Figure 56.1  Activation energy as a function of the inverse of temperature, based on the results of ice-deformation tests at temperatures near the melting point by Morgan (1991).

Figure 56.2  Creep curves (log–log plots of octahedral strain rate as a function of octahedral strain) for ice-deformation tests in various combinations (indicated) of compression and shear. The combined octahedral stress for all tests was 0.4 MPa. The test temperature was −2.0°C. (After Li et al., 1996.) Reproduced by permission of the International Glaciological Society.

Figure 56.3  (a) A series of creep curves (normalized to a minimum strain rate of 1, and with a dashed line indicating a flow enhancement of 3) and crystal orientation fabrics (after Jacka & Li, 2000) from laboratory experiments at −19°C in which compressive stress ranges from 0.8 to 0.1 MPa. (b) A series of creep curves (normalized to a minimum strain rate of 1, and with a dashed line indicating a flow enhancement of 3) and crystal orientation fabrics (after Jacka & Li, 2000) from laboratory experiments at 0.2 MPa compressive stress in which temperature ranges from −5.0 to −21°C. Reproduced by permission of Hokkaido University Press.

Figure 56.4  Crystal mean area plotted as a function of microparticle concentration. Data represented by squares are from ice core DSS, on Law Dome, East Antarctica, and samples are from within the Holocene depth interval. Solid triangles represent DSS Last Glacial Maximum (LGM) samples. Circles represent samples from the beginning of the last glacial period. The cross is a datum from a Dye 3 (Greenland) LGM sample. (After Li et al., 1998.) Reproduced by permission of the International Glaciological Society.

Figure 57.1  (a) Calculated elevation profiles compared with measured elevations for the three representative transects described in the text. (b) Root-mean-square mismatch of calculated and measured elevations, averaged for the three profiles, as a function of stress exponent. For each n value, each profile is least-squares optimized by adjusting softness parameters as described in the text.

Figure 57.2  Relations between steady-state ice volume per divide length (i.e. area of cross-sectional profiles), effective viscosity, stress exponent and two specific controls on viscosity (temperature and enhancement). (a) Black diamonds indicate estimated positions for the three modern ice sheets, according to representative profiles: EA, East Antarctic-type; WA, West Antarctic-type; G, Greenland-type. West Antarctic volumes are those above the flotation level (hence sea-level equivalent). (b and c) Volumes have been normalized to values at 0°C and at E = 1. For 'WA, basal’ the effective temperature or enhancement of basal shear has been varied. For ‘WA, margins’ only the effective temperature or enhancement of the ice-stream shear margins has been varied.

Figure 57.3  Dependence of WAIS-type volume per length on ice stream width (2w), at three different effective temperatures for the shear margins.

Figure 57.4  Dependence of approximate characteristic response-times on effective viscosity, stress exponent, and one control on viscosity (temperature), shown in panel c. Black diamonds in (a, b, c, and d) are estimates for modern EAIS and GIS. The response-time to accumulation-rate change is that for which ice surface slope does not change, corresponding to a situation wherein the ice margin is free to expand. The comparable response times for fixed margins are smaller, by a factor of (n + 2)/(2n + 2). These response times neglect covariant change in temperature, which will slow all responses.

Figure 57.5  Relations between steady-state ice volume per divide length and other variables for a Greenlandic-type ice sheet with a flat bed, allowing for feed-back between accumulation rate and topography. Black squares indicate points for which the calculation was performed. In panel (b), enhancement values are indicated corresponding to the equivalent softness change induced by the temperature change as shown.

Figure 58.1  Modelled steady-state isochrons for an idealized divide with a thickness of 1000 m and accumulation rate of 0.1 cm yr−1, resembling Siple Dome, West Antarctica. The crossover stress used in the model increases from left to right, as reflected in the diminishing size of the isochron arch. Ωchar is the non-dimensional ratio τchar/k. (From Pettit & Waddington, 2003, reprinted from the Journal of Glaciology with permission of the International Glaciological Society.)

Figure 58.2  Horizontal (top panel) and vertical (bottom panel) velocity fields from the best-fitting model assuming steady state. For both panels, Bindschadler Ice Stream is to the right and Kamb Ice Stream is to the left. The dashed lines are velocity contours.

