during the last four glacial cycles ranged from ~120 to 140 m below present sea level to ~5 to 10 m above (fig. 45). Dutton and others (2009) concluded that sea level was at least 18 m higher than present 245,000 to 190,000 years ago (Mindel-Riss or Yarmouth interglacial; MIS 7). A lower sea level accompanies a greater volume of glacier ice on

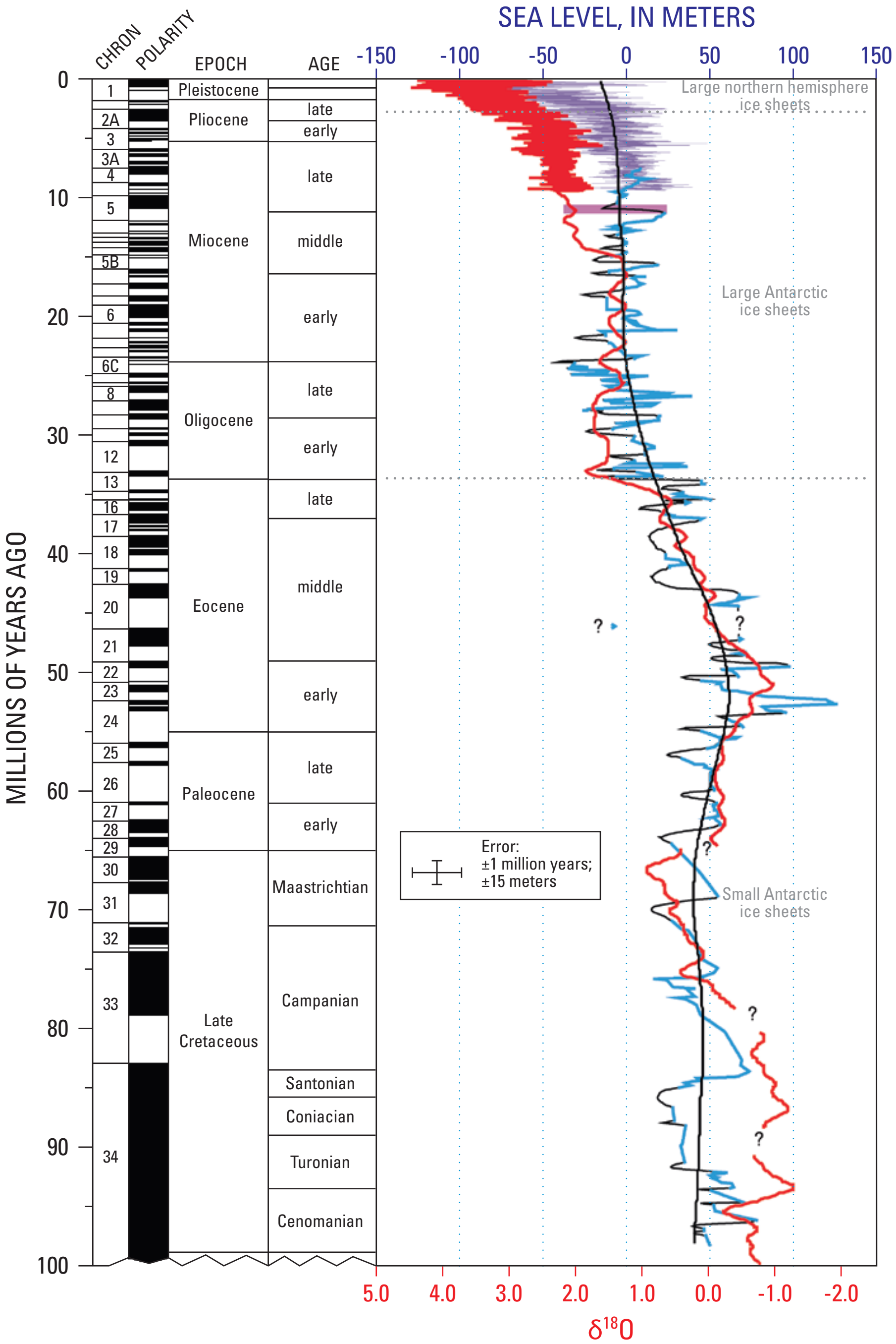


Figure 43.—*Variation in global sea level during the last 100 Ma (light blue and purple lines), from the Late Cretaceous of the Mesozoic Era through the Cenozoic Era to the present time. The red line is a benthic foraminiferal $\delta^{18}O$ synthesis; the purple line is for the interval 0 to 7 Ma. Note the fluctuating decline in sea level (light blue, 7 to 100 Ma) after the peak in the early Eocene. Modified from Miller and others (2005, p. 1,295, fig. 3).*

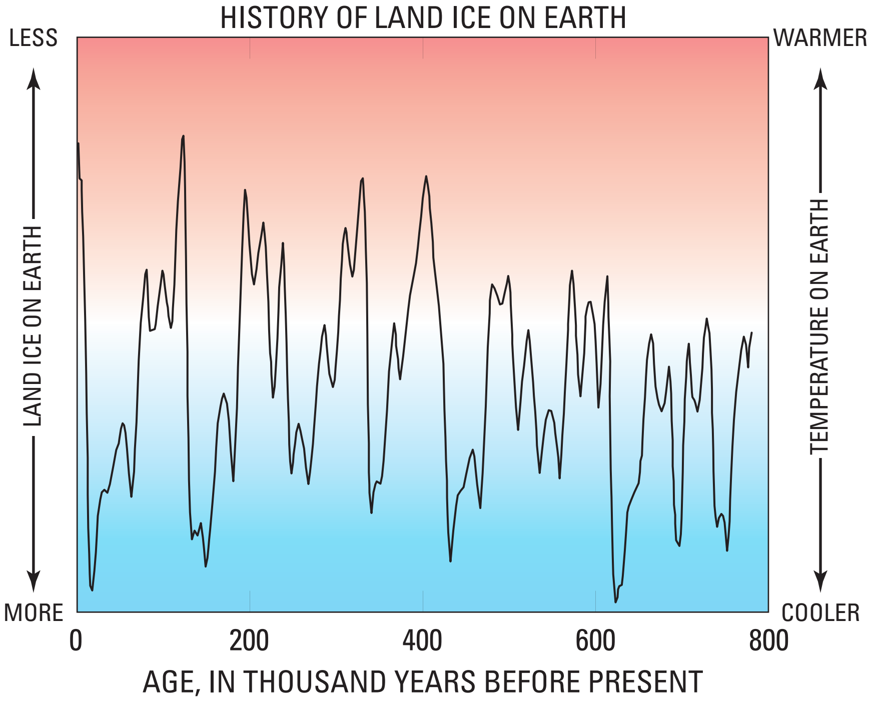


Figure 44.—*Volume of glacier ice on land during the past 800,000 years. Modified from Alley (2000, p. 94, fig. 10.1).*

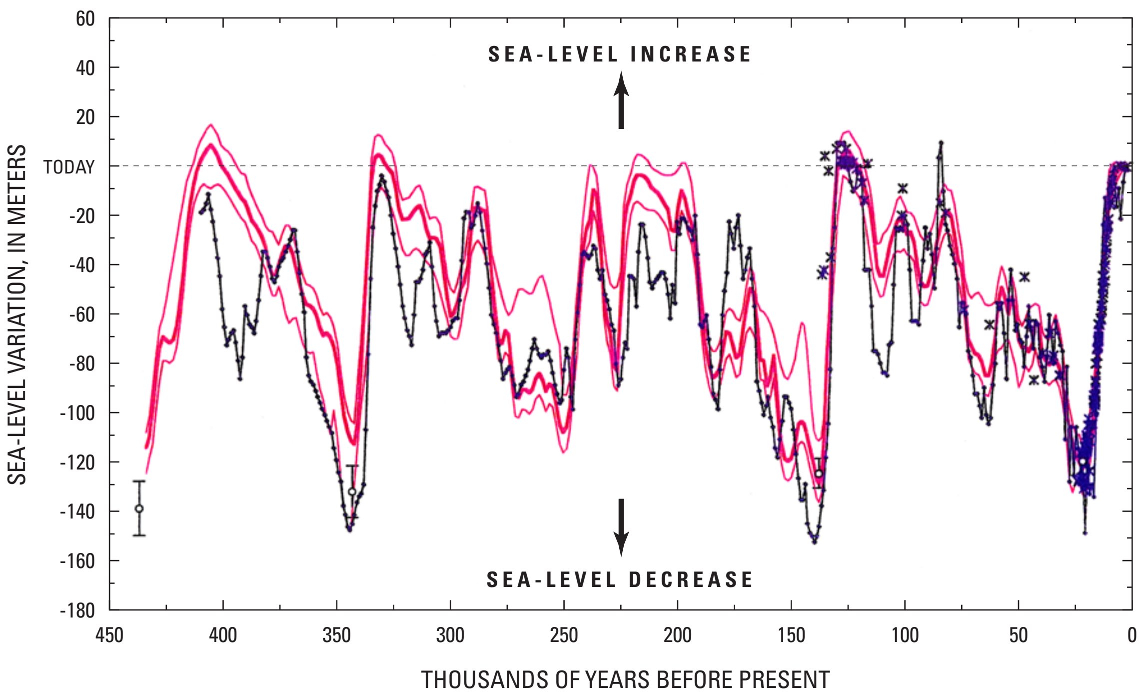


Figure 45.—*Various estimates of changes in global sea level during the last 440,000 years: four glacial cycles of about 100,000 years each. Modified from IGBP Science No. 3 (Alverson and others, 2001, p. 7, fig. 1).*

land during glacials, about three times the current volume of glacier ice on land (Flint, 1957), (see graphic in upper right of pl. 1). A higher sea level represents less glacier ice on land (see graphic in upper right of pl. 1). Figure 44 shows fluctuations (volume of glacier ice on land) during the last 750,000 years. Because "the past is the key to the future," a key issue to be discussed later in this section is the source of melting glacier ice as an interglacial begins. With only 0.5 m of ice on land available from non-ice-sheet glaciers, the other 4.5 to 9.5 m rise in sea level characteristic of previous interglacial intervals during the past 1 Myr must have been derived from the melting of the Greenland and Antarctic ice sheets (Mercer, 1970; Bamber and others, 2009; Ivins, 2009). Because the Greenland ice sheet can account for only 6.5 m of the rise, melt from both ice sheets therefore is necessary to account for a rise of 9.5 m (table 3).

In addition to glacial meltwater added to ocean volume during climate warming, thermal expansion (steric increase) also contributed to the rise (table 13) (Domingues and others, 2008). During the late 20th century, oceanic thermal expansion accounted for 57 percent of the total annual rise in sea level (Church and others, 2007). The other 43 percent was caused primarily by glacial meltwater from non-ice-sheet glaciers (Hall and Williams, 2007; Meier and others, 2007).

Under climatic conditions of a "Super-interglacial" (fig. 27) (Lubick, 2006), however, the sea-level rise might exceed 9.5 m. Kennedy and Hanson (2006, p. 1673) write, "As subsequent glaciations came and went, CO_2 concentration and temperature were tightly linked. When both went down, ice sheets grew and sea levels sank, lower than today's by more than 100 m. When both went up, there were relatively stable warm periods with high sea levels." The preceding statement is based on past variations of atmospheric CO_2 which ranged from about 180 ppm (glacial) to 280 ppm (interglacial).

How will components and elements of the Earth System respond to global climate warming if the CO_2 concentration in the atmosphere passes 400 ppm (predicted by 2012) and continues to increase? Will the increased volume

Table 13.—*Observed rate of global rise in sea level and estimated contributions to that rise from thermal expansion and meltwater from non-ice-sheet glaciers and the Greenland and Antarctic ice sheets*

[Data prior to 1993 are from tidal gages worldwide; data following 1993 are from satellite altimetry. Modified from table SPM–1 in Intergovernmental Panel on Climate Change (2007b, p. 7)]

Source of rise in sea level	Rate of rise in sea level (millimeters per year)	
	1961–1993	1993–2003
Thermal expansion	0.42 ± 0.12	1.6 ± 0.5
Non-ice-sheet glaciers	0.50 ± 0.18	0.77 ± 0.22
Greenland ice sheet	0.05 ± 0.12	0.21 ± 0.07
Antarctic ice sheet	0.14 ± 0.41	0.21 ± 0.35
Sum of contributions from thermal expansion and meltwater from glaciers to rise in sea level	1.1 ± 0.5	2.8 ± 0.7
Observed total sea level rise	1.8 ± 0.5	3.1 ± 0.7
Difference—Observed contributions from thermal expansion and meltwater from glaciers to rise in sea level	0.7 ± 0.7	0.3 ± 1.0

of CH_4 from melting terrestrial and submarine permafrost cause even more rapid warming (Brook and others, 2008)? Will non-ice-sheet glaciers and ice sheets melt faster (Truffer and Fahnestock, 2007) and accelerate eustatic sea-level rise? Is the volumetric discharge rate of glacial meltwater into the oceans during the late Pleistocene and early Holocene that Fairbanks (1989) documented and that Blanchon and Shaw (1995) and Overpeck and others (2006) discussed relevant to the Earth's future?

Sea Level, Ice, and Climate Change

Introduction

In 1940, the Icelandic geologist Sigurdur Thorarinsson (1940) addressed the question of glacier shrinkage in relation to the eustatic rise in sea level. He reviewed the conclusions reached by several scientists with respect to changes in eustatic sea level since the Last Glacial Maximum (LGM) after the loss of so much glacier ice on land. In a review of the available scientific literature on the state of glaciers in all the glacierized regions of Earth, Thorarinsson (1940, p. 147) concluded that "the majority of the glaciers in practically every glacier district of the world are now receding, that is, the present glacier shrinkage is a universal phenomenon." Thorarinsson was referring to non-ice-sheet glaciers. He concluded, from scant evidence, however, "that the inland ices [ice sheets] of Antarctis [Antarctica] and Greenland have probably not been affected to any appreciable extent…" (Thorarinsson, 1940, p. 147). Sixty years later, at the beginning of the 21st century, Thorarinsson's conclusion still would have been accurate in describing the state of the Earth's non-ice-sheet and ice-sheet glaciers.

With better data acquired by airborne and satellite remote sensors during the late 20th century, the state of the Antarctic ice sheet (see Chapter B of this volume) and the Greenland ice sheet (see Chapter C of this volume) began to attract the attention of geologists and glaciologists from many nations. Before remote sensors provided such data, glaciologists had focused on compiling long-term records of fluctuations of termini and of variations in mass balance, and they had compiled inventories of non-ice-sheet glaciers (for example, World Glacier Monitoring Service and the GLIMS consortium).

The first major symposium to comprehensively address changes in glaciers in connection with changes in sea level was that on "Sea Level, Ice, and Climatic Change," sponsored by the International Association of Hydrological Sciences. The symposium volume included several papers on sea level, ice sheets and climate (Allison, 1981). Revelle (1983), Robin (1987), and Sundquist (1990) discussed the relationship of the increase in concentration of carbon dioxide in the Earth's atmosphere to future changes of sea level. The Committee on Glaciology, Polar Research Board, National Research Council (U.S.A.), sponsored a major Workshop on "Glaciers, Ice Sheets, and Sea Level: Effects of a CO_2-Induced Climatic Change" (National Research Council, 1985). The Executive Summary of the Report of the Workshop (National Research Council, 1985, table 1, p. 2) summarized what was known at the time about the mass balance of non-ice-sheet glaciers and ice sheets. Both the Greenland and the Antarctic ice sheets had a positive mass balance, causing a small per annum drop in sea level; the non-ice sheet glaciers had a negative mass balance, resulting in a small per annum rise in sea level. The overall state of knowledge about the Earth's glaciers, however, was inadequate. Significant error limits, not statistically valid, are given in table 1 of the Executive Summary, accompanied by the following statement:

> The glaciers and small ice caps [non-ice-sheet glaciers] of the world, excluding the ice sheets of Antarctica and Greenland, have, in general, been shrinking during the past 100 years (Table 1). However, the data set is temporally and spatially sparse. Most of the glacier wastage that contributes to sea-level rise is thought to be derived from the mountain ranges bordering the Gulf of Alaska, Central Asia, and the Patagonian ice caps [sic, ice fields], but these are areas with few observations. In some regions, even the area of glacier is poorly known.

Although most of the knowledge about changes in mass balance of glaciers globally was limited to non-ice-sheet glaciers, especially mountain glaciers, glaciologists increasingly focused on scientific studies of the Antarctic ice sheet. Alley (1989, p. 45) summed up the prevailing opinion among glaciologists during the latter part of the 20th century on the source of meltwater from glaciers:

> In summary, we know that sea levels have fluctuated over the last hundreds of thousands of years largely because of changes in the volume of major ice sheets. Currently, sea level is rising slowly because of thermal expansion of the oceans and the melting of mountain glaciers. Greenhouse warming in the future can be expected to accelerate mountain glacier melting and begin melting the Greenland ice sheet, but it will also remove some water from the oceans through increased snowfall in Antarctica. The net result would be slow or zero sea-level rise over the next century or longer. However, dramatic sea level rise would occur if the greenhouse warming trend destabilized the marine West Antarctic ice sheet [Editors' note: see more recent papers by Bamber and others, 2009; Ivins, 2009; Hurtley, 2009]. Such an event would result in tremendous economic dislocation and the potential for human disaster. Studies to date have discovered new feedback mechanisms that tend to stabilize the ice sheet, but much work remains to be accomplished before confident predictions can be made.

Sea-Level Rise During the 20th Century

Sea level has been rising since the end of the Little Ice Age, and the rate of rise has accelerated during the past century (Gornitz and others, 1982), according to tide-gauge measurements (fig. 46). Church and White (2006) show the change in rate from 0.8 mm a^{-1} from the 1880s to the 1920s, 2.0 mm a^{-1} from the 1920s to the 1980s, and 3.2 mm a^{-1} from the 1980s to 2000; the average rate of rise is about 1.6 mm a^{-1} during 120 years of tide-gauge observations, a total increase in sea level of 21.6 cm. Miller and Douglas (2004) report a rate of 1.5 to 2.0 mm a^{-1} for the 20th century; they also concluded that the contribution to the increase in rise from glacial meltwater was greater than that from ocean warming. Peltier and Tushingham (1991) corrected tide-gauge measurements at 500 sites for the effect of glacial isostasy and determined that sea level was rising at a rate of 2.4±0.9 mm a^{-1}. Nerem and others (2006) reviewed the state of knowledge with respect to present changes in sea level.

Barnett (1983) analyzed the research previously reported by other scientists and his own analysis of sea-level data from many tide-gauge sources; he concluded that sea level was rising at the rate of 15 cm per century. At the same time, the U.S. Environmental Protection Agency (Hoffman and others, 1983) provided estimates of future sea-level rise through the year 2150. Previous research had estimated a range of >5 cm per century, based on analysis of loss of glacier ice (Thorarinsson, 1940), to 30 cm per century (Emery, 1980) based on tide-gauge records.

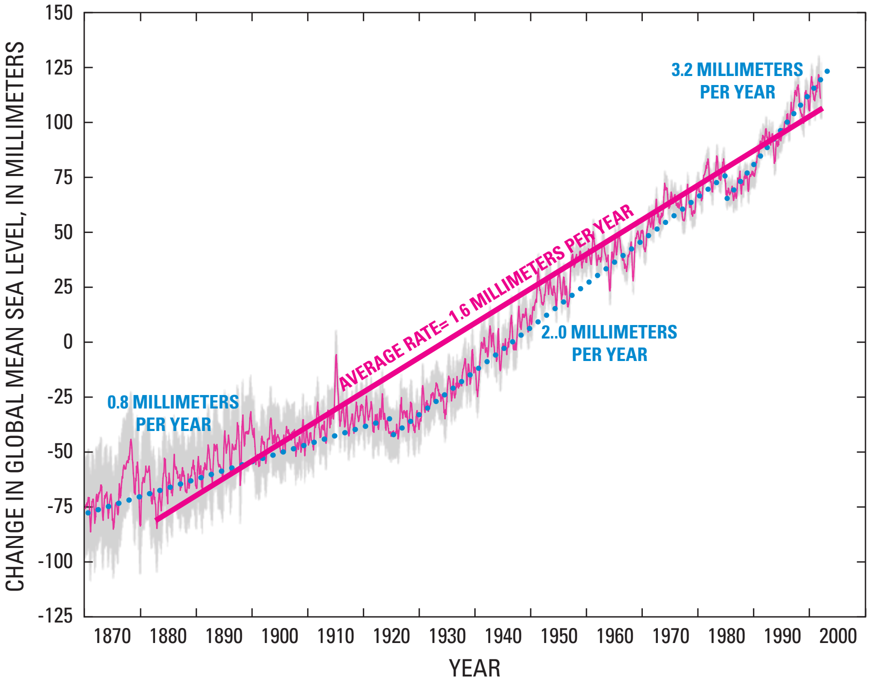


Figure 46.—*Tide-gauge observations from about 1870 to 2005. Note change in rate of rise in sea level, the result of steric increase in volume of a warmer ocean and from runoff of glacial meltwater. Modified from Church and White (2006).*

Glaciologists also began to examine the extent to which changes in volume of glaciers contributed to the rise in sea level. Meier (1984) wrote a pioneering paper on the shrinkage (loss in mass) of non-ice-sheet glaciers for the period 1900 to 1961, focusing on the 10- to 15-cm rise in sea level during the past century; he concluded that melting of non-ice-sheet glaciers contributed one-third to one-half of rise in sea level. The fraction matches the rise in sea level not accounted for by thermal expansion (steric) of the oceans.

For more than two decades since publication of that important paper, Meier and his colleagues (Meier and others, 2007) continued to stress the conclusion, from many analyses of data on fluctuations of glaciers and changes in mass balance, that non-ice-sheet glaciers were the dominant source of glacial meltwater contributing to the rise in eustatic sea level and that the non-ice-sheet glaciers, including ice caps, would remain the dominant source during the 21st century.

In 1985, the National Research Council's (1985) report on the workshop "Glaciers, Ice Sheets, and Sea Level: Effects of a CO_2-Induced Climatic Change" concluded "that the Antarctic Ice Sheet will contribute between 0 and 0.3 meters of sea level change in a CO_2-enhanced environment by the year 2100. The Greenland Ice Sheet will contribute between 0.1 and 0.3 meters (Reeh, 1985), and glaciers and small ice caps [non-ice sheet glaciers] could add a similar amount." Meier (1990a, p. 115) summarized the conclusions reached in a symposium sponsored by the American Geophysical Union on "Sea Level Change," writing: "By 2050, small glaciers and ice caps are expected to contribute +0.16±0.14 m; the Greenland Ice Sheet, +0.08±0.12 m; and the Antarctic Ice Sheet, -0.3±0.2 m." Meier (1990, p. 116) also stated:

> The consensus estimates reported here have large uncertainties, reflecting both our incomplete knowledge of processes and our lack of sufficient observational data. It does appear that a sea-level rise of 1 m by 2050 is unlikely. But even a 30-cm rise will cause social and economic problems in low-lying areas: this modest rise corresponds to a retreat in shoreline of 30 m or more,

> unless artificial protection is established. There would also be intrusions of saltwater in estuaries and groundwater aquifers, some destruction of coastal wetlands and an increased frequency of damage from storm surges.

In 1990, the U.S. Global Change Research Program incorporated sea-level change into its Fiscal Year 1991 Research Plan ("Our Changing Planet") (Williams, 1989).

Warrick and others (1993) edited an important book, "Climate and Sea Level Change: Observations, Projections and Implications," providing an excellent synthesis on the state of knowledge regarding climate and sea-level change. In the "Data" section of the book, Gornitz (1993) and Aubrey and Emery (1993) discussed recent global sea-level changes with respect to land levels. Three chapters in the "Projections" section of the book "looked" into the future with respect to changes in sea level caused by (1) the increase in global mean temperature (Wigley and Raper, 1993); (2) meltwater from non-ice-sheet glaciers (Kuhn, 1993); and (3) possible changes in the mass balance (volume) of ice sequestered in the Antarctic ice sheet and the Greenland ice sheet in a key chapter by Oerlemans (1993b).

In the late 1990s, several glaciologists, analyzed changes in volume (mass balance) of non-ice-sheet glaciers in various geographic regions and began to estimate the contribution of glacial meltwater in those regions to the rise in sea level. Dowdeswell and others (1997) analyzed mass-balance data from non-ice-sheet glaciers in the Arctic, including 40 different ice caps and mountain glaciers; they concluded that these glaciers were adding about 0.13 mm a^{-1} to the rise in eustatic sea level. For the two ice fields and their associated outlet glaciers (Northern and Southern Patagonian Ice fields) of Southern South America, Aniya (1999) concluded that glacial meltwater from these ice fields added an average of 0.038 mm a^{-1} from 1944 to 1945 and 1995 to 1996.

Dyurgerov and Meier (1997b) analyzed available mass-balance data for non-ice-sheet glaciers on Earth for the period 1961 to 1990 and calculated that the non-ice-sheet glaciers contributed an average rate of sea-level rise of 0.25±0.10 mm a^{-1}, even though the sea-level rise was 0.9 mm a^{-1} during those years of high negative mass balances. Dyurgerov and Meier (1997b, p. 392) also noted, "The contribution of [non-ice-sheet] glaciers to sea-level rise has increased greatly since the middle 1980s and even more steeply since the late 1980s, which is in agreement with the rise of global temperature."

Also during the late 1990s, the glaciologists Mark F. Meier and David B. Bahr, and other glaciological colleagues at the Institute for Arctic, [Antarctic], and Alpine Research at the University of Colorado, Boulder, Colo., U.S.A., published a series of seminal papers on the area, volume, and number of non-ice-sheet glaciers on Earth; their findings were based on analysis of databases on glaciers at the WGMS and NSIDC and on statistical projections (scaling methods) (Meier and Bahr, 1996; Bahr, 1997, and Bahr and others, 1997). Two Intergovernmental Panel of Climate Change assessments were published in the 1990s. The Second Assessment Report of the Intergovernmental Panel on Climate Change (IPCC) (1996), as summarized by Meier and Wahr (2002), shows that the low, middle, and high estimates for contributions to rise in sea level during the 20th century were +0.2, +0.4, and +0.7 mm a^{-1} for thermal expansion of the oceans; for non-ice-sheet glaciers, +0.2, +0.35 and +0.5 mm a^{-1}; for the Greenland ice sheet, -0.4, 0, and +0.4 mm a^{-1}; and for the Antarctic ice sheet, -1.4, 0, and +1.4 mm a^{-1}. Therefore, the prevailing scientific judgment at the end of the 20th century was that ocean warming contributed about 50 percent to the observed rise in global sea level and that glacial meltwater from non-ice-sheet glaciers contributed the same percentage. For the Greenland and Antarctic ice sheets, the low and middle estimates were either negative or zero; hence, neither ice sheet was considered to be contributing to the eustatic rise in sea level.

Sea-Level Rise During the Beginning of the 21st Century

Meier and Wahr (2002) also summarized the estimated rates of rise in sea level in the 20th century as reported in the Third Assessment Report of the IPCC published in 2001. The low, middle, and high estimates for sea-level rise from thermal expansion were +0.3, +0.5, and +0.7 mm a^{-1}, and for rise in sea level from glacial meltwater from non-ice-sheet glaciers +0.2, +0.3, and +0.4 mm a^{-1}, nearly the same as those reported in the second IPCC assessment five years earlier. For the Greenland and Antarctic ice sheets, however, the low, middle, and high estimates were 0, +0.05, and +0.1 mm a^{-1}, and -0.2, -0.1 and 0 mm a^{-1}, respectively. However, Meier (2003b) noted that the estimates for the contribution of glacier meltwater and from other sources are less than the rise in sea level that the tide gauges measured, He referred to the discrepancy as an "enigma," suggesting that new technologies, including new satellite sensors, were needed to resolve it. Miller and Douglas (2004) discussed a different discrepancy: that a mass increase of the oceans 2 to 3 times greater than ocean warming can cause (about 0.5 mm a^{-1}) would have been needed in order to account for a rate of sea-level rise during the 20th century of 1.5 to 2.0 mm a^{-1}. As to which geographic area was the source of glacial meltwater from non-ice-sheet glaciers, Meier and Dyurgerov (2002) concluded from their research and that of Arendt and others (2002) that the volume of meltwater from glaciers in Alaska had been underestimated, as had data presented in the Second Assessment Report of the Intergovernmental Panel on Climate Change (IPCC) (2001), and that the Alaskan glaciers were contributing about 50 percent of the total glacial meltwater to sea-level rise. Walsh and others (2005) concurred. Arendt and others (2002) stated that from the mid-1950s to the mid-1990s, the glaciers of Alaska had been contributing about 78 percent more glacial meltwater than the Greenland ice sheet contributed during the same period of time (Szuromi, 2002). Another variable not widely recognized is the volume of surface water stored in reservoirs behind the nearly 30,000 dams built on river systems globally during the 20th century. Chao and others (2008) computed that impounded reservoir water has reduced the rise in sea level by 0.55 mm a^{-1} during the past 50 years (from about 1945 to 2005). If this volume of water were to be "added back in" to the global hydrologic cycle, the rate of sea-level rise during the 80 years since about the late 1920s would be +2.46 mm a^{-1}.

Dyurgerov and Meier (2005, 2006), Dyurgerov (2006), and Bahr and others (2009) continued their research on the contribution of non-ice-sheet glaciers to sea-level rise by adding the contributions from ice caps and mountain glaciers in Greenland that lie outside the Greenland ice sheet and around the Antarctic ice sheet but are independent of either ice sheet. Raper and Braithwaite (2006) lowered by 50 percent the previous estimates of the volume of glacial meltwater from non-ice-sheet glaciers, especially meltwater from ice caps. Meier and others (2005) had previously challenged the validity of the models that Raper and Braithwaite (2005) had stipulated in their previous paper. Meier and others (2007) argued that non-ice-sheet glaciers would be the dominant source of glacial meltwater, with 60 percent of the loss of glacier ice from such glaciers rather than from the Greenland and Antarctic ice sheets; acceleration of glacier melting could result in +0.1 to +0.25 m of additional rise in sea level by the end of the 21st century. Domingues and others (2008) examined the aggregate of the components contributing to eustatic rise in sea level in order to produce a more accurate assessment of the contribution from these variables: on ocean warming and thermal expansion from the upper 700 m and deeper levels, meltwater from the Greenland and Antarctic ice sheets and non-ice-sheet glaciers, and terrestrial storage. Meier and others (2007) reported that the present rate of global sea-level rise is 3.1±0.7 mm a^{-1}; the contribution from ocean warming (steric) is 1.6±0.5 mm a^{-1}; the contribution from loss of non-ice-sheet glaciers and ice sheets is 1.8 mm a^{-1}. Nerem and others (2006) reviewed the state of knowledge with respect to the present changes in sea level.

Summarizing the observed rate of sea-level rise and the sources of the rise that was published in the Fourth Assessment Report of the Intergovernmental Panel on Climate Change (IPCC) (2007a), (table 13), the Intergovernmental Panel on Climate Change (IPCC) (2007b, p. 6–7) wrote:

> Rising sea level is consistent with warming (Figure SPM.1). Global average sea level has risen since 1961 at an average rate of 1.8 [1.3 to 2.3] mm/yr and since 1993 at 3.1 [2.4 to 3.8] mm/yr, with contributions from thermal expansion, melting glaciers and ice caps, and the polar ice sheets. Whether the faster rate for 1993 to 2003 reflects decadal variation or an increase in the longer-term trend is unclear.

Table 13 is from the summary in Bindoff and others (2007), Chapter 5, “Observations: Oceanic Climate Change and Sea Level,” who based some of the information in the table from Lemke and others (2007), Chapter 4, “Observations: Changes in Snow, Ice and Frozen Ground.”

Figure 47 shows globally averaged rise in sea level from 1870 to 2006: the latest data are from satellite altimetry data; the earlier data are from tide-gauge measurements (Church and White, 2006). Figure 48 shows the components contributing to the rise in eustatic sea level from 1993 to 2003, derived from the Fourth Assessment Report of the Intergovernmental Panel on Climate Change (IPCC) (2007a) and Church and others (2007).

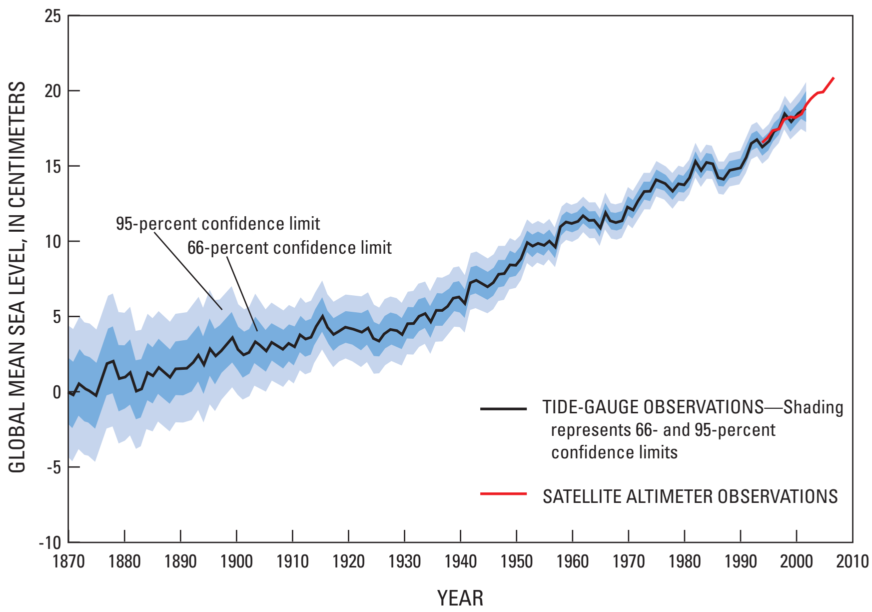


Figure 47.*—Globally averaged rise in sea level from tide-gauge observations (black line) and from satellite altimetry data (from sensors on Topex/Poseidon and Jason-1 satellites) (red line). Modified from Bentley and others (2007, p. 156, fig. 6C.3); original source is Church and White (2006).*

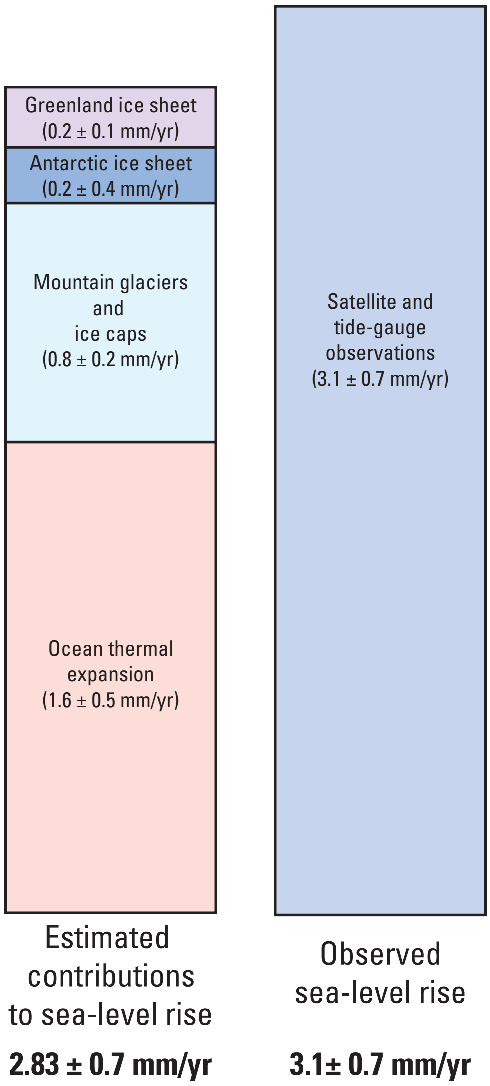


Figure 48.*—Estimated sources of global rise in sea level from 1993 through 2003, estimated from ocean thermal expansion (steric rise) and as meltwater from non-ice-sheet glaciers and from the Greenland and Antarctic ice sheets (2.83±0.7 mm a^{-1}). Satellite data and tide-gauge observations indicate a higher mean rise (3.1±0.7 mm a^{-1}) during the same period. Modified from Bentley and others (2007, p. 157, fig. 6C.4).*

Satellite Remote Sensing of the Greenland and Antarctic Ice Sheets During the 21st Century and Rise in Sea Level

At the beginning of the 21st century, new data about changes in the Greenland and Antarctic ice sheets began to be obtained by sensors on Earth-orbiting satellites. More accurate measurement and modeling of thermal expansion of the oceans also became available (Meehl and others, 2007, table 10.7, p. 820). Analyses of geographic variation in spatial patterns in global rise of sea level by Mitrovica and others (2001) concluded that the Greenland ice sheet had contributed about 0.6 mm a^{-1} of sea-level rise during the 20th century. In 2004, Bamber and Payne (2004) edited a comprehensive book, "Mass Balance of the Cryosphere. Observations and Modelling of Contemporary and Future Changes." In that volume, Bamber and Kwok (2004) reviewed the various Earth-orbiting satellites and the types of sensors that each carried for determining changes in the Earth's non-ice-sheet glaciers and the two ice sheets. The list included the Landsat series of spacecraft (MSS, RBV, TM, ETM+ imaging sensors); SPOT imaging sensors, including stereoimaging; ASTER, with its medium-resolution imaging sensors, including stereoimaging, the primary data source for the GLIMS consortium; the NOAA series of satellites with the Advanced Very High Resolution Radiometer (AVHRR) imaging sensor; synthetic aperture radar (SAR), scatterometer (microwave) and passive microwave sensors on a number of satellites; various satellites with radar altimeters, laser altimeters, such as the Geoscience Laser Altimeter System (GLAS) sensor on ICESat (National Aeronautics and Space Administration, 2002), and MODIS data to measure radiometric temperature of glacier surfaces (Hall and others, 2006, 2008).

Using aircraft and satellite laser altimeter data concerning part of the ice sheet in West Antarctica, Thomas and others (2004) found that glacier-thinning rates were sufficient to add +0.2 mm a^{-1} to sea level. Vaughn (2005) reviewed the use of the ERS-1 radar altimeter measurements of the Antarctic ice sheet and showed that most of West Antarctica had a negative mass balance; most of East Antarctica had a positive mass balance. Vaughn and others (2007) in a subsequent paper reiterated their conclusions with respect to the West Antarctic part of the Antarctic ice sheet. Zwally and others (2005), from an analysis of satellite radar altimetry data of the Greenland ice sheet (10.5 years) and the Antarctic ice sheet (9 years), determined that the Greenland ice sheet was in a slightly positive mass balance (producing a 0.03 mm a^{-1} drop in sea level), but the Antarctic ice sheet had a slightly negative mass balance (producing a 0.08 mm a^{-1} rise in sea level). However, the Arctic Climate Impact Assessment (ACIA) (2005) concluded that the Greenland ice sheet was contributing a +0.13 mm a^{-1} rise in sea level. Alley and others (2005) reviewed the impact of the continuing increase of CO_2 and other greenhouse gases in the Earth's atmosphere and the melting of the Greenland and Antarctic ice sheets. Dowdeswell (2006) noted that the acceleration in flow of several outlet glaciers from the Greenland ice sheet and the increase in areal extent and duration of melting were significant changes; he suggested that the estimates for the contribution of glacial meltwater to the rise in sea level were too low.

From analysis of GRACE data between 2002 and 2005, Velicogna and Wahr (2006b) determined that the decrease in volume of glacier ice of the Antarctic ice sheet was mostly from West Antarctica and that it contributed +0.4±0.2 mm a^{-1} to the rise in sea level. There is no question that the continued acquisition of GRACE data will be critical to developing a continuous and long-term record of changes in mass of both the Greenland and Antarctic ice sheets (Luthcke and others, 2006; Sullivant, 2007; Velicogna, 2009). Cazenave and others (2009) used gravimetric data acquired from the GRACE system of satellites, satellite altimetry, and the Argo satellite to examine sea level during the period 2003 to 2008; they concluded that sea level is rising at a rate of ~2.5 mm a^{-1} and that non-ice-sheet glaciers and accelerated melting from the polar ice sheets have equally contributed to the increase in ocean mass since 2003. Rignot and others (2008) used satellite interferometric SAR data for 85 percent of the coastline of Antarctica from 1992 to 2006 to estimate the total mass flux of ice discharging into the ocean. In East Antarctica, the loss of ice was 4±61 billion metric tons (Gt) a^{-1}; in West Antarctica, ice sheet loss was 132±60 Gt a^{-1} 2006, 59 percent higher than 10 years earlier; on the Antarctic Peninsula the loss was 60±46 Gt a^{-1}, a 140-percent increase. Perkins (2009a) reported on research by Eric Steig and colleagues at the University of Washington (U.S.A.) who analyzed climate data for Antarctica from 1957 to 2006. East Antarctica warmed +0.1°C; in West Antarctica, the average temperature increased about +0.17°C per decade. The British Antarctic Survey (2007) had previously reported a +3°C increase on the Antarctic Peninsula since the 1940s.

From a number of sources, Shepherd and Wingham (2007) reviewed recent contributions from the Antarctic and Greenland ice sheets to the rise in sea level. With reference to the Fourth Assessment Report of the Intergovernmental Panel on Climate Change (IPCC) (2007a), they (Shepherd and Wingham, 2007, p. 1,529) stated,

> It is apparent that the late 20th- and early 21st-century ice sheets at least are dominated by regional behaviors that are not captured in the models on which IPCC predictions have depended, and there is renewed speculation… of accelerated sea-level rise from the ice sheets under a constant rate of climate warming.

Using data from the GLAS sensor on ICESat, Pritchard and others (2009) documented significant dynamic thinning at the margin of the Greenland ice sheet and along the cryospheric coast of Antarctica. Kerr (2006, p. 1,698) reported another similar concern that Michael Oppenheimer of Princeton University voiced as: "The time scale for future loss of most of an ice sheet may not be millennia," as glacier models have suggested, "but centuries." Finally, Steffen and others (2008) summed up the state of knowledge in 2007 about the average rate of sea-level rise to which non-ice-sheet glaciers had contributed since the mid-19th century (about +0.3 to +0.4 mm a^{-1}); in 2007, the mass balance loss was about 400 Gt a^{-1} (about 1.1 mm a^{-1} of sea-level rise). For the Greenland ice sheet, the mass balance loss increased from 100 Gt a^{-1} in the mid-1990s to more than 200 Gt a^{-1} by 2006. Using GRACE data, Van den Broeke and others (2009) quantified a mass loss of about 1,500 Gt during the period 2000 to 2008, or a contribution of 0.46 mm a^{-1} of eustatic sea-level rise, half from runoff and precipitation and half from ice dynamics. Since 2006, however, the increase in rates of summer melt have resulted in a mass loss of 273 Gt a^{-1} or 0.75 mm a^{-1} of sea-level rise. For the Antarctic ice sheet, the mass balance loss was about 80 Gt a^{-1} in the mid-1990s, increasing to 130 Gt a^{-1} in the mid-2000s. Therefore, most of the Earth's non-ice-sheet glaciers and the two ice sheets had a negative mass balance by 2007. About 360 Gt of melted glacier ice will raise global sea level +1 mm.

Contribution to Sea-Level Rise During the 21st Century from the Greenland Ice Sheet and the Antarctic Ice Sheet

The Fourth Assessment Report of the IPCC (2007a) and the Summary of the Intergovernmental Panel on Climate Change (IPCC) (2007b, p. 5) recognize that "losses from the ice sheets of Greenland and Antarctica have *very likely* contributed to sea level rise over 1993 to 2003 (see Table SPM-1)" (table 13) since the Third Assessment Report of the IPCC in 2001. However, later in the same Summary (Intergovernmental Panel on Climate Change [IPCC], 2007b, p. 17), "Contraction of the Greenland Ice Sheet is projected to continue to contribute to sea level rise after 2100….If a negative surface mass balance were sustained for millennia, that would lead to virtually complete elimination of the Greenland Ice Sheet and resulting contribution to sea level rise of about 7 m. [see table 7]…Current global model studies project that the Antarctic Ice Sheet will remain too cold for widespread surface melting and is expected to gain in mass due to increased snowfall." In the Fourth Assessment Report of the IPCC (2007a) and in various summaries of the conclusions reached in the report (Collins and others, 2007; Alley, 2007), there were substantial difficulties in understanding the dynamic processes governing response of the Greenland and Antarctic ice sheets to global climate warming: "Understanding of these processes is limited and there is no consensus on their magnitude" (Intergovernmental Panel on Climate Change [IPCC], 2007b, p. 17; Meehl and others, 2007; Church and others, 2007).

