known for its high amount of annual snow precipitation due to the blocking action of the south face of Ross Island to moist air masses from the south.

Aurora Glacier, descending from the south-east face of Mount Erebus, and Terror Glacier, descending from the south-west face of Mount Terror, cause additional thickening of the ice shelf near the coast, as shown in Figure 1a. No radar data are available from the area of the "rollers", which extend from Cape Crozier south-westwards. Further RES data on the thickness of the Ross Ice Shelf to the south-east of Ross Island have been published in Stern and others (1990).

A comparison of Figure 1a and b reveals considerable, but systematic, differences in ice thickness on a local scale. Most noticeable, the terminus of Terror Glacier appears to have receded east-north-eastwards by about 5 km. The differences in ice thickness were mapped. The estimated ice-thickness difference at each measuring point of the 1985 survey — the corresponding ice thickness in the early 1960s was interpolated from Figure 1b — is indicated in Figure 1c.

The apparent discrepancy between both data sets cannot be attributed to systematic positioning errors of points of measurement. Whatever systematic errors one might assume, no reasonable scheme can be evoked that might explain the considerable and systematic differences in measured ice thickness. In the light of the limited amount of information available from the radar measurements in the 1960s, the recently discovered discrepancy between both data sets should be viewed with caution. On the other hand, the alternative — substantial thinning of the ice shelf in Windless Bight within the last 25 years — should also be considered. What effects could possibly cause such a change?

Strong reflections can potentially be produced by brine-soaked layers near the base of the ice-absorbing EM waves attempting to travel to deeper levels. No direct evidence to this effect is available. Brine-soaking has been shown to exist at the western edge of Windless Bight (Stuart and Bull, 1963; Risk and Hochstein, 1967). Detailed studies on brine-infiltration mechanisms and the limit of brine infiltration along the McMurdo Ice Shelf — as the front part of the Ross Ice Shelf near McMurdo is also called — have been published by Kovacs and Gow (1975), Kovacs and others (1982), and Cragin and others (1983). Their results clearly indicate brine-soaking to be confined to areas west of Windless Bight. Brine-soaking, therefore, is very unlikely to be the underlying cause of the discrepancy between the above data sets.

Bottom melting: Pillsbury and Jacobs (1985) gave an average value of 0.3 m a^{-1} for the basal melting rate of the Ross Ice Shelf, Risk and Hochstein (1967) a rate of ~1 m a^{-1} for the ice shelf near Windless Bight. Jacobs and others (1981) estimated the melting rate of the floating Erebus Glacier ice tongue to be in the range of 0.3–3.0 m a^{-1}. An assumed melting rate of 3.5 m a^{-1} (80 m in 23 years; 1962–85) at the under-side of the floating terminus of Terror Glacier appears to be at the upper limit of the range of acceptable values. If such a high amount of bottom melting occurs today, then one has to assume — at least on a local scale — a considerable build-up in ice thickness in the recent past.

Recent slow-down of glacier movement: the terminus of Terror Glacier has receded. The downward velocity of the glacier has possibly been reduced by decreased snow accumulation in the source area in recent times, such that bottom melting at the glacier terminus currently exceeds the supply rate.

Recent surge event: heat flow from the interior of an active volcano is potentially quite variable with time. An increase in heat flow (e.g. caused by a high-level intrusion into the flank of the volcano) has conceivably changed in recent times the thermal conditions at one or both glacier beds, causing a minor surge event and increased glacier velocities over a restricted period of time in the recent past. At the same time, more than the usual amount of ice would be supplied to the ice shelf. However, such an effect is more likely for Aurora Glacier descending from Mount Erebus (Mount Terror appears to be extinct). After cessation of the surge, bottom-ice melting would take over until a new balance between supply and melting rate is reached.

Only a continuing systematic monitoring effort in the near future might unequivocably resolve the question, whether the ice thickness in Windless Bight currently experiences significant changes or not.