Figure 58.3  Stratigraphic record of Ice age based on basal slip in ice crystals and for creep of isotropic polycrystalline ice at −10°C. (a) Temperature θ = −3°C (from Jacka & Maccagnan, 1984). (b) Temperature θ = −15°C (from Jacka & Li, 2000).

Figure 58.4  Photographs of thin-sections in polarized light and ice fabrics from the Greenland Ice Core at 2806 (top) and 2860 m (bottom) (from De la Chapelle at al., 1998).

Figure 60.1  Schematic diagram depicting four ice-creep regimes: Dislocation creep, characterized by n = 4.0 and p = 0, GBS-limited basal-slip creep, with n = 1.8 and p = 1.4, basal-slip-limited GBS creep, with n = 2.4 and p = 0, and diffusion creep, with n = 1 and p = 2 or 3, depending on whether volume or grain-boundary diffusion, respectively, is rate-limiting. Creep data for the n = 2.4 creep regime for fine-grained ice are in excellent agreement with creep data for single crystals of ice oriented for basal slip (e.g. Wakamah, 1967). The heavy solid line represents the composite flow behaviour of ice given by Equation (3).

Figure 60.2  Plot comparing the composite flow law of Equation (4) (solid lines) with previous laboratory data for coarse-grained ice samples, for T = 268 K. The upper solid line is calculated for d = 0.2 mm, the lower for d = 2 mm. The dotted line represents the Glen flow law; the dotted–dashed line represents data from experiments conducted at high confining pressure (Durham et al., 1992). Data points are from ambient pressure tests: ◆, d = 0.2 mm, Goldsby & Kohlstedt (2001); □, d ≥ 1 mm, Steinemann (1998a); ○, d ≥ 1 mm, Mellor & Smith (1966); ▲, d ≥ 1 mm, Barnes et al. (1971).

Figure 60.3  Deformation map for ice constructed from the flow laws for GBS-limited creep and dislocation creep for a temperature of 268 K. The heavy solid line of negative slope is the boundary between mechanisms. The dotted–dashed lines are strain-rate contours, calculated using the appropriate flow law for the rate-limiting creep mechanism in each creep regime. The box labelled ‘Glen’ bounds the approximate θ versus d conditions of Glen’s experiments (Glen, 1952, 1955). The box with dashed lines on three sides outlines the full range of stresses and grain sizes in the Goldsby and Kohlstedt experiments (most of which were conducted at temperatures < 268 K); the solid vertical line on the right edge of this box marks the approximate range of stresses for experiments on samples with grain sizes of ca. 0.2 mm at 268 K (see fig. 3 of Goldsby & Kohlstedt, 2001). The approximate range of stresses in glaciers and ice sheets is shown by the large rectangle.

Figure 60.4  Plot comparing the composite flow law of Equation (4) (solid line) with data from field studies on glaciers and ice sheets. The upper solid line is calculated for d = 2 mm and T = 273 K, the lower solid line for d = 1 mm and T = 255 K. The dotted-line parallelagram represents the range of conditions of the experiment on the Barnes Ice Cap by Hooke (1973), which yielded n = 1.65. Data points are from: ◆, Gerrard et al. (1952), with data reanalysed as in Nye (1953); ○, Meier (1960); ▲, Gow (1963); □, Holdsworth & Bull (1970).

Figure 60.5  Plot of shear stress and depth versus grain size for the deep ice-core from Byrd Station, Antarctica. Shear stresses τ were esti-
mated from the depth \( h \) using \( \tau = \rho gh \sin(\alpha) \), where \( \rho \) is density, \( g \) is gravitational acceleration and \( \alpha \) is the surface slope (ca. 2.5 \times 10^{-3} \), after Frost & Ashby, 1982). Grain-size versus depth data (solid symbols connected schematically with the solid curve) are from Gow et al. (1968). Superimposed on the figure is a shear stress versus grain size deformation mechanism map constructed for a temperature of ca. 273 K. The heavy solid line is the boundary between GBS-limited creep and dislocation creep; the dotted–dashed lines are strain-rate contours. Temperatures measured at discrete depths along the borehole (from Gow et al., 1968) are indicated in the figure.

Figure 61.1 Deformation in uniaxial compression where polygranization and migration recrystallization are active: (a) the fabric after 0.5 vertical strain, where the initial fabric was isotropic, and (b) normalized strain rate versus strain.

