

The Microstructure of Polar Ice.

Part I: Highlights from Ice Core Research[☆]

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Abstract

Polar ice sheets play a fundamental role in Earth's climate system, by interacting actively and passively with the environment. Active interactions include the creeping flow of ice and its effects on polar geomorphology, global sea level, ocean and atmospheric circulation, and so on. Passive interactions are mainly established by the formation of climate records within the ice, in form of air bubbles, dust particles, salt microinclusions and other derivatives of airborne impurities buried by recurrent snowfalls. For a half-century scientists have been drilling deep ice cores in Antarctica and Greenland for studying such records, which can go back to around a million years. Experience shows, however, that the ice-sheet flow generally disrupts the stratigraphy of the bottom part of deep ice cores, destroying the integrity of the oldest records. For all these reasons glaciologists have been studying the microstructure of polar ice cores for decades, in order to understand

[☆]Dedicated to the memory of Sigfús Jóhann Johnsen (1940-2013).

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the genesis and fate of ice-core climate records, as well as to learn more about the physical properties of polar ice, aiming at better climate-record interpretations and ever more precise models of ice-sheet dynamics. In this Part I we review the main difficulties and advances in deep ice core drilling in Antarctica and Greenland, together with the major contributions of deep ice coring to the research on natural ice microstructures. In particular, we discuss in detail the microstructural findings from *Camp Century*, *Byrd*, *Dye 3*, *GRIP*, *GISP2*, *NorthGRIP*, *Vostok*, *Dome C*, *EDML*, and *Dome Fuji*, besides commenting also on the earlier results of some pioneering ventures, like the *Jungfraujoch Expedition* and the *Norwegian–British–Swedish Antarctic Expedition*, among others. In the companion Part II of this work (Faria et al., this issue), the review proceeds with a survey of the state-of-the-art understanding of natural ice microstructures and some exciting prospects in this field of research.

Keywords: ice, glacier, ice sheet, mechanics, creep, recrystallization, grain growth, microstructure, fabric, texture

1. Introduction

Ice is one of the oldest known minerals (Adams, 1990; Faria and Hutter, 2001) and manifests itself in diverse forms, most commonly as snow, frost, hail, icicles, ice plates, permafrost, firn, and massive polycrystals. Although it is neither as ubiquitous as quartz nor as precious as diamond, ice is highly regarded by its environmental and economic importance, as well as by the exceptionally large deposits of “pure” ice found in continental-sized polar ice sheets (the impurity content of polar ice typically lies in the ppb range; Legrand and Mayewski, 1997). These ice sheets cover virtually all Greenland and Antarctica with more than $2.7 \times$

10 10^{16} m³ of ice, corresponding to ca. 2.5×10^{19} kg of freshwater, or 64 m of sea level
11 rise equivalent (Lemke et al., 2007).

12 Like any usual crystalline solid, ice undergoes creep at sufficiently low stresses
13 and temperatures higher than around half of its pressure melting point (Petrenko
14 and Whitworth, 1999; Durham et al., 2001). Seeing that temperatures naturally
15 occurring on Earth generally lie within that range, it should be no wonder for con-
16 temporary scientists to witness glaciers and ice sheets creeping slowly under their
17 own weight. Notwithstanding, more often than not one still can find expositions
18 in the modern literature attributing the creep of glaciers and ice sheets to an odd
19 fluidity of ice. Such a pseudodoxy is nourished by the charm of the old glaciolog-
20 ical literature (beautifully described by Clarke, 1987 and Walker and Waddington,
21 1988), ancient beliefs (Adams, 1990; Faria and Hutter, 2001), and the long list of
22 real peculiarities of this material, which range from its abnormally low mass den-
23 sity to the persistence of brittle properties up to its melting point (Hobbs, 1974;
24 Petrenko and Whitworth, 1999; Schulson and Duval, 2009).

25 While the creep of large ice masses can itself be considered an unsurprising
26 phenomenon, the microscopic mechanisms that drive it are far from trivial and
27 have been challenging scientists for several decades. Here we review some of
28 these studies, with special emphasis on polar ice from deep ice cores, and present
29 an up-to-date view of the modern understanding of natural ice microstructures and
30 the deformation processes that may have produced them.

31 This work is divided in two correlated publications. Here in Part I, we re-
32 view the advances in the research on natural ice microstructures during the last
33 eight decades, using deep ice cores from Antarctica and Greenland to draw the
34 storyline. In the companion Second Part (Faria et al., this issue) —from now on

35 called *Part II*— we discuss several aspects of our current understanding of nat-
36 ural ice microstructures, including deformation mechanisms, induced anisotropy,
37 grain growth and recrystallization, among others. The whole review ends with a
38 summary of key concepts in the form of a glossary, for quick reference (Appendix
39 A of Part II).

40 For the sake of brevity, we concentrate attention here to a limited number
41 of ice cores only, which we consider most representative of the advances in ice
42 microstructures occurring in a given period. Inevitably, in some situations we
43 have faced the dilemma of choosing between two or more cores equally relevant
44 within the same period. In such cases we have given preference to the core with
45 the largest amount of information available for us. Admittedly, this pragmatic
46 attitude generates a selection bias towards those ice coring projects we have been
47 directly or indirectly involved with. Information about other important polar ice
48 cores, not discussed here (e.g. Law Dome, Taylor Dome, Siple Dome, Talos
49 Dome, WAIS, NEEM and others), is available in the review by Bentley and Koci
50 (2007) and in the Ice Core Gateway of the U.S. National Oceanic and Atmospheric
51 Administration (NOAA; <http://www.ncdc.noaa.gov/paleo/icecore>), among other
52 resources.

53 Summaries of the most relevant microstructural, geophysical, and geographi-
54 cal data about the ice cores discussed here are given in Table B.1 and Figs. A.1–
55 A.3.

56 **Remark 1.** For the description of ice cores we adopt here the convention *from top*
57 *to bottom*, unless explicitly specified otherwise. In usual cases of ordered stratig-
58 raphy, this convention implies inverse chronological order, viz. *from younger to*
59 *older*. It is in this sense that a phrase like “transition from the Holocene to the

60 Last Glacial” may appear, indicating the fact that the Last Glacial is older than the
61 Holocene. Climatologists may feel a bit uncomfortable with this convention, but
62 it is the most logical choice for describing the physical features of an ice core.

63 **2. Early research in natural ice microstructures**

64 It is usually a great injustice to attribute a scientific innovation to a single person,
65 team, or publication. Nevertheless, such a regrettable act is often justified by the
66 fact that the human mind cannot easily grasp history unless the latter is reduced to
67 a plain timeline decorated with milestones. In this vein, we apologetically commit
68 such an injustice here by naming milestones that, in our opinion, exemplify well
69 scientific trends in decisive periods of ice microstructure research.

70 *2.1. The Jungfrauoch Expedition*

71 We start with a field expedition that has not only boosted research in ice mi-
72 crostructures, but also marked a turning-point in the way Glaciology is organized
73 today. Gerald Seligman, a former businessman and skillful ski-mountaineer, was
74 president of the Ski Club of Great Britain and author of an influential treatise on
75 snow structure (Seligman, 1936). That work motivated him to consider the role
76 of ice microstructure in the metamorphism of snow into ice. With this aim he led
77 in 1937 a pioneering party to study this process on the Jungfrauoch, Switzerland,
78 which included John D. Bernal, F. Philip Bowden, T. P. Hughes, Max F. Perutz
79 and Henri Bader (Remark 2).

80 **Remark 2.** It is impossible to overestimate the importance for modern Glaciol-
81 ogy of the constellation of scientists involved in the Jungfrauoch Expedition.

82 Bernal discovered (together with Ralf H. Fowler) the essential principles that de-
83 termine the arrangement of atoms in the ice lattice (Bernal and Fowler, 1933),
84 nowadays known as the *ice rules*. Bowden and Hughes laid the foundations of our
85 modern understanding of the frictional behavior of snow and ice (Bowden and
86 Hughes, 1939; Bowden, 1953). Perutz became one of the pioneers of the modern
87 (non-Newtonian) theory of ice creep (Perutz, 1948, 1949, 1950a,b, 1953). Finally,
88 Bader joined his Ph.D. supervisor Paul Niggli in the Swiss Snow and Avalanche
89 Commission as snow crystallographer in 1935, soon turning into one of the key
90 proponents of a permanent laboratory for snow and avalanche research in Davos,
91 Switzerland, which quickly evolved (in 1943) to the renowned Swiss Federal In-
92 stitute for Snow and Avalanche Research, SRF (Achermann, 2009). Bader left
93 Switzerland prior to SRF's inauguration, however, moving to the Americas in
94 1938 to become, among other things, an international prime mover of polar deep
95 ice coring (Bader, 1962; see also de Quervain and Röthlisberger, 1999; Langway,
96 2008). Seligman, on the other hand, was named in 1936 President of the newly-
97 founded Association for the Study of Snow and Ice, which after the World War II
98 hiatus evolved to the British Glaciological Society (publisher of the influential
99 Journal of Glaciology) and in 1962, still under Seligman's lead, to the (Internat-
100 tional) Glaciological Society.

101 The results of the Jungfraujoeh Expedition have been published in four papers,
102 describing various aspects of the crystallography, metamorphism, mechanics and
103 thermodynamics of snow, firn and ice (Perutz and Seligman, 1939; Hughes and
104 Seligman, 1939a,b; Seligman, 1941). As commented by Seligman (1941) in his
105 general review of the Expedition:

106 The work of earlier investigators and my own had traced the transition

107 of new powdery snow into hard firn snow, but no one had systemati-
108 cally studied how this white, air-filled firn turned into the blue air-free
109 ice of the lower glaciers. This was the ground of the present research.
110 Glacier movement had been supposed to play a part, and this had to
111 be investigated, including of course the flow of the névé. My long-
112 cherished desire to use polarized light to reveal the detailed develop-
113 ment of firn and ice crystals required the help of a crystallographer,
114 which led to unexpected and valuable results. With the exception of
115 a few desultory photographs polarized light had never been used: a
116 surprising omission in glaciological research.

117 Details of these crystallographic investigations on the Jungfraujoeh have been
118 described by Perutz and Seligman (1939). Firn and ice samples were collected
119 from the walls of crevasses or from grottoes and pits dug in the accumulation and
120 ablation zones of the Great Aletsch Glacier and its surroundings. They prepared
121 thin sections and determined crystalline orientations using a technique described
122 by Bader et al. (1939) for snow studies. Among other results, Perutz and Seligman
123 (1939) noticed a conspicuous microstructural contrast between the “small regular”
124 crystallites of firn and the “large irregular” grains of ice. They observed a lattice
125 preferred orientation in the upper meters of firn, with *c*-axes lying perpendicular
126 to the glacier surface and gradually giving way to more isotropic (“random”) *c*-
127 axis distributions below a few tens of meters of depth. In the deeper ice, however,
128 strong lattice preferred orientations could again be observed, suggesting that the
129 effect of glacier flow on the ice microstructure could be to some extent compared
130 to the mechanism of high-temperature creep in other polycrystalline materials,
131 e.g. magnesium (Remark 3). In particular, in places where the ice was subjected

132 to shear, the ice crystallites were oriented with their basal planes parallel to the
133 direction of shear.

134 **Remark 3.** Comparisons between the mechanisms of high-temperature creep in
135 ice and other polycrystalline materials would later pave the way for the painstaking
136 mechanical tests conducted by John W. Glen (1952, 1955) and Samuel Steinemann
137 (1954, 1958), which confirmed the suggestion by Perutz (1949, 1950b) that
138 the flow of glaciers could be modeled by a power law, nowadays known as *Glen's*
139 *flow law*. It is worth noticing that Glen was a Ph.D. student under supervision of
140 Egon Orowan and Max Perutz in Cambridge, while Steinemann was a Ph.D. student
141 under supervision of Paul Niggli and Ernst Brandenberger at the ETH Zurich.

142 According to Seligman (1941), Perutz proposed that grain growth in glaciers
143 could come about through a process of dynamic recrystallization, in which “softer”
144 grains well oriented for simple shear have lower free energy and grow at the expenses
145 of “harder” grains that cannot yield to the imposed stresses.

146 After World War II, several studies similar to those performed by the Jungfrau-
147 joch party were conducted on various glaciers (e.g. Ahlmann and Droessler, 1949;
148 Seligman, 1949; Bader, 1951; Rigsby, 1951, 1958, 1960). These investigations
149 contributed to enriching the records of glacier microstructures, introducing new
150 details, diversity, and complexity to the picture. They failed, however, to provide
151 a consistent description of the microstructural evolution of natural ice. One crucial
152 reason for this failure derives from the fact that the analyzed ice samples had
153 in general no clear spatial or historical relation to each other, being usually collected
154 from distinct pits and similar superficial excavations in the ablation zone of
155 glaciers. From these investigations it soon became evident that a systematic study

156 of natural ice microstructures could only be accomplished by extracting an ice
157 core from the heart of a natural large ice body. Such an enterprise was however
158 a formidable prospect for post-war scientists. New mechanical drilling technolo-
159 gies, specific for ice, had to be developed and the logistics of all equipment and
160 research teams would have to be carefully planned and tested.

161 2.2. *The first shallow ice cores*

162 Eventually, in 1949 two independent international teams set off to distant global
163 locations to start drilling the *first two polar ice cores* for glaciological studies.
164 During the Norwegian–British–Swedish Antarctic Expedition (NBSAE) of 1949–
165 1952, Valter Schytt (1958) and colleagues recovered an ice core of nearly 100 m
166 from the Maudheim site on Quar Ice Shelf, Dronning Maud Land, Antarctica
167 (Remark 4). Nearly simultaneously, within the 1949–1950 activities of the Juneau
168 Ice Field Research Project (JIRP), Henri Bader cored to almost 100 m into the
169 temperate Taku Glacier in Alaska (Miller, 1954; Langway, 2008). Both drilling
170 actions proved to be extremely difficult, and the quality of the recovered ice cores
171 was precarious. Notwithstanding, some physical properties of parts of these cores
172 could be analyzed.

173 In particular, Schytt (1958) studied the crystallography of the whole Maud-
174 heim ice core in depth intervals of approximately 5 m, therefore producing *the first*
175 *microstructural investigation of deep polar ice and of an ice shelf*. He observed
176 a smooth transition of firn into ice at 60–65 m depth, but a clear discontinuity in
177 grain growth with depth below ca. 70 m, with grain sizes increasing six times
178 faster with depth than in the upper 70 m. He interpreted this discontinuity as the
179 boundary between ice produced by in-situ accumulation and ice supplied by the
180 inland ice sheet. In the petrographic analysis, single and multiple maxima could

181 be identified in the c -axis distributions of samples from distinct depths, with no
182 general trend towards a well-established preferred orientation with depth.

183 **Remark 4.** During NBSAE's first winter, drilling was also performed by Bertil
184 Ekström (Schytt, 1958). Unfortunately, by the end of the season Ekström and
185 other two companions, Leslie Quar and John Jelbart, drowned in a track-driven
186 vehicle accident (Mills, 2003). On account of this fatality, three ice shelves around
187 Maudheim Station have been posthumously named after them.

188 **3. The first polar deep ice cores: IGY sites, Camp Century, Bird Station, Dye** 189 **3**

190 After the difficulties faced by the JIRP and NBSAE teams with the pioneering
191 ice cores drilled in Alaska and Antarctica, as well as the subsequent (and equally
192 problematic) drilling campaign on *Central Greenland* by the Expéditions Polaires
193 Françaises, EPF, in 1950–1951 (Langway, 2008), glaciologists in the whole world
194 became aware of not only the great potential, but also the great hurdles of deep
195 ice coring.

196 *3.1. IGY ice cores*

197 Fortunately, the approaching of the Third International Polar Year (IPY) in 1957–
198 1958, which was soon renamed the International Geophysical Year (IGY), helped
199 stimulating the interest in big scientific enterprises in polar regions. Indeed, the
200 U.S. National Academy of Sciences (NAS) Committee for the IGY soon adopted
201 deep core drilling into polar ice sheets for scientific purposes as one of its high-
202 priority, long-term research projects, and subsequently the National Science Foun-
203 dation (NSF) tasked the U.S. Army Snow, Ice and Permafrost Research Estab-
204 lishment (SIPRE), under the leadership of Chief Scientist Henri Bader, with the

205 responsibility for defining, developing, and conducting the entire U.S. ice core
206 drilling and research program under a joint interagency agreement (Bader, 1962).

207 As reported by Langway (1970, 2008), the SIPRE pre-IGY pilot drilling tri-
208 als were conducted at Site-2, Northwest Greenland in 1956 (305 m) and 1957
209 (411 m), being closely followed by two IGY core drillings in Antarctica, the first
210 at Byrd Station, in 1957–1958 (307 m) and the second at Little America V, on the
211 Ross Ice Shelf, in 1958–1959 (264 m). This was a period of great technological
212 improvements not only in drilling, but also in analytical methods (see e.g. Gow,
213 1963a,b; Langway, 1970). The success of the IGY drilling campaigns and the
214 increasing quality of the recovered cores motivated NAS to assign SIPRE with
215 the task of developing a post-IGY deep ice coring system capable of reaching
216 bedrock depths. The outcome of this post-IGY project was a series of celebrated
217 ice cores drilled by B. Lyle Hansen and his team, two of them reaching bedrock
218 in Greenland (Camp Century) and Antarctica (Byrd Station), respectively.

219 3.2. *Camp Century*

220 The first deep polar ice core to *reach the base* of a polar ice sheet was retrieved
221 from Camp Century, Northwest Greenland, in 1963–1966 (after two unsuccessful
222 attempts in 1961–1963) and achieved a final length of 1375 m (Hansen and Lang-
223 way, 1966). For the standards of that time, the physical quality of the core was
224 very good, allowing the first *continuous record* of structure and chemical com-
225 position of a polar ice sheet, stretching from surface to bedrock. More than this,
226 it delivered the definite proof that the combination of ice core drilling with oxy-
227 gen isotope analysis was indeed a valuable method for reconstructing Earth’s past
228 climate (Dansgaard et al., 1969).

229 Measurements of grain sizes and *c*-axis orientations started on the field, in

230 1961, but a thorough microstructural analysis of the whole core was accomplished
231 and published only 16 years later (Herron and Langway, 1982; Fig. A.2; a prelim-
232 inary crystallographic investigation of the bottom 16 m of Camp Century's debris-
233 laden basal ice appeared somewhat earlier, viz. Herron and Langway, 1979). Circa
234 50 horizontal and six vertical thin sections, covering the whole Camp Century core
235 at variable depth intervals, were prepared for crystallographic studies by section-
236 ing thick samples with a microtome. Grain sizes were usually measured from
237 photographs using a semi-automatic particle size analyzer for detecting cross-
238 sectional areas, whereas in difficult cases (e.g. sections contained too large or
239 too complex grains) this method was replaced by counting crystallites within a
240 given area. Crystalline *c*-axis orientations were measured on a Rigsby univer-
241 sal stage (essentially an enlarged version of the conventional four-axis universal
242 stage, especially designed for the larger crystallites found in natural ice; Rigsby,
243 1951, 1958) and presented in a variety of ways, from contoured pole figures to
244 resultant directional vectors and statistical parameters derived from eigenvalues
245 and -vectors.

246 In the upper hundreds of meters of the Camp Century core Herron and Lang-
247 way (1982) observed a thirty-fold increase in the average grain cross-sectional
248 area to more than 100 mm² at 700 m (\approx 3 kaBP, according to Dansgaard and
249 Johnsen, 1969), with grain shapes turning gradually more complex and interlock-
250 ing. Below 850 m the average grain size decreases to less than 60 mm² at 1000 m
251 depth, followed by a drastic size reduction to ca. 2 mm² within a very short depth
252 interval (1136–1149 m depth), which coincides with the climatic transition from
253 the Holocene interglacial to the Last Glacial period (interglacial–glacial transition;
254 Dansgaard and Johnsen, 1969). This sudden reduction in grain size is eventually

255 followed by a gradual increase to about 20 mm² at 1300 m depth, which abruptly
256 gives way to an extremely fine-grained (ca. 0.6 mm²) debris-laden ice at the bot-
257 tom 10 m of the core (Herron and Langway, 1979).