Although many glaciologists ascribed to the consensus that non-ice-sheet glaciers (that is, the estimated 160,000 mountain glaciers and about 70 ice caps (Meier and Bahr, 1996; Meier, 1998a)) and ocean warming (steric increases) were the two primary components of the projected eustatic rise in sea level, other glaciologists began to question that conclusion, one that had its initial basis in Meier's landmark paper in "Science" on "Contribution of Small Glaciers to Global Sea Level" (Meier, 1984). In addition, questions were also being raised by some glaciologists about using "millennia" as the time scale needed for any significant volumetric changes to occur in the Greenland and Antarctic ice sheets (Intergovernmental Panel on Climate Change [IPCC], 2007b, p. 17) and the long-held reference in textbooks to "millennia" (Sugden and John, 1976). Truffer and Fahnestock (2007, p. 1508) declared that it is time to rethink the time scale for changes in ice sheets: "Satellite data show that ice sheets can change much faster than commonly appreciated, with potentially worrying implications for their stability."

Schellnhuber's map of global "tipping points" in climate change (Kemp, 2005) flags the "Instability of Greenland Ice Sheet?" and "Instability of West Antarctic Ice Sheet?" (see also Anderson, 2007). The concept of climate system vulnerabilities and critical thresholds and "tipping points" that would trigger "dangerous climate change" is from "Avoiding Dangerous Climate Change," edited by Schellnhuber and others (2006, p. 1–2). Several factors have come together to focus attention on the possibility of greater contributions to sea-level rise from the Greenland ice sheet (Lowe and others, 2006) and from the Antarctic ice sheet (Rapley, 2006). In the early 1990s, Oerlemans (1993b) addressed possible changes in mass balance of the two ice sheets; Thomas (1986) emphasized the importance of satellite remote sensing in determining changes in the Greenland and Antarctic ice sheets. Paleoclimate evidence addressed the question of ice-sheet instability. Overpeck and others (2006, p. 1747) concluded:

> Sea-level rise from melting of polar ice sheets is one of the largest potential threats of future climate change. Polar warming by the year 2100 may reach levels similar to those of 130,000 to 127,000 years ago that were associated with sea levels several meters above modern levels; both the Greenland Ice Sheet and portions of the Antarctic Ice Sheet may be vulnerable. The record of past ice-sheet melting indicates that the rate of future melting and related sea-level rise could be faster than widely thought.

The National Research Council (2002) addressed the topic of "Abrupt Climate Change: Inevitable Surprises." Other glaciologists updated the identical topic ("Abrupt Climate Change"), more strongly emphasizing changes in glaciers and in rise of sea level (Climate Change Science Program, 2008; Clark and others, 2008a, b; McGheehin and others, 2008; Steffen and others, 2008). Scientists were more publicly—and the popular press was more frequently—raising concerns that changes in the two remaining ice sheets caused by global climate warming could in turn cause a more rapid rise in global sea level (Sabadini, 2002). Ikeda and others (2009, p. 15) identified the most crucial issues in climate science: (1) causes and magnitude of sea-level rise; and (2) decay of glaciers and the Greenland and Antarctic ice sheets. Archer (2009) devoted a chapter to "Sea Level in the Deep Future" in his book, "The Long Thaw: How Humans are Changing the Next 100,000 Years of Earth's Climate." An editorial in "Nature" (Nature, 2008) was entitled "All Eyes North—The Arctic—Particularly Greenland—Needs to Become a Major Focus of Research for Years to Come."

James E. Hansen, one of the first scientists to publicly proclaim that human activities were changing the Earth's climate by annually increasing CO_2 in the Earth's atmosphere (Kerr, 1989), stated unequivocably in a "Scientific American" article (Hansen, 2004, p. 73) "The dominant issue in global warming, in my opinion, is sea-level change and the question of how fast ice sheets can disintegrate." Hansen (2007a) again raised the issue of how rapidly the two ice sheets could melt and pointed out that sea level could rise several meters within a century. Hansen (2007b) also deplored the scientific reticence in clearly addressing the issue of sea-level rise; he called for a panel of scientific leaders to produce a readable report to address "THE" (his emphasis) dominant issue in global climate warming. The new evidence of changes in the Greenland ice sheet (Rignot, Braaten, and others, 2004; Pritchard and others, 2009; Velicogna, 2009) and in the Antarctic ice sheet (Rignot, 1998; Rignot, Cassasa, and others, 2004; Pritchard and others, 2009; Velicogna, 2009) from analysis of data acquired by newly deployed sensors on Earth-orbiting satellites was certainly behind Hansen's concern. Data from the tandem GRACE satellites were especially alarming (Lemonick, 2008; Velicogna, 2009). Other glaciologists also raised their concerns about needing more information about changes in the two ice sheets. For example, Vaughn and Arthern (2007, p. 1,509) noted that "The IPCC report [Intergovernmental Panel on Climate Change (IPCC), 2007a] has highlighted the urgent need to reduce uncertainty over the future of ice sheets in Greenland and Antarctica." Vaughn and Arthern (2007) noted the recent observations (from satellite data) of changes in both ice sheets.

Forecast of Sea-Level Rise During the 21st Century

Warrick (1993) includes a table showing estimates by various scientists of the range in rise of sea level to the year 2100; estimates ranged from 20 cm (Oerlemans, 1989) to 1 m (Thomas, 1986). A graph taken from the Fourth Assessment Report of the Intergovernmental Panel on Climate Change (IPCC) (2007a; Bindoff and others, 2007) (fig. 49) shows sea level for three time periods: 1800 to 1870, for which we have no data but only estimates of the past, ranging from 0.12 m to 0.2 m below present sea level; 1870 to 2007, based on the instrumental record, a rise in sea level from about 0.16 m below present (2007) sea level to about 0.4 m above the 2007 level. The projected rise for the third period, from 2007 to 2100, is relatively linear, similar to that of the late 20th century, about +0.4 m to +0.5 m.

By the late 1990s, sea level was rising at a rate of 3 to 4 mm a^{-1}. In their study, "The Probability of Sea Level Rise," Titus and Narayanan (1995, p. 111) write, "Global warming is most likely to raise sea level… 34 cm by the year 2100. There is also a 10 percent chance that climate change will contribute… 65 cm by 2100…. There is a 1 percent chance that global warming will raise sea level 1 meter in the next 100 years…" Jacobsen (1988, p. 30), reviewing the

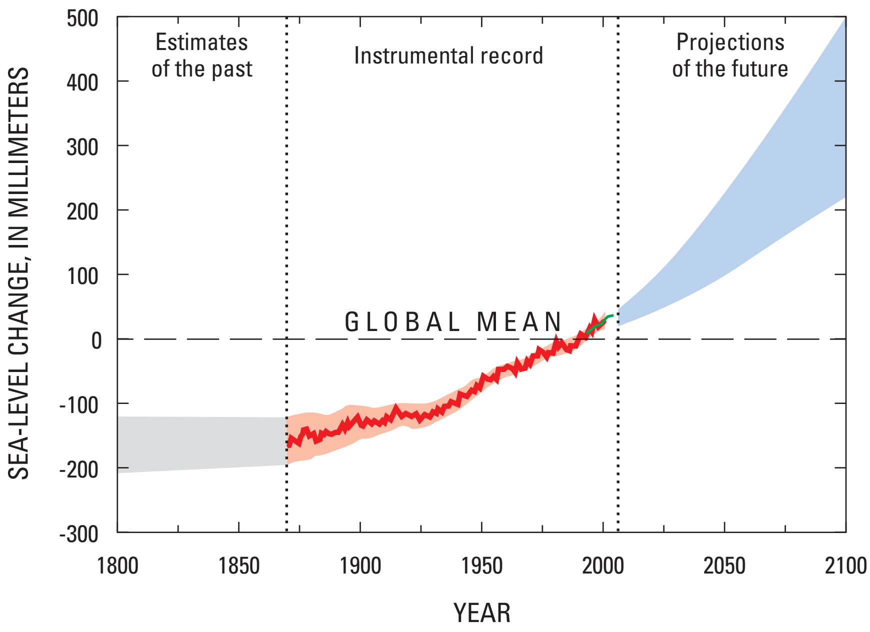


***Figure 49.**—Change in global sea level during three time periods: 1800 to 1870 (estimated), 1870 to 2007 (from the instrumental record), and 2007 to 2100 (projections into the future). Modified from Bindoff and others (2007, p. 409, FAQ 5.1, fig. 1).*

impact of a rise in global mean temperature of 1.5° to 4.5°C as early as 2030, wrote, "If correct, the predicted temperature changes would precipitate a rise in sea level of 1.4 to 2.2 m by the end of the next [21st] century."

Russell (2009) discussed the impact of rising sea level on the Republics of the Maldives and Kiribati, providing a graphic that presented three scenarios for rise in sea level for the year 2100. The lowest range was the rise projected by the Fourth Assessment Report of the Intergovernmental Panel on Climate Change (IPCC) (2007a): +0.2 to +0.6 m, from ocean warming (steric) and from non-ice-sheet glaciers. The highest range was from Pfeffer and others (2008): +0.8 to +2 m, from meltwater from non-ice-sheet glaciers and the two ice sheets. The middle range was from Stefan Rahmstorf of the Potsdam Institute for Climate Impact Research (Anonthaswamy, 2009): +0.5 to +1.4 m from ocean warming and glacial meltwater. The analysis by Siddall and others (2009) of past changes in sea level forecast a rise of 7 to 82 cm in global sea level by 2100. For Ananthaswamy (2009), sea level by 2100 will definitely be higher than that of today. Whether the projected rise in sea level is a fraction of a meter or several meters, the projections agree that the rise in sea level depends on the response of the Greenland and Antarctic ice sheets to global climate warming. Will the actual scenario be "business as usual," similar to the conclusion of the Fourth Assessment Report of the Intergovernmental Panel on Climate Change (IPCC) (2007a), perhaps a maximum of 0.6 m? Or will the response be nonlinear? Satellite remote sensing of areal, surface, and volumetric changes in the Greenland and Antarctic ice sheets, which combine observations with modeling, is the only feasible way of monitoring changes of these two ice sheets and the magnitude of their contribution of glacial meltwater to the rise in eustatic sea level and its regional variation (Mitrovica and others, 2009). Low-lying coastal regions, the deltas of which are heavily populated, and low-lying islands, such as atolls, must be given advance warning.

Such warnings must be based on sound science, so that sound planning will be in place to deal with the impact of inundation of land by sea water. The expected severe impacts include inundation and uprooting of fixed facilities (Jakobson, 1989), and displacement of large human populations (Small and Nicholls, 2003; McGranahan and others, 2007; New Scientist, 2009; Syvitski and others, 2009).

Consequences of Rise in Sea Level

The consequence of the more or less stabilized condition of sea level about 7,000 years ago is that human settlements and populations became concentrated at favorable locations along coastlines, harbors, and mouths of major rivers (and associated deltas). Contributing factors were: the land devoted to agriculture could be expanded; there was access to fisheries; and transporting people and goods by ship was relatively easy and cost effective. This pattern of densely settled low-elevation regions along coasts, including heavily populated deltas, persists. McGranahan and others (2007) pointed out that coastal zones less than 10 m above sea level, representing 2 percent of the Earth's land area, are home to 10 percent of the Earth's population, and are vulnerable to disasters from climate change. About 50 percent of the Earth's global population now lives within 60 km of one of the Earth's oceans (McMichael and others, 1996). Small and Nicholls (2003), using new data on global population density, concluded that 1.2 billion human beings live within 100 km of some coastline and at elevations of 100 m or less; these population densities are nearly three times higher than the average density globally.

As one of the many consequences of human-induced climate change (Douglas and others, 2001; Schlesinger and others, 2007; Diaz and Murname, 2008), the rise in eustatic sea level will have its greatest impact on low-lying coasts and islands (Vellinga and Leatherman, 1989; National Research Council, 1990; Poore and others, 2000; Nichols and others, 2007), especially on those coasts and islands where sediments are easily erodible and from which bedrock is absent. Inundation of such regions and subsequent displacement of populations will occur (National Research Council, 1987; Rowley and others, 2007). Accelerated coastal erosion will also occur (Milliman, 1989; Leatherman and others, 2000). Therefore, global-scale monitoring of sea level from land-based sensors (tide gauges), airborne sensors (geodetic airborne laser altimetry), and satellite sensors will be required in order to prepare for these eventualities (Leatherman and Kershaw, 2002; Leatherman and others, 2003). Woodworth and others (1992) discussed the determination and effects of rise in sea level. Twenty-five years ago, Barth and Titus (1984) had already recognized the challenges to society that the rise in sea level presents.

There is also a need to determine the full range of hazards associated with a rise in eustatic sea level (Marbaix and Nicholls, 2007). The subsidence of deltas in Egypt and Bangladesh (Milliman and others, 1989) and of the Mississippi River delta in the United States is a representative example, because the eustatic sea-level rise in these regions is augmented by an increase in "local" sea level. Syvitski and others (2009) assess the vulnerability of 33 deltas worldwide to flooding and rise in sea level. They concluded that, during the 10 years from 1998 to 2008, 85 percent of these deltas underwent severe flooding, submerging 260,000 km^2 of land for a limited time. These deltas could experience a 50-percent increase if severe flooding during the

21st century meets projected estimates of rise in sea-level and if impoundment of water in dams and the subsequent diversion of sediments continues. Salt-water intrusion in coastal aquifers (Meier, 1990b), accelerated erosion of the coast, and greater vulnerability of human populations to storm surges from typhoons (Bangladesh) and hurricanes (Louisiana, U.S.A.) are magnified in regions of subsiding deltas and low-lying coastal regions. A report by the Climate Change Science Program (2009), "Coastal Sensitivity to Sea-Level Rise: A Focus on the Mid-Atlantic Region," addresses these related issues. Bird (1986?) addressed the impact of sea-level rise on the coasts of Australia, Africa, and Asia. Giese and Aubrey (1987) "estimated" the economic loss to coastal communities from a "relative" rise in sea level in Massachusetts, U.S.A., beginning in 1980 and persisting through 2025. The loss in taxes from ocean-front property, from impacted private residences, from roads and other public works, and from commerce (tourism revenue) was estimated based on inundation of 1,200 to 4,000 ha, worth between $3 billion and $10 billion.

GLACIER MASS CHANGES AND THEIR EFFECT ON THE EARTH SYSTEM[10]

By MARK B. DYURGEROV[11] *and* MARK F. MEIER[2]

Abstract

Glaciers are indicators of climate change; they also have significant impacts on processes of global importance, such as rise in sea level, hydrology of mountain-fed rivers, freshwater balance of oceans, and even the shape and rotation of the Earth. In this section, we discuss the effects of glaciers—all perennial ice masses other than the Greenland and Antarctic ice sheets. Observational results on glacier mass-balance collected since the mid-20th century in many mountain and subpolar regions on Earth present clear evidence that the volumes of most glaciers are decreasing, with substantially increased losses since the mid-1970s and even more rapid loss since the end of the 1980s. Our new estimates are based on a total area of glaciers of 785,000 km^2, somewhat larger than earlier estimates, because of improved information on isolated glaciers and ice caps around the periphery of the large ice sheets.

Glacier wastage (melting) causes rise in sea level; we now estimate that this contribution averages 0.51 mm a^{-1} for the period from 1961 to 2003 but that it rose to 0.93 mm a^{-1} in the decade from 1994 to 2003. Together with recent calculations of loss of ice-sheet volumes, this addition of freshwater to the oceans now accounts for a rise in sea level of about 1 mm a^{-1} that may affect ocean circulation and ocean ecosystems. The magnitude and rate of glacier wastage are critical to the ability to understand and to project changes in sea level. This contribution from glaciers is likely to increase, not decrease, in the future.

Acceleration of glacier wastage also affects other global processes such as spatial and temporal changes in the Earth's gravitational field as well as Earth's oblateness and its rotation rate. A recent (1998) increase in oblateness has been attributed to recent acceleration of glacier wastage. This wastage also results in regional uplift of deglacierized areas.

Glacier mass-balance data (both annual and seasonal) can be used to infer climatic variables such as precipitation and temperature, and the spatial distribution of these mass-balance data can assist in the analysis and modeling of climate change. This potential for inference is especially important in high-mountain and high-latitude areas, where precipitation data are few and biased because meteorological stations are insufficiently distributed geographically. Large differences between snow accumulation (winter balance) and observed precipitation suggest that one should use caution when considering adjustments to the data in any modeling of interactions between glaciers and climate and in projecting future changes in sea level. The increase in air temperature is the major forcing of changes in glaciers. Glacier response to recent climate

[10] Editors' note: Mark B. Dyurgerov died on 5 September 2009. His co-author of this section, Mark F. Meier, and Tatyana Kostyashkina completed work on an important "unfinished" paper; that paper was published in Moscow in October 2010: Dyurgerov, M.B., 2010, Reanalysis of glacier changes: From the IGY to the IPY, 1960–2008: Moscow, Data of Glaciological Studies, Publication 108, 116 p. Many of the graphs used to illustrate this Section are updated in Publication 108. Mark F. Meier died on 25 November 2012.

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warming shows a steepening in gradient of mass balance with altitude caused by increasing ice ablation below the equilibrium line altitude (ELA) and, to a lesser extent, increasing snow accumulation above that altitude. Observational results also show increases of glacier mass turnover and mass-balance sensitivity to air temperature; these changes are *not* predicted by existing models of relationships between glaciers and climate. Sensitivity and turnover have also shown a remarkable decrease in variability starting at the end of the 1980s.

Global acceleration of losses in glacier volume has affected the freshwater cycle at many scales, from global to local. Precipitation over glacier areas averages about 36 percent higher than that over nonglacier land. The glacier contribution to the freshwater inflow to the Arctic Ocean has been increasing; this increase will continue as a result of global warming and will affect many aspects of the Arctic climate system. The glacier input to the Arctic Basin is unique in that much of it flows directly to the ocean rather than to the several major rivers that hydrologists regularly measure. Increasing summer runoff to large Asian rivers and high-elevation glacierized watersheds in both Americas is important for agriculture and for other human needs, but this release of water from storage as ice may diminish in the future as the relatively small high-mountain glaciers shrink in volume and eventually disappear within decades.

Introduction

Glacier variations have been of interest for hundreds of years because they can be sensitive indicators of changes in climate (for example, Forel, 1894; Haeberli, 2004). We now know that glacier variations may also affect global rise in sea level, the hydrology of mountain-fed rivers, the freshwater balance of the oceans, the frequency and intensity of natural disasters, and even the shape and rotation of the Earth. In recent years, the rate of loss of glacier ice to the oceans has accelerated, and this trend is expected to continue as a result of the rise of so-called greenhouse (radiatively active) gases in the atmosphere. At the same time, satellites and other new technology have made it increasingly feasible for glaciologists to measure subtle changes in the Earth System. Therefore, it is appropriate to examine the role of glaciers in global processes.

This section discusses mainly the effect of only the [non-ice-sheet] glaciers of the world—that is, all of the perennial ice masses other than the Greenland and the Antarctic ice sheets. The discussion therefore includes mountain glaciers, ice fields, ice caps, and all other kinds of glaciers—temperate or cold or polythermal—but it excludes the termini of outlet glaciers from the two ice sheets except in some special cases where it is not possible to determine the origin of glacial meltwater.

We consider the area of these glaciers to be about 785,000 km^2. This most recent estimate (Dyurgerov, 2005) is larger than older estimates such as those by Meier and Bahr (1996) or Haeberli and others (1999) not because the area that glaciers cover is expanding but because more reliable methods have been found for estimating the glaciers that lie around the periphery of the great ice sheets (Shumskiy, 1969; Weidick and Morris, 1998) and in other regions. An earlier study (Meier and Bahr, 1996, fig. 4) used incomplete glacier inventories and inadequate scaling analysis (Bahr and Meier, 2000) as its basis for presenting an estimated distribution of glacier sizes. These size distributions can also be used with a volume/area scaling algorithm (Bahr and others, 1997; Macheret and others, 1999) to estimate glacier-volume distributions and thus their likely areal changes with further melting. Our estimated area of

785,000 km^2 for the glaciers we discuss here therefore translates to a volume of about 250,000 km^3, equivalent to a rise in sea level of about 0.7 m. It is virtually impossible to state the number of existing glaciers because the definition of a single glacier is very subjective; the number certainly exceeds 200,000 and may even be increasing as global glacier wastage causes tributary ice masses to split off from a trunk glacier.

The World Glacier Inventory Program (WGI) was an ambitious attempt to measure and classify all of the perennial ice masses [non-ice-sheet glaciers] of the world (Haeberli and others, 1989). However, an inventory of glaciers and ice caps could be completed only in certain areas, such as Europe, and many of the more important regions could not be measured. Therefore, the WGI is inherently biased. That inventory can, however, be extended by techniques that more closely approximate the actual number of glaciers, such as those suggested by the Global Land Ice Measurements from Space (GLIMS) consortium (Kieffer and others, 2000).

Some recent estimates of the total area and volume of glaciers and ice caps are given in table 14. Almost half of this estimated volume occurs around the periphery of the Antarctic and Greenland ice sheets, which previous estimates of area and volume did not include (Meier and others, 2005). Table 14 tabulates these area estimates by regions and by sources (Dyurgerov, 2005, appendix 1).

Differences between the volumes that different authors estimate are large, and they depend on the assumptions used and the method of calculation. Even larger differences between the various estimates shown in table 14 are due to whether an author included the glaciers around the peripheries of the two ice sheets. Some authors assume that these peripheral glaciers will be analyzed as parts of the ice sheets, but these small glaciers are at lower altitudes, in more maritime climates, and are too small to be included realistically in the coarse grids that glaciologists use for modeling the big ice sheets. Vaughn (2006, p. 147) points out that "[the glaciers of the Antarctica Peninsula have] greater

Table 14.—*Total areas and volumes of glaciers around Greenland, Antarctica, and elsewhere as reported in selected recent authoritative sources*

[Areas and volumes are estimated based on possible errors in measurements, the uncertainty of the relation between area and volume, and the scatter in the power-law relation of area and volume. 10^3 km^2, thousands of square kilometers; 10^3 km^3, thousands of cubic kilometers; NA, not applicable]

Source	Area (10^3 km^2)		Volume (10^3 km^3)		Total	
	Global glacierized area, excluding Antarctica and Greenland	Glacierized area in Antarctica and Greenland[1]	Global glacial ice volume, excluding Antarctica and Greenland	Antarctica and Greenland[1]	Area (10^3 km^2)	Volume (10^3 km^3)
Meier and Bahr (1996)	540	140	NA	NA	680	180
Raper and Braithwaite (2006)	522	NA	87	NA	522	87
Ohmura (2004)	521	NA	51	NA	521	51
Dyurgerov (2005)	540	245	NA	NA	785	NA
From Dyurgerov and Meier section of Part A–2 (Glaciers) of this chapter	540 ± 30	245 ± 100	[2]133 ± 20	[2]125 ± 60	785 ± 100	260 ± 65

[1]Excluding the Antarctic and Greenland ice sheets.

[2]Calculated by separating the distribution of glacier area and size for all of Antarctica and Greenland, which Meier and Bahr (1996) calculated, and increasing the total by using the newer (larger) area but the same mean thickness that Meier and Bahr (1996) determined for polar and subpolar glaciers, thus increasing the total area.

similarity to subpolar glacier systems (such as coastal Greenland, Svalbard, Patagonia, and Alaska), which are known to be more sensitive to atmospheric warming, than to the cold ice sheets covering the rest of the Antarctic continent … .”

Much attention has traditionally been paid to variations in the length of glaciers—their advance and retreat (for example, Forel, 1895; Oerlemans, 1994, 2005; Haeberli , 1995). Although useful for demonstrating the changes that have been happening, these data give only crude measures of glaciers' overall changes unless detailed knowledge is available through modeling of their dynamic response to climate change. Determining the “length response time,” the parameter generally used for studies of a given glacier, requires knowledge not only of glacier geometry but also about its mass balance. The parameter is further constrained to relatively small perturbations about a mean length (Jóhannesson and others, 1989; Harrison and others, 2001; Klok and Oerlemans, 2003). This knowledge is available for only a few glaciers; therefore, these histories of advances and retreats are generally of limited use for large-scale syntheses of, for instance, year-to-year climate change or other important issues such as sea-level rise. By themselves, measures of change in glacier area are similarly limited in direct application, but they are extremely important—indeed necessary—for more rigorous analyses when combined with studies of changes in thickness or of mass-balance. The following discussion describes the annual (or net) balance of glaciers, the direct measure of the exchange of ice mass, through atmospheric or hydrologic processes, between land and ocean.

Mass Balance

Of the several ways that glacier mass balance can be measured, two are most common. The first repeatedly measures the elevation of the ice surface; these data on changes in thickness, combined with glacier area and an appropriate density of snow and ice, yield changes in mass. The second takes mass-balance observations on the surface and then sums the measurements over the glacier and during a year to attain the glacier-wide mass changes, the net or annual balance. Changes in mass are usually reported annually, but they can be determined seasonally in some cases; they may also be available only as long-term (multiyear) values. The first method (elevation of the surface) has recently become especially productive because laser altimeters that can be flown in aircraft with global positioning systems (GPS) for spatial orientation have been developed (for example, Echelmeyer and others, 1996; Abdalati and others, 2001; Arendt and others, 2002). Multiple GPS profiles taken from snowmobile traverses can also determine accurate elevations of glacier surfaces. Oddur Sigurðsson (Icelandic Meteorological Office) and others (Shuman, Hall, and others, 2006; Shuman, Sigurðsson, and others, 2009) used snowmobile traverses to measure the surface elevation and area (~146 km^2) of the Drangajökull ice cap, northwestern Iceland, in April 2005.

Most mass-balance data from the world's glaciers have been obtained by traditional surface measurements, which can be used to measure important details such as changes in snow and ice density; however, these traditional measurements are very labor intensive and not without hazard to the field glaciologist. Extensive literature exists on mass-balance methods (for example, Østrem and Brugmann, 1991), terminology (Mayo and others, 1972), international programs (for example, Haeberli, 1995, 2004), and compilations of results by the World Glacier Monitoring Service (WGMS) in Zürich and the National Snow and Ice Data Center (NSIDC) in Boulder, Colo. Datasets by J.G. Cogley (2002 [*http://www.trentu.ca/geography/glaciology.htm*]), by Dyurgerov (2002 [*http://instaar.colorado.edu/other/occ_papers.html*]), by Dyurgerov (2005 [*http://nsidc.org/data/g10002.html*]), and in a major reference book on the subject (Bamber and Payne, 2004) also are noteworthy. In addition, numerous attempts have been made to model mass-balance time series on the basis of climate data (for example, Oerlemans, 1993a), but the present discussion emphasizes observational data in order to avoid circular reasoning in studies on the relation of glacier changes to climate. The locations of the glaciers used in this compilation are shown in figure 50.

Global Compilation of Mass Balances

The area of small [that is, non-ice-sheet] glaciers that we consider (785,000 km^2) includes areas of individual ice caps in West Antarctica that have no direct connection with the ice sheet; these were not included in previous evaluations. There is evidence that these glaciers may have a negative mass balance and may now be contributing meltwater to sea-level rise (Morris, 1999; Schneider, 1999; Morris and Mulvaney, 2004; Skvarca and others, 2004; Cook and others, 2005; Rignot and others, 2005). Dyurgerov (2005, appendix 1) and table 14 present our recent estimates of regional and global areas. In order to assess the global effects of glacier wastage, changes in the volume of these glaciers need to be compiled for large regions. These data are needed for the study of regional climates and regional water (hydrologic) cycles in connection to climate change, their contributions to sea-level

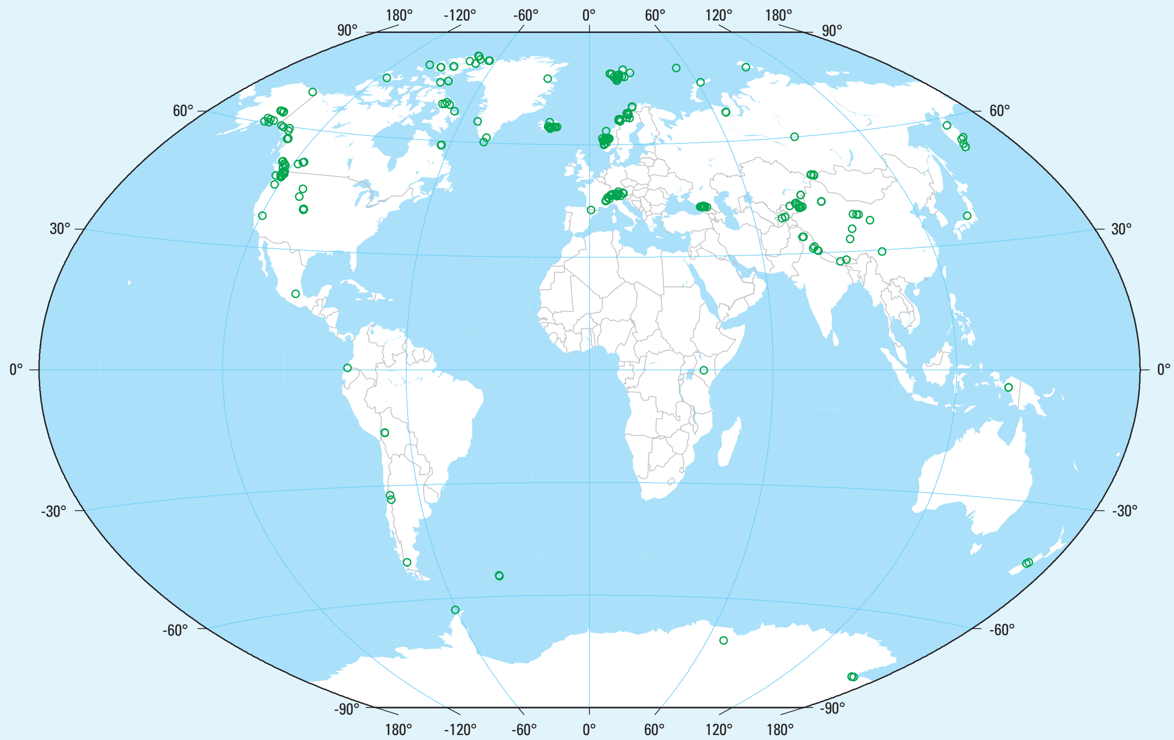


Figure 50.—*Locations of glaciers for which glaciologists have produced mass-balance records.*

change, the impacts of glaciers on the gravitational field/geoid, and other large-scale purposes. Difficulties exist for averaging the data because they are unevenly distributed geographically (for some regions there is only one, or no, time series of change in glacier volume), and unresolved problems persist with spatial extrapolation of glacier data. As is well known, glacier mass-balance data are extremely variable (for example, fig. 51) with respect to many local, regional, and global parameters: longitude, latitude, elevation, aspect ratio, snow/ice temperature, and distance from sources of moisture (Dyurgerov, 2002).

Our recent analysis combined individual time series for changes in glacier volume into larger, climatically homogeneous regions, placing the data into three samples or systems:

- 49 mountain and subpolar systems where sufficient observational data on mass-balance are available from individual glaciers, (Dyurgerov, 2005, appendixes 2 and 4);

- 13 larger scale regional systems that are similar geographically and (or) in terms of climate (Dyurgerov, 2005, appendix 5);

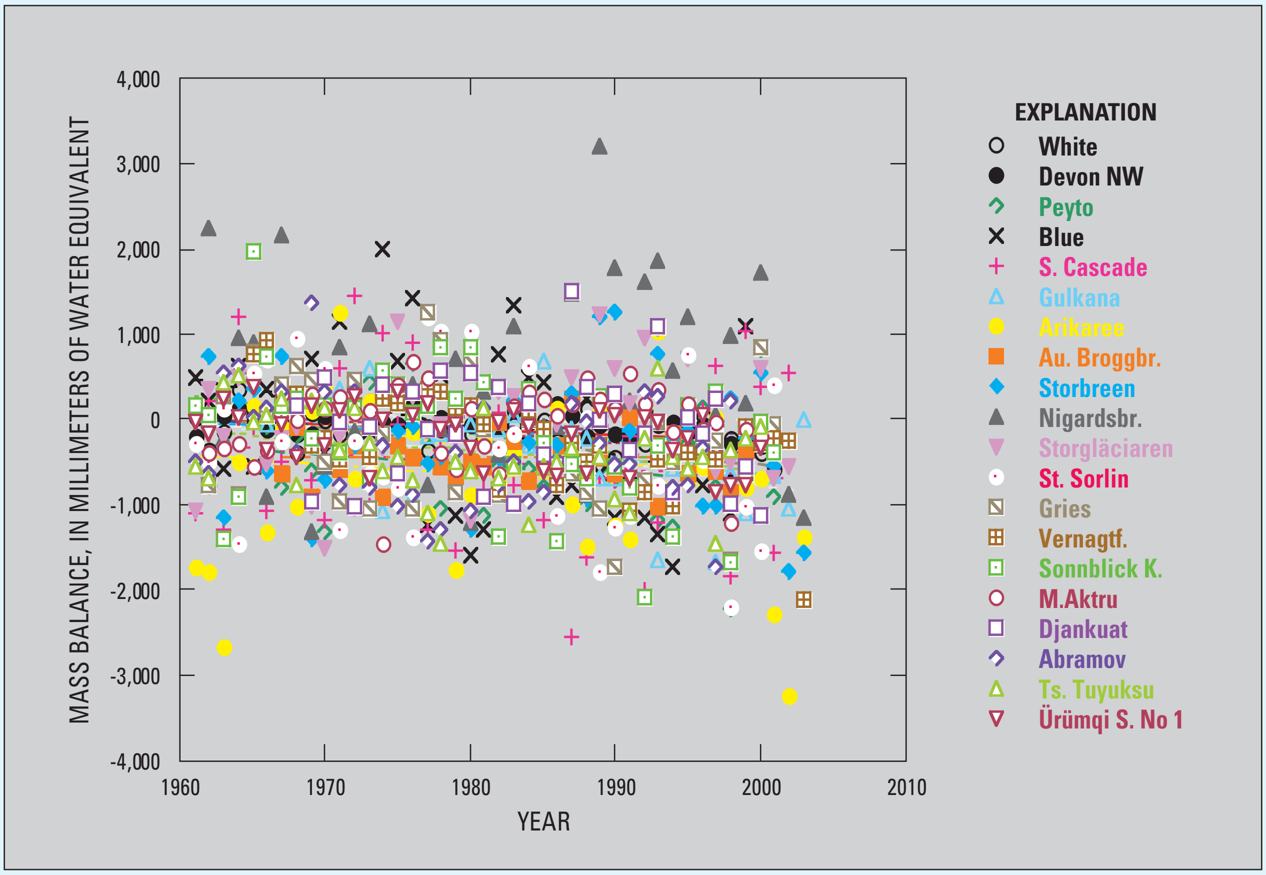


Figure 51.—*Scatterplot showing the significant variability in annual mass balances of 18 selected glaciers with lengthy observational records: White Glacier, Canada; Devon NW (ice cap), Canada; Peyto Glacier, Canada; Blue Glacier, Washington; South Cascade Glacier, Washington; Gulkana Glacier, Alaska; Austre Brøggerbreen, Svalbard (Norway); Storbreen, Norway; Nigardsbreen, Norway; Storglaciären, Sweden; Griesgletscher, Switzerland; Vernagtferner, Austria; Sonnbliekgletscher, Austria; Maliy Aktru Glacier, Russia; Djankuat Glacier, Russia; Abramov Glacier, Kyrgyzstan; Ts. Tuyuksu Glacier, Kazakhstan; and Ürümqi S. No. 1 Glacier, China.*

- 7 globally composite systems, in order to estimate global glacier changes in glacier volume and their contributions to the planetary water cycle (global hydrologic cycle) and to sea-level change (Dyurgerov, 2005, appendix 6).

To calculate changes in glacier volume by systems and regions, we applied a previously introduced scheme (Dyurgerov and Meier, 1997a, b; 2000; 2004) that includes weighting specific mass-balance values by surface area because the sample of observed glaciers is biased toward small (non-ice-sheet) glaciers in many areas of the Earth. We avoided modeling mass balances by using data from meteorological observations because precipitation data are meager for most subpolar and high-mountain regions and, thus, are of limited use for independent analyses of the impact of climate on mass balance.

The main disadvantage of using observational time series, on the other hand, is the necessity for data extrapolation from individual sites to larger areas. This deficiency still exists because no completely reliable approach to extrapolating mass-balance data has yet been found.

Mass-Balance Results

New compilations of time series for mass balance for selected glacier systems are presented in figure 52, for large regions in figure 53, and for the world in figure 54*A*. The details of how these time series were compiled are given in Dyurgerov (2005), but we mention here several interesting aspects of these sequences:

- First, the general trends in the change in volume and their variability are close to those previously calculated and published (Dyurgerov and Meier, 1997a, b; Church and Gregory, 2001; Dyurgerov, 2002).

- Second, very pronounced spikes in the globally averaged annual mass-balance time series are found in connection with the largest explosive volcanic eruptions, in particular Mount Agung, Bali, Indonesia, in 1963; Mount St. Helens, Washington, United States, in 1980; El Chichón, México, in 1982; and Mount Pinatubo, Philippines, in 1991 (fig. 54*B*), with cooling and positive mass balance found for the following 1 to 3 years, regionally and globally (Abdalati and Steffen, 1997; Dyurgerov and Meier, 2000).

- Third, the markedly negative mass balances and acceleration of losses of glacier volume in the late 1980s and 1990s correspond to the unusually high mean temperatures during these years.

- Fourth, the acceleration of change in glacier volume presented here is consistent with other evidence of warming in the Earth System, including reduction of the area and thickness of sea ice (Laxon and others, 2003; see also Part A-4-I on “Sea Ice,” by Parkinson and Cavalieri, in this chapter, p. A345–A380 and p. A489–A496), decreasing areal extent of snow cover by about 0.2 percent a^{-1} in the Northern Hemisphere (Armstrong and Brodzik, 2001; see also Part A–3 on “Global Snow Cover,” by Hall and Robinson, in this chapter, p. A313–A344), increasing temperature and thawing in permafrost (Arctic Climate Impact Assessment [ACIA], 2004, 2005; see also Part A–5 on “Permafrost and Periglacial Environments,” by Heginbottom, and others, in this chapter, p. A425–A496), acceleration in the movement and disintegration of outlet gla-

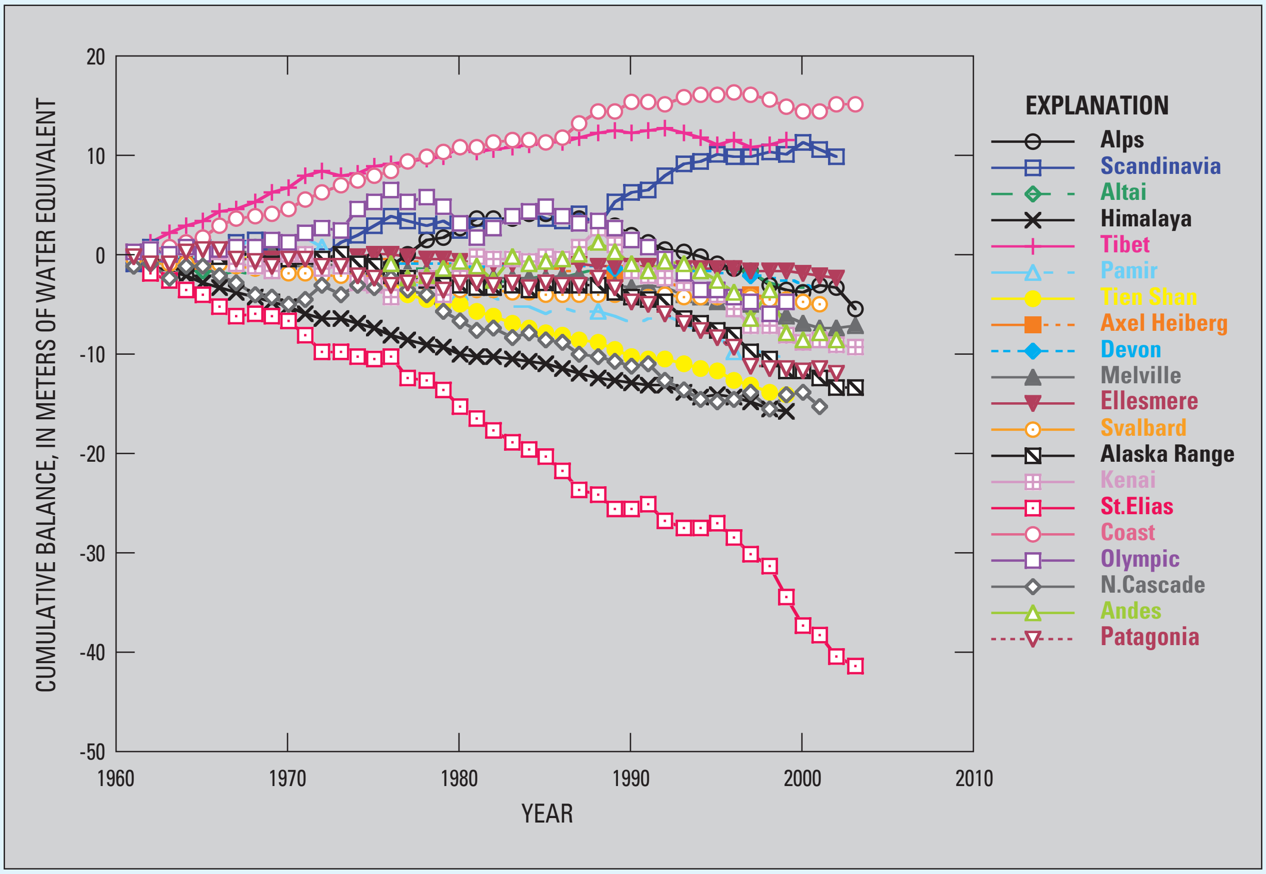
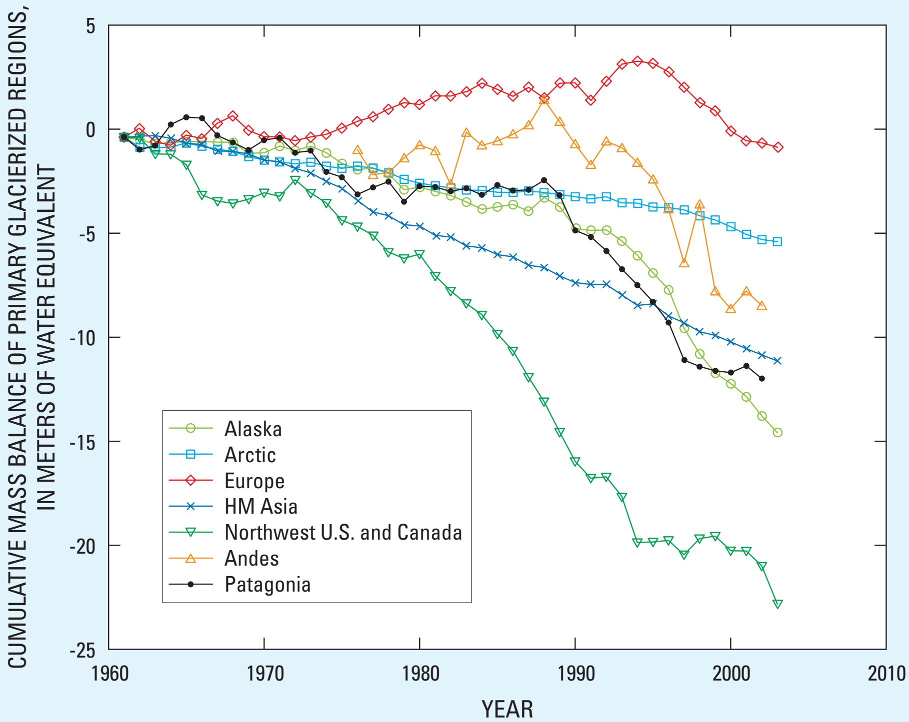


Figure 52.*—Cumulative mass balances of selected glacier systems compiled from individual time series showing differing changes over time until the beginning of the 21st century.*

Figure 53.*—Cumulative mass balances calculated for large glacierized regions. For these calculations, we used the time series of mass balance for all glaciers—more than 300 from time to time and from 30 to 100 with multiyear records (see http://www.nsidc.org). We weighted the annual mass-balance data for individual glaciers by their surface area and then by the aggregate surface area of 49 primary glacier systems (20 of them are shown in fig. 52). By the end of the 1980s, and more clearly during the 1990s, these cumulative curves for large glacierized regions show a significant shift toward accelerated loss of mass.*



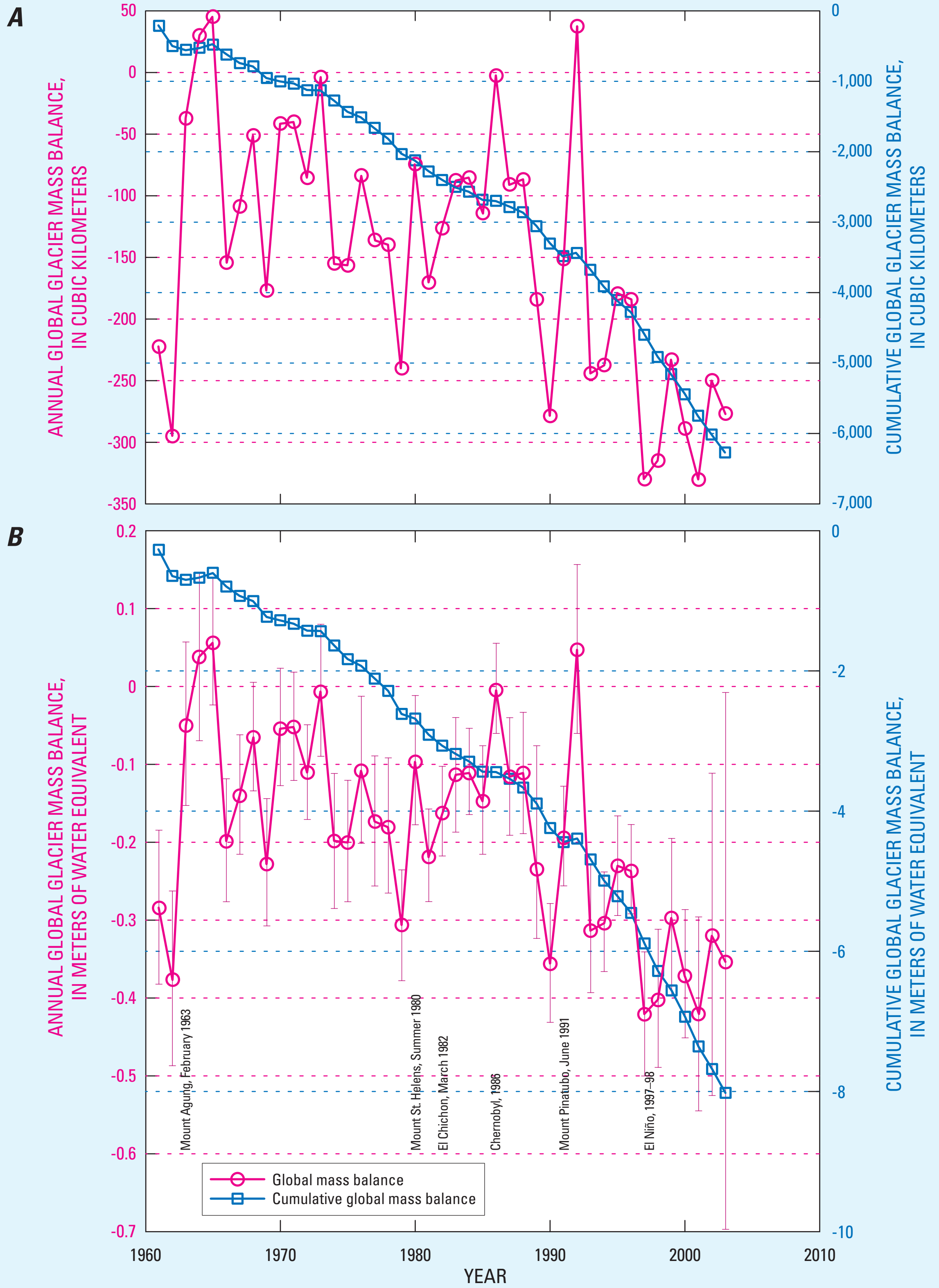


Figure 54.—**A**, *Annual variability in global mass balance of glaciers and cumulative mass-balance values globally.* **B**, *Change in volume and variability computed for the worldwide system of mountain glaciers and subpolar ice caps, which has an aggregate area of 785,000 km². The results of direct mass-balance observations on 300 glaciers worldwide are averaged by area of individual glaciers—49 primary systems, 13 larger regions, 7 continental-size regions, and globally—to construct the single global curve. Vertical bars on B are estimated standard errors.*

ciers and ice shelves in Greenland and Antarctica (Scambos and others, 2000; Zwally and others, 2002; Rau and others, 2004; Thomas and others, 2004; Ferrigno and others, 2006, 2008, 2009), accelerated melting of the Greenland ice sheet (Hall and others, 2004; Steffen and others, 2004) and ice caps and mountain glaciers in Iceland (Cameron, 2005), and rapid disintegration of Alpine glaciers (Paul and others, 2004).