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SIR,

Vapor-pressure dependence on temperature in models of snow metamorphism

The modelling of heat and vapor flows through snow continues to be of interest in work on snow metamorphism and heat transfer. The effect of temperature on the vapor pressure of ice is of interest in several fields and it is worth reviewing how vapor pressure is approximated, and in examining some of the consequences of those approximations. While most are good approximations to the vapor pressure, they are not necessarily good approximations to its derivatives.

The flux of vapor (J) in the vertical direction (z) due to molecular diffusion depends on the density gradient, or

$$J = -D\mathrm{d}\rho/\mathrm{d}z \qquad (1)$$

where D is the diffusion coefficient and vapor density (ρ) is given by the ideal gas law,

$$p = \rho RT \qquad (2)$$

where p is vapor pressure, R is the gas constant for water vapor, and T is the absolute temperature. Given that the temperature gradient is conveniently measured, the flux is usually expressed with $(\mathrm{d}\rho/\mathrm{d}T)(\mathrm{d}T/\mathrm{d}z)$ where ρ is obtained from the ideal gas law and the Clausius–Clapeyron equation,

$$\mathrm{d}p/\mathrm{d}T = L/(v - V)T \qquad (3)$$

where L is the latent heat of sublimation, v is the specific volume of the gas, and V is the specific volume of the solid. When continuity is used to balance the rate of condensation with the gradient of flux, the second derivative is taken (i.e. Palm and Tveitereid, 1979). Given that the gradients of approximations are used, it is necessary to establish the accuracy of the gradients, apart from the accuracy of the approximations themselves.

Washburn (1924) derived the dependence of vapor pressure on temperature by integrating the Clausius–Clapeyron equation using the latent heat at 0°C, the heat capacities of ice and water vapor at 0°C, and measured values of p at 0°, –50°, and –100°C. He rejected the measured values between 0° and –50°C, and stated that the following equation was more accurate:

$$\log_{10}p = -2446/T + 8.231\log_{10}T - 1677E - 5T + 1.205E - 5T^2 - 6.757 \qquad (4)$$

where p is in mm of mercury and T is in K. This formula was used by Dorsey (1940) and by the *International critical tables* (Washburn, 1928) as the standard, and is accepted here as the basis for comparison. Only de Quervain (1963) has used it in snow studies.

A common approximation to this formula can be derived (e.g. Colbeck, 1980) by making the further assumption that the specific volume of water vapor (v) is much greater than the specific volume of ice, then

$$p = p_0\exp(L(T - T_0)/RTT_0) \qquad (5)$$

where p_0 and T_0 are usually taken as the triple point. Giddings and LaChapelle (1962) used the Clausius–Clapeyron equation and an approximation to Equation (5), while Colbeck (1980) used Equation (5) as well as the analogous relationships between pressure and curvature, and temperature and curvature. Taking L as 2838 J g^{-1} (Rossini and others, 1952), R as 0.4619 J g^{-1} K^{-1}, and p_0 as 610.5 Pa at 273.1 K, Washburn's equation can be approximated by Equation (5) with great accuracy. Even the second derivative of this function is extremely close to Washburn's as shown in Figure 1. Thus, Equation (5) is a very close approximation to Washburn's formula for nearly all applications.

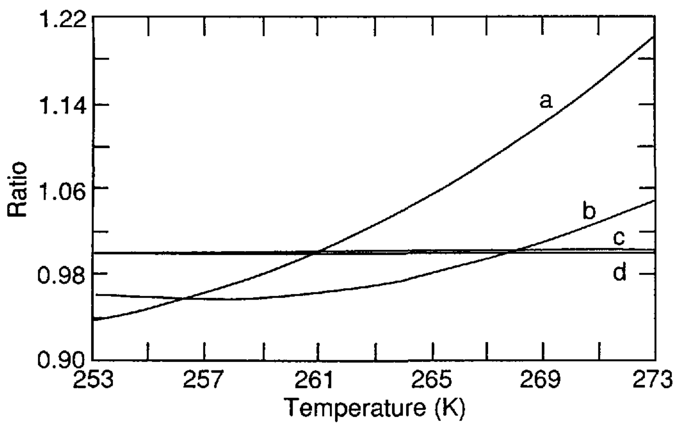

Fig. 1. Ratios of the derivatives versus temperature. (a) Second derivative: Yosida's over Washburn's equation. (b) First derivative: Yosida's over Washburn's equation. (c) and (d) First and second derivatives: Equation (5) over Washburn's equation.