Figure 61.2 Normalized strain rate for different levels of nearest neighbour interaction (NNI) in uniaxial compression and simple shear as a function of cone angle. Isotropic ice has a cone angle of 90° and ice with all crystals aligned has a cone angle of 0°.

Figure 61.3 Normalized strain rate in combined uniaxial compression (\( \sigma \)) and simple shear (\( \tau \)) stress state as a function of stress ratio (\( \tau/\sigma \)) and cone angle: (a) vertical strain rate and (b) shear strain rate (Thorsteinsson & Waddington, 2002).

Figure 62.1 Schematic crack velocity versus stress intensity factor. Critical crack growth (CCG) occurs for \( K \geq K_c \), where \( K \) is the stress intensity factor and \( K_c \) is the fracture toughness. Subcritical crack growth (SCCG) arises for \( K_{th} \leq K < K_c \), with \( K_{th} \) being the threshold of the stress intensity factor.

Figure 62.2 Simulation of crevasse formations in the west face hanging glacier of the Eiger, Switzerland. According to Pralong et al. (2003), the damage is assumed to be isotropic. The crevasses appear in the model as concentration of damage. The dark grey colour represents ice without microcracks (\( D = 0 \)). Light grey corresponds to broken ice (\( D = 1 \)), i.e., ice with a very high density of microcracks. In the simulation, the frontal crevasse grows. The fracture process zone at its tip is depicted in black and corresponds to intermediate values of \( D \). (See www.blackwellpublishing.com/night for colour version.)

Figure 63.1 Schematic diagram illustrating the process of formation of distinctive basal regelation ice by interaction of the glacier at its basal boundary with bed roughness elements. This process, which contributes to basal motion, is known as Weertman (1957) regelation, and is probably most effective for bed roughness elements with dimensions \( 10^{-1}–10^{3} \) m. It produces layers of basal ice with distinctive physical and chemical characteristics.

Figure 63.2 Englacial ice facies overlying an erosional unconformity with distinctively layered debris-laden basal stratified-ice facies in the basal zone at Taylor Glacier, Antarctica.

Figure 63.3 Examples of field evidence for the lower effective viscosity of different ice facies in situ: (a) Competence-contrast boudining at Variegated Glacier, showing the relatively high effective viscosity of the cleaner boudinaged dispersed facies ice, with stratified facies ice layers deformed around it. (b) Velocity profiles in the basal zone at Suess Glacier, showing relatively rapid deformation in the ‘amber’ (dispersed) facies basal ice (from Fitzsimons et al., 2000). (c) Displacement profile in a marginal cliff at Taylor Glacier, Antarctica, showing rapid displacement rates in the lower part of the debris-laden basal ice section.

Figure 63.3 Continued

Figure 63.4 The results of various studies showing the effect of the presence of solid impurities on the creep rate of ice, in relation to the creep rate of debris-free ice. (From Budd & Jacka, 1989.)

Figure 63.5 Temperature dependence of uniaxial compressive strength for (a) debris-laden basal stratified-ice facies and (b) englacial-ice facies from Taylor Glacier, Antarctica. Note that the envelope of strength defined by the maximum value of strength at each increase systematically as temperature decreases for the debris-laden basal stratified-ice facies.

Figure 64.1 Photomontage and structural interpretation of the section through the moraine complex at the Leverett Glacier. Numbered arrows show the location of the key units, thick black lines delineate their boundaries, and thin white lines highlight the orientation and extent of structures within them. Ice flow was from left to right.

Figure 64.2 Structures reflecting glacier–permafrost interactions in western Arctic Canada: (A) pinch-and-swell structure of sand, North Head (trowel for scale) (69°43′N, 134°26′W); (B) ice clasts within glaciotectonite, Pullen Island (area depicted ca. 1.5 m wide) (69°46′N, 134°25′W); (C) buried basal ice, Mason Bay (face is ca. 4 m high) (69°33′N, 134°02′W); (D) ice dyke–sill truncating a composite wedge, North Head (spade for scale).

Figure 65.1 Ice composition, structure and deformation at the base of Suess Glacier in the Taylor Valley, Antarctica.

