Figure 21.1 The global climate system as it pertains to glacier responses to and with climatic change.

Figure 21.2 Relationship between regional climate, glacier mass balance and glacial response for terrestrial and tidewater margins. The solid grey line that leads from the evidence of a glacial event/response back to the climate box is the usual intellectual pathway but this ignores the intermediate elements in the climate/glacier interactions. (From Meier, 1965; Andrews, 1975.)

Figure 21.3 Map showing North Atlantic sites (Iceland as an insert—Sigmunes = *S, Stykkisholmur (Sty) B997-328 and -330; W, White Glacier; *H, core HU87-009; *V, core V28-14; *G, Greenland Summit ice cores; P, core PS2644; solid black line = Cockburn Moraines; black squares = convection sites; L and H, the general locations of the Iceland Low Pressure system and the Azores High Pressure cell, which define the North Atlantic Oscillation.

Figure 21.4 (A) Feed-back loops associated with the effect of increased runoff and freshwater export from the Arctic Ocean (from Mysak & Power, 1992). (B) Feed-backs between ice extent, climate, and glacier response in north Iceland (Stotter et al., 1999).

Figure 21.5 Modern Stykkisholmur data. (A) NAO winter index (Hurrell, 1995) versus the January–April average pressure at Stykkisholmur (Fig. 21.3). (B) Winter pressure versus winter precipitation (mm) at Stykkisholmur. (C) ‘Net mass balance index’ (winter precipitation (mm)/summer degree-days versus Stykkisholmur winter pressure). (D) The ‘net mass balance index’ (C) versus net mass balance data from Icelandic glaciers (Dyurgerov, 2002) for the period 1987–2002.

Figure 21.6 Sketch of ice sheet with terrestrial and marine margins and controls on (a) Little Ice Age (LIA) and (b) last glacial cycle.

Figure 21.7 Data from Iceland on Little Ice Age (LIA) time-scales. (A) Mann et al. (1999), (B) sea ice and δ18O in B997-328, (C) Stykkisholmur winter and summer temperature trends, (D) Sigmunes water column temperature and Stykkisholmur mean annual temperature (MAT).

Figure 21.8 (A) Solar insolation and IRD history of East Greenland shelf, (B) 330 temp estimates and a 1470 filter response (note difference in scales), (C) peaks in haematite grains in core V28-14 from Denmark Strait (Bond et al., 1999) (Fig. 21.3) and intervals of moraine formation around north Iceland glaciers (Wastl et al., 2001). The rectangular grey areas represent groups of moraines between inferred cold events.

Figure 21.9 (a) GISP2 δ18O data, with 1470-yr band-pass filter. (b) Solar radiation at 65°N in July, (c) Multitapered (MTM) spectra. Data for (c) and (d) from Blossvæsshaug and Labrador Sea—D–O events. Notice the different age scales. The arrows in (c) and (d) link possible correlative events (see Fig. 21.3 for location of cores).

Figure 21.10 GISP2 data—accumulation (cm yr−1) versus δ18O at 500-yr intervals for the past 50,000 yr. The figures represent the estimated age (cal. kyr) for a data point. The line is the best fit for the correlation between the two variables.

Figure 21.11 (A) δ18O records on benthic foraminifera for the past 0–1 Myr, and (B) 1.5 to 2.5 Myr (see text for source).

Figure 21.21 Map of Iceland showing key locations referred to in the text. Also shown are critical basal marine core dates, evidence of former ice limits and the main ice caps in existence today. The 200 m bathymetric contour is marked for reference and all dates are in k14C yr BP. (See www.blackwellpublishing.com/knight for colour version.)

Figure 21.22 Modelled present-day ice surface using a two-stepped cooling of 2°C for 1000 yr followed by 200 yr at −1°C. Also shown are the modelled ice-sheet extents associated with 2000 yr of cooling of 3, 4 and 5°C under the present precipitation regime and no sea-level change. (See www.blackwellpublishing.com/knight for colour version.)

Figure 21.23 The optimum LGM modelled ice-sheet surface and its flow regime most compatible with the available offshore evidence corresponding to a cooling of 12.5°C and a 35% decrease in precipitation with an additional 30% suppression applied north of the 65th parallel. (See www.blackwellpublishing.com/knight for colour version.)

Figure 21.24 The long profile of Bárðardalur fjord showing the basal topography, its mountain relief, the modelled optimum LGM ice stream and that reconstructed by Norddahl (1991) from tritium evidence. Superimposed are the two long-profiles modelled using the standard value for geothermal heat flux (G = 54.2 mW m−2) applied across the model domain and zero basal sliding (A = 0). (See www.blackwellpublishing.com/knight for colour version.)