258 Preferred *c*-axis orientations were identified to evolve with depth towards a
259 strong vertical single maximum at the bottom of the core, with a marked enhance-
260 ment within the depth interval 1136–1149 m corresponding to the interglacial–
261 glacial transition. The fine-grained and highly oriented crystallites in the lowest
262 10 m of the core suggest a zone of high deformation on a frozen bed, which is con-
263 sistent with estimated temperature of –13°C at the ice–bedrock interface (Hansen
264 and Langway, 1966; Herron and Langway, 1979).

265 3.3. Byrd Station

266 After successfully finishing core retrieval at Camp Century in July 1966, the same
267 party headed for south and started core drilling at Byrd Station, Antarctica, in
268 November 1966. In less than two field seasons, Hansen and his team managed to
269 recover a total core length of 2164 m, reaching bedrock in January 1968. Shortly
270 after, however, good luck turned its back on them, as they lost their valued drill
271 rig stuck in frozen *subglacial water*, which upwelled into the hole while the drill
272 was pinching the bed (Ueda and Garfield, 1970). Fortunately, the entire ice core
273 was already retrieved and safe, and could provide the most complete portrait of
274 Antarctic ice to that date.

275 Gow and Williamson (1976) performed the crystallographic analysis of the
276 Byrd deep ice core (Fig. A.3). The methods of microstructural investigation were
277 generally similar to those employed on the Camp Century core (Sect. 3.2). From
278 the firn–ice transition zone at 56 m depth down to ca. 600 m (\approx 5.5 kaBP, accord-
279 ing to Hammer et al., 1994) they observed a twenty-fold increase in the average

280 grain cross-sectional area, with the average grain size stabilizing at about 60 mm².
281 Concomitantly, the regular polygonal grain structure just below the firn–ice transi-
282 tion gradually gives way to a complex structure of interlocking grains, frequently
283 showing undulose extinction and similar manifestations of lattice distortion. At
284 1200 m depth the core reaches the glacial–interglacial transition and the grain
285 size stability breaks down with a marked three-fold decrease in grain size within
286 a depth interval of only 100 meters. The resulting fine-grained structure persists
287 for further 500 m, in a zone characterized by intense ash layers and widespread
288 *cloudy bands* (Fig. A.4 and Appendix A of Part II). Below 1600 m depth the fine-
289 grained structure starts becoming disturbed by interdigitations of coarse-grained
290 ice, which eventually overrides the ice microstructure beneath 1800 m depth, with
291 increasingly large crystallites reaching sizes of several thousands of mm² at the
292 bottom of the core.

293 The depth development of *c*-axis preferred orientations in the upper 1800 m
294 of the Byrd deep ice core follows roughly that of Camp Century: a gradual but
295 persistent formation of a vertical single maximum. By analyzing the microstruc-
296 ture of deep ice in greater detail, Gow and Williamson (1976) discovered a *con-*
297 *sistent relation between grain size, c-axis preferred orientations, and impurity*
298 *content*, such that the higher the impurity content, the smaller are the grains and
299 the stronger is the vertical single maximum. As a consequence, the fine-grained
300 cloudy bands in the depth range 1200–1800 m of the Byrd core are generally
301 associated with a strong single-maximum *c*-axis distribution, while the *c*-axis pre-
302 ferred orientations of the coarse-grained ice, intermixed in that depth range and
303 pervasive below 1800 m depth, are characterized by multiple maxima.

304 In many aspects, the Byrd deep ice core established new standards for our

305 understanding of the physics and microstructures of polar ice sheets. First, the
306 observed general evolution of grain sizes and *c*-axis orientations with depth estab-
307 lished the basis for the (overused) *tripartite paradigm* of polar ice microstructure,
308 also known as the “three-stage model” (cf. Sect. 5 and Appendix A of Part II; the
309 formulation below follows De la Chapelle et al., 1998):

- 310 1. in the upper hundreds of meters of an ice sheet, grains grow in the regime
311 of Normal Grain Growth (NGG; Stephenson, 1967; Gow, 1969);
- 312 2. in intermediate depths, NGG is counterbalanced by grain splitting via “poly-
313 gonization” (Alley et al., 1995);
- 314 3. at the bottom of the ice sheet, where the ice temperature raises above ca. -10°C ,
315 dynamic recrystallization with nucleation of new grains (SIBM-N) markedly
316 transforms the microstructure (Duval et al., 1983).

317 Second, the highly oriented fine-grained structure of the impurity-rich glacial
318 ice in the depth interval 1200–1800 m suggested that horizontal simple shearing
319 is considerably strong in that zone. This finding prompted a question, colloqui-
320 ally epitomized by the title of Stan Paterson’s (1991) article, which has pervaded
321 ice core studies ever since: “*Why is glacial ice sometimes soft?*” Actually, the
322 first step towards answering this question has been taken by Gow and Williamson
323 (1976) themselves. They reported the existence and basic properties of *cloudy*
324 *bands* (see Appendix A of Part II), and identified them as one of the major strati-
325 graphic features of glacial ice. They noticed also that the fine-grained structure
326 and high anisotropy of such bands disclose them as localized zones of intense
327 shearing, which may possibly be major contributors to the flow of the ice sheet.
328 Such extensive shearing along discrete strata situated well above bedrock could
329 cause differential layer thinning and seriously distort the stratigraphy, making the

330 dating and interpretation of climate records extremely complicate. Today, cloudy
331 bands continue to challenge our understanding of ice mechanics and microstruc-
332 ture, with novel methods of observation and modeling casting new light on this
333 issue (Takata et al., 2004; Lhomme et al., 2005; Svensson et al., 2005; Gow and
334 Meese, 2007; Faria et al., 2009, 2010).

335 Finally, the danger of unexpected *subglacial water upwelling* into the borehole
336 would not only become a recurrent source of troubles for future deep ice core
337 drillings (see next sections), but also a presage of the unexpected extension and
338 dynamics of the subglacial hydrologic environment (Clarke, 2005; Siegert, 2005;
339 Evatt et al., 2006; Wingham et al., 2006).

340 3.4. Dye 3

341 The successful operations at Camp Century and Byrd Station proved that core
342 drilling down to the bedrock through several kilometers of creeping polar ice was
343 feasible, and that the physical and environmental information recorded in ice cores
344 was invaluable. These results motivated researchers from Denmark, Switzerland
345 and the United States to meet in 1970 in order to plan a new major research pro-
346 gram for ice core drilling in Greenland, named GISP: the Greenland Ice Sheet
347 Program. Originally, GISP was a very ambitious eleven-year program involving
348 three deep ice core drillings down to bedrock, but budgetary restrictions forced
349 the program to reduce deep bedrock drilling to only one location, the Summit, in
350 North-Central Greenland (Langway, 2008). Eventually, however, further finan-
351 cial restrictions compelled the selection of a logistically more convenient site in
352 Southern Greenland, at the U.S.A.F. Distant Early Warning Radar Station Dye 3
353 (Dansgaard et al., 1982). Drilling started at Dye 3 in 1979, after seven years of
354 preliminary field and laboratory studies, and in 1981 the newly designed Danish

355 electromechanical drill ISTUK touched bedrock at 2037 m. Several on site labora-
356 tories (including two equipped science trenches and a clean-room trailer) and new
357 processing procedures established *new standards of organization and efficiency*
358 for deep ice core field studies.

359 Vertical thin sections were sampled by Herron et al. (1985) on site, at approx-
360 imately 100 m depth intervals throughout the core, and prepared them for crystal-
361 lographic analyses following the procedures already adopted in previous ice core
362 studies (e.g. Herron and Langway, 1982). Average grain sizes were determined
363 using the intercept method. Crystalline *c*-axis orientations were measured at 23
364 selected depths using a Rigsby universal stage and were presented in a variety of
365 ways, following nearly the approach already adopted in the Camp Century studies
366 (cf. Sect. 3.2). These *c*-axis observations were also compared with the results of an
367 alternative method for monitoring material anisotropy through ultrasonic velocity
368 measurements of selected ice core samples.

369 Herron et al. (1985) observed (cf. Fig. A.2) a ten-fold increase in the aver-
370 age grain cross-sectional area to ca. 30 mm² at 800 m (\approx 2 kaBP, according to
371 Reeh, 1989), followed by a size reduction in the next 100 m and subsequent grain
372 size stabilization around an average cross-sectional area of 16 mm². Finally, at
373 the interglacial–glacial transition at ca. 1785 m depth (Dansgaard et al., 1982;
374 Gundestrup and Hansen, 1984), the average grain size sharply reduces to less than
375 0.5 mm² within some tens of meters, and then resumes its growth trend with depth
376 down to bedrock, reaching ca. 5 mm² at the bottom of the core (where the tem-
377 perature is around -13°C ; Gundestrup and Hansen, 1984). A general tendency
378 to horizontally elongated grains was observed throughout the core, especially in
379 coarse-grained ice (where the grain aspect ratio can reach 1.3).

380 Crystallographic and ultrasonic measurements of the Dye 3 core revealed a
381 trend similar to previous deep ice cores, especially the Byrd Station core, with a
382 steady reorientation of *c*-axes towards vertical and a marked vertical single max-
383 imum below the interglacial–glacial transition at 1785 m depth. More detailed
384 grain size and *c*-axis measurements conducted by Langway et al. (1988) in glacial
385 ice from 1785–2037 m depth showed that the strong vertical single-maximum *c*-
386 axis distribution persists throughout this lower portion of the core, with grain sizes
387 varying between 0.2 and 7 mm². Smaller grains were found in high-impurity lay-
388 ers and, conversely, larger grains were found in low-impurity strata. In contrast to
389 the Camp Century and Byrd cores (cf. Sects. 3.2 and 3.3), Langway et al. (1988)
390 reported that, in the Dye 3 core, *impurity content* seemed to have a strong influ-
391 ence on grain sizes, but less of an effect on *c*-axis preferred orientations.

392 **4. News from Greenland: GRIP, GISP2, NGRIP**

393 While U.S. polar deep drilling operations could be successfully performed since
394 the late 1950's, thanks in part to exclusive scientific programs organized by the
395 National Academy of Sciences (NAS) and the National Science Foundation (NSF),
396 the nations of post-war Europe had first to organize themselves in a stable politico-
397 economical framework, in order to allow the creation of exclusive European pro-
398 grams capable of financing such complex and expensive scientific enterprises. In
399 this vein, the 1970's and 1980's constituted a period of remarkable changes in the
400 European scientific landscape. The first United Nations Conference on the Envi-
401 ronment, held in Stockholm in 1972, motivated the European Commission (EC) to
402 launch its first Environment Action Program (EAP), the earliest of a series of five-
403 year action programs for dealing with critical environmental issues. In 1974 the

404 European Science Foundation (ESF) was created, and in 1986 the ESF launched
405 its Polar Science Network Program.

406 These specific European programs for climate and environment established
407 the grounds for the creation of successful European deep drilling projects in po-
408 lar regions, through collaborative funding schemes involving the EC, ESF, and
409 several national funding agencies.

410 *4.1. GRIP*

411 The decisive contributions of Denmark and Switzerland to the success of GISP
412 led European glaciologists to propose to ESF the creation of a long term pro-
413 gram for promoting glaciological research. In 1988 the ESF agreed and launched
414 the European Glaciological Program (EGP). The first project within this program
415 was the Greenland Ice Core Project (GRIP), which aimed at drilling to bedrock a
416 deep ice core at the highest point of the Greenland Ice Sheet, the Summit (the site
417 originally selected for GISP, cf. Sect. 3.4), for investigating the climatic and envi-
418 ronmental changes of the past 250,000 years (GRIP community members, 1996).
419 Nearly at the same time, a U.S. companion project called GISP2 would pursue
420 similar objectives at a site just 27 km to the west (cf. Sect. 4.2).

421 Funding of GRIP came initially from national funding agencies of the eight
422 participating European nations (Denmark, Switzerland, France, Germany, United
423 Kingdom, Italy, Iceland and Belgium). This was soon complemented by finan-
424 cial support of the European Commission under the European Program on Cli-
425 matology and Natural Hazards (EPOCH). Drilling and logistic operations were
426 coordinated by the GRIP Operation Center (GOC), which was established for this
427 purpose at the Geophysical Institute of the University of Copenhagen. Drilling
428 started in summer 1990, using an updated version of the ISTUK drill, and stopped

429 in July 1992, after penetrating through 6 m of debris-laden (silty) ice just above
430 bedrock, at a depth of 3028.8 m below surface (Johnsen et al., 1994). Unfor-
431 tunately, due to severe stratigraphic disturbances caused by the ice flow in the
432 lowest 10% of the core, reliable dating has been limited to depths ca. 300 m above
433 bedrock (≈ 110 kaBP; Peel, 1995; Landais et al., 2003), although tentative chrono-
434 logical reconstructions of the disturbed bottom ice do exist (Landais et al., 2003;
435 Suwa et al., 2006).

436 More than 60 vertical and horizontal thin sections were sampled on site at ir-
437 regular intervals, ranging from 10 to 115 m in the upper 770 m, and from 25 to
438 55 m in the rest of the core (Thorsteinsson et al., 1997). The samples were pre-
439 pared for crystallographic analysis following the already standard methods used
440 in previous ice core studies. Further sampling of core depths of special interest
441 was done later, at the storage facility in Copenhagen.

442 Average grain sizes were measured directly, mainly from vertical thin sec-
443 tions, using the linear intercept method. Crystalline *c*-axis orientations were de-
444 termined mostly from horizontal thin sections using a semi-automatic Rigsby uni-
445 versal stage (Lange, 1988). The results were analyzed by a special software and
446 presented in a variety of ways, from point scatter pole figures to median inclina-
447 tions and statistical parameters derived from eigenvalues and -vectors.

448 Thorsteinsson et al. (1997) observed (cf. Fig. A.2) a steady and regular de-
449 velopment of preferred *c*-axis orientations with depth towards a single vertical
450 maximum distribution, which is compatible with the stress regime in an ice dome,
451 viz. dominated by uniaxial vertical compression. In contrast to the Camp Century
452 and Byrd cores (cf. Sects. 3.2 and 3.3), no significant strengthening of the single
453 maximum distribution could be recognized at the interglacial–glacial transition

454 depth.

455 GRIP's grain size development with depth, as observed by Thorsteinsson et al.
456 (1997), are comparable to those previously reported for Camp Century, Dye 3 and
457 Byrd: an eight-fold increase in average grain cross-sectional area below 100 m
458 depth to ca. 10 mm² at 700 m depth (\approx 3.5 kaBP, according to Dansgaard et al.,
459 1993), followed by a stable mean grain size in the remaining part of the Holocene
460 interglacial ice. At the interglacial–glacial transition the average grain size re-
461 duces to half, and continues to decrease with depth to ca. 3 mm² at 1980 m. Fur-
462 ther down, grain size starts to moderately increase again, reaching ca. 15 mm²
463 at 2790 m depth, in early glacial ice close to the transition to the Eemian inter-
464 glacial. In the bottom 250 m of the core, where the climate records are disturbed
465 by the ice flow (Taylor et al., 1993; Peel, 1995; see also Sect. 4.2), the average
466 grain size varies dramatically between less than 12 mm² and more than 300 mm²
467 (Thorsteinsson et al., 1995), revealing a conspicuous correlation with impurity
468 concentration changes (which in turn are related to climatic contrasts). A general
469 tendency to horizontally elongated grains was observed throughout the core, with
470 grain aspect ratios lying in the range 1.1–1.4.

471 The similarity of GRIP's grain size profile with previous deep ice cores was
472 interpreted as a corroboration of the tripartite paradigm of polar ice microstructure
473 (“three-stage model”; see Sect. 3.3), even though the *c*-axis preferred orientations
474 found in the deepest 250 m of the GRIP core did not correspond to the expected
475 LPO in the recrystallization regime.

476 4.2. *GISP2*

477 After several years of planing, the U.S. Greenland Ice Sheet Project II (GISP2)
478 was officially initiated in late 1988 by the Division of Polar Programs (DPP, now

479 Office of Polar Programs) of NSF. It was developed as the first project of the new
480 Arctic System Science Program (ARCSS), a DPP initiative focusing on environ-
481 mental change in the Arctic. The scientific activities of GISP2 were coordinated
482 by the GISP2 Science Management Office at the Climate Change Research Center
483 of the University of New Hampshire, while logistics and drilling were organized
484 by the Polar Ice Coring Office (PICO) at the University of Nebraska (1987–1989)
485 and the University of Alaska Fairbanks (1989–1993).

486 The objectives of GISP2 were essentially similar to those of its companion
487 European project GRIP (see Sect. 4.1): drilling down to bedrock a deep ice core
488 at Summit, the location originally selected for GISP (cf. Sect. 3.4), in order to
489 investigate climatic and environmental changes back to the Eemian interglacial.
490 The fact that the GRIP and GISP2 drilling sites were so near (just 28 km apart)
491 implied a great advantage not only for logistics, but also for the ice core analy-
492 sis, since the records of the two cores could be used to validate each other. The
493 harmony and partnership between European GRIP and U.S. GISP2 scientists was
494 not only paramount for facilitating the logistics and validation procedures, but it
495 became also a paragon for future international drilling projects.

496 Drilling started in summer 1989 and terminated in July 1993, after drilling
497 3053.4 m of ice and almost 1.6 m of bedrock material (Gow et al., 1997). As in
498 the case of the GRIP core, severe stratigraphic disturbances caused by the ice flow
499 in the lowest 10% of the core limited reliable dating to depths ca. 300 m above
500 bedrock (\approx 110 kaBP; Peel, 1995), although tentative chronological reconstruc-
501 tions of the disturbed bottom ice do exist (Suwa et al., 2006).

502 More than 500 vertical and horizontal thin sections were sampled at 20 m in-
503 tervals from 94 to 1501 m depth, and thereafter at 10 m intervals down to 3053 m,

504 together with some additional sections for particular studies (Gow et al., 1997).
505 The samples were prepared for crystallographic analysis following standard tech-
506 niques applied in previous ice core studies. Crystalline *c*-axis orientations were
507 determined with a usual Rigsby universal stage , and presented as point scatter
508 pole figures. Average grain sizes were measured from photographs of the sec-
509 tions between crossed polarizers using two distinct methods: linear intercepts for
510 vertical sections, and measurements of the 50 largest grains in horizontal sections.

511 The GISP2 grain size analysis presented by Gow et al. (1997) is very inter-
512 esting, in the sense that its comparison of different methods reveals the degree of
513 subjectivity which ice core microstructure studies are often exposed to (Fig. A.2).
514 The linear intercept method led Woods (1994), Alley and Woods (1996), and Gow
515 et al. (1997) to identify four regimes of grain size development with depth, which
516 are to some extent similar to those reported for Camp Century, Dye 3, Byrd, and
517 GRIP. In Regime 1 the average grain cross sectional area undergoes a tenfold in-
518 crease within 600 m (which corresponds to a roughly linear growth with age),
519 reaching ca. 9 mm² at 700 m below surface (\approx 3.2 kaBP, according to Meese
520 et al., 1997). In the subsequent Regime 2, the mean grain size remains somewhat
521 stable, with a very slight decreasing trend. This stability is abruptly terminated
522 in Regime 3, which starts at the interglacial–glacial transition (at around 1680 m
523 depth) with a more than twofold grain size reduction within nearly 200 m. There-
524 after, mean grain size follows a slight increasing trend that extends over more
525 than 1000 m. Nevertheless, this impurity-rich glacial ice remains generally fine-
526 grained. At a depth of about 2750 m (close to the transition to the Eemian in-
527 terglacial), however, the first layers of clear, coarse-grained ice begin to appear,
528 betokening critical stratigraphic disturbances (Peel, 1995; cf. Sect. 4.1) and the

529 emergence of Regime 4. With thicknesses varying between tens to hundreds of
530 millimetres, such coarse-grained ice strata become very frequent around 2950 m
531 depth, making the ice close to bedrock very clear, with crystallites as large as
532 1000 mm² of cross-sectional area. The basal 13 m of the ice sheet are nonetheless
533 composed of fine-grained silty ice.

