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Hydrographic and climatic changes influencing the proglacial Druzhby drainage system, Vestfold Hills, Antarctica

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Abstract: Freshwater drainage systems, fed by melting of nearby inland ice and perennial snowdrifts, exist in the south-eastern part of arid, ice-free, coastal Vestfold Hills. Most important is the complex Druzhby system. Intriguing questions arise about the conditions controlling its seasonal development and about the consequences of possible changes of those conditions. A water-balance was calculated for 1990–91. The system went through four seasonal phases of which only one displayed a fully developed external drainage. Evidently, the relative duration of those phases can vary considerably from one year to another. The system depends critically on the water supply from ice-dammed Chelnok Lake, which could readily be drained by a minor retreat of Sørsdal Glacier. Exposed to excessive evaporation, the large Crooked Lake would then become internally drained and reach a new equilibrium in ~830 years. A Crooked Lake sediment core can be interpreted as suggesting this occurred during the Holocene. The idea, inferred from striae, of a late Holocene Chelnok Glaciation reaching the northern shores of Crooked Lake is questioned. Instead, it is suggested that the Chelnok striae originate from local basal melting of the ice sheet draining southward into a deglaciated Sørsdal trough. At present, runoff is determined by opposing short and long-term climatic influences.

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Key words: climatic change, Crooked Lake, Druzhby, hydrography, Vestfold Hills, water-balance

Introduction

The Vestfold Hills are a large (410 km^2), ice-free coastal region on the north-eastern shore of Prydz Bay, East Antarctica (68°35'S, 78°10'E) (Fig. 1). This hummocky lowland of Precambrian rocks is bordered to the east by the continental ice sheet and to the south by an outlet glacier, Sørsdal Glacier. Mean annual temperature is -10.2°C, with a highest recorded temperature of +13.0°C and a lowest recorded of -38.3°C. In the eastern parts of the area, frequent east–north-easterly katabatic winds occur at an average speed of 5.0 m s^{-1} (Pickard 1986a). A great number of lakes of various sizes exist in the area. Some of them are saline or hypersaline, others are fresh. The lakes are affected by high evaporation and a scarce supply of seasonal meltwater from adjacent snowpatches and glacier ice (Adamson & Pickard 1986a, Bronge 1989, Pickard 1986b). Freshwater drainage systems similar to those of the Vestfold Hills exist in several Antarctic ice-free areas, such as the Bunger Hills (Wisniewski 1983, Colhoun & Adamson 1989, Kaup *et al.* 1993), Wohltat Mountains (Loopmann *et al.* 1986), the Schirmacher Oasis (Richter 1983, Loopmann & Klokov 1986, Richter & Haendel 1989) and the Dry Valleys of Victoria Land. Most thoroughly investigated is the hydrology of the lakes and rivers of the Dry Valleys (Chinn 1993). The largest freshwater system in the Vestfold Hills is the Druzhby drainage system (Tierney 1975, Adamson & Pickard 1986a), which was gauged by Colbeck (1977) and by Gallagher & Burton (1988) at Ellis Rapids, the main outlet (Fig. 1). A tributary system, the Chelnok, was gauged by Bronge (1989). Changing climatic conditions as well as fluctuating positions of the margin of Sørsdal Glacier have had a profound influence on the hydrography of the area (Adamson & Pickard 1983, 1986a). The aim of this study is to quantify the hydrological inputs and outputs of the Druzhby drainage system, to state the relative importance of the different source areas contributing meltwater to the system, and to assess the conditions necessary for the existence of the Druzhby system in its present form. In order to calculate a water-balance for the drainage system, the various sources of inputs and outputs must be identified and quantified as far as possible. Thus, the fieldwork was focused on measuring of stream flow, evaporation and glacier ablation.

Physical settings and methods

The Druzhby drainage system is comprised mainly of Druzhby Lake, Pauk Lake, Crooked Lake and Chelnok Lake (Fig. 1). Pauk Lake receives meltwater from the ice sheet and from large perennial snowdrifts along the ice margin. It drains into Druzhby Lake through a chain of smaller lakes. Chelnok Lake is a collecting basin for meltwater from Sørsdal Glacier, and the outlet, Tierney Creek, runs via Tierney Falls through two smaller lakes before it enters Crooked Lake. With an areal extent of 8.54 km^2 and a maximum recorded depth of 140 m (Adamson & Pickard 1986a), it is one of the largest Antarctic lakes. Crooked Lake also receives meltwater from the ice sheet via "Corner Lakes" (unofficial name) and