One reason for our larger values of mass loss of glaciers is that we use a somewhat larger total glacier area. Another is that we incorporate data that were new or not previously available, including the results of new measurements in the most recent years, showing more negative mass balances; in particular, (1) new results for changes in mass balance and in volume from the Northern and Southern Patagonian Ice Fields (Rignot and others, 2003), (2) updated mass-balance results for Alaskan glaciers (Arendt and others, 2002), (3) recalculation of mass balance of individual ice caps around the Greenland ice sheet (Weidick and Morris, 1998), and (4) new mass-balance data for glaciers in South America that have not been available before (Casassa and others, 2002).

Impacts of Global Wastage on Sea Level

The trend of rising sea levels is one of the most troublesome and geographically far-reaching aspects of global-environmental change. Societal and economic impacts of eustatic (global) rise in sea level are already evident, and the consequences of continued rise—perhaps even of accelerating rise—are substantial (Douglas and others, 2001). Beach erosion and shoreline retreat affect valuable real estate (Giese and Aubrey, 1987) and the livelihood of waterfront communities. Retreating shoreline may reduce some coastal wetlands or even eliminate coastal-wetland ecosystems if the rise is sufficiently rapid. Saltwater incursion into coastal aquifers and the advance of the saltwater wedge in estuaries may be locally harmful. More than 100 million people live within 1 m of mean sea level (Douglas and Peltier, 2002), and the problem is especially urgent for the inhabitants of low-lying small islands.

Changes in sea level are caused by warming and freshening of ocean water, changes in storage of surface and ground water, and the loss of mass in glacier ice, among other processes (table 12). The meltwater contribution from glaciers has been recognized and studied for many years (for example, Thorarinsson, 1940; Meier, 1984; Church and Gregory, 2001).

Our new time series for changes in glacier volume has been expressed in terms of sea level (fig. 55) and shows that the glacier contribution from 1961 to 2003 is somewhat larger than that estimated in our previous calculations (Dyurgerov and Meier, 1997b; Dyurgerov, 2001, 2002) and those presented in Church and Gregory (2001). This contribution of 0.49 mm a^{-1} is a significant fraction of the 1.5±0.5 mm a^{-1} contribution that is listed as the total 20th century rise in sea level in the Third Assessment Report of the Intergovernmental Panel on Climate Change (IPCC) (Church and Gregory, 2001). The rate of that rise raises the question of how adding that volume of frozen water (glacier ice) from land to ocean (as glacier melt) affected the role of warming and thermal expansion (thermosteric rise) or freshening (hyalosteric rise) of ocean water (Antonov and others, 2002), since these latter two processes together normally produce a eustatic (global) rise in sea level. Satellite geodetic observations and analyses during the early 2000s suggested

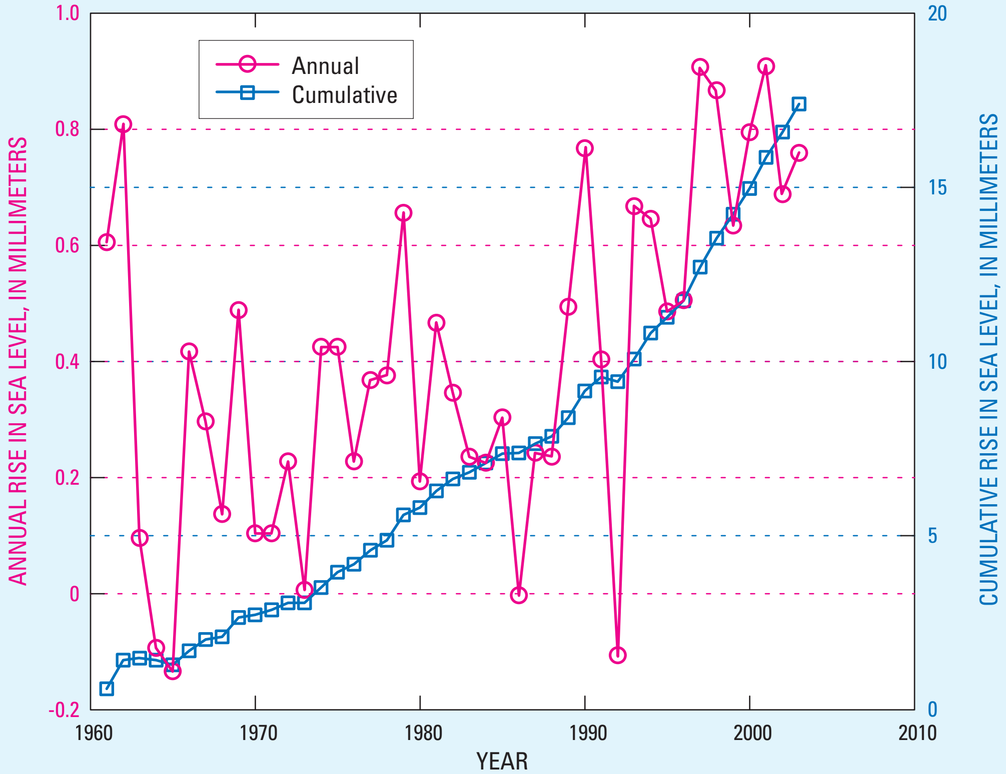


Figure 55.—*Glacier contribution to rise in sea level from mountain glaciers and subpolar ice caps, which have an aggregate area of 785,000 km^2. Observational data from figure 54 have been used to express change in glacier volume in terms of contribution to rise in sea level by dividing the change in glacier volume (mass balance), in cubic kilometers of water, by 362 x 10^6 km^2, which is the surface area of the world oceans. Air-temperature data from the National Center for Environmental Prediction (NCEP)/National Center for Atmospheric Research (NCAR). At* http://www.cdc.noaa.gov/cdc/reanalysis/reanalysis.shtm.

that the total rise during the 1900s might be 2.5 to 3.1 mm a^{-1} and that thermosteric processes could entirely account for it (for example, Cabanes and others, 2001); this result from those satellite data produced an enigma ("the attribution problem") (Munk, 2002; Meier and Wahr, 2002).

This enigma may be resolved, at least partially, by some re-analyses of the tide-gauge and satellite observations. Miller and Douglas (2004) and Lombard and others (2004) suggest that 20th-century average rise in sea level is 1.5 to 2.0 mm a^{-1} and that only 0.5 to 0.8 mm a^{-1} of this rise is due to ocean warming. These authors base their conclusion on the ocean temperature data published by Levitus and others (2005); they add that the remaining, significant fraction can be attributed to eustatic inputs, such as glacier melt of about 1.3±0.5 mm a^{-1}. This contribution is likely to be mostly from glacier and ice-sheet melt because the only appreciable other sources are changes in land hydrology, which appear to result in both positive and negative effects on sea level in roughly equal fractions.

Our recent estimates of the average contribution to the rise in sea level from glaciers (1961–2003) is 0.51 mm a^{-1}, rising to 0.93 mm a^{-1} in the decade 1994 to 2003 (fig. 55). If this five-decade average is added to recent contributions from the Greenland ice sheet (0.12–0.215 mm a^{-1}; Krabill and others, 2000; Zwally and Giovinetto, 2000; Box and others, 2004; Thomas, 2004)[12] and the Antarctic ice sheet (0-0.14 mm a^{-1}; Bentley, 2004; Thomas and others, 2004),[12] glacier and ice-sheet melt may now account for 0.8 to 1.1 mm a^{-1},[12] most of the recently suggested eustatic contribution to sea level. This addition of freshwater to the ocean has significant effects on ocean circulation, ocean ecosystems, and sea-level change.

Glacier-ice melt has markedly accelerated during recent years and is likely to continue at a high rate into the future. Although many small glaciers will disappear, much of the current meltwater runoff is from large glaciers (for example, Bering Glacier and Malaspina Glacier (fig. 56) in Alaska), which will be slow to decrease in area, and additional glacier ice area will *add* to the hydrological cycle as cold glaciers continue to warm and begin producing runoff. Thus, it is important to understand the future effect of glaciers on sea-level rise and on ocean freshening as global warming progresses, so that we can project these future effects with some confidence.

Figure 56.—*Malaspina Glacier, with an area of about 5,000 km^2, is one of the two largest glaciers in Alaska. With annual losses averaging about 1 m of water equivalent, it and its neighbors are major contributors to current and future rise in sea level. These glaciers are so massive and so thick that their areas and volumes will not appreciably shrink during the 21st century. Painting by Mark F. Meier, 2004.*

Impact on the Earth's Gravitational Field

The transfer of mass between land and ocean influences changes in the Earth's shape. Earth's gravitational field is related to our discussion of variations in sea level. Glacier wastage plays a role here, too, both globally and regionally. The Earth's oblateness (the main component is known as J_2) varies at many time scales and influences Earth's rotation rate (length of day, *lod*) and the movement of Earth's rotational axis (polar wander). Using modern geodetic satellites, such as TOPEX and GRACE, geologists can measure the temporal and spatial changes in Earth's gravity field, J_2, and *lod* with remark-

[12] Most recent estimates, based on geodetic airborne laser altimetry measurements along profiles, have dramatically increased these values 0.4 to 0.6 mm a^{-1} for the Greenland ice sheet and the same for the much larger Antarctic ice sheet (Rignot, 2005); hence, meltwater from the two ice sheets and other glaciers has contributed about 2 mm a^{-1} to the rise in sea level.

able precision, thus permitting us to examine the role of glaciers in this system. In turn we can use the geodetic data as a check, both regionally and globally, on our estimates of wastage of glacier ice. However, some questions remain (Munk, 2002).

J_2, averaged over a long time, is decreasing as a result of tidal friction, postglacial rebound, mass transfers, and other effects (Munk and Revelle, 1952; Peltier, 1988; Munk, 2002; Cox and Chao, 2002). This decrease causes the Earth's rotation rate to speed up in order to preserve angular momentum. In 1998, however, J_2 began to increase (fig. 57). Dickey and others (2002) suggested that this increase might be caused, at least in part, by the acceler-

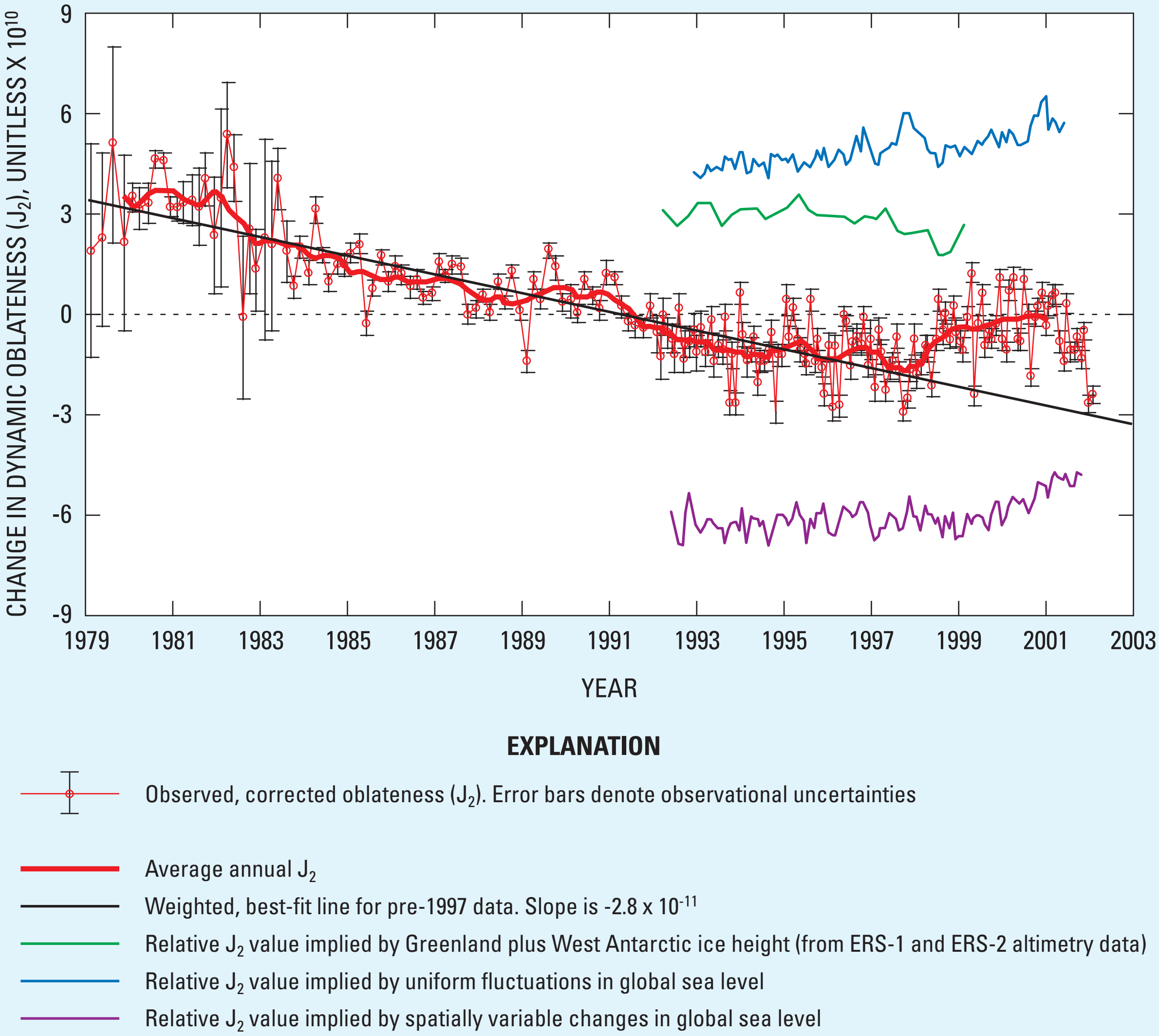


Figure 57.*—Time series of the Earth's oblateness (J_2). J_2 is decreasing over a long term, due to tidal friction, postglacial rebound, and other effects; as J_2 decreases, the speed of the Earth's rotation is increasing in order to preserve angular momentum. Dickey and others (2002) suggest that the marked increase in J_2 around 1998 might have been caused by the recent acceleration of glacier wastage. The offset green line represents the change in J_2 due to changes in surface height of the Greenland and Antarctic ice sheets; the blue line represents the changes implied by uniform fluctuations in sea level; and the purple line represents the changes implied by spatially variable changes in sea level. These changes, however, do not account for the 1998 increase in J_2. Modified from Cox and Chao (2002, p. 832, fig. 2). Figure courtesy of Science magazine. Used with permission.*

ated wastage of mountain glaciers. The long-term trend in J_2 is -2.7×10^{-11} a^{-1}. Ivins and Dyurgerov (2004) suggest that glacier wastage (very negative mass balances) during recent years could cause an addition of $+3.0 \times 10^{-11}$ a^{-1} to this trend ($+1.8 \times 10^{-11}$ a^{-1} due to Northern Hemisphere glaciers), not quite enough to reproduce the jump observed in 1998 but demonstrating that glacier wastage at high latitudes is important in understanding changes in the Earth's geophysical system.

These changes in gravity may be regionally significant as well. Regional uplift in southeastern Alaskan has been attributed to recent glacier wastage (Larsen and others, 2004). Sauber and others (1995) noted the effect of the recent surge of the large Bering Glacier on the local gravity field; this led Mark Meier (Institute for Arctic and Alpine Research, written commun., 1998) to suggest that gravity studies on nearby bedrock might be a useful tool for measuring mass changes in large glaciers that are inaccessible.

Glacier–Climate Interactions

Glaciers as Indicators of Climatic Change

Glacier mass-balance data (both annual and seasonal) can be used to infer climate variables such as precipitation and air temperature, and the spatial (geographic) distribution of these data can be used to assist in the analysis of climate and of climate modeling. First, however, we must explain the concept of "reference-surface balances" (Elsberg and others, 2001). These balances integrate mass-balance observations over an unchanging "reference surface" instead of over the area of a glacier that changes with time ("conventional balances"). Elsberg and others, (2001, p. 649) propose that "a [reference-surface] balance, which deliberately omits the influences of changes in area and surface elevation, is better correlated to climatic variations than the conventional one, which incorporates those influences."

Most of the conventional balances we mention here are computed over changing areas, although the areas are not necessarily measured annually. In order to test the applicability of the reference-balance method, we compare the two methods in a plot of the two calculated balances against air temperature (fig. 58). These show that there is a real, but small, difference. Over the years from 1968 to 1999, the difference amounts to about 2 percent. Applying this percentage to the total glacier area on Earth, we find a difference of about 3.6 km^3 a^{-1}, equivalent to about 0.01 mm a^{-1} of sea-level rise, a small correction. We recognize that our result, using conventional methods, may slightly understate the amount of glacier mass exchange with the climate.

Seasonal glacier-balance data, including winter balance b_w and summer balance b_s, provide estimates of distributions of precipitation and of summer temperature in high-mountain and high-latitude areas where observational climatic data are both scarce and biased. Very few long-term climatic stations are in operation above 3,000 m in altitude, and those that operate grossly underestimate the actual precipitation when compared to observed glacier winter balances (fig. 59). This is also true at high latitudes and is largely due to the difficulty of measuring precipitation in the form of blowing and drifting snow.

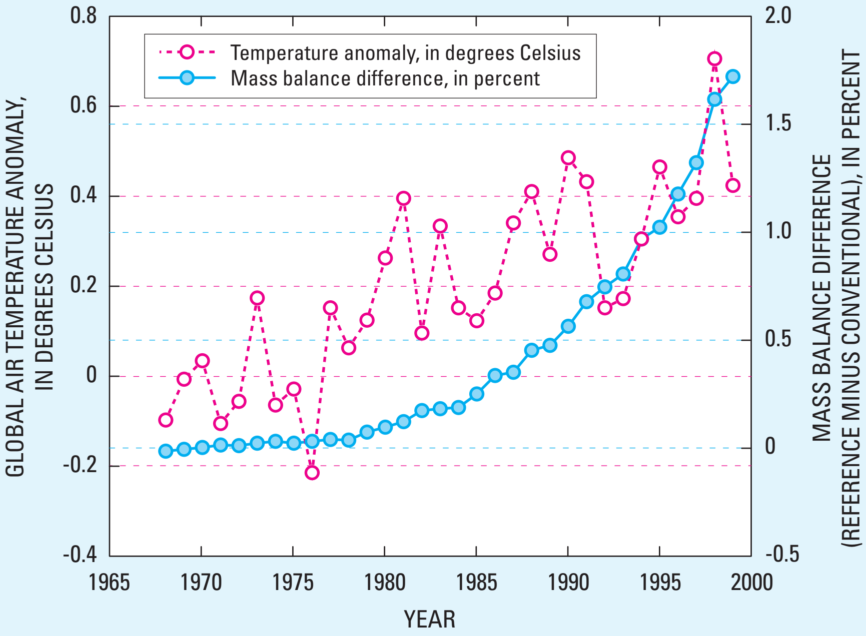


Figure 58.—*The differences between reference mass balances (glacier area considered constant) and conventional mass balances (glacier area changing in time, per observations), calculated for the mass-time series of 33 Northern Hemisphere benchmark glaciers. This difference, expressed in percent, shows an increase during 35 years (1965–2000) and may be indirectly related to the increase in positive air-temperature anomalies (temperature anomalies are from Hansen and others, 1999).*

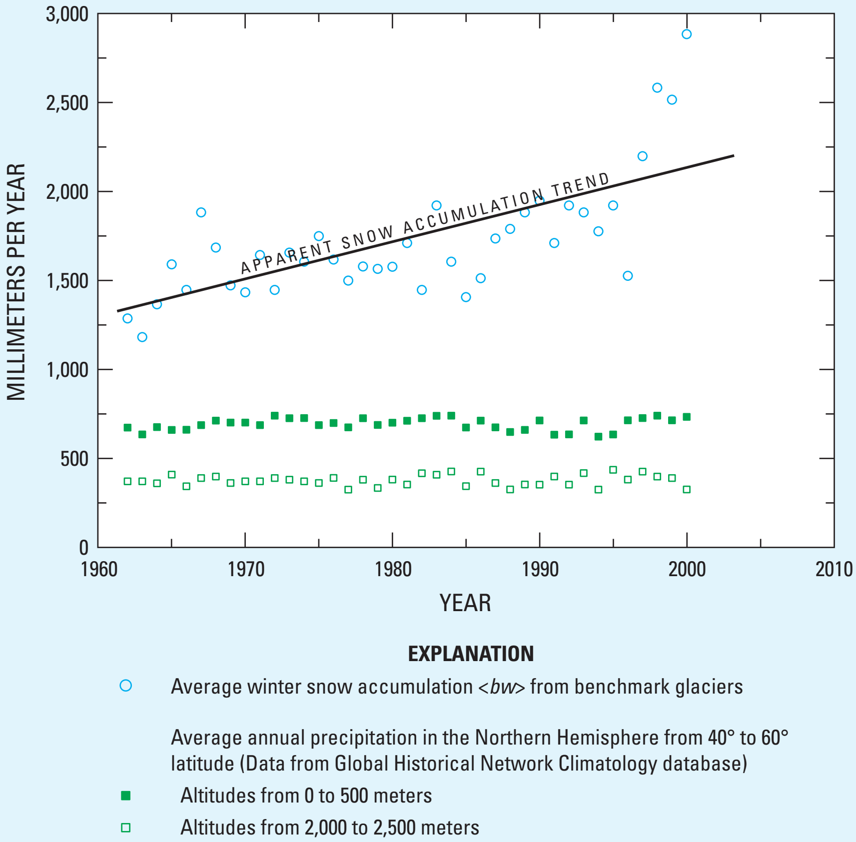


Figure 59.—*Winter snow accumulation, <**b**$_w$>, from benchmark glaciers (averaged for all available observations; see Dyurgerov, 2002), and annual precipitation, <**Pa**>, averaged for Northern Hemisphere latitude 40° to 60°, at two altitudinal ranges, 0 to 500 m and 2,000 to 2,500 m. From the Global Historical Network Climatology database. The apparent trend from 1960 to 2000 indicates winter snow accumulation sharply increasing from 1998 to 2000.*

Winter balance and precipitation obey rather different spatial statistics. Figure 60 shows the correlation with distance for both winter balance and observed precipitation. The correlation of point measurements of precipitation is not high even at short distances of separation, and it decreases to zero at a distance of 2,000 km. Winter balance, on the other hand, shows a high correlation at mesoscale distances, and it drops to only about 0.5 at 2,000 km, suggesting that accumulation values are well correlated spatially and are less influenced by local variations. Neither variable shows appreciable correlation at larger separations.

Another important measure of glacier-climate interactions is the change in mass-balance components with elevation (fig. 61) and the vertical gradient in mass balance, *db/dz* (see, for example, fig. 62) (Shumskiy, 1947; Meier and Post 1962; Kuhn, 1981, 1984; Dyurgerov and Dwyer, 2001). Observational results show that *db/dz* is changing with time; in particular, *db/dz* has become steeper in years with warmer climate conditions (fig. 62) because glaciers are losing mass at low altitudes in response to higher temperatures and gaining mass at high altitudes in response to increasing snow accumulation (Dyurgerov and Dwyer, 2001). This pattern indicates an increase in the intensity of the global hydrological cycle during these times of global warming (see also "Intensification of the global hydrologic cycle," by Huntington, in this chapter, p. A35–A51) and also results in acceleration of glacier flow, other conditions being equal. We suggest that *db/dz* is an important metric of glacier interaction with climate and that its change in many glaciers at the same period of time is evidence of large-scale climatic change.

Along with this change in the mass-balance gradient, the altitude of the equilibrium line (ELA) has been increasing, and the accumulation area ratio (*AAR*), the glacier area above the ELA divided by the total glacier area, has been decreasing (fig. 63). One interesting indication of recent shifts in glacier mass balances is seen in the averaged *AAR* data plotted as standard departures (fig. 63); this shows an accelerated decrease in the *AAR* in about 1977 and again in the early 1990s, in common with other evidence of increased glacier wastage. Measurements of *AAR* can be made from satellite images, so this metric is especially useful.

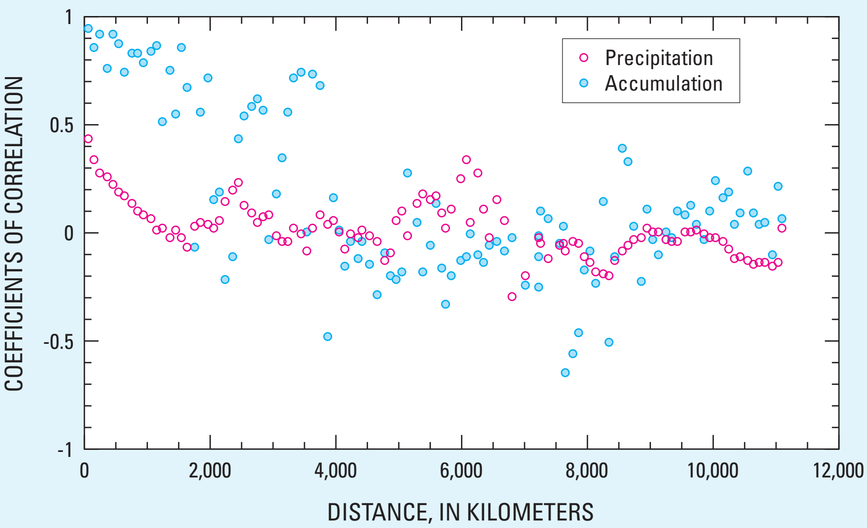


***Figure 60.**—Spatially distributed patterns of autocorrelations computed for annual snow accumulation on glaciers and for annual precipitation at 1,000- to 1,500-m elevation in the Northern Hemisphere. From meteorological stations, National Climate Data Center (NCDC) database. The winter snow accumulation, b_w, is the maximum amount of snow accumulation measured at the glacier surface at the end of accumulation season. These b_w are usually 20 to 30 percent less than the annual amount of snow accumulation.*

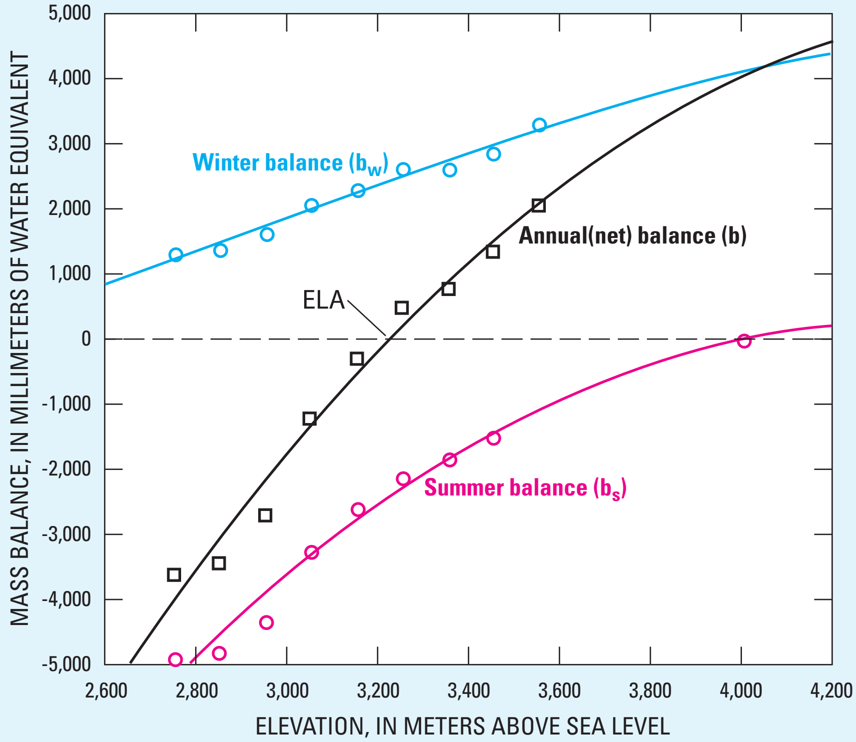


Figure 61.—*Changes in mass balance for winter (b_w), summer (b_s), and in annual/net mass balance (b) for a single glacier, Djankuat Glacier, Central Caucasus, Russia, at increasing elevation. Observational data have been averaged from 1968 to 1997. Meltwater runoff (≈b_w) is about zero at elevations near 4,000 m where annual mass balance equals winter balance, which is annual snow accumulation. The equilibrium-line altitude (ELA) is the elevation on a glacier that separates the ablation area from the accumulation area (Paterson, 1994).*

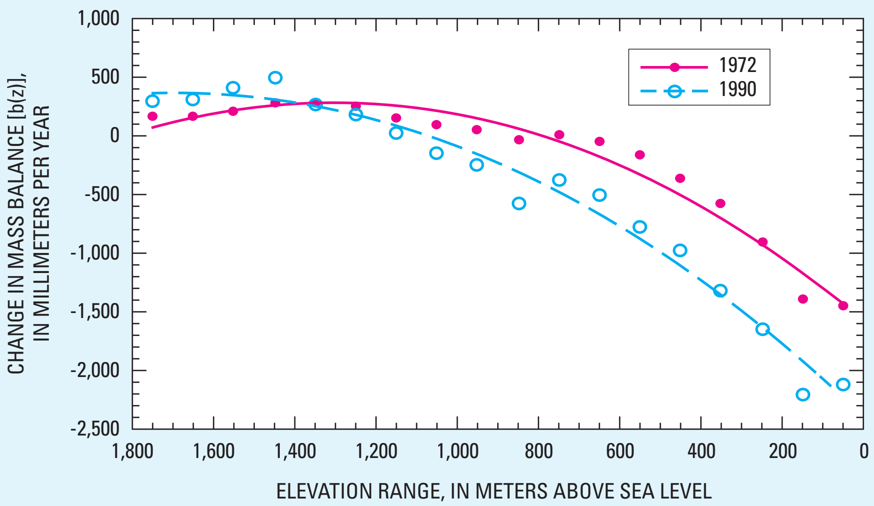


Figure 62.—*Change in mass-balance gradient between cold (1972) and warm (1990) years. Data on mass balance were averaged for 21 Northern Hemisphere glaciers and adjusted to the same elevation.*

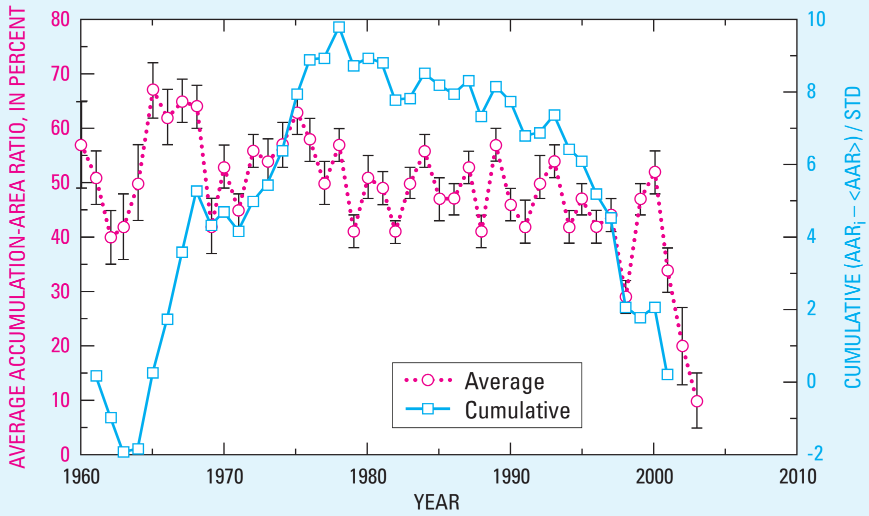


Figure 63.—*Variability of AAR (AAR$_i$), and the change with time of the accumulation-area ratio <AAR> in terms of standardized cumulative departure. AAR$_i$ averages data for all time series longer than 5 years; bars are standard errors. The change with time of <AAR> shifts to a decrease at the end of 1970s; data after 2001 are incomplete. AAR$_i$ is the mean of the annual values for AAR for all time series; it averages <AAR> during the period from 1961 to 2001.*

The current increases in winter balances at high elevations have been especially rapid in the last decade (1990–2000); they have not been paralleled by appreciable increases in precipitation as measured at lower altitude meteorological stations (fig. 59). This points to a significant increase in the intensity of the hydrologic cycle at high elevations. The increase in b_w is even more remarkable considering the simultaneous decrease in glacier accumulation areas.

An appropriate measure of this change in both major components of glacier mass balance is the glacier mass turnover, the average of the absolute values of b_w plus b_s (Meier, 1984) (fig. 64*A*). Another important measure is the sensitivity of mass balance to air temperature, db/dT (fig. 64*B*). Both of these observations, which demonstrate changes in the warm decades of the 20th century, have not been predicted by, or used in, glacier-climate models (for example, Church and Gregory, 2001; Zuo and Oerlemans, 1997). An interesting result of these observations is that the variability of mass turnover and of the sensitivity of mass balance to air temperature shows a sharp change at the end of 1980s, followed by major decreases in variability. This is temporal variability, but it represents spatial (geographic) variability as well, because the calculation is based on dozens of time series in different geographical locations.

The spatial-temporal changes in glacier mass balances appear to be forced by changes in air temperature, which has increased globally, most notably since the late 1970s. Figure 65A presents air temperature as a function of time and latitude, showing zonal anomalies. Figure 65B presents mass-balance standardized departures calculated for large glacier regions. This shows that an acceleration in volume wastage in some regions started as early as the 1970s (for example, in Central Asia) and was completed by the end of the 20th century in other regions—for example, in the Arctic.

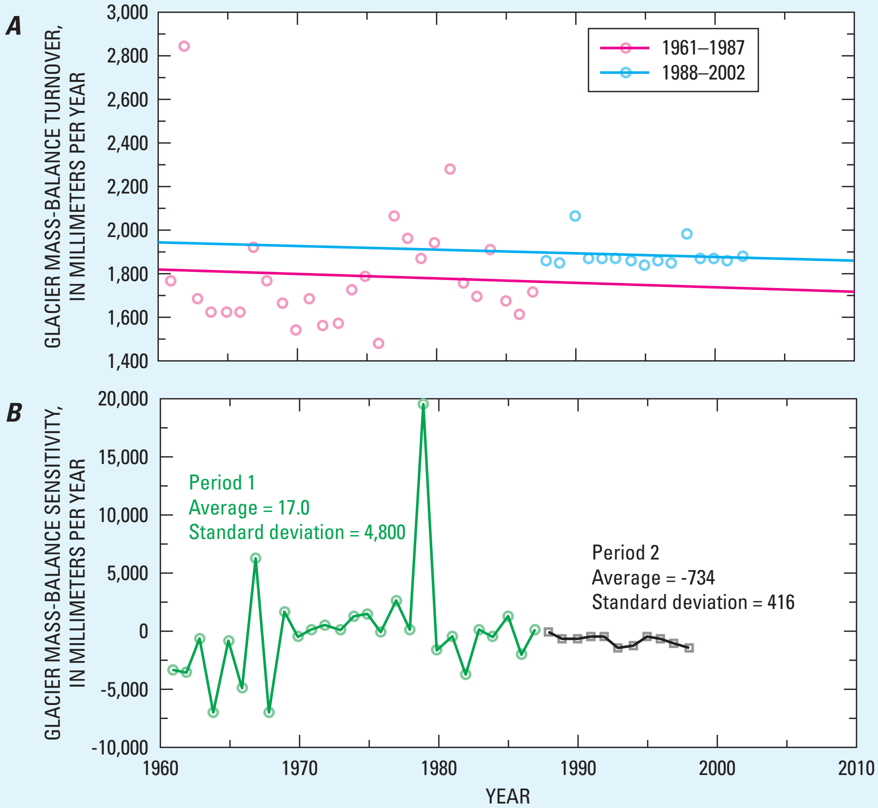


Figure 64.—**A**, *Glacier mass-balance turnover dramatically increased after 1987, and annual variability decreased at the same time.* **B**, *The mass-balance sensitivity to the globally averaged air temperature also has increased, accompanied by a decrease in annual variability at about the same time. Northern hemisphere glacier mass balances are used to calculate sensitivity to annual air temperature. Long-term annual mass-balance time series averaged for about the same 40 benchmark glaciers have been used to calculate averages (Dyurgerov, 2001). Note that this is a different measure of sensitivity than that used by the Intergovernmental Panel on Climate Change (IPCC) (Church and Gregory, 2001); the IPCC measure involves a change between two steady states.*

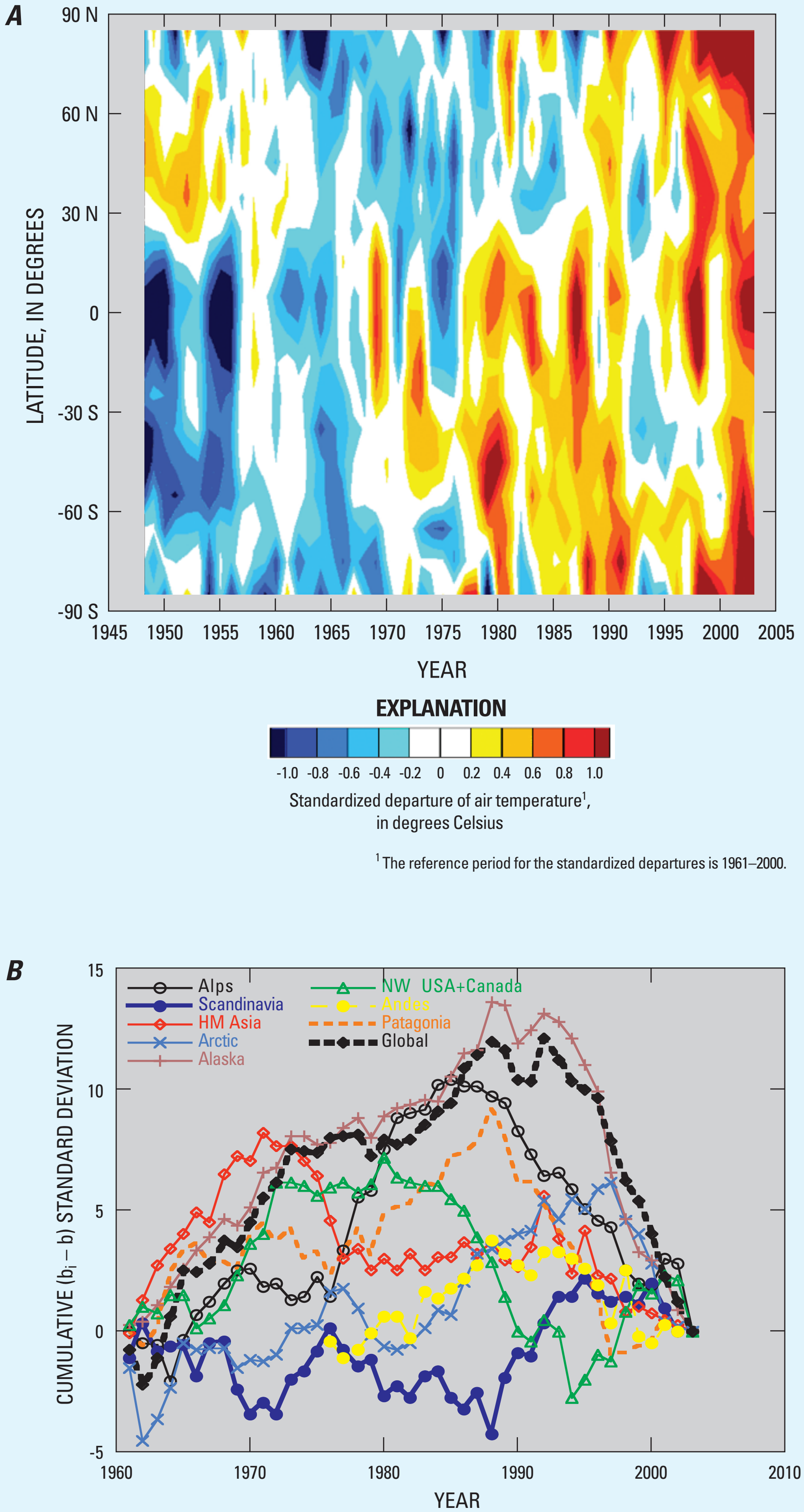


***Figure 65.*—A**, *Temperature as a function of time and latitude showing zonal anomalies. Data from National Center for Atmospheric Research (NCAR) reanalysis dataset calculated by McCabe (in Dyurgerov and McCabe, 2006). At* http://www.cdc.noaa.gov/cdc/reanalysis/reanalysis.shtml. **B**, *Shifts in timing towards acceleration in wastage of glacier volume are expressed in standardized cumulative departures. These graphs show that different glacier systems have responded to large-scale changes in climate at differing times, from the early 1970s until the end of the 1990s. This process of change in glacier volume in response to climate has taken about three decades; bi is the regional and global mass balances for individual years, i; <b> is average mass balance during the period 1961–2003.*

The large differences between observations of snow accumulation and measured precipitation indicate that one must use extreme caution and make necessary adjustments in any use of precipitation data for modeling glacier mass balance and for projecting the contributions of glacier meltwater to changes in sea level.

These glacier data give realistic information on the fundamental processes of climate-glacier interrelations, many of which are not realized or predicted by existing models. Because glaciers are major contributors to global and regional hydrologic cycles, observations on them deserve more attention as we improve monitoring of the evolving Earth System.