534 In contrast, the grain size dataset produced by Gow et al. (1997) via mea-
535 surements of the 50 largest grains in each sample revealed a somewhat different
536 picture. Four key regimes could still be identified, which are qualitatively simi-
537 lar to those determined with the linear intercepts method, but grain size magni-
538 tudes, variability, and rates of change, as well as the depths delimiting the key
539 regime zones, are different. In the upper zone, which corresponds to Regime 1
540 and extends from 100 to 1000 m depth, mean grain size increases steadily from
541 4.5 mm² to 22–50 mm². It remains within this wide range throughout the second
542 zone, which corresponds to Regime 2. Thus, as observed with the linear inter-
543 cept method, the stability of Regime 2 is abruptly terminated at around 1680 m
544 depth (the interglacial–glacial transition), with a more than twofold grain size
545 reduction to 11–21 mm² within nearly 200 m, which marks the beginning of
546 Regime 3. Below 2300 m the average grain size shows again a slight increase,
547 reaching ca. 25 mm² in the end of the third zone, at 2990 m depth. Below that
548 depth and down to 13 m above bedrock one finds the fourth zone, corresponding
549 to Regime 4, where grains become huge, often exceeding 1000 mm² of cross-
550 sectional area. Gow et al. (1997) remarked that, in their opinion, the 50 largest
551 grains method produced a grain size profile more similar to that observed at the
552 Byrd core (Sect. 3.3).

553 As in the case of GRIP, the similarity of GISP2's grain size profile with previ-

554 ous deep ice cores was interpreted as a corroboration of the tripartite paradigm of
555 polar ice microstructure (“three-stage model”; see Sect. 3.3).

556 Crystallographic measurements of the GISP2 core revealed a development of
557 preferred *c*-axis orientations with depth roughly similar to those already observed
558 in other deep ice cores (GRIP, Byrd, Dye 3, Camp Century), but with some impor-
559 tant differences in the details. Gow et al. (1997) report a progressive reorientation
560 of *c*-axes towards the vertical, including a strong clustering of *c*-axes beneath the
561 interglacial–glacial transition (at 1680 m depth). In the bottom 300 m of the core,
562 where stratigraphic disturbances become critical and layers of coarse-grained ma-
563 terial start to appear, the *c*-axes in the coarse-grained strata show significant de-
564 viations from the strong vertical single maximum, tending to exhibit a broad or
565 girdle-like *c*-axis distribution around the vertical. It should be remarked, however,
566 that Thorsteinsson et al. (1997) observed no sharp contrast in the *c*-axis distribu-
567 tions in the interglacial–glacial transition zone of the GRIP core, and that the zone
568 of recrystallized, coarse-grained basal ice at Byrd Station (where pressure melt-
569 ing conditions occur at the bed) is much thicker than at the GRIP and GISP2 sites,
570 where bottom ice temperatures are about -9°C .

571 An interesting feature of the crystallographic observations of the GISP2 core
572 was the discovery of *crystal striping* below ca. 2200 m depth (Alley et al., 1997),
573 identified in thin sections as stripes of crystallites with *c*-axis preferred orienta-
574 tions very distinct from the surrounding ice matrix, and believed to be formed
575 during the process of folding. In fact, visual stratigraphy analyses of the GISP2
576 core revealed that first signs of wavy strata already appear at around 2200 m,
577 centimeter-sized overturned folds are found below 2400 m, and clear evidences
578 of large-scale stratigraphic disturbances (affecting at least meters of core) occur at

579 the bottom 10% of both GRIP and GISP2 cores (Taylor et al., 1993; Gow et al.,
580 1997).

581 4.3. NGRIP

582 In spite of the of the many scientific breakthroughs and invaluable climatic in-
583 formation provided by the two Greenlandic deep ice cores from the Summit area
584 (GRIP and GISP2), the severe disturbances in the Eemian climate records of these
585 two cores posed an unwelcome setback for polar paleoclimatology. This disap-
586 pointing situation prompted the search for a new drilling site, which should con-
587 tain undisturbed ice from the Eemian interglacial period. Based on radio-echo
588 sounding profiles and geophysical models (Dahl-Jensen et al., 1997), a site on an
589 ice ridge 325 km north-northwest of the Summit was eventually selected for what
590 would be known as the North Greenland Ice Core Project (NGRIP, or NorthGRIP).

591 Support for NGRIP came from diverse funding agencies in Denmark (SNF),
592 Belgium (FNRS-CFB), France (IPEV and INSU/CNRS), Germany (AWI), Ice-
593 land (RannIs), Japan (MEXT), Sweden (SPRS), Switzerland (SNF) and the USA
594 (NSF, Office of Polar Programs). This established NGRIP as a truly *multi-continental*
595 (America, Asia and Europe) deep ice core drilling program, which was directed
596 and organized by the Niels Bohr Institute of the University of Copenhagen (Dahl-
597 Jensen et al., 2002).

598 Drilling started in summer 1996, and bedrock was reached at 3085 m depth
599 in July 2003 (NorthGRIP members, 2004). Thanks to an unexpectedly intense
600 geothermal heat flux in North Greenland (within the range 50–200 mW/m²; Dahl-
601 Jensen et al., 2003), it turned out that the basal melting rate at NGRIP (> 7 mm/a)
602 is high enough to lubricate the bed, therefore minimizing stratigraphic distur-
603 bances caused by simple-shearing flow at the bottom of the ice sheet. Conse-

604 quently, in contrast to the serious stratigraphic disruptions observed at the bottom
605 of GRIP and GISP2 (Sects. 4.1 and 4.2), the NGRIP paleoclimate records back to
606 the transition to the Eemian interglacial are unusually thick and well preserved.
607 Unfortunately, the price paid for such nice paleoclimate records is very high: the
608 intense geothermal heat flux melted away most of the Eemian ice, limiting the
609 NGRIP age to 123 kaBP (NorthGRIP members, 2004).

610 An important feature of the NGRIP core is that it became the first deep ice
611 core to have part of its visual stratigraphy (within the depth interval 1330–3085 m)
612 recorded with a new German–Danish *automated Ice-core Line-Scanner* (ILS; Dahl-
613 Jensen et al., 2002; Svensson et al., 2005; see Fig. A.4). It was also the first
614 deep ice core to have some thick sections investigated with a prototypical version
615 of the automated optical microscopy and image analysis method later known as
616 *Microstructure Mapping* (Kipfstuhl et al., 2006; also Fig. A.4). Additionally, it
617 turned into the first *Greenlandic* deep ice core to be crystallographically investi-
618 gated by means of an *Automatic Fabric Analyzer* (AFA; the first polar ice core to
619 be investigated with this technique was Dome F, cf. 6.2; see also Fig. A.4). Ac-
620 tually, two different AFAs have been used (for a description of the main methods
621 of crystallographic analysis, from the Rigsby stage to modern AFAs, see the re-
622 view by Wilen et al. 2003): the Japanese model developed by Wang and Azuma
623 (1999) was employed for *c*-axis studies in the depth range 100–2930 m, while
624 grain sizes were investigated between 115 and 880 m depth with the Australian
625 model developed by Russell-Head and Wilson (2001).

626 Vertical thin sections for *c*-axis studies were prepared by Wang et al. (2002)
627 at 55–66 m intervals between 100 and 1370 m depth, and further 300 samples
628 were extracted from the depth range 1370–2930 m. Observed *c*-axis preferred

629 orientations were presented in a variety of ways, e.g. as point scatter pole figures,
630 eigenvalues, and statistical measures, viz. degree of orientation, spherical aper-
631 ture and the Woodcock value (after Woodcock, 1977). Based on these analyses,
632 four crystallographic zones could be identified (cf. Fig. A.2). In Zone 1, rang-
633 ing from 100 to 750 m depth, nearly random distributions of *c*-axis orientations
634 are observed. In Zone 2 a broad vertical single maximum develops between 750
635 and 1300 m depth. This turns into a vertical girdle distribution in Zone 3, which
636 ranges from 1300 to 2500 m. Finally, a strong vertical single maximum prevails
637 over the girdle below 2500 m. The formation of a vertical girdle distribution of
638 *c*-axes in Zone 3 has been interpreted by Wang et al. (2002) as an evidence for
639 extension flow transverse to the NGRIP ridge, The plane of the vertical girdle ly-
640 ing in the direction of the ridge, perpendicular to the axis of horizontal extension.
641 The change from the girdle to a strong single maximum at about 2500 m depth
642 suggests the prevalence of simple shear in the lowest part of the ice sheet.

643 NGRIP Grain sizes have been studied only in the upper 900 m of the deep ice
644 core, corresponding to approximately the last 5.3 kaBP. Svensson et al. (2003b)
645 sampled 15 twin pairs of vertical thin sections evenly distributed in the depth
646 interval 115–880 m, and determined the following parameters for each grain: area,
647 width, height, flattening, roundness and *c*-axis orientation. In spite of its limited
648 depth range and number of samples, the NGRIP grain size record have become
649 one of the most studied grain size datasets from a Greenlandic deep ice core,
650 owing to its quality and level of detail.

651 In the general NGRIP grain size analysis, Svensson et al. (2003b) found that
652 the mean cross-sectional area of the grains increases with depth towards a con-
653 stant value of ca. 10 mm², and their shape becomes increasingly irregular. The

654 grain cross-sectional area distribution develops from a single log-normal to a bi-
655 modal log-normal distribution. Owing to this, a standard Normal Grain Growth
656 (NGG) model was not suitable for fitting the entire grain cross-sectional area pro-
657 file. Instead, an extended, empirical grain growth model was proposed, under
658 the assumption that below a certain depth it was the grain volume rather than the
659 cross-sectional area that grows linearly with time. In a companion paper, Svens-
660 son et al. (2003a) investigated the microstructure of a continuous 1.1 m long sec-
661 tion from around 301 m depth of the NGRIP core with the aim of relating crystal-
662 lite properties and impurity concentrations. A strong seasonal variation in grain
663 cross-sectional areas was noticed, with the smallest grains appearing in spring,
664 when the concentration of Ca^{2+} has its maximum, therefore suggesting a relation
665 between grain sizes and dust concentration (according to Whitlow et al., 1992;
666 Legrand and Mayewski, 1997; Kuramoto et al., 2011, the major source of Ca^{2+}
667 in Greenland is mineral dust, which is transported mainly from Asian sources by
668 turbulent events in early spring). In contrast to grain sizes, lattice orientations did
669 not display a detectable seasonal variation.

670 The issue of grain growth in the NGRIP core was revisited by Mathiesen et al.
671 (2004), who found that the grain size distributions of all measured depths could be
672 collapsed into a single curve by rescaling. They proposed a modified NGG equa-
673 tion with an additional “grain fragmentation” term (viz. grain splitting via rotation
674 recrystallization; RRX; cf. Appendix A of Part II) and found that the curve that
675 fitted all depths was a steady-state Bessel function, which is significantly different
676 from log-normal distribution previously proposed by Svensson et al. (2003a,b) for
677 the same dataset. Some years later, Durand et al. (2008) complemented the study
678 by Mathiesen et al. (2004) with an investigation of the relation between neighbour-

679 ing grains in the NGRIP core. They found evidences that rotation recrystallization
680 (RRX) already occurs in the upper part of the NGRIP core, seemingly at a nearly
681 constant rate, therefore contradicting deformations models (e.g. Montagnat and
682 Duval, 2000) based on the tripartite paradigm (cf. Sect.3.3).

683 Recently, Roessiger et al. (2011) compared the NGRIP grain size data with
684 the results of grain growth simulations and proved that simple NGG models with
685 an extra grain splitting term may fit well the observed data, but their physical
686 meaning is doubtful. The ice microstructure in the upper hundreds of meters of
687 polar ice sheets is usually *not in equilibrium*, and this causes noticeable effects
688 on the growth of grains that are only spuriously reproduced by such simplistic
689 models.

690 **5. News from Antarctica: Vostok, EDC**

691 Today, Vostok and Dome C are two of the three sites in Central Antarctica
692 occupied by all-year research stations (the other being the U.S. Amundsen–Scott
693 Station at the Geographic South Pole). Both lie in the Eastern Indian Sector of
694 the Antarctic Ice Sheet, circa 560 km apart. Deep ice core drilling activities have
695 been occurring in both sites since the early 1970s, in part under the auspices of the
696 International Antarctic Glaciological Project (IAGP), a large program of collabo-
697 rative glaciological studies involving Australia, France, the United Kingdom, the
698 USA, and the Soviet Union, which was carried on from the late 1960s to the mid
699 1980s and focused on an extensive part of the East Antarctica (Radok, 1977, 1985;
700 Turchetti et al., 2008). It was only by the turn of the millennium, however, that the
701 respective drilling teams reached terminal depths, with ice older than 400 kaBP.

702 *5.1. Vostok*

703 Research at Vostok has a long tradition, which dates back to the setting up of the
704 Soviet Vostok Station in December 1957, during the International Geophysical
705 Year (IGY). Since its its beginning, Vostok has existed as a year-round research
706 base of the Complex Antarctic Expedition (CAE). In 1959 CAE was renamed the
707 Soviet Antarctic Expedition (SAE) and in 1992, after the collapse of the Soviet
708 Union, the Russian Antarctic Expedition (RAE). Beneath Vostok Station and al-
709 most 4 km of ice lies one of the biggest lakes in the world, *Lake Vostok*, with a
710 surface area larger than 14,000 km² and a mean water thickness of about 125 m
711 (Kapitsa et al., 1996).

712 Deep drilling started at Vostok in April 1970, with Borehole 1, which reached
713 952.4 m in May 1972, just before a failure of the winch-brake mechanism that
714 led to the irrecoverable fall of the TELGA-14M electrothermal drill into the hole
715 (Talalay, 2012). After further four boreholes, a number of branch-holes, several
716 drills, and decades of drilling experience, the KEMS-132 electromechanical drill
717 finally reached the transition from meteoric ice into accretion ice (frozen from
718 the beneath subglacial lake) at 3538 m depth in Borehole 5G-1 (age of meteoric
719 ice estimated to be around 420 kaBP, according to Petit et al., 1999). Shortly
720 afterwards, drilling came to a halt in January 1998, at a depth of 3623 m, about
721 140 m above the ice–lake interface (Vasiliev et al., 2007).

722 This episode marked the completion of almost three decades of deep ice coring
723 at Vostok, but it did not establish the end of drilling itself. Somewhat like the
724 building of a Gothic cathedral, drilling at Vostok seemed to be a thrilling, never-
725 ending enterprise: after decades of deep ice coring, the main objective of the
726 Vostok program became getting to Lake Vostok, using the existing Borehole 5G

727 for access. To this aim, drilling was resumed in 2005 and after all sorts of technical
728 difficulties and the need to open another branch-hole (Borehole 5G-2), lake water
729 was finally hit in February 2012 at 3769.3 m depth (Jones, 2012; Talalay, 2012;
730 Vasiliev et al., 2012).

731 Originally, the plan was to replace the electromechanical ice-coring drill by
732 a coreless thermal drill at some point close to the ice–water interface (Vasiliev
733 et al., 2011). Eventually, however, no change to thermal drilling was made, and
734 ice-coring continued to the very end (Talalay, 2012; Vasiliev et al., 2012). Con-
735 tamination of the lake was most likely avoided by a vigorous surge of lake water
736 into the hole as soon as the drill broke into the lake (Vasiliev et al., 2011; Jones,
737 2012). After raising almost 600 m into the borehole (equivalent to several cubic
738 metres of subglacial water) the water must have frozen, sealing the lake beneath it
739 (Talalay, 2012). Preliminary results about microbial life in the frozen lake water
740 remain elusive (Schiermeier, 2012, 2013), mainly because of probable contam-
741 ination of the frozen water by the drilling fluid (a potentially toxic mixture of
742 kerosene and HCFC-141b; Talalay, 2012). Further exploration of the lake using a
743 variety of probes, cameras and water samplers is planned for the coming seasons.

744 Comprehensive crystallographic studies of Vostok ice have been performed
745 in the 2083 m long core (≈ 150 kaBP according to Petit et al., 1999) retrieved
746 from Borehole 3G-1 in the period 1980–1982 (Lipenkov et al., 1989; Fig. A.3).
747 Changes in grain size with depth were determined in 110 horizontal thin sections
748 by counting grains within a given area. Grain shapes were estimated by the meth-
749 ods of directed and random secants expressed in terms of the coefficients of planar
750 and linear dimensional orientation (Underwood, 1970). Crystalline *c*-axis orien-
751 tations were measured with a usual Rigsby universal stage and presented as point

752 scatter pole figures.

753 Lipenkov et al. (1989) found that the mean cross-sectional area of the ice
754 grains experiences a 30-fold increase with depth down to the interglacial–glacial
755 transition at about 1870 m, followed by a marked 60% reduction to ca. 12 mm²
756 within a depth interval of less than 150 m. Grain sizes are systematically smaller
757 in ice from colder periods (which are richer in impurities) than from warmer pe-
758 riods, indicating a *correlation between grain size and climate records/impurity*
759 *concentration*. Horizontal grain elongation is noticeable below 100 m and under-
760 goes a considerable increase between 350 and 500 m depth. Inspection of thin
761 sections between crossed polarizers suggested the near absence of interpenetrat-
762 ing crystallites and just some indications of undulose extinction below 900 m.
763 These observations led Lipenkov et al. (1989) to assume that the whole core was
764 in the first of the three stages of the tripartite paradigm of polar ice microstructure
765 (cf. Sect. 3.3), namely Normal Grain Growth (NGG) driven by reduction of the
766 grain boundary energy.

767 Crystallographic analyses revealed a quasi-uniform distribution of *c*-axis ori-
768 entations in the upper 350 m of the core, and the gradual formation of a vertical
769 girdle below 454 m depth. Lipenkov et al. (1989) could identify no significant
770 correlation between *c*-axis preferred orientations and impurity concentration or
771 climate records. On the other hand, the grain elongation along a horizontal direc-
772 tion perpendicular to the plane of the vertical girdle was interpreted as resulting
773 from basal glide induced by a tensile stress in the direction of the elongated grains,
774 so that *horizontal simple shearing is probably of little significance* along the core
775 and the general ice flow regime above the lake may be comparable to that of ice
776 shelves.

777 5.2. *EDC*

778 The EPICA Dome C (EDC) deep ice core is one of the two sister cores drilled by
779 the European Project for Ice Coring in Antarctica (EPICA), an eleven-year (1996–
780 2006) joint scientific program of the European Science Foundation (ESF) and the
781 European Commission (EC). A major part of the EPICA funding came from a
782 series of EC projects and from national contributions by ten participating countries
783 (Belgium, Denmark, France, Germany, Italy, The Netherlands, Norway, Sweden,
784 Switzerland, and the United Kingdom). The scientific activities of EPICA were
785 coordinated by a Steering Committee, which included representatives of all ten
786 participating nations (Oerter et al., 2009).

787 The main objective of EPICA was drilling down to bedrock two ice cores
788 for paleo-climate and -atmosphere records from deep ice of the Antarctic inland.
789 In contrast to its companion EDML core (see Sect. 6.1), the EDC drilling site
790 was chosen due to its remarkably low accumulation rate, which was expected to
791 provide a rather long climate record with very old ice at the bottom of the ice sheet
792 (EPICA community members, 2004). Additionally, the EDC core should yield a
793 long record of the atmospheric influences characteristic of the Indian Sector of
794 Antarctica. A decisive advantage of Dome C was that its site had already been
795 well studied and documented by numerous field surveys and ice-coring ventures
796 (Lorius et al., 1979; Young, 1979; Duval and Lorius, 1980; Jouzel et al., 1989)
797 executed within the frames of the International Antarctic Glaciological Project
798 (IAGP).