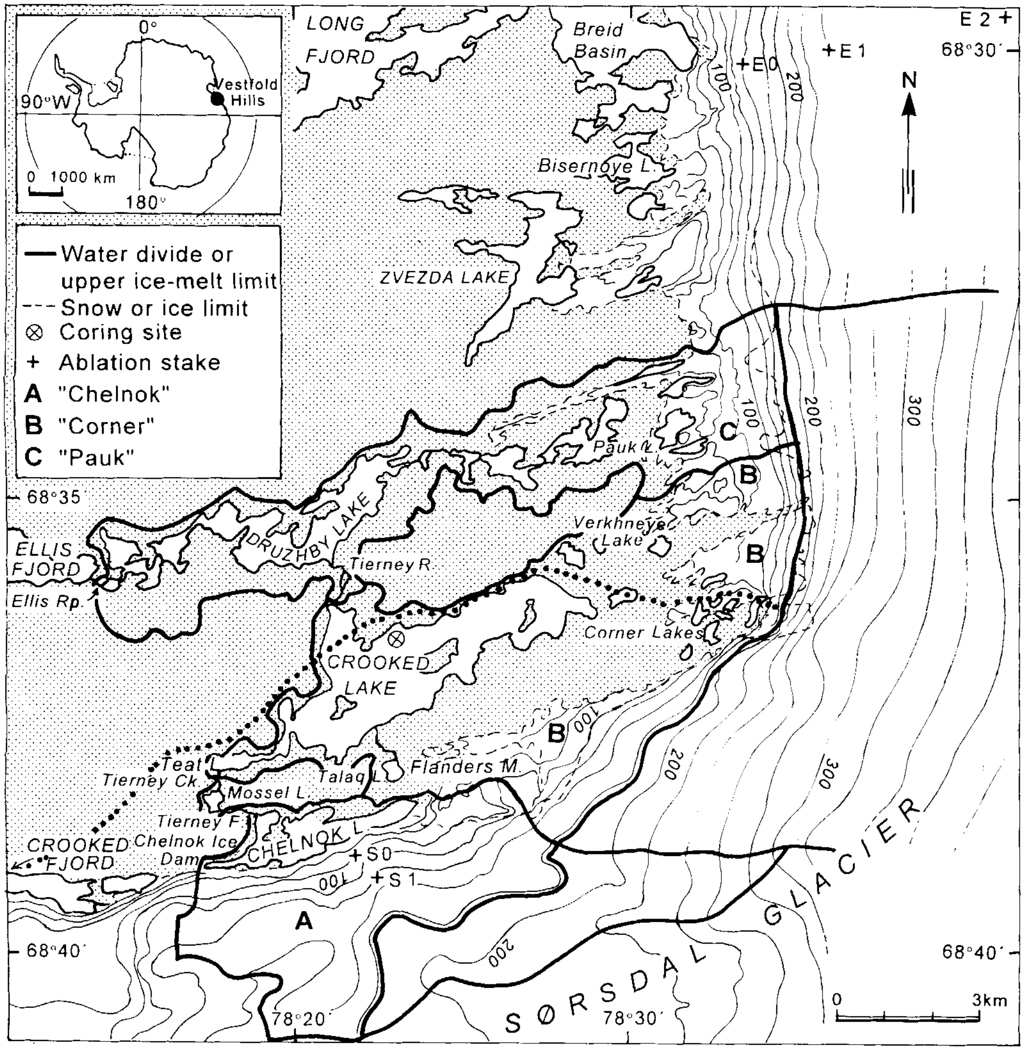

Fig. 1. Map of south-east Vestfold Hills, with the Druzhby system and its drainage basins. Shaded area is ice-free. Letters A, B and C refer to glaciated or permanently snow-covered parts of the sub-basins, where surface melt occurs. The dotted line indicates the northern limit of Chelnok striae. The letters E and S followed by a number refer to ablation stakes. Stake S2 is not shown (68°41.72'S, 78°27.34'E).

Table I. Areas within the Druzhby drainage basin.

	Area (km^2)	Σ (km^2)	Lakes (%)	Ice-covered (%)
Druzhby-Pauk drainage basin	24.94		28	
(ice-free part)		31.79		22
"Pauk" area	6.85			100
Druzhby-Pauk lakes	6.26		100	
Trajer & other lakes	0.65		100	
Crooked drainage basin (ice-free part)	37.43		31	
Mossel drainage basin	0.56	52.62		28
"Crooked" area	14.63			100
Crooked Lake	8.54		100	
Teat Lake	0.25		100	
Mossel Lake	0.17		100	
Talg, Corner, Verkhneye lakes & others	2.80		100	
Chelnok drainage basin	3.39		58	
(ice-free part)		22.92		85
"Chelnok" area	19.53			100
Chelnok Lake	1.98		100	
Druzhby drainage basin		107.33	31	38

directly from Sørsdal Glacier. From Crooked Lake an outlet, Tierney River, leads to Druzhby Lake which in turn has a short outlet, Ellis Rapids, to the sea. The surface of Crooked Lake drops below the level of its outlet threshold during the winter due to ablation from the frozen surface (Colbeck 1977).

The flow variations of Ellis Rapids only reflect faintly what is happening on the ice sheet and the glacier due to the reservoir effects of all the lakes along the watercourses. The discharges from Pauk Lake and Chelnok Lake most closely reflect the runoff from the ice. After exclusion of areas with internal drainage, the terrestrial part of the Druzhby drainage basin is much smaller than that delineated by Colbeck (1977). The drainage basin boundaries on land can be assumed to follow closely the shorelines of the fjords, lakes and creeks (Adamson & Pickard 1986a, Gallagher & Burton 1988). The major exceptions are places where large snowdrifts form. A large part of the Druzhby drainage basin is part of the ice sheet and Sørsdal Glacier (Fig. 1, Table I). The upper boundaries are defined by the altitude above which sublimation is the only ablation process, and hence no meltwater is

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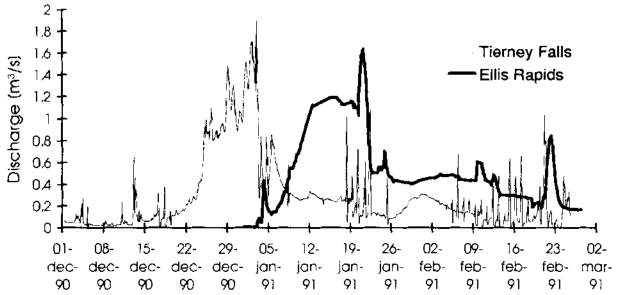

Fig. 2. Hydrographs for Ellis Rapids and Tierney Creek, summer 1990–91.