Figure 21.25 (a) Modelled basal velocity from the optimum LGM experiment overlain with the orientation of observed ice directional features (adapted from Bourgeois et al., 2000) superimposed with the direction of the corresponding modelled basal vector for the same location. (b & c) Rose diagrams of modelled basal velocity vectors and observed orientations of ice directional features for the two locations indicated. (See www.blackwellpublishing.com/knight for colour version.)

Figure 21.26 Modelled time-series from LGM through to early Holocene of: (a) predicted mean ELA, (b) ice sheet area, (c) ice volume and (d) bulk freshwater runoff.

Figure 21.27 (a) The modelled Younger Dryas ice-sheet geometry and associated flowlines compared with empirical reconstructions for: (b) the northern fjords, (c) the northeast and (d) the southeast based on geomorphological mapping and dating of end moraines, trim lines and raised shorelines (Norddahl & Pétursson, in press). (See www.blackwellpublishing.com/knight for colour version.)

Figure 21.31 Time series of Arctic monthly sea-ice extent anomalies, 12-month running anomalies and least-squares linear fit for the Arctic basin, based on data from 1979 to 2002. (Courtesy of National Snow and Ice Data Center, Boulder, CO.)

Figure 21.32 Time series of the winter (December through to March) index of the Arctic Oscillation from 1950 to 2003 and linear least-squares regression line.

Figure 21.33 Mean annual circulation of the Arctic sea-ice cover, based on data from the International Arctic Buoy Programme, with overlay of sea-level pressure from 1979 to 2002. Contours are shown every 1 hPa.

Figure 21.41 Locations of cores mentioned in the text.

Figure 21.42 Records of climate, IRD and petrological components in IRD: (A) oxygen isotope composition in GISP2 ice core at Summit Greenland; (b–f) climate and IRD records from DSDP site 609. H refers to Heinrich event. Note that in (b) cold peaks point down; in (c–f) cold peaks point up. Dashed lines mark start of Heinrich events as defined by increases in detrital carbonate. Precursory events can be seen in both haematite-stained grains and Icelandic glass. See text and Bond et al. (1999) for details.

Figure 21.43 Sr and Nd isotope composition of IRD in core SU90-09 as fingerprints of IRD sources just before and during Heinrich events 1 and 2. Compositions of layers just below each Heinrich event are precursors and consistent with European sources, including Iceland. Compositions of Heinrich layers are consistent with a Laurentide source. See Grousset et al. (2001) for details.

Figure 21.44 Results from the Kaspi et al. (2004) model suggesting that, owing to an atmospheric coupling of ice sheets as sea ice cools the North Atlantic, millennial oscillations eventually will become phase locked, producing (a) synchronous ice-sheet growth and decay. For certain model parameters (b) ice sheet growth and decay have the same periodicity, but smaller ice sheets lead larger ones, thereby producing precursory-like events seen in the North Atlantic marine records.

Figure 21.45 (a) The precise 1470-yr timing of D–O events from Rahmstorf (2003). Dashed lines mark out exact 1470-yr intervals.
Markers with solid dots denote onset of D–O cycles as defined by warming above a threshold defined by amplitude and rate. (b) Deviation in years of onset of D–O cycles from the exact 1470-yr pacing. (c) Spectral analyses (multi-tapered MTM method) of haematite-stained grain cycles in Fig. 24.2. The heavy line in raw data is the average spectrum; light lines are the upper and lower 90% confidence limits. The heavy dashed line is the mean AR1 (red noise) spectrum; the light dashed line is the upper 90% confidence limit. The prominent peak centred on 1800 yr indicates a broad-band process (not periodic) that is different from red noise. The process could be periodic, however, and is distorted in the record by errors in the marine age model.

**Figure 24.6**  The record of haematite-stained grains from the Holocene in cores MC52 and VM29-191. Vertical lines mark out an exact 1470-yr cycle. The dashed line is the result of filtering the haematite-stained grain record with a Gaussian filter centred on 1470 yr and bandwidth of 1000 to 2000 yr. That note that although the Holocene cycle pacing is not exactly 1470 yr, it falls within the same range as its glacial counterpart in Fig. 24.5c.

**Figure 24.7**  Model results from Braun et al. (2005) using the CLIMBER-2 model forced with the DeVries solar cycle (2100 yr) and the Gleissberg solar cycle (88 yr) under glacial boundary conditions. Vertical dashed lines demark an exact 1470-yr pacing. Even though weak, the solar forcing in the model produces ocean surface-temperature anomalies of several degrees C with an exact pacing of 1470 yr.