799 Italy and France provided the logistics for the EDC drilling. In early 2005
800 the new all-year facility Concordia Station became operational at Dome C, re-
801 placing an older French–Italian summer camp on the same site. Drilling started

802 in 1996 but the drill got stuck at 788 m depth and the borehole was abandoned
803 in 1999. This first core has been named EDC96 (EPICA community members,
804 2004). Drilling of the second core, EDC99 (sometimes also called EDC 2), started
805 in 1999, circa 10 m apart from the EDC96 borehole. It stopped in December 2004
806 at a depth of 3260 m, around 15 m above bedrock (Jouzel et al., 2007), after seis-
807 mic soundings suggested the presence of melt water just below. Ice at the bottom
808 of the core is estimated to be *older than 800 kaBP* (Jouzel et al., 2007; Parrenin
809 et al., 2007).

810 The microstructure of the EDC96 and EDC99 cores have been investigated
811 with several methods, including digital image analyzes of thin sections photographed
812 between crossed polarizers and, similar to NGRIP (cf. Sect. 4.3), two types of *Au-*
813 *tomatic Fabric Analyzers* (AFAs): the Japanese model (Wang and Azuma, 1999)
814 and the Australian model (Russell-Head and Wilson, 2001), cf. Fig. A.4.

815 Wang et al. (2003) studied grain sizes, shapes, and *c*-axis orientations of 33
816 vertical and horizontal thin sections from 100 to 1500 m depth of the ice cores
817 EDC96 (100–575 m) and EDC99 (575–1500 m). Grain sizes and *c*-axes were
818 analyzed with the Japanese AFA (Wang and Azuma, 1999), although further in-
819 formation about grain sizes and shapes have also been produced through digital
820 image analyzes of sections photographed between crossed polarizers. In addition,
821 fine microstructure details have been studied in some thick sections using a pre-
822 liminary version of the automated optical microscopy and image analysis method
823 that would later become known as *Microstructure Mapping* (μ SM; Kipfstuhl et al.,
824 2006, ; cf. Fig. A.4). In another study, Weiss et al. (2002) investigated grain
825 growth in EDC shallow (Holocene) ice through digital image analyzes of ca. 100
826 vertical thin sections from 100–580 m depth, photographed between crossed po-

827 larizers. Following a similar procedure, the EDC99 grain size dataset was later
828 extended by EPICA community members (2004) down to 3139 m with a peri-
829 odicity of 10 m. All these data have later been complemented by Durand et al.
830 (2009), who studied grain sizes and *c*-axis orientations of the EDC99 ice core in
831 the depth range 214–3133 m using the Australian AFA (Russell-Head and Wilson,
832 2001). Sampling was performed at 50 m intervals in the depth ranges 214–313 m
833 and 511–1500 m (which overlap with the previous study by Wang et al., 2003)
834 and every 11 m elsewhere.

835 The outcome of these studies (summarized in Fig. A.3) is that *c*-axis pre-
836 ferred orientations at EDC evolve with depth from a nearly isotropic distribution
837 close to the firn–ice transition at 100 m to a strong vertical single maximum at
838 the bottom of the core (Wang et al., 2003; Durand et al., 2009). For the upper
839 1500 m of EDC, Wang et al. (2003) could show that the gradual clustering of
840 *c*-axes towards the vertical (which is expected for an ice dome undergoing uni-
841 axial compression) agrees well with equivalent datasets from GRIP and Dome F
842 (cf. Sects. 4.1 and 6.2), when plotted together with respect to a common nor-
843 malized depth (i.e. depth/total ice thickness). Furthermore, a simple model of
844 strain-induced *c*-axis rotation based on the assumption that basal dislocation glide
845 is the dominant deformation mechanism (Azuma, 1994) satisfactorily reproduces
846 the anisotropy evolution with depth in all these cores. Below 1500 m at EDC, Du-
847 rand et al. (2009) showed that the more or less steady evolution of *c*-axis preferred
848 orientations becomes punctuated by enhanced clustering of *c*-axes around the ver-
849 tical, in fine-grained layers with increased impurity concentration. Such a sharp
850 enhancement is particularly noticeable at around 1750 m depth, which marks the
851 MIS5e–MIS6 transition from the last interglacial to the penultimate glacial period

852 (ca. 130 kaBP). Durand et al. (2009) attributed this enhancement to a combina-
853 tion of several factors: a change in ice rheology (possibly caused by small grain
854 sizes or high impurity concentration), a suitable *c*-axis distribution, and the oc-
855 currence of noticeable horizontal simple shearing already at intermediate depths.
856 This combination of factors explains why such an anisotropy enhancement has
857 only been observed at intermediate depths: in shallower EDC ice simple shearing
858 is negligible, while in deeper EDC ice the clustering of *c*-axes is already so strong
859 that further enhancement is no longer noticeable.

860 Grain size measurements in ice samples from both EDC cores revealed a gen-
861 eral grain growth behavior comparable to that observed in the Vostok and Dome F
862 cores (cf. Sects. 5.1 and 6.2): a roughly steady increase of mean grain size with
863 depth (Wang et al., 2003), punctuated by sharp size reductions at critical climatic
864 transitions (Durand et al., 2009). It is suggested that the fine-grained ice observed
865 at such climatic transitions is caused by the *pinning of grain boundaries by dust*
866 *particles*, which exist in high concentrations in glacial periods (Weiss et al., 2002;
867 Durand et al., 2009). In EDC Holocene ice the mean grain cross-sectional area
868 doubles with depth, reaching ca. 5 mm² shortly before the interglacial–glacial
869 transition at about 450 m depth. At this transition the mean grain size reduces to
870 nearly 3.5 mm² and remains approximately constant down to 600 m. Below this
871 depth, average grain size starts increasing again up to 50 mm² at about 1750 m
872 depth (MIS5e–MIS6 transition), and then drops to half. Below that depth, grains
873 resume growth, but now showing an increasing variability with depth. At the bot-
874 tom of the core, grains become rather big, reaching several hundreds of mm² in
875 cross-sectional area.

876 In the upper 580 m of the EDC core, Weiss et al. (2002) found that the *mean*

877 *grain size data could not be properly fitted* with a parabolic Normal Grain Growth
878 (NGG) law. Rather, they proposed a nearly cubic NGG model where the grain
879 volume (instead of its cross-sectional area) increases almost linearly with time.
880 Crystalline misorientation analyzes performed by Durand et al. (2009) with the
881 Australian AFA and by Wang et al. (2003) with optical microscopy revealed ev-
882 idence for rotation recrystallization (RRX) already at very shallow depths. In
883 particular, Wang et al. (2003) remarked that one out of two grains in the EDC core
884 seemed to have sub-grain boundaries, irrespective of depth.

885 **6. Recent Antarctic deep ice cores: EDML, Dome F**

886 Dronning Maud Land (DML) is a large territory in East Antarctica, between 20°W
887 and 45°E. It comprises about one sixth of the Antarctic continent, including the
888 second highest ice dome of the Antarctic ice sheet, Dome F. Two deep ice cores
889 have been retrieved from DML. The first was drilled on Dome F, at the Japanese
890 Dome Fuji Station. The second (EDML) was drilled on an ice ridge stemming
891 from Dome F, at the German Kohnen Station, which lies circa 1000 km northwest
892 from Dome Fuji.

893 *6.1. EDML*

894 As already mentioned in Sect. 5.2, the EPICA Dronning Maud Land (EDML)
895 ice core is one of the two sister cores drilled by the European Project of Ice
896 Coring in Antarctica. In contrast to its companion EDC core, the main criteria
897 for choosing the EDML drilling site were (i) a high accumulation rate, which
898 should yield a high temporal resolution of the climate records, and (ii) its loca-
899 tion in the Atlantic sector of Antarctica, in central DML, in order to allow direct

900 comparison with the climate records of Greenlandic ice cores (EPICA commu-
901 nity members, 2006). In contrast to EDC, however, the Central DML region
902 was rather unexplored prior to the EPICA investigations. Therefore, the selec-
903 tion of a precise drilling location required four years (1995–1999) of intensive
904 pre-site surveys. Eventually, the EPICA Steering Committee chose the location
905 for the EDML drilling site and the German Kohnen Station was established there
906 in January 2001. Logistics and drilling were organized by the Alfred Wegener
907 Institute (AWI), Germany. Deep drilling started in EDML in January 2002 using
908 the NGRIP drill apparatus. It finished in January 2006 at 2774 m depth, nearly
909 10 m above bedrock, after subglacial water poured into the borehole (Oerter et al.,
910 2009).

911 A distinctive feature of the EDML core is that it became the first ice core to
912 have *continuous and thorough records of visual stratigraphy and microstructure in*
913 *microscopic resolution*. These records have provided unprecedented details about
914 the mechanics and microstructure of polar ice, as well as their interactions with
915 climate proxies (Faria et al., 2010). A large amount of microstructural features, in-
916 cluding grain sizes and shapes, subgrain boundaries, air-bubble sizes, shapes, and
917 counts, slip bands, *c*-axis orientations and cloudy bands (Fig. A.4; see also Part II),
918 have been measured using the automated optical microscopy and image analysis
919 method known as Microstructure Mapping (μ SM) (Kipfstuhl et al., 2006), as well
920 as a new version of the Australian Automatic Fabric Analyzer (AFA), model G20
921 (Wilson et al., 2003). Visual stratigraphy was determined with an automated Ice-
922 core Line-Scanner (ILS) similar to the one used at NGRIP (cf. Sect 4.3; Faria
923 et al., 2010, in preparation).

924 Vertical thick sections of fresh ice and firn were cut at approximately 10 m in-

925 tervals throughout the core (10–2774 m depth) and prepared for μ SM studies us-
926 ing controlled sublimation polishing and etching, as described by Kipfstuhl et al.
927 (2006). In order to minimize relaxation effects, all samples were cut, prepared and
928 mapped in the field, shortly (0–2 days) after drilling. In addition to the standard
929 μ SM images, a modified microscopy set-up was used to produce a second series of
930 μ SM micrographs highlighting air bubbles of samples from the bubbly-ice zone,
931 viz. 90–1200 m depth (Ueltzhöffer et al., 2010; Bendel et al., 2013). Supplemen-
932 tary vertical thick sections have also been prepared for depths of special interest.
933 Additionally, about 150 vertical and horizontal thin sections from the depth inter-
934 val 54–2564 m have been prepared in at least 50 m increments for AFA analyzes
935 using standard techniques employed in previous ice core studies.

936 Similar to other Antarctic deep ice cores, the EDML mean grain size has
937 a general tendency to increase with depth (Fig. A.3). However, in the case of
938 EDML the influence of impurities seems more marked. In particular, three peri-
939 ods of pronounced Antarctic cold, known as the Marine Isotope Stages MIS2 (last
940 glacial), MIS4, and MIS6 (penultimate glacial), left their imprints on the EDML
941 microstructure in the form of exceptionally fine-grained ice. In the upper 700 m
942 of the EDML core, the mean grain cross-sectional area increases with depth from
943 1.5 mm^2 at 100 m to 4.5 mm^2 at about 700 m. Below that depth, which coincides
944 with the interglacial–glacial transition (i.e. MIS1–MIS2 transition, according to
945 the EDML1 chronology; Ruth et al., 2007), mean grain size reduces markedly,
946 reaching ca. 0.8 mm^2 at 900 m depth and remaining small for further 150 m.
947 Grains become bigger again during the warm period of MIS3 and grow in aver-
948 age to more than 6 mm^2 at about 1700 m depth. During the colder period MIS4
949 (approx. 1700–1850 m depth), grains get as small as 0.5 mm^2 in average, and

950 then resume growth reaching an average size around 20 mm^2 at 2370 m depth.
951 Below that point, the most extreme grain-size reduction in EDML takes place,
952 with grains becoming smaller than 0.3 mm^2 in average within just some tens of
953 meters. This change coincides with the most striking change in impurity con-
954 tent, caused by the transition from the last interglacial to the penultimate glacial
955 (MIS5e–MIS6 transition, ca. 130 kaBP). Within the depth range 2385–2405 m
956 the grain boundaries display a characteristic ordered pattern, resembling a “brick
957 wall” (Faria et al., 2006, 2009, in preparation), which offers a patent evidence of
958 strain accommodation by *microscopic grain-boundary sliding via microshear* (cf.
959 Drury and Humphreys, 1988; Bons and Jessell, 1999). The resulting change in the
960 ice rheology is corroborated by a corresponding change in the visual stratigraphy,
961 characterized by remarkably strong, flat and horizontal cloudy bands, as well as
962 by an abrupt reduction in the borehole diameter by ca. 5% within a period of less
963 than two years, caused by an accidental lack of drilling-fluid pressure (Faria et al.,
964 2006). Between 2400 and 2500 m depth, ice remains generally fine-grained, but
965 the grain size variability increases and the visual stratigraphy becomes severely
966 disrupted. Below 2500 m the ice temperature exceeds -10°C and grain sizes in-
967 crease dramatically, reaching thousands of mm^2 below 2600 m (Weikusat et al.,
968 2009b).

969 EDML *c*-axis preferred orientations show the depth evolution typical for an
970 ice ridge (Fig. A.3, cf. Sect. 4.3): an almost uniform distribution in the upper
971 450 m, followed by the continual development of a great circle girdle distribution
972 down to 1700 m depth, characteristic of horizontal extension flow transverse to the
973 ridge. Below that depth, a changeover region is formed towards an elongated ver-
974 tical single maximum, which ends with a sudden collapse of *c*-axes into a strong

975 vertical single maximum at 2050 m depth, where horizontal simple shearing sup-
976 posedly becomes dominant. Below 2564 m depth grains become too large for
977 meaningful determination of *c*-axis distributions (Eisen et al., 2007; Faria et al.,
978 2010).

979 In contrast to the tripartite paradigm invoked to explain the microstructure
980 evolution of certain polar ice cores (e.g. Byrd, GRIP, GISP2; cf. Sects. 3.3, 4.1
981 and 4.2), *dynamic recrystallization is active at all depths* in EDML, as confirmed
982 by detailed analyzes of grain shapes, subgrain boundary densities, and neighbor-
983 ing grain misorientations, as well as comparison with microstructures produced in
984 ice creep tests (Hamann et al., 2007; Faria et al., 2009; Weikusat et al., 2009a,b).
985 In fact, dynamic recrystallization markedly affects the ice microstructure already
986 in the firn zone (Kipfstuhl et al., 2009), as it is triggered by the highly *heteroge-*
987 *neous deformation* of polar ice on the polycrystalline and intracrystalline scales
988 (Faria et al., 2009; cf. Sect. 2.2 of Part II). The complexity of subgrain formation
989 and rotation recrystallization (RRX) has also been exposed by high-resolution lat-
990 tice orientation analyses via X-ray Laue diffraction (Miyamoto et al., 2011) and
991 Electron Backscatter Diffraction (EBSD; Weikusat et al., 2010): diverse types of
992 subgrain boundaries could be identified, many of them formed by non-basal dis-
993 locations. These results show that, while basal dislocations are the main agents
994 of intracrystalline deformation in polar ice, *non-basal dislocations* play a decisive
995 role in heterogeneous strain accommodation through the formation of subgrain
996 boundaries (Weikusat et al., 2011; cf. Sects. 2.2, 3.3 and 4.1 of Part II).

997 6.2. *Dome F*

998 Japanese research in Antarctica has a long tradition that goes back to Nobu Shi-
999 rase's 1910–1912 expeditions (Shirase, 2011). Modern Japanese Antarctic re-

1000 search started in conjunction with the International Geophysical Year (IGY, 1957–
1001 1958), through the first Japanese Antarctic Research Expedition (JARE-1) of 1956
1002 (Geographical Survey Institute of Japan, 2007). In 1968, JARE-9 scientists started
1003 collecting glaciological, climatological and geochemical data on the ice sheet in
1004 East Dronning Maud Land (East DML). These studies were carried on in sub-
1005 sequent JARE expeditions, culminating decades later with the development of
1006 the Dome Fuji Ice Coring Project, aiming at a comprehensive study of past and
1007 present glaciological/climatological features of the Antarctic ice sheet in the East
1008 DML (Dome-F Deep Coring Group, 1998). The Project was planned and executed
1009 by JARE, as part of the International Geosphere–Biosphere Program (IGBP) of
1010 the International Council for Science (ICSU).

1011 Dome Fuji Station was constructed in 1994 on the summit of East DML
1012 (Dome F), the second highest ice dome in Antarctica, 3810 m above sea level,
1013 on a relatively flat bedrock with an elevation of about 800 m. Deep drilling
1014 started in August 1995 and reached a depth of 2503 m in December 1996 (Dome-F
1015 Deep Coring Group, 1998). Climate records down to this depth seemed intact and
1016 the age of the ice was estimated to be around 340 kaBP (Watanabe et al., 1999b;
1017 Kawamura et al., 2007).

1018 As reported by Motoyama (2007), drilling stopped temporarily at 2503 m
1019 depth due to a shortage of antifreeze supply, and efforts were made to maintain the
1020 borehole open by reaming. During this process, the drill got stuck and the bore-
1021 hole had to be abandoned. Persisting in the aim of full penetration to bedrock,
1022 a new deep ice core drilling project commenced at Dome Fuji in 2001. A com-
1023 pletely new drill system was developed and drilling started in the austral summer
1024 2003 at the Dome Fuji 2 site, circa 43 m north of the abandoned borehole. In

1025 January 2007 the JARE team reached the final depth of 3035.2 m, after finding
1026 small rocks and signs of frozen subglacial water, both indicating close proximity
1027 to the bedrock. A first, preliminary dating suggests that the age of the ice at the
1028 bottom of the Dome Fuji 2 core may be around 720 kaBP.

1029 Vertical thin sections of the Dome Fuji 1 core were sampled by Azuma et al.
1030 (1999) at 20 m intervals from 100 to 2250 m depth, and thereafter at 10 m intervals
1031 down to 2503 m. They were prepared for crystallographic analyses following
1032 standard techniques employed in previous ice core studies. A major feature of
1033 the Dome Fuji 1 core is that it became the first deep ice core to have its *c*-axis
1034 orientations, as well as grain sizes and shapes, investigated with an *Automatic*
1035 *Fabric Analyzer* (AFA), which was developed by Wang and Azuma (1999). The
1036 results of these analyses were presented in a variety of ways, including mean grain
1037 size, aspect ratio and elongation direction, as well as *c*-axis point scatter pole
1038 figures, median inclinations, eigenvalues, mean orientations and misorientation
1039 angles.

1040 Azuma et al. (1999, 2000) observed (Fig. A.3) that the mean grain size of
1041 Dome Fuji 1 deep ice core is ca. 3 mm² at 112 m depth and remains nearly con-
1042 stant down to 420 m (interglacial–glacial MIS1–MIS2 transition, according to
1043 Watanabe et al., 1999a,b). A slight decrease is observed in the depth range 420–
1044 700 m followed by a roughly steady increase in deeper ice, with larger variations,
1045 reaching a maximum value of about 83 mm² at 2490 m. It was found that grain
1046 size variations correlate well with the $\delta^{18}\text{O}$ profile (Remark 5), including two con-
1047 spicuous decreases in mean grain size at about 1830 and 2300 m depth, which cor-
1048 respond to two interglacial–glacial transitions (MIS5e–MIS6 and MIS7e–MIS8,
1049 dated 130 and 245 kaBP, respectively, cf. Watanabe et al., 1999a,b). Grain elon-

1050 gation is nearly constant with depth down to ca. 800 m, and experiences a slight
1051 increase in the mean aspect ratio from 1.7 to 1.9 within the depth range 800–
1052 1500 m. Below that depth range, the aspect ratio fluctuates markedly ($\pm 10\%$)
1053 about 1.9. No signs of nucleation recrystallization (SIBM-N; cf. Appendix A of
1054 Part II) could be identified down to 2500 m.