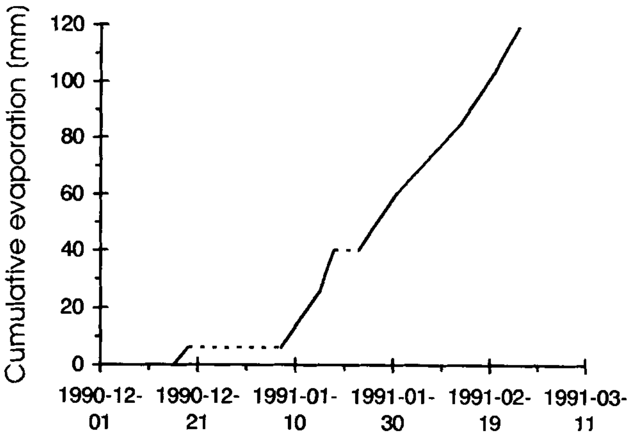

Fig. 3. Cumulative evaporation at Mossel Lake, summer 1990–91.

generated. Below these boundaries there is blue ice only, but further up on the ice (*c.* 5 km) snowpatches begin to appear. The patches become more numerous and larger the farther inland they are located. Along the margin of the ice sheet there is a zone of perennial snowdrift, formed by the prevailing north-north-easterly winds on the lee-side of the sloping ice front. The contours on the topographic map of the Vestfold Hills and on the Russian Antarctic Map (Yegorova *et al.* 1966) give a fairly accurate view of the glacier topography. The ice flow may have altered some minor features of it, but the general shape of the ice cannot have changed substantially in the relatively short period since these maps were surveyed (~30 years). To the east of the margin, the ice sheet surface rises relatively steeply upwards and soon attains high altitudes (Fig. 1). This contrasts with Sørsdal Glacier which rises more gently upwards from a north-facing, marginal ice cliff, about 15 m high. In the marginal parts of the ice sheet there are several deeply incised meltwater gullies (Adamson & Pickard 1986a, Gore 1993). In some areas of Sørsdal Glacier there are crevasses, capable of diverting meltwater streams in other directions than the contour lines would indicate. Field observations have shown that this phenomenon is restricted to relatively small areas only. The areal extents of the lakes and the drainage basins are shown in Table I. An area of special interest is Flanders Moraine (Fig. 1), a stagnant, or nearly so, body of ice covered by a 1–3 m thick layer of till, and subject to thermokarstic melting processes (Pickard 1983, Adamson & Pickard 1986a, Pickard 1986b, Fitzsimons 1989, Lundqvist 1989). Meltwater from this area is diverted both to the Chelnok sub-basin and to the rest of the Druzhby drainage basin (Fig.1).

To determine runoff from the entire Druzhby drainage basin and from the Chelnok sub-basin, a gauging station was established just above Ellis Rapids, where stage height was continuously recorded by an Ott mechanical recorder. Another gauging station was established in Tierney Creek, just downstream from Tierney Falls. There, a Campbell CR-10 datalogger was used for monitoring the stage height throughout the season. An APC pressure sensor was connected to it. The sensor body was firmly attached to the bottom of the creek in a position preventing it from being influenced by hydrodynamic pressure. The sensor had a measuring range of 0–200 hPa, which corresponds to 0–2 m water depth. Discharge tests were made using an integrating conductivity meter and the instantaneous salt dilution method (Østrem 1964, Bronge & Openshaw 1996).

An evaporation pan was attached to the bottom in the shallow near-shore water of Mossel Lake (Fig. 1), in a position sheltered from waves.

Ice ablation was measured from two lines of stakes, the east (E) and south (S), each consisting of three white-painted wooden stakes, aligned approximately perpendicularly to the margins of the ice sheet and Sørsdal Glacier, respectively (Fig. 1).

Results and discussion

Discharge

Stage height was logged in Tierney Creek from 1 December 1990–25 February 1991. During this period 25 discharge tests were carried out. A detransformed logarithmic linear regression made on these measurements gave the rating equation:

$$Q = 0.00038 \times h^{2.36483} \qquad (1)$$

with discharge, Q, in m³ s⁻¹ and stage height, h, in cm. Equation (1) applied to the stage height record gave the season's hydrograph (Fig. 2). The seasonal runoff was calculated to be 2.19×10^6 m³, which corresponds to an annual mean specific value of 3.03 l s^{-1} × km². This can be compared with the 3.59×10^6 m³ runoff calculated for the 1987–88 season (Bronge 1989).

Flow through Ellis Rapids did not start until 1 January 1991, and the recording went on until 27 February 1991 when the gauging station was closed because the creek began to freeze and further recordings would have become unreliable. Following Equation (1), a rating equation was calculated, using 11 discharge tests:

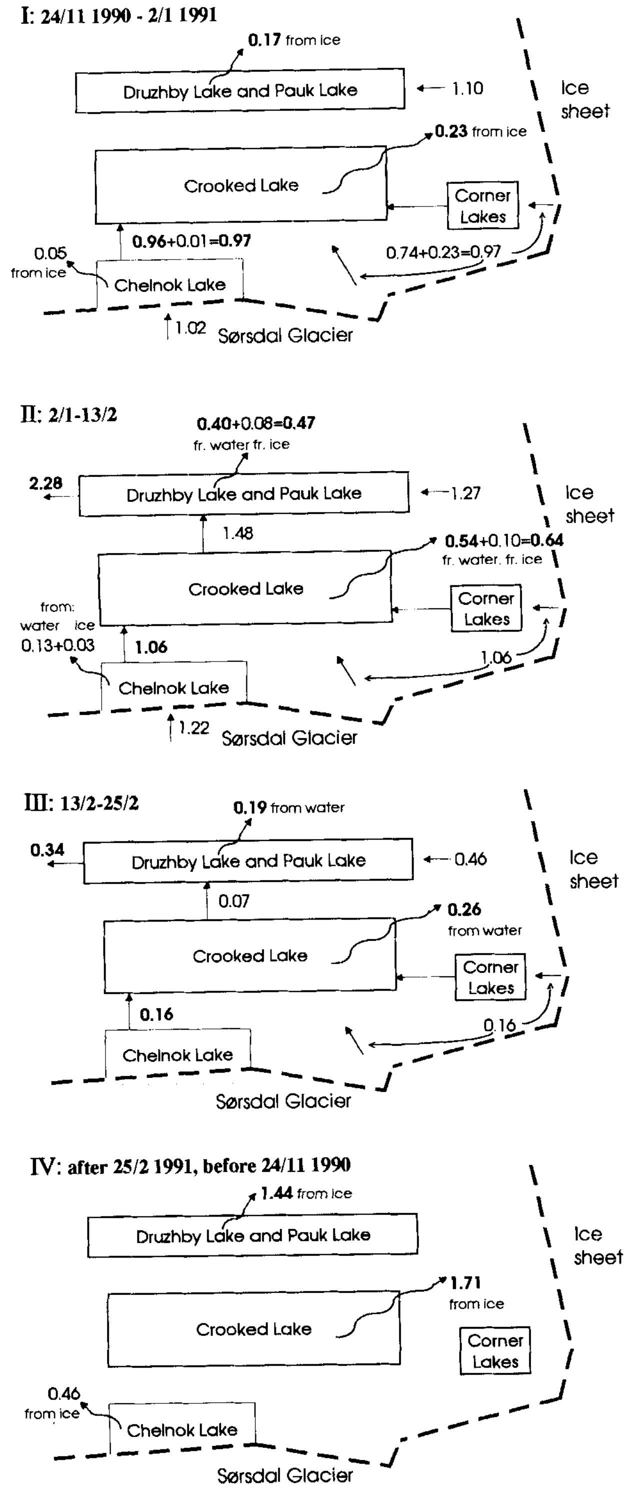

Fig. 4. Seasonal development of the Druzhby drainage system, runoff season 1990–91. Amounts of water given in 10^6 m³. Measured values in bold.

$$Q = 0.00157 \times h^{1.95947} \quad (2)$$

The resulting hydrograph is shown in Fig. 2. In this case, the seasonal runoff amounted to 2.62×10^6 m³, corresponding to an annual mean specific value of 0.78 l s^{-1} × km². Gallagher & Burton (1988) found that the discharge at Ellis Rapids during the 1985–86 summer was almost zero. From satellite passive-microwave data for 1978–87, Zwally & Fiegles (1994) found that the extent and duration of surface melting on Antarctic ice shelves and ice sheet margins was at its lowest in the 1985–86 summer (November–February). On the other hand, Colbeck (1977) recorded an annual runoff of 4.49×10^6 m³ from 16 February 1975–15 February 1976. He also observed that the flow continued beneath the ice of the rapids until May despite very low air temperatures. Similarly, the flow during 1991 can be expected to have continued some time after 27 February due to recession of Druzhby Lake. With the assumption of one month's receding flow after that date, the seasonal runoff would have to be adjusted with another 0.17×10^6 m³ up to 2.79×10^6 m³. Accordingly, the annual runoff recorded during 1990–91 at Ellis Rapids and Tierney Falls respectively, differ considerably from what Colbeck (1977), Gallagher & Burton (1988) and Bronge (1989) found for the seasons 1975–76, 1985–86 and 1987–88 respectively. Thus, the flows of the Vestfold Hills rivers display a high interannual variability. Similar conditions were found in the Dry Valleys from hydrological data recorded over 20 years (Chinn 1993). In fact, this holds for the entire rim of the Antarctic continent, according to Zwally & Fiegles (1994).

Evaporation

The evaporation pan was in operation from 16 December 1990–25 February 1991. There were two breaks in the record, one of them when the pan was submerged by the rising lake water (19 December 1990–7 January 1991), the other when the water in the pan froze through and cracked it (18 January–23 January 1991). The record gave an average of 2.54 mm d^{-1} evaporated from open water areas (Fig. 3), corresponding to an annual value of 928 mm. This value falls well within the range of earlier measurements. Values between 650 mm yr^{-1} and 1570 mm yr^{-1}, depending upon the method used, were obtained by R. Taylor (personal communication, 21 March 1989), and an amount of 1220 mm yr^{-1} was calculated by Bronge (1989). The figures differ, not only due to various methods applied, but also due to interannual variability.

Precipitation

Precipitation in Antarctica is very difficult to measure due to the frequent strong winds (Colbeck 1977, Burton & Campbell 1980, Schwerdtfeger 1984, Streten 1986). Liljequist (1970) calculated an approximate annual value of 300 mm water equivalent for the Vestfold Hills area. Fitzsimons (1990) assumed it to be less than 250 mm. To what extent this value is applicable to the non-glaciated part of the area and the exposed ice cliff is somewhat uncertain, since the wind causes a very uneven distribution of the snow in the terrain.

Nevertheless, a great deal of the drainage basin consists of a relatively smooth glacier surface and lake ice. During the winter of 1988 the glacier surface was blown bare (A. Tymms, personal communication, 24 May 1989). This makes it reasonable to make the simplified assumption that precipitation was 250 mm over the relatively rugged non-glaciated area, and zero over the glaciated part of the drainage basin. The snow appeared in patches on the frozen lakes and it was estimated to cover about 50% of their surfaces.

Flow pattern

In early November 1990 the temperature had risen enough to allow ice melt, and on about 24 November, flow started in Tierney Creek and other small streams near the ice margin. During the intervening time Chelnok Lake was refilled to its outlet threshold. Similarly, before flow could start in Tierney River, Crooked Lake had to refill to its outlet threshold to compensate for the ablation loss during the preceding winter. Therefore, it was not until 2 January 1991 that any discharge took place in Tierney River, and flow continued until 13 February, when discharge was very low. On 1 January, flow started in the Ellis Rapids, induced by water coming from Pauk Lake and from snowmelt along Druzhby Lake. In mid-February inflow to Crooked Lake from Tierney Creek was quite low, and the flow at Ellis Rapids was merely caused by recession of the lakes upstream. A flow peak occurring in both records on 21–22 February was caused by strong winds blowing downstream, forcing water through the rivers.