Glacier Hydrology and Its Impact on Ocean Salinity

Freshwater runoff from glaciers has distinctive characteristics: a natural regulation that buffers the effect of warm/dry years or cool and wet years, seasonal water storage and release in summer when it is generally most needed, a high sediment discharge and a marked daily variation in flow that renders stream channels unstable, and the possibility of temporarily storing bodies of water adjacent to or under the ice, causing damaging floods or sudden release in certain regions (Meier, 1969b). In addition, this runoff is a component of the exchange of freshwater with the ocean, which, in turn, affects ocean circulation. Thus, glacier runoff affects water resources, agriculture, hydroelectric power, the environment, the economy, and even ocean circulation.

Two main characteristics may be used to define glacier hydrologic impacts. The first is the area covered by glacier ice relative to the area of the entire watershed. For example, for the Antarctic ice sheet, the ratio is approximately 1.0; for Asia, it is <0.003; for the Vernagtferner Basin (Alps), it is about 0.8. The second variable is the glacier mass turnover, which is the average of the absolute values of snow accumulation in water-equivalent (b_w) and of surface ablation (b_s), as discussed previously. Note that it is important to use conventional balances, rather than reference-surface balances, in hydrologic studies. In this context, we shall not discuss iceberg calving.

Glacier Impact on the World Ocean

First, we hypothesize glaciers to be in a steady state: glacier area is constant so that mass balance is zero, and annual meltwater runoff is approximately equal to annual precipitation, which is the accumulation of snow on glacier surface plus the liquid precipitation that occurs generally in summer. We used available observational results of b_w in the Northern Hemisphere and to these values added summer precipitation, which is estimated to be about 1/3 of b_w. The resulting estimate of average precipitation on glaciers is about 912 mm a^{-1} for about 40 glaciers worldwide from 1961 to 2003. We also made another estimate of annual precipitation and accumulation, using the mass turnover; these observational results produce an average value of 970 mm a^{-1}. The average value of these two determinations is 940 mm a^{-1}. Average precipitation on Northern Hemisphere land is about 690 mm (Gleick, 1993; averaged from six different calculations, table A6). The ratio of the two values is 1.36; that is, precipitation over glaciers is, on the average, about 36 percent higher than that over unglacierized land.

In actuality, glaciers have not been in a steady state during the last 30 to 40 years since about 1970. Acceleration in mass losses is reported here and in many other publications. These negative mass balances represent the loss of perennial ice and water storage, which together augment runoff. We used the same results presented in figure 55. To obtain the annual total glacier runoff (excluding iceberg calving), we multiplied the winter balance rate by 1.36. The upper curve in figure 66 shows the trend and variability in glacier freshwater flux to the world ocean. On a worldwide scale, the total glacier meltwater runoff (excluding the Greenland and Antarctic ice sheets) is small relative to the river runoff (about 40×10^3 km^3 a^{-1}; Gleick, 1993). Regionally, however, glacial meltwater runoff is substantial and can dominate other sources of freshwater, not only during the summer melt period but also over long periods of time.

Glacier Meltwater Flux to the Arctic Ocean

The Arctic Ocean Basin is unique in the Northern Hemisphere in that glaciers, mostly on arctic archipelagos, contribute water directly to the ocean rather than to the large rivers flowing into the ocean (Dowdeswell, and Hagen, 2004). Only some glaciers in western Europe, Iceland, Scandinavia, and part of Alaska contribute meltwater to major rivers. These glaciers feeding gaged rivers constitute only about 8 percent of the aggregate glacier area (315×10^3 km^2) in the pan-Arctic region (Dyurgerov and Carter, 2004).

The following analysis is based on the nine major river basins in Siberia and North America that are regularly gaged (fig. 67, excluding the contribution from western Europe, basin 10). Because the glacier area in these nine basins is less than 4 percent of the total pan-Arctic glacier area, river-discharge data cannot provide an integrative measure of water balance for the entire land area of the pan-Arctic. In this region therefore, glaciologists must study separately

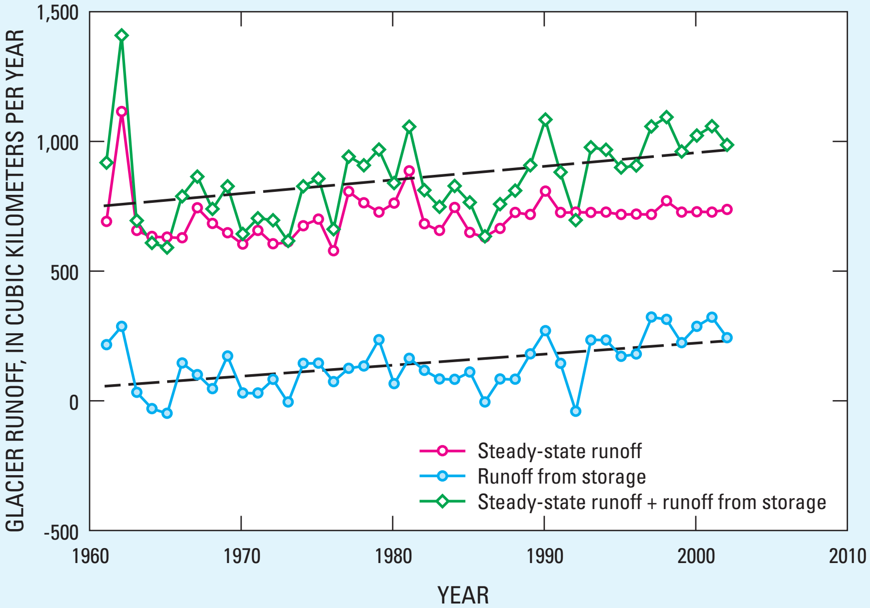


Figure 66.*—The components of meltwater runoff from glaciers. Observational results for winter mass balance were used to calculate the steady-state runoff for 40 glaciers worldwide during the period 1961–2003, steady-state runoff increased by one-third to account for accumulation in four summer months to approximate the annual value. These annual values of mass gain were multiplied by the glacier area of 785,000 km^2 to obtain the annual rate of flux of total freshwater volume from glaciers to the world ocean, excluding flux from the Greenland and Antarctic ice sheets (steady-state components in this figure). Runoff from storage is (equal to) glacier mass balance.*

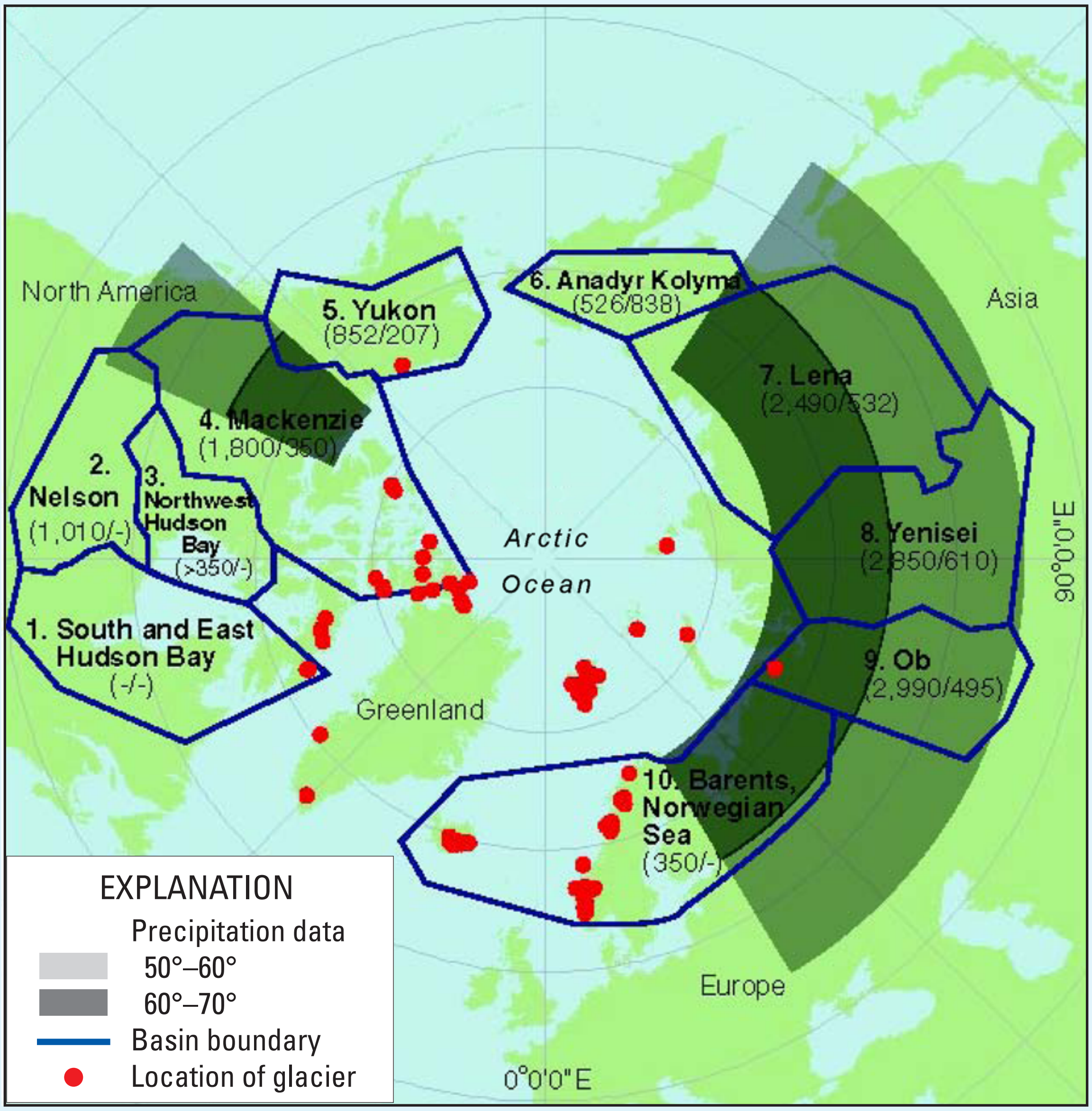


Figure 67.*—The pan-Arctic drainage area includes the Arctic archipelagoes as well as continental watersheds. The 10 large pan-Arctic river basins (boundaries are shown) with available annual runoff data were used to calculate river runoff to the Arctic Ocean. Red dots show benchmark glaciers with mass-balance records that we used to calculate fresh water runoff from glaciers to the Arctic Ocean. Two precipitation data zones are also shown. Basins are numbered from 1 to 10. First number in each pair of parentheses indicates basin area in thousands of square kilometers; second number indicates discharge in cubic kilometers per year (Lammers and others, 2000). Note that the Yukon River Basin (as defined by Lammers and others (2000)) includes the entire Brooks Range, including the benchmark McCall Glacier (see Chapter K, p. K471–p. K476, and figs. 440 and 441 on p. K474 and p. K475, respectively) and many other river basins that drain to the Arctic Ocean; however, accurate discharge data are not available for basins 1, 2, 3, and 10. Basin 10 does not include calculations of river discharge and glacier meltwater production to reduce the overlap between gauged and ungauged data.*

the two major components of freshwater inflow to the ocean from rivers and from glaciers. This separation, however, allows us to estimate the impacts of glaciers and to compare them with the impacts of rivers or of change in the hydrologic cycle in the Arctic.

River-runoff data show little trend in runoff in the pan-Arctic (fig. 68*A*). During the reference period (1961–1990), no increase in river runoff became evident; only after 1995 did an increase appear (fig. 68*B*). This increase may have resulted from several ongoing changes in water storage, such as degradation of permafrost, damming of reservoirs, or glacier wastage in the watersheds. These changes are poorly known but are believed to be small (Yang and others, 2003; Zhang and others, 2003; Holmes and others, 2005).

The glacier contribution of freshwater to the Arctic Ocean, on the other hand, has been steadily increasing (fig. 69). This contribution comes primarily from glaciers in the Canadian, Russian, and Svalbard archipelagos and from individual ice caps around the Greenland ice sheet. The increase in glacier contribution, combined with a small increase in river runoff (fig. 68*B*) since

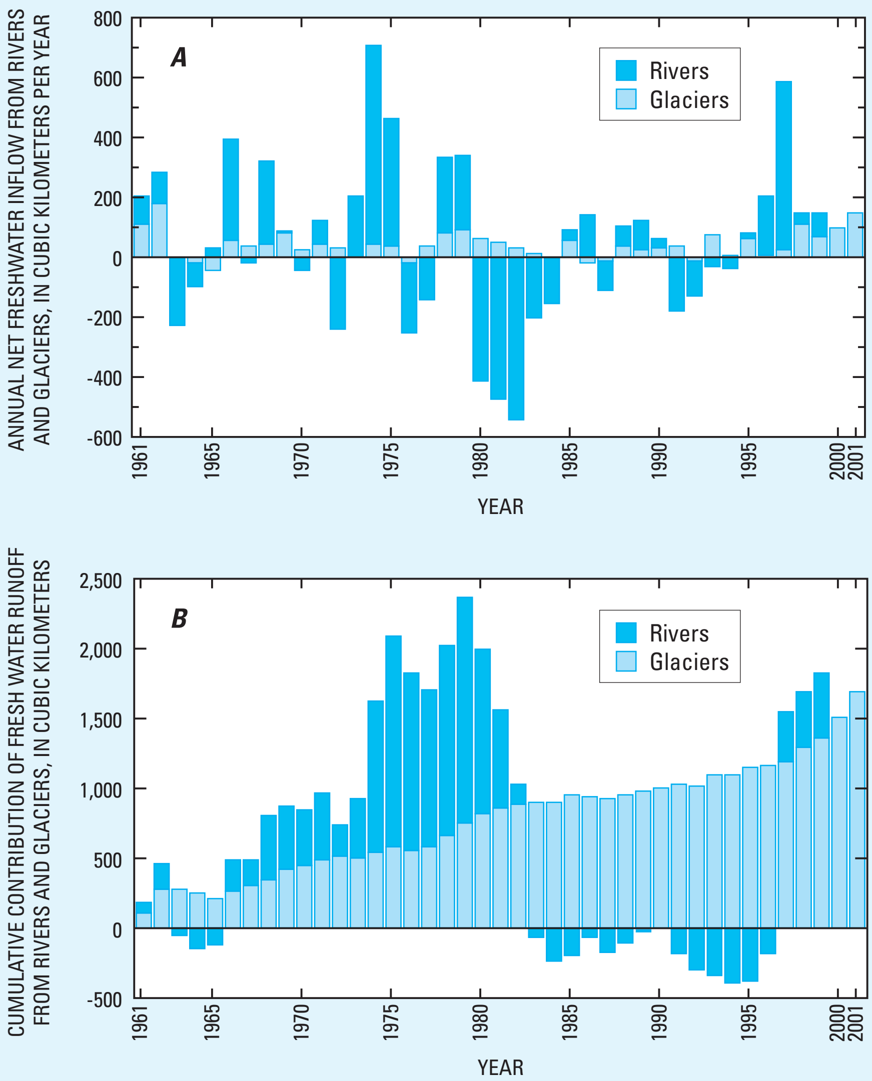


Figure 68.—**A**, *Annual net inflow from pan-Arctic rivers and glaciers, not including the Greenland ice sheet.* **B**, *Cumulative contribution from rivers (standardized departures) and glaciers during the period from 1961 to 2001.*

the end of 20th century, support the conclusion derived from modern hydrographic records on freshening of North Atlantic waters (Bard, 2002). This, in turn, suggests that deep Arctic ventilation has steadily changed during the last 40 years (Dickson and others, 2002). This freshening and ventilation can be used to model future climate, changes in sea-ice area, bioproductivity, and other environmental changes in the Arctic. Fluctuations in freshwater runoff from both rivers and glaciers have been large and are consistent with large fluctuations of sea-ice thickness. Evidence of increases in ice thickness on the Arctic Ocean from 1998 to 2002 suggests decreases in seawater salinity due to increases of inflow from rivers and glaciers.

It is likely that freshwater inflow to the Arctic Ocean from glaciers will continue to increase as a result of climate warming. Larger areas of ice caps will produce more runoff. Monitoring these ice caps will allow development of a long-term environmental strategy. The continuing decline in observational networks on glaciers and ice caps, and their ensuing runoff, increased the difficulty of monitoring and forecasting of the hydrologic cycle in the Arctic.

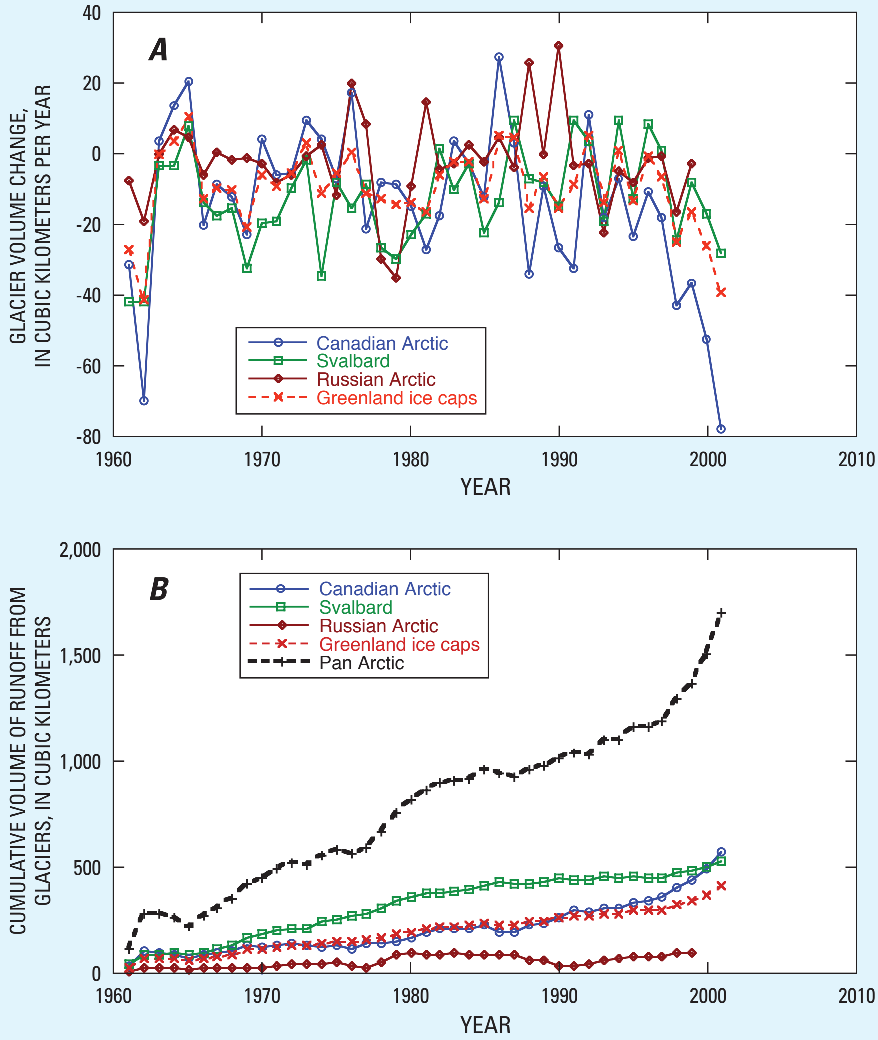


***Figure 69.*—A**, *Change in volume of glaciers, calculated for large Arctic archipelagoes during the study period from 1960 to 2010.* **B**, *Cumulative values of the annual contribution of runoff during the study period from 1960 to 2010 for the same glacier areas.*

Glaciers in High-Mountain Regions

Areas of high-population in and near the mountains in Asia and the Americas derive much of their usable water from snowmelt and icemelt. Glacier contribution to annual total runoff of the important rivers Amu Darya, Syr Darya, and Ily[13] in (the former Soviet Union part of) Central Asia is from 5 to 40 percent and as much as 70 percent in the upper basins (Kemmerikh, 1972). The glacier contribution to total annual runoff averaged over the Central Asia region increases from 15.6 percent to 40 percent during the ablation season (Kemmerikh, 1972; Krenke, 1982).

In the highly glacierized watersheds of the Aral Sea (Syr Darya and Amu Darya rivers), annual and summer air temperature increased by 0.017 to 0.043 C a^{-1}, and annual precipitation slightly decreased from 1962 to 1991 compared with the previous period of 1932 to 1961 (Konovalov and Williams, 2005). As a result, glacier area was reduced from 8,754 km^2 to 6,920 km^2 (21 percent), and volume was reduced from 517 km^3 to 375 km^3 (27 percent). As a result of these changes, the glacier meltwater contribution to river runoff and flow into the Aral Sea was reduced (Konovalov and Williams, 2005).

The Yellow (Huang He), Yangtze, Brahmaputra, and other rivers of eastern and southern Asia have their source in glacierized mountains; the contributions of glacier runoff to these rivers is, however, poorly known. In the Las Cuevas river basin (Mendoza, Argentina), glacier runoff may reach 73 to 90 percent of the total flow and 75 percent for the San Juan River Basin in the driest years. The contribution of meltwater runoff from glaciers has increased during the last 40 years because of a decrease in precipitation (Milana and Maturana, 1999; Leiva and others, 2007).

The glacier area in the tropical regions is relatively small, and the hydrological significance of glaciers here may only be local, except perhaps in Perú. The glacier area defined for tropical South America from 1950 to 1980 was 2,700 to 2,800 km^2, decreasing to less than 2,500 km^2 by the end of 20th century (Kaser and Osmaston, 2002). (See "Glaciers of South America," Chapter I in this series [*http://pubs.usgs.gov/pp/p1386i*].) More than 70 percent of this area is concentrated in the Cordillera Blanca of Perú (Kaser and Osmaston, 2002); these glaciers provide significant river runoff. Small glaciers also occur in the tropics in East Africa (in Kenya, Tanzania, Uganda, and Zaire) and Irian Jaya, Indonesia (Kaser and Osmaston, 2002). (See "Glaciers of the Middle East and Africa, Chapter G in this volume [*http://pubs.usgs.gov/pp/p1386g*] and "Glaciers of New Zealand and Irian Jaya, Indonesia, Chapter H in this volume [*http://pubs.usgs.gov/pp/p1386h*].) A specific feature of glacier runoff in the tropics is that there is no substantial seasonality in glacier melt (owing to very small seasonality in air temperature), as well as little annual variation in precipitation. This means that glaciers show little change from year to year in their contribution to river runoff (Kaser and Osmaston, 2002). Tropical glaciers show a remarkably small difference between the seasonal course of the coefficients of precipitation and the coefficients of runoff (ratio between monthly and annual mean values) compared to, for instance, the European Alps, where the differences between these coefficients are large (Kaser and Osmaston, 2002).

[13] The geographic place-names used in this section conform to the usage authorized for foreign names by the U.S. Board on Geographic Names as listed on the GEOnet Names Server (GNS) Web site: *http://gnswww.nga.mil/geonames/GNS/index.jsp* . Any geographic place-names not listed on the Web site are shown in italics.

Local Hydrologic Impact: An Example

Glaciers may play substantial roles on the local scale. One example from the United States is the high- mountain watersheds in the Front Range of the Rocky Mountains of Colorado, where several tiny glaciers are sources of water for agriculture and the City of Boulder, Colo. The largest of these is the Arapaho Glacier (0.28 km^2 in area). Mass balance was measured in different periods in the 1960s and 1970s. The balance was partially reconstructed for other years up to 2002, using snow-accumulation measurements at the end of winter at neighboring Arikaree Glacier, Colo., and ablation of snow and ice was calculated by means of a positive degree-day model with air temperatures from several neighboring automatic meteorological stations. The mass change of the small Arapaho Glacier shows a large downward trend, which is consistent with but more exaggerated than trends calculated for glaciers in the Pacific Northwest and with the global mass change (fig. 70).

In summer 2003, mass-balance components were measured every 2 to 3 weeks to determine when the glacier began to release water from long-term storage and when the transient mass balance crossed the zero value and became negative. This happened at the beginning of August 2003 (fig. 71), and Arapaho Glacier continued to contribute meltwater from its storage to the local watershed until 22–23 October 2003. The aggregate amount of this water from storage was 2 m averaged over the glacier area; the amount of precipitation at the same period was only several centimeters. This water from ice storage is associated with the negative mass balance resulting from dry and warm weather conditions, periods when glaciers are especially valuable sources of freshwater (Meier, 1969).

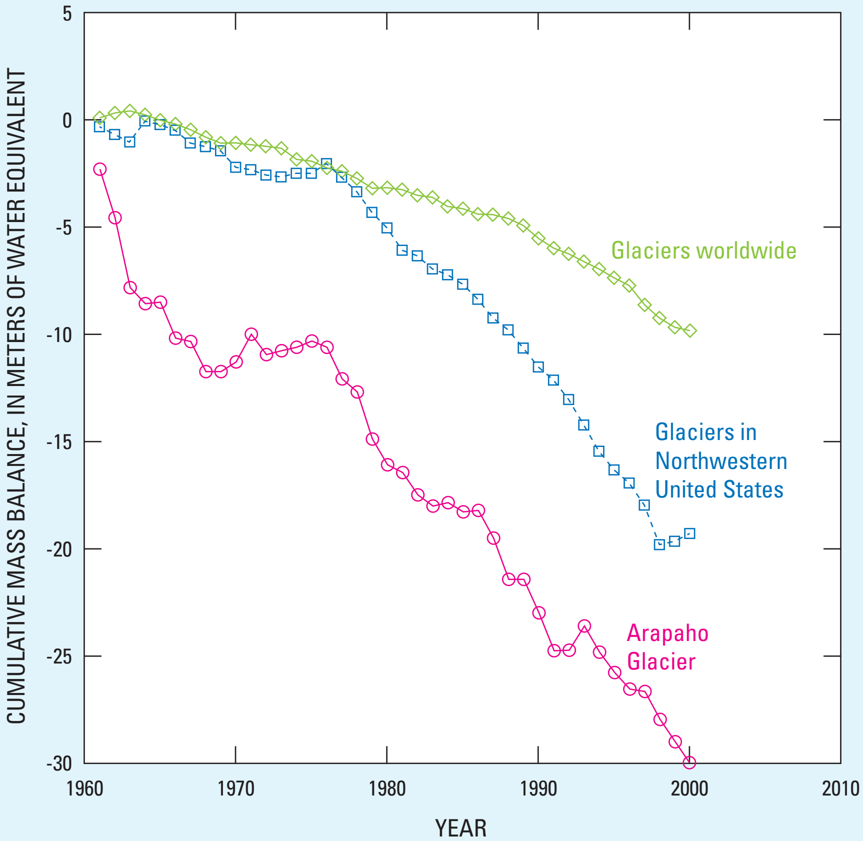


Figure 70.*—Loss of mass from Arapaho Glacier, Colo., mean loss of mass from glaciers in the Pacific Northwest, and mean loss of mass in glaciers worldwide.*

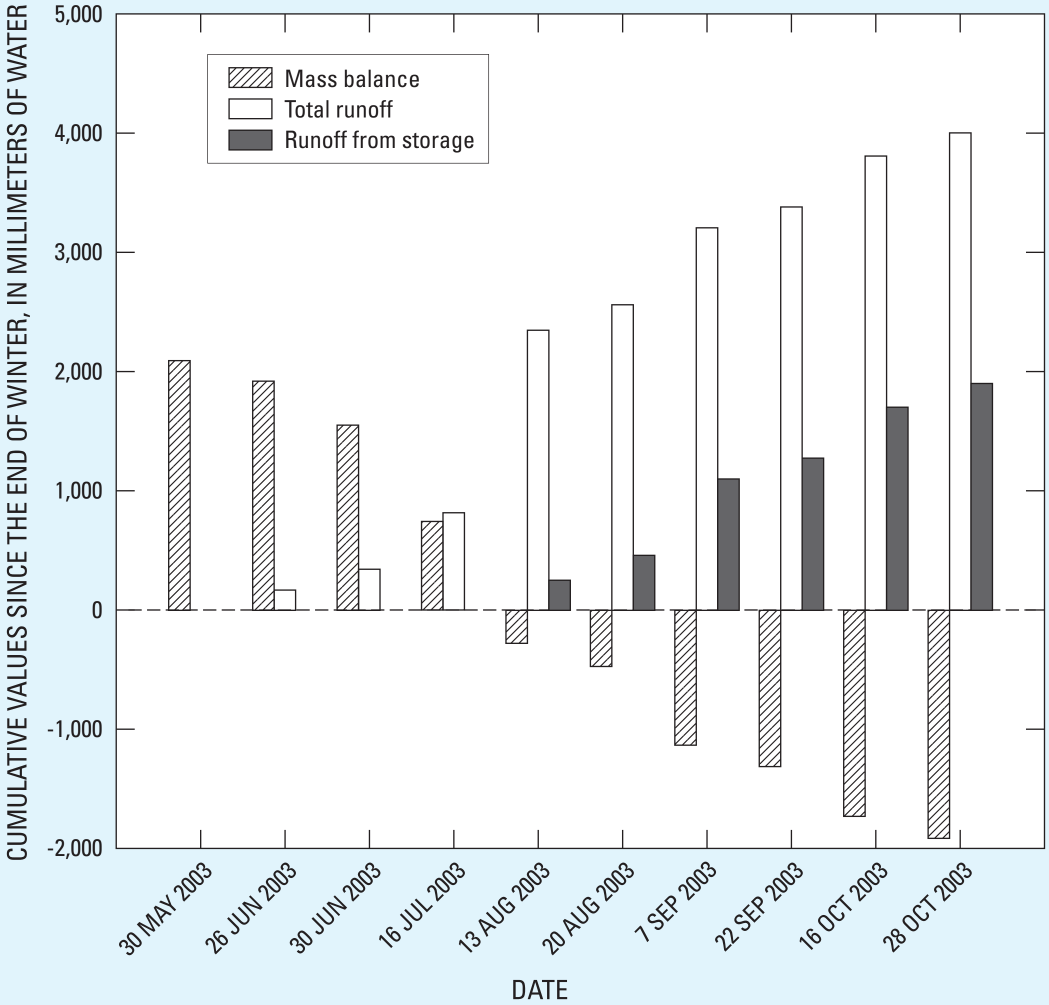


Figure 71.*—Mass balance, meltwater runoff, and runoff derived from storage for Arapaho Glacier, Colo., during the 2003 ablation season (May–October); cumulative values begin with May 2003 data and end with October 2003 data.*

Other Effects of Glaciers on the Environment

Large-Scale Glacier-Induced Events

A series of extremely warm years, with a marked increase in glacier melting and acceleration in iceberg calving, increases the freshwater contribution to the ocean. In unpredictable ways, this increase may affect ocean thermohaline circulation, water salinity, and sea-ice extent; the trajectory of oceanic currents may also be altered (Bigg, 1996). The possibility of unexpected large-scale, glacier-related events in connection with changes in climate cannot be entirely ruled out. "Once dramatic change has occurred, returning to a previous state may not be merely a matter of reversing the freshwater forcing, as once the ocean circulation has entered a particular state it may be resistant to change. The ocean state can show bifurcation." (Bigg, 1996).

A different combination of extreme climatic events—for example, global climate cooling caused by several explosive volcanic eruptions closely spaced in time—may trigger glaciers to gain mass and start to advance with poorly predicted regional effects. Our late 20th- and 21st-century civilization has never dealt with such events, so we have no recent experience with large-scale glacier advances of glaciers (Nesje and Dahl, 2000).

Iceberg Calving

Another component of glacier mass balance that is not measured or reported in most conventional glaciological studies is iceberg calving. This process is important for many glaciers in coastal Alaska, southern South America, and the sub-Arctic and subantarctic islands. Many other glaciers end in lakes (lacustrine calving) or discharge ice by breakoffs onto land; these are generally smaller components of the overall mass balance. Such "mechanical ablation" needs to be considered in order to obtain a more accurate representation of the relation between climate and mass balance and to estimate the effect of calving on sea-level rise. The calving of floating glaciers, of course, has no effect on sea level, but most glaciers have grounded termini. Iceberg calving in high latitudes of both hemispheres is of interest and affects various human activities such as shipping, coastal and offshore installations, and fisheries; icebergs could be used for freshwater supply in the regions with shortage of freshwater (Swithinbank and others, 1980; Camirand and others, 1981).

Glazovskiy (1996) estimated the rate of calving for glaciers in the Russian Arctic. Brown and others (1982) compiled calving rates for the Alaskan coast. Iceberg calving into the lake below the Mendenhall Glacier (Juneau Ice Field, Alaska) amounted to 5.7 percent of the surface ablation from 1999 to 2000 (Motyka and others, 2001). Unpublished estimates of total calving for Patagonia and Tierra del Fuego have been made by Per Holmlund (University of Stockholm, written commun., 1997) and for Svalbard by Jan Øve Hagen (University of Oslo, written commun., 1997). Mark Meier (Institute for Arctic and Alpine Research, unpublished data) has made preliminary estimates of calving discharge for the Canadian Arctic and for small glaciers around Greenland and Antarctica by estimating the width of calving fronts and individual calving rates on the basis of analogous climatic regions. J.G. Cogley, in his 2002 database [*http://www.trentu.ca/geography/glaciology.htm*], has included the characteristics of potential calving glaciers, both tidewater and lacustrine, from World Glacier Monitoring Service sources. Various other measurements of individual glacier calving, of all types, have been made. Warrick and others (1996) combined the estimates of Glazovskiy, Brown, Holmlund, Hagen, and Meier to suggest that iceberg discharge into the sea from glaciers at present (in the early 21st century) totals between 22 and 80 $km^3\ a^{-1}$, but this result is untested and uncertain. Probably mass loss by calving is larger considering that Columbia Glacier, Alaska, alone discharges more than 7 $km^3\ a^{-1}$ (O'Neel and others, 2005).

Glacier Archaeology and Palaeontology

Glacier recession has produced a wave of interest to study newly exposed archeological evidence (Dixon and others, 2005). This interest accelerated with the discovery in 1991 of a Neolithic "Ice Man" in the Tirolean Alps (Spindler, 1994; Fowler, 2000). Since that time, the field of glacier archaeology has expanded, and many new discoveries have been made, especially with the new use of space images and GIS (Dixon and others, 2005).

Ice Recession

The exposure of newly deglacierized terrain due to disappearing glaciers has mixed socioeconomic effects. More than 700 km^2 of land has become freed from glaciers in the eastern Arctic and perhaps even more in the Canadian Arctic archipelago (Burgess and Sharp, 2004). This new exposure may have ecological and economic interest because it may expand plant cover or alter traditional animal migration routes, and it may uncover new mineral resources. Gold mining has been developed in the internal Tien Shan, Ak-Shirak range, by artificial removal of glacier ice (Khromova and others, 2003).

The loss of ice climbs because of glacier recession has affected the Alps, Himalaya, and other mountain regions (for example, Bowen, 2002). The degradation of scenic impact because of disappearing glaciers is expected to damage real estate and tourism in alpine countries (for example, News Wales, 2004, [*http://www.newswales.co.uk/?section=Environment&F=1&id=6872*]).

Glacier Hazards

Regional- and local-scale glacier hazards are pervasive. These are common in most glacierized high-mountain areas in the world. Their impact on society depends on the density of human structures and settlements in those regions. Among such glacier-related hazards are (1) lahars and other types of debris flows, (2) glacier outburst floods (jökulhlaups), (3) ice and debris avalanches, and (4) rockfalls and rockslides.

Lahars are mudflows or debris flows originating on the flanks of a volcano caused by the mobilization of tephra deposits during high-precipitation events or explosive volcanic events. Some lahars originate with glaciers and are lubricated by glacier meltwater. An example is the lahar initiated by the explosive eruption of Mount St. Helens, Washington, in 1980, that moved down valley with speeds from 16 to 80 $km\ h^{-1}$, damaged 27 bridges and 200 homes, destroyed nearly 300 km of highways and 24 km of railways, and damaged wildlife and fisheries (Lipman and Mullineaux, 1981). This lahar was lubricated, in part, by the rapid melting of glaciers on the mountain (Brugman and Meier, 1981), which provided a jökulhlaup component.

Debris and glacier-ice avalanches are especially dangerous on local scales because they can occur spontaneously, often without any advance warning. Earthquakes, steam explosions (from geothermal or volcanic activity), and intense rainstorms can trigger debris avalanches from parts of a glacier. Since 1702, more than 20 catastrophic events have resulted from ice avalanches that have caused outburst floods from glacier lakes in the Cordillera Blanca or Perú. Two of them, in 1962 and 1970, destroyed several villages and caused the deaths of more than 25,000 inhabitants (Morales-Arnao, 1998, p. I67–I71; [*http://pubs.usgs.gov/pp/p1386i/peru/hazards.html*]). On 20 September 2002, a collapse of a hanging glacier from the slope of *Mount Dzhimarai-Khokh* onto the *Kolka Glacier* triggered an avalanche of ice and debris that went over the Maili glacier terminus and then slid more than 24 km farther. (See also the section on the Kolka Glacier in "Glaciers of the Former Soviet Union" in the "Glaciers of Asia" in Chapter F of this series [*http://pubs.usgs.gov/pp/p1386f*].) The avalanche buried small villages in the Russian Republic of North Ossetia, killing dozens of people (Desinov and others, 2002; Haeberli and others, 2004). Smaller, but significant, rockfalls occur on unstable slopes on high and steep mountain and in valleys near glacier termini. Glaciers undercut headwalls and valley walls, making slopes steep and unstable, especially so after glaciers have thinned and (or) retreated up valley, leaving lateral moraines on steep slopes. Some moraines and debris-covered slopes have cores of slowly melting ice and rocks that often slide down from these unstable slopes (Sigurðsson and Williams, 1991).

Jökulhlaups (glacier outburst floods) are the abrupt release of water stored under, within, or adjacent to glaciers (see also p. A266–A273 of this Part for additional discussion of jökulhlaup hazard). The Icelandic term "jökulhlaup" is sometimes restricted to the classic outburst of water bodies formed beneath ice caps due to volcanic heating from below or to eruptions through the ice but is more often applied to glacier outburst floods in general, whatever the cause. The classic jökulhlaups from Vatnajökull and Mýrdalsjökull ice caps in Iceland are discussed in greater detail in the "Glaciological Hazards: Glacial Lakes and Jökulhlaups" section of the "Glaciers of Iceland," (Chapter D) in this volume. A recent, dramatic jökulhlaup from Vatnajökull has been discussed; discharge reached 50,000 $m^3\ s^{-1}$ for several hours (Haraldsson, 1997; Björnsson, 1998, 2004; Jóhannesson, 2002). Volcanic or geothermal heating may even cause jökulhlaups from steep volcanos with highly crevassed glaciers, such as Mount Rainier, Wash. (Driedger and Fountain, 1989).

Ice-dammed lakes can form where a glacier blocks a valley, acting as a dam for the water. When the water reaches a certain height behind the dam, water will begin to create passages through or under the ice dam. The water flow increases by melting the channel walls; the larger the channel, the greater the flow. This process may occur rapidly, resulting in a glacier outburst flood. Glacier-dammed lakes form in various situations as Post and Mayo (1971) describe in detail in relation to glacier outburst floods in Alaska. Glacier-dammed lakes are numerous there, and many glacier outburst floods are related to glacier surges, earthquakes, avalanches, landslides, and other processes (Post and Mayo, 1971).

An inventory of glacial lakes (considering all lakes above 3,500 m above sea level, not necessarily dammed at present) has recently being compiled for Nepal and Bhutan (Mool, Bajracharya, and Johsi, 2001; Mool, Wangda, and others, 2001). In Nepal, 2,323 glacial lakes with an aggregate area of 75 km^2

have been inventoried out of the total of 3,252 glaciers in Nepal that cover an area of 5,323 km^2. In Bhutan, 2,674 lakes with an aggregate area of 107 km^2 out of 677 glaciers in Bhutan with an area of 1,317 km^2 have been inventoried (Mool, Wangda, and others, 2001). In both countries, glacial-lake outburst flood events (12 noted in Nepal up to 2001) have caused extensive damage to roads, bridges, villages, and infrastructure as well as loss of human lives.

In the Zailiyskiy Alatau and Kungey-Ala-Too ranges, western Tien Shan, about 110 lakes were inventoried with the aggregate volume estimated at about 5 to 10×10^6 m^3 of water (Kotlyakov, 1997). These lakes have always been the main sources of damaging mudflows for Kazakhstan mountain regions. The number of glacial lakes, not always dammed, in the high mountains of Asia may exceed 100,000. Their possible impact on irrigation, as well as their being a source of hazards, have not yet been assessed thoroughly.

Conclusions and Research Directions

Conclusions

Because the amount of glacier ice in the world varies with time, it affects many other components of the Earth System. It is very important to synthesize results from <u>all</u> glaciers and ice caps (in addition to the two huge ice sheets), including the glaciers peripheral to the ice sheets in Greenland and Antarctica. The global area of these glaciers is about 785,000 km^2, with an estimated volume of about 250,000 km^3; these values are somewhat larger than previous estimates. Although this area is small compared to that of the ice sheets, the small glaciers and ice caps cycle mass through their system rapidly and have significant effects on Earth System processes. Annual mass balances of these glaciers have been predominately negative since scientific observations began. However, the rate of ice wastage in the most recent decade has almost doubled. This freshwater addition to the ocean affects its circulation, especially in the Arctic Ocean; it also affects the gravity field of Earth, perhaps accounting for the partial slowing of its rotation rate observed since 1998. Seasonal-mass balance observations are especially useful; they show a slight increase in the winter balance at high elevations and a more negative summer balance at lower elevations. These seasonal balances give insight into precipitation and other climate variables in the high mountains and high latitudes, where meteorological observatories and climate records are few. Increased runoff from glaciers has affected water supplies, agriculture, and the frequency of natural hazards such as floods and debris flows in high-mountain regions. But as the small glaciers have disappeared, the ice reservoirs have become depleted and vital summer runoff reduced.

Global effects on the environment are now better known, and many of them have been quantified because of the recent application of new technologies. Human society has also become more sensitive to these global changes because of growing populations, especially in previously unpopulated regions. Changes in glacier regime link to changes in climate and the hydrologic cycle through energy and water balances, but glaciers transform the resulting balances through changes on their own surfaces, usually in complex and not necessarily predictable ways. Therefore, after half of a century of scientific observations, we conclude that no alternative method of study can yet replace direct observations on glaciers.

Research Directions

1. Continued direct mass-balance observations on all previously chosen "benchmark" glaciers, because long records are essential and irreplaceable. Additions to this network are desirable, especially in Alaska, the Russian Arctic, the Patagonian ice fields, New Zealand, and the high mountains in Central Asia (in particular, in the Pamirs, Karakorum, Himalaya, and Tibet). New observational networks on sub-Antarctic islands and on individual glaciers bordering and noncontiguous with the Greenland and Antarctic ice sheets are especially needed.

2. Development of long-term observational networks on large glaciers that are currently underrepresented in observational networks, such as those in Alaska (for example, Bering or Malaspina Glaciers), Asia (for example, *Fedtchenko Glacier* in the Pamirs) and Patagonia (for example, in the Southern Patagonian Ice Field). Individual large glaciers in the vicinity of the Greenland and Antarctic ice sheets, including their floating parts, are especially important for monitoring.

3. A shift to studies of the regime and changes of glacier systems, in addition to those of individual glaciers. Large, complex valley glaciers and ice caps, some of which include hundreds of outlet glaciers, are becoming more scientifically important entities for study, even though they are more complex. New methodologies for integrating results over larger spatial scales are needed.

4. Enhanced applications of satellite images in glacier research, such as the Global Land Ice Measurements from Space (GLIMS) consortium [*http://www.glims.org/*]. (see also section on "Global Land Ice Measurements from Space (GLIMS)" in this chapter.) A list of highly recommended parameters for direct observations could be expanded to include quantitative measurements applicable to climate-change studies, such as altitudes of transient snowlines (TSLs) and of equilibrium-line altitudes (ELAs) and seasonal mass-balance components. The modern concept of worldwide glacier monitoring has the objective of integrating in situ measurements with remotely sensed data and numerical modeling (Bishop and others, 2004).

5. Inclusion of timely quantitative data obtained from satellites, as well as ground observations, on Web sites such as those operated by the World Glacier Monitoring Service (WGMS) [*http://www.geo.unizh.ch/wgms*], the National Snow and Ice Data Center (NSIDC) [*http://nsidc.org*], NASA's National Snow and Ice Data Center Distributed Active Archive Center (NSIDC DAAC) [*http://www.earth.nasa.gov/data/daac/nsidc_daac.html*], and other glaciological institutions so that these results will be available quickly to the broad scientific community.

Monitoring Changes in Length, Area, and Mass Volume of Glaciers

Introduction

In 1894 the International Glacier Commission was founded in Zürich, Switzerland, during the 6th International Geological Congress (Radok, 1997; Haeberli, 1998). Thus began an international program to systematically measure fluctuations in the termini of glaciers and to produce accurate topographic maps of glaciers. The monitoring and mapping of changes in the area and volume of glaciers have evolved from direct field measurements, including the compilation of maps using plane table surveying methods, and, since the late 1940s, measuring changes in mass balance and more recently by using data acquired by sensors carried on aircraft and satellites. Aerial photogrammetric methods revolutionized the production of accurate topographic maps, including maps of glaciers, but ground surveys were still required to establish geodetic control for the stereopairs of vertical aerial photographs. Remote sensing technologies on satellites combined with geographic information system (GIS) methods have extended the monitoring of changes in glaciers from the local and regional scale to the global scale.