1055 **Remark 5.** The oxygen isotope ratio $\delta^{18}\text{O}$ is commonly used as a proxy for paleo-
1056 temperature. Experience shows that *in Antarctic ice cores* the inverted $\delta^{18}\text{O}$ depth
1057 profile correlates with the concentrations of most impurities, in such a way that
1058 impurity concentration is generally higher (and $\delta^{18}\text{O}$ values lower) in colder peri-
1059 ods (EPICA community members, 2006; Faria et al., 2010).

1060 As expected for an ice dome, crystallographic *c*-axis orientations gradually
1061 change with depth from a random orientation distribution pattern near the surface
1062 to a strong vertical single maximum at 2500 m depth. Curiously, Azuma et al.
1063 (1999, 2000) found that the clustering of *c*-axes tends to be *weaker at depths with*
1064 *high impurity concentration and small grain sizes*, a result that is not incompat-
1065 ible with the observations from Dye 3 and GRIP cores (Sects. 3.4 and 4.1), but
1066 stays in direct contrast to the results from Camp Century, Byrd and GISP2 cores
1067 (cf. Sects. 3.2, 3.3 and 4.2). A possible explanation of this phenomenon has been
1068 put forward by Azuma et al. (1999, 2000): they propose that *diffusion creep* could
1069 sometimes become significant in polar ice under conditions of low temperature
1070 and low deviatoric stress, provided that the impurity concentration is high enough
1071 and the mean grain size sufficiently small, as it happens in the high-impurity layers
1072 of the Dome Fuji core.

1073 **7. Conclusion and afterword**

1074 Compared to glaciers and other natural ice bodies, polar ice sheets offer many
1075 advantages for the study of natural ice microstructure evolution. In particular, the
1076 history of stress and temperature conditions experienced by a piece of polar ice is
1077 generally much longer, simpler and more steady than it would be in a glacier. This
1078 facilitates considerably the interpretation of deformation and recrystallization mi-
1079 crostructures. Therefore, polar ice cores have become invaluable for investigations
1080 of the microstructure evolution of natural ice.

1081 In spite of all these advantages, it becomes evident from this review that un-
1082 derstanding the microstructural evolution of polar ice has been a challenging task
1083 for many decades. Even today, our knowledge about this subject is still imper-
1084 fect and incomplete, as discussed in detail in the companion Part II of this work
1085 (Faria et al., this issue). The conclusions drawn from the analyses of different
1086 ice cores have not always been consonant, as summarized in Table B.2. Such a
1087 difficulty can be attributed to several causes, ranging from the high variability of
1088 natural phenomena and the occasional subjectivity of certain methods (as revealed
1089 by the grain size studies by Gow et al., 1997 and Kipfstuhl et al., 2009) to the fact
1090 that most deep ice cores retrieved in the last decades are *climate-motivated cores*,
1091 meaning that they are generally extracted from rather singular sites (e.g. domes or
1092 ridges) that provide best-quality paleoclimate records, but rather unrepresentative
1093 (and sometimes even pathological) physical data.

1094 These facts give support to the thesis, which is being endorsed by an increasing
1095 number of glaciologists, that further progress in ice-core physics demands the pro-
1096 duction of *physically motivated deep ice cores* (Faria, 2009), viz. cores extracted
1097 from sites that are representative of the most common physical processes taking

1098 place in polar ice sheets (e.g. flow instabilities, changes in ice rheology, subglacial
1099 processes, etc.). To achieve this aim, *multidisciplinary collaborations* are essen-
1100 tial (like those promoted by the ESF Research Networking Programme Micro-
1101 Dynamics of Ice, Micro-DICE). By joining forces with geologists, geophysicists
1102 and other Earth and engineering scientists, glaciologists may have much stronger
1103 arguments to convince funding agencies and policy makers of the necessity of
1104 an international and multidisciplinary drilling program for physically motivated
1105 deep ice cores. Such a program would not only generate invaluable results for
1106 ice physics, geology and materials science: it would also provide the basis for
1107 prodigious advances in the study of palaeoclimate records of climate-motivated
1108 ice cores.

1109 *Afterword.* During the preparation of this manuscript we received with great sad-
1110 ness the news on the passing of our colleague and friend Sigfús Jóhann Johnsen
1111 (1940-2013), to whom we dedicate this work. Sigfús played a fundamental role
1112 in, and made great contributions to, all Greenland deep ice core drillings (since
1113 Dye 3) and many ice-core studies (since Camp Century). He was a leading name
1114 in the development of deep ice core drilling in Greenland and Antarctica, partici-
1115 pating in 36 ice coring expeditions onto the Greenland ice sheet as “drill-master”
1116 and scientific expert (IGS, 2013). We have learned the art of deep ice coring from
1117 him and have applied this knowledge in uncountable ice-core projects. Sigfús
1118 never retired. He tried to go onto the Greenland ice sheet to perform drilling and
1119 research even after having been advised not to do so, because of his debilitated
1120 health. He was a dedicated ice core scientist and a truly hero of glaciology. We
1121 all miss his charismatic personality, which has been cherished in all corners of the
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1134 **References**

- 1135 Achermann, D., 2009. Die Schnee- und Lawinenforschung in der Schweiz.
1136 Ph.D. thesis, University of Zurich, Zurich.
- 1137 Adams, F. D., 1990. The Birth and Development of the Geological Sciences.
1138 Dover, New York, originally published by Williams & Wilkins, Baltimore,
1139 1938.
- 1140 Ahlmann, H. W., Droessler, E. G., 1949. Glacier ice crystal measurements at Keb-
1141 nekajse, Sweden. *J. Glaciol.* 1 (5), 269–274.
- 1142 Alley, R. B., Gow, A. J., Meese, D. A., 1995. Mapping c-axis fabrics to study
1143 physical processes in ice. *J. Glaciol.* 41 (137), 197–203.

- 1144 Alley, R. B., Gow, A. J., Meese, D. A., Fitzpatrick, J. J., Waddington, E. D.,
1145 Bolzan, J. F., 1997. Grain-scale processes, folding, and stratigraphic distur-
1146 bance in the Greenland Ice Sheet Project 2 ice core. *J. Geophys. Res.* 102,
1147 26819–26830.
- 1148 Alley, R. B., Woods, G. A., 1996. Impurity influence on normal grain growth in
1149 the GISP2 ice core, greenland. *J. Glaciol.* 42 (141), 255–260.
- 1150 Augustin, L., Panichi, S., Frascati, F., 2007. EPICA Dome C 2 drilling operations:
1151 performances, difficulties, results. *Ann. Glaciol.* 47, 68–72.
- 1152 Azuma, N., 1994. A flow law for anisotropic ice and its application to ice sheets.
1153 *Earth Planet. Sci. Lett.* 128, 601–614.
- 1154 Azuma, N., Wang, Y., Mori, K., Narita, H., Hondoh, T., Shoji, H., Watanabe, O.,
1155 1999. Textures and fabrics in the Dome F (Antarctica) ice core. *Ann. Glaciol.*
1156 29, 163–168.
- 1157 Azuma, N., Wang, Y., Yoshida, Y., Narita, H., Hondoh, T., Shoji, H., Watanabe,
1158 O., 2000. Crystallographic analysis of the Dome Fuji ice core. In: Hondoh, T.
1159 (Ed.), *Physics of Ice Core Records*. Hokkaido University Press, Sapporo, pp.
1160 45–61.
- 1161 Bader, H., 1951. Introduction to ice petrofabrics. *J. Geol.* 59 (6), 519–536.
- 1162 Bader, H., 1962. Scope, problems, and potential value of deep core drilling in ice
1163 sheets. Special Report 58, U.S. Army Cold Regions Research and Engineering
1164 Laboratory, Hanover, NH.

- 1165 Bader, H., Haefeli, R., Bucher, E., Neher, J., Eckel, O., Thams, C., 1939. Der
1166 Schnee und seine Metamorphose. Vol. 3 of Beiträge zur Geologie der Schweiz
1167 - Geotechnische Serie - Hydrologie. Kümmerly and Frey, Bern.
- 1168 Bell, R. E., Studinger, M., Tikku, A. A., Clarke, G. K. C., Gutner, M. M.,
1169 Meertens, C., 2002. Origin and fate of Lake Vostok water frozen to the base
1170 of the East Antarctic ice sheet. *Nature* 416, 307–310.
- 1171 Bendel, V., Ueltzhöffer, K. J., Freitag, J., Kipfstuhl, S., Kuhs, W. F., Garbe, C. S.,
1172 Faria, S. H., 2013. High-resolution variations in size, number, and arrangement
1173 of air bubbles in the EPICA DML ice core. *J. Glaciol.* 59 (217), 972–980.
- 1174 Bentley, C. R., Koci, B. R., 2007. Drilling to the beds of the Greenland and Antarc-
1175 tic ice sheets: a review. *Ann. Glaciol.* 47, 1–9.
- 1176 Bernal, J. D., Fowler, R. H., 1933. A theory of water and ionic solution, with
1177 particular reference to hydrogen and hydroxyl ions. *J. Chem. Phys.*, 515–548.
- 1178 Bons, P. D., Jessell, M. W., 1999. Micro-shear zones in experimentally deformed
1179 octachloropropane. *J. Struct. Geol.* 21, 323–334.
- 1180 Bowden, F. P., 1953. Friction on snow and ice. *Proc. Roy. Soc. London A* 217,
1181 462–478.
- 1182 Bowden, F. P., Hughes, T. P., 1939. The mechanism of sliding on ice and snow.
1183 *Proc. Roy. Soc. London A* 172, 280–298.
- 1184 Clarke, G. K. C., 1987. A short history of scientific investigations on glaciers. *J.*
1185 *Glaciol. Special Issue*, 4–24.

- 1186 Clarke, G. K. C., 2005. Subglacial processes. *Annu. Rev. Earth Planet Sci.* 33,
1187 247–276.
- 1188 Dahl-Jensen, D., Gundestrup, N., Gogineni, S. P., Miller, H., 2003. Basal melt at
1189 NorthGRIP modeled from borehole, ice-core and radio-echo sounder observa-
1190 tions. *Ann. Glaciol.* 37, 207–212.
- 1191 Dahl-Jensen, D., Gundestrup, N. S., August 1987. Constitutive properties of ice
1192 at Dye3, Greenland. In: Waddington, E. D., Walder, J. S. (Eds.), *The physical*
1193 *basis of ice sheet modelling*. No. 170 in IAHS Publication. pp. 31–43.
- 1194 Dahl-Jensen, D., Gundestrup, N. S., Keller, K., Johnsen, S. J., Gogineni, S. P.,
1195 Allen, G. T., Chuah, T. S., Miller, H., Kipfstuhl, S., Waddington, E. D., 1997.
1196 A search in north Greenland for a new ice-core drill site. *J. Glaciol.* 43 (144),
1197 300–306.
- 1198 Dahl-Jensen, D., Gundestrup, N. S., Miller, H., Watanabe, O., Johnsen, S. J., Stef-
1199 fensen, J. P., Clausen, H. B., Svensson, A., Larsen, L. B., 2002. The NorthGRIP
1200 deep drilling programme. *Ann. Glaciol.* 35, 1–4.
- 1201 Dansgaard, W., 2004. *Frozen Annals: Greenland Ice Cap Research*. Narayana
1202 Press, Odder, Denmark.
- 1203 Dansgaard, W., Clausen, H. B., Gundestrup, N. S., Hammer, C. U., Johnsen, S. J.,
1204 Hvidberg, C. S., Steffensen, J. P., Kristinsdottir, P. M., Reeh, N., 1982. A new
1205 Greenland deep ice core. *Science* 218 (4579), 1273–1277.
- 1206 Dansgaard, W., Johnsen, S. J., 1969. A flow model and time scale for the ice core
1207 from Camp Century, Greenland. *J. Glaciol.* 8 (53), 215–223.

- 1208 Dansgaard, W., Johnsen, S. J., Clausen, H. B., Dahl-Jensen, D., Gundestrup, N. S.,
1209 Hammer, C. U., Hvidberg, C. S., Steffensen, J. P., Sveinbjornsdottir, A. E.,
1210 Jouzel, J., Bond, G., 1993. Evidence for general instability of past climate from
1211 a 250-kyr ice-core record. *Nature* 364, 218–220.
- 1212 Dansgaard, W., Johnsen, S. J., Møller, J., Langway, Jr., C. C., 1969. One thousand
1213 centuries of climatic record from Camp Century on the Greenland ice sheet.
1214 *Science* 166 (3903), 377–381.
- 1215 De la Chapelle, S., Castelnau, O., Lipenkov, V., Duval, P., 1998. Dynamic recrystallization and texture development in ice as revealed by the study of deep ice
1216 cores in antarctica and greenland. *J. Geophys. Res.* 103, 5091–5105.
- 1217
1218 de Quervain, M., Röthlisberger, H., 1999. Henri bader (1907–1998). *ICE News*
1219 *Bull. Int. Glaciol. Soc.* 120 (2), 20–22.
- 1220 Dome-F Deep Coring Group, 1998. Deep ice core drilling at Dome Fuji and
1221 glaciological studies in East Dronning Maud Land, Antarctica. *Ann. Glaciol.*
1222 27, 333–337.
- 1223 Drury, M. R., Humphreys, F. J., 1988. Microstructural shear criteria associated
1224 with grain-boundary sliding during ductile deformation. *J. Struct. Geol.* 10, 83–
1225 89.
- 1226 Durand, G., Persson, A., Samyn, D., Svensson, A., 2008. Relation between neigh-
1227 bouring grains in the upper part of the NorthGRIP ice core: implications for
1228 rotation recrystallization. *Earth Planet. Sci. Lett.* 265 (3), 666–671.
- 1229 Durand, G., Svensson, A., Persson, A., Gagliardini, O., Gillet-Chaulet, F., Sjolte,

- 1230 J., Montagnat, M., Dahl-Jensen, D., 2009. Evolution of the texture along the
1231 EPICA Dome C Ice Core. *Low Temp. Sci.* 68, 91–105.
- 1232 Durham, W. B., Stern, L. A., Kirby, S. H., 2001. Rheology of ice I at low stress
1233 and elevated confining pressure. *J. Geophys. Res.* 106 (6), 11031–11042.
- 1234 Duval, P., Ashby, M. F., Anderman, I., 1983. Rate-controlling processes in the
1235 creep of polycrystalline ice. *J. Phys. Chem.* 87, 4066–4074.
- 1236 Duval, P., Lorius, C., 1980. Crystal size and climatic record down to the last ice
1237 age from Antarctic ice. *Earth Planet. Sci. Lett.* 48 (1), 59–64.
- 1238 Eisen, O., Hamann, I., Kipfstuhl, S., Steinhage, D., Wilhelms, F., 2007. Direct ev-
1239 idence for radar reflector originating from changes in crystal-orientation fabric.
1240 *The Cryosphere* 1, 1–10.
- 1241 EPICA community members, 2004. Eight glacial cycles from an Antarctic ice
1242 core. *Nature* 429 (6992), 623–628.
- 1243 EPICA community members, 2006. One-to-one coupling of glacial climate vari-
1244 ability in Greenland and Antarctica. *Nature* 444 (7116), 195–197.
- 1245 Evatt, G. W., Fowler, A. C., Clark, C. D., Hulton, N. R. J., 2006. Subglacial floods
1246 beneath ice sheets. *Phil. Trans. R. Soc. London A* 364, 1769–1794.
- 1247 Faria, S. H., 2009. The multidisciplinary ice core. *Low Temp. Sci.* 68, 35–37.
- 1248 Faria, S. H., Freitag, J., Kipfstuhl, S., 2010. Polar ice structure and the integrity of
1249 ice-core paleoclimate records. *Quat. Sci. Rev.* 29 (1), 338–351.

- 1250 Faria, S. H., Hamann, I., Kipfstuhl, S., Miller, H., 2006. Is Antarctica like a birth-
1251 day cake? Preprint 33/2006, Max Planck Institute for Mathematics in the Sci-
1252 ences, Leipzig.
- 1253 Faria, S. H., Hutter, K., 2001. The challenge of polycrystalline ice dynamics.
1254 In: Kim, S., Jung, D. (Eds.), *Advances in Thermal Engineering and Sciences*
1255 *for Cold Regions*. Society of Air-Conditioning and Refrigerating Engineers of
1256 Korea (SAREK), Seoul, pp. 3–31.
- 1257 Faria, S. H., Kipfstuhl, S., Azuma, N., Freitag, J., Hamann, I., Murshed, M. M.,
1258 Kuhs, W. F., 2009. The multiscale structure of Antarctica. Part I: inland ice.
1259 *Low Temp. Sci.* 68, 39–59.
- 1260 Faria, S. H., Kipfstuhl, S., Lambrecht, A., in preparation. The EPICA-DML deep
1261 ice core. Springer, Heidelberg.
- 1262 Faria, S. H., Weikusat, I., Azuma, N., this issue. The microstructure of polar ice.
1263 Part II: state of the art. *J. Struct. Geol.*
- 1264 Fristrup, B., 1959. Recent investigations of the Greenland Ice Cap. *Geografisk*
1265 *Tidsskrift* 58, 1–29.
- 1266 Geographical Survey Institute of Japan, 2007. 50 years of Antarctic research ex-
1267 peditions by the Geographical Survey Institute. *Bull. Geogr. Survey Inst.* 54,
1268 1–24.
- 1269 Glen, J. W., 1952. Experiments on the deformation of ice. *J. Glaciol.* 2 (12), 111–
1270 114.

- 1271 Glen, J. W., 1955. The creep of polycrystalline ice. *Proc. Roy. Soc. London A*
1272 228, 519–538.
- 1273 Gow, A. J., 1963a. The inner structure of the Ross Ice Shelf at Little America V,
1274 Antarctica, as revealed by deep core drilling. In: IAHS Red Book 61. Interna-
1275 tional Association of Hydrological Sciences, pp. 272–274.
- 1276 Gow, A. J., 1963b. Results of measurements in the 309 meter bore hole at Byrd
1277 Station, Antarctica. *J. Glaciol.* 4 (36), 771–784.
- 1278 Gow, A. J., 1968. Deep core studies of the accumulation and densification of snow
1279 at Byrd Station and Little America V, Antarctica. Research Report 197, U. S.
1280 Army CRREL, Hanover, NH.
- 1281 Gow, A. J., 1969. On the rates of growth of grains and crystals in south polar firn.
1282 *J. Glaciol.* 8 (53), 241–252.
- 1283 Gow, A. J., Meese, D., 2007. Physical properties, crystalline textures and c-axis
1284 fabrics of the Siple dome (Antarctica) ice core. *J. Glaciol.* 53 (183), 573–584.
- 1285 Gow, A. J., Meese, D. A., Alley, R. B., Fitzpatrick, J. J., Anandakrishnan, S.,
1286 Woods, G. A., Elder, B. C., 1997. Physical and structural properties of the
1287 Greenland Ice Sheet Project 2 ice core: a review. *J. Geophys. Res.* 102, 26559–
1288 26575.
- 1289 Gow, A. J., Williamson, T., 1976. Rheological implications of the internal struc-
1290 ture and crystal fabrics of the West Antarctic ice sheet as revealed by deep core
1291 drilling at Byrd Station. *Geol. Soc. Am. Bull.* 87, 1665–1677.