During the year the system develops through four phases, corresponding to the four seasons (Fig. 4). At the beginning of the melt-season, the system acts as several separate internally drained systems. During spring (*phase I*) the system is recharged with meltwater until Crooked Lake starts spilling over through Tierney River and substantial amounts of water begin flowing at Ellis Rapids. This event marks the beginning of the hydrological summer, during which the system is fully developed (*phase II*). Similarly, the cessation of flow from Crooked Lake marks the beginning of autumn, a period of declining flows (*phase III*), which culminates in the change back to separate internally drained systems once again. Finally, during winter, the system is hydrologically inactive, though subjected to considerable ablation loss (*phase IV*). The magnitude of the seasonal runoff at Ellis Rapids 1990–91 ranged between the extremes recorded by Colbeck (1977) and Gallagher & Burton (1988) during the seasons 1975–76 (high discharge) and 1985–86 (low discharge), respectively. This indicates that conditions in 1990–91 were representative of the seasonal development of the Druzhby drainage system. The effect of the fluctuating climate from one year to another will show up as a change in the relative duration of the hydrological phases outlined in Fig. 4. Thus, a summer with a climate favourable for glacier-melt will expand *phase II* at the expense of *phases I* and *III*. If the opposite occurs, the system will never enter *phase II*, i.e. it will not develop into one single drainage system. This occurred in 1985–86 (Gallagher & Burton 1988) and in 1993–94 (J. Gibson, personal communication, 25 May 1994).

Water-balance

During winter 1990, ablation had lowered the surface of Crooked Lake 0.20 m below the level of its outlet threshold. This meant that an inflow of 1.71×10^6 m³ water was required to refill the lake up to its outlet level. Of this, 0.97×10^6 m³ came from Chelnok Lake, while the remaining 0.74×10^6 m³ must have come from "Corner Lakes", Talg Lake and a number of minor affluents. An additional 0.23×10^6 m³ must also have come from these miscellaneous sources to

Table II. Water balance for the Druzhby drainage basin 1990–91.

Development phase		Water out: (m³×10⁶)	Σ (m³×10⁶)		Water in: (m³×10⁶)	Σ (m³×10⁶)	Comments
I-III:	Q	2.62	2.62	M_G	1.02		
	E_L	0.19		M_G	0.97		
	E_L	0.26		M_G	1.10		
	E_L	0.16		M_G	1.22		
	E_L	0.47		M_G	1.06		
	E_L	0.64		M_G	1.27		
	E_L	0.17		M_G	0.16		
	E_L	0.23		M_G	0.46	7.26	
	E_L	0.05	2.17	M_S	1.14	1.14	Remaining snow equivalent to 2 cm.
				S	-1.44		Initial lowering of Druzhby,
				S	-1.71		Crooked & Chelnok lakes.
				S	-0.46	-3.61	
			4.79			4.79	
IV:	E_S	12.87	12.87	P	12.87	12.87	E_S and P estimated values of snow sublimation and drifting snow respectively.
	A_W	*		F_G	*		*Winter ablation and net inflow of ice can not be quantified.

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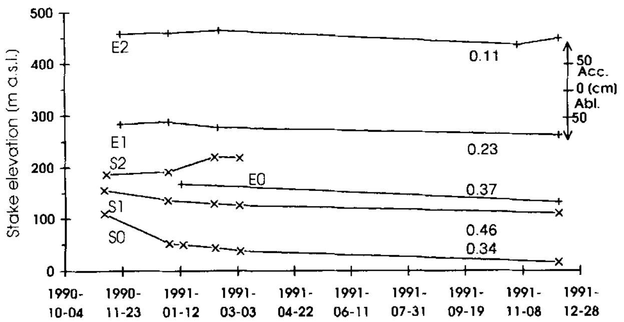

Fig. 5. Stake diagram showing stake lines E (east) and S (south) 1990-91. Numbers on the lines refer to surface lowering in m y^{-1}. The two first levels of stake E0 are inferred, not measured.

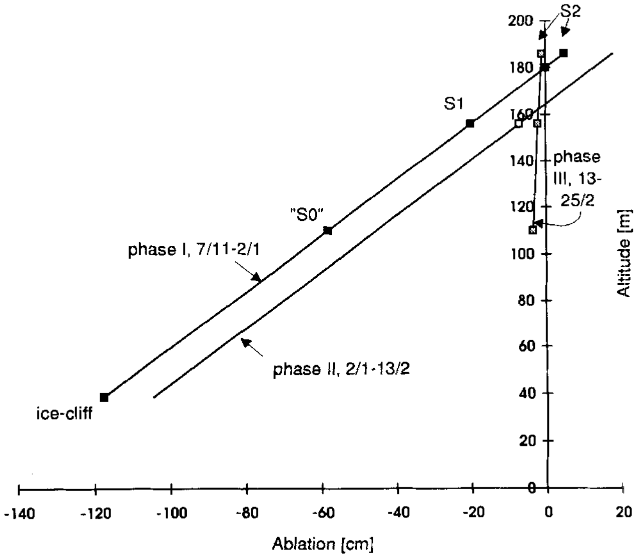

Fig. 6. Ablation as a function of altitude along stake line S (south).

compensate for ablation of the still frozen Crooked Lake during *phase I* (Fig. 4). That is, during *phase II,* 50% of the inflow to Crooked Lake came from Tierney Creek while the other 50% was derived from other sources. Accordingly, during *phase II,* 1.06×10^6 m³ was discharged at Tierney Creek, and an equal amount was contributed from the other sources. On 19 January 1991 the greater part of Crooked Lake was free of ice. Prior to that date, during *phase II,* 0.10×10^6 m³ was ablated from the frozen lake. During the remainder of *phase II,* 0.54×10^6 m³ evaporated from the open lake. This left 1.48×10^6 m³ of excess water to be discharged through Tierney River. Similarly, Druzhby Lake and Pauk Lake lost 0.47×10^6 m³ through ablation and evaporation during *phase II.* With 2.28×10^6 m³ discharged at Ellis Rapids, this gives an inflow of 1.27×10^6 m³ to Pauk Lake derived from surface melting on the ice sheet and from melting snowdrifts. During *phase III* the inflow to Crooked Lake was almost balanced by the evaporation from the lake, so an excess of only 0.07×10^6 m³ flowed through Tierney River. The bulk of the discharge at Ellis Rapids took place during *phase II* due to outflow from Crooked Lake, a pattern also observed by Colbeck (1977). Subsequent to the refill of Crooked Lake, 47% of the discharge at Ellis Rapids was derived from the tributary Chelnok Lake drainage basin. This fraction remained the same during both *phases II* and *III.* If the period of refilling Crooked Lake is included, the value increases to 84%. Gore's (1992) estimate of the seasonal discharge during the 1989–90 summer from "Corner Lakes" to about half the discharge at Ellis Rapids gains support from the 1990–91 record. It is obvious that the relatively small Chelnok catchment area plays a very important role for the contribution of water to the Druzhby drainage system.