Figure 65.2 Photograph of the stratified ice facies in the basal ice zone of Suess Glacier 26 m into the tunnel. The solid facies occurs at the roof of the tunnel at this location.

Figure 65.3 Boundinage structures 1.7 m above the base of Suess Glacier. The boudins have formed from a 10–mm-thick layer of sand. The glacier flow direction is from right to left.

Figure 65.4 A folded layer of gravelly sand 1.0 m above the bed of Suess Glacier. The heads of the bolts are 10 mm in diameter and the glacier flow direction is from right to left.

Figure 65.5 Sheared recumbent folds 0.8 m above the base of Suess Glacier. The glacier flow direction is from right to left.

Figure 65.6 Photograph of a 4 m shaft excavated through part of the basal zone of Wright Lower Glacier showing layers of sand interbedded with ice. The glacier flow direction is from top to bottom.

Figure 65.7 The basal ice solid facies with well-preserved planar bedding in Wright Lower Glacier.

Figure 65.8 Crack in the solid facies in Wright Lower Glacier. The glacier flow direction is from right to left.

Figure 65.9 Ice from above and below intruding into a crack in the solid facies in Wright Lower Glacier. The glacier flow direction is from right to left.

Figure 65.10 Photograph of a tunnel in the left margin of Taylor Glacier showing the ‘key-hole’ structure produced from more rapid deformation associated with debris-rich ice. The tunnel walls were vertical initially. Photograph taken 11 months after the tunnel was excavated. The glacier flow direction is from the ladder toward the reader.

Figure 65.11 Circular strain markers deformed into ellipses after 10 days in debris-rich basal ice in Taylor Glacier. The glacier flow direction is from left to right.

Figure 66.1 Instrument used to record ice displacement. The steel support is ca. 50 cm high.

Figure 66.2 Results for site 1: (a) ice velocity for anchor P3; (b) air temperature. Gaps in the record occur after each complete turn of the potentiometer, when the resistance returns to zero.

Figure 66.3 Results for site 2: (a) displacements of anchors P2 and P3; (b) air temperature.

Figure 66.4 Stick-slip motion at site 2. Part of the velocity time series for (a) the upper anchor P3 and (b) the lower anchor P2.

Figure 67.1 Long section of Tianfleuor Glacier, Switzerland, illustrating its constituent ice zones (UZ = Upper Zone; LZ = Lower Zone; BZ = Basal Zone) as reconstructed from core characteristics. (After Hubbard et al. (2003) with the permission of the International
Fig. 67.2 Results of the modelled response of Tsanfleuron Glacier to a 75 m rise in ELA. The current measured glacier surface profile is given as a solid line and the modelled steady-state profile is given as a dashed line: (a) multilayer rheology model and (b) single-layer rheology model. (After Hubbard et al. (2003) with the permission of the International Glaciological Society.)

Fig. 67.3 Measured relationship between horizontal velocity of a surface stake and water level in a borehole at Findelengletscher, Switzerland. (After Iken & Bindschadler (1986) with the permission of the International Glaciological Society.)

Fig. 67.4 Displacement of initially vertical segmented rods emplaced for ca. 5 days in subglacial sediments beneath Breiðamerkurjökull, Iceland. (After Boulton & Hindmarsh (1987) with the permission of the American Geophysical Union.)

Fig. 67.5 Deformation profiles in glacier ice (a) above a basal zone of low traction corresponding to the location of a major melt-season basal channel and (b) above adjacent ice characterized by higher basal traction. (After Willis et al. (2003) with the permission of the International Glaciological Society.)

Fig. 68.1 Map of the study site on Lauteraagletscher, showing the locations of the surface and borehole measurement sites and the strain grid. The two ellipses are strain rate ellipse during the indicated time periods of the days from 22 to 27 August 2001. These ellipses represent the deformation of the circle (broken line) subjected to the computed surface strain rate in units of $10^{-4}$ day$^{-1}$. The dash–dot line at the top indicates the water level corresponding to the ice overburden.

Fig. 68.2 Records of water level in a borehole, horizontal flow speed, vertical displacement of the surface relative to the mean elevation, borehole length and air temperature. The dash–dot line at the top indicates the water level corresponding to the ice overburden.