- 1292 GRIP community members, 1996. Greenland Ice Core Project: An ESF Research
1293 Programme. Final report. Tech. rep., European Science Foundation, Strasbourg.
- 1294 Gundestrup, N. S., Hansen, B. L., 1984. Bore-hole survey at Dye 3, south Green-
1295 land. *J. Glaciol.* 30, 282–288.
- 1296 Haefeli, R., von Sury, H., 1975. Strain and stress in snow, firn and ice along the
1297 EGIG profile of the Greenland ice sheet. In: IAHS Red Book 114. International
1298 Association of Hydrological Sciences, pp. 342–352.
- 1299 Hamann, I., Weikusat, C., Azuma, N., Kipfstuhl, S., May 2007. Evolution of ice
1300 crystal microstructures during creep experiments. *J. Glaciol.* 53 (182), 479–489.
- 1301 Hammer, C. U., Clausen, H. B., Langway, Jr., C. C., 1994. Electrical conductivity
1302 method (ECM) stratigraphic dating of the Byrd Station ice core, Antarctica.
1303 *Ann. Glaciol.* 20, 115–120.
- 1304 Hansen, B. L., Langway, Jr., C. C., 1966. Deep core drilling and core analysis at
1305 Camp Century, Greenland, 1961–1966. *Antarct. J. U.S.* 1 (5), 207–208.
- 1306 Herron, S. L., Langway, Jr., C. C., 1979. The debris-laden ice at the bottom of the
1307 Greenland Ice Sheet. *J. Glaciol.* 23 (89), 193–207.
- 1308 Herron, S. L., Langway, Jr., C. C., 1982. A comparison of ice fabrics and textures
1309 at Camp Century, Greenland and Byrd Station, Antarctica. *Ann. Glaciol.* 3,
1310 118–124.
- 1311 Herron, S. L., Langway, Jr., C. C., Brugger, K. A., 1985. Ultrasonic velocities and
1312 crystalline anisotropy in the ice core from Dye 3, Greenland. In: Langway, Jr.,
1313 C. C., Oeschger, H., Dansgaard, W. (Eds.), *Greenland Ice Core: Geophysics,*

- 1314 Geochemistry, and the Environment. Vol. 33 of Geophys. Monograph. Ameri-
1315 can Geophysical Union, Washington, DC, pp. 23–31.
- 1316 Hobbs, P. V., 1974. Ice Physics. Clarendon, Oxford.
- 1317 Holtzschler, J. J., de Q. Robin, G., Glen, J. W., 1954. Depth of polar ice caps.
1318 Geogr. J. 120 (2), 193–202.
- 1319 Hughes, T. P., Seligman, G., 1939a. The bearing of snow permeability and re-
1320 tentivity on the density increase of firn and ice-band formation in glaciers. In:
1321 IAHS Red Book 26, Commission of Snow and Glaciers, Q1-R3. International
1322 Association of Hydrological Sciences, pp. 1–7.
- 1323 Hughes, T. P., Seligman, G., 1939b. The temperature, melt water movement and
1324 density increase in the névé of an Alpine glacier. Mon. Not. R. Astron. Soc.,
1325 Geophys. Suppl. 4, 615–647.
- 1326 Hvidberg, C. S., Keller, K., Gundestrup, N. S., Tschering, C. C., Forsberg, R.,
1327 1997. Mass balance and surface movement of the Greenland Ice Sheet at Sum-
1328 mit, Central Greenland. Geophys. Res. Lett. 24 (18), 2307–2310.
- 1329 IGS, June 2013. Sigfús Jóhann Johnsen 1940–2013. <http://http://www.igsoc.org/news/sigfusjohnsen/>,
1330 retrieved on 01 Sep 2013.
- 1331 Johnsen, S. J., Dahl-Jensen, D., Dansgaard, W., Gundestrup, N. S., 1995. Green-
1332 land palaeotemperatures derived from GRIP bore hole temperaure and ice core
1333 isotope profiles. Tellus 47B, 624–629.
- 1334 Johnsen, S. J., Gundestrup, N. S., Hansen, S. B., Schwander, J., Ruffli, H., 1994.

- 1335 The new improved version of the ISTUK ice core drill. *Mem. Natl. Inst. Polar*
1336 *Res., Spec. Issue 49*, 9–23.
- 1337 Jones, N., 2012. Russians celebrate Vostok victory. *Nature* 482 (7385), 287.
- 1338 Jouzel, J., Masson-Delmotte, V., Cattani, O., Dreyfus, G., Falourd, S., Hoffmann,
1339 G., Minster, B., Nouet, J., Barnola, J. M., Chappellaz, J., Fischer, H., Gal-
1340 let, J. C., Johnsen, S., Leuenberger, M., Loulergue, L., Luethi, D., Oerter, H.,
1341 Parrenin, F., Raisbeck, G., Raynaud, D., Schilt, A., Schwander, J., Selmo, E.,
1342 Souchez, R., Spahni, R., Stauffer, B., Stenni, J. P. S. B., Stocker, T. F., Tison,
1343 J. L., Werner, M., Wolff, E. W., 2007. Orbital and millennial Antarctic climate
1344 variability over the past 800,000 years. *Science* 317 (5839), 793–797.
- 1345 Jouzel, J., Raisbeck, G., Benoist, J. P., Yiou, F., Lorius, C., Raynaud, D., Petit,
1346 J. R., Barkov, N. I., Korotkevitch, Y. S., Kotlyakov, V. M., 1989. A comparison
1347 of deep Antarctic ice cores and their implications for climate between 65,000
1348 and 15,000 years ago. *Quat. Res.* 31 (2), 135–150.
- 1349 Kapitsa, A. P., Ridley, J. K., Robin, G. D. Q., Siegert, M. J., Zotikov, I. A., 1996.
1350 A large deep freshwater lake beneath the ice of central East Antarctica. *Nature*
1351 381 (6584), 684–686.
- 1352 Kawamura, K., Parrenin, F., Lisiecki, L., Uemura, R., Vimeux, F., Severinghaus,
1353 J. P., Hutterli, M. A., Nakazawa, T., Aoki, S., Jouzel, J., Raymo, M. E., Mat-
1354 sumoto, K., Nakata, H., Motoyama, H., Fujita, S., Goto-Azuma, K., Fujii, Y.,
1355 Watanabe, O., 2007. Northern Hemisphere forcing of climatic cycles in Antarc-
1356 tica over the past 360000 years. *Nature* 448, 912–916.

- 1357 Kipfstuhl, S., Faria, S. H., Azuma, N., Freitag, J., Hamann, I., Kaufmann, P.,
1358 Miller, H., Weiler, K., Wilhelms, F., 2009. Evidence of dynamic recrystalliza-
1359 tion in polar firn. *J. Geophys. Res.* 114, B05204.
- 1360 Kipfstuhl, S., Hamann, I., Lambrecht, A., Freitag, J., Faria, S. H., Grigoriev,
1361 D., Azuma, N., 2006. Microstructure mapping: A new method for imag-
1362 ing deformation-induced microstructural features of ice on the grain scale. *J.*
1363 *Glaciol.* 52 (178), 398–406.
- 1364 Kuramoto, T., Goto-Azuma, K., Hirabayashi, M., Miyake, T., Motoyama, H.,
1365 Dahl-Jensen, D., Steffensen, J. P., 2011. Seasonal variations of snow chemistry
1366 at NEEM, Greenland. *Ann. Glaciol.* 52 (58), 193–200.
- 1367 Landais, A., Chappellaz, J., Delmotte, M., Jouzel, J., Blunier, T., Bourg, C., Cail-
1368 lon, N., Cherrier, S., Malaizé, B., Masson-Delmotte, V., Raynaud, D., Schwan-
1369 der, J., Steffensen, J. P., 2003. A tentative reconstruction of the last interglacial
1370 and glacial inception in Greenland based on new gas measurements in the
1371 Greenland Ice Core Project (GRIP) ice core. *J. Geophys. Res.* 108 (D18), 4563.
- 1372 Lange, M. A., 1988. A computer-controlled system for ice-fabric analysis on a
1373 Rigsby stage. *Ann. Glaciol.* 10, 92–94.
- 1374 Langway, Jr., C. C., 1970. Stratigraphic analysis of a deep ice core from Green-
1375 land. Special Paper 125, Geological Society of America, Boulder, CO.
- 1376 Langway, Jr., C. C., 2008. The history of early polar ice cores. Technical Re-
1377 port ERDC/CRREL TR-08-1, U.S. Army Engineer Research and Development
1378 Center, Cold Regions Research and Engineering Laboratory, Hanover, NH.

- 1379 Langway, Jr., C. C., Shoji, H., Azuma, N., 1988. Crystal size and orientation
1380 patterns in the Wisconsin-age ice from Dye 3, Greenland. *Ann, Glaciol.* 10,
1381 109–115.
- 1382 Legrand, M., Mayewski, P. A., 1997. Glaciochemistry of polar ice cores: a review.
1383 *Rev. Geophys.* 35, 219–143.
- 1384 Lemke, P., Ren, J., Alley, R. B., Allison, I., Carrasco, J., Flato, G., Fujii, Y.,
1385 Kaser, G., Mote, P., Thomas, R. H., Zhang, T., 2007. Observations: changes in
1386 snow, ice and frozen ground. In: Solomon, S., Qin, D., Manning, M., Chen, Z.,
1387 Marquis, M., Averyt, K. B., Tignor, M., Miller, H. L. (Eds.), *Climate Change*
1388 *2007: The Physical Science Basis. Contribution of Working Group I to the*
1389 *Fourth Assessment Report of the Intergovernmental Panel on Climate Change*
1390 (IPCC). Cambridge University Press, Cambridge.
- 1391 Lhomme, N., Clarke, G. K. C., Marshall, S. J., 2005. Tracer transport in the Green-
1392 land Ice Sheet: constraints on ice cores and glacial history. *Quat. Sci. Rev.* 24,
1393 173–194.
- 1394 Lipenkov, V. Y., Barkov, N. I., Duval, P., Pimienta, P., 1989. Crystalline texture of
1395 the 2083m ice core at vostok station, antarctica. *J. Glaciol.* 35 (121), 392–398.
- 1396 Lorius, C., Merlivat, L., Jouzel, J., Pourchet, M., 1979. A 30,000-yr isotope cli-
1397 matic record from Antarctic ice. *Nature* 280 (5724), 644–648.
- 1398 Lythe, M. B., Vaughan, D. G., 2001. BEDMAP: a new ice thickness and subglacial
1399 topographic model of Antarctica. *J. Geophys. Res.* 106 (B6), 11335–11351.

- 1400 Mathiesen, J., Ferkinghoff-Borg, J., Jensen, M. H., Levinsen, M., Olesen, P., Dahl-
1401 Jensen, D., Svensson, A., 2004. Dynamics of crystal formation in the Greenland
1402 NorthGRIPice core. *J. Glaciol.* 50 (170), 325–328.
- 1403 Meese, D. A., Gow, A. J., Alley, R. B., Zielinski, G. A., Grootes, P. M., Ram,
1404 M., Taylor, K. C., Mayewski, P. A., Bolzan, J. F., 1997. The Greenland Ice
1405 Sheet Project 2 depth–age scale: methods and results. *J. Geophys. Res.* 102,
1406 26411–26423.
- 1407 Merlivat, L., Ravoire, J., Vergnaud, J. P., Lorius, C., 1973. Tritium and deuterium
1408 content of the snow in Groenland. *Earth Planet. Sci. Lett.* 19 (2), 235–240.
- 1409 Miller, M. M., 1954. Juneau Icefield Research Program. Juneau Icefield Research
1410 Program Report 7, American Geographic Society.
- 1411 Mills, W. J., 2003. *Exploring Polar Frontiers. Vol. 1. ABC-CLIO, Santa Barbara,*
1412 *CA.*
- 1413 Miyamoto, A., Weikusat, I., Hondoh, T., 2011. Complete determination of ice
1414 crystal orientation and microstructure investigation on ice core samples en-
1415 abled by a new x-ray laue diffraction method. *J. Glaciol.* 57 (201), 67–74, awi-
1416 n18929.
- 1417 Montagnat, M., Duval, P., 2000. Rate controlling processes in the creep of po-
1418 lar ice, influence of grain boundary migration associated with recrystallization.
1419 *Earth Planet. Sci. Lett.* 183, 179–186.
- 1420 Motoyama, H., 2007. The second deep ice coring project at Dome Fuji, Antarc-
1421 tica. *Sci. Drilling* 5, 41–43.

- 1422 Motoyama, H., Furukawa, T., Nishio, F., 2008. Study of ice flow observations
1423 in Shirase drainage basin and around Dome Fuji area, East Antarctica by dif-
1424 ferential GPS method. *Nankyoku Shiryo (Antarctic Record)* 52, 216–231, (in
1425 Japanese).
- 1426 Nolan, M., Motkya, R. J., Echelmeyer, K., Trabant, D. C., 1995. Ice-thickness
1427 measurements of Taku Glacier, Alaska, U.S.A., and their relevance to its recent
1428 behavior. *J. Glaciol.* 41 (139), 541–553.
- 1429 NorthGRIP members, 2004. High-resolution record of the Northern Hemisphere
1430 climate extending into the last interglacial period. *Nature* 431, 147–151.
- 1431 Obbard, R., Baker, I., 2007. The microstructure of meteoric ice from Vostok,
1432 Antarctica. *J. Glaciol.* 53 (180), 41–62.
- 1433 Oerter, H., Drücker, C., Kipfstuhl, S., Wilhelms, F., 2009. Kohnen station, the
1434 drilling camp for the EPICA deep ice core in Dronning Maud Land. *Polar-*
1435 *forschung* 78 (1/2), 1–23.
- 1436 Parrenin, F., Barnola, J.-M., Beer, J., Blunier, T., Castellano, E., Chappellaz, J.,
1437 Dreyfus, G., Fischer, H., Fujita, S., Jouzel, J., Kawamura, K., Lemieux-Dudon,
1438 B., Loulergue, L., Masson-Delmotte, V., Narcisi, B., Petit, J.-R., Raisbeck, G.,
1439 Raynaud, D., Ruth, U., Schwander, J., Severi, M., Spahni, R., Steffensen, J. P.,
1440 Svensson, A., Udisti, R., Waelbroeck, C., Wolff, E., 2007. The EDC3 chronol-
1441 ogy for the EPICA Dome C ice core. *Clim. Past* 3, 485–497.
- 1442 Paterson, W. S. B., 1983. Deformation within polar ice sheets: an analysis of the
1443 Byrd Station and Camp Century borehole-tilting measurements. *Cold Reg. Sci.*
1444 *Technol.* 8 (2), 165–179.

- 1445 Paterson, W. S. B., 1991. Why ice-age ice is sometimes “soft”. *Cold Reg. Sci.*
1446 *Technol.* 20 (1), 75–98.
- 1447 Peel, D. A., 1995. Profiles of the past. *Nature* 378, 234–235.
- 1448 Pelto, M. S., Miller, M. M., 1990. Mass balance of the Taku Glacier, Alaska from
1449 1946–1986. *Northw. Sci.* 64, 121–130.
- 1450 Perutz, M. F., 1948. A description of the iceberg aircraft carrier and the bearing of
1451 the mechanical properties of frozen wood pulp upon some problems of glacier
1452 flow. *J. Glaciol.* 1 (3), 95–104.
- 1453 Perutz, M. F., 1949. The flow of ice and other solids. in: Joint meeting of the
1454 british glaciological society, the british rheologists’ club and the institute of
1455 metals. *J. Glaciol.* 1 (5), 231–240.
- 1456 Perutz, M. F., 1950a. Direct measurement of the velocity distribution in a vertical
1457 profile through a glacier. *J. Glaciol.* 1 (7), 382–383.
- 1458 Perutz, M. F., 1950b. In: *Glaciology—the flow of glaciers.* The Observatory
1459 70 (855), 64–65.
- 1460 Perutz, M. F., 1953. The flow of glaciers. *Nature* 172 (4386), 929–932.
- 1461 Perutz, M. F., Seligman, G., 1939. A crystallographic investigation of glacier
1462 structure and the mechanism of glacier flow. *Proc. Roy. Soc. London A* 172,
1463 335–360.
- 1464 Petit, J. R., Jouzel, J., Raynaud, D., Barkov, N. I., Barnola, J.-M., Basile, I., Ben-
1465 der, M., Chappellaz, J., Davisk, M., Delaygue, G., Delmotte, M., Kotlyakov,

- 1466 V. M., Legrand, M., Lipenkov, V. Y., Lorius, C., Pépin, L., Ritz, C., Saltzman,
1467 E., Stievenard, M., 1999. Climate and atmospheric history of the past 420,000
1468 years from the Vostok ice core, Antarctica. *Nature* 399, 429–436.
- 1469 Petrenko, V. F., Whitworth, R. W., 1999. *Physics of Ice*. Oxford University Press,
1470 Oxford.
- 1471 Radok, U., 1977. International Antarctic Glaciological Project: past and future.
1472 *Antarct. J. U.S.* 12 (1), 32–38.
- 1473 Radok, U., 1985. The Antarctic ice. *Sci. Amer.* 253 (2), 98–105.
- 1474 Reeh, N., 1989. The age-depth profile in the upper part of a steady-state ice sheet.
1475 *J. Glaciol.* 35 (121), 406–417.
- 1476 Reeh, N., Clausen, H. B., Dansgaard, W., Gundestrup, N., Hammer, C. U.,
1477 Johnsen, S. J., 1978. Secular trends of accumulation rates at three Greenland
1478 stations. *J. Glaciol.* 20 (82), 27–30.
- 1479 Rignot, E., Mouginot, J., Scheuchl, B., 2011. MEaSURES InSAR-based Antarc-
1480 tica ice velocity map. version 1.
- 1481 Rigsby, G. P., 1951. Crystal fabric studies on Emmons Glacier Mount Rainier,
1482 Washington. *J. Geol.* 59 (6), 590–598.
- 1483 Rigsby, G. P., 1958. Fabrics of glacier and laboratory deformed ice. In: IAHS
1484 Red Book 47, *Physics of the Movement of Ice*. International Association of
1485 Hydrological Sciences, pp. 351–358.
- 1486 Rigsby, G. P., 1960. Crystal orientation in glacier and in experimentally deformed
1487 ice. *J. Glaciol.* 3 (27), 589–606.

- 1488 Ritz, C., 1989. Interpretation of the temperature profile measured at Vostok, East
1489 Antarctica. *Ann. Glaciol.* 12, 138–144.
- 1490 Robin, G. D. Q., 1955. Ice movement and temperature distribution in glaciers and
1491 ice sheets. *J. Glaciol.* 18 (18), 523–532.
- 1492 Roessiger, J., Bons, P. D., Grier, A., Jessell, M. W., Evans, L., Montagnat,
1493 M., Kipfstuhl, S., Faria, S. H., Weikusat, I., 2011. Competition between grain
1494 growth and grain-size reduction in polar ice. *J. Glaciol.* 57 (205), 942–948.
- 1495 Russell-Head, D. S., Wilson, C. J. L., 2001. Automated fabric analyzer system for
1496 quartz and ice. *Geol. Soc. Aust. Abstracts* 64, 159.
- 1497 Ruth, U., Barnola, J. M., Beer, J., Bigler, M., Blunier, T., Castellano, E., Fis-
1498 cher, H., Fundel, F., Huybrechts, P., Kaufmann, P., Kipfstuhl, S., Lambrecht,
1499 A., Morganti, A., Oerter, H., Parrenin, F., Rybak, O., Severi, M., Udisti, R.,
1500 Wilhelms, F., Wolff, E., 2007. “EDML1”: a chronology for the EPICA deep ice
1501 core from Dronning Maud Land, Antarctica, over the last 150 000 years. *Clim.*
1502 *Past* 3, 475–484.
- 1503 Schiermeier, Q., 2012. Hunt for life under Antarctic ice heats up. *Nature*
1504 491 (7425), 506–507.
- 1505 Schiermeier, Q., July 2013. Claims of Lake Vostok fish get frosty response.
- 1506 Schlosser, E., Oerter, H., 2002. Shallow firn cores from Neumayer, Ekströmisen,
1507 Antarctica: a comparison of accumulation rates and stable-isotope ratios. *Ann.*
1508 *Glaciol.* 35, 91–96.