The annual water-balance for the drainage basin is given by the equation:

$$A_W + E_S + Q + E_L = M_G + M_S + S + P + F_G \qquad (3)$$

where A_W is winter ablation, E_S is sublimation from snow, Q is runoff, E_L is evaporation and ablation from lakes, M_G is glacier melt, M_S is snowmelt, S is water storage in lakes, P is precipitation and F_G is net inflow of ice. In Table II the water-balance for 1990–91 is shown with the measured and calculated amounts of water. On the income side of the calculation, an amount of 1.14×10^6 m³ water remains. This volume can be attributed to snow transformed to meltwater, and corresponds to 20 mm w. eq. averaged over the drainage basin. Most of the winter snow is sublimated and is tentatively assigned the value of 12.87×10^6 m³. With the 1.14×10^6 m³ added, this corresponds to the assumed 250 mm of annual snowfall. This means that the only significant contribution to the annual water-balance during *phase IV* is water-storage in lakes, S, which is negative. Winter ablation and net inflow of ice do not necessarily need to be equal; according to Adamson & Pickard (1986a) and Gore (1993) a slight recession is going on at parts of the ice margin.

Glacier ablation

The ablation stakes (Fig. 1) could not all be drilled into the ice at once. Stakes S1, S2, E1, and E2 were established on 7, 9, 21 and 21 November 1990, respectively. Stakes E0 and S0 could not be placed until 14 January 1991, when most of the runoff had already taken place. The ablation at S0 prior to that date could be inferred from the other two stakes on Sørsdal Glacier, but attempts to do the same for E0 give ambiguous results. Therefore, calculation of the runoff from the ice sheet had to be based upon the ablation values recorded from stake line S. All the stakes except S2 could be retraced and measured in the next summer, on 9 December 1991. Some winter ablation (sublimation) was recorded then at all stakes (Fig. 5). During the melt-season (*phases I* and *II*) a considerable amount of superimposed ice was recorded at S2. Winter ablation was probably not sufficient to prevent

Table III. Ablation and runoff from the glaciated parts of the Druzhby drainage basin.

Area, time		Ablation [m³×10⁶ w.eq.]	Runoff [m³×10⁶]	Melt/ablation ratio
"Chelnok"	phase I	3.84	1.02	0.27
	phase II	4.20	1.22	0.29
"Crooked"	phase I	4.15	0.97	0.23
	phase II	4.57	1.06	0.23
"Pauk"	phase I	2.81	1.10	0.39
	phase II	3.22	1.27	0.39

that stake from being submerged by the ice the following summer. Generation of meltwater was restricted to *phase I* and *phase II*. The water-yielding part of the ablation area was restricted to a rather narrow marginal zone between the ice-free area and the contour line of 180 m elevation (Figs. 1 & 6). The relationship between ablation and altitude changed very little from *phase I* to *phase II*, when melting took place. It became very different once melting had ended for the season and ablation comprised sublimation only (Fig. 6). The sparse set of observations does not allow a more detailed study of the ablation, but nevertheless it can be used to calculate the major input of meltwater to the Druzhby drainage system. The recorded annual ablation values from stakes S0, S1 and E0 (0.34 m, 0.46 m and 0.37 m, respectively) are consistent with earlier data from Pickard (1983) and Pickard *et al.* (1984) who reported annual ablation values of 0.46 m and 0.3–0.4 m, respectively, measured on the ice sheet and on Sørsdal Glacier at about 120 m altitude. Parts of the edge of Sørsdal Glacier forms an ice cliff facing north. At "Chelnok" Lake the cliff foot lies at an altitude of 39 m a.s.l. Chinn (1987a) stated that unconfined ice cliffs may be subjected to 2 to 4 times higher ablation rates than level ice surfaces. From Fig. 6 it is also evident that most meltwater must be contributed by the very marginal part of Sørsdal Glacier. The glaciated part of the Druzhby drainage basin can be divided into three sub-areas, here named "Chelnok", "Crooked" and "Pauk" (Fig. 1). When the ablation is calculated for each of them and compared with their runoff during the runoff season (*phases I* and *II*, Fig. 4), it turns out that the sublimation constitutes a major part of the ablation (Table III). The ablation values for the "Chelnok" and "Crooked" sub-areas include water coming from the 0.05 m annual surface melt of Flanders Moraine (Pickard 1983).

The ratios between melt and ablation are average values for the three sub-areas and remain constant throughout the melting season (Table III). Closest to the ice-free area, melting forms a major part of the ablation and at 180 m a.s.l. it becomes insignificant (see also Chinn (1987a)). Most of the "Pauk" area consists of perennial snowdrifts and only minor areas are blue ice. This explains the relatively large proportions of meltwater supplied from the "Pauk" area (Table III). Before flow started at Tierney Falls (prior to 24 November) the ablation closest to Chelnok Lake amounted to 1.72×10^6 m³ water equivalents. The melt/ablation ratio of "Chelnok" *phase I*, 0.27 (Table III), is used to calculate the runoff from the "Chelnok" area of Sørsdal Glacier from 7–24 November. It turns out to have been 0.46×10^6 m³ corresponding to a 0.23 m lowering of Chelnok Lake below its outlet threshold during the preceding winter, a value similar to the 0.30 m sublimated from the frozen lakes of the Schirmacher Oasis (Richter & Haendel 1989) and the 0.30 m from the Dry Valleys lakes (Chinn 1993). A *phase Ia* could therefore be introduced as a period of refill of Chelnok Lake.