The original—and continuing—impetus to measure and to monitor areal and volumetric changes in glaciers, especially regarding the response of glaciers to changes in climate, arose in the disinterested pursuit of the geological sciences. It involved research by glaciologists, geologists, hydrologists, physical geographers, and geophysicists from many nations. To advance glaciological knowledge, the International Glaciological Society was established in Cambridge, England, in 1936 (Wood, 1986). Today, the IGS publishes two journals, "Journal of Glaciology" and "Annals of Glaciology," and occasional books. The Iceland Glaciological Society was founded in 1950 and publishes the journal "Jökull" ["Glacier"].

In Switzerland, Norway, Iceland, Canada, and in other nations with glacierized drainage basins, where the development of hydroelectric power was given high priority, state funding of glaciology and hydroglaciological research and the establishment of institutions staffed by glaciologists, hydrologists, and other scientists also received a high priority. Other nations that had the national resources to support a broad range of scientific research and a strong network of internationally known universities, such as the United Kingdom, Sweden, Denmark, France, United States, and Russia, also supported glaciological research. Antarctic treaty nations that have been active in Antarctica since the International Geophysical Year (1958–1959) have strongly supported glaciological research; Antarctica has been an important "training ground" for generations of glaciologists from these and other nations. During the International Geophysical Year three World Data Centres for Glaciology were established: in the United States in Boulder, Colorado (the predecessor to the current National Snow and Ice Data Center (NSIDC)); in Cambridge, England, U.K., at Scott Polar Research Institute; and in Moscow, Soviet Union (now Russia) at the Institute of Geography. A fourth World Data Center was later added in Lanzhou, China, in what is now the Cold and Arid Regions Environmental and Engineering Reseach Institute.

The international center for coordinating the compilation and archiving of datasets of glaciers worldwide is the World Glacier Monitoring Service (WGMS), established in 1986 and located in Zürich, Switzerland. WGMS merged the Permanent Service on the Fluctuations of Glaciers, established in 1967, and the World Glacier Inventory (WGI), established in 1976 (Müller and Scherler, 1980), into a unified institution (Haeberli and others, 1989a). Glaciologists from many nations established international standards for measuring fluctuations of glaciers (Kasser, 1967; United Nations Educational, Scientific, and Cultural Organization [UNESCO], 1969), measuring changes in mass balance, and preparing inventories of glaciers (United Nations Educational, Scientific, and Cultural Organization [UNESCO] 1970a; Müller and others, 1977; Müller, 1978, International Association of Hydrological Sciences, 1980; Scherler, 1983).

At the same time that Swiss and other glaciologists were developing international standards for measuring fluctuations of glaciers and for preparing glacier inventories, the U.S. glaciologist Mark F. Meier was Chair of the Working Group of the International Commission of Snow and Ice (ICSI) of the International Association of Scientific Hydrology (IASH). Under Meier's aegis, the Working Group produced "A Guide for Compilation and Assemblage of Data for Glacier Mass Balance Measurements." United Nations Educational, Scientific, and Cultural Organization (UNESCO) published the Working Group's report "Combined Heat, Ice, and Water Balances at Selected Glacier Basins" (1970b). Meier (1962) had previously published definitions for terms to be used for calculating the mass budget of a glacier. The Norwegian glaciologist Gunnar Østrem and Alan D. Stanley, both of whom were working in Canada, published a field manual for mass-balance measurements of glaciers; Østrem and Stanley (1966) was subsequently revised in a second edition (Østrem and Stanley, 1969), and a third edition appeared two decades later (Østrem and Brugman, 1991).

Early work on the preparation of glacier inventories was also done in Canada by Ommanney (1969a, b, 1970) on Axel Heiberg Island. Stanley (1970) and Müller (1970) carried out pilot studies for glacier inventories in the Rocky Mountains and Mount Everest region, respectively. Müller was affiliated with McGill University, Montréal, and later with the Swiss Federal Institute of Technology, and Director of the Permanent Service on the Fluctuations of Glaciers (PSFG). All three pilot studies are included in (United Nations Educational, Scientific, and Cultural Organization (UNESCO) (1970a) "Perennial Ice and Snow Masses. A Guide for Compilation and Assemblage of Data for a World Inventory."

WGMS has published nine volumes in the Fluctuations of Glaciers series. The first volume was by Kasser (1967), covering the five-year period 1959–1965. Successive five-year periods through volume 9 (2000–2005) cover 45 years of observations of changes in glaciers in many glacierized regions of the Earth. In addition to these internationally valued volumes, work has continued on the World Glacier Inventory (Haeberli, Bösch, and others, 1989).

During the International Geophysical Year in 1957, the U.S. glaciologist William O. Field, Director of the American Geographical Society, established the World Data Center (WDC-A) for Glaciology in Boulder, Colo., as a data center devoted to archiving glaciological information. From 1971 to 1976, the USGS operated the center in Tacoma, Washington, under the management of the USGS glaciologist Mark F. Meier. The WDC for Glaciology became part of NOAA's Environmental Data and Information Service in 1976 and, under the management of Roger G. Barry, also became affiliated with the University of Colorado. In 1982, NOAA established the National Snow and Ice Data Center [*http://nsidc.org*] as the primary U.S. data center for archiving all aspects of research on the Earth's cryosphere and for distributing the archived data. Various satellites and international programs acquired glaciological data: the Climate and Cryosphere (CliC) Program, the Global Energy and Water Cycle Experiment (GEWEX), the Global Climate Observing System (GCOS) (including the Global Terrestrial Observing System (GTOS) [*http://www.fao.org/gtos/*], and the Global Terrestrial Network for Glaciers (GTN-G) [*http://www.fao.org/gtos/gt-netGLA.html*], under the World Climate Research Programme (WCRP). The WGMS is the manager of GTN-G, within the GCOS/GTOS. During the 10 years since its launch on 21 November 2000, the Advanced Spaceborne Thermal Emission and reflection Radiometer (ASTER) sensor on the Terra Satellite has acquired medium-resolution (15-m pixels) multispectral images, some in stereo, of the Earth's glaciers. Zemp and others (2008, p. 11) state the Global Terrestrial Network's goal: "GTN-G aims to combine (a) field observations with remotely sensed data, (b) process understanding with global coverage, and (c) traditional measurements with new technologies by using an integrated and multi-level monitoring strategy." The ASTER data of glaciers are analyzed and archived at NSIDC (and are archived at other data centers); analyses of the data are also archived and reported to the WGMS. The scope of the Landsat MSS, RBV, TM, and ETM+ images of glaciers was extended and complemented by the addition of ASTER images starting in late 2000. Acquisition of Landsat 1 MSS images of glaciers began in July 1972 and still continues, with partly functioning Landsat 5 and Landsat 7 still (2012) in operation. A comprehensive review of Global Land Ice Measurments from Space (GLIMS), by Bruce H. Raup and Jeffrey S. Kargel, appears as a separate section on p. A247–A260, in this part.

In addition to the GLIMS consortium and the various programs operating under the umbrella of the World Climate Research Programme, other projects are also directed at mapping glaciers from a global perspective. GlobGlacier [*http://www.globglacier.ch/*], a project with the Earth Space Agency's Living Planet Programme, is directed at fulfilling data gaps in the databases at NSIDC and WGMS (Paul, Kääb, and others, 2009b).

Without the compilation and archiving of long-term datasets on key variables in the cryosphere of the Earth System, it would not be possible to document changes in such variables. The WGMS [*http://www.wgms.ch*], its predecessor institutions, and the NSIDC continue to maintain databases on changes in non-ice-sheet glaciers. For some glaciers, these data extend from the end of the Little Ice Age, or for more than 100 years in many glacierized regions. Zemp and others (2008, p. 54) noted that "from the end of the Little Ice Age [non-ice-sheet] glaciers around the globe have been shrinking significantly, with strong glacier retreats in the 1940s, stable or growing conditions around the 1970s, and again increasing rates of ice loss since the mid 1980s. The early mass-balance measurements indicate strong ice losses as early as the 1940s and 1950s, followed by a moderate ice loss between 1966 and 1985, and accelerating ice losses until [the] present." The WGMS continues its leadership in archiving datasets obtained from various sources that monitor glaciers worldwide in the 21st century (Haeberli and others, 1998), now joined by the glaciological databases maintained at the NSIDC [*http://www.nsidc.org*].

The database on changes in and the inventories of non-ice-sheet glaciers that WGMS and NSIDC maintain have been invaluable sources of data for relating changes in climate to variations in glaciers by Swiss and other glaciologists (for example, Haeberli, Müller, and others, 1989; Haeberli, 1995; Haeberli and Hoelzle, 1995; Haeberli and others, 2002, 2007; Zemp and others, 2005) and other scientists (for example, Oerlemans, 1994, 2001, 2005; Bahr, 1997; Bahr and others, 1997; Dyurgerov and Meier, 1997a, b).

The primary focus of the WGMS, its predecessor institutions, and NSIDC is to measure changes in termini and in mass balance and to compile inventories of mountain glaciers, the most numerous type of glacier on Earth, estimated as totaling 160,000 in all geographic regions (Meier and Bahr, 1996; Bahr and others, 1997; Meier, 1998a). Bulletin No. 10 (2006–2007) of the "Glacier Mass Balance Bulletin," is the latest compilation to be published by the WGMS (Haeberli, and others, 2009); Bulletin No. 1 (1988–1989) was the first to be published in 1991. A few outlet glaciers from ice caps and ice fields are also included in the WGMS database. The estimated 70 ice caps on Earth (Meier and Bahr, 1996; Bahr and others, 1997; Meier, 1998a, b), especially the larger ones, were not generally included. Nor was much attention directed at the two ice sheets, the Antarctic ice sheet and the Greenland ice sheet. Both ice sheets are composites of a still-undetermined large number of individual glaciers: outlet, ice stream, and ice shelf. The reason for the lack of attention to the two ice sheets was that such ice masses were far too large for glaciologists to monitor changes in area and volume using the equipment then available. The advent of satellite remote sensing technology has banished that impediment.

The achievements of the WGMS (and of the GLIMS consortium) are authoritative. Directing their efforts at non-ice-sheet glaciers, both institutions emphasize measuring annual changes in glacier termini and in mass balance, measuring changes in area and volume of glaciers, producing maps of glaciers, and making detailed inventories of glaciers. At the beginning of the 21st century, WGMS successfully met its ambitious objectives, including making its database available in digital format to all users. Its database on glaciers is also linked to databases maintained by its principal collaborator in the United States, the National Snow and Ice Data Center (NSIDC) at the University of Colorado, Boulder, Colo. (Mullins and others, 2002). NSIDC is also the headquarters of the Global Land Ice Measurements from Space (GLIMS) consortium. With respect to the glacier databases that WGMS and NSIDC (including GLIMS) archive, Zemp and others (2008, p. 54) write:

> The internationally coordinated collection of information about ongoing glacier changes since 1894 and the efforts towards the compilation of a world glacier inventory have resulted in unprecedented data sets. Several generations of glaciologists around the world have contributed their data to the present state of knowledge. For the second half of the 20th century, preliminary estimates of the global distribution of glaciers and ice caps covering some 685,000 km^2, are available, including detailed information on about 100,000 glaciers, and digital outlines for about 62,000 glaciers. The database on glacier fluctuations includes 36,240 length change observations from 1803 glaciers as far back as the late 19th century, as well as about 3,400 annual mass balance measurements from 226 glaciers covering the past six decades.

Hall, Williams, and others (2003) emphasize that knowledge about changes in glaciers in various geographic regions is important for assessing current and future water resources (for example, in the Himalaya and the Andes), glaciological hazards, and the importance of glaciers as indicators of climate change and as primary contributors to change in global sea level (see

"Sea Level Variability and Volume of Glacier Ice on Land" on p. A172–A191). The authors also point out that major gaps exist in our knowledge about changes in ice caps, ice fields, and their associated outlet glaciers and, most importantly, about changes in the Greenland and Antarctic ice sheets and their associated outlet glaciers, ice streams, and ice shelves.

If all of the non-ice-sheet glaciers were to melt, the potential maximum rise in sea level is about 0.5 m (table 3). For the Greenland ice sheet, the maximum rise in sea level would be about 7 m, and for the Antarctic ice sheet the rise would be about 70 m (bottom right graphic on pl. 1). Therefore, knowledge about changes in the Antarctic and the Greenland ice sheets is imperative in order to determine the volume of glacial meltwater that each of the two ice sheets contributes to the potential magnitude of rise in sea level during the 21st century and beyond. The only feasible way to study and monitor changes in the Greenland and Antarctic ice sheets—and the large ice caps—is to use satellite remote sensing technologies and GIS methods. Various sensors on Earth-orbiting satellites have provided glaciologists with the data to study and monitor changes in all the Earth's glaciers—non-ice-sheet and ice sheet combined. Landsat 1 MSS images, first acquired on 23 July 1972, provided the impetus for preparing this 11-chapter USGS Professional Paper 1386-A–K, "Satellite Image Atlas of Glaciers of the World" (Williams and Ferrigno, 2005). The GLIMS consortium was initially based on ASTER, a sensor on the Terra spacecraft launched on 18 December 1999 and later expanded to include image data from other satellite sensors and to incorporate GIS technology into the research (Bishop and others, 2004).

Conventional Glacier Monitoring

With support from the International Commisson on Snow and Ice (ICSI), the United Nations Education, Science and Cultural Organization (UNESCO), the United Nations Environment Programme (UNEP), the International Union of Geodesy and Geophysics (IUGG), and the International Association of Hydrological Sciences (IAHS), glaciologists from several nations assembled themselves into working groups in order to develop a set of international guidelines for measuring various parameters of glaciers in the field. The guidelines covered measurement of annual changes in glacier length (change in position of the terminus), preparation of or use of existing maps of glaciers (Gunning, 1966), measurement of mass balance of a glacier (winter, summer, and annual change) (United Nations Educational, Scientific, and Cultural Organization [UNESCO] (1969, 1970a, b), and preparation of glacier inventories (United Nations Educational, Scientific, and Cultural Organization [UNESCO], 1970a; Müller and others, 1977; Müller, 1978). Because the guidelines were prepared before Landsat was launched on 23 July 1972, it was assumed that measurements of glaciers would be based on methods that were traditional at the time: direct field measurements, maps, ground photographs, and vertical and oblique aerial photographs (either individual or stereopairs). Figure 72 shows a cross-sectional view (fig. 72*A*) and a plan view (fig. 72*B*) of a mountain glacier with some of the parameters of a glacier to be measured.

The original set of guidelines, "Variations of Existing Glaciers. A Guide to International Practices for Their Measurement" (United Nations Educational, Scientific, and Cultural Organization [UNESCO], 1969) was a product of the International Hydrological Decade (IHD) (1965–1974). The set was

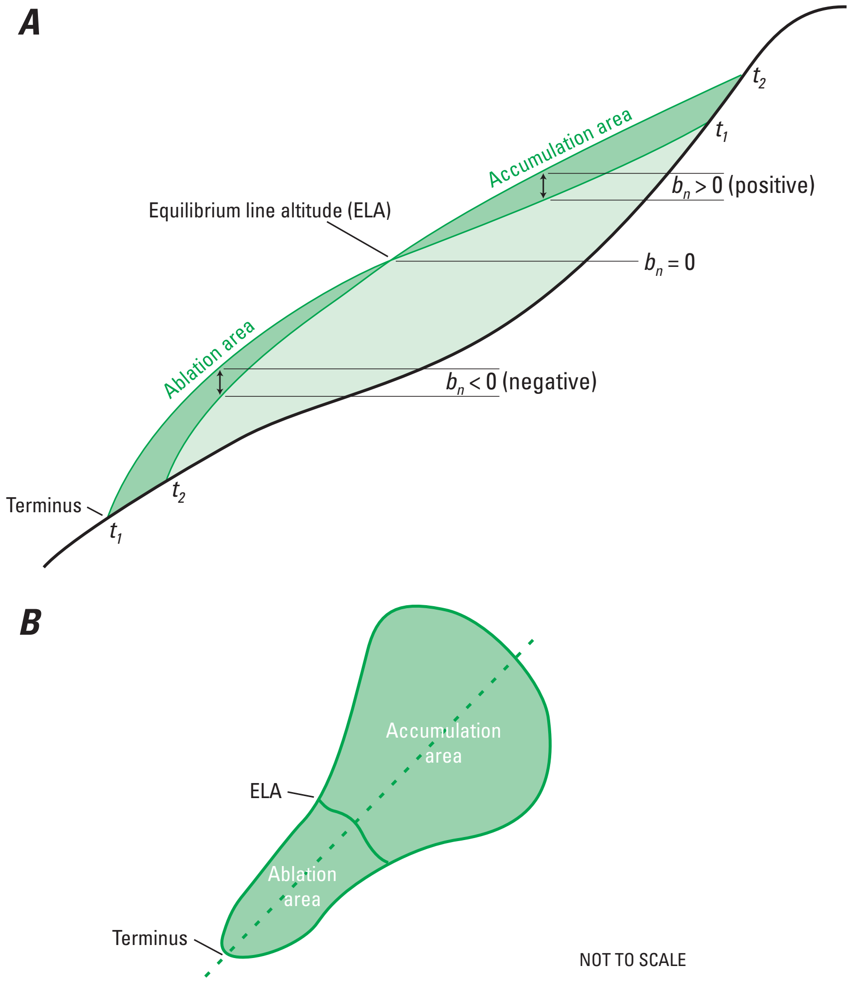


***Figure 72.**—Some glaciological parameters of a mountain glacier shown in, **A**, cross-sectional view and, **B**, plan view. The accumulation area plus ablation area represents the total glacier area. The positive net mass balance (b_n) of the accumulation area and the negative net mass balance of the ablation area are separated by the equilibrium line altitude (ELA) where net mass balance is zero (b_n=0). The accumulation area ratio (AAR) is determined by dividing the accumulation area by the total area of the glaciers. The length of a glacier is measured from the terminus along its mid-line (dashed line) to its uppermost margin. Modified from Andrews (1975, p. 33, fig. 3-1A).*

divided into three parts: I, Mountain glaciers; II, Ice sheets, ice caps and calving glaciers; and III, Data submission and publication; guidelines for basic observations, for more comprehensive measurements, and for other data needs were provided for each category.

Five years after the launch of Landsat, the Temporary Technical Secretariat for the World Glacier Inventory (WGI) published guidelines for preparing preliminary glacier inventories directly from satellite images (Scherler, 1983). A World Glacier Inventory Workshop held in Riederalp, Switzerland, in September 1978 (International Association of Hydrological Sciences, 1980) had recognized that satellite images would have to be used to conduct at least a “global inventory” of glaciers, even though necessarily a preliminary one, especially if ice caps and ice fields (and associated outlet glaciers) and the Greenland and Antarctic ice sheets were to be included (Part II of United Nations Educational, Scientific, and Cultural Organization [UNESCO], 1969).

Accurately measuring a glacier and extracting reliable information about its area, volume, topography, thickness, the position of its terminus, its mass balance, and other glaciologically important parameters, even for a small mountain glacier (such as Place Glacier (area <4 km^2), Coast Range, British Columbia, Canada) required gathering data from direct field observations, supplemented by ancillary data collected or compiled by others or by other instruments, such as maps, aerial photographs, satellite images, radio-echo-sounding surveys, and many other sources and instruments. For larger glaciers, such as ice caps and ice sheets, other techniques and technologies must be employed.

Glaciologists can use remote-sensing and GIS technologies for measuring five glaciological parameters: area and length, changes in surface elevation, velocity, and changes in volume. The measurement of mass balance in the field, the glacier parameter that provides a direct correlation with climate change, is labor-intensive and costly. Their cost and the requirement that such measurements be continued for decades explain why so few of the Earth's glaciers have even short-term records of mass-balance, let alone records for a decade or more. Zemp and others (2008) cite 3,400 annual mass balance measurements for 226 glaciers since the first records were begun by Valter Schytt at Storglaciären, Sweden, in 1945, although prototype mass-balance measurements were begun on the Aletschgletscher, Switzerland, in 1927 (United Nations Educational, Scientific, and Cultural Organization [UNESCO], 1970a) and even earlier, in the early 20th century, at point locations on the Silvrettagletscher and Aletschgletscher in the Swiss Alps (Frank Paul, written commun., 2010). Dyurgerov and Meier (1997a, p. 379) state, "There are only about 40 glaciers with continuous mass balance measurements for more than 20 yr [years], but more than 100 with 1 to 5 yr [years] of mass balance records." For example, at the present time (2012), the USGS continues to carry out annual mass-balance measurements on one glacier in Washington (South Cascade Glacier) and two in Alaska (Gulkana and Wolverine Glaciers).

Parameters such as slope, elevations, and changes in surface elevation can be determined from accurate topographic maps or digital-elevation models (DEMs) compiled from other sources (for example, ASTER stereoscopic images). Calculation of glacier area (fig. 72*B*) may be difficult because of snow pack covering the glacier margin, especially at higher elevations. A cover of morainic debris may obscure the location of the terminus (figs. 72*A*, *B*); it is difficult to distinguish whether visible glacier ice, draped with morainic debris, is still part of or detached from the glacier proper. Unless a glacier is advancing or has a calving front or wall in a lake, fjord, or other marine environment, accurately determining the position of a glacier terminus on aerial photographs or satellite images presents a major problem.

Unless a glacier map is compiled under the aegis of a glaciologist, most maps have deficiencies that degrade their utility for extracting accurate parameters for glaciers, such as area, length, and elevation of the surface. To compound these difficulties, there is virtually no modern map coverage of the glacierized terrain of Alaska, the most glacierized state in the United States. Instead of modern maps at a scale of 1:25,000 (as is the case with the other 49 States and all territories of the United States), for example, the existing maps at a scale of 1:250,000, 1:63,360, and 1:62,500 do not have adequate map detail, nor can they provide topographic accuracy for detailed glacier studies. In the past, most topographic maps were compiled using aerial photogrammetric methods with plotters that use stereoscopic pairs of vertical aerial photographs acquired at high solar elevation angle (for example, late June in the Northern Hemisphere), not the optimum time to map glaciers, because residual snow pack masks glacier margins. Two other related problems relate to snow cover: (1) aerial photogrammetrists and cartographers are not able to differentiate snow pack from glaciers, so snow cover is often shown as "glacier" and (2) the white "floating dot" used to determine topographic contours in a stereoplotter, such as the PG-2, does not work in the accumulation area of a glacier because of the “white” on “white” problem (lack of contrast). Therefore, all topographic elevations in the upper part of a glacier are inaccurate, and therefore topographic contours are not reliable (National Committee on Climate and Glaciers, 1991). Therefore, it may be impossible to measure slope, elevation of the surface of a glacier, and changes in surface elevation, except in cases where the glacier data were collected by geodetic airborne laser altimeter surveys or were from the Shuttle Radar Topography Mission (SRTM) data concerning a specific glacier. Geodetic airborne laser altimeter surveys have been used to successfully assess changes in volume of several glaciers in Alaska (Arendt and others, 2002); for limitations in the availability and use of SRTM data for global glaciological studies, see p. A241–A242.

Maps of Glaciers

Throughout the chapters of this volume that are devoted to specific geographic areas, special maps of glaciers have been prepared by glaciologists, working closely with photogrammetrists and cartographers in the case of maps prepared with aerial photogrammetric methods. James B. Case prepared the map of the Little Jarvis Glacier, Alaska (“Glaciers of Alaska,” Chapter K, p. K47, of this volume (Molnia, 2008)) under the auspices of the American Geographical Society (AGS). The AGS used aerial photogrammetry for its glacier mapping project (nine glaciers); it was sponsored by the U.S. National Committee for the IGY (1957–1958) (American Geographical Society (1960). In some cases, the glacier map is prepared by glaciologists with conventional field-mapping techniques (for example, plane-table mapping with a telescopic alidaide (surveyor) and a stadia rod (rodman)).

The International Hydrological Decade (IHD) (1965–1974), which was directed at problems associated with water resources problems, especially in developing countries, provided an additional catalyst. Snowmelt and glacier melt from glacierized hydrological basins needed to be accounted for in the seasonal and total annual volume of runoff, important to optimum generation of hydroelectric power and thus to the availability of water resources for agricultural, industrial, and personal uses downstream. Canada, Norway, Switzerland, and the United States were the leaders in the preparation of specialized topographic maps of glaciers, and the first three nations led in preparing glacier inventories.

Canada’s National Committee for the International Hydrological Decade convened a special workshop on glaciers in 1970. Of the 32 attendees, 29 glaciologists represented various Canadian governmental and academic institutions; one glaciologist from Sweden and two from the United States attended (Demers, 1970). Ommanney (1969b) completed a glacier inventory of Axel Heiberg Island, Northwest Territories; Ommanney (1970) used his inventory of the Steacie Ice Cap area from the inventory as one of three pilot inventories included in the United Nations Educational, Scientific, and Cultural Organization (UNESCO) (1970a) “Perennial Ice and Snow Masses.” The Ommanney (1969b) inventory of the glaciers of Axel Heiberg Island included four 1:500,000-scale Glacier Inventory maps (planimetric) in the Glacier Atlas of Canada series in a back pocket. The work by Ommanney (1969b, 1970) was done under the auspices of the Inland Waters Branch of the Department of Energy, Mines and Resources. Reid and Charbonneau (1979) published glacier surveys in Alberta; topographic maps of the Athabaska and Saskatchewan Glaciers, at a scale of 1:10,000 were included in a back pocket. Five more topographic maps of glaciers (Sentinel, Sphinx, Nadahini, Kokanee, and Bugabee Glaciers, at scales of 1:2,500 or 1:5,000) were included in the back pocket of their Glacier Surveys in British Columbia (Reid and Charbonneau, 1980). The two Reid and Charbonneau (1979, 1980) publications were prepared under the auspices of the Water Resources Branch, Inland Waters Directorate, Environment Canada.

C. Simon L. Ommanney (written commun., 2009) wrote that the mapping of the Salmon Glacier and other glaciers on Axel Heiberg Island was done by the Photogrammetric Research Section of the National Research Council of Canada; this work led to the first Glacier Mapping Symposium in 1966 (Gunning, 1966). C. Simon L. Ommanney (written commun., 2009) also remarked that glaciologist Gunnar Østrem's "personal interest and drive" resulting from the extensive glacier-mapping program he spearheaded in Norway led him to organize a second "Glacier Mapping and Surveying Symposium" 20 years later (Østrem, 1986). In 1985, at the University of Iceland, Reykjavík, Iceland, the International Glaciological Society convened a Symposium of Glacier Mapping and Surveying; the proceedings volume was published as volume 8 in the Annals of Glaciology series (Østrem, 1986).

The Norwegian glaciologists Gunnar Østrem and Nils Haakensen, working closely with photogrammetrists and cartographers, prepared detailed maps of several glaciers in Norway that pioneered new cartographic standards for maps of glaciers. Thomas J. Schultz, Standards Team Leader (Maps), sent an e-mail to Richard S. Williams, Jr. on 31 October 1998. On 27 November 1998, Richard S. Williams, Jr. sent a memorandum with five recommendations in response to his e-mail that concerns the "Depiction of Glaciers on U.S. Geological Survey Maps." The five recommendations produced here were based on evaluation of 10 maps of glaciers prepared by Norwegian glaciologists under the auspices of the Norwegian Water Resources and Energy Directorate, emphasizing the "new map" of Midtre Folgefonni in that series [Østrem, 1987(?)].

1. USGS maps should depict glacier margins and true contours on the glacier in green. Crevasses and hydrography (for example, glacier streams, lakes, and others) should be in blue, as they are now. Green for glaciers follows the international convention pioneered by Norwegian glaciologists and clearly separates glaciers from [other] hydrologic features.

2. The margin of the glacier (for example, ice cap) and certainly the terminus of a glacier (for example, valley glacier, outlet glacier from ice cap or ice field, and others) should have the date of the source material used by the cartographer of photogrammetrist to delineate the margin or terminus. The collar information should stipulate the date and type of source material, so that it can be accessed by map users, if they want to analyze the original data. This is the single biggest deficiency of USGS maps: stipulation of date and type of source material used for the compilation and (or) revision of a map. It is a particular problem on maps that show "Revised" but don't show what was revised (for example, glaciers, hydrography, culture, contours, and others) and the date and type of source material used for the revision.

3. Do not show “form lines” on a glacier; they are very misleading. Contours on the glacier should be real; in most cases, such contours are limited to the lower parts of the glacier, where sufficient contrast on the ice permits the photogrammetrist or cartographic technician to accurately plot contours. It is virtually impossible to accurately plot contours on the snow-covered up-glacier area because of lack of contrast.

4. If a revised map depicts a different position of the terminus, date the new terminus. You can also include the position of the prior terminus and date it. Do not show “form lines” on the glacier. It’s generally too costly to put new contours on the glacier, so just leave it blank (devoid of contours). If the prior terminus was down valley from the present terminus, it’s appropriate to use a green tint for where the glacier was. Conversely, an advancing glacier could also have a green tint in the area between the former glacier terminus and the new one. As noted, both termini positions should carry the date of the source material.

5. For smaller-scale maps, glaciers should be shown in a solid green tint. For larger-scale maps, glacier margins should be in green, either solid (clearly visible), small dashes (snow-covered margins), or large dashes (estimated position).

Mapping the Antarctic ice sheet required the development of a specialized set of cartographic symbols to portray glacier ice features inland and along its complex and dynamic cryospheric coast. The Working Group on Geodesy and Cartography of the Scientific Committee on the Antarctic Research (SCAR) (1980) stipulated standard symbols to be used on maps of Antarctica. With a few modifications, these standards were used on the USGS I-2600 map series, “Coastal-Change and Glaciological Maps of Antarctica” (Williams, and others, 1995; Ferrigno and others, 2002; Ferrigno and others, 2008) (see fig. 18).

Although maps of glaciers are still prepared by terrestrial surveying methods and aerial photogrammetric techniques, such conventional techniques have been superceded or complemented by modern surveying methods. On-the-ground surveys use laser-surveying methods and data from the global positioning system (GPS) of satellites; the latter provides precise latitude and longitude (x, y) and elevation (z). Airborne photogrammetry uses digital images linked to GPS; geodetic airborne laser altimetry survey (single line or scanning) is also linked to GPS. Satellite imaging is two dimensional or three dimensional, as with the specially processed stereoimages from ASTER. Elevation data of many glaciers can be acquired with laser altimeters (for example, ICESat’s GLAS) and radar altimetry (for example, SRTM, with its two radars on the Space Shuttle) from Earth-orbiting platforms. All of the digital data on glaciers—whether image, laser-altimeter, radar , or other source, whether acquired during ground, airborne, or satellite surveys—are compiled into maps by using Geographic Information System (GIS) technologies (for example, ArcGIS) and visualized as digital elevation models (DEMs) and other GIS visualizations. The maps include precise latitude, longitude, and elevation contours. Defining the terminus and outline of a glacier, however, still confronts two problems: (1) masking of the glacier margin by snowpack and (2) morainic debris in the lower part of the glacier in which the spectral reflectance of the surficial debris is identical to that of the surrounding terrain.

History of the Compilation of Glacier Inventories

A major goal of glaciologists has been to inventory all of the Earth's glaciers, from the smallest mountain glaciers to the two large ice sheets. Meier (2003a) briefly summarized the history of the preparation of glacier inventories. Glaciers attracted the attention of natural historians in several countries, one of whom was the late 18th century Icelander Sveinn Pálsson, who wrote a comprehensive treatise on glaciers, "Icelandic Ice Mountains" (Pálsson, 2004 [1795]), with accompanying maps and other illustrations. The international effort to monitor glaciers began in 1894 following the International Geological Congress. Until the International Geophysical Year (1957–1959), emphasis was placed on recording fluctuations of glacier termini, an effort that still continues. However, the IGY fostered the measurement of changes in mass balance of glaciers because of the relationship of change in glacier volume to changes in climate but also because of the need for the compilation of glacier inventories (Annals of the IGY, 1959). Even before the IGY recommendation, the Comité Spécial de l'Année Géophysique Internationale (CSAGI) suggested that glacierized nations do a "census" of their glaciers (International Union of Geodesy and Geophysics [IUGG], 1956). The World Glacier Monitoring Service (WGMS) in Zürich, Switzerland, is the international center for collecting, archiving, and making available to the glaciological community long-term datasets on fluctuations of glaciers, country by country, including (1) measurements of changes in position of glacier termini (advance or retreat) and mass balance and thickness and (2) glacier inventories. Meier (2003a, p. 87) remarks that "WGMS has joined with NSIDC [National Snow and Ice Data Center, Boulder, Colo., U.S.A.] to publish [in print and web-accessible formats] digital fluctuation and mass balance data and to digitize the glacier inventory data." Meier (2003a, p. 87) further noted, "Now the challenge—and the opportunity— is to incorporate, standardize, and determine the accuracy of remote sensing information." Cogley (2009) proposed a more complete World Glacier Inventory, which he calls World Glacier Inventory-Extended Format (WGI-XF), nearly doubling the number of glacier records in the current inventory (approximately 70,000) to more than 131,000 glacier records. Paul, Barry, and others (2009) recommended using digital data to compile glacier inventories.

Ohmura (2009), in his review of the status of the World Glacier Inventory (WGI), stated that 46 percent of the estimated area of non-ice-sheet glaciers on Earth has been inventoried and that about 95,000 glaciers out of an estimated total of 207,000 still need to be inventoried, most of them being in Alaska, Canada, and South America. If inventories made under the Global Land Ice Measurements from Space (GLIMS) consortium are combined with WGI data, 54 percent have been inventoried. Ohmura's 2009 total of 95,000 is probably a low estimate. There are an estimated 100,000+ glaciers in Alaska alone, only 600 of which have been named officially (Molnia, 2008).

According to the International Association of Hydrological Sciences (1980), glacier inventories can be preliminary (Scherler, 1983; Williams, 1986b) or comprehensive (United Nations Educational, Scientific and Cultural Organization [UNESCO], 1970a; Müller and others, 1977; Müller, 1978; Ommanney, 2002b, p. J83–J110; Paul, 2007; Cogley, 2009; Ohmura, 2009). Jiskoot (2003) prepared a map showing the status of inventories of glaciers in the 41 nations and other geographic entities (for example, Antarctica) that currently (2012) have glaciers (fig. 73). A table accompanying the map listed the nations with complete inventories (26), including Norden (Norway and Sweden), Alpine countries (Switzerland (Kääb and others, 2002), Austria, Italy, and France), Russia, China (Shi Yafeng and others, 2008, 2009), and other countries with only a small number of glaciers (for example, México and Kenya). Countries with no inventories (3) include Antarctica. Nations with incomplete inventories (12) include the United States, Canada, Argentina, Chile, Greenland, and Iceland, all having significant glacierized areas. In Iceland, however, a full preliminary inventory was completed in 2008 (Sigurðsson and Williams, 2008). C. Simon L. Ommanney describes his preliminary inventory of glaciers of Canada in the “Glacier Atlas of Canada” series in “Mapping Canada’s Glaciers” in Chapter J of this volume (Ommanney, 2002b, p. J83–J110; see also Ommanney, 1969a, b, 1970, 1980). In the United States, preliminary inventories have been completed of the Brooks Range, Alaska (1,001 glaciers, 723 km^2) (Post and Meier, 1980; Brown, 1989), the Sierra Nevada, California (Raub and others, 1980), and maps of glaciers in the western United States (Fountain and others, 2007) have been made. Although Iran, Turkey, and Antarctica are shown as having no glacier inventories, in Chapter G of this volume (1386), “Glaciers of the Middle East and Africa,”

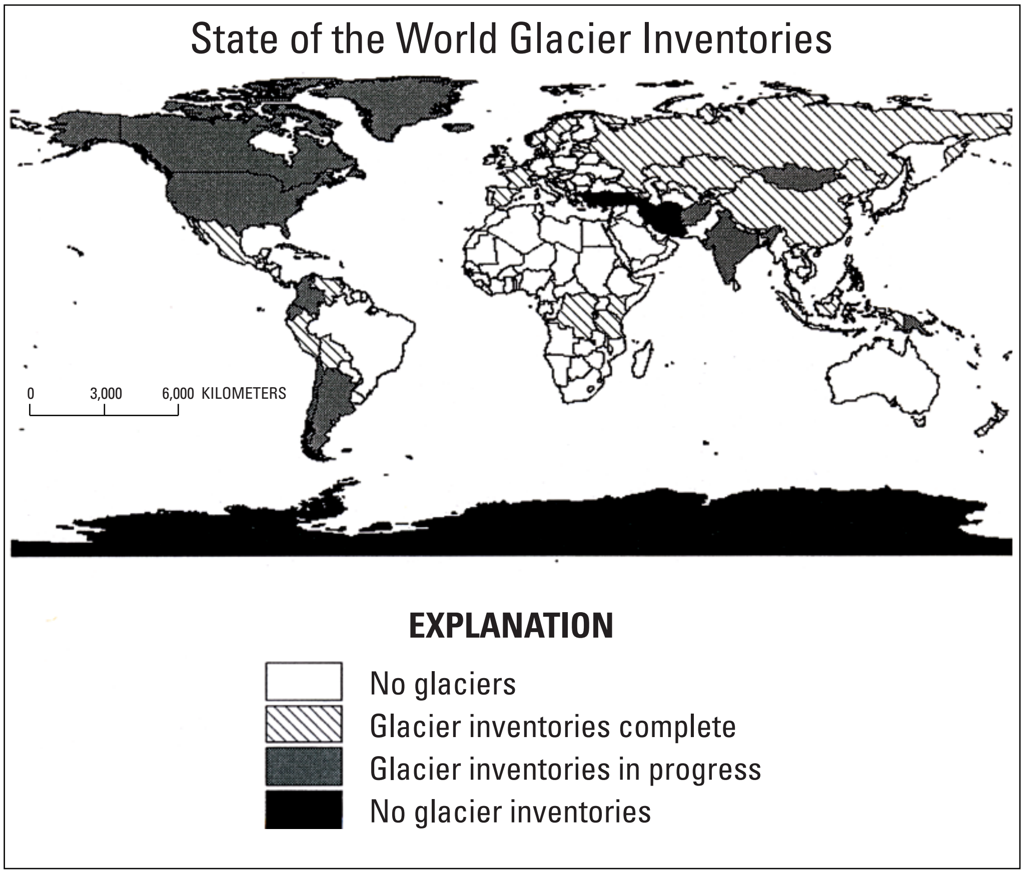


***Figure 73.**—The status in 2003 of the compilation of inventories of glaciers in the 41 nations and other geographic entities (for example, Antarctica) that are currently glacierized (have glaciers). Modified from Jiskoot (2003, figure on p. 97).*

Kurter (1991) prepared a preliminary inventory of the "Glaciers of Turkey," and Ferrigno (1991) prepared a preliminary inventory of the "Glaciers of Iran". Moussavi and others (2009) published a new glacier inventory of Iran that showed 27.12 km^2 of glacier area in five geographic locations.

As part of the USGS effort to map changes in the cryospheric coast of Antarctica, the "Coastal-Change and Glaciological Maps of Antarctica" series of maps, the accompanying booklet for each map in the I-2600 map series contains a section on "Glacier Inventory" and tables that provide an inventory of named and unnamed (identified by the WGMS regional prefix (AN7), followed by the latitude and longitude (degrees and minutes)) glaciers and other glaciological features. One of the objectives in the mapping effort was to compile a "Preliminary Inventory of the Named and Unnamed Glaciers of Antarctica," map by map (see, for example, Ferrigno and others, 2006, for a preliminary inventory of the Trinity Peninsula area and of the South Shetland Islands, Antarctica). Swithinbank (1980b, 1985) had previously addressed the problem of compiling an inventory of the glaciers that comprise the Antarctic ice sheet. With the other principal authors of the I-2600 map series, he was able to begin to compile a preliminary glacier inventory of Antarctica.

C. Simon L. Ommanney (written commun., 2009) noted that a number of scientists have completed inventories of glaciers on the Antarctic Peninsula, in the South Shetland Islands, in East Antarctica, in West Antarctica, and on the subantarctic islands. Rabassa and others (1982) completed an inventory of glaciers on James Ross and Vega Islands (see also Ferrigno and others, 2006). Inventories have also been done of the Antarctic Peninsula and King George Island in the South Shetland Islands (Braun and others, 2001; see also Ferrigno and others, 2006); in the Wohlthat Mountains, Queen Maud Land, East Antarctica (Srivastava and others, 1988), and mountain glaciers in the northern Prince Charles Mountains, East Antarctica (Krebs and Mabin, 1997); mountain glaciers in northern Victoria Land, Antarctica (Diolaiuti and others, 2003); and glaciers in the subantarctic (Donoghue and Allison, 2008). Numerous subglacial lakes in Antarctica have been discovered that have also been inventoried (Siegert and others, 2005).

Mass Balance of Glaciers

Conventional methods for determining the mass balance of a specific glacier have been well developed by glaciologists since the pioneering glaciological work by three Swedish glaciologists: Hans Wilhelmsson Ahlmann (1953), C.C. Wallén, and Valter Schytt, who applied the concept to Storglaciären, Kebnekaise, Sweden, in 1945 (Holmlund and Jansson, 1999). Since the pioneering research of the Swedish glaciologists, mass-balance measurements have been made of 226 glaciers during various spans of time. The mass-balance record for Storglaciären is still the longest series for an entire glacier of all of the Earth's glaciers. This glaciological parameter of mass balance is considered to be the most important variable, because the relationship is so direct for assessing the relationship of changes in mass balance of a glacier to changes in climate. At the beginning of the 21st century, the mass balance status of glaciers worldwide, especially the mass balance of the Greenland and Antarctic ice sheets (Rignot and Thomas, 2002), is of central concern because prolonged negative mass balances of glaciers will cause water in the global hydrologic cycle, which is sequestered in glacier ice on land, to be transferred to the oceans, resulting in a rise in global sea level. Remote sensing technologies are the only feasible way for determining changes in mass balance of the Greenland and the Antarctic ice sheets (Bamber and Payne, 2004); see also the earlier section on "Sea Level Variability and Volume of Glacier Ice on Land," (p. A172–A191) and the next section on "Selective Studies on Remote Sensing of Glaciers" (p. A239–A245).

For a schematic diagram of the relationship between input and output for an ice sheet and ice cap and for a valley glacier, see figure 74 (see also fig. 13). At the present time, 22 countries continue to carry out mass-balance measurements on glaciers within their national domains.