Yosida (1950) suggested and many have used an approximation to Equation (5). This is derived by setting T in the numerator of the exponential of Equation (5) equal to an average value in the temperature range of interest. The relationship is then simplified to

$$p = p_0\exp(c(T - T_0)) \qquad (6)$$

where c has generally been taken as 0.0857 K^{-1} and p_0 as 611 or 642 Pa (Yen, 1962; de Quervain, 1973; Perla, 1978; Palm and Tveitereid, 1979; Sommerfeld, 1983; Powers and others, 1985). Yen (1962) has even reduced this to a linear approximation and used its derivative, which is in error by more than 200% in his range of temperatures. While the correlation coefficient between Yosida's and Washburn's expressions is very high, the derivatives are of most interest and they are not so well correlated. As shown in Figure 1, the second derivative is off by 20% in this range of temperatures and would be further in error if a wider range of temperatures was of interest, in polar firn for example. The first derivative is used more frequently and incurs only a 5% error in this temperature range. Thus, for most problems in seasonal snow, Yosida's approximation is adequate, whereas for studies using the second derivative the approximation is questionable. A considerable simplification in the mathematics would have to be achieved before Yosida's approximation could be justified, whereas Yen's linearization of the vapor-pressure–temperature relationship is highly suspect, even when it allows analytical solutions.

In summary, Equation (5) is a very good approximation to Washburn's equation and can also be used to represent its derivatives. Yosida's approximation continues to receive widespread use because it is also a good approximation, but its most common use is to represent the gradients, not the vapor-pressure–temperature relationship itself. In this regard, its use should be restricted to the first derivative over a narrow range of temperatures. Use of a linear approximation is always discouraged.

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SIR,

Extraordinary melt-water run-off near Søndre Strømfjord, West Greenland

During late January 1990, two melt-water rivers started to flow from the western margin of the Greenland ice sheet into Søndre Strømfjord despite air temperatures of below −30°C. Description of this unusual event is based upon local observations made by S. Malmquist (personal communication, May 1990).

The Ørkendalen river started flowing in the last week of January at one-third of its normal summer level (Fig. 1). One week later, melt water started to flow from Sandflugtsdalen river at approximately one-quarter of its normal summer level (Fig. 1). Both rivers continued to flow for a further 3 weeks during which air temperatures were consistently below −30°C. Discharge from these rivers over-ran the ice-covered fjord for a distance of 10 km (Fig. 1). A heavy freezing fog resulted from the exposure of relatively warm river water to sub-zero air temperatures. The fog was observed leading from the ice margin along the river channels towards the fjord by overflying trans-Atlantic aircraft.

From the above information, it was possible to quantify the volume of water involved in this event. Based on estimated "normal" summer discharges totalling $140\,m^3\,s^{-1}$ for Ørkendalen and Sandflugtsdalen rivers, an estimated $90 \times 10^6\,m^3$ of water were involved in this event. This figure probably underestimates the total volume of water drained but provides an approximation on which discussion can be based.

Until now, river flows have only been documented within the normal summer melt season (*c.* mid-May–*c.* mid-October). Not only are the flows described above outwith the usual period but they are in excess of those witnessed by the author in late October 1986 and early June 1987. As such, this event probably represents the release of stored melt water, as it cannot represent ice-surface ablation given the sub-zero temperatures.