Hydrographic changes

What makes the Druzhby drainage system sensitive and prone to great hydrographic changes, is the important role played by Chelnok Lake and its drainage area, funnelling large amounts of water into the system. The present state of the system could be altered by glaciological or climatic changes or both. Parts of the Sørsdal Glacier edge are at present slowly retreating (Adamson & Pickard 1986a, Gore 1993). A retreat of probably only some 50–100 m of the edge of Sørsdal Glacier at Chelnok Ice Dam (Fig. 1) would either drain Chelnok Lake completely or reduce its size. In both cases Tierney Creek would disappear. Hence, when an ordinary year such as 1990–91 is considered, it may be noted that, during the time corresponding to *phases I* and *II* (Fig. 4), there would only be a net inflow to Crooked Lake of $0.97 - 0.23 + 1.06 - 0.64 = 1.16 \times 10^6$ m³, which is less than the 1.71 $\times 10^6$ m³ required to fill the lake to its outlet threshold. Thus *phases II* and *III* in their present form would not set in. Instead, during part of the summer, the system would degrade to a Druzhby-Pauk system with a substantially reduced flow through Ellis Rapids and to a "binary", internally drained Crooked-Corner system. Such "binary" systems can be found in several places in the Vestfold Hills along the transitional zone between saline and fresh lakes. Crooked Lake would be lowered by the excessive evaporation until reaching a new equilibrium between inflow and ablation/evaporation from the lake surface. The time required for this could be calculated if the lake's bathymetry were known. At present (1995), however, no complete bathymetric survey has been made of Crooked Lake. The bathymetric map of nearby Ellis Fjord (Gallagher & Burton 1988), shows that large parts of its floor lie at the lower end of its depth range of 0–117 m b.s.l. The greatest depth recorded in Crooked Lake is 140 m (Adamson & Pickard 1986a, Pickard 1986a). Thus, if the lake is tentatively approximated by a paraboloid, 140 m deep, a calculation shows that equilibrium would occur in about 830 years.

A sediment core taken at a depth of 38.33 m near the northern shore of the lake (Fig. 1) yielded 593 mm of laminated fine-grained sediments overlying till. The organic content was too low to permit any radiocarbon datings. Instead, if the laminae are regarded as varves and are counted, an age of about 370 years BP is obtained as the beginning of sedimentation at this site (author, unpublished

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data). This age can be regarded as a minimum age, since years of low water inflow must be poorly represented or completely absent in the sedimentary record. This sedimentary record is shorter than the records from cores taken in Pauk Lake and Bisernoye Lake, situated closer to the ice-margin than Crooked Lake (author, unpublished data). One reason for that could be that the scenario described above has actually taken place some time during the Holocene. A minor retreat of Sørsdal Glacier would then have drained Chelnok Lake with a lowered Crooked Lake as a consequence. Any fine-grained sediments would then be exposed to wind-erosion and removed. This in turn would have reset the sedimentary clock. A subsequent readvance would then have reversed the process, eventually leading to the present state of the Druzhby system. This course of events could fit into the scenario described by Fitzsimons (1991), who suggested an early Holocene retreat of the margin of Sørsdal Glacier beyond its present position, followed by an advance. However, it could be argued that the short sedimentary record from Crooked Lake is a result of the Chelnok Glaciation, a late Holocene ice advance suggested by Adamson & Pickard (1983, 1986a,b) on the basis of glacial striae (Fig. 1). That advance would have overrun most of the area now occupied by Crooked Lake. At present, the margins of the ice sheet and Sørsdal Glacier are cold-based. Ice, frozen to its bed, has limited power to make striae. Instead of grinding the bed, the ice sticks to it when advancing. This argues against the existence of the Chelnok Glaciation. In addition to that, the direction of the striae, 005°, make it very hard to give a glacio-dynamic explanation. A more north-westerly direction would fit better with the Chelnok Glaciation hypothesis. The ice could have become wet-based (bed-eroding) in either of two ways or due to both of them:

1) by thickening enough to attain basal melting, or

2) by an increase of the mean air temperature.

The Chelnok striae could therefore have been created beneath an extended ice sheet during a period warmer than the present. A crude estimate of the ice thickness required to bring about basal melting, given today's annual mean temperature (-10.2 °C) is ~400 m (Paterson 1981). A rapid retreat of Sørsdal Glacier due to marine flooding could have made part of the ice sheet drain in a south–south-westerly direction into the Sørsdal trough, creating the Chelnok striae. This agrees with the results of Lundqvist (1989), Colhoun (1991), Fitzsimons (1991), Hirvas & Nenonen (1991), Hirvas *et al.* (1993) and Hirvas *et al.* (1994). A careful examination of the striae may corroborate this hypothesis.

In contrast to the Druzhby system, the Zvezda system (Fig. 1), second largest in this area, is not likely to respond significantly to a moderate ice retreat or climatic change, because it is not ice-dammed. However, "Corner Lakes" (Gore 1991, 1992) display, in a smaller spatial and temporal scale, a similar sensitive dependence on occasional water inflow, snow/ice retreat and evaporation as the entire Druzhby system. In fact, conditions of this kind are prevalent along the ice-proximate parts of level, ice-free coastal areas in East Antarctica, such as Bunger Hills (Wisniewski 1983, Adamson & Pickard 1986b, Colhoun & Adamson 1989, Kaup *et al.* 1993) and Schirmacher Oasis (Richter 1983, Loopmann & Klokov 1986).