The original field manuals for making mass-balance measurements on a glacier (Østrem and Stanley, 1966) and the later revision (Østrem and Stanley, 1969) were updated by Østrem and Brugman (1991). The North American Committee on Climate and Glaciers (Fountain and others, 1991) also presented a synopsis of mass-balance standards for glaciers. Committee members were from the United States (USGS) and Canada (Environment Canada and the University of Toronto). Paterson (1994) and Hooke (2005)

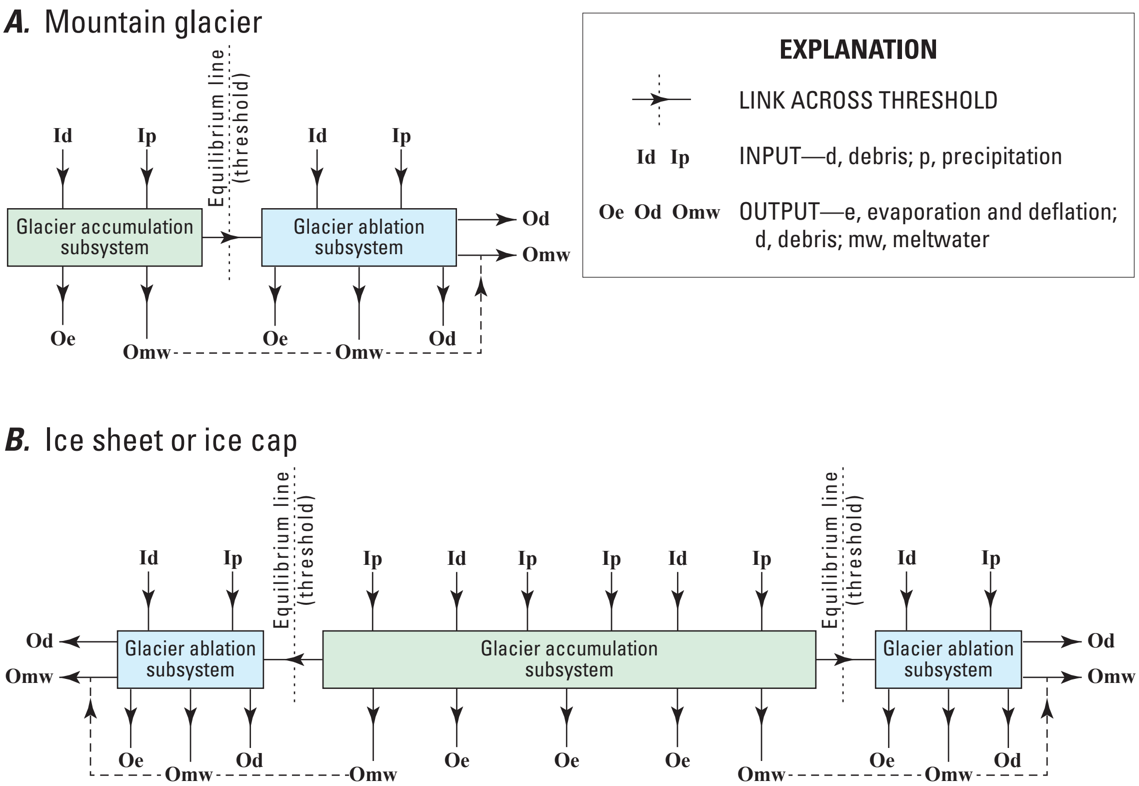


Figure 74.**—Schematic diagrams of input (accumulation) and output of mass (ablation),* **A, mountain glaciers and,* **B***, for ice caps and ice sheets. The accumulation and ablation areas of mountain glaciers, A, tend to be approximately similar, except for metastable tidewater glaciers. In a steady-state or in an increase-in-mass state, ice caps (miniature ice sheets) and ice sheets have much larger accumulation areas than ablation areas, which are toroidal in shape because of their parabolic bilateral symmetry in cross section (see fig. 4B). Under conditions of climate warming, an increase in elevation (altitude) of the equilibrium line (ELA) reverses this relationship of accumulation and ablation areas (for example, the toroidal ablation area is larger and the circular accumulation area is smaller). If the ELA rises above the surface of the ice cap or ice sheet, the accumulation area no longer exists, and the ice cap or ice sheet will rapidly disappear (melt away). Modified from Sugden and John (1976, p. 6, fig. 1.5). See also figure 13, which shows the sensitivity of both types of glacier systems to change in elevation (altitude) of the equilibrium line.*

address the theory behind mass-balance measurements. Fountain and Vecchia (1994) and Fountain and others (1997) addressed strategies for measuring mass balance. Geografiska Annaler (1984, 1999) published two issues of the journal specifically devoted to the measurement of mass balance of glaciers. Dyurgerov (2002) addressed glacier mass balance and regime.

The mass balance (b) of a glacier is the difference between accumulation and ablation (figs. 72, 74; see also fig. 13) and is determined by the amount of snow gained during the winter (b_w) and the loss of snow and ice during the summer (b_s) in volume of water equivalent. Net balance (b_n) is the mass balance at the end of the hydrologic (balance) year (end of ablation). Some glaciers are measured at a fixed date each year at the end of the melt (ablation) season; this is referred to as the annual balance. Measuring a glacier's mass balance (b) for the winter (b_w), summer (b_s), and net balance for the year (b_n), is extremely costly in terms of labor (measurement of snow/firn density in cores, survey of height of the array of stakes on the snow and ice surface), transportation to and from glaciers, and support for the field team, often for several weeks. Other methods have been used, such as measuring changes of surface elevation on successive maps prepared by aerial photogrammetric methods (Haakensen, 1986; Østrem and Tvede, 1986; see also map of Ålfotbreen in Geografiska Annaler, 1999) or plane-table mapping and by airborne (or space-borne) laser altimetry surveys. But Frank Paul (written commun., 2010), sums up the common agreement among glaciologists:

> It might be important to stress here that the geodetic determination of the mass balance (DEM differencing) is much more than just another method to determine mass balance. Actually it is THE method to determine the overall mass change of a glacier and calibrate the field-based measurements determined at individual stakes. The stake measurements provide the process understanding and the high temporal resolution, but without the frequent (one per decade) geodetic calibration large systematic errors can accumulate that might reveal a completely wrong picture of the mass change over time…

Satellite photogrammetry, such as using sequential stereoscopic ASTER images of glaciers, permits differencing between digital-elevation models (DEMs) to calculate overall changes in mass balance (changes in volume combined with estimates of the water equivalent value of change in volume of snow, firn, and ice). Shuttle Radar Topography Mission (SRTM) data can also be used for comparison with pre-SRTM and post-SRTM maps of glaciers, whether compiled by conventional mapping or from other sources. Geodetic airborne laser altimetry surveys of Midtre Lovénbreen, a valley glacier in Svalbard, provided geodetic ground-control points for the preparation of a DEM of the glacier compiled by aerial photogrammetric methods (Barrand and others, 2009). Surface-elevation, volume, and other changes in the Bering Glacier System were calculated using images from ASTER, MODIS, ERS-1 and ERS-2 synthetic aperture radar (SAR), geodetic airborne laser altimetry surveys, DEMs produced by airborne interferometric SAR (InSAR), SRTM data, DEMs from ASTER images, laser altimetry from ICESat's GLAS, and DEMs compiled from conventional USGS maps and Natural Resources Canada maps (produced by aerial photogrammetric methods). Another example of the use of data from ground-based, aerial, and satellite sources to compile accurate maps of a glacier is the Drangajökull ice cap in northwestern Iceland. The earliest map of Drangajökull was made in 1913–1914, using plane table-surveying techniques. The Defense Mapping Agency, in cooperation with the National Land Survey of Iceland, used unidentified source material to compile an unpublished 1:50,000-scale topographic map (manuscript Map

1517II) of Drangajökull, although the accuracy of the margin of the ice cap is suspect. On 19–20 April 2005, multiple traverses of the ice cap were made on snowmobiles equipped with GPS devices; a fixed GPS station was operated nearby. Using GIS technology, an accurate topographic map of Drangajökull was compiled that will be used for comparison with the previous two maps. The area of Drangajökull was 204 km^2 in 1913–1914; on 16 August 2004 its area was 146 km^2, a reduction in area of about 28 percent in 90 years. An attempt to use satellite laser altimetry (GLAS on ICESat) to measure elevation of profiles across the ice cap was not successful because the terrain was too steep and cirrus clouds were probably present (Shuman, Sigurðsson, and others, 2009).

Because no density of snow and firn is measured, an estimated value is given to obtain the water equivalent of change in volume. Østrem (1975) endeavored to monitor mass balance on previously studied and mapped Norwegian glaciers by analyzing changes in the position of the equilibrium line altitude (ELA) on Landsat images. Aniya and others (1996) later applied the same concept to their inventory of outlet glaciers of the Southern Patagonian Ice Field, South America. Changes in volume (mass balance) of ice sheets, seasonally and longer, can be measured from data acquired by the tandem GRACE satellites and DEM differencing.

Selective Studies on Remote Sensing of Glaciers

Soon after the launch of ERTS-1 (Landsat 1) on 23 July 1972, images of the glacierized regions of the Earth began to be archived at the USGS EROS Data Center in Sioux Falls, S. Dak., U.S.A., and at other data centers. For the first time, glaciologists had access to images of all non-ice-sheet and ice-sheet glaciers equatorward of about 81° North and South latitudes. Acquiring images that are cloud-free and when snow cover is minimal (end of mass-balance year) of all glaciers within these latitudes could require a decade of acquiring images on approximately a monthly basis during the late summer and early fall (in the Northern Hemisphere). For nearly 31 years, successive Landsats (1–5 and 7), with improved sensors (MSS, RBV, TM, and ETM+), in visible and infrared wavelength bands of the electromagnetic spectrum and with medium spatial resolution, have acquired new images of the Earth's glacierized regions (Williams and Ferrigno, 1991a), which have been archived and made available to glaciologists and other scientists measuring and monitoring changes on the Earth's surface as an important component of the U.S. Global Change Research Program (Williams, 1991). Although Landsat 1 and 2 recorded RBV images with a pixel resolution of 80 m (same as the MSS image), Landsat 3 RBV images have a pixel resolution of 30 m, providing much higher spatial resolution of images of glaciers (Ferrigno and others, 1983; Williams, 1983b, fig. 31–185, p. 1,849). The Landsat series of spacecraft, with its imaging sensors, was later joined by other Earth-orbiting spacecraft that provided data of glaciers in different spectral bands (for example, RADARSAT and ERS-1 images, with SAR sensors), with higher spatial resolution and stereoscopy (for example, ASTER images) and non-imaging data (for example, GLAS on ICESat), gravity instruments on GRACE, and data from the Shuttle Radar Topography Mission (SRTM). All of these datasets from various sensors on Earth-orbiting satellites revolutionized the study of glaciers by providing a variety of data (imaging and non-imaging) that enable glaciologists to analyze changes in glaciers on a global basis. In particular, they provided the capability of studying changes in larger glaciers, such as ice caps and associated outlet glaciers; the outlet glaciers, ice streams, and ice shelves of the Antarctic ice sheet; and the outlet glaciers and ice streams of the Greenland ice sheet. In addition, image data of virtually all the Earth's glaciers allowed us to fulfill the long-held goal of compiling a global inventory of glaciers.

Glaciologists quickly seized the opportunity to analyze ERTS 1 and 2 (Landsat 1 and 2) MSS and RBV images of glaciers. Could areas of glaciers be determined, could changes in area and length be measured, and could topographic ice divides between outlet glaciers be identified on ice caps and ice sheets? What other glaciological features and parameters, such as surface velocity, mass balance, and location of subglacial volcanoes, could be determined? Could a preliminary global inventory of glaciers be compiled from satellite image data? Publications by Krimmel and Meier (1975) and Østrem (1975) began to answer these questions.

Meier (1979) wrote the first technical booklet on "Remote Sensing of Snow and Ice." Hall and Martinec (1985) published the first book on "Remote Sensing of Ice and Snow." The National Research Council (1989) addressed the "Prospects and Concerns for Satellite Remote Sensing of Snow and Ice." Twenty years later, Rees's (2005) book on "Remote Sensing of Snow and Ice" was published. In the same year, Kääb (2005b) published a technical monograph on "Remote Sensing of Mountain Glaciers and Permafrost Creep." Cao, Meisheng, and others (2006) published "Remote Sensing of the Cryosphere" in Chinese. The most recent book on the subject, "Remote Sensing of Glaciers," was edited by Pellikka and Rees (2009).

The International Glaciological Society (IGS) became the international focal point for publishing scientific papers on the analysis of image and non-image data from the remote sensing of glaciers. The IGS sponsored the first Symposium on Remote Sensing in Glaciology, in Cambridge, England, U.K., in 1974; papers were published in a special 1,482-page issue of the "Journal of Glaciology" in 1975 (v. 15, no. 73) (Glen and others, 1975). Three more IGS-sponsored symposia on the same topic followed and were published in volumes of the "Annals of Glaciology" (v. 9, v. 17, and v. 34). The second symposium was also held in Cambridge, England, U.K., in 1986 (Williams, 1987a); the third in Boulder, Colo., in 1992 (Steffen, 1993); and the fourth in College Park, Maryland, in 2001 (Winther and Solberg, 2002). Another topically related symposium sponsored by the IGS, "Role of the Cryosphere in Global Change," was held in Columbus, Ohio, in 1994; papers were published in v. 21 of "Annals of Glaciology" (Rothrock, 1995). Knight (2006) and Slaymaker and Kelly (2007) wrote and edited books on the same topic more than two decades after the book by Hall and Martinec (1985).

The glaciers of Iceland became a focus for long-term research with satellite remote sensing technologies, including a series of papers published within two years of the launch of ERTS-1 (Landsat 1) on the Vatnajökull ice cap (Williams and Thorarinsson, 1974; Williams and others, 1974; Thorarinsson and others, 1974; Williams, 1987b). The USGS published in 1976 a collection of short papers that provided examples of environmental information obtained from analysis of ERTS images, including mapping changes in Antarctica (Williams, and Carter, 1976). By the late 1970s, the USGS had initiated a project to publish a multi-volume "Satellite Image Atlas of Glaciers" in its Professional Paper series (Ferrigno and Williams, 1980), which became the 11-volume USGS Professional Paper 1386-A–K, "Satellite Image Atlas of Glaciers of the World" (Williams and Ferrigno, 2005), involving in the international endeavor 113 glaciologists worldwide as its authors. Meier (1980) published a review paper on remote sensing of snow and ice, followed by a series of papers (Williams, 1983a, b, 1985; Swithinbank, 1985; Hall, 1988), book chapters (Williams, 1983a, 1986a), and books (Hall and Martinec, 1985; National Research Council, 1989).

During the 1990s and the early years of the 21st century, satellite images were used to measure changes in the termini of mountain glaciers (Hall, Bayr, and others, 2003; Hall, Scharfen, and others, 2003; Hall, Williams, and others, 2003) and outlet glaciers from ice caps (Hall and others, 1992). Mapping glacier facies on Vatnajökull, Iceland (Williams and others, 1991), was attempted, although an error was made, incorrectly delineating a "slush zone" on Landsat 4 and 5 TM band 4 images that was not present. Success was achieved in comparing satellite-derived measurements with ground-based measurements of the fluctuations of outlet glaciers of Vatnajökull (Williams, Hall, and others, 1997). Williams and Hall (1993) contributed a chapter on glaciers in the "Atlas of Earth Observations Related to Global Change."

Instruments on satellites other than Landsat were also used to study the polar regions (Massom, 1991). In a later volume on "Polar Remote Sensing," Massom and Lubin (2006) focused on the Greenland and Antarctic ice sheets, as greater emphasis was increasingly being directed at changes in the two ice sheets. ERS-1 synthetic aperture radar (SAR) imagery examined properties and ice dynamics of the Greenland ice sheet (Fahnestock and others, 1993). Interferometric SAR (InSAR) was used to monitor the velocity of the Rutford Ice Stream, Antarctica (Goldstein and others, 1993). Satellite radar altimetry was used to delineate the topography of the Amery Ice Shelf, Antarctica (Brooks and others, 1983). Bindschadler and others (1989) mapped the surface topography of the Greenland ice sheet with satellite radar altimetry. Digital elevation models, derived from the Shuttle Radar Topography Mission were compared with preexisting maps of the Northern and Southern Patagonian Ice Fields to estimate changes in volume of the largest 63 outlet glaciers (Rignot and others, 2003).

The Shuttle Radar Topography Mission (SRTM) acquired elevation data of the Earth's land areas between lat 60° N. and lat 54° S. during an 11-day mission between 11 and 22 February 2000 by the Space Shuttle (Space Transportation System (STS)-99). Lead agencies of the U.S. Government involved in the project were a civilian agency, the National Aeronautics and Space Administration (NASA), and a U.S. Department of Defense (DOD) agency, the U.S. National Geospatial-Intelligence Agency (NGA). Unclassified SRTM data are archived at and distributed by the EROS Data Center of the USGS [*http://dds.cr.usgs.gov/srtm/*].

For global glaciological studies, SRTM data have two major deficiencies. First, no glacier in the polar regions, including the Greenland and Antarctic ice sheets, falls within the range of data coverage. Therefore, more than 97 percent of glacier area and more than 99 percent of glacier volume cannot be studied using this source. Second, for regions outside the territory of the United States, the U.S. National Geospatial-Intelligence Agency (NGA) of the U.S. Department of Defense (DOD) restricted civilian sampling of SRTM data to 90-m spatial resolution (equal to the pixel resolution of Landsat MSS images and equivalent to Level 3 Digital Terrain Elevation Data (DTED) used in the compilation of 1:250,000-scale maps); for the United States, 30 m-pixel data are available. From a global perspective, therefore, these two deficiencies of SRTM data mean that for glaciers in the polar regions, where most of the volume and area of the Earth's glaciers are situated, and for glaciers located outside the United States in the regions between lat 60° N. and lat 54° S., no data are available. Many mountain glaciers cannot be accurately measured.

Echelmeyer and others (1996) carried out airborne laser altimetry surveys of glaciers in Alaska. Aðalgeirsdóttir and others (1998) used airborne laser altimetry to analyze changes in elevation and volume of the Harding Icefield and associated outlet glaciers, Kenai Mountains, Alaska. Sapiano and others (1998) analyzed 13 changes in elevation, volume, and position of terminus for 9 glaciers. The most comprehensive survey to date was the airborne laser altimetry carried out for 67 glaciers in Alaska to measure changes in thickness compared with data from preexisting maps (Arendt and others, 2002). Garvin and Williams (1993) carried out geodetic airborne laser altimeter (GALA) surveys of three outlet glaciers—two in Iceland (Breiðamerkurjökull and Skeiðarárjökull) and one in Greenland (Jakobshavn Isbræ). Shuman, Hall, and others (2006) and Shuman, Sigurðsson, and others (2009) compared ICESat GLAS profiles of Drangajökull with ground GPS-generated profiles. Krabill and others (1995) have carried out repeated airborne laser altimeter surveys across the Greenland ice sheet to measure changes in thickness. Knoll and Kerschner (2009) used data from an airborne laser scanner to make a glacier inventory for the South Tyrol region of Italy. A digital-elevation model of the Hofsjökull ice cap, Iceland, was produced from InSAR data (Barton and others, 1999); other remote sensing techniques, including SAR, InSAR, and Landsat data, to measure decadal-scale changes in Hofsjökull were also evaluated (Hall and others, 2000).

By the end of the 20th century, two workshops had been convened to assess the current and future applications of remotely sensed and other data to monitor changes in glaciers. A "Workshop on the Long-Term Monitoring of Glaciers of North America and Northwestern Europe" was convened by the USGS in Tacoma, Washington, U.S.A., in 1996 (Williams and Ferrigno, 1997). Williams and others (1997) addressed the use of various satellite sensors to measure changes in the area and volume of ice caps and ice sheets. Williams and Hall (1998) contributed a similar paper to a WGMS-sponsored Expert Meeting (Workshop) held in Zürich, Switzerland, that resulted in the publication of "Into the Second Century of Worldwide Glacier Monitoring: Prospects and Strategies" (Haeberli and others, 1998). At the beginning of the 21st century, in the year 2002, the National Snow and Ice Data Center convened a "Workshop on Assessing Global Glacier Recession," in Boulder, Colo., U.S.A. (Casey, 2003). The results of the workshop were summarized by Barry (2003).

Also, at the beginning of the 21st century, instruments on newly launched Earth-orbiting satellites began to acquire data on glaciers that provided important new information to glaciologists. One of the instruments, Advanced Spaceborne Thermal Emission and reflection Radiometer (ASTER) on Terra provided image data in several spectral bands (fig. 75*A*), including stereoscopic images (fig. 75*B*), that permitted the compilation of a more robust inventory of glaciological parameters for the Earth's non-ice-sheet glaciers (Kääb and others, 2003; Bishop and others, 2004) (see "Global Land Ice Measurements from Space (GLIMS)," by Bruce H. Raup and Jeffrey S. Kargel (following this section on p. A247–A260)). The spectral bands recorded by the Landsat Thematic Mapper could be used to discriminate between fresh snow, firn, glacier ice, and debris-covered glacier ice in the visible and near-infrared bands (fig. 76). ASTER covered twice as many spectral bands in the visible and near-infrared, short-wavelength infrared, and thermal infrared (fig. 75*A*).

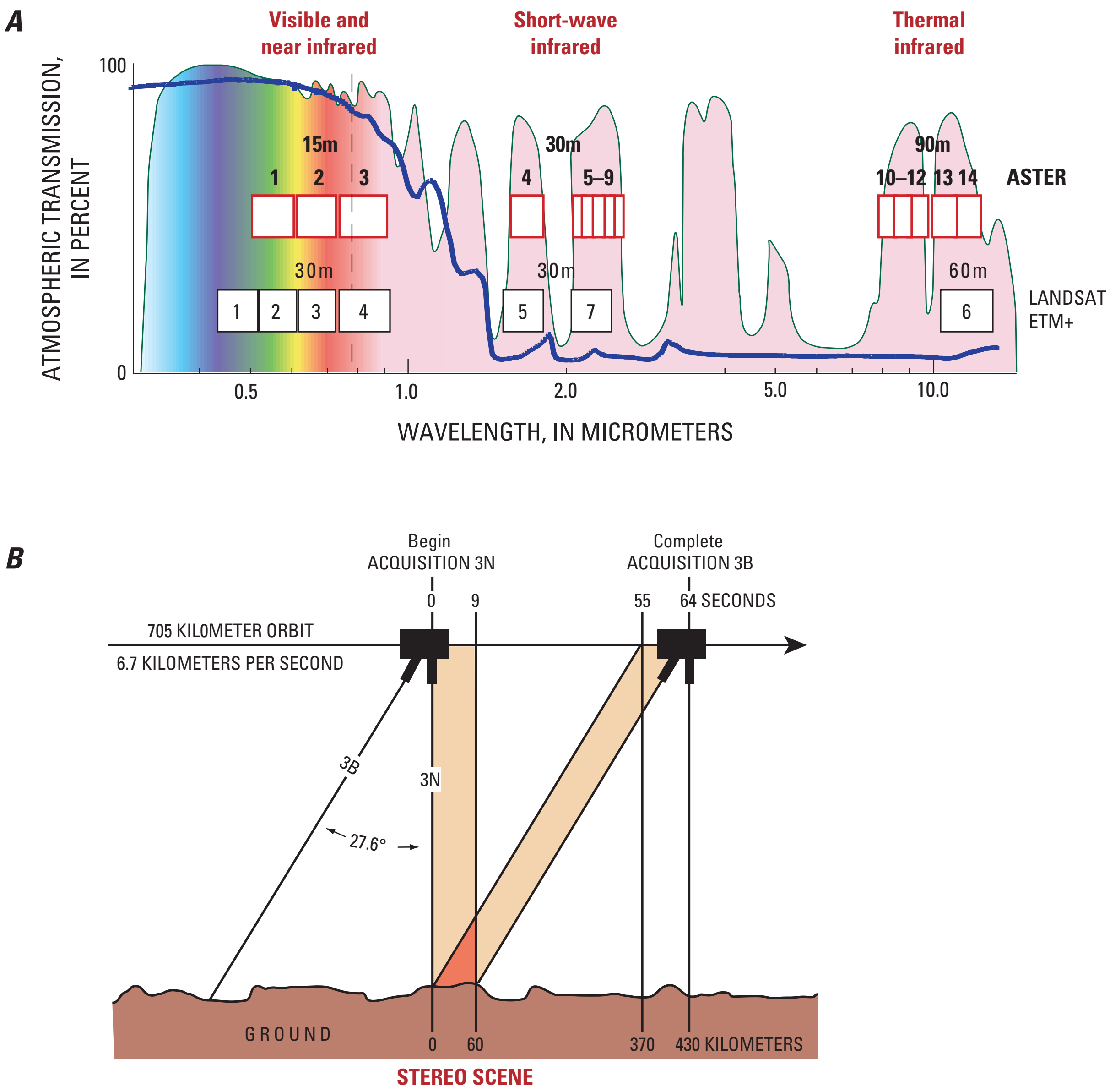


Figure 75.—**A**, *Advanced Spaceborne Thermal Emission and Reflection Radiometer (ASTER) spectral bands, for comparison with Landsat Enhanced Thematic Mapper (ETM+). The rectangular boxes (red: ASTER; black: ETM+) indicate the sensor channels, with their respective spatial resolution indicated above the boxes. The superimposed colored curve represents the atmospheric transmission in percent dependency on wavelength. The vertical dashed line marks the approximate margin of long-wavelength visible light. Abbreviations for bands of the spectrum: VNIR (visible and near infrared), SWIR (short-wave infrared), and TIR (thermal infrared).* **B**, *ASTER stereo geometry and timing of the nadir-band 3N and the back-looking sensor 3B. An ASTER nadir scene, approximately 60 km in length, and a correspondent scene looking back by 27.6° off-nadir angle and acquired about 60 seconds later together form a stereo scene. Modified from Kääb and others (2003, p. 44, figs. 1, 2).*

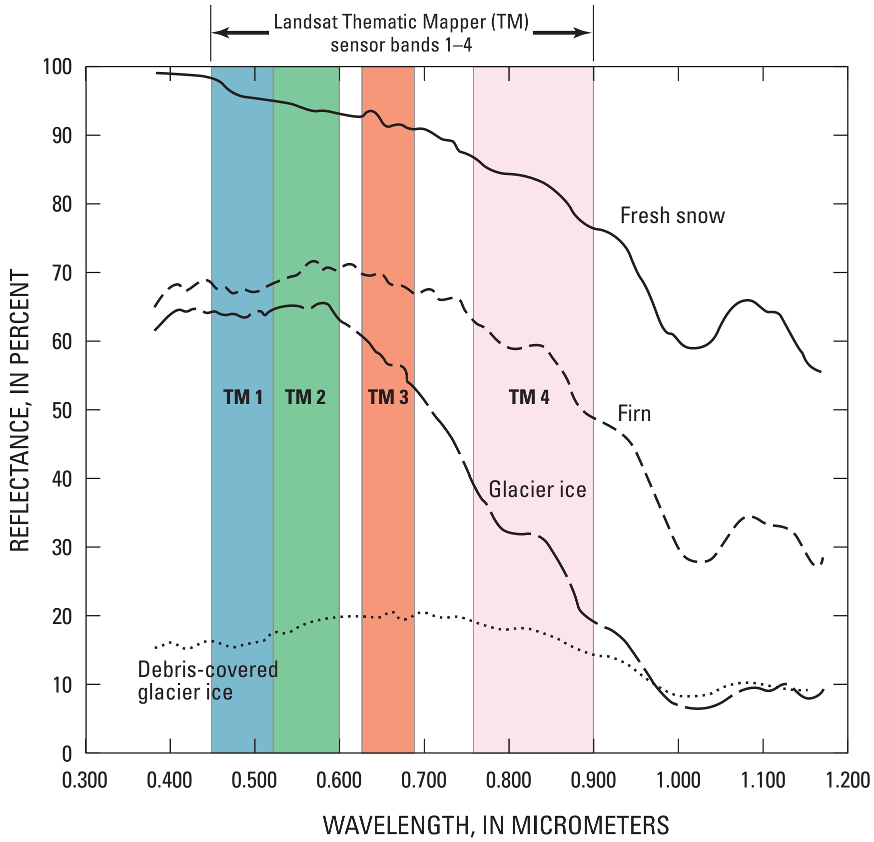


Figure 76.*—Fresh (newly fallen) snow, firn (denser than snow but not yet ice), bare glacier ice, and debris-covered glacier ice discriminated on four spectral bands (1–4) recorded by the Landsat Thematic Mapper (TM) sensor. Modified from Hall and others (1989, p. 28, fig. 7); see also Hall and others (1988, p. 317, fig. 4).*

Swiss glaciologists have combined analyses of satellite images from GLIMS (Landsat, SPOT, and ASTER) with geographic information systems (GIS) technology to compile sophisticated inventories of the glaciers of Switzerland (Paul and others, 2002; Paul, Barry, and others, 2009; Kääb and others, 2002; Rau and others, 2007). Paul (2007) published his doctoral dissertation, “The New Swiss Glacier Inventory 2000—Application of Remote Sensing and GIS,” an outstanding “model” of a glacier inventory compiled from remote sensing and GIS technologies. In 2007, Paul, Maisch, and others (2007) modeled the future areal extent of glaciers in the Swiss Alps using topographic-elevation data. Kääb (2005), and Kääb and others (2005) also reviewed the variety of remote sensors that has been used to study and analyze mountain glaciers and their importance for monitoring in the future (Paul, Kääb, and others, 2007). Racoviteanu and others (2009) discussed some of the challenges associated with mapping glaciers with ASTER images under the GLIMS consortium. The other two satellites are the Ice, Cloud, and land Elevation Satellite ICESat (National Aeronautics and Space Administration, 2002), with its Geoscience Laser Altimeter System (GLAS), and the tandem Gravity Recovery And Climate Experiment (GRACE), both designed to measure changes in the elevation of the surface (ICESat/GLAS) and changes in the volume (GRACE) of the Greenland ice sheet (Luthcke and others, 2006) and the Antarctic ice sheet (Velicogna, 2009). An analysis of Antarctic elevation data determined from ICESat traverses of low-slope surfaces during cloud-free conditions shows a ±14 cm accuracy for the GLAS profile (Shuman, Zwally, and others, 2006). GRACE measurements of the Greenland ice sheet from 2002 to 2006 show a loss of ice of 150 to 250 $km^3\ a^{-1}$ for each of the four years (Sullivant, 2007). Hall and others (2008) compared surface temperature and melt duration from MODIS data with mass loss from GRACE data on the Greenland ice sheet (fig. 21). In fact, most glaciological studies now incorporate remotely sensed data.

Satellite Remote Sensing of Glaciers in the 21st Century

By 2013, all 11 chapters in the USGS "Satellite Image Atlas of Glaciers of the World" will have been published and available in both print and digital (online) formats. This series constitutes the first global survey using satellite images and ancillary data—a preliminary inventory—of the glaciers of the Earth, both non-ice-sheet and ice-sheet glaciers. The overlapping GLIMS consortium will, in cooperation with the WGMS and the NSIDC, continue its goal of producing a more comprehensive inventory of the Earth's glaciers using a variety of imaging sensors on satellites in combination with GIS technology. Swiss glaciologists completed an update of their complete inventory of Swiss glaciers using remote sensing and GIS methodologies (Kääb and others, 2002; Paul and others, 2002; Paul, 2007).

There is also a need to have an operational Landsat series to acquire and archive ETM+ images of the Earth's glaciers. Williams (1991) argued that the Landsat Program was the most important satellite in the U.S. Global Change Research Program for the 1990s because of its continuity of image acquisition. The National Research Council (2007, p. xiii) stated that "for over 30 years, Landsat observations have provided the best means of examining the relationship between human activities and their terrestrial environment." A fully operating ASTER instrument must be part of the instrument package on a future satellite so that imaging of the Earth's glaciers in both single and stereo-image modes can be continued. ICESat-II, with an improved GLAS sensor, is needed to provide repeat laser profiles of the Greenland and Antarctic ice sheets. Cryosat-2, with its capability for acquiring radar altimetry profiles of the ice sheets with its Synthetic Aperture Radar/Interferometric Radar Altimeter (SIRAL), was initially planned for launch in 2009[14] (Committee on Earth Observation Satellites [CEOS], 2008). Both ice sheets will draw increased attention because of their potentially large contribution to the accelerating rise in global sea level.

Periodic measurement of changes in the Earth's cryosphere and its response to global warming (Armstrong and others, 2002), especially changes in the Greenland and Antarctic ice sheets (Bamber, 2006), requires an international commitment to maintain an operational system of Earth-orbiting satellites that carry the relevant sensors for acquiring data on key glaciological parameters. Under the Programme for Monitoring of the Greenland Ice Sheet (PROMICE), scientists with the Geological Survey of Denmark and Greenland and other institutions are carrying out an ambitious effort that combines ground-based observations (for example, network of automatic weather stations), airborne remote sensing surveys (for example, airborne geodetic laser altimeter and radio-echo sounding), satellite remote sensing (for example, imaging sensors on Landsat, Advanced Spaceborne Thermal Emission and reflection Radiometer (ASTER) images on the Terra satellite part of the GLIMS consortium and GlobGlacier (Paul, Kääb, and others, 2009), and synthetic aperture radar (SAR) inferometry (InSAR) data from

[14] Editors' note: CryoSat-2 was successfully launched on 8 April 2010.

ESA's ERS-2 spacecraft (Ahlstrøm and others, 2009)). Non-imaging sensors, on satellites such as GRACE, are needed to measure changes in volume of the ice sheets. A World Meteorological Organization report (Key and others, 2007) recommend an integrated global observing strategy for the cryosphere.

The National Research Council (2008) documented the importance of Earth observations from space for the past five decades. The National Research Council (2007, p. ES02) recommended, "The U.S. government, working in concert with the private sector, academe, the public, and its international partners, should renew its investment in Earth-observing systems and restore its leadership in Earth Science and applications." These recommendations were part of an important study entitled "Earth Science and Applications from Space: National Imperatives for the Next Decade and Beyond" (National Research Council, 2007). An editorial in "Nature" (Nature, 2007, p. 761) declared "Data sets encapsulating the behavior of the Earth system are one of the greatest technological achievements of our age… [with particular relevance to the hiatus (since 2003) in acquiring Landsat images]. Landsat images documented changes on the Earth's surface. There is only one Earth, with only one history, and we get only one chance to record it…. A record not made is gone for good." The National Research Council (2007, p. 6-6) stated, "Landsat dominance in high [sic, medium] resolution environmental monitoring may be over." As of late 2009, an ETM+ sensor on Landsat 8 had still not been placed in orbit.[15] However, the USGS EROS Data Center provided free access (download) to its archive of ortho-rectified Landsat images, a return to the original altruistic mission of the ERTS (later renamed Landsat) Program: "For the Benefit of All Mankind." This major decision by the USGS provides a wealth of image data of the Earth to scientists involved in studies of global environmental change, including glaciologists. Chuvieco (2008) discussed the importance of satellite remote sensing in monitoring changes in the global environment. Witze (2007) reviewed the scope of the problem that the United States and other nations face in monitoring the environment of the Earth from space. Her review provided a table listing 26 key environmental variables—10 of the atmosphere, 6 of the oceans, and 10 of the land (including glaciers)—that need to be monitored by specialized sensors on a large variety of satellites operated by many nations.

[15] Editors' note: The ETM+ sensor for Landsat 8 has been replaced by a new instrument, the Operational Land Imager (OLI); the Landsat 8 spacecraft is called the "Landsat Data Continuity Mission," with a planned launch in December 2012. Preparation for Landsat 9 is underway, with a planned launch in December 2017.

Global Land Ice Measurements from Space (GLIMS)

By BRUCE H. RAUP[16] *and* JEFFREY S. KARGEL[17]

Introduction

Glaciers are the Earth's chief repositories of freshwater, holding 2.17 percent of Earth's total water; 97.2 percent of Earth's water is in the oceans (saline) (Williams, 1986a). In many geographic locations, glaciers are almost ideal natural reservoirs because they store water in frozen form during cold and cloudy (wet) periods, and they slowly release it during sunny (dry) periods when liquid water is most needed downstream in a hydrologic basin (Yang and Hu, 1992). However, some glaciers store and suddenly release enormous volumes of water, causing widespread damage, injuries, and loss of life downstream if those areas are populated (Richardson and Reynolds, 2000; Cenderelli and Wohl, 2001; Kääb, Wessels, and others, 2003; Kääb, 2005a, b).

It is clear that the world's climate is undergoing a rapid change to warmer conditions (Intergovernmental Panel on Climate Change [IPCC], 2007a). Because of this observed global increase in temperature, glaciers also are changing (MacCracken and others, 2001; Bush and others, 2004; Lemke and others, 2007; Meehl and others, 2007).

Glaciers respond to and integrate effects of climate change over their entire surface area, but the response can be complex. Many glaciers are changing in a linear and easily predictable correlation with climatic perturbations. Some glaciers have undergone rapid disintegration or have melted away entirely: for example, Brewster Glacier in New Zealand, Meren Glacier in Irian Jaya, Indonesia (Allison and Peterson, 1989; Kincaid and Klein, 2004); three mountain glaciers in Iceland (Sigurðsson and Williams, 2008); numerous small glaciers in Glacier National Park, Mont., U.S.A. (Key and others, 2002; Hall and Fagre, 2003); and in the Alps (Paul, Kääb, and Haberli, 2007). Feedback processes and nonlinear responses can result in unexpectedly rapid responses of some glaciers, not only small cirque and valley glaciers (Paul and others, 2004; Barry, 2006) but also large ice caps and ice sheets (Oppenheimer, 1998) and ice shelves (Scambos and others, 2000, 2004; Ferrigno and others, 2006, 2008, 2009). Furthermore, some glaciers—especially calving tidewater and surge-type glaciers—are inherently unstable and undergo episodic or cyclic advances and retreats that are more or less independent of climate change. Changes in the amount of the Earth's surface covered by glaciers, in particular the Antarctic and the Greenland ice sheets, can have a significant impact on global climate through changes in albedo, changes in the area of land relative to that of water, and other effects. Therefore, to improve current understanding of past, present, and future climate change, scientists require a much more accurate assessment of glaciers, the area of Earth they cover, and their relation to climate change.

In general, glaciers are melting worldwide, retreating (Oerlemans, 1994, 2005), and losing mass (Haeberli and others, 1999; Cameron, 2005; Kaser and others, 2006) at historically unprecedented rates. However, it is critical

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to document and understand recent glacier fluctuations in order to forecast changes and their impacts on society. The two largest remaining ice masses, the Antarctic and the Greenland ice sheets, are important because their melting would have a major impact on the rate and magnitude of rise in sea level (Gregory and Oerlemans, 1998; Bindoff and others, 2007) and because their presence is important to global climate. The reduction in mass of mountain glaciers, ice caps, and ice fields ultimately will limit the amount of water available seasonally within a hydrologic basin. Glaciers are also important indicators of past glacial conditions and past climatic fluctuations. Glaciers are a critical dynamic element of potential multinational and multi-billion-dollar regional hydrologic hazards (Micklin, 2004; Kääb, Reynolds, and Haeberli, 2005). Therefore, scientists must understand the state and the dynamics of, and impending future changes of, glaciers in these regions (Bishop and others, 2004; Kargel and others, 2005). It is also important to document the areal extent and distribution of low-latitude glaciers, especially ice caps (Williams and others, 1991; Hall and others, 2000). Many of these types of glaciers are not likely to be around several decades from now (Dyurgerov and Meier, 1997a; Thompson and others, 2002), making it all the more important that scientists extract the climatic information in ice cores withdrawn from them (See "Ice cores, high-mountain glaciers, and climate," by Thompson, in this Part on p. A157–A171) and know the geographic distribution of these low-latitude glaciers now. Some of the many important issues affected by glaciers and glacier change include global rise in sea level, water resources, landscape changes (for example, newly deglacierized terrain (glaciers no longer present), and hazards such as outburst floods (jökulhlaups) from glacier lakes (also called "glacier lake outburst floods" (GLOFs) in the Alps and Himalaya).

Glacier Remote Sensing

Glaciologists, working individually and within large, well-coordinated national and international programs and projects, have faced the daunting challenge of inventorying the distribution of the world's glaciers, characterizing their state and dynamics, and quantifying basic glaciological parameters, such as length and area. Among them are William O. Field's two-volume compilation of "Mountain Glaciers of the Northern Hemisphere" (Field, 1975) and his index maps in "Atlas of Mountain Glaciers in the Northern Hemisphere" (Field, 1958) and Mercer's "Southern Hemisphere Glacier Atlas" with index maps (Mercer, 1967); the production of the World Glacier Inventory via the World Glacier Monitoring Service (WGMS) (Haeberli, 1998); and the publication of the "World Atlas of Snow and Ice Resources" (Kotlyakov, 1997a).

Two tasks under the U.S. Geological Survey's Glacier Studies Project (GSP) [*http://www.glaciers.er.usgs.gov*] address the geographic distribution and the fluctuations of glaciers. The first task is the publication of a "Satellite Image Atlas of the Glaciers of the World" (Williams and Ferrigno, 2005; revised in 2007, 2010), an 11-volume series of USGS Professional Papers (1386–A through K) published in print and digital formats by the USGS between 1988 and 2013 (expected completion). The publications have the stated objective of using Landsat images, acquired between 1972 and 1981 to establish a global baseline of the areal extent of all of the Earth's glaciers with the pixel resolution (79 m) of the Landsat Multispectral Scanner (MSS) sensor and also using ancillary data (for example, aerial photographs, maps, previously published inventories, among other source materials). In later chapters ("volumes") of USGS Professional Paper 1386, Landsat TM and ETM+, Advanced Spaceborne Thermal Emission and reflection Radiometer (ASTER), MODerate resolution Imaging Spectroradiometer (MODIS), and other satellite images were used. Ground photographs and oblique and vertical aerial photographs of glaciers complement the satellite images. The second task is the publication (in print and (or) digital formats) of a series of "Coastal-Change and Glaciological Maps of Antarctica" (Ferrigno and others, 2002; Williams, Ferrigno, and Foley, 2005, revised 2009, 2010). For the 11-volume glacier atlas, 113 scientists from more than 25 nations analyzed Landsat images and ancillary information, comparing their Landsat-analysis results with the documented historical knowledge of changes in glacier area in their respective countries and regions. Both tasks evolved from the USGS's early recognition that satellite remote-sensing technology was the only feasible means of monitoring changes in the Earth's glaciers globally (Williams, 1985). In the 40 years since the first civil Earth Resources Technology Satellite (ERTS, later renamed "Landsat" by the National Aeronautics and Space Administration) began to image the world's glaciers in July 1972, satellite imaging of glaciers has improved dramatically in spatial resolution (for example, 1-m-pixel IKONOS® images versus 15-m-pixel Landsat 7 ETM+ panchromatic band and ASTER visible-near-infrared multispectral bands, versus the 79-m pixels of early Landsat multispectral images). Such imaging has also extended its areal coverage, frequency of coverage, and overall quality: radiometric resolution has improved, saturation over snow and ice features has been reduced, and scan lines and artifacts such as were present in old vidicon images have been reduced or eliminated.

Imaging technology from Earth-orbiting satellites continues to improve, with successive instruments tending to have higher spectral, radiometric, and spatial resolution and geometric fidelity. Spectral bandwidth and sensor response function for selected multispectral instruments are some aspects of this technological evolution (fig. 77). The experimental instruments Advanced Land Imager (ALI) (10-m-pixel resolution) and Hyperion (a hyperspectral instrument; 10-nm sampling interval (220 spectral bands) from 356 nm to 2,577 nm at 30-m-pixel resolution), both on the Earth Observing 1 (EO–1) platform of NASA's New Millennium Program, are included in figure 77 for reference. EO–1 was launched on 21 November 2000. ASTER on the NASA Earth Observing System (EOS) Terra Satellite is a multispectral instrument with a sufficient number of spectral bands to resolve the reflectance features of snow at approximately 2.2 micrometers (μm). This makes ASTER suitable for discriminating new snow from firn, bare glacier ice, dirty ice, partially vegetated and debris-contaminated ice, and even from lake water, as was the case with Landsat TM bands (Hall and others, 1988, 1989; fig. 76). Sensors on EO–1 (ALI), EO–1 (Hyperion), Landsat 4 and 5 (TM), Landsat 7 (ETM+), and MODIS on the Terra and Aqua satellites also include spectral-band coverage within all or part of the 2.1- to 2.6-μm section of the electromagnetic spectrum.