- 1509 Schulson, E. M., Duval, P., 2009. Creep and Fracture of Ice. Cambridge University
1510 Press, Cambridge.
- 1511 Schytt, V., 1958. Snow and ice studies in Antarctica. Ph.D. Thesis, University of
1512 Stockholm. In: Norwegian–British–Swedish Antarctic Expedition, 1949–1952,
1513 Scientific Results 4, Glaciology II. Norsk Polarinstitut, Oslo.
- 1514 Seddik, H., Greve, R., Placidi, L., Hamann, I., Gagliardini, O., 2008. Application
1515 of a continuum-mechanical model for the flow of anisotropic polar ice to the
1516 EDML core, Antarctica. *J. Glaciol.* 54 (187), 631–642.
- 1517 Seligman, G., 1936. Snow Structure and Ski Fields. Macmillan, London.
- 1518 Seligman, G., 1941. The structure of a temperate glacier. *Geogr. J.* 97 (5), 295–
1519 315.
- 1520 Seligman, G., 1949. The growth of the glacier crystal. *J. Glaciol.* 1 (5), 254–268.
- 1521 Shirase, N., 2011. Antarctica: The Japanese South Polar Expedition of 1910–12.
1522 Bluntisham Books and Erskine Press, Bluntisham.
- 1523 Siegert, M. J., 2005. Lakes beneath the ice sheet. *Annu. Rev. Earth Planet. Sci.*
1524 33, 247–276.
- 1525 Siegert, M. J., Kwok, R., 2000. Ice-sheet radar layering and the development of
1526 preferred crystal orientation fabrics between Lake Vostok and Ridge B, central
1527 East Antarctica. *Earth Planet. Sci. Lett.* 179, 227–235.
- 1528 Stafford, J. M., Wendler, G., Curtis, J., 2000. Temperature and precipitation of
1529 Alaska: 50 year trend analysis. *Theor. Appl. Climatol.* 67, 33–44.

- 1530 Steinemann, S., 1954. Results of preliminary experiments on the plasticity of ice
1531 crystals. *J. Glaciol.* 2, 404–412.
- 1532 Steinemann, S., 1958. Experimentelle Untersuchungen zur Plastizität von Eis.
1533 *Beitr. Geol. Schweiz, Hydrol.* 10, 1–72, also as Ph.D. Thesis, Swiss Federal
1534 Institute of Technology (ETH) Zurich.
- 1535 Stephenson, P. J., 1967. Some considerations of snow metamorphism in the
1536 antarctic ice sheet in the light of ice crystal studies. In: Oura, H. (Ed.), *Physics*
1537 *of Snow and Ice*. Vol. 1. Hokkaido University Press, Sapporo, pp. 725–740, pro-
1538 ceedings of the International Conference on Low Temperature Science, 1966,
1539 Sapporo, Japan.
- 1540 Suwa, M., von Fischer, J. C., Bender, M. L., Landais, A., Brook, E. J., 2006.
1541 Chronology reconstruction for the disturbed bottom section of the GISP2 and
1542 the GRIP ice cores: Implications for Termination II in Greenland. *J. Geophys.*
1543 *Res.* 111 (D2), D02101.
- 1544 Svensson, A., Baadsager, P., Persson, A. r., Schø tt Hvidberg, C., Siggard-
1545 Andersen, M.-L., 2003a. Seasonal variability in ice crystal properties at North-
1546 GRIP: a case study around 301 m depth. *Ann. Glaciol.* 37, 119–122.
- 1547 Svensson, A., Nielsen, S. W., Kipfstuhl, S., Johnsen, S. J., Steffensen, J. P., Bigler,
1548 M., Ruth, U., Röthlisberger, R., 2005. Visual stratigraphy of the North Green-
1549 land Ice Core Project (NorthGRIP) ice core during the last glacial period. *J.*
1550 *Geophys. Res.* 110, D02108.
- 1551 Svensson, A., Schmidt, K. G., Dahl-Jensen, D., Johnsen, S. J., Wang, Y., Kipfs-

- 1552 tuhl, J., Thorsteinsson, T., 2003b. Properties of ice crystals in NorthGRIP late-
1553 to middle-Holocene ice. *Ann. Glaciol.* 37, 113–118.
- 1554 Takata, M., Iizuka, Y., Hondoh, T., Fujita, S., Fuji, Y., Shoji, H., 2004. Strati-
1555 graphic analysis of Dome Fuji Antarctic ice core using an optical scanner. *Ann.*
1556 *Glaciol.* 39, 467–472.
- 1557 Talalay, P. G., 2012. Russian researchers reach subglacial Lake Vostok in Antarc-
1558 tica. *Adv. Polar Sci.* 23 (3), 176–180.
- 1559 Taylor, K. C., Hammer, C. U., Alley, R. B., Clausen, H. B., Dahl-Jensen, D., Gow,
1560 A. J., Gundestrup, N. S., Kipfstuhl, J., Moore, J. C., Waddington, E. D., 1993.
1561 Electrical conductivity measurements from the GISP2 and GRIP Greenland ice
1562 cores. *Nature* 366, 549–552.
- 1563 Thorsteinsson, T., Kipfstuhl, J., Eicken, H., Johnsen, S. J., Fuhrer, K., 1995. Crys-
1564 tal size variations in Eemian-age ice from the GRIP ice core, Central Greenland.
1565 *Earth Planet. Sci. Lett.* 131, 381–394.
- 1566 Thorsteinsson, T., Kipfstuhl, J., Miller, H., 1997. Textures and fabrics in the GRIP
1567 ice core. *J. Geophys. Res.* 102, 26583–26599.
- 1568 Turchetti, S., Naylor, S., Dean, K., Siegert, M., 2008. On thick ice: scientific
1569 internationalism and Antarctic affairs, 1957–1980. *Hist. Technol.* 24 (4), 351–
1570 376.
- 1571 Ueda, H. T., Garfield, D. E., 1970. Deep core drilling at Byrd Station. In: IAHS
1572 Red Book 86. International Association of Hydrological Sciences, pp. 53–62.

- 1573 Ueltzhöffer, K. J., Bendel, V., Freitag, J., Kipfstuhl, S., Wagenbach, D., Faria,
1574 S. H., Garbe, C. S., 2010. Distribution of air bubbles in the EDML and EDC
1575 ice cores from a new method of automatic image analysis. *J. Glaciol.* 56 (196),
1576 339–348.
- 1577 Underwood, E. E., 1970. *Quantitative Stereology*. Addison–Wesley, Reading.
- 1578 Van den Broeke, M. R., König-Langlo, G., Picard, G., Munneke, P. K., Lenaerts, J.
1579 T. M., 2009. Surface energy balance, melt and sublimation at Neumayer station
1580 (East Antarctica). *Antarct. Sci.* 22 (1), 87–96.
- 1581 Vasiliev, N. I., Lipenkov, V. Y., Dmitriev, A. N., Podolyak, A. V., Zubkov, V. M.,
1582 2012. Results and characteristics of 5G hole drilling and the first tapping of lake
1583 Vostok. *Ice and Snow* 4 (120), 12–20, in Russian.
- 1584 Vasiliev, N. I., Talalay, P. G., Bobin, N. E., Chistyakov, V. K., Zubkov, V. M.,
1585 Krasilev, A. V., Dmitriev, A. N., Yankilevich, S. V., Lipenkov, V. Y., 2007. Deep
1586 drilling at Vostok station, Antarctica: history and recent events. *Ann. Glaciol.*
1587 47, 10–23.
- 1588 Vasiliev, N. I., Talalay, P. G., Vostok Deep Ice Core Drilling Parties, 2011. Twenty
1589 years of drilling the deepest hole in ice. *Sci. Drilling* 11, 41–45.
- 1590 Victor, P.-E., Bauer, A., 1961. Correspondence on “Air temperature and precipi-
1591 tation on the Greenland Ice Sheet. *J. Glaciol.* 3 (29), 949.
- 1592 Vittuari, L., Vincent, C., Frezzotti, M., Mancini, F., Gandolfi, S., Bitelli, G.,
1593 Capra, A., 2004. Space geodesy as a tool for measuring ice surface velocity in
1594 the Dome C region and along the ITASE traverse. *Ann. Glaciol.* 39, 402–408.

- 1595 Walker, J. C. F., Waddington, E. D., 1988. Descent of glaciers: some early specu-
1596 lations on glacier flow and ice physics. *J. Glaciol.* 34 (118), 342–348.
- 1597 Wang, Y., Azuma, N., 1999. A new automatic ice-fabric analyzer which uses
1598 image-analysis techniques. *Ann. Glaciol.* 29, 155–162.
- 1599 Wang, Y., Kipfstuhl, S., Azuma, N., Thorsteinsson, T., Miller, H., 2003. Ice-
1600 fabrics study in the upper 1500 m of the Dome C (East Antarctica) deep ice
1601 core. *Ann. Glaciol.* 37, 97–104.
- 1602 Wang, Y., Thorsteinsson, T., Kipfstuhl, J., Miller, H., Dahl-Jensen, D., Shoji, H.,
1603 2002. A vertical girdle fabric in the NorthGRIP deep ice core, North Greenland.
1604 *Ann. Glaciol.* 35, 515–520.
- 1605 Watanabe, O., Fujii, Y., Kamiyama, K., Motoyama, H., Furukawa, T., Igarashi,
1606 M., Kohno, M., Kanamori, S., Ageta, Y., Nakawo, M., Tanaka, H., Satow, K.,
1607 Shoji, H., Kawamura, K., Matoba, S., Shimada, W., 1999a. Basic analyses of
1608 Dome Fuji Deep Ice Core Part I: Stable oxygen and hydrogen isotope ratios,
1609 major chemical compositions and dust concentration. *Polar Meteorol. Glaciol.*
1610 13, 83–89.
- 1611 Watanabe, O., Kamiyama, K., Motoyama, H., Fujii, Y., Shoji, H., Satow, K.,
1612 1999b. The paleoclimate record in the ice core at Dome Fuji station, East
1613 Antarctica. *Ann. Glaciol.* 29, 176–178.
- 1614 Wegener, K., 1955. Die Temperatur in grönlandischen Inlandeis. *Geofis. Pura*
1615 *Appl.* 32, 102–106.
- 1616 Weidick, A., Bennike, O., 2007. Quaternary glaciation history and glaciology of

- 1617 Jakobshavn Isbræ and the Disko Bugt region, West Greenland: a review. Geo-
1618 logical Survey of Denmark and Greenland Bulletin 14.
- 1619 Weikusat, I., de Winter, D. A. M., Pennock, G. M., Hayles, M., Schneijdenberg,
1620 C. T. W. M., Drury, M. R., June 2010. Cryogenic EBSD on ice: preserving a
1621 stable surface in a low pressure SEM. *J. Microsc.* 242 (3), 295–310.
- 1622 Weikusat, I., Kipfstuhl, S., Azuma, N., Faria, S. H., Miyamoto, A., 2009a. Deform-
1623 ation microstructures in an Antarctic ice core (EDML) and in experimentally
1624 deformed artificial ice. *Low Temp. Sci.* 68, 115–123.
- 1625 Weikusat, I., Kipfstuhl, S., Faria, S. H., Azuma, N., Miyamoto, A., 2009b. Sub-
1626 grain boundaries and related microstructural features in EDML(Antarctica)
1627 deep ice core. *J. Glaciol.* 55 (191), 461–472.
- 1628 Weikusat, I., Miyamoto, A., Faria, S. H., Kipfstuhl, S., Azuma, N., Hondoh, T.,
1629 2011. Subgrain boundaries in Antarctic ice quantified by X-ray Laue diffrac-
1630 tion. *J. Glaciol.* 57 (201), 85–94.
- 1631 Weiss, J., Vidot, J., Gay, M., Arnaud, L., Duval, P., Petit, J. R., 2002. Dome
1632 concordia ice microstructure: impurities effect on grain growth. *Ann. Glaciol.*
1633 35, 552–558.
- 1634 Wesche, C., Eisen, O., Oerter, H., Schulte, D., Steinhage, D., 2007. Surface topog-
1635 raphy and ice flow in the vicinity of the EDML deep-drilling site, Antarctica. *J.*
1636 *Glaciol.* 53 (182), 442–448.
- 1637 Whitlow, S., Mayewski, P. A., Dibb, J. E., 1992. A comparison of major chemical
1638 species seasonal concentration and accumulation at the South Pole and Summit,
1639 Greenland. *Atmos. Environ.* 26A (11), 2045–2054.

- 1640 Wilen, L. A., DiPrinzio, C. L., Alley, R. B., Azuma, N., 2003. Development,
1641 principles, and applications of automated ice fabric analyzers. *Microsc. Res.*
1642 *Technique* 62, 2–18.
- 1643 Wilhelms, F., Sheldon, S. G., Hamann, I., Kipfstuhl, S., 2007. Implications for
1644 and findings from deep ice core drillings - an example: The ultimate tensile
1645 strength of ice at high strain rates. In: Kuhs, W. F. (Ed.), *Physics and Chemistry*
1646 *of Ice* (The proceedings of the International Conference on the Physics and
1647 *Chemistry of Ice* held at Bremerhaven, Germany on 23-28 July 2006). No. 311
1648 in *The Royal Society of Chemistry Special Publication*. pp. 635–639.
- 1649 Wilson, C. J. L., Russell-Head, D. S., Sim, H. M., 2003. The application of an
1650 automated fabric analyzer system to the textural evolution of folded ice layers
1651 in shear zones. *Ann. Glaciol.* 37, 7–17.
- 1652 Wingham, D. J., Siegert, M. J., Shepherd, A., Muir, A. S., 2006. Rapid discharge
1653 connects Antarctic subglacial lakes. *Nature* 440, 1033–1036.
- 1654 Woodcock, N. H., 1977. Specification of fabric shapes using an eigenvalue
1655 method. *Geol. Soc. Amer. Bull.* 88, 1231–1236.
- 1656 Woods, G. A., 1994. Grain growth behavior of the GISP2 ice core from central
1657 Greenland. Technical Report 94-002, Pennsylvania State University, Earth Sys-
1658 tem Science Center, University Park, PA.
- 1659 Young, N. W., 1979. measured velocities of interior East Antarctica and the state
1660 of mass balance within the I.A.G.P. area. *J. Glaciol.* 24 (90), 77–87.

1661 **Appendix A. FIGURE CAPTIONS**

Figure A.1: Maps of Antarctica and Greenland indicating the drilling sites of the ice cores described in this work. The gray zones in Antarctica indicate ice shelves (i.e. floating ice) and the cross marks the Geographic South Pole.

Figure A.2: Summary of the main features of the grain-size profiles and lattice preferred orientations (LPOs) of Greenlandic deep ice cores (for Antarctic ice cores, see Fig. A.3). Grain sizes are described by the average grain cross-sectional area and the LPOs by *c*-axis pole figures. The symbols *||||* and *~~~~* denote ice frozen to bed and detected subglacial water at bed, respectively (cf. Table B.1). Notice that the profiles and pole figures summarized here are mere outlines that do not display the details and variability of the original data, available in the following references. Camp Century: Herron and Langway (1982). Dye 3: Herron et al. (1985); Langway et al. (1988). GRIP: Thorsteinsson et al. (1997). GISP2: Gow et al. (1997). NGRIP: Wang et al. (2002); Svensson et al. (2003b).

Figure A.3: Summary of the main features of the grain-size profiles and lattice preferred orientations (LPOs) of Antarctic deep ice cores (for Greenlandic ice cores and further explanations, see Fig. A.2). Notice that the profiles and pole figures summarized here are mere outlines that do not display the details and variability of the original data, available in the following references. Byrd Station: Gow and Williamson (1976). Vostok: Lipenkov et al. (1989). EDC: EPICA community members (2004); Durand et al. (2009). EDML: Seddik et al. (2008); Weikusat et al. (2009b). Dome F: Azuma et al. (1999, 2000).

Figure A.4: Modern methods for visualization of multiscale structures of polar ice on the field. *Top left*: Microstructure Mapping (μ SM) mosaic image of two cloudy bands in the bubble–hydrate transition zone (EDML, 954 m depth). The high concentration of microinclusions make the cloudy bands appear darker than the surrounding ice in this image. Contrasting differences in average grain size and shape are evident between the cloudy and “clean” ice. Bright objects are air hydrates preserved inside the ice, while black objects are air bubbles, or decomposing air hydrates on the sample surface. Scale bar: 1 mm. *Right*: Linescan image of a one-meter long ice core piece (EDML, 1092–1093 m depth). Notice that linescan images are produced by light scattering from the side, against a dark background, and are therefore negative pictures of the core. Brighter bands indicate stronger light scattering due to a higher concentration of impurities (viz. cloudy bands). (From Faria et al., in preparation). *Bottom left*: Automatic Fabric Analyser (AFA) mosaic trend image of a thin section of Greenlandic ice (NEEM, 822.3 m depth). The color and brightness describe respectively the azimuth and colatitude of the c -axis orientation. Scale bar: 10 mm.

1662 **Appendix B. TABLES**

1663 (See next page)

Table A.1: Overview of ice cores mentioned in the text. The cores EDC96 and Dome F 1 are not listed below because their data are identical to those given for EDC99 and Dome F 2, respectively (with the exception of core length and drilling period, which can be found in the main text). A question mark within brackets indicates a tentative input derived from indirect information. In some cases, certain data appear with slight variations in the literature, depending on the employed methods and definitions. In such situations we tried to identify and adhere to the more detailed sources.

ice core acronym	project / expedition name	site	position	elevation a.s.l. (m)	core length (m)	vert. ice thickness (m)	drilling period	topography	surface speed (m/a)	ice-equiv. acc. rate (cm/a)	surface T ann. aver. ($^{\circ}\text{C}$)	borehole bottom T ($^{\circ}\text{C}$)	bed conditions
NBSAE ^[1]	Norwegian-British-Swedish Antarctic Expedition	Maudheim, Quar Ice Shelf, Antarctica	71°03'S, 10°55'W	37	< 100	≈ 270	1949–1952	ice shelf	≈ 130	≈ 40	≈ -17	≈ -15	floating on sea
JIRP ^[2]	Juneau Ice Field Research Project	Taku Glacier, Alaska	58°37'N, 134°15'W	1086	89	≈ 300	1949–1950	tidewater glacier	> 200	40.4	-0.9	(?) ≈ 0	frozen to bed
EPF Camp VI ^[3]	Expéditions Polaires Françaises	Camp VI, Greenland	69°42'N, 48°16'W	1598	125	1350	1950–1951	flank (?)	≈ 140	59.7	-13	≈ -12	(?)
EPF Station Centrale ^[4]	Expéditions Polaires Françaises	Station Centrale, Greenland	70°54'N, 40°37'W	3000	150	3000	1950–1951	flank (?)	≈ 18	35.9	-28	≈ -25	(?)
Camp Century ^[5]	Camp Century Ice Core Project	Camp Century, Greenland	77°11'N, 61°08'W	1884	1375	1375	1963–1966	flank (?)	5.5	35.0	-24	-13	frozen to bed
Byrd ^[6]	Byrd Ice Core Project	Byrd Station, Antarctica	80°01'S, 119°31'W	1530	2164	2164	1966–1968	flank (?)	12.7	39.1	-28	-1.8	subglacial water
Dye 3 ^[7]	Greenland Ice Sheet Project (GISP) Dye 3	Dye 3 Station, Greenland	65°11'N, 43°49'W	2480	2037	2037	1979–1981	flank	12.5	53.5	-20	-12	frozen to bed (?)
GRIP ^[8]	Greenland Ice-core Project	Summit Station, Greenland	72°35'N, 37°38'W	3231	3029	3029	1989–1992	dome	0.25	23.0	-32	-9	frozen to bed
GISP2 ^[9]	Greenland Ice Sheet Project 2	GISP2 Station, Greenland	72°35'N, 38°28'W	3200	3053	3053	1989–1993	dome	1.7	21.5	-31	-9	frozen to bed
NGRIP ^[10]	North Greenland Ice-core Project	NGRIP Station, Greenland	75°06'N, 42°19'W	2917	3085	3085	1996–2004	ridge/divide	1.33	19.5	-32	-2.4	subglacial water
Vostok 5G ^[11]	Vostok Ice Core Project	Vostok Station, Antarctica	78°28'S, 106°48'E	3488	3769	3755	1990–2011	over a lake	1.5	2.5	-57	-3	subglacial lake
EDC99 ^[12]	EPICA Dome C	Concordia Station, Antarctica	75°06'S, 123°21'E	3233	3260	(?) 3273	1999–2005	dome	0.015	4.8	-55	-2.3	subglacial water (?)
EDML ^[13]	EPICA Dronning Maud Land	Kohnen Station, Antarctica	75°00'S, 0°04'E	2892	2774	2782	2001–2006	ridge/divide	0.74	7.0	-45	-3	subglacial water
Dome F 2 ^[14]	Dome Fuji Ice Core Project	Dome Fuji Station, Antarctica	77°19'S, 39°42'E	3810	3035	3090	2003–2007	dome	0.068	3.5	-58	-2	not frozen to bed

^[1] Schytt (1958); Lythe and Vaughan (2001); Schlosser and Oerter (2002); Van den Broeke et al. (2009); Rignot et al. (2011); <http://www.spri.cam.ac.uk/resources/expeditions>. ^[2] Miller (1954); Pelto and Miller (1990); Nolan et al. (1995); Stafford et al. (2000); <http://geonames.usgs.gov/pls/gnispublic>. ^[3] Holtzschner et al. (1954); Robin (1955); Wegener (1955); Frisrup (1959); Meriväat et al. (1973); Haeferli and von Sury (1975); Weidick and Bennike (2007). ^[4] Holtzschner et al. (1954); Robin (1955); Frisrup (1959); Victor and Bauer (1961); Meriväat et al. (1973); Haeferli and von Sury (1975); Dansgaard (2004). ^[5] Hansen and Langway (1966); Paterson (1983); Herron and Langway (1982); Dansgaard (2004); <http://gombessa.tripod.com/scienceleadstheway/id9.html>. ^[6] Gow (1968); Gow and Williamson (1976); Paterson (1983); Gow et al. (1997); Dansgaard (2004); <http://www.ncdc.noaa.gov/paleo/icecore/antarctica>. ^[7] Reeh et al. (1978); Paterson (1983); Gundestrup and Hansen (1984); Dahl-Jensen and Gundestrup (1987). ^[8] Johnsen et al. (1995); Hvidberg et al. (1997); Thorsteinsson et al. (1997); <http://www.esf.org/activities>. ^[9] Gow et al. (1997); Hvidberg et al. (1999); <http://www.ncdc.noaa.gov/paleo/icecore>; <http://www.gisp2.sr.unh.edu>. ^[10] Dahl-Jensen et al. (2002); Svensson et al. (2003b); <http://www.ncdc.noaa.gov/paleo/icecore/greenland/summit>. ^[11] Lippenkov et al. (1989); Ritz (1989); Kapitza et al. (1996); Siegert and Kwok (2000); Bell et al. (2002); Oerter and Baker (2007); Vasiliev et al. (2007, 2011); Jones (2012). ^[12] EPICA community members (2004); Vituani et al. (2004); Augustin et al. (2007); Durand et al. (2009). ^[13] Wésche et al. (2007); Wilhelms et al. (2007); Oerter et al. (2009). ^[14] Dome-F Deep Coring Group (1998); Watanabe et al. (1999b); Motoyama et al. (2008).