Climatic fluctuations

At present, ice recession in the Vestfold Hills occurs primarily by sublimation and evaporation (Table III) and the amounts of meltwater released are relatively small (Bronge 1989, Colhoun 1991). The almost total absence of glaciofluvial deposits in the area and the pristine condition of tills (Lundqvist 1989, Colhoun 1991, Hirvas & Nenonen 1991, Hirvas, *et al.* 1993, Hirvas *et al.* 1994) strongly indicate that the Holocene deglaciation was of this "dry" nature, and not of a West Greenland type with large amounts of meltwater in action.

The ablation from the blue ice areas bordering the Vestfold Hills increases when insolation is high and is impeded under overcast conditions (Bronge 1989). This means that ablation is dependent on the latitude of the cyclonic tracks. When the cyclonic tracks migrate northwards, coastal Antarctica obtains a more continental, colder climate. Increased ablation and less accumulation of snow and superimposed ice may then make the ice edge retreat, and even increase the summer runoff to the freshwater drainage systems. A more maritime climate, influenced by the migrating lows, gives higher temperatures, increased accumulation of snow and ice (Jouzel *et al.* 1989) and a subdued runoff (Bronge 1989). Due to the increased snow accumulation, the level of the internally draining lakes will then rise. This has actually been observed in the Vestfold Hills during recent decades (H. Burton, personal communication, September 1990), and the same thing has been recorded from the lakes of the Dry Valleys (Chinn 1987b, 1993). Similarly, accumulation of snow and superimposed ice has increased in Wilkes Land (Morgan *et al.* 1991) and on James Ross Island (Peel 1992), but not on Ritscherflya, Dronning Maud Land (Isaksson *et al.* 1996). Furthermore, Gore (1993) recorded snow and ice fields in the southeastern Vestfold Hills consistently increasing in size since 1947.

According to the discussion above, the present state of the Druzhby system could be described as follows:

A southerly migration of the cyclonic tracks during the latest decades have caused lakes in the western Vestfold Hills to rise and the summer runoff from the ice sheet and the Sørsdal Glacier to decrease. Oxygen isotope studies of the Vostok ice core have revealed that accumulation in the interior of Antarctica has been higher during warmer stages, like the Eemian, than during glacials (Jouzel *et al.* 1989). This kind of change will be reflected in the behaviour of the marginal parts of the ice sheet, though with a considerable time-lag. The response time of Sørsdal Glacier is not known, but the glacier's present state of slow retreat (Adamson & Pickard 1986a, Gore 1993) was certainly induced during a previous period, colder than now. Such periods during the latest two millenia were inferred from lake sediment cores of Nicholson Lake (Bronge 1992) and other freshwater lakes in the Vestfold Hills (author, unpublished data) and from ice cores of Law Dome (Morgan 1985).

Thus, a change of climate to summers with predominantly overcast weather, leading to less melting of ice and snow, would cause an annual runoff deficit for Crooked Lake. Such summers do occur within the range of the present climate, but not frequently enough not to be compensated for by the excessive melt during ordinary to "warm" summers. An increase of ice melt, on the other hand, would not in a short perspective change the characteristics of the drainage system, only prolong the duration of *phase II* at the expense of *phases I* and *III*. In a longer perspective though, increased melt can make Sørsdal Glacier retreat, inducing the course of events discussed above. Statistical examination of additional, consecutive hydrologic and climatic data from Vestfold Hills might allow determination of the magnitude of climate change necessary to make Crooked Lake an internally drained basin once again. To compensate for the loss of the Tierney Creek flow, an increase of 47% of the water derived from melt would be necessary. In that case Crooked Lake would start spilling over at the very beginning of *phase III*.

Conclusions

The annual evaporation from open water was found to be 928 mm which is consistent with earlier results. Most of the ablation from the ice sheet and Sørsdal Glacier occurred by sublimation. During the melt season though, about 1/3 of the ablation was melt, which took place predominantly in ice lying below 180 m a.s.l. The hydrology of the area is characterized by excessive evaporation. During 1990–91 the Druzhby drainage system went through four phases of development, which will recur each year with variable relative duration. The water balance reveals that the hydrography of the area is quasi-stable. Since the lakes are lowered below their outlet thresholds due to winter ablation, they must be refilled by the small affluents before the system becomes externally draining for the season. The system is sensitive to disturbances in water supply, which may be caused by a retreat of Sørsdal Glacier, a change of climate or both. The Druzhby system could readily be deprived of half its inflow of water by an emptying of ice-dammed Chelnok Lake to Crooked Fjord. This would cause the Druzhby system to split into an externally drained Druzhby-Pauk system and an internally drained Crooked-Corner system. A sediment core from Crooked Lake indicates that sedimentation at 38.3 m depth started at least 370 BP. The idea of a recent (<3000 BP) Chelnok Glaciation overriding most of Crooked Lake and creating the Chelnok striae (005°) can be questioned. The reasons for that are twofold:

1) it is difficult to explain in terms of glacier dynamics,

2) to make an extensive system of striae, the ice must have been thick and/or warm enough to melt underneath.

The striae are explained here as being older and the result of local bottom-melting of a thick ice sheet, further advanced than now, which began to drain south–south-westwards into the Sørsdal trough, opened up by a rising sea-level. The thin sediments of Crooked Lake can, in accordance with this hypothesis, be explained by a recent partial drying up of that lake due to loss of water supply following a collapse of Chelnok Ice Dam. The solution of these problems requires a glaciological programme, including mass-balance and ice flow measurements for Sørsdal Glacier, a hydrologic-climatologic programme for the Druzhby system and its subsystems, and, a careful examination of the Chelnok striae. Climatic data from the Vestfold Hills and other Antarctic areas show that the hydrography of freshwater drainage systems like the Druzhby system is sensitive to short and long-term climatic fluctuations. Sometimes the influences from them amplify each other, sometimes they counteract each other. At present, runoff from the ice seems to be impeded by less insolation due to an increased influence from the migrating cyclonic lows, whereas the ice edge is under slow retreat due to decreased accumulation further inland during a previous colder period of more continental climate.

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