**Figure 25.1**  Northern hemisphere glacier mass balance sensitivity to annual air temperature. Long-term annual mass-balance time series averaged for about the same 40 bench mark glaciers have been used to calculate averages (Dyurgerov, 2001). Note, this is a different measure of sensitivity than that used by the IPCC (Church et al., 2001b), which involves a change between two steady states.

**Figure 27.1** Basal melt rates, B, seaward of Antarctic (hollow circle) and Greenland (solid circle) grounding lines, versus thermal forcing, AT, from the ocean, which is the difference between the nearest in situ ocean temperature data and the seawater freezing point at a depth of 0.88 times the maximum grounding line ice thickness. The regression indicates that a 1°C increase in effective ocean temperature increases melt rate by 10 myr$^{-1}$. PIG, Pine Island; THW, Thwaites; SMI, Smith; KOH, Kohler; DVR, DeVicq; LAN, Land; BYR, Byrd; DAV, Davie; NIN,annis; MEK, Mertz; TOT, Totten; DEN, Denman; SCO, Scott; LAM, Lambert; SHI, Shirase; JUT, Jutulstraumen; STA, St. Anthony; WILLS; SLE, Slessor, REC, Recovery; INS, Institute; RUT, Rutford; CAR, Carlson; EVA, Evans; OBT, Ostenfeld; PET, Petermann; RYD, Ryder; ZAC, Zacharias Iström; 79N, 79 north glaciers.

**Figure 28.1**  Map of North America and Greenland showing four ice sheets at their maximum extent in the last glaciation. Oscillations in the various routings of precipitation and ice-melt runoff from the southern part of the Laurentide Ice Sheet between 18 ka and 7 ka and their oceanic impact are the subject of this chapter. (See www.blackwellpublishing.com/knight for colour version.)

**Figure 28.2** A similar view as in Fig. 28.1 showing four continental watersheds which played significant roles in directing runoff to the oceans. These watersheds are the Mississippi Valley catchment draining to the Gulf of Mexico, the Great Lakes–St Lawrence Valley catchment discharging to the North Atlantic Ocean, the Athabasca–Mackenzie Valley catchment draining to the western Arctic Ocean, and the Hudson Bay catchment discharging via Hudson Strait to the northern Atlantic Ocean (western Labrador Sea). Outlets that were significant controls on runoff routing include Chicago (A) from southern Lake Michigan basin, Wabash Valley (B) from western Lake Erie basin, Mohawk Valley (C) connecting Lake Ontario basin with Hudson River Valley, Hudson River Valley (D) draining to Long Island Sound and the Atlantic Ocean, Hudson Shelf Valley (E), a possible extension of the Hudson Valley route (R. Thielar, personal communication, 2004), southern Lake Agassiz outlet (F), first eastern Agassiz outlet (G), northwestern Agassiz outlet (H), second eastern Agassiz outlet (I), and Kinojevis outlet (J) to Ottawa and St Lawrence River valleys. See Fig. 28.4a for names and locations of Great Lakes basins. (See www.blackwellpublishing.com/knight for colour version.)

**Figure 28.3**  History of Lake Agassiz baseline runoff and outburst floods to oceans from southern LIS after Licciardi et al. (1999) and Teller et al. (2002). Triangles illustrate Lake Agassiz outburst flood fluxes assuming each lake draw down was completed in 1 yr after the opening of a lower outlet by ice retreat. Flood discharges are added to baseline runoff. (a) Total runoff showing ice melt and precipitation components plus outburst floods through all routes. (b) Total (ice melt + precipitation) runoff plus outburst floods (trianles) to the North Atlantic Ocean via first the Hudson River and after 11 ka (13,000 cal. yr) the St Lawrence Valley routes; YD and horizontal arrows indicate period of the Younger Dryas cold event. In an alternative scenario the large Agassiz flood shown at the onset of the Younger Dryas event would not have discharged by this route, and baseline runoff would have been reduced (thin grey line) (see also Fig 28.3d). (c) Total (ice melt + precipitation) runoff via the Mississippi Valley route to Gulf of Mexico. Note the strong antiphased relationship in discharge with that of the route via Hudson River–St Lawrence. (d) Total (ice melt + precipitation) runoff to western Arctic Ocean. The two outburst floods (black triangles representing floods after 12,000 cal. yr) uncovered first eastern (lower) outlet (G) to Superior basin in the Great Lakes, thereby initiating a large outburst flood as the lake draw down was shown to the other outlet. The flood discharge and Agassiz basin runoff were diverted to the North Atlantic Ocean via the Great Lakes–St Lawrence route. Alternatively, if ice retreat was sufficient, Agassiz discharge may have switched first to the southwestern outlet as in Fig. 28.5c. (c) By 10 ka (11,400 cal. yr) advanced of the Superior ice lobe (Marquette advance, MA) had closed the first eastern Agassiz outlet and diverted the Agassiz basin runoff briefly back to the Mississippi Valley route to Gulf of Mexico. Then, with opening of the northwestern outlet (H) to the Athabasca–Mackenzie Valley as ice retreated along the southwestern margin of LIS, runoff and a draw-down outburst flood were directed to the western Arctic Ocean. (See www.blackwellpublishing.com/knight for colour version.)
Figure 28.6 Interaction of ice retreat, opening of lower outlets, and differential rebound resulting in the switching of Lake Agassiz discharge from one outlet to another (after Teller et al., 2001). See text for explanation.