Fig. 68.3 Englacial flow fields in a longitudinal cross-section computed for the times indicated in the figures. (a) Horizontal flow speed ($m$ day$^{-1}$), (b) Vertical strain rate ($10^{-4}$ day$^{-1}$). Positive value indicates tensile strain rate.

Fig. 69.1 Modelled horizontal surface velocity across Haut Glacier d’Arolla at 70 m resolution for $n = 3$ and $A = 0.063$ yr$^{-1}$ bar$^{-3}$. The contours are plotted at 2.5 m yr$^{-1}$ intervals. Over lain within circles are the winter 1995 velocity vectors and a bivariate analysis of modelled against observed velocities.

Fig. 69.2 (a) The magnitude and orientation of modelled surface-parallel principal stresses at 140 m resolution; inward compression, outward arrows indicate extension. The shaded areas represent zones of maximum surface stress and indicate areas of potential ice failure. (b) The distribution of surface crevasses across Haut Glacier d’Arolla as observed from aerial photography and ground mapping.

Fig. 69.3 (a) The relative magnitude and orientation of surface-parallel modelled and (b) measurement derived principal stresses and strains, respectively, in the region of the high-density strain network indicated in (a).

Fig. 69.4 (a) Distribution of measured annually averaged (August 1995 to August 1996), horizontal velocity within a half cross-section at Northing 91700. (b) The modelled composite velocity distribution within the glacier cross-section composed of time–weight averages of 20/52 ‘winter’ no sliding, 31/52 ‘normal summer’ sliding and 1/52 enhanced ‘spring event’ sliding.

Fig. 69.5 Comparison of modelled and observed annual surface velocity distribution from 1995 to 1996 overlain on the subglacial channel network reconstructed from hydraulic potential analysis (Sharp et al., 1993).

Fig. 69.6 Modelled temporal-evolution of Haut Glacier d’Arolla surface centre-line in 10 yr intervals from the mapped 1880 (maximum) extent through to the glacier’s predicted demise in 2070. Observed 1940 and 1992 long-profiles are overlain for reference.

Fig. 70.1 Satellite image (AVHRR) of the Siple Coast region in West Antarctica (A) and a velocity map (B) for those parts of this region where RADARMAP interferometry data are available (Joughin et al., 2002). The satellite map shows names of the major ice streams and of other important glaciological features (source: http://igloo.gsfc.nasa.gov/wais/articles/images/ismap2.jpg). Velocity scale is logarithmic. Velocity data provided by Dr I. Joughin (JPL-Caltech).

Fig. 70.2 Sensitivity of ice-stream velocity to changes in ice-stream width (A) and basal or driving stress (B) based on Equation (1a–c). The range of ice-stream widths in (A) covers tributaries (ca. 20 km), ice plains (ca. 100 km) and ice-stream trunks in between (Joughin et al., 1999; Raymond, 2000; Tulaczyk et al., 2000b). The stress difference in (B) is defined as the difference between driving stress and basal resistance. The stress change rate represents variations in either driving stress or basal resistance. Ice accelerations result from an increase in driving stress or a decrease in basal resistance. Velocity changes are plotted for the case of acceleration (positive per cent per year) but equivalent decelerations can be obtained by simply switching signs. The velocity change isolines are plotted on a linear scale in (A) and on a logarithmic scale in (B).

Fig. 71.1 Flowline-model domains, from divides down Mercer Ice Stream (A), Whillans Ice Stream (B), Kamb Ice Stream (C) and Bindschadler Ice Stream (D) and across the Ross embayment to the continental-shelf edge. Topographic data from BEDMAP data set (sponsored by the Scientific Committee on Antarctic Research and coordinated by the British Antarctic Survey, Cambridge). (Based on fig. 1 in Parizek et al., 2002, 2003.)

Fig. 71.2 History of basal heat budget for flowlines (A), (B), (C) and (D), respectively. Total heat budgets are calculated assuming catchment sheet is twice as wide as stream. Shaded regions indicate a positive budget. (Based on fig. 3 in Parizek et al., 2002, 2003.)

Fig. 71.3 Last Glacial Maximum (22.1 ka) and modern ice-sheet reconstructions with isotherms and basal melting indicated for flowlines C (with grounding-line forcing) and D (without grounding-line forcing). (Based on fig. 2 in Parizek et al., 2002, 2003.)