Possible sources of stored melt water along this section of the ice-sheet margin include ice-dammed lakes and englacial or subglacial reservoirs. Ice-dammed lake drainage, although common within this region (Sugden and others, 1985; Russell, 1989; Russell and others, 1990), is unlikely to have resulted in this unseasonal outburst as there does not appear to be a suitably large lake located between Ørkendalen and Sandflugtsdalen melt-water streams (Fig. 1). The total volume of water drained during this event is 2.5 times that drained from an ice-dammed lake (Russell, 1989) and 300 times that noted by Russell and others (1990) for a small ice-dammed lake. Although this ice margin is likely to have been frozen to the bed during the winter months, sub- and/or englacial melt water originating at great distances from the ice margin may have still been travelling towards the ice margin down an equipotential gradient (Shreve, 1972). On meeting the cold, impermeable ice margin, melt water may have been stored as a sub- or englacial reservoir under considerable pressure. The release of water within such a reservoir is likely to have been maintained by high water pressures. The drainage of such a sub- or englacial reservoir located at a considerable distance from the ice margin may provide an explanation for the unusual events noted in Søndre Strømfjord in January and February 1990.

Although similar events may occur, unnoticed, during the summer melt flows, these winter discharge events may have important geomorphological effects upon the glacier margin and the pro-glacial river channels. The volume of water released in this event was far greater than that so far noted for any of the ice-dammed lakes within this area, constituting a significant part of run-off from this section of the ice margin.

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SIR,

Comments on: "6000-year climate records in an ice core from the Høghetta ice dome in northern Spitsbergen"

Fujii and others (1990) have recently presented an estimate of climatic conditions in northern Svalbard during the last 6000 years, based on their interpretation of an 85.6 m long ice

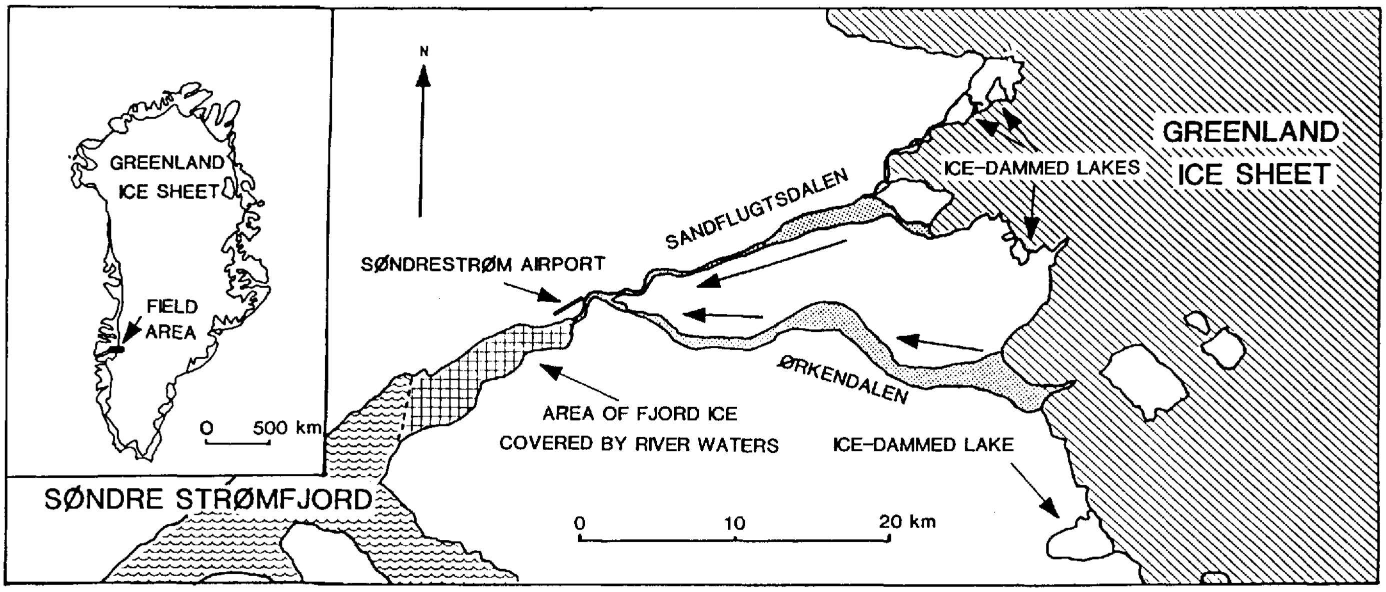

Fig. 1. Location map showing the melt-water routeways into Søndre Strømfjord, West Greenland.