Figure 28.7 Schematic diagram of ocean–ice interaction and oscillatory switching of Mississippi runoff to and from Hudson River Valley (after Clark et al., 2001). See text for explanation.

Figure 28.8 Schematic diagrams of oscillations of Lake Agassiz area, level and outflow routings as a result of ice retreat, climatic feedbacks and differential rebound. In these diagrams, the external forcing of relatively high summer insolation which drove general ice retreat (Kutzbach et al., 1998) is represented in box 1 with the heavy border. (a) Lake–atmosphere–ice interactions suggest decreased precipitation over the adjacent ice sheet as the lake enlarges, based on numerical modelling by Hostetler et al. (2000). This model accounts for ice retreat and the first opening of eastern outlets following early growth of Lake Agassiz. (b) Following concepts in Teller (1987), a large proglacial lake possibly provided more melt and nourished growth of the adjacent ice sheet. See text for further explanation. (c) Possible feed-backs and effects using the modelling results of Krinner et al. (2004) for Eurasian glacial lakes. Large lakes induced cool climate that suppressed summer ice melting, and thereby enhanced growth of the adjacent ice sheet. Positive feedbacks to lake enlargement as in Figs 28.8B and 28.8C may account for ice readvances such as the Marquette Advance (Fig. 28.5c) which closed the eastern outlet about 10 ka (11,400 cal. yr).

Figure 28.9 Curve illustrating the response of North Atlantic Deep Water (NADW) formation to the introduction of an arbitrary volume unit of freshwater that reduces surface salinity at a time when there is a specific ‘Rate of NADW formation’ (after Stocker, 1996; Bond et al., 1999). For example, if salinity was reduced by 1 unit on the ‘Salinity’ scale, when NADW formation was high (point a), there would have been only a small response in NADW formation rate (point b). This situation is like the present interglacial mode, which is relatively insensitive to freshwater additions. If salinity was reduced by 1 unit when NADW formation was low and salinity was also relatively low (point c), as probably occurred during a full glacial, there would only be a small response in the rate of NADW formation (point d). In contrast, if freshwater additions reduced the salinity by 1 unit when salinity was near the threshold value at point b, it would reduce the rate of NADW formation dramatically to point c. This probably is what the North Atlantic Ocean was like during the transition from glacial to interglacial conditions, and may have led to the large change in NADW formation associated with the Younger Dryas that is attributed to the 9500 km$^3$ outburst from Lake Agassiz.

Figure 29.1 Gauge locations and corresponding hydrological record durations.

Figure 30.1 Schematic diagram of a glacier-influenced continental margin. (Modified from Dowdeswell et al., 2002b.)

Figure 30.2 Morphology of high-latitude continental margins from swath-bathymetric data. (a) Sediment scarp marking a palaeo-grounding line formed during the retreat of ice offshore of the Larsen Ice Shelf, Antarctica (arrowed) (modified from Evans et al., 2005). (b) Irregular pattern of scour produced by the keels of drifting icebergs impinging on sea-floor sediments. (i) Iceberg scours on the East Greenland continental shelf offshore of the Scoresby Sund fjord system. (ii) Iceberg scours on the continental shelf west of the Antarctic Peninsula. Note the irregular pattern of the scours in plan view and the morphological contrast with the subglacially produced, megascycle glacial lineations (arrowed) to either side.

Figure 30.2 Continued (d) Submarine channels in the Greenland Basin, east of Greenland at 73°–76°N. (Modified from Ó Cofaigh et al., 2004.)

Figure 30.3 Oblique view of the Bear Island Fan, Norwegian–Svalbard margin, with debris flows observed on 6.5 kHz long-range side-scan sonar imagery shown. (Modified from Taylor et al., 2002.)