One of the important newer developments with ASTER is the addition of an aft-looking band 3 camera (fig. 75*B*). With nadir and aft imaging in that band (800 nm), ASTER enables along-track stereo imaging, the image pairs being acquired about 1 minute apart. Civilian high-resolution thermal-infrared imaging began with Landsat 3 (a single 120-m-pixel resolution band); thermal imaging also has improved in recent years, with ASTER having six thermal bands at a spatial resolution of 90-m pixels and with Landsat 7 Enhanced Thematic Mapper Plus (ETM+) having a single thermal band with a spatial

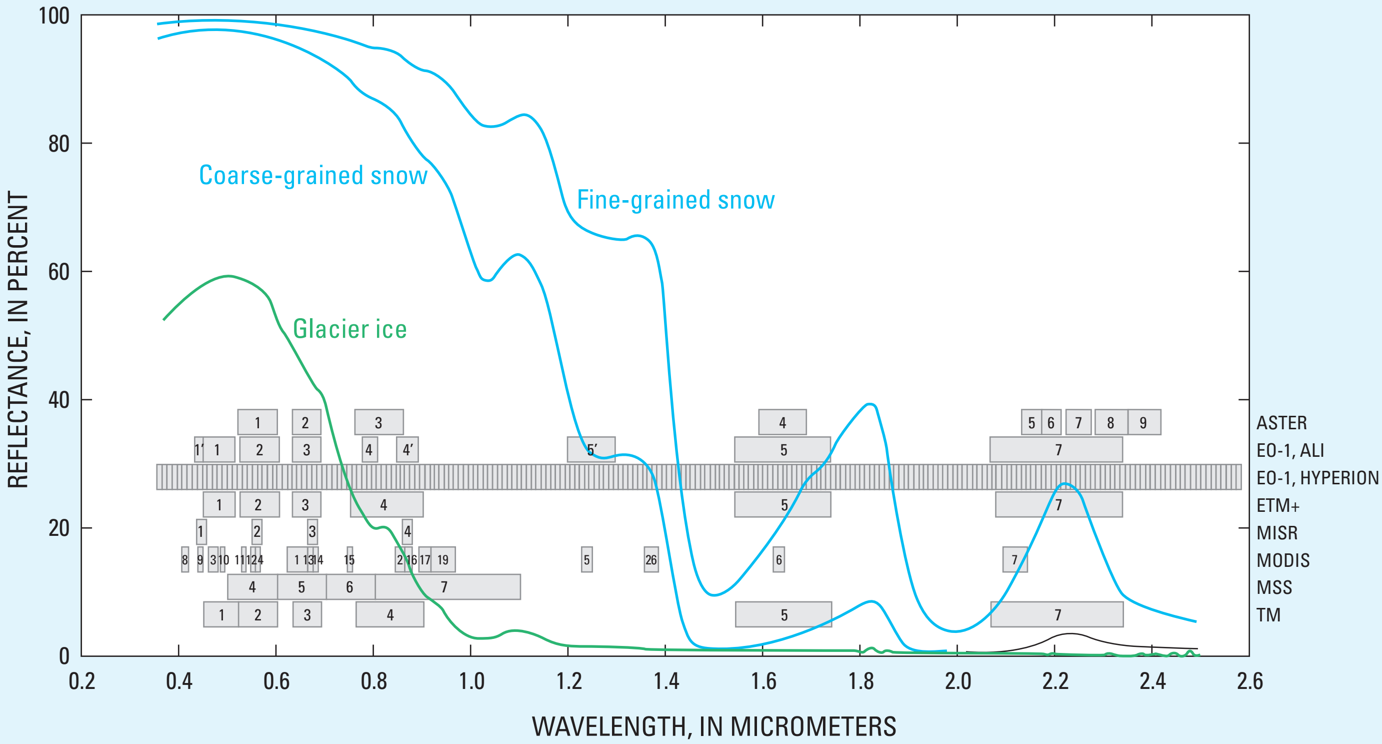

Figure 77.—*Spectral bands that selected instruments on Earth-orbiting satellites record. The bandwidths of past, present, and experimental instruments are shown against the spectral-reflectance curves of bare glacier ice, coarse-grained snow, and fine-grained snow. The numbers in the gray boxes are band numbers of the electromaganetic spectrum. The Hyperion sensor records 220 bands between 0.4 µm and 2.5 µm; in the diagram the individual bands are not numbered. Instruments are abbreviated as: ASTER, Advanced Spaceborne Thermal Emission and reflection Radiometer; ALI, Advanced Land Imager; ETM+, Enhanced Thematic Mapper Plus (Landsat 7); MISR, Multiangle Imaging SpectroRadiometer; MODIS, MODerate-resolution Imaging Spectroradiometer (36 bands, of which 19 are relevant to discrimination of snow and ice); MSS, MultiSpectral Scanner (Landsats 1–3); and TM, Thematic Mapper (Landsats 4 and 5). Several of these instruments also have bands in the thermal infrared. Landsat bands 1 through 3 generally saturate over snow (Ferrigno and Williams, 1983), but ASTER bands, with their adjustable gains, were designed to avoid the saturation problem (Raup and others, 2007).*

resolution of 60 m. Civilian panchromatic imaging has increased dramatically in pixel resolution on more recent satellites: to 15 m for Landsat 7 ETM+, 5 m or better for Satellite Pour l'Observation de la Terre (SPOT) 5, 1 m for IKONOS®, and 0.61 m panchromatic and 2.4 m multispectral for QuickBird2. The products of some of these commercial systems are expensive; therefore, the NASA policy of permitting registered users of Global Land Ice Measurements from Space (GLIMS) to have no-cost access to glacier images from ASTER, including global digital elevation models (GDEM), expedites coordinated glacier monitoring on a global scale. In addition, the USGS EROS Data Center has made images in the Landsat archive available for free downloading over the Internet. The free availability of digital elevation models (DEMs) from the SRTM mission, which covers glaciers between lat 60° N. and lat 54° S., adds to the wealth of satellite image data of glaciers available in photographic and digital formats.

The technology available for analyzing these images to determine the areal extent and dynamics of glaciers has also grown dramatically. In the early days, assessments of glacier extent, surface features (Williams and others, 1991), and ice motion were done manually; but increasingly sophisticated and accurate semiautomatic and automatic methods of glacier analysis have evolved over time to allow glaciologists to extract information difficult or impossible (and expensive) to obtain from the ground or by airborne surveys (Lucchitta and Ferguson, 1986; Scambos and others, 1992; Williams and others, 1997; Paul and others, 2002; Taschner and Ranzi, 2002; Hall, Bayr, and others, 2003; Paul, Huggel, and Kääb, 2004; Rignot, Braaten, and others, 2004; Rignot, Cassasa, and others, 2004; Kääb, 2005a, b; Kargel and others, 2005; Khromova and others, 2006; Raup and others, 2007). Technology and semiautomated applications include mapping glaciers and associated features such as proglacial and supraglacial lakes (Sidjak and Wheate, 1999; Kääb and others, 2002; Wessels and others, 2002; Wheate and others, 2002; Hall, Scharfen, and others, 2003; Khromova and others, 2003; Racoviteanu and others, 2009) (fig. 78); mapping ice-flow fields (Scambos and others,1992; Kääb, 2005a; Kääb, Lefauconnier, and Melvold, 2005) (fig. 79); preparing glacier inventories (Paul and others, 2009); measuring changes in areal extent of glaciers and mapping changes in flow speeds or strain-rate fields (Holdsworth and others, 2002; Skvarca and others, 2003; Scambos and others, 2004; Kääb, 2005a; Stearns and others, 2005); measuring acceleration of glacier retreat (Hall and others, 1992; Kääb and others, 2002; Paul and others, 2004; Ferrigno and others, 2005, 2006, 2007, 2008, 2009, 2010); and measuring changes in surface elevation and volume of glaciers (for example, the Bering Glacier system (Muskett and others, 2009)).

These improvements notwithstanding, the relatively small cadre of glaciologists dedicated to interpreting and understanding glaciological data—especially data pertaining to changes in glaciers—makes the increased efficiency afforded by technology all the more critical. It also requires organization and coordination, which can enhance efficiency and geographic coverage of glacier observations, provide for checks on accuracy and uniformity of reported data, and reduce excessive redundancy of effort. This is the context in which the GLIMS consortium [*http://glims.org/*] was developed as a team-member project for the ASTER instrument (Kieffer and others, 2000; Kargel and others, 2005).

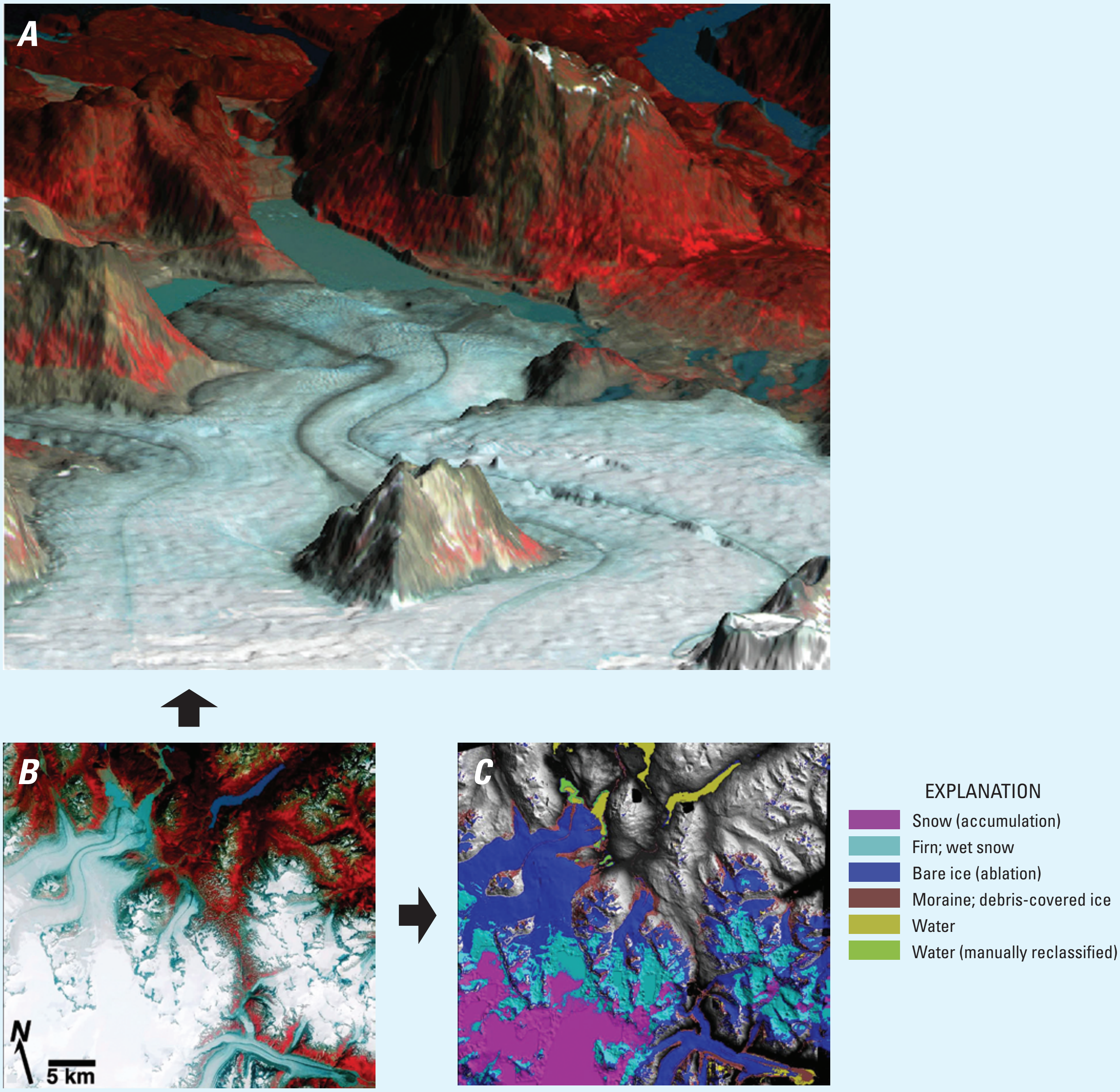

Figure 78.—**A**, *Advanced Spaceborne Thermal Emission and reflection Radiometer (ASTER) image (false color infrared composite) draped over an ASTER digital elevation model (DEM) showing the terminus of Llewellyn Glacier, northwestern British Columbia, and proglacial lake; perspective view looking to the northeast.* **B**, *Part of an ASTER scene showing Llewellyn (top left) and Tulsequah (lower right) Glaciers.* **C**, *Map showing water and glacier features and created from partial ASTER scene shown in B using an enhanced maximum likelihood supervised classification of three derived bands (after Sidjak and Wheate, 1999). The ASTER DEM is the shaded relief base image. Areas of no visible relief are null areas. Image and caption courtesy of Rick Wessels, U.S. Geological Survey, Alaska Volcano Observatory, Anchorage, Alaska. ASTER image (ID SC:AST_L1A.003:2003795732, bands 3, 2, 1 (RGB), acquired on 8 August 2001). From U.S. Geological Survey, EROS Data Center, Sioux Falls, S. Dak.*

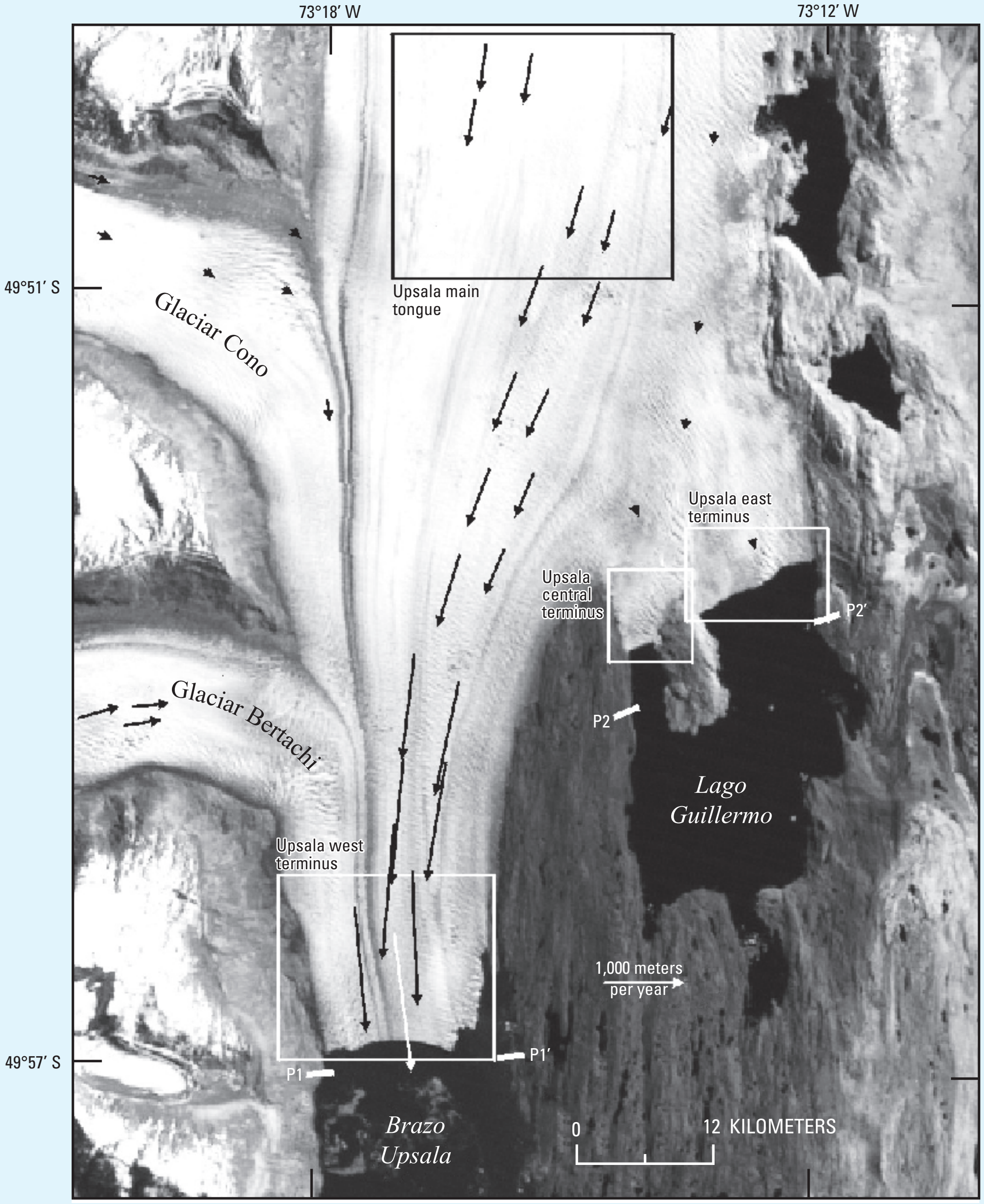

Figure 79.*—Annotated section of Landsat 7 Enhanced Thematic Mapper (ETM+) image (band 8), acquired on 14 October 2001, showing Glaciar Upsala, Southern Patagonian Ice Field, Argentina, with overlay of representative velocity vectors. The vectors were determined automatically using the IMCORR IMage CORRelation software (Scambos and others, 1992) and were simplified for this figure. P1-P1′ and P2-P2′ indicate position of depth profiles in Brazo Upsala (arm of Lago Argentino) and Lago Guillermo, respectively. Rectangles indicate areas of velocity analysis, and are given in Skvarca and others (2003). Modified from Skvarca and others (2003, p. 185, fig. 1).*

GLIMS

Goals

GLIMS makes use of state-of-the-art imaging and image-analysis technology and draws from a global network of nearly 100 scientists in 28 countries. GLIMS has three goals: (1) repeat imaging of the world's estimated 160,000 glaciers by ASTER at up to 15-m-pixel resolution, in 14 bands (spanning the visible through thermal infrared), and in stereo in one of the bands (Raup and others, 2000); (2) build and populate a Web-accessible database of glacier data (Raup and others, 2007); and (3) assess the images for areal extent of glaciers and changes (Kääb, Huggel, and others, 2003; Paul, Kääb, and others, 2004; Rau and others, 2004). GLIMS relies primarily on processed imagery from ASTER, and many of its activities fall under the auspices of the ASTER Science Team; however, it also makes use of imagery from the Landsat series and several other satellites, historic map archives, and field studies. Many members of the GLIMS consortium and other glaciologists are dedicated to practical application of glacier-analysis results to problems involving glacier inventories, climate change, landscape evolution, current and future water-resource availability, and current and future glaciological hazards (Haeberli and others, 2004; Kotlyakov and others, 2010). This work is being undertaken in direct collaboration with the WGMS, Zürich, Switzerland, and is a logical extension of the "Satellite Image Atlas of Glaciers of the World," the "Coastal-Change and Glaciological Maps of Antarctica," the "World Atlas of Snow and Ice Resources," and the World Glacier Inventory efforts.

History and Ties to Other Activities

The idea of GLIMS was conceived by Hugh H. Kieffer, a USGS geophysicist; activities began as a small project of the ASTER Science Team with major goals, at first with no dedicated funding (Kieffer and others, 2000). However, the observational needs for multispectral and thermal glacier observations were considered during the engineering and development phases of ASTER. To make progress toward glaciological analysis on a global scale, it was evident that a global consortium effort was needed. The GLIMS Coordination Center at the USGS Flagstaff Science Center (FSC), Flagstaff, Ariz., U.S.A., gained some initial funding through NASA as a Pathfinder Project at about the time of ASTER's launch on the Terra satellite in 1999. The development of the consortium was a major early activity (Kieffer and others, 2000); however, Kieffer retired from the USGS in 2003. Therefore, the ASTER Science Team selected Jeffrey S. Kargel of the USGS (now with the University of Arizona) as an ASTER Science Team member and GLIMS Principal Investigator. GLIMS gained more NASA funding beginning in 2004 through separate but explicitly linked, competitively reviewed proposals submitted by USGS/FSC, the National Snow and Ice Data Center (NSIDC) at the University of Colorado, Portland State University, the University of Maine, and the University of Nebraska at Omaha. Also, in 2004, various projectwide coordination tasks, database development, and glacier-analysis tasks were funded, notably at NSIDC. In 2005, Kargel resigned from USGS and transferred GLIMS leadership to the University of Arizona in Tucson. Foreign institutions are funded at various levels through their own government agencies and other sources. The foreign investigators have substantial science leverage because the sum of foreign support substantially exceeds the sum of all U.S. support for GLIMS.

Researchers from around the world contribute results from their analysis of ASTER and other data. To organize and archive the results, the GLIMS Glacier Database, global in extent, has been implemented at NSIDC in Boulder, Colorado, U.S.A. Parameters measured and stored in the GLIMS Glacier Database are compatible with and expanded from those of the World Glacier Inventory. GLIMS fits into broader Earth system monitoring efforts through collaborative links; GLIMS represents a "Tier 5" monitoring activity within the Global Terrestrial Network for Glaciers (GTN–G) framework (part of the Global Hierarchial Observing Strategy, or GHOST [*http://www.gosic.org/publications/ghost.pdf*], which integrates in situ measurements, remote sensing, and modeling). GLIMS also collaborates with national and regional glacier monitoring activities such as Omega [*http://omega.utu.fi/*] and the GlobGlacier project, a program funded by the European Space Agency started in autumn 2007 [*http://dup.esrin.esa.it/projects/summaryp98.asp*]. GLIMS is one of the major end users of GlobGlacier output and is therefore acting as a significant guide to the development of that project.

Glacier Analysis

Automated or semiautomated classification of surface units is a key step in extracting the required information from satellite data. One of the major uses of ASTER data in GLIMS is for tracking changes in glacier extent between successive ASTER images or between other datasets and an ASTER overpass. ASTER (and other data) can also be used to track ice motion. The motion of large glaciers can be detected and mapped using ASTER images acquired only a few weeks apart. Use of good time-series data permits measurement of not only the rate of change but also the rate at which change is accelerating or decelerating.

As part of GLIMS-related work, Regional Centers (collaborating GLIMS institutions) have used satellite imagery to quantify glacier change in many regions of the world. The changes tracked range from short-duration, fast-evolving changes in length or area to large changes in thickness, in cases where adequate DEMs can be produced. GLIMS collaborators have produced glacier-change studies focusing on geographic areas such as the Cordillera Blanca mountain range, Perú (Raup and others, 2007), the Antarctic Peninsula (Rau and others, 2006), and the Stubai Alps, Austria (Schicker, 2006). Several GLIMS collaborators have used DEMs to quantify changes in mass balance of glaciers in their regions, including western British Columbia (Schiefer and others, 2007), western Himalaya (Berthier and others, 2007), and Cordillera Blanca, Perú (Racoviteanu and others, 2007; Vignon and others, 2003). Temporal coverage is increasing in some regions. For example, ASTER data have been combined with data from other sources to create a time series to study change through several decades (fig. 80). Debris-covered glaciers pose a special challenge to automated mapping from satellite imagery, but several GLIMS groups have made progress on this problem (for example, Paul, Huggel, and Kääb, 2004; Taschner and Ranzi, 2002).

GLIMS has made a significant start at building a global database of glacier outlines, with an estimated 35 to 40 percent of all glaciers represented. Challenges remain and are well discussed by Paul and others (2007). Detailed

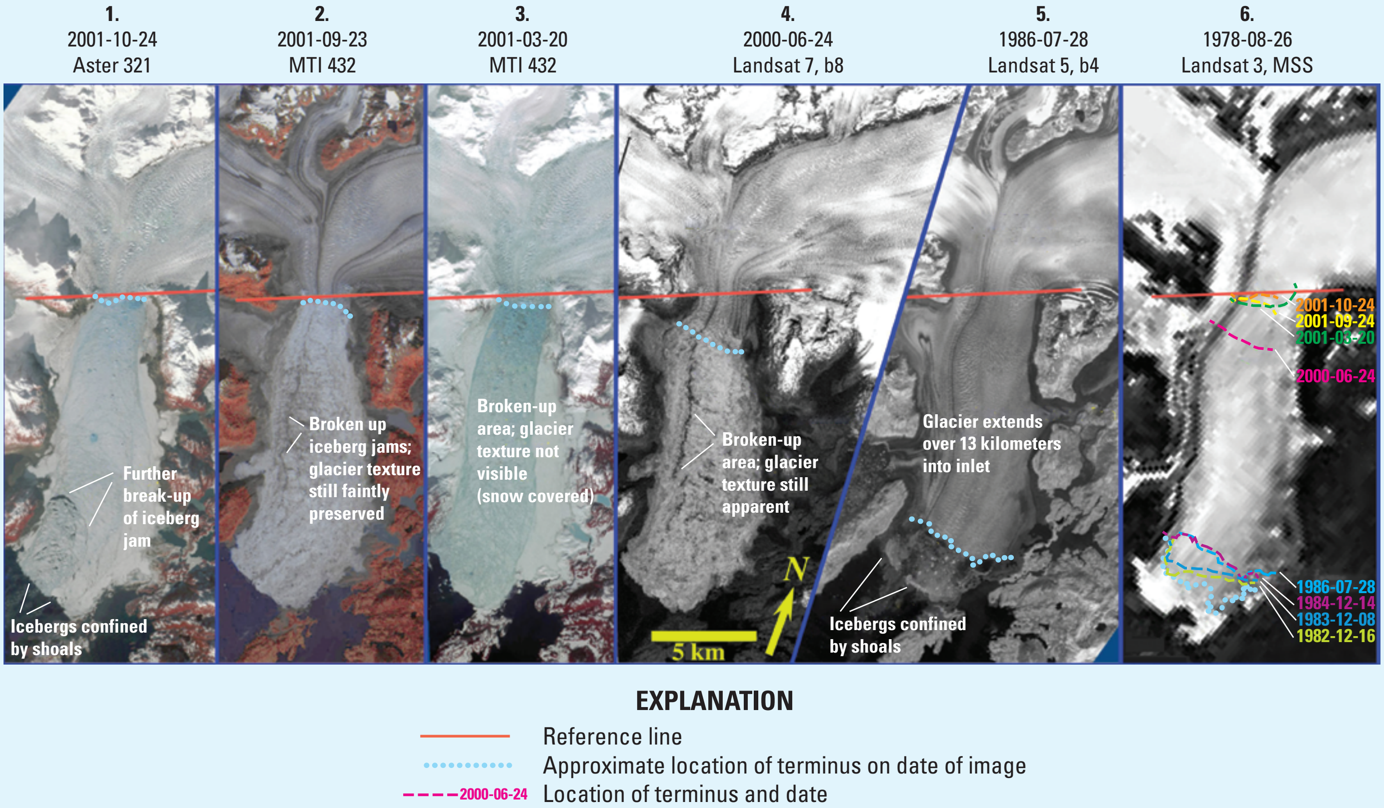

Figure 80.*—The rapid retreat of Columbia Glacier, Alaska, during 1978–2001 observed and documented using the Advanced Spaceborne Thermal Emission and reflection Radiometer (ASTER), and other imaging systems, such as Landsats 3, 5, and 7 and the U.S. Department of Energy's Multispectral Thermal Imager (DOE MTI). Image panels left to right: (1) ASTER L1B bands 3, 2, 1, SC: AST_L1A.003 (2004702997) as red, green, blue, acquired on 24 October 2001 at 21:30:18 UTC; (2) MTI DOE (false-color infrared) acquired on 23 September 2001 at 20:25:40 UTC; (3) MTI DOE (color infrared), acquired on 20 March 2001 at 20:41:29 UTC; (4) ETM+ image ID 7067017000017650, band 8, acquired on 24 June 2000 at 20:53 UTC; (5) TM5 image ID 5067017008620910 (Path 67 , Row 17) band 4, acquired on 28 July 1986 at 20:23 UTC; and (6) Landsat 3 MSS acquired on 26 August 1978 at 19:27 UTC ID: p073-17-3m 19780826. The panel on the right includes sketches of 1982–2001 locations of the terminus. Images and caption courtesy of Rick Wessels, U.S. Geological Survey, Alaska Volcano Observatory, Anchorage, Alaska. See also Krimmel (2008, p. K54–K74, figs. 42–56).*

presentations on selected scientific aspects of GLIMS are published in Kargel and others (2005). An overview of analysis methods is given in Raup and others (2007). In 2009, the International Glaciological Society published a special thematic volume in their Annals of Glaciology series on "World Glacier Inventory" (Jacka and others, 2009); many of the papers published in the volume (v. 50, no. 53) present results of analysis of ASTER image and GDEM data (for example, Racoviteanu and others, 2009; Paul, Barry, and others, 2009).

Continued population of the GLIMS Glacier Database remains the primary goal of GLIMS. Among other future directions, we are striving to understand recent glacier variations and to forecast likely future changes on the basis of climate models (Bush and others, 2004; Lemke and others, 2007). We also need to understand computed assessments of uncertainty in climate models (general circulation models (GCMs)) and the sensitivity of model results to input parameters (Stainforth and others, 2005; Harrison and Stainforth, 2009).

GLIMS Glacier Database

Each collaborating GLIMS institution (Regional Center or RC) oversees the creation and analysis of data for a particular region appropriate to its expertise. These data are submitted to the GLIMS Glacier Database at NSIDC.

Database Contents

The NSIDC GLIMS project has created a geospatial and temporal database composed of glacier outlines and various scalar attributes. These data are derived from high-resolution, optical satellite imagery, primarily the ASTER instrument on the NASA Earth Observing System (EOS) Terra satellite, the TM sensor on the Landsat 5 satellite, and the ETM+ sensor on the Landsat 7 satellite. Historical data (maps and photographs) are used to document changes from earlier periods. Each glacier "snapshot" is from a specific time, and the database is designed to store multiple snapshots representative of different times. The GLIMS Glacier Database currently (2012) contains the outlines for more than 100,000 glaciers (fig. 81), high-resolution imagery (the base imagery for the analysis results) for some areas, and information about GLIMS participants and institutions. The database also includes metadata for approximately 200,000 ASTER images acquired over glacierized terrain. Metadata on ASTER imagery are downloaded daily from USGS Center for Earth Resources Observation and Science (EROS) [*http://edc.usgs.gov/*], parsed, and ingested into the database. Also stored in the database are browse (reduced-resolution) images corresponding to the ASTER metadata; these are viewable online. Statistics on the number and quality of ASTER scenes, grouped by RC, have been generated. This information is crucial to

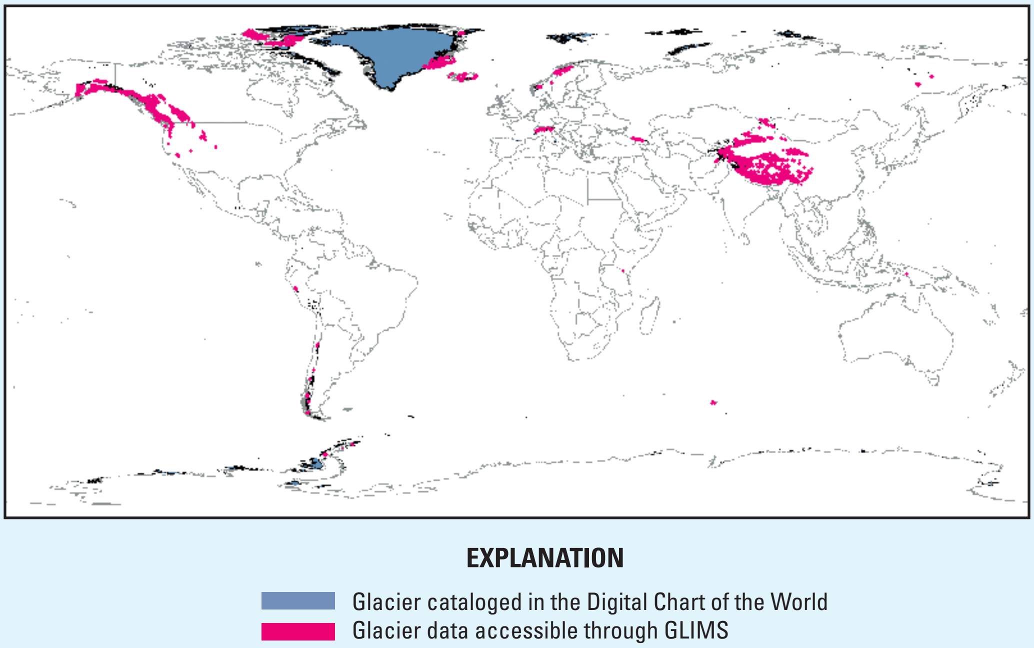

Figure 81.*—Base map of the world's glaciers (blue), including the Greenland ice sheet and the Antarctic ice sheet, and glaciers assessed by GLIMS (red). The blue patches for glaciers represent the dataset of the Earth's glaciers, an augmented version of the Digital Chart of the World (DCW) described in Raup and others (2000). At* http://glims.org.

understanding where suitable images currently exist and for determining what Regions require an increased priority in scheduling time for ASTER to acquire images for their study.

Database Access

NSIDC has implemented a Web-based interface to the GLIMS Glacier Database that enables exploration of the data via interactive maps. There is also a text-based interface that allows queries based on glacier name, size, and many other scalar attributes. The GLIMS MapServer Web site allows users to map and query several thematic layers, including glacier outlines, ASTER image footprints, GLIMS Regional Center locations, and the World Glacier Inventory (fig. 82) [*http://glims.org/*]. Glacier query results can be downloaded into

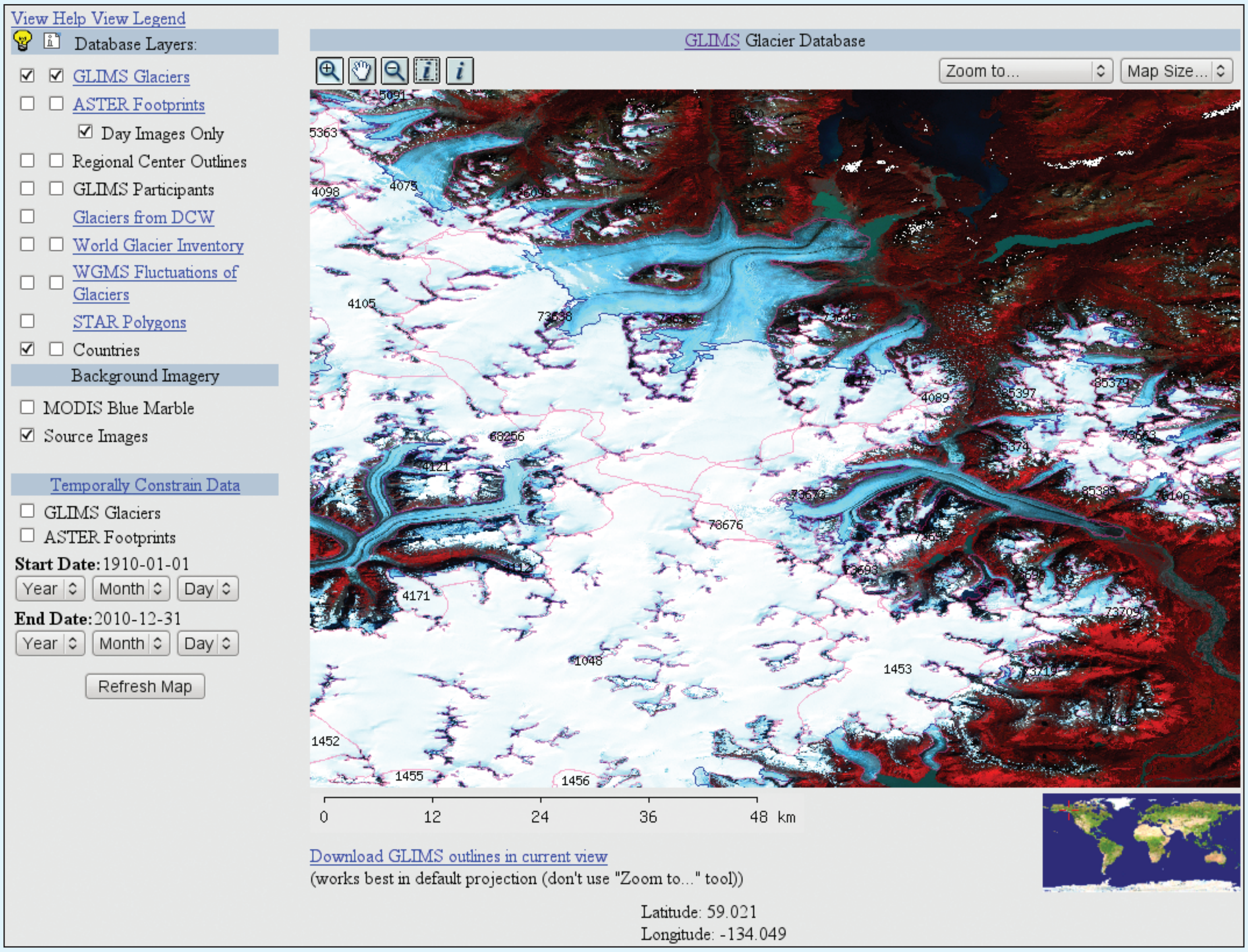

Figure 82.—*Computer screen display of the GLIMS MapServer showing database layers and options for temporarily constraining data. Elements in image include glacier boundaries (red), boundaries (purple) of internal rock (nunataks), transient snow lines (blue), and glacier analysis IDs (numbers). The Llewellyn Glacier is in the upper center of the image; it is an outlet glacier from the Canadian side of the Juneau Icefield. (See also fig. 78; and Chapter K of "Satellite Image Atlas of Glaciers of the World," p. K104, fig. 85.)*

a number of GIS-compatible formats, including Environmental Systems Research Institute, Inc. (ESRI) Shapefiles, MapInfo tables, Generic Mapping Tools (GMT), Geographic Mark-up Language (GML), and Keyhole Markup Language (KML) for viewing in virtual globes such as Google Earth. When the ASTER footprint layer is queried, metadata for those images are displayed, and links to browse imagery are provided. Both the glacier's layer and the ASTER footprints layer can be temporally constrained so that the resulting map will show features from the specified time period only. The data are stored in a spatially enabled database (PostGIS), which has sophisticated functions for analysis and query of spatial data. The MapServer application provides data from the GLIMS Glacier Database to other Open Geospatial Consortium (OGC)-compliant services via OGC-standard protocols, thereby increasing the utility of this glacier dataset (fig. 83). The database is accessible at [*http://nsidc.org/glims/*]. This Web site provides an overview of the GLIMS Glacier Database and links to the interactive map interface and text-based interface.

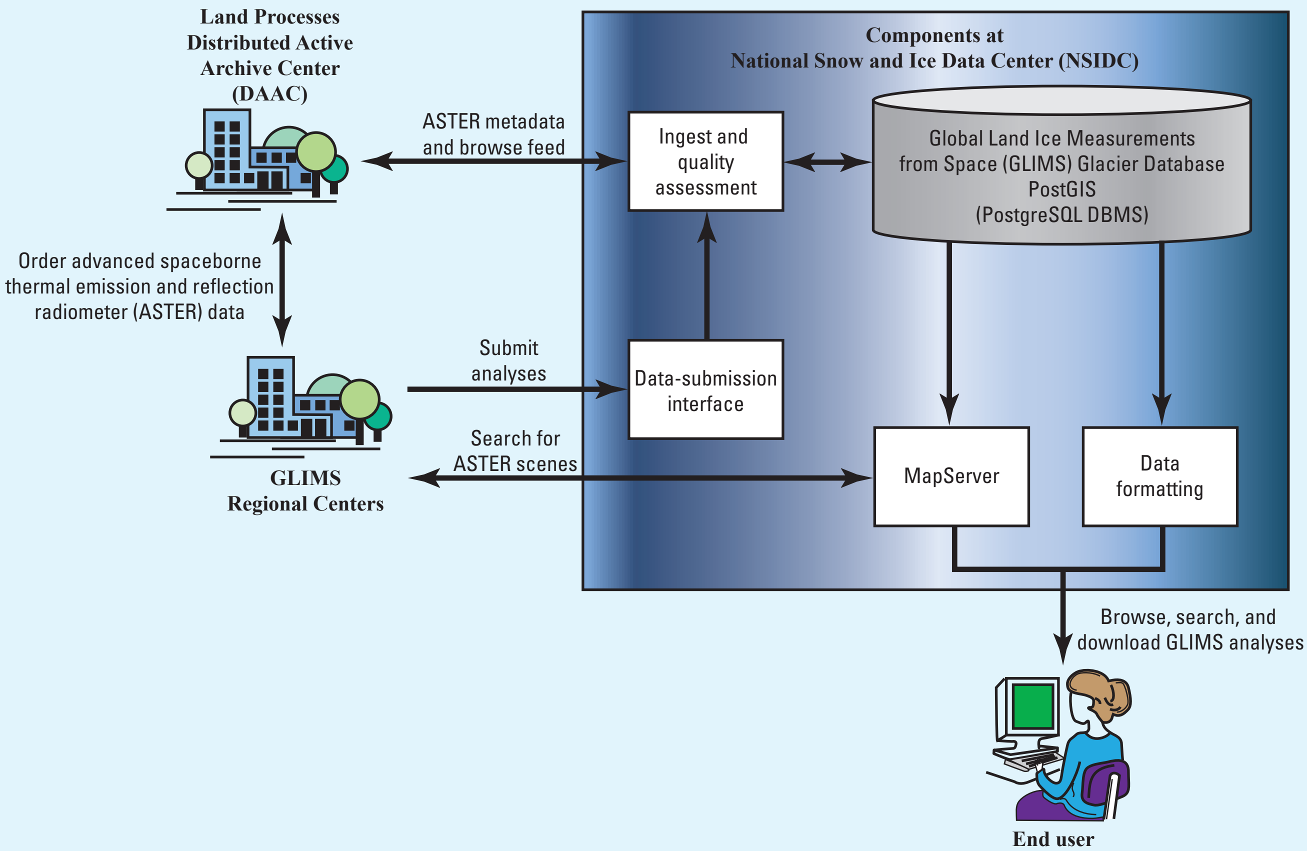

Figure 83.—*Components of the GLIMS Glacier Database, its public interfaces, and links to GLIMS Regional Centers (RCs). RCs can use the GLIMS MapServer to search for appropriate ASTER scenes, which can be ordered from the Land Processes Distributed Active Archive Center (LP DAAC) at the USGS Center for Earth Resources Observation and Science. RCs produce the results of their analysis using GLIMSView and other tools, and then transfer the results to NSIDC. Using a Web browser, anyone can search GLIMS data graphically or with text-based constraints, and GLIMS data can be downloaded in a variety of formats, including Environmental Systems Research (ESRI) shapefiles, Geography Markup Language (GML), Generic Mapping Tools (GMT), and Keyhole Markup Language (KML) for viewing in virtual globes such as Google Earth.*

Conclusions

The GLIMS consortium represents an international collaborative effort to build a database of glacier outlines and associated metadata suitable for studying glacier change. The project has involved directing ASTER data acquisition for glaciers, standardization of data models, definitions, processing protocols, and data-transfer methods to create a database with appropriately consistent data. This project expands on previous glacier inventory work, such as the World Glacier Inventory, which the World Glacier Monitoring Service coordinates. We are making progress in most regions; the database contains data for all major glacierized land masses. The database is accessible via the World Wide Web, and data maybe downloaded in a choice of formats. Through their GLIMS work, Regional Centers are producing many scientific results detailing glacier change in many regions of the world. As this work continues, we expect the database to contain multitemporal data for an increasing number of glaciers, with increasing geographic coverage as well. With glacier change apparently accelerating globally, comprehensive mapping of Earth's glaciers has never been more important.

Acknowledgments

The authors acknowledge vital contributions to the GLIMS consortium by the U.S. GLIMS collaborators and the international GLIMS consortium members, listed on the [*http://glims.org/*] Web site. We thank Rick Wessels, Alaska Science Center, USGS, for his contributions to figures 78 and 80. The GLIMS Glacier Database and Web site rely heavily on various open-source software packages, including GNU/Linux, PostgreSOL, PostGIS, and MapServer. GLIMS work at NSIDC is supported by NASA grants NNG04GF51A and NNG04GM09G. Jeff Kargel was supported through NASA ASTER Science Team funding.

Glaciological Hazards: Global, Regional, and Local

Introduction

Hazards associated with glaciers can be global (eustatic rise in sea level), regional (icebergs from tidewater glaciers and ice shelves), and local (surge-type glaciers, jökulhlaups, ice avalanches, debris avalanches, lahars, and landslides and rockslides). The rise in eustatic sea level is the most important glaciological hazard because of its global impact (Lambeck, 2002). Hazards are, of course, defined in terms of the impact of various glaciological processes on human populations, causing injuries or death and causing damage to the infrastructure, such as roads, bridges, buildings, and so on. Although several glaciological hazards were discussed earlier in "Glacier Mass Changes and Their Effect on the Earth System," by Mark B. Dyurgerov and Mark D. Meier, on p. A220–A222, this section discusses glaciological hazards from a geographic (global, regional, and local) point of view. Historical and modern accounts are also provided on the occurrence of the two types of jökulhlaups (lacustrine and volcanic and geothermal) in various geographic regions. Paleojökulhlaups resulting from retreat and melting of continental ice sheets in North America, Eurasia, and South America during the late Pleistocene and early Holocene epochs will also be discussed.

Twenty-five years ago, the Icelandic geophysicist/glaciologist Helgi Björnsson and other geologists participated in a "Workshop on Alaskan Hydrology"; one of the subjects addressed was "Natural Hazards Caused by Glaciers" (Björnsson and others, 1986). In 1991, the USGS held a two-day "Polar Research Program Strategies Workshop" that involved 26 USGS scientists. Many scientific topics that addressed the polar regions included glaciological hazards (Williams, 1995). More recently, an "International Workshop on Glacier Hazards, Permafrost Hazards, and Glacier Lake Outburst Floods (GLOFs) in Mountain Areas: Processes, Assessment, Prevention, Mitigation" was held in Vienna, Austria, on 10–13 November 2009 [*http://www.geo.uio.no/remotesensing/gaphaz/home.html*]. Periodic imaging with satellite sensors of potential glaciological hazards in remote valleys of the Himalaya and the Andes was a high priority of the Vienna workshop because monitoring systems are absent in such high mountain regions. The workshop was organized by a Scientific Working Group of the International Association of Cryospheric Sciences (IACS) and the International Permafrost Association (IPA), Glacier and Permafrost Hazards in Mountains (GAPHAZ).

Global: Eustatic Rise in Sea Level

The eustatic rise in sea level resulting from discharge of glacier meltwater into oceans and seas (and steric increase in ocean volume from climate warming) has a global impact on all of the Earth's insular and continental coastlines: deltas, barrier islands, other unconsolidated sediments, bedrock, and cryospheric. An excellent review of the projected rise in sea level by the end of the 21st century is provided by Fletcher (2009). The variation in sea level and changes in volume of glacier ice on land (ice above sea level) during the Quaternary Period are discussed earlier in Part A-2.

The islands and coastal regions most impacted by changes in sea level are those that are topographically low. In the context of global warming, low-lying islands and coastal regions include coral reefs, such as atolls, and deltas, barrier islands, and coastal plains. Any rise in sea level will cause inundation of these coasts (fig. 84) (New Scientist, 2009; Williams and Gutierrez, 2009). In regions of low topographic relief, the vertical rise in sea level is magnified horizontally across the landscape, making human populations in such areas increasingly vulnerable to inundation from seasonal floods and surges from intense storms, in addition to loss of land to rising sea level (Small and Nicholls, 2003; McGranahan and others, 2007; Syvitski and others, 2009).