Table A.2: Summary of the most essential features of the deep ice cores discussed in this work.

ice core	features
NBSAE	<ul style="list-style-type: none"> • First deep ice core from Antarctica. • Pioneering microstructural investigations of deep polar ice and ice-shelf ice.
JIRP	<ul style="list-style-type: none"> • First deep ice core from a polar glacier.
EPF	<ul style="list-style-type: none"> • First deep ice cores from Greenland.
Camp Century	<ul style="list-style-type: none"> • First deep ice core to reach the base of a polar ice sheet. • First continuous record of structure and chemical composition of a polar ice sheet. • Clustering of <i>c</i>-axes is stronger in fine-grained layers with high impurity content (cloudy bands).
Byrd	<ul style="list-style-type: none"> • Established the tripartite paradigm. • Systematic study of cloudy bands. • Consistent relation between grain sizes, <i>c</i>-axis orientations, and impurity content: the higher the impurity content, the smaller the grains and the stronger the <i>c</i>-axis clustering. • First problems with subglacial water upwelling.
Dye 3	<ul style="list-style-type: none"> • Established new standards of organization and efficiency for deep ice core field studies. • Strong correlation between impurity content and grain size, but no clear relation to <i>c</i>-axis preferred orientations.
GRIP	<ul style="list-style-type: none"> • First multi-national European deep ice-core drilling project. • Clear correlation between impurity content and grain size, but no definite relation to <i>c</i>-axis preferred orientations, in agreement with Dye 3. • Microstructural similarity with GISP2 and previous cores was invoked as corroboration of the tripartite paradigm.
GISP2	<ul style="list-style-type: none"> • Discovery of crystal striping and its relation to folding. • Microstructural similarity with GRIP and previous cores was invoked as corroboration of the tripartite paradigm.
NGRIP	<ul style="list-style-type: none"> • First multi-continental deep ice-core drilling project. • First deep ice core to be partially analyzed with an automated Ice-core Line-Scanner (ILS). • First deep ice core to be partially analyzed with the method of Microstructure Mapping (μSM). • Evidences that the microstructure of polar ice even at shallow depths is usually not in equilibrium, in contradiction to the premises of the tripartite paradigm.
Vostok	<ul style="list-style-type: none"> • Longest ice core ever drilled. • First deep ice core from the top of a large subglacial lake. • Remarkably regular microstructure points to an ice flow above the lake comparable to that of ice shelves (insignificant horizontal simple shearing) and negligible dynamic recrystallization. • Further support to the correlation between grain size and impurity content.
EDC	<ul style="list-style-type: none"> • Oldest ice to date. • Further discrepancies between the observed ice microstructure and the tripartite paradigm. • Evidence of grain-boundary pinning by dust particles, giving further support to the correlation between grain size and impurity content.
EDML	<ul style="list-style-type: none"> • First ice core to have continuous and thorough records of visual stratigraphy (via ILS) and microstructure (via μSM) in microscopic resolution. • Evidence of strain accommodation by microscopic grain-boundary sliding via microshear in warm cloudy bands at the MIS5e–MIS6 transition. • Evidence of dynamic recrystallization throughout the core, including the firm layer, in direct contradiction to the tripartite paradigm. • Evidence that non-basal dislocations play a decisive role in heterogeneous strain accommodation through the formation of subgrain boundaries.
Dome F	<ul style="list-style-type: none"> • Deepest and oldest ice from the Atlantic Sector of Antarctica to date. • First deep ice core (Dome F 1) to be crystallographically investigated with an Automatic Fabric Analyzer (AFA). • Clustering of <i>c</i>-axes is weaker in layers with high impurity content and small grain sizes, in direct contrast to Camp Century and Byrd. • Possible activation of diffusion creep at low temperatures and stresses within cloudy bands.

Figure 1
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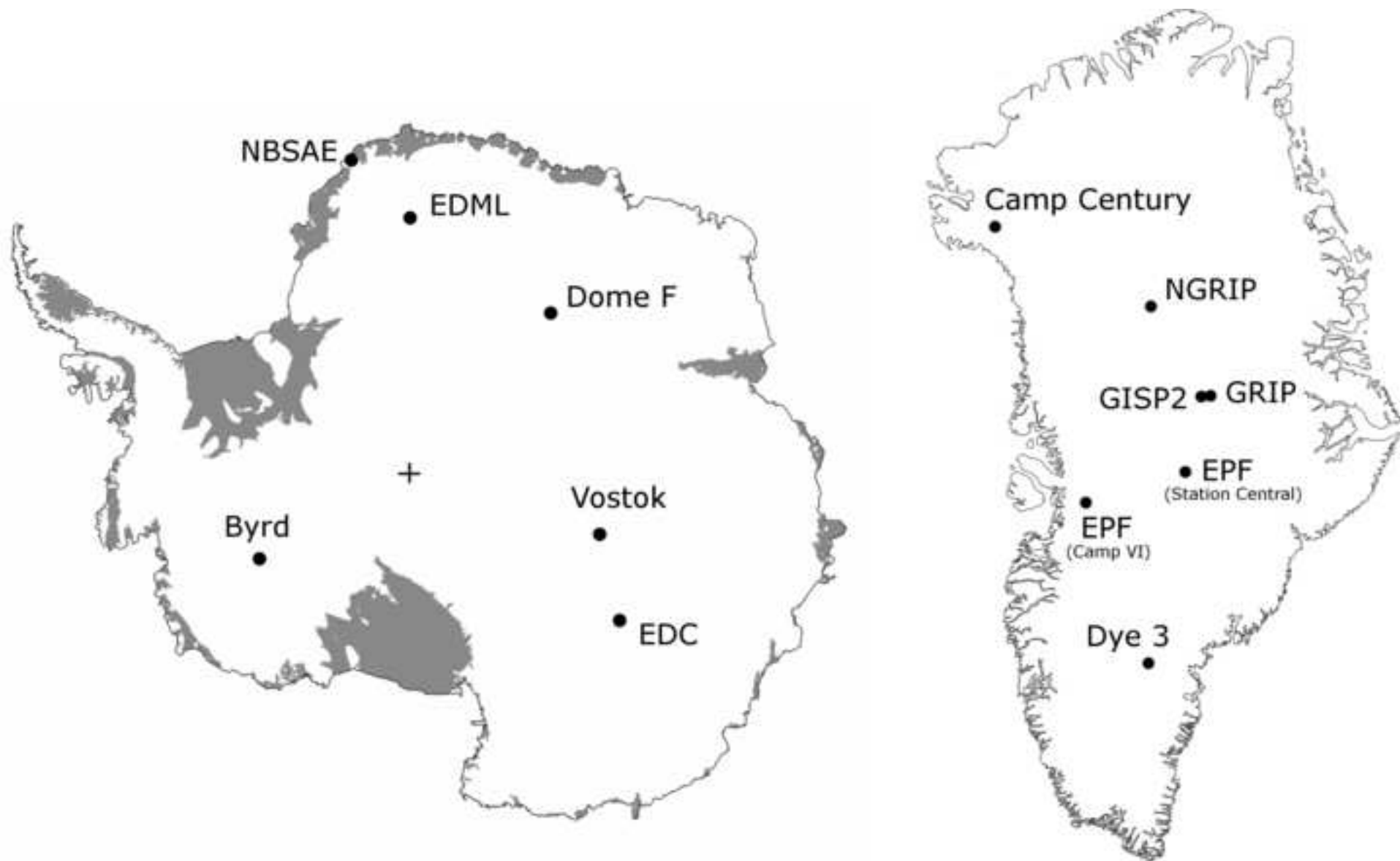


Figure 2
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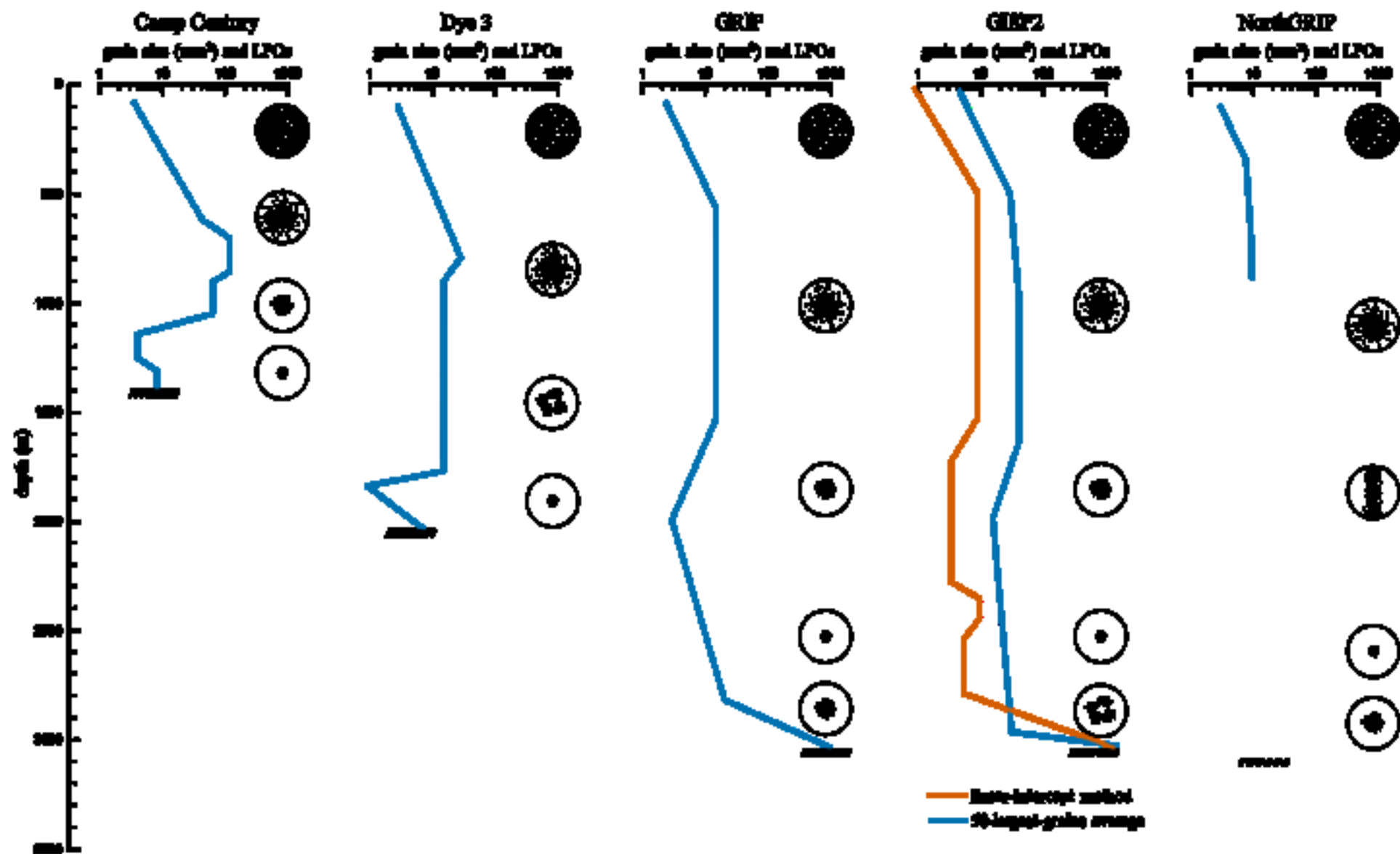


Figure 3
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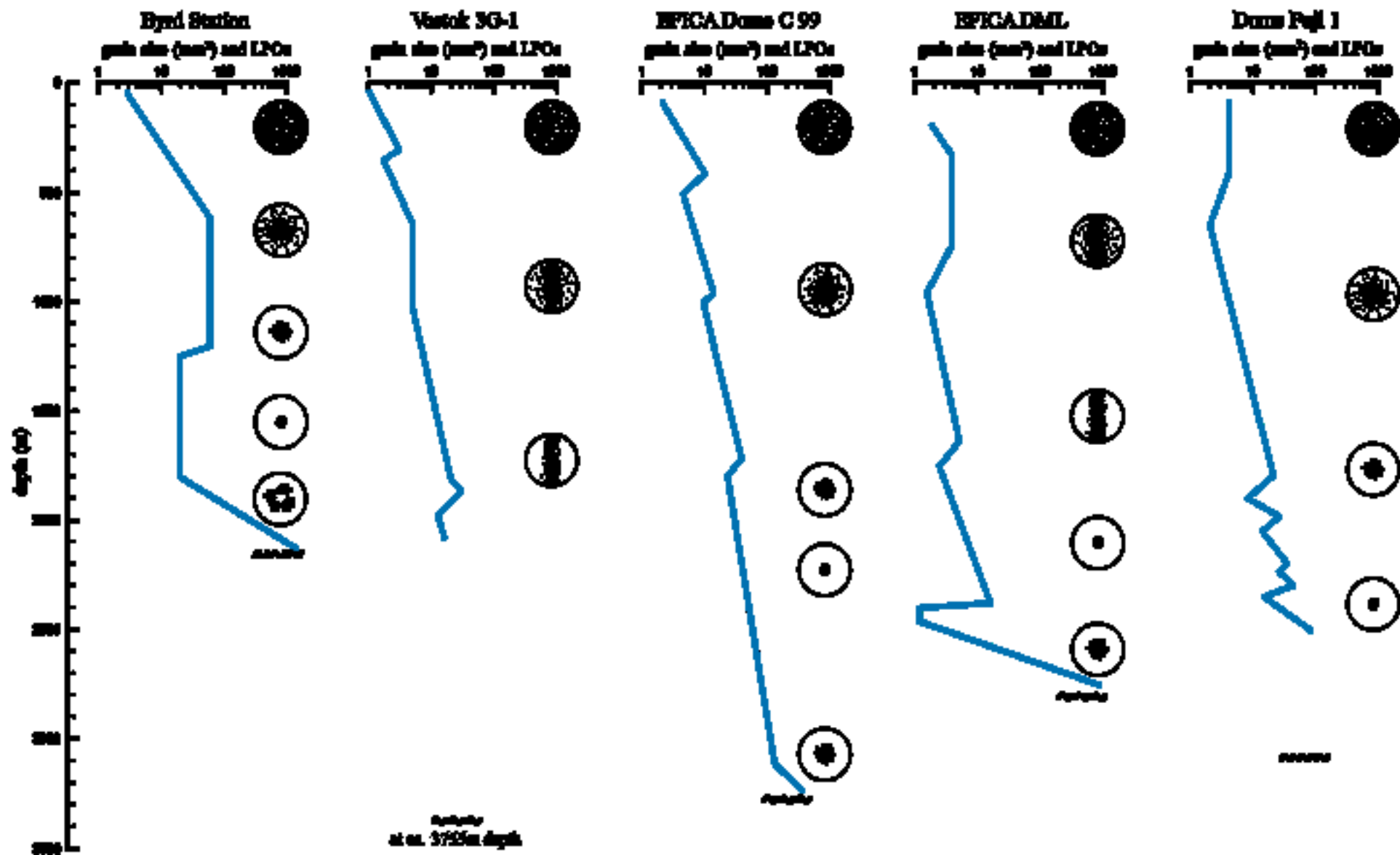
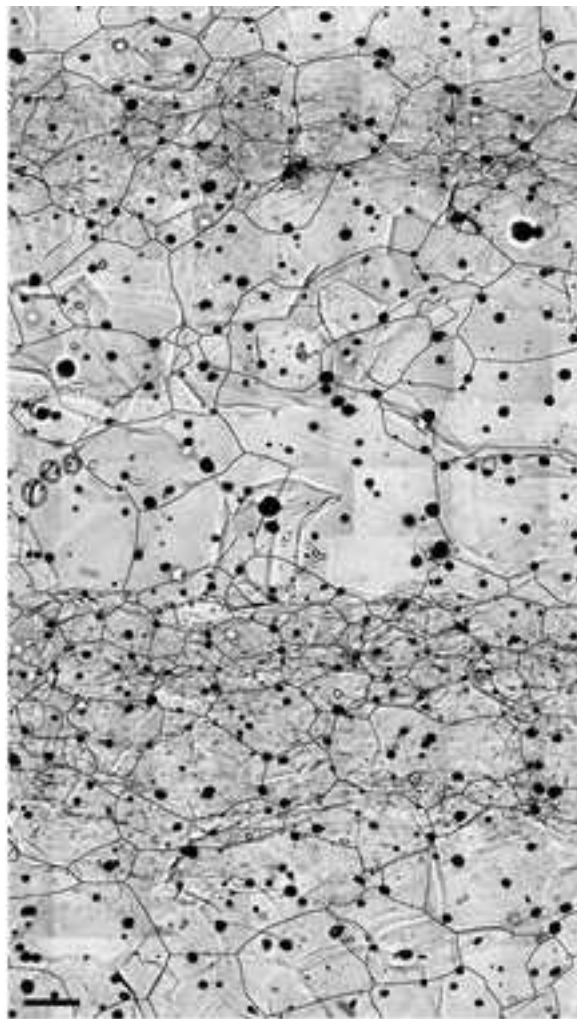


Figure 4
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