Figure 31.1 Location of the West Shetland margin; bathymetric contours in metres.

Figure 31.2 (a) Sea-bed relief image of the outer shelf illuminated from the northwest illustrating the northern area of morainal ridges of various scales; (b) sea-bed relief image illuminated from the northeast over the southern end of the West Shetland Slope and Faroe–Shetland Channel, illustrating the debris flows of the Rona Wedge, and the smooth to irregular (Judd Deeps) floor of the Channel; (c) BGS high-resolution seismic (1 kJ sparker) profile (79/14-23) across the large (M1–M3) and small (m1–m3) morainal ridges illustrated in (a); (d) BGS high-resolution seismic (1 kJ sparker) profile (83/04-31) and interpreted line drawing of the morainal ridges in the continental environment (Environment Canada, 2003). Abbreviations: TWTT, two-way travel time; msecs, milliseconds; SBP, sea-bed pulse.

Figure 31.2 Continued (c) Irregular pattern of scours produced by the keels of drifting icebergs impinging on sea-floor sediments. (i) Iceberg scours on the East Greenland continental shelf offshore of the Scoresby Sund fjord system. (ii) Iceberg scours on the continental shelf west of the Antarctic Peninsula. Note the irregular pattern of the scours in plan view and the morphological contrast with the subglacially produced, megascycle glacial lineations (arrowed) to either side.

Figure 32.1 Vertical profiles of relative humidity. Measured values represent half-hourly averages. The line labelled ‘Snowdrift’ was taken at 9 January, 1800–1830 hours, during the peak of the storm. The other, labelled ‘No drift’, was taken at 11 January, 2030–2100 hours, during a period with weak winds and no snowdrift. The level $z_{sat}$ is indicated at the right vertical axis for the ‘Snowdrift’ profile.

Figure 32.2 Temporal variation of relative humidity, in per cent (at 5 cm and at 2 m), wind speed, in m s$^{-1}$ (2 m), $z_{sat}$, in cm, transport rate, in $10^2$ kg m$^{-1}$ s$^{-1}$, snowdrift and surface sublimation, in W m$^{-2}$, during a 5-day period in January 1998.

Figure 32.3 Dependence of snowdrift and surface sublimation rates on wind speed (2 m). Symbols represent grouped data, and the error bars represent the standard deviations of the average values.

Figure 34.1 Daily mean values of the surface energy balance derived from an automatic weather station on the tongue of Morteratsch glacier for the year 2000.

Figure 34.2 Mean modelled annual surface energy flux ($\chi$), net shortwave radiation ($S_{net}$), longwave radiation ($L_{net}$), sensible ($H_{se}$) and latent ($H_{le}$) turbulent heat flux averaged over 100 m height intervals for the year 2000.

Figure 34.3 Modelled mean specific mass balance, ablation and snow accumulation for 100 m height intervals for the year 2000 (solid) and for a 1°C increase in air temperature (dotted).

Figure 35.1 $d$-8D diagram in meteoric ice from the Vostok ice-core. (Reproduced by permission of American Geophysical Union from
Souchez et al. (2002). Copyright American Geophysical Union.

Figure 35.2  $d$–$\delta D$ diagram in lake ice from Taylor Valley, Antarctica. Numbers are increasing with depth. (Reproduced by permission of American Geophysical Union from Souchez et al. (2000). Copyright American Geophysical Union.)

Figure 35.3  Double-diffusion freezing at the grounding line in Terra Nova Bay. (a) Sketch of the suggested mechanism. (b) $\delta D$–$\delta ^{18}O$ diagram of the ice samples. Black circles, ice samples; open circles, initial water samples computed from $\delta$ values of ice samples and equilibrium fractionation coefficients. The straight line represents the best-fit line for the waters. Insert shows a diagram for glacier ice samples in the area. (Reproduced by permission of International Glaciological Society from Souchez et al. (1998). Copyright International Glaciological Society.)

Figure 35.4  $\delta D$–$\delta ^{18}O$ diagram for different ice types in Beacon Valley. (Reproduced by permission from Marchant et al. (2002). Copyright Geological Society of America.)

Figure 36.1  Vostok source region temperatures derived from the 400-kyr-old East Antarctic ice-core record (lower panel). Following the classic approach for interpreting the deuterium excess, $T_d$ assumes that the effective source temperature can be deduced from the excess alone. However, $T_{\text{source}}$ takes into account the combined effects of condensation-site temperature changes and source temperature changes. In the upper panel the difference between the two computations is shown indicating that these differences are growing and shrinking following the glacial–interglacial 100kyr cycle.