Although atolls generally have small human populations, their area is limited, and the topographic elevation of even the highest elevation is usually only a few meters above sea level. Any significant inundation will therefore force these populations to relocate to higher terrain elsewhere (Whipple, 2009). Residents of three island archipelagos, two in the Pacific Ocean (the Republic of Kiribati and Tuvalu (Patel, 2006)) and one in the Indian Ocean (the Republic of Maldives (Schmidle, 2009)), are already working on plans to relocate in the future (Russell, 2009).

However, it is in the heavily populated deltas on Earth that the greatest hazard to human populations and associated infrastructure will occur (Rowley and others, 2007). With the introduction of agriculture at the beginning of the Holocene Epoch, about 12,000 years ago (Diamond, 1997), groups of human beings found that deltas were favored landscapes for growing crops for three reasons: (1) they had relatively flat land, gently sloping toward the sea; (2) the existing rivers and distributary streams provided a source of water independent of rainfall; (such landscapes could also be easily modified to create a network of irrigational channels between the existing courses of streams); and (3) seasonal floods deposited a new layer of sediment with its associated nutrients. Deltas, therefore, became the locus of increasingly large populations of human beings. They were certainly the basis on which many ancient civilizations developed, as in Egypt (Nile River delta) and in Mesopotamia (Tigris-Euphrates delta). The dense populations of many modern societies inhabit other major delta regions associated with food production, such as rice. Among such societies are those of Bangladesh and West Bengal, India (Ganges-Brahmaputra delta), China (Yangtze River and Yellow River deltas), Pakistan (Indus River delta), Vietnam (Mekong River and Red River deltas), Egypt (Nile River delta), and the United States (Mississippi River delta).

Deltas began to form when the continental ice sheets on land in North America, Eurasia, and South America had completed melting. Sea level then began to stabilize between 8000 and 6500 years ago (Stanley and Warne, 1997) because rivers met the ocean at grade and deposited sediment in suspension and in bed load. Therefore, future rise in sea level from melting of glacier ice on land and from steric increase in volume will inundate the Earth's deltas, including crop lands, and it will submerge the built infrastructure, ultimately forcing the local population to migrate inland to higher terrain (Appenzeller, 2007; Syvitski, 2008). Coastal aquifers will also become more saline, so groundwater resources will be reduced for coastal populations. On the larger deltas, such as the Ganges-Brahmaputra delta, with a current population of about 150 million people, tens of millions of human beings will be displaced and will need to be relocated. Additional infrastructure will have to be constructed to meet the needs of the relocated population.

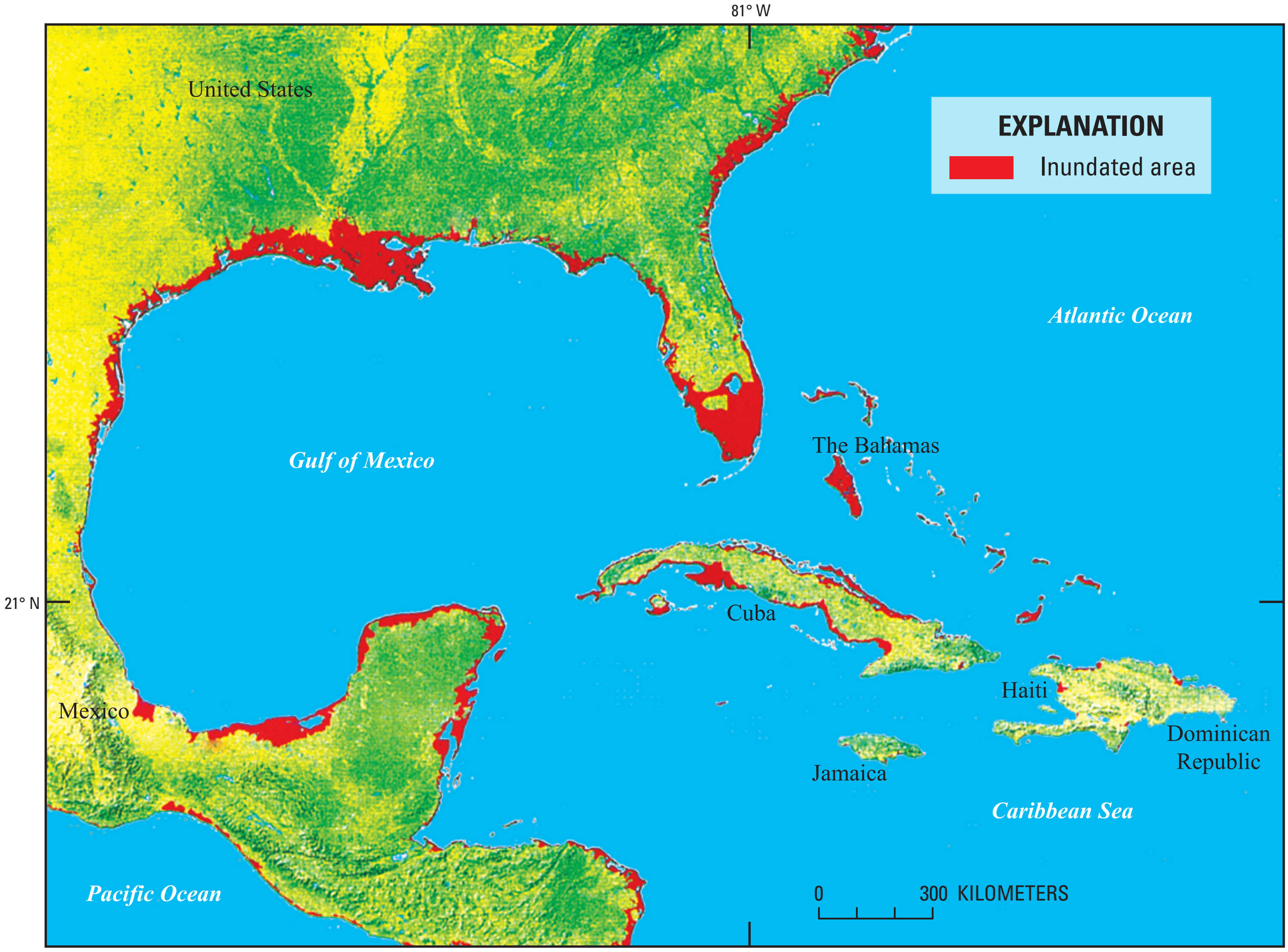

Figure 84.*—Inundation of low-lying coastal regions and islands in the Gulf of Mexico, Caribbean Sea, Pacific Ocean, and Atlantic Ocean projected as occurring if sea level rises 6 m. Note especially the loss of land in the southeastern United States, Florida, the Louisiana delta, and other areas of the Gulf Coast of the United States and México, especially the inundation of most of the Bahamian islands, which will force its citizens to migrate. Inhabitants of all areas marked in red would be displaced inland. During the last interglacial, most of the Greenland ice sheet melted (Koerner, 1989). The resulting rise in sea level was about 6 m. From Rowley and others (2007, p. 105, fig. 1).*

In more developed regions, houses, public buildings, roads, and other infrastructure will be inundated along coastal regions, including barrier islands and sea islands. Virtually all of the world's major coastal cities, including those inland on estuaries, have major airports. Using model projections, Yin and others (2009) examined the impact of a rapid rise in sea level on the densely populated northeastern coastal regions of the United States. To protect (for a time) or to relocate the most valuable elements of the infrastructure will require a large expenditure of funds. As an example, after the catastrophic North Sea Storm in 1953 that flooded polders (lowland reclaimed from the sea) and the built infrastructure and that killed 1,835 people, the Netherlands spent $8 billion in the Delta Project during a 44-year period to provide protection from future inundation by severe storms. The 1953 storm surge raised sea level 5 to 6 m locally. However, neither the engineered barriers in the Netherlands nor the storm barriers that protect many cities (for example, London, England, U.K., Providence, Rhode Island, U.S.A.), were designed to accommodate a significant rise in sea level, with or without a severe storm surge adding to the rise. If the Greenland ice sheet melted completely, its meltwater would add about 6 m to the global sea level (table 3). That rise equals the 1953 storm surge in the North Sea, but the rise from the Greenland meltwater would be global in its effect.

Regional: Icebergs

Outlet glaciers from ice caps and ice fields, ice streams from ice sheets whose termini reach the sea, and ice shelves produce large volumes of icebergs from ice walls (grounded ice) and from ice fronts (floating ice). Tabular icebergs are commonly generated from the cryospheric coast of Antarctica, especially from its ice shelves (for example, Ross, Ronne, Filchner, and Amery Ice Shelves and many other smaller ones). Smaller tabular icebergs break off small ice shelves (for example, Ward Hunt Ice Shelf) on the cryospheric coast of Ellesmere Island, Nunavut, Canada, where they have been used as "ice islands" for scientific observations in the Arctic gyre (for example, Fletcher's Ice Island). See Ellesmere Island Ice Shelves and Ice Islands, by Martin O. Jeffries, in Chapter J of this volume. Most other icebergs from Greenland, Alaska, Svalbard, and other glacierized islands in the Arctic are irregular in shape and drift with the local currents. An iceberg from Greenland drifting south off the coast of Newfoundland was rammed by the *H.M.S. Titanic* on the night of 14 April 1912. The *Titanic* sunk nearly 3 hours later on 15 April 1912, with the loss of 1,517 lives. The *S.S. Exxon Valdez*, transiting Prince William Sound from Valdez, Alaska, to the Gulf of Alaska, was holed after running into a skerry (Bligh Reef) on 24 March 1989, discharging 40 million litres (10.8 million U.S. gallons) of oil into the Sound. The ship had navigated a course much farther east than was usual to avoid icebergs in Prince William Sound that originated from the calving terminus of the rapidly receding Columbia Glacier, a tidewater glacier (Post, 1977; Tangborn and Post, 1998). The University of Alaska at Fairbanks awarded an honorary doctor of science degree to the USGS glaciologist Austin Post in 2004 for his contributions to the field of glaciology; part of the citation to the degree read: "Icebergs are thought to have played a role in the spill as the Exxon Valdez had shifted course to avoid them…".

Local: Tidewater Glaciers

Tidewater glaciers are glaciers whose terminus extends into the ocean directly, into a fjord, an ice margin lake, or a strait that is connected to the sea. The termini of all marine tidewater glaciers are affected by the rise and fall of the local tide; hence their name. Glaciers can also terminate in a lake, unconnected to the sea; the calving process from the face of the termini of marine tidewater glaciers as well as of lacustrine-terminating glaciers is the same. Both types produce icebergs, but the marine tidewater glaciers are the ones that generate icebergs that pose a hazard to ships.

Tidewater glaciers generally, and with specific reference to the 60 current and past tidewater glaciers of Alaska, are discussed in Chapter K of this volume (Molnia, 2008, p. K52–K74); Krimmel, in a separate section in Chapter K (p. K54–K74), describes the tidewater-glacier cycle and the history of the "Columbia and Hubbard Tidewater Glaciers."

Examples of tidewater glaciers are discussed in 8 of the 10 geographic-area chapters in this volume, including B, "Antarctica"; C, "Greenland"; D, "Glaciers of Iceland" (for example, Breiðamerkurjökull, which calves into Jökulsárlón, an ice margin lake that is connected to the North Atlantic Ocean through a short stretch of river (Jökulsá); E, "Glaciers of Europe," specifically Svalbard and Jan Mayen; F, "Glaciers of Asia," specifically Russian Arctic islands; I, "Glaciers of South America" (for example, Glaciar San Rafael is an outlet glacier from the Northern Patagonian Ice Field that calves into Laguna San Rafael, which is connected by a narrow strait to Estero Elefantes and through the Archipiélago de los Chonos to the Pacific Ocean); J, "Glaciers of North America," specifically glaciers on Ellesmere Island, Nunavut, Canada; and K, "Glaciers of Alaska."

Local: Surge-Type Glaciers

Surge-type glaciers are glaciers in which, after an extended period of quiescence, the lower part of the glacier suddenly moves forward ("surges") for a kilometer or more during a few weeks or months at a speed 10 to 100 times greater than the glacier's normal flow (Post, 1969, p. 229). The rapid increase in speed of the glacier is the result of dynamic instability in a glacier. An imperfectly understood slow motion of the glacier increases its gradient, and the resultant instability is an inability to maintain a balanced velocity. At the completion of the surge, the receiving area of the surging glacier is broken up only "skin deep;" its core of ice is flowing viscously and therefore is a more or less coherent mass. The lower part of the glacier becomes stagnant, however, and melts *in situ* during the period of quiescence. After a decade or several decades, dynamic instability recurs, and the glacier again surges.

The classic study of a surge-type glacier in Alaska, "Variegated Glacier," was done by Kamb and others (1985), who concluded that high water pressure at the base of the glacier was the process that caused a hundredfold increase in glacier speed. Björnsson and others (2003), in their analysis of surge-type glaciers, concluded that glaciers whose surfaces are too gentle in slope (so that they flow too slowly to keep up with the rate of accumulation) create an instability that a surge resolves. Whatever mechanism(s) or process(es) are discovered to govern surge-type glaciers must also be able to account for the vast differences in size of surge-type glaciers, in the speed of their advance, in the slope of the glacier surface, in their total advance, and in the recurrence interval of the surge. For example, two glaciers, Búrfellsjökull and Teigardalsjökull, are located in northern Iceland; they are the smallest surge-type glaciers in the world, each with areas of about 1 km^2 (Sigurðsson and Williams, 2008). For the vast disparity in size of surge-type glaciers, compare the 1973 oblique aerial photograph (fig. 85*A*) of Eyjabakkajökull, a surge-type glacier on the northern margin of the Vatnajökull ice cap, with the 1978 Landsat 3 return beam Vidicon (RBV) image (fig. 85*B*) of the same glacier. To the left of Eyjabakkajökull is the eastern part of Brúarjökull, another surge-type glacier with historic short-term advances of 8 km (1963–1964) to 10 km (1890) (Thorarinsson, 1969).

Surge-type glaciers have been described in Antarctica (Stearns and others, 2008); in Greenland (Weidick, 1988; Jiskoot and Juhlin, 2009), including Jakobshavn Isbræ, which Mark Meier (1972) referred to as "a continuously surging glacier." In Iceland, 21 surge-type glaciers were identified (not including other glaciers about which the authors are uncertain) (Thorarinsson, 1964, 1969; Björnsson and others, 2003; Sigurðsson and Williams, 2008). In several mountain ranges in western Canada and Alaska, 204 surge-type glaciers were identified (Post, 1969). In Svalbard (Chapter E, "Glaciers of Europe" in this volume), 86 surge-type glaciers are listed. Several surge-type glaciers are mentioned in "Glaciers of Asia" (Chapter F), especially in the Pamirs; in South America (Chapter I, "Glaciers of South America"), especially several outlet glaciers from the Southern Patagonia Ice Field in Argentina and Chile); in North America (Chapter J, "Glaciers of North America"); and in Alaska (Chapter K, "Glaciers of Alaska"), 205 are noted.

Local: Jökulhlaup

Introduction

Following the convention established by Thorarinsson (1939a), who described two types of jökulhlaup (glacilimnogen and volcanogen), glaciologists and volcanologists use the term jökulhlaup, an Icelandic loanword for "glacier outburst floods," to describe two types of large-magnitude floods in glacierized regions: lacustrine and volcanic and geothermal. Lacustrine jökulhlaups result from the failure of terminal and (or) lateral moraines that impound glacial lakes. Many examples of large-volume floods, including jökulhlaups, are documented in the book "The Extremes of the Extremes: Extraordinary Floods," edited by Snorrason and others (2002). Volcanic and (or) geothermal jökulhlaups result from subglacial volcanic and (or) geothermal activity that melts large volumes of glacier ice. In the case of subglacial volcanic activity, the eruption usually penetrates the surface of the glacier, leaving a linear depression or a concentric collapse cauldron, often open to the atmosphere for a time.

A jökulhlaup may occur soon after a volcanic eruption begins, if the meltwater follows a path beneath the glacier to its terminus or margin. Sometimes, however, the jökulhlaup may occur days or weeks following the start of the eruption, when sufficient volume of water has accumulated in a subglacial depression, eventually creating a channel beneath the glacier and (or) "lifting" the glacier from it bed. Subglacial geothermal activity generates meltwater that

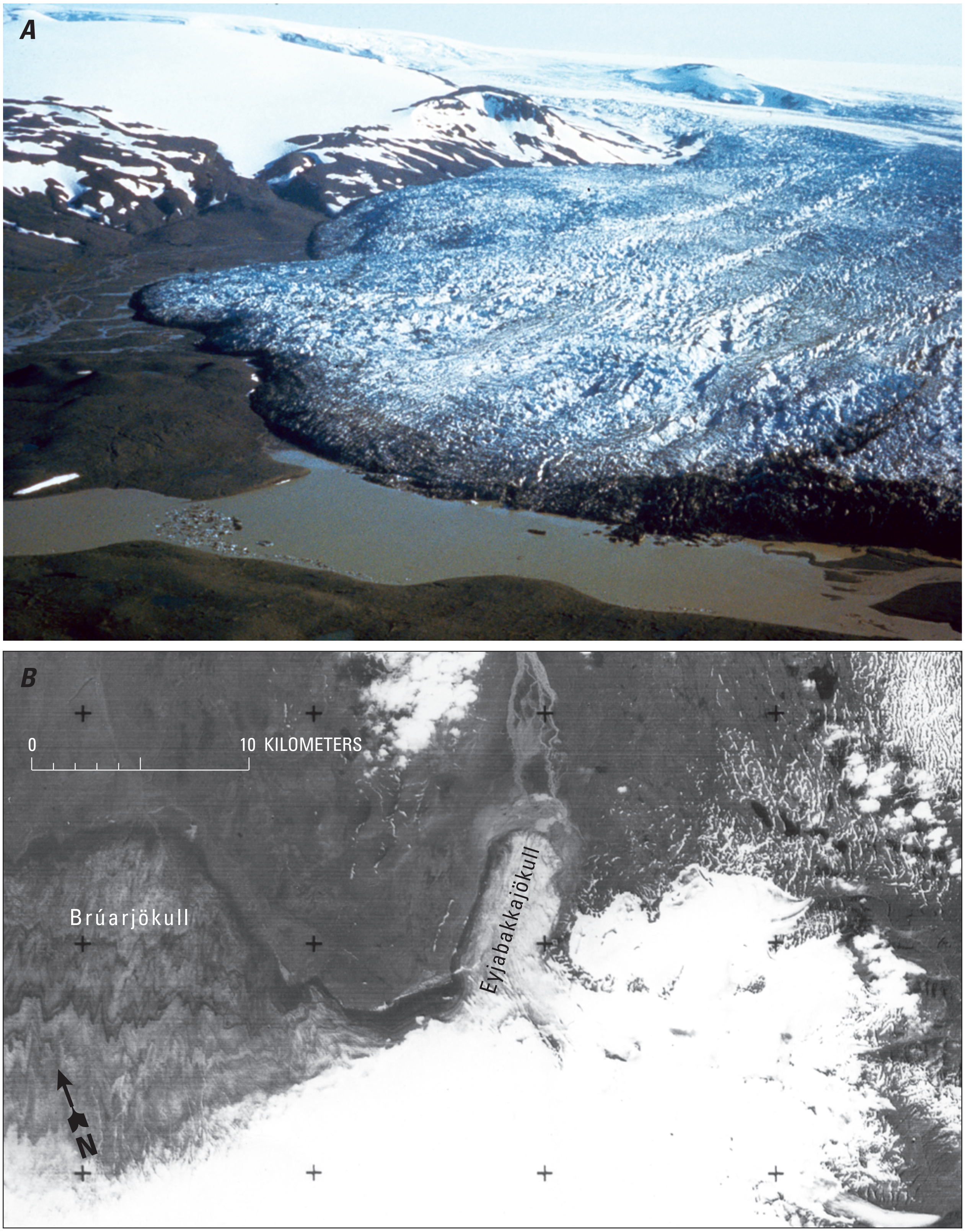

Figure 85.—**A**, *Oblique aerial photograph looking south across the terminus of the surge-type glacier, Eyjabakkajökull, as it appeared on 25 July 1973 after it had completed a 2.8-km surge. Photograph by Richard S. Williams, Jr., U.S. Geological Survey.* **B**, *Part of Landsat image 30157-11565-D, acquired on 9 August 1978, of Eyjabakkajökull, after the melting and retreat of the glacier's terminus more than 5 years after its surge. The fractured ice in its lower part, including stagnation during that interval, shows the kind of detail that Landsat 3 return-beam vidicon (RBV) images offer (Williams, 1986a, p. 8, fig. 3).*

is trapped below a subsidence in the glacier surface (for example, concentric fractures bounding a subsidence cauldron). Eventually the hydraulic head (level) of the accumulation of subglacial water will reach a pressure sufficient to flow beneath (subglacial), within (englacial), and on the surface (supraglacial) of the glacier (Haraldson, 1997).

In preparation for discussing the global geographic distribution of historic and modern jökulhlaup events caused by the failure of ice-dammed or moraine-dammed lakes[18] and by subglacial volcanic activity, the melting and retreat of the continental ice sheets during the late Pleistocene and the early Holocene Epochs that created very large impoundments of glacial meltwater will be reviewed.

Paleojökulhlaups During the Late Quaternary Era

During the late Pleistocene and early Holocene, impoundment of glacial meltwater along the margins of continental ice sheets, thinning of glacier ice, and regional recession of ice sheets, often over irregular subglacial topography, created conditions for ice-dam failure (undercutting and (or) overtopping of the ice dam and overflow of the proglacial lake). The discharge of enormous volumes of water followed in several cases (>1 sverdrup; 1 sverdrup is defined as 10^6 m^3 s^{-1}), or what can be described as "megajökulhlaups," defined as a jökulhlaup with a discharge ≥1 sverdrup (Martini and others, 2002). All known historic or modern jökulhlaups, whether lacustrine or volcanic in origin, have water discharges that are usually <0.5 sverdrup (Baker, 2002a, b).

O'Connor and Costa (2004, p. 3, in table 1) provided a list of 27 Quaternary jökulhlaups (and riverine floods) whose peak discharges were ≥0.1 sverdrup. Their table spans the period from the late Pleistocene to the late 20th century. The six largest were megajökulhlaups, that occurred during the late Pleistocene, and were the result of the failure of ice dams: four in Russia, one in Mongolia, and one in the northwestern United States. The largest was the Kuray megajökulhlaup, Altai, Russia (Carling and others, 2002), with a peak discharge of 18 sverdrups. The Missoula megajökulhlaup, in the northwestern United States, ranked second, with a peak discharge of 17 sverdrups. The megajökulhlaup in Mongolia had a peak discharge of 4 sverdrups, and the three others in Altai, Russia, had peak discharges of 2, 2, and 1.9 sverdrups. Four other late Pleistocene jökulhlaups resulting from the failure of ice dams were: Lake Regina, Canada and the United States (0.8 sverdrups); Wabash River, Indiana (0.27 sverdrups); Lake Agassiz, bordering Canada and the United States (0.13 sverdrups); and Porcupine River, Alaska (0.13 sverdrups). Late Pleistocene jökulhlaups are also reported to have occurred in southern South America (Bell, 2008), as the Patagonian ice sheet thinned, retreated, and finally separated into two smaller ice sheets, the remnants of which are the present-day Northern and Southern Patagonian Ice Fields (see Chapter I, Glaciers of South America, in this volume).

The best-documented megajökulhlaup was the drainage of glacial Lake Missoula, about 15,000 years ago, with a peak discharge of 17 sverdrups (O'Connor and Costa, 2004). First described by the geomorphologist J Harlan Bretz (1923) as the "Spokane Flood," the second largest megajökulaup known

[18] Glaciologists and glacial geologists studying glaciers in the Alps and Himalayas have used "glacier lake outburst flood (GLOF)" instead of "jökulhlaup"; the latter term takes precedence, having first been used by the Icelandic glaciologist Sigurdur Thorarinsson in 1939. "Jökulhlaup" is defined in the "Glossary of Geology" (Neuendorf and others, 2005, p. 345); GLOF is not included in the nearly 39,300 terms defined in the fifth edition of the "Glossary."

has been described by many geologists in terms of the erosional and depositional impact on the terrain. The Lake Missoula megajökulhlaup flowed from east to west from its impoundment in Montana along the southern margin of the Cordilleran ice sheet and across Idaho and Washington along the drainages of the Snake and Columbia Rivers. In southeastern Washington, the "Channeled Scablands" and the braided pattern of outflow channels are notable landscape features that were formed (Bretz, 1969; Weis and Newman, 1976; Baker and Nummedal, 1978; Baker, 2002a, b). Waitt (1985) postulated multiple jökulhlaups from glacial Lake Missoula.

During the early Holocene, impounded glacial meltwater along the thinning and northward retreating southern margin of the Laurentide ice sheet in eastern Canada was discharged northward into Hudson Bay. Two overflow episodes of proglacial Lake Agassiz, with peak discharges of 1.2 and 0.2 sverdrups, discharged freshwater into the Labrador Sea and the North Atlantic Ocean. This jökulhlaup, about 8,200 years ago, caused an abrupt cooling event by altering oceanic circulation (Colman, 2002; Clarke and others, 2003; Kleiven and others, 2008).

One other jökulhlaup of note during the early Holocene occurred in northern Iceland about 2,500 years ago. A subglacial volcanic eruption from the northern part of the Vatnajökull ice cap, originating from the Bárðarbunga caldera (Tómasson, 2002) or from the double caldera at Kverkfjöll (Thorarinsson and others, 1974), caused a jökulhlaup along the Jökulsá a Fjöllum which had a peak discharge of 0.3 sverdrup (Tómasson, 2002) to 0.7 sverdrup (Waitt, 2002). As with the channeled scablands of Washington, erosional landscape features, such as a large abandoned waterfall and associated headward erosion at Ásbyrgi, document a large jökulhlaup in the past (Malin and Eppler, 1981; Tómasson, 2002).

Historic and Modern Jökulhlaups

Jökulhlaups from Subglacial Volcanic and (or) Geothermal Activity

One of the largest jökulhlaups in modern time was the 12 October 1918 volcanic eruption of Katla, the source of repeated subglacial eruptions in the Mýrdalsjökull ice cap, Iceland, during historic times. The 1918 eruption caused a jökulhlaup with a peak discharge estimated at 0.3 sverdrup (Tómasson, 1996, 2002), the tenth largest glaciological event (accompanied by a subglacial volcanic eruption from the Katla caldera) listed by O'Connor and Costa (2004). The 1795 treatise of the Icelandic glaciologist Sveinn Pálsson on the glaciers of Iceland documented numerous jökulhlaups from historic annals; most were volcanic (subglacial) in origin and had flowed across the sandur on Iceland's south coast (Pálsson, 2004 [1795]). From Pálsson's descriptions and other historic documents, it is obvious that the 1721 and 1755 jökulhlaups from Katla were considerably larger in volume than the 1918 jökulhlaup from Katla.

Icelandic (Thorarinsson, 1956; Björnsson, 1975; Haraldsson, 1997) and non-Icelandic scientists (Wadell, 1935) have thoroughly studied jökulhlaups that result from subglacial volcanic eruptions and subglacial geothermal activity in Iceland. They have especially studied the October 1996 volcanic eruption on the eastern margin of the Bárðarbunga volcanic caldera, northwestern Vatnajökull, Iceland, which included infilling of the subglacial lake in the Grímsvötn caldera, and a subsequent jökulhlaup from the terminus of the Skeiðarárjökull outlet glacier in November 1996 (Haraldsson, 1997; Jóhannesson, 2002). The 1996 Skeiðarárhlaup had a peak discharge of 50,000 $m^3 s^{-1}$ or 0.05 sverdrup.

All glacierized regions of the world with active volcanoes have the potential for jökulhlaups, from either volcanic or geothermal activity (Kotlyakov and others, 1985). Geographic regions where glacier-capped and (or) subglacial volcanoes are present are discussed in this volume and in the chapters in this volume that have documented volcanic eruptions, including B, "Antarctica" (Corr and Vaughn, 2008); D, "Glaciers of Iceland"; E, "Glaciers of Europe" (Beerenberg Volcano on the northern part of Jan Mayen); I, "Glaciers of South America" (Volcán Lautaro, Southern Patagonian Ice Field, Chile and Argentina (Arko and Skvarca, 1964)); Nevado del Ruiz Volcano, Colombia (Naranjo and others, 1986); J, "Glaciers of North America" (active volcanoes in western Canada); and K, "Glaciers of Alaska" (Aleutian Islands and mainland Alaska (Benson and Motyka, 1979)).

Jökulhlaups from Ice-Dammed, Subglacial, Englacial, and Supraglacial Lakes

Water can be impounded (fig. 86, modified from Roberts, 2005) on the margin of glaciers, at the base, within, and on the surface of glaciers (fig. 87). The impoundments are formed from glacial meltwater, percolating rainfall, and runoff from the surrounding landscape. Other than jökulhlaups resulting from subglacial volcanism and (or) geothermal activity, most jökulhlaups are the result of the failure of ice dams that impound water at the margin of a glacier or where the glacier has advanced to dam a valley or the arm of a fjord.

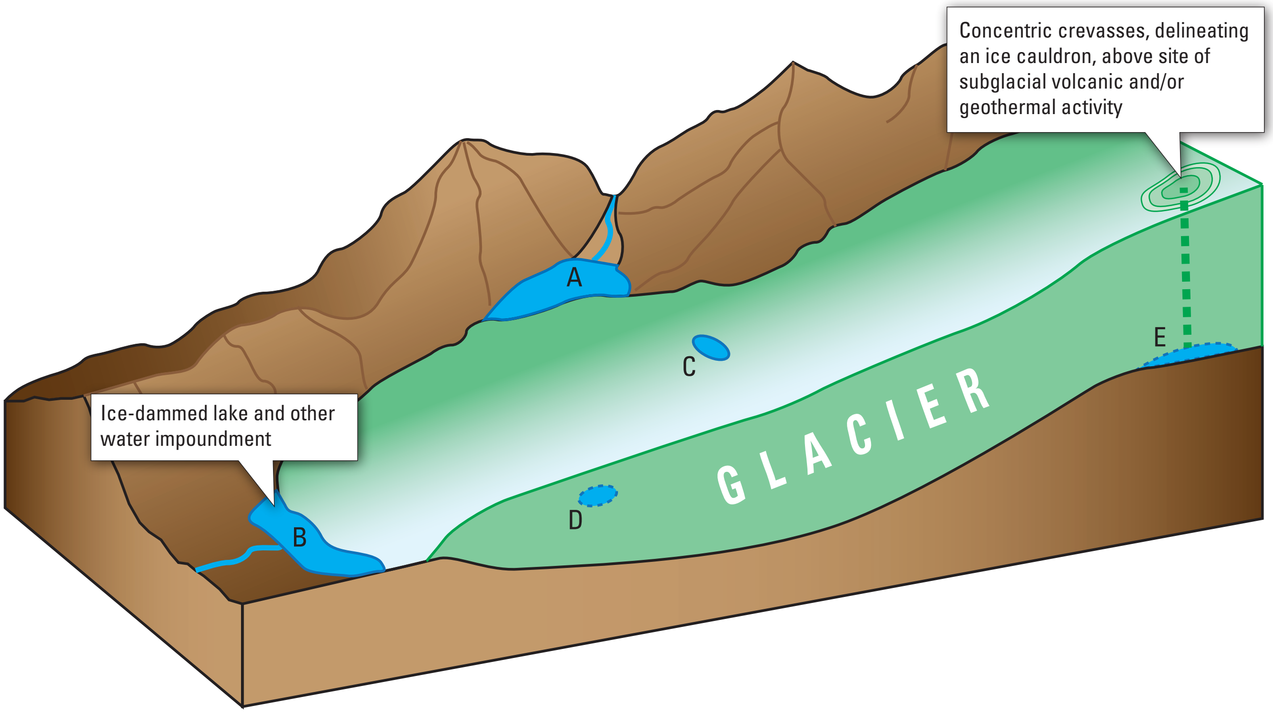

***Figure 86.**—Diagram showing the location of impounded water on, within, under, and adjacent to a glacier; the sudden release of water from such impoundments produces a jökulhlaup.* **A**, *Ice-margin lake caused by damming of a tributary valley or distributary glacier from the main trunk of a valley or outlet glacier;* **B**, *Proglacial lake at the terminus of a valley glacier or outlet glacier from an ice cap or ice field.* **C**, *Supraglacial lake (see fig. 87).* **D**, *Englacial lake.* **E**, *Subglacial lake. Modified from figure in Roberts (2005, p. 3 of 21, fig. 1).*

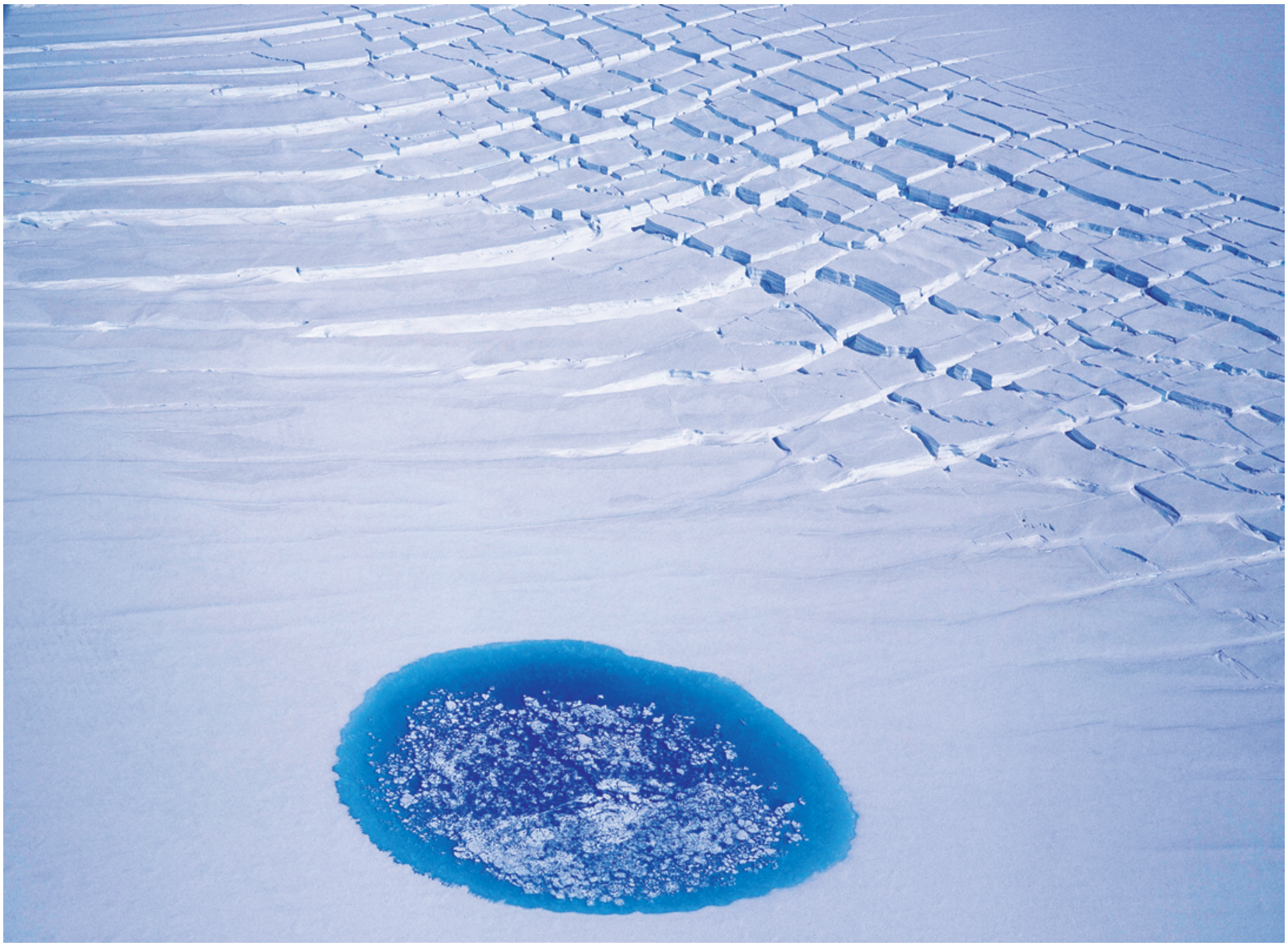

Figure 87.—*Oblique aerial photograph of a supraglacial lake taken on 27 July 1995, one year after the surge of Síðujökull, an outlet glacier of the Vatnajökull ice cap, Iceland. Photograph by Oddur Sigurðsson, Icelandic Meteorological Office.*

Lacustrine-type jökulhlaups have been reported in Antarctica, where subglacial jökulhlaups have resulted in an increase in velocity of the Byrd Glacier during 2005 and 2007 (Fricker, 2008; Stearns and others, 2008). In Greenland, a jökulhlaup from an ice-dammed lake at the margin of Russell Glacier occurred on 31 August 2007 (Mernild and others, 2008). Some scientists have suggested that increased surface melt on the Greenland ice sheet is causing acceleration in velocity of outlet glaciers as the water reaches the base of the ice (Zwally and others, 2002). Das and others (2008) have also implicated sudden drainage (jökulhlaups) of water from supraglacial lakes reaching the bottom through crevasses and moulins.

In Europe, jökulhlaups from both ice-dammed and moraine-dammed proglacial lakes have occurred historically and in modern times (Haeberli and others, 2001). Jökulhlaups from ice-dammed and moraine-dammed proglacial lakes in the Himalaya have also occurred during historic and modern times as mountain glaciers thinned and retreated after the end of the Little Ice Age (late 19th century) (Wilford, 1988), including Nepal (Mool, Bajrachary, and

Joshi, 2001) and Bhutan (Mool, Wangda, and others, 2001). In South America, jökulhlaups have also occurred from proglacial lakes and ice-dammed arms of fjords on the margin of the Southern Patagonian Ice Field, as reported in "Glaciers of South America," Chapter I of this volume. For example, on the western margin of the southern Patagonian Ice Field, an advance of the outlet glacier Glaciar Brüggen (Pío XI) dams an arm of the fjord; when the ice dam fails, a jökulhlaup occurs from emptying of Lago Greve. On the eastern margin of the Southern Patagonian Ice Field is Glaciar Perito Moreno; when it advances across Canal de los Témpanos, it dams the connection between Brazo Rico and Lago Argentino. When the ice dam fails, a jökulhlaup discharges water from Brazo Rico into Lago Argentino.

In the glacierized regions of western Canada, many ice-dammed lakes are known to have been the source of jökulhlaups, such as the ice-dammed Tulsequah Lake, a proglacial lake at the terminus of Tulsequah Glacier (Clarke and others, 1984). Other examples of jökulhlaups from ice-dammed lakes in the region are described in "Glaciers of North America," Chapter J of this volume. Glacier-capped volcanoes in the Pacific Northwest of the United States and in western Canada have the potential for producing jökulhlaups caused by volcanic activity. Jökulhlaups not associated with volcanic activity have been reported from Mount Rainier, Wash. (Richardson, 1968). Driedger (1988) considers the source of the water to be the melting of snow and glacier ice and the downward percolating rainfall that is impounded in englacial or subglacial reservoirs before release.

Jökulhlaups that have occurred in Iceland ("Glaciers of Iceland," Chapter D of this volume) and in Alaska ("Glaciers of Alaska," Chapter K of this volume) have been well documented in the scientific literature in both glacierized regions. Thorarinsson (1939b, 1956), in his review of the historical literature of Iceland and from contemporary accounts, points out that many of the ice-dammed lakes of Iceland have been the source of jökulhlaups. The unnamed distributary glacier from the Skeiðarárjökull outlet glacier on the southern margin of the Vatnajökull ice cap dams and calves into the proglacial lake Grænalón; Grænalón generally produces a jökulhlaup in late summer that drains along the surface of the western margin of Skeiðarárjökull (see fig. 63 in Sigurðsson and Williams, 2008, p. 90). Björnsson (1977), in his comprehensive study of marginal and supraglacial lakes in Iceland, lists 25 such lakes, 13 lateral (such as Grænalón), 7 proglacial (along termini of outlet glaciers), 3 surface lakes around nunataks, and 2 within ice cauldrons; 17 of the lakes have had at least one jökulhlaup. Preusser (1976), from a review of literature on jökulhlaups in Iceland for his doctoral dissertation, describes examples of volcanic and lacustrine jökulhlaups, including the various types of the latter.

The most comprehensive review of ice-dammed lakes and jökulhlaups (both lacustrine and volcanic) in Alaska was published by Post and Mayo (1971). They documented 750 ice-dammed lakes, all of which could be the source of jökulhlaups; they also included descriptions of historic jökulhlaups, including frequent jökulhlaups from Lake George along the Knik River, one of which—the 1961 jökulhlaup— had a peak discharge of 0.01 sverdrup (Post and Mayo, 1976, p. 211). The damming of Russell Fiord, an arm of Disenchantment Bay, by the advance of the tidewater Hubbard Glacier, is discussed in "Glaciers of Alaska," Chapter K, p. K54–K60; p. K70–K74; p. K168–K174, of this volume. In 1986, four years in advance of the forecast by Post and Mayo (1971, p. 6), the Hubbard Glacier closed off the entrance to Russell Fiord, creating a short-lived "Russell Lake." When the ice dam broke, the peak discharge of the jökulhlaup into Disenchantment Bay was 0.1 sverdrup (Mayo, 1989), the largest known lacustrine jökulhlaup in historic and modern time.

Other Local Glaciological Hazards: Ice Avalanches, Debris Avalanches, Lahars, and Landslides and Rockslides

Valley glaciers and outlet glaciers from ice caps and ice fields slowly create over-deepened subaerial and subglacial valleys and steep valley walls, primarily through erosion. Such glaciers also deposit lateral moraines along the valley walls and terminal moraines at their termini. As the glacier thins and retreats under conditions of climate warming, the glacier ice no longer supports the lateral moraines, which become increasingly unstable on the valley walls. Heavy precipitation events or earthquakes can cause rock formations and morainic deposits from the steep valley falls to suddenly detach and slide downslope to the deglacierized valley or onto the surface of glaciers as rock slides in Russia ("Glaciers of Asia," Chapter F), in Iceland (Kjartansson, 1967; Sigurðsson and Williams, 1991) and as debris avalanches in Alaska (see Chapter K, "Glaciers of Alaska," of this volume). Retreat of the glacier upvalley from its terminal moraine often leaves a proglacial lake impounded behind the moraine; additional thinning and retreat of the glacier and rainfall will slowly expand the area and depth of the lake, setting the stage for the terminal moraine to be overlapped, thus causing a jökulhlaup. The recession of glaciers in the western United States and Canada since the late 19th century has created conditions favorable to an increase in these types of glaciological hazards as the valleys become deglacierized (Moore and others, 2009).

Depending on the relief of the terrain, the amount of water impounded, and the availability of unconsolidated glacial and other sediments, the jökulhlaup could be mostly water or it could contain varying percentages of sediment. On glacierized volcanoes, where volcanic tephra is available, melting of the ice cap during a volcanic eruption (for example, Nevado del Ruiz, Colombia (see Chapter I, "Glaciers of South America," p. I11–I13; p. I18–I22; of this volume), especially when combined with heavy rainfall, can cause a debris avalanche of ice, water, and volcanic sediment that can travel many tens of kilometers or more along river valleys descending from the volcano (Naranjo and others, 1986). Unconsolidated material on the flanks of a volcano mobilized by water from a jökulhlaup or from heavy precipitation is called a "lahar," an Indonesian loanword. For example, during the 18 May 1980 eruption of Mount St. Helens, Wash., the initial massive debris avalanche extended downstream along the North Fork Toutle River for more than 27 km; it was followed by a large mudflow formed from melting of glaciers and snow, and it transported a mixture of glacier ice, tephra, glacial sediment, and water that traveled about 85 km downstream along the South Fork Toutle River, the Toutle River, and the Cowlitz River and into the Columbia River, blocking the shipping channels (Brugman and Post, 1981). Numerous debris avalanches from geothermally active sections of the summit crater on Mount Baker, Wash., onto Boulder Glacier have been reported (Frank and others, 1975).

"Glaciers of South America," Chapter I of this volume, notes that the glacierized parts of the Andes, especially proglacial, moraine-dammed lakes in the Cordillera Blanca, Perú, are especially dangerous (Morales-Arnao, 1998). Failure of the moraine dam by overtopping caused by an increase in lake depth from glacial meltwater, ice avalanches into such lakes, and earthquakes have caused many debris avalanches (aluviones) during historic and modern times. Widespread destruction and loss of life downstream are the results. Lliboutry and others (1977) list 35 named and unnamed proglacial lakes in the Cordillera Blanca.

"Glaciers of Asia," Chapter F of this volume, discusses the rock and ice avalanche in the Kolka-Karmadon area of North Osetiya, Russia (Caucasus Mountains) (Kotlyakov and others, 2010). Chapter F also presents various glaciological hazards in the Himalaya. Analyses of images from satellite sensors are being used to assess potential glaciological hazards and to carry out post-event assessments in glacierized regions (Kääb, Wessels, and others, 2003).

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